

## How does Venus lose heat?

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**Abstract.** The tectonics and volcanism of the terrestrial planets are controlled by the loss of heat from the planetary interior. On the Earth, about 70% of the heat flow through the mantle is attributed to the subduction of cold lithosphere. In order to understand the tectonic and volcanic processes on Venus it is necessary to understand how heat is transported through its mantle. In this paper, three alternative end-member hypotheses are considered. The first is the steady state loss of heat through the mantle to the surface in analogy to the Earth. However, without plate tectonics and subduction on Venus, a steady state requires either a very high plume flux or very rapid rates of lithospheric delamination. The required plume flux would be equivalent to about 80 plumes with the strength of the Hawaiian plume. The required delamination flux implies a 50% delamination of the entire Venus lithosphere every 10 m.y. Neither appears possible, so that it is concluded that Venus cannot transport heat through its mantle to its surface on a steady state basis. The second hypothesis is that there has been a strong upward concentration of the heat-producing elements into the crust of Venus; the heat generated is then lost by conduction. Surface measurements of the concentrations of the heat-producing elements place constraints on this model. If everything is favorable this hypothesis might be marginally acceptable, but it is considered to be highly unlikely. The third hypothesis is that heat is lost by episodic global subduction events followed by long periods of surface quiescence. The near-random distribution of craters suggests that the last subduction event occurred about 500 Ma. This model implies a thick thermal lithosphere ( $\approx 300$  km) at the present time, which is consistent with a variety of surface observations. Lava lakes on the Earth are considered as analogies to plate tectonics; they also exhibit episodic subduction events.

### Introduction

Studies of the surface of Venus during the Magellan mission have provided a wealth of data on its tectonic and volcanic processes [Solomon *et al.*, 1992]. The radar images of the surface are complemented by global topography and gravity anomaly data. It is now clear that plate tectonics, as it is known on the Earth, does not occur on Venus. At the present time, Venus is a one-plate planet. Nevertheless, there are tectonic features on Venus that certainly resemble major tectonic features on the Earth. Beta Regio has many of the features of a continental rift on this planet. It has a domal structure with a diameter of about 2000 km and a swell amplitude of about 2 km. It has a well-defined central rift valley with a depth of 1–2 km, and there is some evidence for a three-armed planform (allocogen). Alta, Eistla, and Bell Regiones have similar rift zone characteristics [Senske *et al.*, 1992; Grimm and Phillips, 1992]. Aphrodite Terra, with a length of some 1500 km, is reminiscent of major continental collision zones on this planet, such as the mountain belt that extends from the Alps to the Himalayas. Ishtar Terra is a region of elevated topography with a horizontal scale of 2000–3000 km. A major feature is Lakshmi Planum, which is an elevated plateau similar to Tibet with a mean elevation of about 4 km. This plateau is surrounded by linear mountain belts, Akna, Danu, Freyja, and Maxwell montes, reaching elevations of 10 km, similar in scale and elevation to the Himalayas.

Mountain belts on the Earth are generally associated with plate tectonic processes. The global mid-ocean ridge system stands  $\sim 2.5$  km above the ocean basins. This topography is attributed to the thermal compensation of oceanic lithosphere. Mountain belts associated with subduction zones (e.g., the Andes) and with continental collisions (e.g., the Alps and Himalayas) are associated with crystal thickening and Airy compensation. Volcanism on the Earth is associated with zones of plate accretion (mid-ocean ridges), plate destruction (island arcs), and hot spots (mantle plumes). Magma generation at both ocean ridges and hot spots is attributed to pressure release melting; magma generation at island arcs is still poorly understood. Clearly, any comprehensive understanding of tectonism and volcanism on Venus requires an understanding of how heat is transported in the absence of plate tectonics.

On the Earth some 70% of the heat transfer through the mantle is attributed to the subduction of the cold oceanic lithosphere at ocean trenches. The remainder is primarily attributed to the ascent of hot mantle plumes with a minor contribution from the partial subduction (delamination) of the lower continental lithosphere. Without active plate tectonics the evolution of Venus is significantly different than the Earth. Three end-member models have been proposed, each of which will be discussed in turn.

The first model is the uniformitarian model. In this model the transport of heat through the mantle and lithosphere of Venus is in a near steady state balance with the heat generated by the heat-producing elements and the secular cooling of the planet. This requires a relatively thin, stable lithosphere with heat transport to its base by mantle convection. Heat transport

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through the lithosphere must be conduction or another unspecified mechanism.

The second model is the catastrophic model. In this model the present loss of heat to the surface of Venus is not in balance with its internal heat generation. The global lithosphere stabilized about 500 Ma, and the interior of the planet has been heating up since then. Heat is lost in episodes of global subduction of the thickened lithosphere.

The third model is the differentiated planet. In this model the heat-producing elements have been almost entirely fractionated into the crust, and the heat generated is lost by conduction to the surface.

### Uniformitarian Model

Before considering a uniformitarian model for Venus, a brief discussion of how heat is transported through the mantle of the Earth will be given. A comprehensive review has been given by *Turcotte and Schubert* [1982]. The total heat loss at the surface of the Earth is close to  $Q_E = 3.55 \times 10^{13}$  W with an estimated error of less than 5%. With a total surface area of  $A_E = 5.1 \times 10^8$  km<sup>2</sup> the mean surface heat flow is  $q_E = 70$  mW m<sup>-2</sup>. The origin of this heat is the decay of the radioactive isotopes of uranium, thorium, and potassium and the secular cooling of the planet. The Urey number  $Ur$  is defined to be the ratio of radioactive heat generation to the total heat loss;  $1 - Ur$  is the fraction attributed to secular cooling. Estimates for the Urey number for the Earth fall in the range  $0.6 < Ur < 0.8$ .

Heat transport in the ocean basins is dominated by plate tectonics, whereas heat is lost conductively through the stable continental lithosphere. The oceans and marginal basins have an area  $A_o = 3.1 \times 10^8$  km<sup>2</sup> (60%) and a heat loss  $Q_o = 2.42 \times 10^{13}$  W (68%). Of this total, *Sclater et al.* [1980] attribute 90% to the subduction of the conductively cooled lithosphere ( $Q_{os} = 2.18 \times 10^{13}$  W) and 10% to basal heating ( $Q_{om} = 0.24 \times 10^{13}$  W). The cooling of the oceanic lithosphere generally follows a half-space cooling model to ages of about 100 Ma; the subsequent flattening of topography is attributed to basal heating of the lithosphere by secondary convective processes such as the impingement of mantle plumes.

