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VOLCANISM AND TECTONICS ON VENUS

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ABSTRACT

We review recent developments in the study of volcanism and tectonics on Venus. Venus's crust is basaltic, dry, and probably about 30 km thick. The mantle convects, giving rise to plumes, and has a similar composition and mean temperature ($\sim$1300°C), but a higher viscosity ($\sim 10^{20}$ Pa s), than that of the Earth. Inferred melt generation rates constrain the lithospheric thickness to between 80 and 200 km. The elastic thickness of the lithosphere is about 30 km on average. The present-day lack of plate tectonics may be due to strong faults and the high viscosity of the mantle. Most of the differences between Earth and Venus processes can be explained by the absence of water.

Venus underwent a global resurfacing event 300–600 Ma ago, the cause and nature of which remains uncertain. The present-day surface heat flux on Venus is about half the likely radiogenic heat generation rate, which suggests that Venus has been heating up since the resurfacing event.

INTRODUCTION

Over five years ago, the NASA spacecraft Magellan began sending back information from Venus (Saunders et al 1992). During this time, a great deal of work concerning crustal and tectonic processes on the planet has been produced, much of which has been reviewed elsewhere (Phillips & Hansen 1994, Kaula 1994, Bindschadler 1995). The aim of this article is to synthesize much of the recent work and to show that a consistent description of present-day geological processes on Venus can be produced. Parts of this description are not in dispute; others, however, are still controversial.

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The main part of the paper discusses our understanding of present-day processes on Venus, starting at the surface and progressing deeper into the planet. Our contention is that, thanks in particular to the short-wavelength gravity measurements provided by Magellan, the gross tectonic attributes of Venus (lithospheric thickness, mean mantle temperature, and viscosity) are reasonably well established. The second section examines what little is known about how Venus has evolved and compares it with our understanding of the Earth's evolution. The final part summarizes our conclusions and suggests outstanding issues that have yet to be resolved.

PRESENT-DAY GEOLOGICAL PROCESSES

Geological investigations of Venus are based on four main data sets: synthetic-aperture radar (SAR) images of the surface (Saunders et al 1992), surface altimetry (Ford & Pettengill 1992), compositional measurements of the surface from Soviet landers (Surkov 1977, Surkov et al 1983, Basilevsky et al 1986), and the gravity field measured at spacecraft altitude (Konopliv & Sjogren 1994). Of these, only the last provides a direct measurement of processes happening at depth.

This section begins by discussing surface observations and the inferences that can be made from them. We then examine inferences made about processes happening within the crust, the mantle, and the core of Venus.

Surface Observations and Processes

As is well known, surface conditions on Venus are hostile: The temperature is about 450°C, and the atmosphere is thick (about 90 atmospheres), mainly composed of CO_2, and contains only 1–100 ppm water (Prinn & Fegley 1987). The atmospheric deuterium:hydrogen (D:H) ratio, which is about 120 times the terrestrial value, has been used to suggest that Venus originally possessed at least a few tenths of a percent of a terrestrial ocean, which evaporated, so that the H_2 was lost to space (Donahue & Hodges 1992).

On the basis of the Soviet geochemical measurements, the surface composition of Venus appears to be essentially basaltic (McKenzie et al 1992a). This observation is supported by SAR images that show many volcanic flows similar to basaltic flows seen on Earth. However, judging by the morphology of some flows imaged, there are lavas that are considerably more viscous than typical terrestrial basalts (Pavri et al 1992, McKenzie et al 1992b). On Earth, the composition of melt is controlled by the temperature of the mantle that is melting. Because the mantle is compressible, its actual temperature depends on pressure. It is therefore convenient to use a potential, rather then a real, temperature in melting calculations. The potential temperature T_p of the mantle is defined as the temperature it would have if the pressure were changed from its present value to 0 GPa isentropically, without allowing the material to melt. An important conclusion, made from FeO percentages from the Venera and Vega landers, is that the potential temperature of the mantle from which the sampled melts formed is estimated to be similar to that of present-day mantle potential temperatures on Earth, i.e. about 1300°C (McKenzie et al 1992a). Although the basalts may be up to 300–600 Ma old, Nimmo & McKenzie (1997) show that the mantle temperature is unlikely to have increased in temperature by more than 200°C since then. Hence, 1500°C is a reasonable upper limit on the present-day mantle potential temperature.

Unlike the Earth's topography, the topography of Venus is unimodal: 80% of the planet lies within ±1 km of the mean planetary radius (MPR) (Pettengill et al 1980). In the frequency domain, topography at wavelengths longer than 2000 km is strongly correlated with the gravity, which is again unlike that of Earth (Phillips & Lambeck 1980).

The global SAR images allowed two particularly important conclusions to be reached:

1. The impact crater record, which gives a mean production age of 300–600 Ma, is statistically uniform over the entire surface of the planet in contrast to that of the Earth. Moreover, no more than 10% of the craters are volcanically embayed from the outside (Herrick & Phillips 1994, Strom et al 1994). This uniformity may be due to gradual resurfacing (Phillips et al 1992) but is more likely due to a global resurfacing event ending at 300–600 Ma, with only minor resurfacing thereafter (Schaber et al 1992, Bullock et al 1993, Strom et al 1994, Strom et al 1995). The time over which the resurfacing took place was short, probably less than 100 Ma (Strom et al 1994, Namiki & Solomon 1994). There may, however, be subtle differences in age between different tectonic provinces (e.g. Herrick 1994, Herrick & Phillips 1994, Phillips & Izenberg 1995), e.g. tesserae are older (Ivanov & Basilevsky 1993) and large volcanic structures are younger (Basilevsky 1993, Namiki & Solomon 1994, Price et al 1996) than average.

2. In general, neither SAR images nor topography measurements revealed features that were similar to terrestrial expressions of plate tectonics. The high-resolution Magellan images showed no signs of features similar to spreading ridges (McKenzie et al 1992a, Solomon et al 1992) and only rare examples of possible fracture zones and abyssal hills (McKenzie et al 1992a). Arcuate trenches known as chasmata that appear to be similar to trenches at terrestrial subduction zones are seen, notably in the Aphrodite region (Schubert & Sandwell 1995). Whether localized subduction is occurring at these locations is not clear (Sandwell & Schubert 1992a, Solomon et al 1992, Hansen

& Phillips 1993, Brown & Grimm 1995). Volcanic features (see below) are not clustered in the linear bands that are characteristic of plate boundaries on Earth but are broadly distributed (Head et al 1992). The same is true of plains deformation, which can be coherent over thousands of square kilometers (Solomon et al 1992). The lack of linear bands of volcanism and deformation again suggests that plate tectonics is not currently operating. Rift valleys similar to terrestrial ones are apparent (McGill et al 1981, Foster & Nimmo 1996), but the rifting appears to be related to convective upwelling (e.g. McGill et al 1981) rather than plate motions. These observations, together with the spatial uniformity of the crater record, suggest that creation and destruction of the surface on a 1000-km scale has not been a global process over the last 300–600 Ma.

The SAR images also allowed identification of a range of volcanic and volcano-tectonic features, some of which have clear Earth analogues and others of which do not [for a review see Head et al (1992)]. In the former category, there are volcanoes (some up to 1000 km in diameter), dike swarms that in some cases radiate for thousands of kilometers (McKenzie et al 1992c, Ernst et al 1995), and volcanic flows that range in area up to values similar to those of terrestrial flood-basalt provinces (Lancaster et al 1995). Some of the flows, termed festoons (Head et al 1992), show lobate flows that are ridged, which suggests an unusually high viscosity.

