PROCESSES OF CRUSTAL FORMATION AND EVOLUTION ON VENUS: AN ANALYSIS OF TOPOGRAPHY, HYPSOMETRY, AND CRUSTAL THICKNESS VARIATIONS*

JAMES W HEAD

Department of Geological Sciences, Brown University, Providence, Rhode Island, U.S.A.

Abstract. On Venus, present evidence indicates a crust of predominantly basaltic composition and a relatively young average age for the surface (several hundreds of millions of years). Estimates of crustal thickness from several approaches suggest an average crustal thickness of 10–20 km for much of the lowlands and rolling plains and a total volume of crust of about 1×10^{10} km^3, approximately comparable to the present crustal volume of the Earth (1.02×10^{10} km^3). The Earth's oceanic crust is thought to have been recycled at least 10–20 times over Earth history. The near-coincidence in present crustal volumes for the Earth and Venus suggests that either: (1) the presently observed crust of Venus represents the total volume that has accumulated over the history of the planet and that crustal production rates are thus very low, or (2) that crustal production rates are higher and that there is a large volume of "missing crust" unaccounted for on Venus which may have been lost by processes of crustal recycling.

Known processes of crustal formation and thickening (impact-related magma ocean, vertical differentiation, and crustal spreading) are reviewed and are used as a guide to assess regional geologic evidence for the importance of these processes on Venus. Geologic evidence for variations in crustal thickness on Venus (range and frequency distribution of topography, regional slopes, etc.) are outlined. The hypothesis that the topography of Venus could result solely from crustal thickness variations is assessed and tested as an end-member hypothesis. A map of crustal thickness distribution is compiled on the basis of a simple model of Airy isostasy and global Venus topography. An assessment is then made of the significance of crustal thickness variations in explaining the topography of Venus. It is found that the distinctive unimodal hypsometric curve could be explained by: (1) a crust of relatively uniform thickness (most likely 10–20 km thick) comprising over 75% of the surface, (2) local plateaus (tessera) of thickened crust (about 20–30 km) forming less than 15% of the surface, (3) regions of apparent crustal thicknesses of 30–50 km (Beta, Ovda, Thetis, Atla Regiones and Western Ishtar Terra) forming less than 10% of the surface and showing some geologic evidence of crustal thickening processes (these areas can be explained on the basis of geologic observations and gravity data as combinations of thermal effects and crustal thickening), and (4) areas in which Airy isostasy predicts crustal thicknesses in excess of 50 km (the linear orogenic belts of Western Ishtar Terra, less than 1% of the surface).

It is concluded that Venus hypsometry can be reasonably explained by a global crust of generally similar thickness with variations in topography being related to (1) crustal thickening processes (orogenic belts and plateau formation) and (2) local variations in the thermal structure (spatially varying thermal expansion in response to spatially varying heat flow). The most likely candidates for the formation and evolution of the crust are vertical differentiation and/or lateral crustal spreading processes. The small average crustal thickness (10–20 km) and the relatively small present crustal volume suggest that if vertical crustal growth processes are the dominant mechanism of crustal growth, than vertical growth has not commonly proceeded to the point where recycling by basal melting or density inversion will occur, and that therefore, rates of crustal production must have been much lower in the past than in recent history. Crustal spreading processes provide a mechanism for crustal formation and

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evolution that is consistent with observed crustal thicknesses. Crustal spreading processes would be characterized by higher (perhaps more Earth-like) crustal production rates than would characterize vertical differentiation processes, and crust created earlier in the history of Venus and not now observed ("missing crust") would be accounted for by loss of crust through recycling processes. Lateral crustal spreading processes for the formation and evolution of the crust of Venus are interpreted to be consistent with many of the observations derived from presently available data. Resurfacing through vertical differentiation processes also clearly occurs, and if it is the major contributor to the total volume of the crust, then very low resurfacing rates are required.

Although thermal effects on topography are clearly present and important on both Venus and the Earth, the major difference between the hypsometric curves on Earth (bimodal) and Venus (unimodal) is attributed primarily to the contrast in relative average thickness of the crust between the two terrains on Earth (continental/oceanic; 40/5 km = 35 km, 8:1) and Venus (upland plateaus/lowlands; about 30/15 km = 15 km, 2:1) (35 − 20 km = a difference of 20 km). The Venus unimodal distribution is thus attributed primarily to the large percentage of terrain with relatively uniform crustal thickness, with the skewness toward higher elevations due to the relatively small percentage of crust that is thickened by only about a factor of two. The Earth, in contrast, has a larger percentage of highlands (continents), whose crust is thicker by a factor of eight, on the average, leading to the distinctive bimodal hypsometric curve.