The continents and continental margins have an area  $A_c = 2 \times 10^8$  km<sup>2</sup> (40%) and a heat loss  $Q_c = 1.13 \times 10^{13}$  W (32%). Of this total, *Turcotte and Schubert* [1982] attribute 40% to radioactive isotopes in the continental crust ( $Q_{cc} = 0.45 \times 10^{13}$  W) and 60% to either basal heating or delamination of the continental lithosphere ( $Q_{cm} = 0.68 \times 10^{13}$  W). Inherent in this balance is the assumption that the continental lithosphere has, on average, a steady state balance between basal input, internal heat generation, and surface heat loss. The continental lithosphere is not a thickening thermal boundary layer. Observational evidence for this steady state balance is the lack of thickening sediment piles associated with continental cooling and subsidence.

From the above values the total heat flux from the mantle is estimated to be  $Q_m = 3.10 \times 10^{13}$  W. Of this total, 70% is attributed to the subduction of cold lithosphere ( $Q_s = 2.18 \times 10^{13}$  W) and 30% to other transport processes in the mantle ( $Q_m = 0.92 \times 10^{13}$  W). Two primary candidates have been put forward as other mantle transport processes, mantle plumes and lithospheric delamination. Other mechanisms of secondary convection have also been proposed.

Quantitative studies of the mantle heat transport associated

with plumes have been carried out by *Davies* [1988] and *Sleep* [1990]. These studies relate the rate of creation of plume swells to the mantle heat flux. *Davies* [1988] estimates that the mantle flux due to plumes is  $Q_p = 2.5 \times 10^{12}$  W, and *Sleep* [1990] estimates the value to be  $Q_p = 2.3 \times 10^{12}$  W. *Sleep* [1990] considered 37 plumes from both the oceans and the continents. The heat flux associated with the Hawaiian plume is  $Q_{pH} = 0.36 \times 10^{12}$  W, 15% of the total plume flux and twice as large as any other plume.

Taking the mean of the two estimates given above, the plume flux is  $Q_p = 2.4 \times 10^{12}$  W. However, this is only 26% of the mantle heat flux of  $Q_m = 0.92 \times 10^{13}$  W that is attributed to processes other than subduction. An important question is the process or processes that contribute the other 74%. Of course, the original estimate of  $Q_m = 9.2 \times 10^{12}$  W could be in error. However, it is difficult to accept that it is a factor of 4 too large. There are three possible sources for the discrepancy:

The first possible source is plumes that do not generate well-defined volcanic hot spots and swells. The smallest hot spots considered by *Sleep* [1990] have a heat flux of about  $Q_p = 10^{10}$  W so that we would require some 700 of these to make up the missing flux of  $Q_p = 6.8 \times 10^{12}$  W. It should be noted, however, that the lack of subsidence in continental cratons indicates heating from below even without surface evidence for plumes. *McKenzie* [1984] has proposed that epeirogenic uplift is due to intrusive volcanism near the base of the continental crust without surface volcanics. Thus the studies by *Davies* [1988] and *Sleep* [1990] may represent only a fraction of the actual mantle plume flux.

The second possible source is delamination of the continental lithosphere. The continental lithosphere is gravitationally stable, and there is no evidence that it can be subducted as a whole. However, the mantle portion is gravitationally unstable, and there is considerable evidence that the mantle lithosphere and lower crust separate from the upper crust and descend into the mantle [*Bird*, 1979]. This is delamination, and it will contribute to the mantle heat flux in the same way that subduction of the oceanic lithosphere does. There is observational evidence for lithospheric and lower-crustal delamination in the Himalayas, Alps, and Colorado Plateau. However, the rates are so small that the associated mantle heat flux is generally assumed to be negligibly small.

The third possible source is other secondary convective processes in the mantle. In order for secondary convection to convect significant heat it must contain either ascending hot rock or descending cold rock. But ascending hot rock is generally associated with mantle "plumes" and descending cold rock with descending subducted or delaminated lithosphere. Thus it is not clear that it is appropriate to discuss heat transport associated with secondary convection except in the context of either plumes or delamination (subduction). The loss of heat to the surface of the Earth is illustrated in Figure 1a and tabulated in Table 1. The values given are preferred values but are subject to the errors discussed above.

The first hypothesis for Venus is that it is in a near steady state balance between radioactive heat production and secular cooling and the surface heat loss. Many of the arguments for such a hypothesis have recently been given by *Phillips and Hansen* [1994]. For a near steady state heat loss model, a logical estimate for the present total required surface heat flow on Venus is obtained by scaling the Earth's heat loss ( $Q_E = 3.55 \times 10^{13}$  W) to Venus using the masses of the two planets

( $M_E = 5.97 \times 10^{24}$  kg,  $M_V = 4.87 \times 10^{24}$  kg). The result is that the present heat loss from Venus is  $Q_V = 2.91 \times 10^{13}$  W and the mean surface heat flow is  $q_V = 63$  mW m<sup>-2</sup>. It should be emphasized that this calculation implicitly assumes that Venus and the Earth have similar concentrations of the heat-producing elements. This is certainly reasonable in terms of present models of planetary accretion, and we will return to this point when we discuss the measurements of the surface concentrations of the heat-producing elements on Venus. With the steady state hypothesis and a thermal conductivity  $k = 3.3$  W m<sup>-1</sup> °K<sup>-1</sup> the mean surface thermal gradient on Venus is  $dT/dy = 19^\circ\text{K km}^{-1}$ . Assuming a linear conduction gradient through the lithosphere and a  $\Delta T = 850^\circ\text{K}$ , the mean lithosphere thickness is  $y_L = 45$  km. The mechanical arguments against such a thin lithosphere on Venus have been given by *Turcotte* [1993]; here we are concerned with the thermal implications. It should be noted that the lithosphere can be thickened if a significant fraction of the planet's heat-producing elements are concentrated in it; this model will be discussed in a later section.

If Venus behaves like the Earth, the relative contributions of radioactive heat generation and secular cooling are likely to be nearly equal on the two planets. The expected secular cooling of Venus for the last 500 m.y. is given in Figure 2 for  $Ur = 0.8, 0.6$ . A typical rate of cooling is  $70^\circ\text{K b.y.}^{-1}$ .