Features that lack terrestrial analogues include coronae (and the possibly related arachnoids and novae), tesserae, and channels. Coronae (and the associated objects) are circular features 60–2000 km in diameter, which generally consist of concentric and radial fractures arranged about a central depression (Stofan et al 1992, Squyres et al 1992a). Tesserae are elevated plateau areas that are characterized by very high reflectivity and intense deformation, both compressional and extensional (Solomon et al 1992, Bindschadler et al 1992a). Channels are meandering troughs that appear to be similar in morphology to terrestrial rivers, though they lack tributaries. They are usually no more than 5 km wide but can be up to 7000 km long (Baker et al 1992), and they are likely to have been formed by a low viscosity fluid, probably basalt.

The SAR-imaging system showed many examples of dark or shadowed planes that were interpreted to be either fault planes or talus slopes. Estimates of the total amount of extension across rift zones can be made by making assumptions about the typical angles of the planes that are imaged (Ford et al 1993, Connors 1995). The total width of dark planes across both Ganis Chasma (Atla Regio) and Devana Chasma (Beta Regio) is about 30 km, so, if the rifts are assumed to be symmetrical, the likely amount of extension is from 5–33 km, with the most likely value about 15 km. Because the rifts are at least 150 km across, the stretching factor, β (the ratio of final width to initial width), is about 1.2 or less. Similar conclusions have been reached by Magee & Head (1995).

In summary, from the point of view of investigating crustal and mantle processes, the most important surficial observations are

- surface temperature of 450°C;
- basaltic crustal composition, from a mantle of potential temperature of ~1300°C;
- absence of water at the surface;
- approximately uniform surface age of 300–600 Ma, which suggests a global resurfacing event ended at that time;
- absence of plate tectonics;
- rather small amounts of crustal extension across rifts.

Crustal Processes

The crust of Venus is generally accepted to be basaltic in composition. The thickness of the crust, however, is not well known. Estimates of crustal thickness prior to 1995 were often based on an unrealistically weak rheology (see below). Estimates of crustal thickness using gravity are sensitive to assumptions about crustal density. They produce estimates of about 20–50 km at plume sites (Phillips 1994, Smrekar 1994), 20–40 km in the plains (Konopliv & Sjogren 1994, Grimm 1994a), 35–60 km around tessera sites (Kucinskas & Turcotte 1994, Simons et al 1994, Grimm 1994a), ~30 km in unrifted material around Devana Chasma (McKenzie 1994), and about 25 km on average (Simons et al 1994, Konopliv & Sjogren 1994). Tesserae are thought to be high-standing because they have a thicker crust, though it is not clear whether this crustal thickening is due to upwelling and magmatism (Grimm & Phillips 1991, Phillips & Hansen 1994) or downwelling and compression (Bindschadler & Parmentier 1990, Lenardic et al 1991, Bindschadler et al 1992b). At depths of 30–70 km, depending on temperature, basalt transforms to much denser eclogite (Ito & Kennedy 1971, Ringwood 1975). This transition is important because it limits the topography that can be supported by isostasy (Namiki & Solomon 1993, Jull & Arkani-Hamed 1995), provides an additional driving force for recycling crustal material to the mantle, and probably provides an upper limit on crustal thickness. The exact P-T boundaries and reaction rate of the transition are not well known.

One of the most important results of the last few years has been the realization that very dry basaltic material may be much stronger than was hitherto thought. Mackwell et al (1995) showed that dry diabase is closer in strength to dry olivine than to diabase that has been less thoroughly dried. This stronger rheology is particularly important when investigating the conditions under which lower crustal flow may occur.

The strength of the lower crust is important because it controls coupling between crust and mantle (e.g. Buck 1992) and may also reduce local crustal thickness variations, thereby limiting the height of regions supported by crustal thickening (Bird 1991). Lower crustal flow is only important if the time scale for such flow to occur is short compared to the average age of the Venusian crust, i.e. ~500 Ma. Lenardic et al (1995) and Kidder & Phillips (1996) both concluded that the Mackwell et al (1995) rheology implied that a time scale on the order of 10^9 years would be required for crustal thickening to take place through lower crustal flow.

Although the thickness of the crust is not well constrained, the same is not true for the apparent elastic thickness (or rigidity) of the lithosphere. Estimates of the elastic thickness (T_e) come from two sources: flexural modeling to match observed topography and comparison between gravity and topography in the frequency domain. Both flexural (Sandwell & Schubert 1992b, Johnson & Sandwell 1994, Brown & Grimm 1996) and spectral (McKenzie 1994, Smrekar 1994, Phillips 1994, McKenzie & Nimmo 1997) estimates produce values of between 10 and 50 km, with most estimates clustering around 30 km. Figure 1 is a plot of the admittance for Ulfrun, compared with that of Hawaii. Both cases show how the short-wavelength ($<$500 km) admittance is well fit by an elastic thickness of around 30 km. Tesserae are generally agreed to be mainly isostatically supported (e.g. Bindschadler et al 1992b, McKenzie 1994, Kucinskas & Turcotte 1994) and hence have an elastic thickness lower than average for Venus, though the noisy tessera topography makes estimation difficult (McKenzie & Nimmo 1997).

Given the elastic thickness of the lithosphere, it is possible to investigate the stresses arising from deformation. For instance, Foster & Nimmo (1996) used a simple model of the stresses required to support topography of an elastic plate to estimate the stresses occurring on faults in Devana Chasma, and they concluded that the faults could withstand stresses of at least 80 MPa, about eight times more than can faults on Earth. The most likely explanation for strong faults is the absence of water, which can reduce fault strength through pore fluid pressure.

In summary, although the crustal thickness is uncertain, it probably averages about 30 km. The elastic thickness is also around 25–30 km, except in tessera areas, where it is probably smaller. Faults within the crust are capable of withstanding stresses much higher than can those on Earth, probably because of the absence of water.

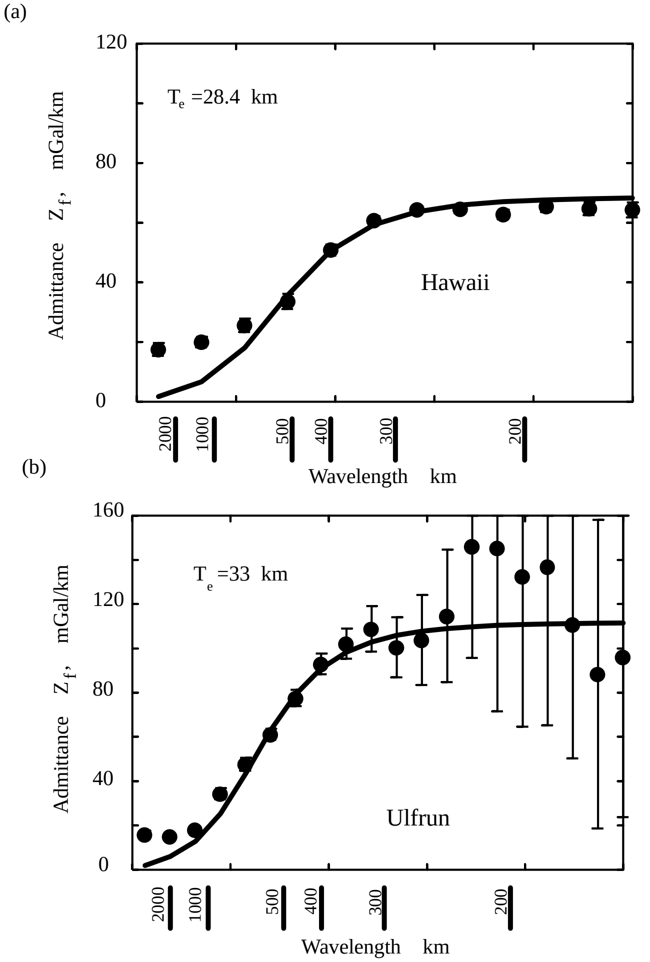


Figure 1 Admittance plots for Ulfrun (−20° to 40°N, −145° to −105°E) and Hawaii from McKenzie & Nimmo (1997). *Dots* are admittance values calculated from observed and calculated two-dimensional gridded line-of-sight acceleration (McKenzie & Fairhead 1997, McKenzie & Nimmo 1997). *Vertical bars* show one standard deviation. The *solid curve* is the theoretical admittance for a specified elastic thickness.