Data necessary to firmly establish the dominant type of crustal formation and thickening processes operating and to determine the exact proportion of the topography of Venus that is due to thermal effects versus crustal thickness variations include: (1) global imaging data (to determine the age of the surface, the distribution and age of regions of high heat flux, and evidence for the nature and global distribution of processes of crustal formation and crustal loss), and (2) high resolution global gravity and topography data (to model crustal thickness variations and thermal contributions and to test various hypotheses of crustal growth).

1. Introduction

Venus and Earth have many similarities to each other relative to the smaller terrestrial planetary bodies but one of the major differences is observed in the global hypsometry. The Earth is characterized by two modes representing the continents and ocean basins, and Venus by a single distinctive mode slightly skewed toward higher elevations (Pettengill *et al.*, 1980; Masursky *et al.*, 1980). In an assessment of mechanisms of lithospheric heat transfer on Venus (Solomon and Head, 1982), Morgan and Phillips (1983) tested the hypothesis that conductive heat loss is an efficient heat loss mechanism and that most of the topography of Venus could result from spatially varying thermal expansion in response to spatially varying heat flow. They developed a model relating surface elevation to lithospheric thickness and heat flow and found that about 93% of the mapped topography of Venus could be explained solely by plausible lithospheric thickness variations and that about thirty-five hot spots could account for the heat loss of the planet. The remaining topography (at high elevations) could be accounted for by crustal thickness variations.

In this paper processes of crustal formation and evolution are examined and the role of crustal thickness variations in the production of topography is assessed. First, evidence for crustal composition, age, average crustal thickness, and total crustal volume on Venus is reviewed, and it is shown that the present crustal volume of Venus is comparable to that of the present Earth, and that unless rates of crustal production on Venus have been considerably less in the past, a volume of crust should have been produced that is not now accounted for ("missing crust"). Secondly, known processes of crustal formation and thickening are assessed for Venus. Thirdly, geologic evidence for variations in crustal thickness on Venus are outlined and specific examples of regions of apparent crustal thickening are assessed. These observations are compared to a simple model of Airy isostasy using global Venus topography and the end-member hypothesis that the topography of Venus could result solely from crustal thickness variations is examined. Finally, an assessment is made of the significance of crustal thickness variations in explaining the topography and regional geology of Venus. On the basis of these considerations and geologic observations, it is concluded that crustal thickness variations linked to crustal formational and modificational processes can account for many of the observed topographic variations. This factor, combined with regional variations in heat flux (lithospheric thickness variations) particularly in the Equatorial Highlands, can explain the distinctive hypsometric curve of Venus. Models to account for these observations are proposed and predictions are made which can be tested by further analysis and new data.

2. Estimates of Crustal Thickness on Venus

On the basis of data from *in situ* geochemical analysis of Venus surface materials in several different locations, surface rocks are interpreted to be predominantly basaltic in composition (Surkov *et al.*, 1984, 1987). Estimates of crustal thickness are derived from analysis of observed deformation wavelengths and modelling of crust/mantle rheology and geometry, and from theoretical considerations. Beginning with the knowledge that crust and mantle material have different rheological characteristics and that the crust of Venus is largely basaltic, and taking into account a range of plausible thermal gradients, various observed structures can be used to estimate the thickness of the crust. For example, on the basis of observed multiple wavelengths of tectonic features combined with theoretical models of deformation of the lithosphere, Zuber (1987) developed constraints on the near-surface thermal gradient, and using rheological models of the crust and upper mantle, estimated the thickness of the crust. Regions that exhibit multiple wavelengths of deformation are consistent with a lithosphere that consists of a relatively thin crust (strong near the surface and weaker at depth) underlain by an upper mantle that is much stronger than the lower crust. Regions that exhibit single wavelengths of deformation are consistent with a lithosphere characterized by a thick crust that does not contain a region of upper mantle strength or small strength contrasts in the lithosphere (Zuber, 1987). For a range of geologic features primarily in the lowlands and rolling plains, Zuber (1987) estimated an upper limit of crustal thickness of 30 km, and in a later study which incorporated growth rate similarity requirements, refined the upper limit of allowable crustal thickness to about 20 km (Zuber and Parmentier, 1990). Using observed impact crater depths and models of viscous relaxation of crater relief, Grimm and Solomon (1988) calculated thermal gradients and found that the observed crater depths require a layer of strength comparable to that of the mantle at relatively shallow depths. On the basis of this analysis, they derived upper bounds to crustal thickness of 10–20 km in regions of low to intermediate elevations. Other estimates in this range include 5–15 km from the geometry of tectonic features (Banerdt and Golombek, 1988), and less than 34 km from models of gravity-driven crustal decollements (Smrekar and Phillips, 1988).