Since plate tectonics is not available to transport heat through the mantle of Venus, alternative mechanisms for heat transport must be found if a near steady state hypothesis is to

**Table 1.** Preferred Values of Heat Loss for the Earth and for the Three Hypothesized Models for Venus

$Q, 10^{13}$ W	Earth	Venus		
		ss	e	c
Total heat loss	3.55	2.91	2.91	2.00
Subduction	2.18	...	2.62	...
Plumes	0.90	0.88	0.29	...
Delamination	0.02	2.03	...	...
Conduction	0.45	...	...	2.00

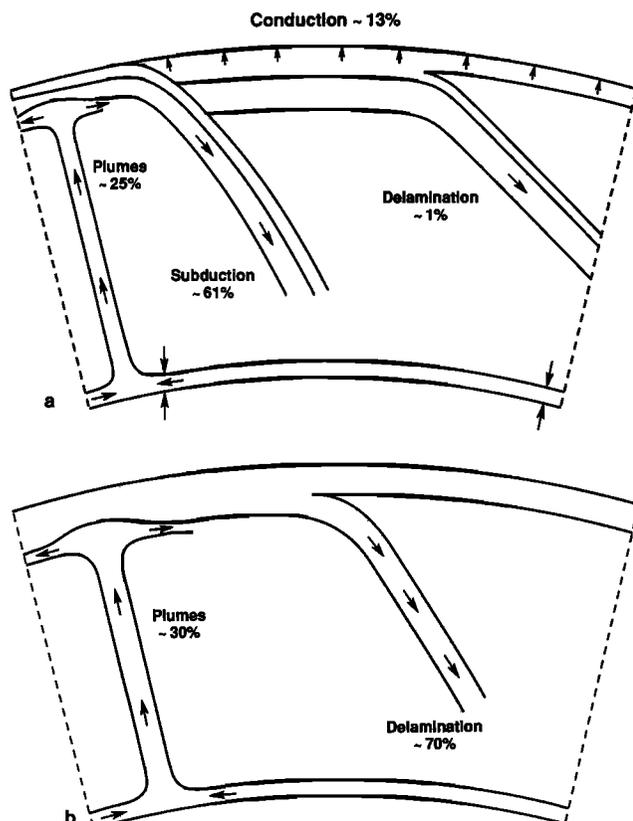
Here, ss denotes steady state; e, episodic subduction; and c, crustal conduction.

be viable. Without the participation of cold subducted lithosphere the temperature differences associated with ascending hot material and descending cold material through the mantle of Venus will be considerably less than on the Earth. In order to transport the same amount of heat we require larger material fluxes through the mantle. The implication is that the Rayleigh number will be higher on Venus; since the Rayleigh number is principally sensitive to the viscosity, the conclusion is that the mean viscosity in the mantle of Venus must be considerably less than on the Earth in order to provide the necessary steady state heat transport. Based on the above discussion of the Earth we consider two mechanisms, plumes and lithospheric delamination. The former could be attributed to instabilities in a hot basal boundary layer, and the latter to partial instabilities in the cold surficial boundary layer, i.e., the lithosphere.

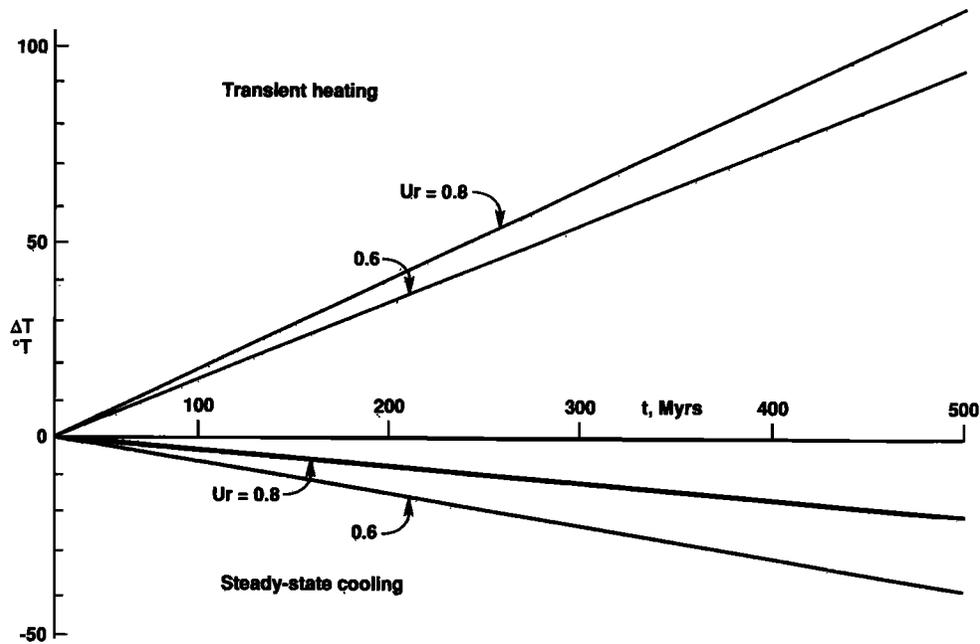
The topography and associated gravity anomalies of the equatorial highlands on Venus have been attributed to the dynamic processes associated with mantle plumes by several authors [*Phillips et al.*, 1991; *Kiefer and Hager*, 1991, 1992]. The extensive distributions of coronas that cover the planet have also been attributed to mantle plumes [*Stofan et al.*, 1991, 1992]. It is certainly reasonable to accept that there is observational evidence for the impingement of mantle plumes on the base of the Venusian lithosphere [*Koch*, 1994]. The question is whether the plumes are sufficiently large and numerous enough to transport the required heat. To carry a substantial fraction of the total heat ( $Q_V = 2.9 \times 10^{13}$  W) would require about 80 plumes with the strength of the Hawaiian plume ( $Q_{PH} = 3.6 \times 10^{11}$  W). With a thin, hot lithosphere, such a large number of strong plumes would be expected to have surface signatures including extensive volcanism; these signatures are not seen. Quantitative studies of the strength of active plumes on Venus indicate that the present plume flux is less than on the Earth [*Smrekar and Phillips*, 1991].

Also, plumes are associated with the instability of hot basal boundary layers. For a fluid heated from within, heat is transported primarily by instabilities in near-surface cold boundary layers [*Parmentier et al.*, 1994]. For Venus this means, for the steady state model, that heat would be transported primarily by delamination of the cold lithosphere. A mechanism for lithospheric delamination on Venus has been proposed by *Buck* [1992]. He decouples the upper crust from the upper mantle with a low-viscosity lower crust, a lower crustal asthenosphere. The upper mantle participates in a plate tectonic subduction cycle, but the upper crust behaves as a scum that floats and does not participate in the subduction. A similar model has been proposed by *Arkani-Hamed* [1993].

The equivalent surface heat flux due to the subduction or



**Figure 1.** (a) Illustration of how heat is transported to the surface of the Earth by subduction, plumes, delamination, and lithospheric conduction. Estimates for the relative contributions are given. (b) Illustration of how steady state heat transport on Venus might be accommodated.