Mantle Processes

The composition of the mantle of Venus is uncertain. However, given that the surface is broadly basaltic, it is likely that the mantle composition is similar to that of Earth. The abundance of radiogenic elements is of particular importance because it determines the internal heat generation rate. Because the abundances of U and Th are similar to those on Earth, and the K/U ratio is also similar, the total abundance of radiogenic elements on Venus is probably about the same as that on Earth (Kaula 1994).

Whether the mantle of Venus was originally wet is not clear (Prinn & Fegley 1987). Because water is incompatible, and Venus shows signs of extensive melt generation, the mantle is likely to be dry unless there is a mechanism that returns water to it (such as the subduction of hydrated oceanic crust on Earth).

On Earth, apart from plate tectonics, the most obvious surface expressions of mantle circulation are areas such as Hawaii and the Cape Verde Islands, beneath which are located rising plumes of mantle material. Such plumes are recognized by a regional uplift of $\sim$1 km over a wavelength of $\sim$1000 km, usually with a volcanic construct and/or rift valley superimposed at the center and a gravity anomaly smaller than that which would occur if the topography were uncompensated. Using these criteria, the pre-Magellan observations were good enough to allow the identification and modeling of several possible plume sites on Venus (McGill et al 1981, Smrekar & Phillips 1991, Kiefer & Hager 1991), many of which have been studied in more detail after Magellan (Bindschadler et al 1992b, Grimm & Phillips 1992, McKenzie 1994, Smrekar 1994, Phillips 1994, Stofan et al 1995, Nimmo & McKenzie 1996). Convective plumes are easy to model experimentally and numerically, so observations made at plume sites can constrain mantle properties, such as lithospheric thickness, strain rate, and heat flux.

PLUME ACTIVITY Gravity data over plume sites give an apparent depth at which compensation occurs of 200 km or more for some areas (Smrekar & Phillips 1991, Grimm & Phillips 1992, Phillips & Hansen 1994, Smrekar 1994). These observations suggest that the gravity anomalies are supported by density contrasts in the mantle, i.e. by ongoing convection. Volcanic activity within the last 30 years has also been suggested as the cause of order-of-magnitude variations in the measured concentrations of atmospheric SO_2 during the various planetary missions (Esposito 1984, Robinson et al 1995). The non-monotonic profiles of the channels on a 1000-km wavelength suggest that the topography of Venus has changed since their emplacement, possibly because of time-dependent convection. The $\sim 0.5^\circ$ offset between the spin and maximum inertia axes may be due to rising plumes within the mantle (Spada et al 1996). Therefore, though plate tectonics is not a significant process on Venus at the moment, the planet nonetheless seems likely to be geologically active.

The dimensionless number that determines the vigor of convection is the Rayleigh number, *Ra*. For reasonable values of heat flux and mantle viscosity on Venus, the Rayleigh number is probably at least 10^7. Although at long ($\sim$3000 km) wavelength, the convection planform as inferred from the gravity consists of circular upwellings and sheet-like downwellings, at shorter wavelengths the planform (i.e. horizontal pattern of convection) becomes more complex (McKenzie 1994). High Rayleigh number convection, such as appears to be occurring on Venus, is typically time dependent, an inference also suggested by the non-monotonic profiles of the channels and the fact that not all sites of volcanic activity are obviously associated with present-day plumes. Given the variation in the wavelength of convective features seen, and the uncertainties in relating wavelength to the depth to which convection extends, it is difficult to use the observed planform to infer whether the mantle on Venus is convecting in one or two layers. Because the temperature structure of the upper thermal boundary layer is the main control on surface gravity and topography (Parsons & Daly 1983), and this structure is independent of depth of convection (McKenzie et al 1974), modeling plume gravity and topography provides almost no constraint on the depth to which convection extends (Nimmo & McKenzie 1996).

LITHOSPHERIC THICKNESS Attempts to model the plume sites referred to above have led to disagreements about the thickness of the lithosphere, with some authors (Turcotte 1993, Kucinskas & Turcotte 1994, Phillips 1994, Moore & Schubert 1995, Solomatov & Moresi 1996) suggesting that it may be 200–400 km thick (though thinned to perhaps 100 km over plumes) and others (Smrekar & Phillips 1991, Smrekar 1994, Schubert et al 1994, Nimmo & McKenzie 1996) suspecting it to be closer to 100 km everywhere. Although some authors have used apparent depths of compensation to infer lithospheric thickness (e.g. Kucinskas & Turcotte 1994, Simons et al 19994, Moore & Schubert 1995), this approach is not always valid. For instance, Robinson et al (1987) and Solomatov & Moresi (1996) have shown that a viscosity varying with depth can make the apparent depth of compensation much smaller or even negative. Similarly, Moresi & Parsons (1995) used a temperature-dependent viscosity to show that a thick conducting lid was not required to explain large apparent depths of compensation.

Some authors (e.g. Phillips 1994) have invoked thinning of the lithosphere due to plume action to account for the regional uplift of up to 3 km observed at plume sites, whereas others (Moresi & Parsons 1995, Nimmo & McKenzie 1996) assume that uplift is dynamic and due to the motion of mantle material. The thinning models require extension factors above the larger plumes, such as Beta and Atla, to be 2 or more to account for the observed uplift. However, from crustal extension estimates, the maximum amount of crustal thinning that has occurred leads to an extension factor of about 1.2 (see above). One way in which total lithospheric thinning could exceed this value is if a ductile lower crust were to have decoupled surface extension from deeper layers, but, as has been shown above, such flow is unlikely even for areas of abnormally thick crust.

As noted above, the likely Rayleigh number for the mantle on Venus suggests that convection is strongly time dependent, with rising and sinking plumes superimposed on a larger scale planform that is also time dependent. If these expectations are correct, the largest convective plumes on Venus may be hard to model because they involve large scale flow with thin thermal boundary layers. Solomatov & Moresi (1996) comment that Beta may well be such a transient phenomenon. Because the effects of a plume of 100-km wavelength would not be visible at the top of a 200-km thick lithosphere, the smallest scale of convective feature visible at the surface provides an upper bound on the lithospheric thickness. Although most attention has been focused on the largest plume sites, such as Beta and Atla, Figure 9 of McKenzie (1994) shows that there are at least half a dozen likely plume sites with widths of 1500 km or less. One of these, in eastern Phoebe, is illustrated in Figure 2. These Hawaii-scale features provide a better constraint on lithospheric thickness than larger ones.