Larger values for crustal thickness (generally in excess of 100 km) have been proposed or are permissible on the basis of: (1) parameterized convection models (Solomatov *et al.*, 1987) which are subject to boundary or initial condition assumptions, (2) gravity data, yielding apparent depths of compensation of long wavelength topography (Phillips *et al.*, 1981) which are thought to be overestimates due to components of dynamic compensation, (3) the depth of the basalt/eclogite phase change (Anderson, 1980) which is applicable only as an upper limit, and (4) the assumption that there is no crustal spreading and that crustal growth is predominantly vertical (Kaula, 1990).

In summary, on the basis of the analyses cited above that emphasize observations and interpretation of the present surface features of Venus (deformational structures – Zuber, 1987; Zuber and Parmentier, 1990; Banerdt and Golombek, 1988; and craters – Grimm and Solomon, 1988), average crustal thicknesses in the Venus lowlands and rolling uplands (the majority of the surface) most plausibly lie in the range of about 10–20 km, although higher estimates cannot be ruled out.

3. Processes of Crustal Formation and Thickening

Planetary crusts are of three principal types (Taylor, 1989): (1) *primary*, which form as a result of accretional heating (e.g., the lunar highland crust); (2) *secondary*, which form as a result of partial melting of planetary mantles (e.g., the lunar maria, and the terrestrial oceanic crust); and (3) *tertiary*, which form by the reprocessing of secondary crusts (Figures 1a, b). According to Taylor (1989), the Earth's continental crust is the only presently known example of a tertiary crust. Although a primary crust may have once existed on Venus, the young age of the majority of the surface presently observed with high resolution data (Ivanov *et al.*, 1986; Schaber *et al.*, 1987; Basilevsky *et al.*, 1987; Campbell *et al.*, 1989) and the strong evidence of partial melting and basaltic composition (Surkov *et al.*, 1984, 1987) for the rocks comprising these young surfaces, support the idea that at least a major part of the crust of Venus is secondary.

A. Crustal formation processes

Two types of processes are known to operate in the formation of secondary crusts which derive from partial mantle melting and vertical ascent of partial melt products. The first of these, *horizontal or lateral crustal growth*, is best exemplified

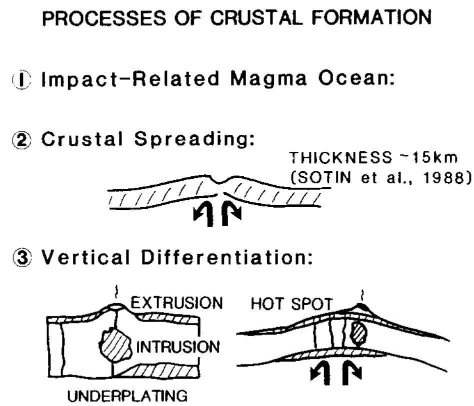

Fig. 1(a).

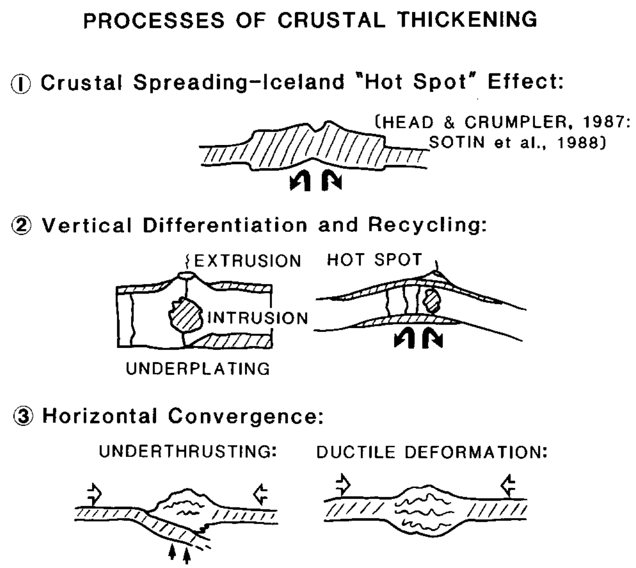

Fig. 1(b).