**Figure 2.** The change in temperature  $\Delta T$  of the interior of Venus is given as a function of time  $t$ . The steady state cooling results are applicable for a uniformitarian model, and the rate of cooling is given by the Urey number  $Ur$ . The transient heating would be applicable to the episodic subduction hypothesis during periods of lithospheric stability and conductive thickening; no heat loss to the surface is assumed.

delamination of cold lithosphere is easily estimated. The energy associated with a global subduction event is given by [Turcotte and Schubert, 1982, p. 281]

$$E_s = 8\pi r_V^2 \rho c (T_m - T_s) \left( \frac{\kappa t}{\pi} \right)^{1/2} \quad (1)$$

where  $r_V$  is the radius of Venus,  $T_s$  the surface temperature,  $T_m$  the mantle temperature, and  $\kappa$  the thermal diffusivity. This relation assumes that the lithosphere has thickened conductively from zero thickness for a period  $t$ . This is clearly a limiting case in that any basal heating of the lithosphere has been neglected. The equivalent mean surface heat loss  $Q_s$  is obtained by dividing (1) by  $t$  with the result

$$Q_s = 8\pi r_V^2 \rho c (T_m - T_s) \left( \frac{\kappa}{\pi t} \right)^{1/2} \quad (2)$$

Taking  $r_V = 6050$  km,  $T_m - T_s = 880^\circ\text{K}$ , and  $\kappa = 1 \text{ mm}^2 \text{ s}^{-1}$ , the heat loss is given as a function of  $t$  in Figure 3. If the total required heat loss from the interior of Venus ( $2.91 \times 10^{13}$  W) were associated with subduction, then the lithosphere would have to subduct, on average, at an age of  $t = 85$  Ma.

The above result can also be used to determine the efficiency of delamination in transporting heat through the lithosphere of Venus. If the temperature separating delaminating lithosphere from stable lithosphere is  $T_D$ , then we define

$$\vartheta_D = (T_D - T_s) / (T_m - T_s)$$

and the mean surface heat loss due to delaminating lithosphere is given by

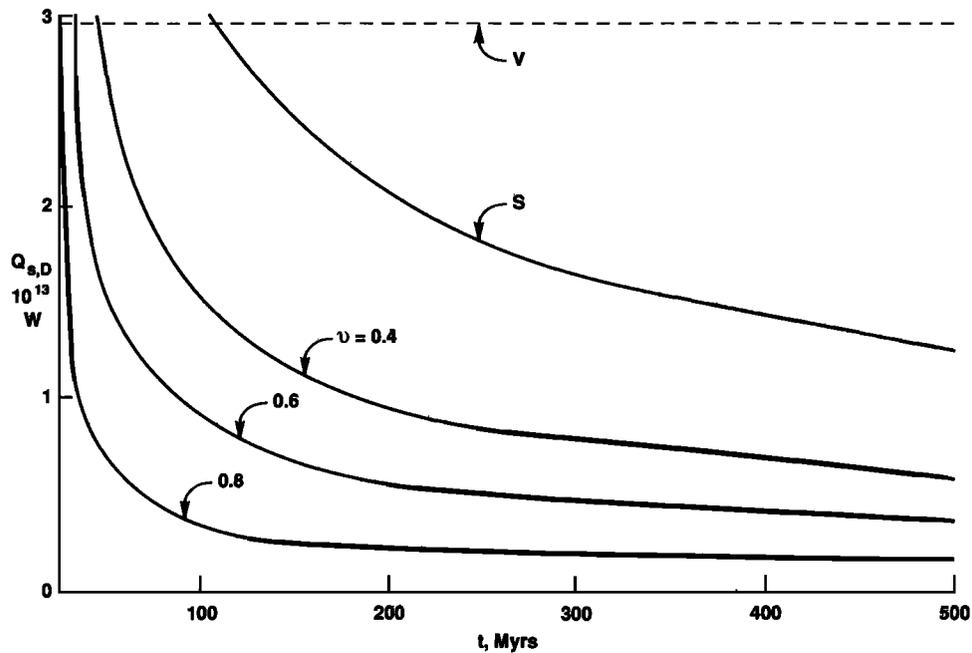
$$Q_D = 8\pi r_V^2 \rho c (T_m - T_s) (1 - \vartheta_D) \left( \frac{\kappa}{\pi t} \right)^{1/2} \quad (3)$$

The equivalent mean surface heat losses  $Q_D$  for  $\vartheta_D = 0.4, 0.6, 0.8$  are given in Figure 3 as a function of the delamination interval  $t$ . If the entire mantle heat flux is attributed to delamination, then it must occur sufficiently often to transmit  $Q_D = 2.91 \times 10^{13}$  W. With  $T_D = 1470^\circ\text{K}$  ( $\vartheta_D = 0.8$ ) the entire lithosphere would have to delaminate, on average, at intervals of  $t_D = 1.2$  m.y., with  $T_D = 1290^\circ\text{K}$  ( $\vartheta_D = 0.6$ ) we require  $t_D = 6$  m.y., and with  $T_D = 1110^\circ\text{K}$  ( $\vartheta_D = 0.4$ ) we require  $t_D = 19$  m.y. It seems inconceivable that global delamination events could take place at such short intervals. Also, global delamination events would be expected to disrupt the upper crust, resulting in intensive volcanism, volcanism that is not observed. This is not to say, however, that delamination is not occurring. Delamination may play an important role in creating the high plateau topography of Ishtar Terra. Nevertheless, it seems inconceivable that delamination could make a significant contribution to the global heat flow. This is consistent with our understanding of the role of delamination on the Earth.

A schematic steady state model for Venus is illustrated in Figure 1b and is tabulated in Table 1. However, since a large plume flux and/or extensive lithospheric delamination is a prerequisite for any steady state hypothesis, it is necessary to conclude that uniformitarian models are not applicable. We next turn to an alternative catastrophic model.

### Catastrophic Model

On the Earth, plate tectonics continuously creates new oceanic crust. Thus the age of the surface rocks has considerable variability. There is conclusive observational evidence that this is not the case for Venus. Based on the near-random distribu-



**Figure 3.** The mean surface heat flux  $Q_{s,D}$  is given for lithospheric subduction ( $S$ ) and delamination ( $D$ ) at intervals  $t$ . The curve  $S$  is for total subduction of the lithosphere. The parameter  $\vartheta$  is a measure of the fraction of the lithosphere that participates in delamination;  $\vartheta = 0$  is equivalent to subduction. The heat loss denoted by  $V$  is the required steady state value obtained by scaling from the Earth.

tion of craters in its surface, *Schaber et al.* [1992] postulated that a global resurfacing event occurred on Venus about 500 Ma. Further statistical tests have confirmed this hypothesis [*Bullock et al.*, 1993; *Strom et al.*, 1994]. Although there has certainly been some surface volcanism since that time, it is clear that a large fraction (~80–90%) of the surface was covered by volcanism during a relatively short period of time [*Namiki and Solomon*, 1994]. This observation is direct evidence that the geologic evolution of Venus is far more catastrophic than the Earth [*Herrick*, 1994]. An essential question is whether the thermal evolution of Venus is also strongly time dependent.