Our belief is that the mechanical boundary layer of Venus is not more than 200 km thick. This view is based on convection models that are required to match inferred melt generation rates, as well as gravity and topography anomalies, of the Hawaii-scale features referred to above (Nimmo & McKenzie 1996). Many of these sites show signs of melt production that are overprinting fractures and are therefore presumably recent; moreover, for the 4–20% resurfacing to have occurred over the last 500 Ma (Bullock et al 1993, Strom et al 1994, 1995), a melt contribution from plumes is probably required. A thick conductive lid inhibits melt production: Figure 3*a* shows that the mantle potential temperature (T_p) must exceed 1700°C to cause melt generation below a mechanical boundary layer that is thicker than 200 km. In general, peak plume temperatures on Earth are not more than 250°C hotter than the background temperature (White & McKenzie 1995). Therefore, even if the background mantle T_p on Venus were 1500°C, it is unlikely that any melt would be generated beneath a mechanical boundary layer thicker than 200 km. Smrekar & Parmentier (1996) have also used melt generation as an additional constraint on mantle properties, and they concluded that a lithospheric thickness of 100–150 km underlaid by a depleted mantle layer was consistent with their observations.

A less robust constraint is given by the values of admittance over plumes. As the mechanical boundary layer thickness increases, so does the admittance. For the observed range of admittance on Venus of 40–60 mgal/km over plume areas (Smrekar 1994, Phillips 1994, McKenzie & Nimmo 1997), Figure 3*b* shows that the mechanical boundary layer is unlikely to be thicker than 300 km.

The inferred rate of melt production also provides a lower bound on lid thickness. Cratering statistics suggest that the average amount of surface melt production over the last 300–600 Ma was no more than 0.15–0.4 km^3 year^{-1} (Bullock et al 1993, Strom et al 1994, Price & Suppe 1994). Because there are

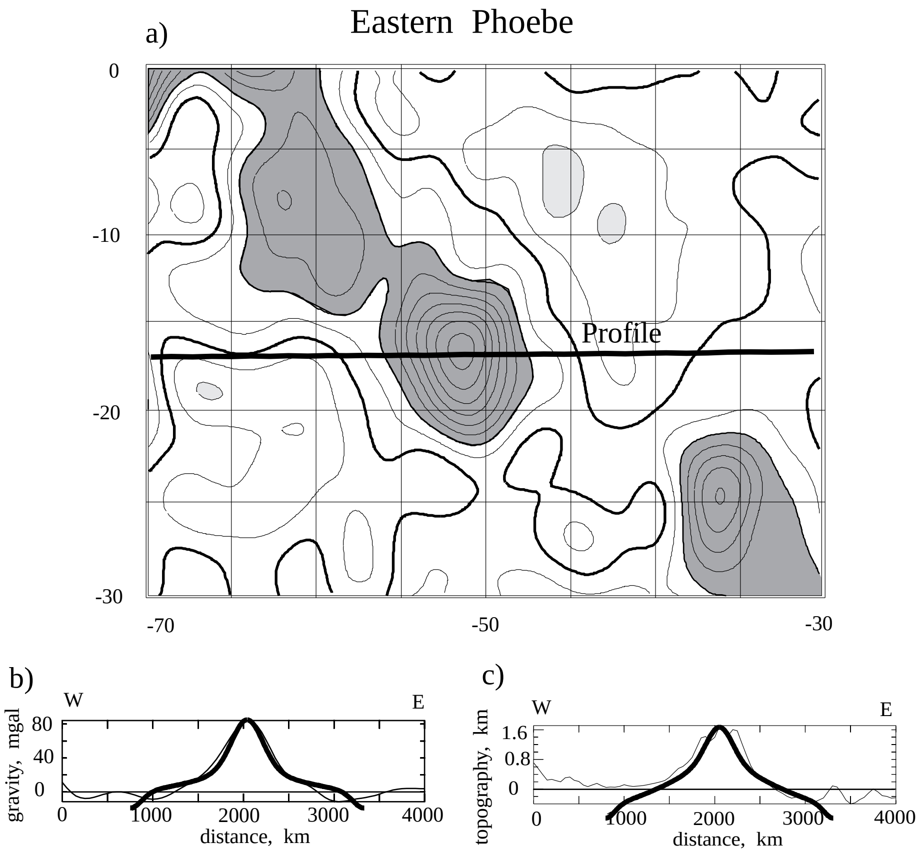
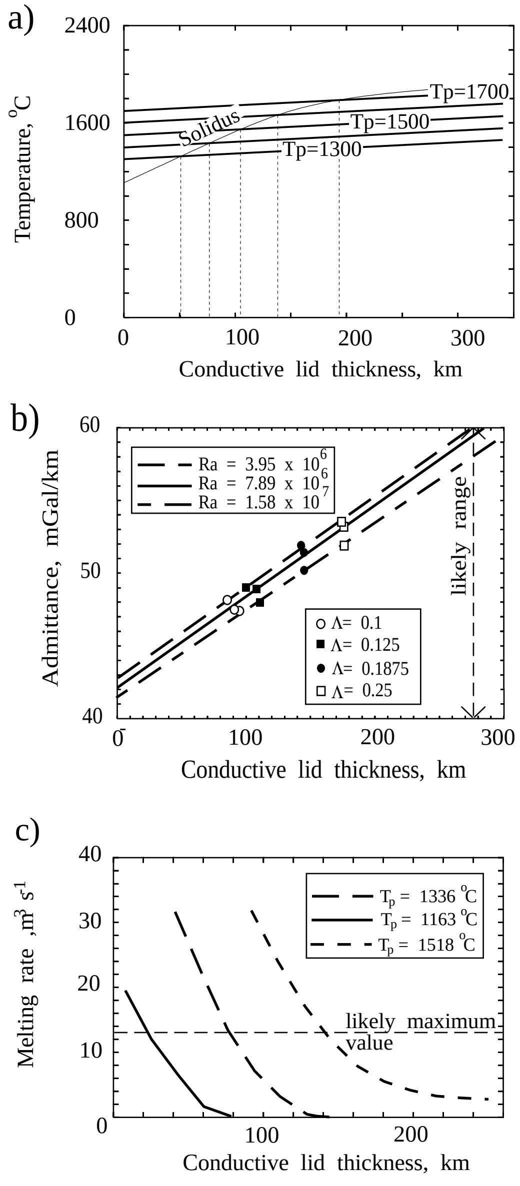


Figure 2 (*a*) Contour map of the gravity anomaly at a height of 70 km. The contour interval is 10 mgal; the zero contour is *bold*; *dark shaded areas* are above +20 mgal, and *light shaded areas* below −20 mgal. The *bold line* shows the location of the profiles, shown as *fine lines* in (*b*) and (*c*). The *heavy profiles* are calculated from the plume model V7 of Nimmo & McKenzie (1996), whose principal difference from V15 is its lid thickness, of 145 km instead of 111 km (modified from Nimmo & McKenzie 1996, figure 5).

at least 10 plumes on Venus at present (McKenzie 1994, Stofan et al 1995), the maximum melt generation rate for each plume is 0.4 km^3 $year^{-1}$ (13 m^3 s^{-1}), even if only 10% of the melt generated reached the surface. Figure 3*c* shows that the mechanical boundary layer cannot be much thinner than about 80 km, unless the mantle T_p is less than 1300°C.

It is possible that lithospheric thinning has taken place beneath plume areas and that the lithospheric thickness elsewhere is greater, as has been suggested by


a)
2400
1600
800
0
Temperature, °C
Tp=1700
Tp=1500
Tp=1300
Solidus
0
100
200
300
Conductive lid thickness, km
b)
60
50
40
Admittance, mGal/km
Ra = 3.95 x 10^6
Ra = 7.89 x 10^6
Ra = 1.58 x 10^7
Λ= 0.1
Λ= 0.125
Λ= 0.1875
Λ= 0.25
likely range
0
100
200
300
Conductive lid thickness, km
c)
40
30
20
10
0
Melting rate, m^3 s^-1
Tp = 1336 °C
Tp = 1163 °C
Tp = 1518 °C
likely maximum value
100
200
Conductive lid thickness, km

Moore & Schubert (1995), Phillips (1994), and Kucinskas & Turcotte (1994). Indeed, Solomatov & Moresi's (1996) temperature-dependent viscosity model produces a stagnant lid that is significantly thinned under the plume axis. However, on Earth, petrological and geophysical observations at Hawaii were fit by Watson & McKenzie (1991) without requiring the mechanical boundary layer to have been thinned. Finally, as we argue above, surface observations of rifts give no indication of extensive ($\beta > 1.2$) thinning. Therefore, the lithospheric thickness of $<$200 km obtained in this section probably can be applied to areas of Venus that do not possess plumes.