by terrestrial crustal spreading centers, in which melting beneath rise crests produces basaltic magmas and creates a crustal layer 5–6 km thick which spreads laterally, and together with a portion of the upper mantle forming the lithosphere, is usually subducted. On the Earth such a layer makes up 59% of the surface area and about 16.5% of the total present crustal volume (Table I) (Taylor and McClennan, 1985), and is presently created at rates of about 20 km^3/yr (Parsons, 1981). If the oldest rocks in the ocean basins are about 200 m.y. old, then at rates of 20 km^3/yr about 0.4×10^{10} km^3 of oceanic crust has been produced during this period. The present preserved volume

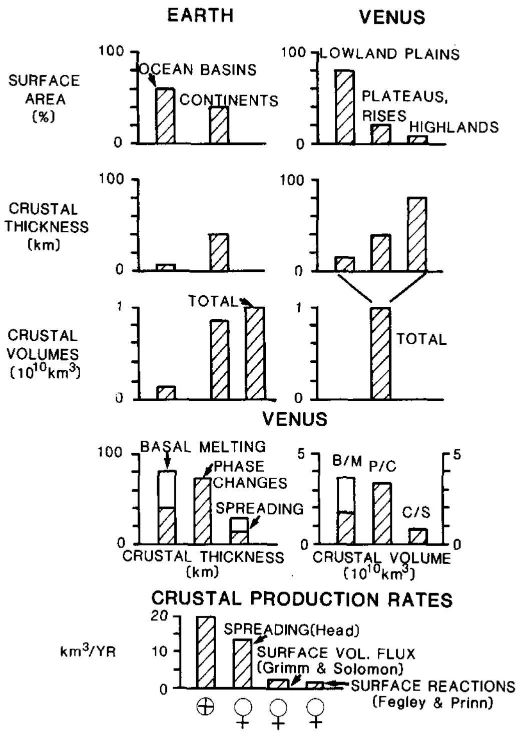

Fig. 1(c).

Fig. 1. Schematic illustrations of processes of crustal formation (a) and thickening (b); (c) Graphical summary of estimates of surface area, crustal thickness, crustal volumes, and various crustal production rates. The surface area of continents, ocean basins, and the three subdivisions of Venus terrâin are shown, and the estimated thickness of crust for each is illustrated. Total crustal volumes are also shown, and are approximately equivalent for the Earth and Venus. For Venus, crustal thicknesses for vertical recycling by basal melting (B/M) and phases changes (P/C) are shown (range of values depending on geotherm) and for lateral crustal spreading (C/S) estimates (Sotin *et al.*, 1989). Values of crustal volumes for these two estimates are also shown. Crustal volumes required for recycling by basal melting or phase changes are 2–4 times greater than presently observed crustal volumes. Crustal volumes characteristic of crustal spreading are approximately equal to presently observed crustal volumes. A range of crustal production rates are illustrated (see Table IV and text).

$$\begin{aligned}&(\text{Area of Earth's ocean basins; } 301 \times 10^6 \text{ km}^3)\\&\quad \times (\text{Average thickness; } 5.5 \text{ km}) = 0.17 \times 10^{10} \text{ km}^3\end{aligned} \tag{1}$$

is only 42.5% of the total crustal volume created in the last 200 m.y., implying that the other 47.5% (0.23×10^{10} km^3) has been subducted. Approaching it from a different perspective, we find that if no subduction is permitted, and the present area of the ocean floor is filled with 5.5 km thick oceanic crust at rates of 20 km^3/yr, generation of the presently observed oceanic crustal volume would be accomplished in 85 m.y.

$$\frac{(\text{Present volume of Earth's ocean crust; } 0.17 \times 10^{10} \text{ km}^3)}{(\text{Crustal production rate; } 20 \text{ km}^3/\text{yr})} = 85 \text{ m.y.} \tag{2}$$

TABLE I

Some values for crustal characteristics of Venus and Earth

Total Earth Crust:
Surface area: 510×10^{6} km^2
Total present volume: 1.02×10^{10} km^3

Terrestrial Oceanic Crust:
Surface area covered: 59%
Average crustal thickness: 5–6 km
Total present volume: 0. 17×10^{10} km^3 (B)
Percent total crustal volume: 16. 5%
Creation rate: 20 km^3/yr.
Volume produced in 200 m.y. at 20 km^3/yr: 0. 4×10^{10} km^3 (A)
Volume subducted in 200 m.y. (A–B): 0. 23×10^{10} km^3
Of the oceanic crust created in last 200 m.y., 57.5% has been subducted, 42.5% remains.
Creation time of the present volume of oceanic crust if created at 20 km^3/yr: 85×10^{6} yr.
Total volume created in 4×10^{9}yrs (recycled 20 times): 8.0×10^{10} km^3