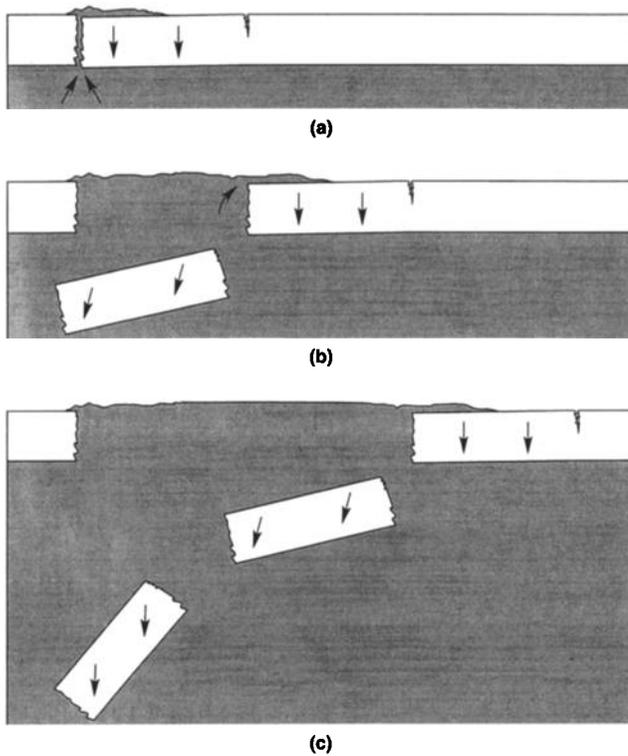
If it is hypothesized that the present heat loss from the interior of Venus is not in balance with the internal heat generation, several alternative models can be postulated. In one model, plate tectonics simply ceased some 500 Ma [*Arkani-Hamed and Toksoz*, 1984; *Arkani-Hamed et al.*, 1993; *Arkani-Hamed*, 1994]. According to this hypothesis the global lithosphere stabilized at that time and will remain stable. The principal problem with this hypothesis is that the interior temperature will increase without plate tectonics or alternative heat transport mechanisms. The rate of temperature increase is given in Figure 3 for two values of the Urey number,  $Ur = 0.6$  and  $0.8$ . Over 500 m.y. the mantle temperature has increased about  $100^\circ\text{K}$ ; the corresponding decrease in viscosity is about an order of magnitude.

An alternative model is that this temperature increase will eventually trigger another episode of global volcanism and tectonics. Without sufficient surface heat loss the lithosphere on Venus will thicken during the period of surface quiescence. One suggestion is that the thickening lithosphere eventually becomes sufficiently unstable that a global subsidence event is triggered [*Turcotte*, 1993]. Heat stored in the interior during the period of surface quiescence is lost to the subducted litho-

sphere and during the period of volcanic activity and tectonism that would follow. An alternative explanation is that mantle convection is episodic due to chemical differentiation [*Parmentier and Hess*, 1992; *Herrick and Parmentier*, 1994].

The limiting behavior of the lithosphere during surface quiescence would be that the lithosphere has been thickening conductively since that time with no significant convective heat flux to its base. In this limit the thickness of the lithosphere is now near 300 km. Such a thick lithosphere is consistent with a number of observations: (1) it provides support for the high topography, up to 10 km; (2) it is consistent with the high observed geoid-topography ratios, up to 33 m/km [*Smrekar and Phillips*, 1991]; (3) it is consistent with the observed unrelaxed craters [*Grimm and Solomon*, 1988]; and (4) it is consistent with the thick, elastic lithospheres inferred from flexural studies [*Sandwell and Schubert*, 1992a].

There is also direct observational evidence that episodic subduction is an applicable mechanism for heat transfer in a convecting system with a very viscous (rigid) upper thermal boundary layer. A natural analog for mantle convection is the thermal convection in a lava lake. Atmospheric cooling creates a "solid" crust which is gravitationally unstable with respect to the molten magma beneath. Episodic subduction has been observed in lava lakes. *Wright et al.* [1968] describe the behavior of the Makaopuhi lava lake during the eruption of the Kilauea volcano in March 1965. They describe a particularly graphic episode of episodic subduction of the stabilized upper thermal boundary layer (p. 3191): "During the night of March 5 the entire, apparently stable crust of the lava lake foundered in a spectacular overturn" and "Crustal foundering was observed repeatedly during the eruption." These authors (pp. 3191–3193) also describe in some detail the subduction (foundering) mechanism:



**Figure 4.** Schematic illustration of an episodic subduction event. The process described for a Hawaiian lava lake is the model, but the process may also be applicable to Venus.

Most major foundering began at a point or line near the margin of the lake. The foundering propagated along a series of arcs of increasing circumference away from the point of origin. The beginning of foundering was marked by the appearance of a crack in the lake surface, which exposed fresh magma. Either from loading behind or from upward pressure in front of the crack a second fracture opened ahead of the first, and the crust bounded by the two cracks tilted upward along the edge bounded by the second crack and then slid backward beneath lava welling out of the first crack. Formation of a third crack ahead of the second was followed by tilting of the crust between the second and third cracks; the crust then slid beneath the new lava behind the second crack; and so forth.

A schematic illustration of this process is given in Figure 4.

Although there is no direct evidence for a global subduction event on Venus, there is evidence that the present lithosphere is on the verge of subduction. *Sandwell and Schubert* [1992a, b] proposed a model in which the large coronas on Venus are incipient circular subduction zones. The foundering lithosphere is replaced by ascending hot mantle in a manner similar to back arc spreading on the Earth.

The initiation of subduction is a long-standing problem in plate tectonics. The problem has been reviewed, and a new model has been proposed by *Fowler* [1993]. The simplest approach to the instability of a thermal boundary was given by *Howard* [1966]. Instead of applying the Rayleigh stability criterion to a convecting layer, he applied it to the boundary layer. Thus the boundary layer becomes unstable when

$$Ra_c = \rho_o g \alpha (T_m - T_s) y_L^3 / \mu \kappa \quad (4)$$

While it is certainly not clear that this result should be applicable to a mechanically rigid lithosphere, it is of interest to test it on the Earth. Taking  $\rho_o = 3300 \text{ kg m}^{-3}$ ,  $g = 10 \text{ m s}^{-2}$ ,  $\alpha =$

$3 \times 10^{-5} \text{ }^\circ\text{K}^{-1}$ ,  $T_m - T_s = 1300^\circ\text{K}$ ,  $\mu = 10^{21} \text{ Pa s}$ ,  $\kappa = 10^{-6} \text{ m}^2 \text{ s}^{-1}$ , and  $Ra_c = 1700$ , we find from (4) that  $y_L = 110 \text{ km}$ . For a conductively thickening thermal boundary layer the thickness is related to its age by

$$y_L = 2.32(\kappa t_L)^{1/2} \quad (5)$$

With  $y_L = 110 \text{ km}$  we obtain  $t_L = 100 \text{ Ma}$ . These estimates for the typical thickness and age of subducting lithosphere on the Earth are quite good. *Turcotte and Schubert* [1982, p. 166] estimate  $t_L = 121 \text{ Ma}$ .