Figure 4 is a summary diagram of a convective model from Nimmo & McKenzie (1996), based on fitting the observed gravity and topography anomalies of a Hawaii-scale plume. It shows the lid thickness and the melting zone and compares them with those of Hawaii, from the model of Watson & McKenzie (1991). The radial extent of the melting zone is similar to the Hawaiian model.

MANTLE VISCOSITY Watson & McKenzie (1991) estimated the mantle viscosity beneath Hawaii to be 3×10^{19} Pa s, a similar value to other estimates of viscosity immediately beneath the lithosphere (e.g. Robinson et al 1988). Using a similar model, the values obtained by Nimmo & McKenzie (1996) of around 3×10^{20} Pa s suggest that Venus lacks a low viscosity layer, which is in agreement with other studies (Phillips 1990, Kiefer & Hager 1991, Smrekar & Phillips 1991). The strong correlation between gravity and topography at long wavelengths (Sjogren et al 1980) has also been used to suggest that this low viscosity zone is absent (Kiefer et al 1986), as has the likelihood of strong mantle coupling implied by the coherent plains deformation (Phillips 1986, Squyres et al 1992b). The fact that large apparent depths of compensation are observed also suggests the absence of a low viscosity zone (Phillips 1986), since one

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Figure 3 (*a*) Plot of depth vs mantle temperature. *Thick lines* are the adiabatic geotherms of different mantle potential temperatures, assuming a thermal expansivity of 4×10^{-5}, gravitational acceleration of 8.9 $m^2\ s^{-1}$, and a heat capacity of 1200 J $kg^{-1}\ K^{-1}$. The *thin line* is the peridotite solidus. *Dashed lines* show the minimum mechanical boundary layer thickness required to stop melting for each mantle potential temperature (T_p). (*b*) Admittance as a function of conductive lid thickness. *Dots* represent different models of Nimmo & McKenzie (1996). *Ra* is the Rayleigh number, and λ the dimensionless lid thickness. (*c*) Melt generation rate plotted against conductive lid thickness for a high potential temperature case (*short dashed line*; model V19 of Nimmo & McKenzie 1996), an intermediate case (*long dashed line*; model V15) and a low potential temperature case (*solid line*; model V12). The *dotted lines* indicate the likely range of melt generation rate for Hawaii-scale features (see text). For the high temperature case, $T_p = 1518°C$; for the intermediate case, $T_p = 1336°C$; and for the low temperature case, $T_p = 1163°C$, where T_p is the potential temperature at the center of the convecting region.

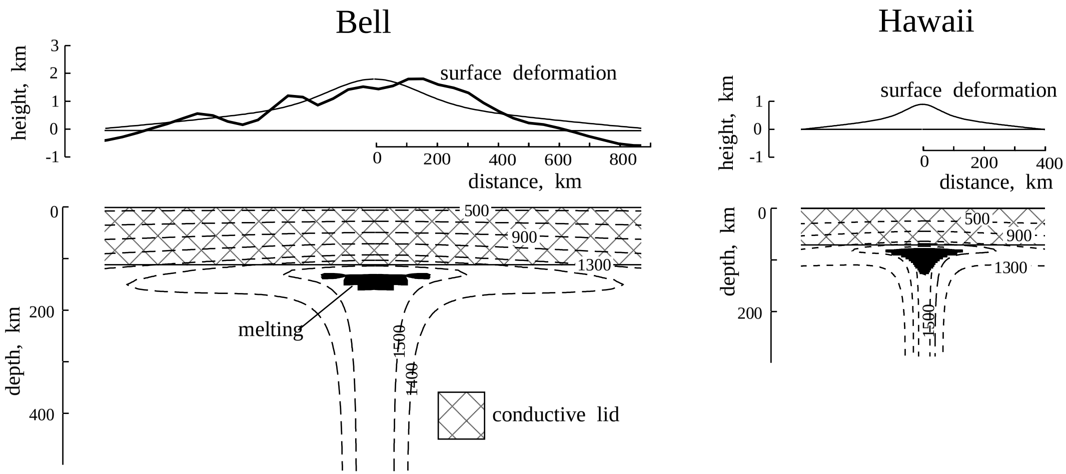

Figure 4 Summary diagram of model V15 (Nimmo & McKenzie 1996) and the best fit model for Hawaii of Watson & McKenzie (1991), showing the regions of melting and the surface deformation associated with the convection. The *thick bold line* is a profile across Bell (30°N, 47°E) using the $l = m = 75$ spherical harmonic topography. The *diagonal lines* indicate the mechanical boundary layer and melt generation occurs in the region in which the instantaneous melt production rate exceeds 0 [see Watson & McKenzie (1991) for more details]. *Dashed lines* are temperature contours in degrees Celsius; vertical exaggeration ~1.7:1. The depth of the convective layer is not well determined (see text). Main parameters used are as follows (quantities in parentheses are for Earth, the others for Venus): heat flux, 22 (51) mW m^{-2}; potential temperature, 1336°C (1280°C); Rayleigh number, 1.58×10^7 (1.78×10^7); mantle viscosity, 3.3×10^{20} (3.0×10^{19}) Pa s; lid thickness, 111 (72) km.

effect of such a zone is to greatly reduce the apparent depth of compensation (Robinson et al 1987, Kiefer & Hager 1991). The reason for the lack of a low viscosity zone may be associated with the fact that the mantle on Venus is probably drier than that on Earth (McKenzie 1977, Kaula 1990, Campbell & Taylor 1983). On Earth, the presence of small amounts of water in olivine reduces viscosity by two to three orders of magnitude (Hirth & Kohlstedt 1996).

The convection models cited above all used an assumption of constant viscosity. Solomatov & Moresi (1996) used a temperature-dependent viscosity and concluded that the viscosity in the interior of the mantle was probably 10^{20}–10^{21} Pa s, which is in good agreement with the value obtained by Nimmo & McKenzie (1996).

SURFACE HEAT FLUX Early models of the thermal state of the mantle assumed that the heat loss on Earth (82 mW m^{-2}) could be scaled to the heat loss on Venus. In fact, on Earth, the main cause of surface heat loss is plate tectonics (Sclater et al 1980); the heat generated by mantle radioactive decay is only about 40 mW m^{-2}. We argue above that the internal heat generation on Venus is likely to be similar to that of Earth; the question remains, however, as to the surface heat flux in the absence of plate tectonics. Recent convective models (Solomatov & Moresi 1996, Smrekar & Parmentier 1996, Nimmo & McKenzie 1996) have all produced values in the range 8–25 mW m^{-2}.

The validity of the heat flux values may be checked by their compatibility with elastic thickness estimates. The elastic thickness of terrestrial oceanic lithosphere follows an isotherm of between 450° and 600°C (Watts & Daly 1981), which corresponds to a homologous temperature [temperature/solidus temperature (in degrees Kelvin)] for wet mantle of 0.55–0.66 (Green 1973). A homologous temperature of 0.6 represents about 680° and 620°C for dry peridotite and dry basalt, respectively. Assuming that the homologous temperature is important in determining elastic thickness, for a T_e of 30 km, the above temperature range gives a heat flux of about 18–25 mW m^{-2}.