Terrestrial Continental Crust:
Surface area covered: 41%
Average crustal thickness: 40 km
Total present volume. 0.84×10^{10} km^3
Percent total crustal volume: 83. 5%

Venus Crust:
Surface area: 460×10^{6} km^2
Present volume if average crustal thickness is 15 km: 0.69×10^{10} km^3
Total volume created in 4×10^{9} yrs (average age 500 m.y., recycled 8 times): 5.52×10^{10} km^3
Estimated average crustal production rate (present volume of average 15 km thick crust, average 500 m.y. old): 13.8 km^3/yr.
Mean volcanic flux (Grimm and Solomon, 1987): 2 km^3/yr (4 km/b.y. average thickness volcanic resurfacing rate).
Mean volcanic flux (Fegley and Prinn, 1989): 1 km^3/yr (2 km/b.y. average thickness volcanic resurfacing rate).
Mean volcanic flux from heat pipe mechanism (Turcotte, 1990): 200 km^3/yr (400 km/b.y. average thickness volcanic resurfacing rate).
Upper limit to present crustal volume (Grimm and Solomon, 1988): 1×10^{10} km^3
Crustal thickness constrained by basal melting: 40–80 km depending on conductive geotherm.
Crustal volume if this process is operating: $1.84–3.68 \times 10^{10}$ km^3
Crustal thickness constrained by phase changes (minimum): 75 km.
Crustal volume if this process is operating: 3.45×10^{10} km^3

If there were no continents on Earth, it would take about 140 m.y. to cover the planet with oceanic crust at this rate. If present conditions were typical over Earth history (present oceanic crust surface area and crustal production rate), than an extremely large volume of oceanic crust will have been produced during Earth history. For example, if 20 km^3/yr of crust were created over 4.5 b.y., the total volume of crust produced by this process would be 8.0×10^{10} km^3, 47 times the present crustal volume in the ocean basins. In other words, if the present area of the oceanic crust is produced in about 85 m.y., it could have been recycled 47 times in 4.5 b.y. Therefore, horizontal or lateral crustal growth (crustal spreading) can result in rapid renewal of the surface and can produce very large volumes of crust over the geologic history of a planet (Table I).

Theoretical assessment of terrestrial spreading centers under Venus conditions indicates that the major difference between the Earth and Venus would be the influence of the enhanced surface temperature on upper mantle temperatures on Venus, and the resulting increase in crustal thickness (Sotin *et al.*, 1989). Crust on Venus produced at spreading centers is predicted to be about 15 km thick (Sotin *et al.*, 1989; Hess and Head, 1990a), in contrast to the average crustal thickness on Earth of about 5–6 km. If such a crustal thickness (15 km) were typical of Venus as a whole, then the total present crustal volume would be about 0.69×10^{10} km^3 (Table I), somewhat lower than the upper limit to present crustal volume of 1×10^{10} km^3 proposed by Grimm and Solomon (1988). If crustal recycling processes operate as they do on the Earth (primarily by way of subduction), and the age of the surface of Venus is between 250 and 1000 m.y. (taken from estimates for about 35–40% of the surface of Venus; Ivanov *et al.*, 1986; Schaber *et al.*, 1987; Campbell *et al.*, 1989; Campbell and Head, 1990), then we can use present estimates of total crustal volume to calculate average rates of crustal production, and the total amount of crust that might have been produced over the history of Venus (Table II). If the total present crustal volume was produced in 250 m.y., then crustal production rates are 27.6–40 km^3/yr, and the total volume of crust produced in 4.5 b.y. is $12.4–18 \times 10^{10}$ km^3. If the total present crustal volume was produced in 1000 m.y., then crustal production rates are 6.9–10 km^3/yr, and total volumes are $3.1–4.5 \times 10^{10}$ km^3. Present crustal volume produced in 500 m.y. gives Earth-like production rates of 13.8–20 km^3/yr, and total volumes over the history of the planet of $6.2–9.0 \times 10^{10}$ km^3.