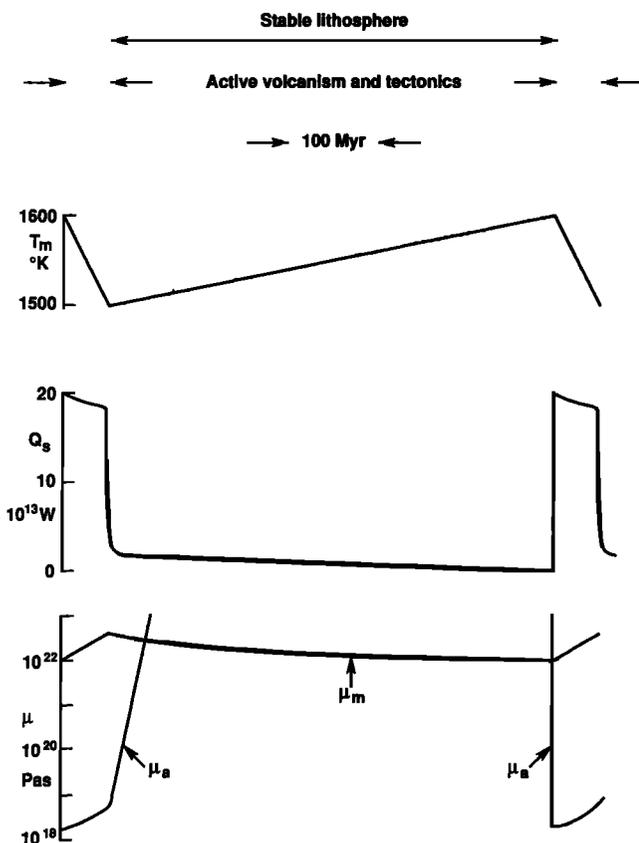
As an approximate model for episodic subduction on Venus we assume that the global lithosphere thickens conductively for 500 m.y.; basal heating of the lithosphere is neglected. The corresponding thickness of the subducted lithosphere from (5) is  $y_L = 290 \text{ km}$ . The mean heat loss due to the global subduction of this lithosphere is, from (3),  $Q_s = 1.23 \times 10^{13} \text{ W}$ . This represents 42% of the total required flux ( $2.91 \times 10^{13} \text{ W}$ ) during the interval. The 58% deficit must be made up during a period of active volcanism and tectonics. While the global lithosphere is stable, the only cooling to the interior is due to the heating of the previously subducted lithosphere. This heating is likely to be spread over several hundred million years. From the results given in Figure 2 the net increase in temperature of the mantle during the 500 m.y. of stable lithosphere is estimated to be about  $60^\circ\text{K}$ ; the corresponding decrease in mantle viscosity is about a factor of 5. We utilize (4) to estimate the mantle viscosity at the time of the global subduction event. With  $y_L = 290 \text{ km}$ , we find that  $\mu = 10^{22} \text{ Pa s}$ .

We assume that the period of rapid plate tectonics lasts for 50 m.y. In order to make up the deficit in heat loss during the period of lithospheric stability, the mean loss of heat is  $Q = 20 \times 10^{13} \text{ W}$ . From (3) this corresponds to an age of subduction  $t_s = 1.9 \text{ Ma}$ . From (5) the corresponding thickness of the subducted lithosphere is  $y_{LS} = 18 \text{ km}$ . And from (4) the mantle viscosity corresponding to this lithosphere thickness is  $\mu = 3 \times 10^{18} \text{ Pa s}$ . Thus the mantle viscosity must be a factor of 3500 larger during the time of global subduction than during the time of rapid plate tectonics. This must be the case even though the mean mantle temperature varies by less than  $100^\circ\text{K}$ .

The strong variation in mantle viscosity is easily attributed to the variation of viscosity with depth. At the time of global subduction the relevant viscosity is at a depth of 290 km; at the time of rapid plate tectonics the relevant viscosity is at a depth of 18 km. Since the mantle solidus has a much steeper gradient than the mantle adiabat, the mantle at a depth of 300 km is well below the solidus. This is the explanation for pressure release volcanism. The episodic model is illustrated schematically in Figure 5. *Sleep* [1994] has suggested that a hemispheric subduction event on Mars resulted in the hemispheric dichotomy on that planet.

## Conduction Model

As discussed previously, 13% of the Earth's heat loss is attributed to thermal conduction through the continental crust. Differentiation of the incompatible elements into the continental crust has led to a substantial enrichment of the heat-producing elements. Within the continental crust there is an upward enrichment so that most of the heat-producing elements are concentrated in the upper 10 km. Prior to the general acceptance of plate tectonics and mantle convection, it was



**Figure 5.** Schematic representation of a cycle of episodic subduction on Venus. The upper mantle temperature  $T_m$ , total surface heat loss  $Q_s$ , and viscosities of the asthenosphere  $\mu_a$  and upper mantle  $\mu_m$  are shown.

widely believed that virtually all the heat-producing elements in the Earth were concentrated in the continental crust. This upward concentration was required because conductive processes could not get the heat out of the mantle without melting it. The third hypothesis we consider for Venus is a similar upward concentration of its heat-producing elements into the crust of that planet.

In order to do this we must consider the constraints on the concentrations of the heat-producing elements both in the mantle of Venus and in its crust. Overall concentrations of uranium, thorium, and potassium in Venus are estimated from terrestrial values. Values for the crust of Venus are obtained directly from data collected by the Vega and Venera landers. Before doing this, however, the concentrations of the heat-producing elements in the silicate mantle of the Earth will be discussed. As upper and lower limits to the Urey number we take  $Ur = 0.8$  and  $0.6$ . Published estimates of the corresponding concentrations of the heat-producing elements and the rates of heat generation are given in Table 2. The values of mean heat generation for the bulk silicate Earth range from  $H = 7 \times 10^{-12}$  to  $5.2 \times 10^{-12} \text{ W kg}^{-1}$ . Mean values for chondritic meteorites are also given in Table 2. It is seen that the Earth is enriched in the refractory elements U and Th relative to the volatile element K. In the absence of other constraints we assume that the range of concentrations associated with the Earth are also applicable to Venus.