STRAIN RATE The strain rate associated with convective activity is important because it will affect the way the lithosphere responds to stresses. A study of faulting across impact craters suggests that the present globally averaged strain rate is no larger than 10^{-18}–10^{-17} s^{-1} (Grimm 1994b). However, the absence of craters in areas of volcanic or tectonic activity (Ivanov & Basilevsky 1993, Namiki & Solomon 1994, Price et al 1996) (which are probably areas of high strain rate) means that this latter value is probably an underestimate. In rifted areas, extension factors of about 1.2 from fault width measurements (see above) and impact crater shape changes (Price et al 1996) suggest that the strain rate is 10^{-16} s^{-1} if the rifts have been active for the last ~100 Ma (Price et al 1996).

Whether the convective velocity determines the velocity of the lithosphere depends on the viscosity structure, strength of the lithosphere, and geometry of circulation (Kiefer 1993). In the absence of a low viscosity zone, the lithosphere may be tightly coupled to the convecting mantle, as has been inferred from tessera morphology (Bindschadler et al 1992a) and plains deformation (Squyres et al 1992b). The convective velocity u is given by the following (McKenzie et al 1974):

$$u \sim Ra^{\frac{1}{2}}\kappa/d. \qquad (1)$$

because convective cells tend to have an aspect ratio ~1, the convective shear stress acts over a horizontal length of d. Assuming that the lithospheric and convective velocities are similar in the absence of a low viscosity zone, the convective strain rate $\dot{\epsilon}_{con}$ is

$$\dot{\epsilon}_{con} \sim Ra^{\frac{1}{2}}\kappa/d^2. \qquad (2)$$

For Ra of $\sim 10^7$ and d of 700 km, the maximum likely strain rate is therefore between 10^{-14} and 10^{-15} s^{-1}, which is not dissimilar to present-day strain rates on Earth. On the other hand, for a strongly temperature-dependent viscosity mantle, the surface layer becomes "stagnant" and the strain rate may be closer to 10^{-17}–10^{-18} s^{-1} (Solomatov & Moresi 1996).

In summary, a certain amount of agreement seems to exist about the following points:

- The present-day mantle temperature of Venus is similar to that of Earth;
- the lithospheric thickness is similar (80–200 km);
- the viscosity of the mantle is about an order of magnitude higher;
- the surface heat flux of Venus is about 8–25 mW m^{-2} and is consistent with the observed elastic thickness of the lithosphere.

Core

Comparatively little is known about the core of Venus. Its existence may be inferred from the similar densities of Venus and Earth, but because the mean moment of inertia is unknown (Mottinger et al 1985), neither the core size nor density can be deduced. Whether it is solid or liquid is also unclear, but recent estimates of the tidal Love number from Magellan's orbit by Konopliv & Yoder (1996) suggest that it may be liquid. The lack of a solid inner core may explain the absence of a magnetic field around Venus (Stevenson et al 1983), though the lower rotation rate compared to that of Earth may also be important.

Summary of Present-Day Processes on Venus

The previous sections show that some processes on Venus are similar to those on Earth and others are different. Considering the similarities first, Venus has a basaltic crust, an average effective elastic thickness of about 25–30 km, a mechanical boundary layer thickness probably around ∼100 km, and a mantle potential temperature of about 1300°C, all of which are similar to terrestrial oceanic lithospheric characteristics. As with Earth, mantle convection is vigorous ($Ra \sim 10^7$) and produces plumes (which are probably time dependent) that cause uplift, rifting, and volcanism.

Considering the differences, the surface appearance of Venus is dissimilar to Earth's because the lack of water has greatly reduced erosion and sedimentation and also because of the apparent absence of plate tectonics (at least since the resurfacing event 300–600 Ma ago). The viscosity of the mantle beneath the lithosphere on Venus is probably an order of magnitude higher than that on Earth, and the surface heat flux on Venus is about half the heat generation rate. Venus is therefore currently heating up, whereas Earth is cooling down, which implies that the thermal histories of the two planets must be different. The crustal thickness of Venus is uncertain, but it is probably about 30 km. The morphology of rifts such as Devana Chasma suggests that faults on Venus are capable of supporting stresses about eight times greater than faults on Earth.

The atmosphere of Venus is extremely dry, perhaps because of a different impact history during planetary formation. Dry rocks retain greater strength at high temperatures than do wet rocks, and dry melts have viscosities that are orders of magnitude higher than those of wet melts. Dry melting occurs at higher temperatures than wet melting, and phase transitions happen much more slowly in the absence of water. Because pore fluid pressure reduces the shear stresses that faults can withstand, faults under dry conditions are stronger than faults in wet conditions.

Assuming that the mantle of Venus is as dry as the atmosphere, and has been throughout the last 0.5 Ga, then the following observations can all be explained simply by the absence of water: high mantle viscosity beneath the lithosphere; fault strength that is about eight times that on Earth; and similar T_e to terrestrial oceanic lithosphere (despite a surface temperature of 450°C).

Perhaps surprisingly, most of the differences between Earth and Venus can be explained by one simple factor. The main conclusion of this review is that the fundamental importance of water to processes on Earth has not been fully recognized.

EVOLUTION OF GEOLOGICAL PROCESSES

The previous section shows that a consistent description of present-day processes on Venus can be developed. As with Earth, however, the route by which

Venus reached its present-day state is much less well understood. This section discusses four related topics: first, what little has been deduced about the history of Venus from the surface record; second, why Venus does not currently possess plate tectonics; third, the nature of the resurfacing event; and finally, how all these topics may relate to the thermal history of the planet.

Surface Record of Evolution

A knowledge of a planet's stratigraphy is a prerequisite for understanding its evolution, which in turn requires good age data. On Venus, relative age relationships may be obtained by normal geological observations, but absolute age determinations depend on cratering counts.

Regional examinations of overlapping relationships have as yet produced little information, although a global stratigraphic survey (Basilevsky & Head 1995) concluded that the resurfacing event involved tectonic and volcanic activity, with declining surface volcanism thereafter. Pre-resurfacing units may survive in tessera regions. At least in some areas, the emplacement of volcanic plains was geologically instantaneous over areas of up to 10^7 km^2 (Basilevsky & Head 1996).

Despite the possible variations in surface age discussed above, $>$80% of the surface is likely to have had all previous craters removed in a relatively short time compared to 500 Ma, probably between 10 and 100 Ma (Phillips et al 1992, Strom et al 1994, Namiki & Solomon 1994). Only the last episode of resurfacing is recorded: Whether it was a single event or a process that finally stopped over a 10–100 Ma period is impossible to tell. Following the last resurfacing, based on the small number of craters that are volcanically embayed from outside or tectonized, less than 20% of the planet was resurfaced over the next 300–600 Ma (Schaber et al 1992, Namiki & Solomon 1994, Strom et al 1994). The nature of the resurfacing event is discussed below.

Plate Tectonics

Previous authors have recognized that plate tectonics on Earth can take place because the driving forces (especially slab pull) exceed the resistive forces (primarily fault friction, flexure, and viscous drag) (e.g. Forsyth & Uyeda 1975, McKenzie 1977). The apparent lack of plate tectonics on Venus is of interest, and the balance of forces is likely to be different.

One possible explanation for the present-day lack of plate tectonics on Venus is that the high surface temperature and thick basaltic crust make the lithosphere less dense than the underlying convecting mantle, thereby inhibiting subduction. The higher surface temperature means that the average lithospheric temperature is about 200°C higher than that on Earth, assuming its base is defined by the same isotherm. This temperature difference alone would reduce the density contrast by about 20 kg m^{-3}. Furthermore, for a crust with a thickness greater than about 20% of the lithospheric thickness, it is likely that the lithosphere would be positively buoyant. The likelihood of a positively buoyant Venusian lithosphere has been noted by other authors in the past (Anderson 1981, Burt & Head 1992, Herrick 1994).