If average crustal spreading rates occurring on Venus are in the range of those interpreted for the Aphrodite Terra region (1–3 cm/yr full spreading rates; Sotin *et al.*, 1989; Crumpler and Head, 1988,1990; Crumpler, 1990), then a range of possible ridge lengths for various total ages of the crust can be calculated,

$$L = \frac{V}{T_c \times R_s \times A} \tag{3}$$

where L is total ridge length, V is total volume of the crust, T_c is crustal thickness, R_s is spreading rate, and A is age to production of the crust. Ridge length values

TABLE II

Total volume of crust produced over Venus history

	Present volume = 1×10^{10} km^3		Present volume = 0.69×10^{10} km^3 (ave. 15 km thick)	
Present crustal volume produced in:	Rate (km^3/yr)	Total volume $\times 10^{10}$ km^3 (in 4.5×10^9 yrs)	Rate (km^3/yr)	Total volume $\times 10^{10}$ km^3 (in 4.5×10^9 yrs)
250 m.y.	40	18.0	27.6	12.4
500 m.y.	20	9.0	13.8	6.2
1000 m.y.	10	4.5	6.9	3.1

for a 15 km thick crust covering the planet ($V = 0.69 \times 10^{10}$ km^3) and produced in 250 m.y., 500 m.y., and 1000 m.y., at rates of 1 and 3 cm are given in Table III. Values for spreading center length average 71×10^3 km and range from 15.3×10^3 km for crust produced in 1000 m.y. at rates of 3 cm/yr, to 184×10^3 km for crust created in 250 m.y. at 1 cm/yr. Total effective ridge length on Earth at present is about 60×10^3 km (Forsyth and Uyeda, 1975), close to the average of the estimates of ridge lengths for Venus. For comparison, evidence for crustal spreading has been presented for the 16×10^3 km length of Western Aphrodite Terra (Head and Crumpler, 1987; Crumpler and Head, 1989).

A second type of process known to operate in the formation of secondary crusts is *vertical crustal growth*. This is in contrast to lateral spreading and represents the *in situ* vertical differentiation and accumulation of melt products. Various aspects of vertical crustal growth might include extrusion onto a crustal surface, intrusion into the crust, or underplating onto the base of the crust (Figure 1a). Such processes of vertical crustal formation and thickening are linked to localized sources of melting and differentiation, and to the thermal gradient, which may be either normal, or anomalous, as at a hot spot (Figure 1a). If the present Earth oceanic crust was produced by such vertical processes rather than crustal spreading, and if the whole crust were produced in 85 m.y. (Equation 2; Table I), then the average vertical resurfacing rate would be about 65 m/m.y. or 65 km/b.y., producing a total crustal thickness of over 290 km over the history of the planet. An extreme example of the hot-spot case is the "heat pipe" mechanism proposed by Turcotte (1989) for Venus in which heat is lost advectively by magma transport through narrow "pipes", resulting in less conductive heat loss and a correspondingly thicker lithosphere. This mechanism (Turcotte, 1989) predicts extremely high crustal production rates (200 km^3/yr for Venus, compared to about 20 km^3/yr for Earth) and very high vertical resurfacing and crustal growth rates (435 m/m.y. and 435 km/b.y. globally). One of the major reasons that crustal production rates are so high on Earth is that plate tectonic processes and the present balance of forces in which slab pull dominates (Forsyth and Uyeda, 1975) create an environment favoring buoyant upwelling of mantle, melting, and crustal production at rise crests. If plate movement on Earth ceased, more heat would be lost by conduction and crustal production rates would decrease accordingly. For example, if the present oceanic crustal volume of the Earth were created in 4.5 b.y. solely by vertical differentiation processes, then the crustal production rate would be

TABLE III

Calculated ridge lengths (10^3 km, Equation 3) for $V = 0.69 \times 10^{10}$ km^3, $T_c = 15$ km

	Total age to production of crust (106 yrs)		
Spreading Rate (10^{-5} km/yr)	250	500	1000
1	184	92	46
3	62	30. 6	15.3

0.38 km^3/yr, less than one-fiftieth of the present rate. If the present total crustal volume of the Earth's crust (oceanic and continental) were produced by vertical crustal growth, than the average rate of crustal production would be 2.26 km^3/yr, about one-ninth the present rate. If the present Venus crust (for example, a global crust of 15 km average thickness; Table I) is produced by vertical crustal growth over the history of the planet, than crustal production rates would be 1.53 km^3/yr, and vertical crustal growth rates would be 3.3 m/m.y. and 3.3 km/b.y.

Since the types of sources and anomalies operating in vertical crustal recycling are local to regional in nature (e.g., vents, volcanoes, hot spots), global crustal thickening is limited by the ability of the magma to migrate laterally before crystallization, and thus requires a large number of sources operating nearly simultaneously, or changes in the location of sources over time in order to produce increased average crustal thickness.