Mid-ocean ridge basalts (MORB) are taken as direct melt products of the Earth's mantle. However, the upper mantle is certainly depleted in incompatible elements relative to the bulk silicate earth due to the concentration of these elements in the continental crust. *Sun and McDonough [1989]* argue that normal (N type) MORB represents the melting of this depleted source region; as evidence they give the consistent depletion of

**Table 2.** Concentrations of Heat-Producing Elements and Rate of Heat Generation  $H$  for a Variety of Planetary Basalts and Source Rocks

	Uranium (U), ppm	Thorium (Th), ppm	Potassium (K), ppm	Th/U	K/U	Heat Production ( $H$ ), $10^{-12} \text{ W kg}^{-1}$	Reference
Chondrite	0.008	0.029	545	3.6	68,000	3.5	SM89
Bulk silicate earth ( $Ur = 0.8$ )	0.029	0.116	290	4.0	10,000	7	TS82
Bulk silicate earth ( $Ur = 0.6$ )	0.021	0.085	250	4.0	11,900	5.2	SM89
<i>Earth Basalts</i>							
N type MORB	0.047	0.12	600	2.6	12,800	10	SM89
E type MORB	0.18	0.60	2,100	3.3	11,700	41	SM89
OIB	1.02	4.20	12,000	4.1	11,800	255	SM89
<i>Moon Basalts</i>							
Low-Ti olivine (sample 12002)	0.22	0.75	415	3.4	1,900	43	H91
Low-Ti olivine (sample 15545)	0.13	0.43	330	3.3	2,500	25	H91
Low-Ti pigeonite (sample 12064)	0.22	0.84	580	3.8	2,600	46	H91
Low-Ti pigeonite (sample 15597)	0.14	0.53	500	3.8	3,600	30	H91
High-Ti, low K (sample 70215)	0.13	0.34	415	2.6	3,200	23	H91
High-Ti, high K (sample 10049)	0.81	4.03	3,000	5.0	3,700	197	H91
Low-Ti aluminous (sample 14053)	0.59	2.1	830	3.6	1,400	117	H91
<i>Venus Basalts</i>							
Vega 1	0.64	1.5	4,500	2.3	7,000	118	S87
Vega 2	0.68	2.0	4,000	2.9	5,900	134	S87
Venera 8	2.2	6.5	40,000	3.0	18,000	531	S87
Venera 9	0.60	3.65	4,700	6.1	7,800	172	S87
Venera 10	0.46	0.70	3,000	1.5	6,500	74	S87
Venera 13	...	...	33,000	...	...	...	S84
Venera 14	...	...	1,700	...	...	...	S84

References are as follows: SM89, *Sun and McDonough [1989]*; TS82, *Turcotte and Schubert [1982]*; H91, *Heiken et al. [1991, pp. 261–263]*; S87, *Surkov et al. [1987]*; and S84, *Surkov et al. [1984]*.

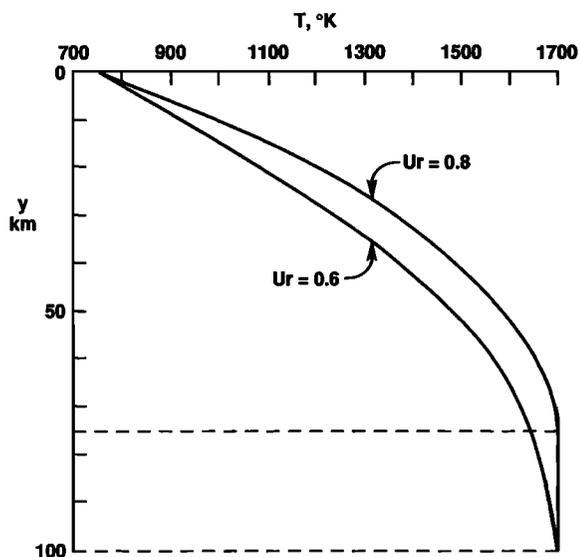


Figure 6. Temperature profiles through a thick crust with a uniform concentration of heat-producing elements and zero mantle heat flow.

the light rare earth elements in these rocks relative to chondrites. On the other hand, they point out that enriched (E type) MORB have generally chondritic rare earth distributions and should represent the partial melting of a source region that has near-bulk earth concentrations. The concentrations of the heat-producing elements in E type MORB as given in Table 2 are  $\approx 7$  times the estimated bulk earth values, leading to a reasonable 12% basaltic component in the undepleted mantle with complete transfer of incompatible elements to the liquid fraction. Also given in Table 2 are typical concentrations of heat-producing elements in ocean island basalts (OIB). Clearly, these basalts have enriched concentrations of the incompatible elements.

Before considering results for Venus, we turn to the Moon. Concentrations and heat production are given for seven mare basalts in Table 2. Certainly, lunar basalts are very complex, and there are significant differences between the interior of the Earth and the Moon. However, the more "primitive" lunar basalts (i.e., olivine and pigeonite) have concentrations of U and Th rather similar to E type MORB. The depletion in K is consistent with the depletion of all volatile elements in the Moon, presumably associated with the giant impact (or other processes) responsible for its formation.

The concentrations of heat-producing elements and heat production for five landing sites on Venus (Vega 1, 2 and Venera 8, 9, 10) are given in Table 2 [Surkov *et al.*, 1987]. Potassium values are given for two additional sites. Venera 8 and 13 sampled upland plains with weakly differentiated melanocratic alkaline gabbroids with high potassium content. Venera 14 sampled a lowland area covered with tholeiitic basaltic tuff with a low potassium content. Young shield structures were sampled by Venera 9 and 10 with lavas close in composition to tholeiitic basalts but with a calc-alkaline trend.

The Venus data are certainly less reliable than the laboratory studies of terrestrial and lunar rocks. However, the results are generally consistent with values expected for basic rocks. Five landers give values that can be associated with moderately radiogenic basaltic rocks, 2–3 times higher than the E type

MORB but lower than typical OIB. Two landers give values more typical of silicic rocks on the Earth. In terms of modeling, an essential question is whether the surface values are typical of crustal values at depth. Certainly, fractionation and crystallization are likely to lead to an upward concentration of the incompatible heat-producing elements. In the continents of the Earth the concentrations of the heat-producing elements decay exponentially with depth on a scale of 10 km. But it is impossible to estimate such variations in the crust of Venus.