An important uncertainty in calculating lithospheric density is whether a ∼30-km thick crust of basalt would transform to denser assemblages of minerals (e.g. Burt & Head 1992). Although the boundaries of the basalt-eclogite transition are reasonably well known (Green & Ringwood 1967, Ito & Kennedy 1971), the rate at which the reactions take place is not. On Earth, deformation and the presence of water can greatly speed up the reaction (Mattey et al 1994), and in their absence, basalt may remain metastable to great depths (Rubie 1990).

A second explanation for the absence of plate tectonics is that the high mantle viscosity might cause a large viscous drag force on any plates that are moving (McKenzie 1977). On Earth, this mantle drag is greatly reduced by the effect of the low viscosity zone (e.g. Forsyth & Uyeda 1975), probably because of partial melting of wet mantle; such an effect does not appear to operate on Venus.

Another consequence of the high mantle viscosity, however, is that convection currents might exert enough viscous drag on an overlying stationary plate to initiate plate motion. Phillips (1986) has calculated the stresses in a stationary elastic plate caused by convection in a fluid with a constant viscosity beneath the plate and a rigid upper boundary condition. He found that in the presence of a low viscosity zone, the convective stresses never exceed 1 MPa. However, Venus lacks a low viscosity zone, in which case convective stresses associated with the large plumes are probably at least 100 MPa. Solomatov & Moresi's (1996) convective models produced stresses in the lithosphere of about 100–300 MPa. This drag force is an order of magnitude greater than the other forces involved in driving and opposing subduction, but only occurs over distances determined by the convective wavelength, which is typically 1000–3000 km. The potentially strong coupling between lithosphere and convecting mantle has been recognized by a number of authors (Phillips 1990, Bindschadler & Parmentier 1990, Bindschadler et al 1992b) and may be responsible, for instance, for the formation of tesserae and coronae.

Finally, the high strength of faults on Venus compared to those on Earth increases the frictional resistance to plate motion by an order of magnitude (McKenzie 1977). This force probably exceeds all likely driving forces except possibly convective drag.

In conclusion, the combination of high mantle viscosity, high fault strength, and thick basaltic crust may all have contributed to the difficulty of initiating and sustaining subduction on Venus. Convective motion associated with plumes may exert sufficient force locally to cause plate motion, but only on a plume length scale (∼1000 km) rather than on the 10,000-km scale seen on Earth.

The Nature of the Resurfacing Event

Virtually the only constraints on the resurfacing event come from the crater record (see above), which suggests that the last global resurfacing took place in less than 100 Ma and was followed by reduced levels of activity. In addition, if the conclusions regarding the present heat flux are correct, then Venus must have been heating up since the resurfacing event because the present rate of heat loss is less than the inferred radiogenic heat generation rate.

Various models have been suggested to explain the above observations, many of which require catastrophic events (see below). Rather than investigate these suggestions further, it is instructive to examine two potential resurfacing mechanisms that are constrained by terrestrial observations, namely resurfacing by plate tectonics and resurfacing by plume activity.

PLATE TECTONIC RESURFACING Plate tectonics, which is taken to mean the creation of lithosphere at spreading ridges and its destruction at trenches, could certainly account for the rapid resurfacing. Assuming that the last resurfacing event took 100 Ma to complete, the average surface age during this time must have been 50 Ma. This age is similar to the average age of oceanic lithosphere on Earth (Sclater et al 1980).

The main topographic features of plate tectonics—ridges, trenches, and transform faults—tend to disappear once plate tectonics has stopped, either because they are no longer thermally maintained or because they are topographic lows that are likely to be preferentially filled in by volcanism. Other characteristics of plate tectonics, such as linear bands of faulting, are also likely to be obscured over the succeeding 500 Ma. Thus, the lack of evidence for present-day plate tectonics on Venus cannot be used as evidence that it was not the cause of the resurfacing event. However, it is not clear whether plate motions could stop within the 100-Ma time scale required by the cratering record.

PLUME-RELATED RESURFACING The other suggested mechanism of resurfacing is plume melt generation. Because less than 20% of the planet has been resurfaced over the last 500 Ma, whereas the resurfacing event took less than 100 Ma, the global surface melt production rate must have been a factor of at least 25 higher than the present-day rate for the 100 Ma before the resurfacing event ended. Such a rate of melt generation could be achieved by increasing the number of plumes, increasing the potential temperature by about 200°C (see Figure 3*c*), reducing the lid thickness, or reducing the solidus temperature (probably by the addition of water). However, it is not clear that any of these mechanisms can produce a rapid reduction in melt generation over 100 Ma: For instance, it is difficult to see how the plume population could decrease by 96%

over 100 Ma, and reduction of the upper mantle temperature by 200°C over 100 Ma would require an average heat flux of 180 mW m^{-2}, about 10 times the present-day rate.

On the basis of these calculations, plume activity alone seems an improbable mechanism to explain the style of resurfacing required by the crater observations. Plate tectonic activity is more plausible, but the mechanism that might cause the motions to cease is not clear.

Thermal Evolution and the Resurfacing Event

Given the disagreement over the current thermal state of Venus and the nature of the resurfacing event, it is not surprising that there is little agreement as to its thermal history. We argue above that the present low surface heat flux means that Venus has been heating up since the resurfacing event. Nimmo & McKenzie (1997) showed that the effect of a sudden drop in surface heat flux can cause the upper mantle temperature to increase by 100°–500°C over 1 Ga if it convects independently of the lower mantle, but that the rate is only ∼50°C/Ga for a single layer mantle. Herrick (1994) also concluded that a change in surface boundary conditions could lead to a rapid rise in mantle temperature and catastrophic resurfacing. These scenarios imply periodicity of behavior, which has also been suggested by other authors. Parmentier & Hess (1992) and Herrick & Parmentier (1994) have suggested that competition between compositional stratification and thermal buoyancy may cause episodic overturn of the upper mantle. Turcotte (1993) suggested that ∼500-Ma periods of surface quiescence (during which the lithosphere thickens to 300 km) are interspersed with short periods of plate tectonics.

On the other hand, many authors have taken the resurfacing event to indicate an irreversible change in behavior. Steinbach & Yuen (1992) have suggested that as Venus cooled, the tendency for mantle layering to occur decreased (Christensen & Yuen 1985), which may have caused hot lower mantle material to sweep upwards, causing extensive melting and resurfacing or producing large-scale deformation of a highly non-Newtonian lithosphere (Weinstein 1996). To avoid invoking a catastrophe, Solomon (1993) proposed that steady cooling of the mantle might lead to a rapid decrease in the rate of tectonic resurfacing, since the strain response to a stress is exponentially dependent on temperature. This, however, does not really explain the widespread volcanic plains, which are all of about the same age. Arkani-Hamed et al (1993) suggested that the vigor of convection decreases as the Rayleigh number declines and the lithosphere ceases to be incorporated into the circulation, which ends resurfacing. Similarly, Solomatov & Moresi (1996) suggested that the reduction in convective stresses with time might cause deformation in the stagnant lid to change from brittle (plate tectonics) to ductile (little surface movement).