B. Crustal thickening processes

Processes of crustal thickening can occur by variations on the above themes and by deformational mechanisms (Figure 1b). In the crustal spreading environment, anomalous crustal growth can occur when a hot spot coincides with a linear spreading center. In the case of Iceland, such a superposition has caused elevated upper mantle temperatures and enhanced melting, producing a crustal plateau of about 15–20 km thickness, a factor of 2–3 greater than the average oceanic crustal thickness (Bath, 1960; Bott, 1965; Palmason, 1971). The presence of numerous hot spots and oceanic plateaus on Earth (Nur and Ben-Avraham, 1982) suggests that Iceland is not an isolated example, and that this process has operated in the past both on rise crests and off. Upon the cessation of the mantle anomaly in the case of a fixed hot spot coinciding with a stationary spreading center, crustal thickness would return to normal, and the plateau would be split and separated by further spreading (Head and Crumpler, 1990; Crumpler and Head, 1990b). For the Earth, this phenomenon, both on and off spreading centers, also has a potentially important influence on the crustal thickening process related to horizontal convergence, because the collision of oceanic plateaus and continental crust is known to cause extensive deformation and accretion of oceanic crustal fragments onto the continents (Ben-Avraham and Nur, 1987).

On Venus, it has been suggested (Head and Crumpler, 1987) that several plateaus along the Western Aphrodite rise (Ovda and Thetis Regio) could be Icelandic-like plateaus superposed on a crustal spreading center. Sotin *et al.* (1989) showed that the gravity and topography data are consistent with a model in which Ovda Regio is an Iceland-like plateau being produced above a region of anomalously high temperature along a zone of crustal spreading. In this case, elevation of upper mantle temperatures of about 100°C would produce a crustal thickness of about 30 km, about a factor of two above the nominal crustal thickness at a Venus spreading center (Sotin *et al.*, 1989). Thus, plateaus with elevations of 2–4 km could be produced from the isostatic effects of increased crustal thickness associated with this crustal spreading mechanism (Figure 1b, top).

Vertical differentiation processes (extrusion, intrusion, underplating, hot spots; Figures 1a, b) may also operate to thicken and recycle the crust. For example, broad sources of mantle upwelling may cause melting and extensive surface volcanism and intrusion. More localized mantle plumes ('hot spots') could be distributed over the planet (Morgan and Phillips, 1983) and produce numerous localized additions to the crust, accumulating a global crust with time. If these processes persist, the crust may thicken sufficiently so that (1) it melts at the base, or (2) becomes negatively buoyant due to phase changes, and thus becomes recycled by either mechanism. In order to assess the significance of vertical recycling, it is important to derive estimates of present resurfacing rates, to calculate resurfacing rates required to match present estimates of crustal thickness, and to calculate thicknesses and rates required to produce recycling by basal melting or negative buoyancy.

Estimates of volcanic resurfacing rates (extrusional component) have been made for Venus (Table I; Figure 1b) using impact crater densities and characteristics (Grimm and Solomon, 1987). Grimm and Solomon (1987) determined an upper limit to the mean volcanic flux (consistent with the age, density, and morphometry of preserved impact craters) of 2 km^3/yr, one-tenth of the terrestrial crustal production rate (Tables I, IV). This is equivalent to an average global volcanic resurfacing rate in thickness of 4 km/b.y. If one assumes that this is equivalent to an average volcanic resurfacing rate over the history of the planet, this mechanism would produce a crust of only about 18 km thickness in 4.5 b.y., a thickness in the range of that of the current crust of Venus (10–20 km). Fegley and Prinn (1989) used SO_2 reaction rates, atmospheric observations and surface composition measurements to estimate the rate of volcanism on Venus. Their value (about 1 km^3/yr) is one-half

TABLE IV.

Crustal volume and thickness production rates

Mechanisms	km^3/yr	km/b.y.	Total thickness (km) in history of planet (4.5×10^9 yrs)
Heat Pipe Mechanism (Turcotte, 1989)	200	400	1800
Grimm and Solomon (1987)	2	4	18
Grimm and Solomon (1987) with 5:1 intrusion:extrusion ratio	10	20	90
Grimm and Solomon (1987) with 10:1 intrusion:extrusion ratio	20	40	180
Fegley and Prinn (1989)	1	2	9
Total present crustal volume (Grimm and Solomon, 1988)	2.2	4.8	21.6
Thickness of 40 km due to basal melting at 25°C/km	4.1	8.9	40
Thickness of 80 km due to basal melting at 15°C/km	8.2	17.8	80
Thickness of 75 km due to phase changes at 10°C/km	7.7	16.7	75

that of Grimm and Solomon (1987) and this rate would produce a crust of only about 9 km thickness over the history of the planet.