As a limiting case we assume that all the heat-producing elements in Venus are concentrated uniformly in a crust of thickness  $y_c$ . Secular cooling is also neglected, so that the heat flow to the base of the crust is taken to be zero. In this limit the temperature distribution in the crust is [Turcotte and Schubert, 1982, p. 145]

$$T = T_s + (T_m - T_s) \left( \frac{y}{y_c} \right) \left( 2 - \frac{y}{y_c} \right) \quad (6)$$

where  $T_s$  is the surface temperature and  $T_m$  the uniform temperature of the mantle. In addition, we require

$$q_s = \rho_c H_c y_c \quad (7)$$

$$T_m - T_s = \frac{1}{2} q_s y_c / k_c \quad (8)$$

where  $H_c$  is the rate of heat production in the crust and  $q_s$  is the surface heat flow.

For a steady state heat balance and  $Ur = 0.8$  we require  $q_s = 50 \text{ mW m}^{-2}$ ; with  $Ur = 0.6$ ,  $q_s = 38 \text{ mW m}^{-2}$ . We also take  $T_s = 750^\circ\text{K}$ ,  $k = 2 \text{ W m}^{-1} \text{ }^\circ\text{K}^{-1}$ , and  $\rho_c = 2900 \text{ kg m}^{-3}$ . If the crust is thick, the temperature within it will exceed its liquidus (assumed to be undesirable). If the crust is thin, the heat production  $H$  will be large (exceeding the observed values). Solutions for the two cases in which the basal temperature approaches the liquidus ( $T_m \approx 1700^\circ\text{K}$ ) are given in Figure 6. For  $Ur = 0.8$  we have  $y_c = 75 \text{ km}$  and  $H_c = 230 \times 10^{-12} \text{ W kg}^{-1}$ , and for  $Ur = 0.6$  we have  $y_c = 100 \text{ km}$  and  $H_c = 130 \times 10^{-12} \text{ W kg}^{-1}$ . Comparing the rates of heat generation with the Venusian values given in Table 2, we see that the values for  $Ur = 0.6$  are generally consistent.

Thus it is possible to construct a model for the upward concentration of the heat-producing elements that has a mantle temperature below the solidus and rates of heat generation compatible with the surface observations. However, this does require extreme assumptions: (1) almost complete transfer of the heat-producing elements to the crust, (2) negligible secular cooling of Venus, and (3) uniform concentrations of the heat-producing elements through the crust.

Presumably, if this model is to be valid, the crust of Venus would have thickened with time with little crustal recycling. An expected consequence of this process would be the systematic depletion of the mantle heat-producing elements with time; this should lead to a reduction in the content of the heat-producing elements in the most recent volcanics and a gradual decay of volcanism with time. A valid question would be whether this type of decay could be consistent with the crater counts.

## Conclusions

The loss of heat from the interior of a terrestrial planet drives the surface tectonics and volcanism of the planet. On the Earth the plates of plate tectonics are thermal boundary layers

of mantle convection cells. The subduction of these cold plates is responsible for 70% of the heat transfer through the Earth's mantle. There is no evidence for active subduction on Venus, so an alternative mechanism for mantle heat transport must be provided. Three limiting cases have been considered in this paper.

The Earth is in a near steady state balance between the sources of heat, radioactive decay of U, Th, and K and secular cooling, and the surface heat loss. A uniformitarian hypothesis is that Venus also has a steady state balance. However, without active subduction the vertical transport of heat through the Venusian mantle must be accomplished either by ascending hot plumes or descending cold delaminated lithosphere. If the heat was transported by plumes, about 80 Hawaiian size plumes would be required. If the heat was transported by lithospheric delamination, a 60% delamination of the entire Venusian lithosphere would be required every 20 m.y. There is no evidence from Magellan data for either the required plume flux or the required delamination flux. Heat transfer considerations argue strongly against a uniformitarian model for Venus.

Mechanical considerations provide independent evidence against the uniformitarian hypothesis. This hypothesis requires a mean lithospheric thickness for the planet of about 50 km. With the high surface temperature, such a thin lithosphere is inconsistent with the high topography, large gravity anomalies, lack of crater relaxation, and large observed flexural rigidities.

Crater statistics suggest that a global volcanic resurfacing event occurred on Venus about 500 Ma. This suggests that a rigid global lithosphere then stabilized and has thickened conductively since. One hypothesis is that plate tectonics simply ceased at that time; however, without plate tectonics the interior temperature would increase. An alternative hypothesis is that subduction on Venus is episodic. Episodes of global catastrophic subduction are followed by periods of surface stability. Lava lakes in Hawaii are considered to be analogies for this process. These lakes experience episodic subduction events as the solid surface crust thickens and becomes gravitationally unstable.

According to the episodic hypothesis the global lithosphere becomes sufficiently unstable as it thickens so that a global subduction event occurs. This is followed by a period of active volcanism and tectonics with high surface heat loss that cools the interior. With the cooling of the interior the vigor of the volcanism and tectonics decreases, leading to the stabilization of the global lithosphere. This lithosphere thickens conductively, and the interior heats up until another global subduction event occurs. The coronas on Venus are taken as incipient subduction zones and as evidence that the lithosphere is on the verge of a global subduction event. Heat loss from the interior of Venus is easily explained by a global subduction event followed by an episode of extensive volcanism and tectonics. And the present thick lithosphere associated with the episodic hypothesis explains a variety of mechanical problems as discussed above.

A third hypothesis for Venus is that the heat-producing elements have been transferred to the crust and Venus is now a "dead" planet. Although available constraints cannot absolutely rule out this hypothesis, it is difficult to envision how a global volcanic resurfacing event can be consistent with a planet that was slowly "dying." The basic conclusion of this

paper is that the bulk of the presently available evidence favors the episodic subduction hypothesis.

An important question to answer is, Why does the Earth have plate tectonics and Venus does not? One suggestion is that while subduction can occur on Venus, seafloor spreading cannot. It is recognized that the temporal evolution of plate tectonics requires intraplate deformation. On the Earth, almost all this deformation takes place in the continental portions of the surface plates. The western United States is an example. The rheology of continental lithosphere is soft relative to oceanic lithosphere due to the silicic composition of the continental crust. Without continents, it is suggested that this intraplate deformation cannot occur on Venus, and thus Venus cannot have a global system of seafloor spreading centers which are necessary complements to subduction in plate tectonics. Without seafloor spreading, episodic global subduction events recycle the cold, unstable lithosphere.

An obvious question is, What happened on Venus prior to the last resurfacing? It is the basic theme of this paper that episodes of global subduction have occurred throughout much of the evolution of Venus. But are there any relict fragments of this early history? It is difficult to explain the high mountains of Ishtar Terra unless they are of silicic composition. They may be continentlike relicts of the past on Venus which were completely reworked during the resurfacing event but did not participate in the last global subduction because of their gravitational stability.

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