The influence that the resurfacing event must have had on the thermal history of Venus depends on the resurfacing mechanism. Perhaps the most crucial question is the extent to which resurfacing involved recycling of the strong surface layer. Regimes in which the lithosphere can be incorporated into the convective circulation cool much faster than those in which it forms a rigid lid on top. On Earth, this recycling occurs through plate tectonics, which accounts for about 75% of the surface heat loss (Sclater et al 1980). Moreover, subducted sediments probably contribute substantial amounts of water to the depleted upper mantle. Whether such recycling ever occurred on Venus is unclear: Surface observations suggest that subduction is not occurring at present, but the strong coupling between lithosphere and mantle means that it may have been a possibility in the past. Alternatively, the basalt-eclogite transition might provide a recycling mechanism.

We conclude above that the surface heat flux of Venus is one third to one quarter of that of Earth, and the arguments above regarding the thermal evolution of the planet assume that this value is correct. However, as-yet unrecognized mechanisms may exist, analogous to “black smokers” on Earth, that might increase the present-day surface heat flux of Venus. In order to greatly increase the surface heat flux, a large fraction of the mechanical boundary layer must regularly be replaced with hotter material, as happens at mid-ocean ridges on Earth. Such a globally operating process is unlikely to occur on Venus without leaving some identifiable signs in the Magellan data set. The only regions where such a process might perhaps remain undetected are the extensive regions of tesserae, such as Aphrodite.

In conclusion, whether any of the proposals discussed in this section are correct is not yet clear. The probability of understanding the thermal evolution of Venus from the observations presently available seems low, not least because the resurfacing event appears to have removed most of the pre-existing surface features. Although disappointing, this conclusion is not surprising when one considers the Earth. Despite a rock record extending over 4 Ga, the details of the Earth’s thermal evolution before the formation of the oldest oceanic crust are still uncertain.

FUTURE WORK

That Venus appears to have suffered at least one global catastrophic event challenges the usual geological assumption of uniformitarianism. However, a careful examination of Earth’s geological history may reveal analogous events.

The amount of water present has dominated the evolution of Venus and the Earth. Water is important in determining mantle viscosity, fault strength, crustal strength, and melt transport. Accordingly, investigating the water budget of

Venus is of great interest, especially the initial total abundance of water; how it was partitioned between mantle, crust, and atmosphere; and how it evolved to its present low levels.

Given the potential importance of the basalt-eclogite transition, more experimental information about the kinetics of the transformation, and the effect that water has on it, would be very valuable. Such experiments might help to constrain the thickness of the crust and the composition of tesserae, neither of which are well known. Of particular interest is the question of whether tesserae are compositionally analogous to terrestrial continents (e.g. Jull & Arkani-Hamed 1995). If they are, then the formation of such areas from silica-rich melts (which are very viscous when dry) presents an interesting problem.

In addition to tesserae, various types of features, most notably coronae, appear to have no terrestrial analogues. Accordingly, and despite much work, no detailed understanding has emerged as to how these landforms arose. Modeling using more realistic rheologies may provide an explanation of these features.

CONCLUSIONS

We now have a satisfactory model of the present dynamics of Venus, which is not much less detailed than similar models for Earth. As outlined above, this model of Venus is based on the following evidence: The internal temperature is constrained by the MgO concentration of 8–11.5% and the *mg*# [$= 100 \times$ [MgO]/([MgO]+[FeO])] of surface basalts of 60–75 (McKenzie et al 1992a); given this temperature, the mechanical boundary layer thickness is constrained by melt generation rate and the admittance. The mantle viscosity is required to be about 3×10^{20} Pa s by the magnitude of the plume-related gravity anomalies on Venus. The effective elastic thickness is estimated, from the relationship between gravity and topography at wavelengths between 500 and 2000 km, to be about 30 km. Assuming that the elastic thickness is controlled by temperature, the observed T_e values are consistent with the temperature gradient inferred from the plume models for either dry basalt or peridotite.

Figure 5 compares the likely temperature structure of the inferred Venusian mantle with that of a continental craton on Earth. The lithospheric thicknesses are comparable, although the thermal boundary layer on Venus is thicker because of the higher mantle viscosity.

Our understanding of the evolution of Venus is much less advanced, as is that of the Earth's evolution. The nature of the resurfacing event is still unclear, but it must mean that the evolution of Venus has differed from that of the Earth, a conclusion also implied by the absence of plate tectonics on Venus. Differing water budgets on the two planets may have been important in determining their different histories.

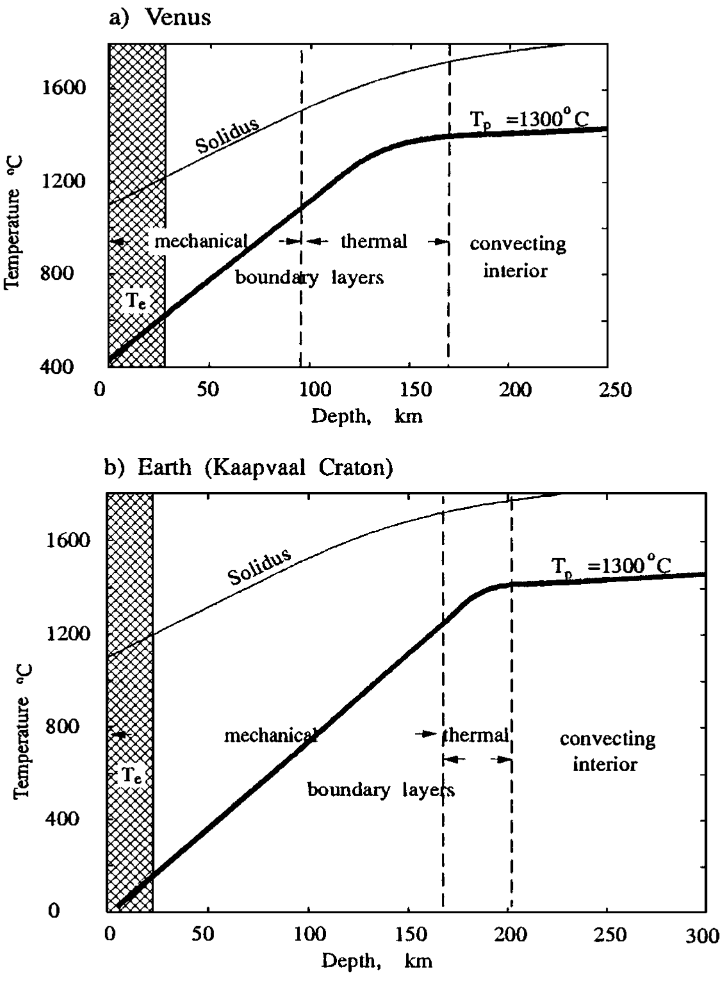

Figure 5 Comparison of the lithospheric structure of Venus with that of the Kaapvaal Craton of South Africa, reproduced from Nimmo & McKenzie (1997). The *bold line* is the geotherm; the *thin line* is the solidus; *diagonal lines* represent the maximum likely elastic thickness of the lithosphere. The values of the parameters used are as follows (quantities in parentheses are for Earth, the others for Venus): mantle viscosity, 3×10^{20} (10^{19}) Pa s; surface temperature, 430°C (0°C); mechanical boundary layer thickness, 95 (165) km; and interior potential temperature 1300°C (1300°C). The heat flux calculated from the thermal structure is 22 (23) mW m^{-2}.

Of particularly interest is that Venus is, in many ways (size, composition, mantle temperature, and lithospheric and elastic thickness), similar to Earth, and yet it lacks plate tectonics and has suffered apparently catastrophic changes in surface behavior. Some of these differences may plausibly be explained by the absence of water, but others are more puzzling. Hopefully, our increased understanding of Venus will contribute to a greater understanding of the way the Earth functions and will allow us to understand processes operating within other silicate bodies, such as Mars and the Moon.

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