On the basis of an analysis of viscous relaxation of impact crater relief, Grimm and Solomon (1988) calculated an upper limit to the present total crustal volume of Venus of 10^{10} km^3. If this present volume represents the total volume of crust produced on Venus over the last 4.5 b.y. (which would include contributions from extrusion, intrusion, and underplating), then the average rate of crustal generation would be about 2.2 km^3/yr, and about 4.8 km of crustal thickness per billion years. Although this number (2.2 km^3/yr) is about the same as that derived as an upper bound for recent volcanic resurfacing rates (2 km^3/yr; Grimm and Solomon, 1987), the latter number measures only the extrusional component, and not contributions from intrusion and underplating. On Earth, ratios of intrusion to extrusion are about 5:1 for oceanic localities and about 10:1 for continental localities (Crisp, 1984). Applying these ratios to the Venus case using the upper bound for recent volcanic resurfacing rates (2 km^3/yr) calculated by Grimm and Solomon (1987) gives values for the 10:1 intrusion-extrusion ratio of 20 km^3/yr crustal production rates, 40 km thickness/b.y., and 180 km for the average thickness of crust produced over 4.5 b.y. (Table IV). The 5:1 intrusion-extrusion ratio gives values of 10 km^3/yr crustal production rates, 20 km thickness/b.y., and thus 90 km for the average thickness of crust produced over 4.5 b.y. These values correspond to total Venus crustal volumes produced over 4.5 b.y. of 4.5×10^{10} km^3 for the 5:1 case and 9×10^{10} km^3 for the 10:1 case. These values are far in excess of the present estimated crustal volume ($0.7–1 \times 10^{10}$ km^3) (Table I; Figure 1b). In summary, estimates of recent volcanic resurfacing rates are sufficient to produce a 10–20 km thick crust over the history of the planet, but require little to no intrusive component. If reasonable values of intrusion to extrusion ratios for Earth are used (5–10:1), the total volume of crust produced over the history of Venus would be very large (a factor of about 5–10 greater than that presently thought to exist), creating a volume of "missing crust" not presently observed.

Several possible explanations for this discrepancy exist and include:

(1) *Underestimation of crustal thickness*: A possible explanation for this discrepancy is that the estimates of 10–20 km crustal thicknesses over large areas of the planet are incorrect and too low, although this seems unlikely because of the variety of observations and techniques used (see initial discussion).

(2) *Variable rates of volcanism*: Another possible explanation is that recent average surface volcanic rates (Grimm and Solomon, 1987) are very high relative to the total history of Venus and that earlier rates were much lower. This seems unlikely in terms of the broad trends in the thermal evolution of the terrestrial planets in which high heat fluxes and extensive volcanism are favored earlier in the history of a planet (Toksoz *et al.*, 1978).

(3) *Incorrect intrusion/extrusion ratios*: A third possible explanation is that the ratios of intrusion to extrusion extrapolated from Earth (5–10:1) are incorrect and too high, and that the real ratio is closer to 1:1. This would mean that the present

rate of crustal production is inherently a factor of 10–20 lower than that of the Earth, and only the existing volume of crust has been produced over the history of the planet. Thermal evolution models favoring abundant volcanism and the lack of evidence of large and abundant crater remnants dating from early history argue against this explanation, although precise estimates of the ratio of intrusion to extrusion (Head and Wilson, 1986) have not been determined. Preliminary estimates suggest that the ratios will tend to be lower than on Earth (Wilson and Head, 1990).

(4) *Presence of crustal recycling*: Another possible explanation is that the currently observed crustal volume is not equivalent to the total crustal volume produced over the last 4.5 b.y. and that something of the order of 5–10 times the presently observed crustal volume has been destroyed and/or recycled. Mechanisms of crustal recycling are undetermined, but they should be consistent with present crustal thickness estimates of about 10–20 km. This would appear to disfavor global recycling mechanisms linked to vertical thickening leading to melting and/or density inversion, which require average crustal thicknesses well in excess of 20 km (Figure 1c).

(5) *Crust largely produced by crustal spreading processes*: A fifth possible explanation is that the crust is not being produced by surface volcanism (Fegley and Prinn, 1989) and vertical resurfacing (Grimm and Solomon, 1988) alone, but rather is being produced primarily by horizontal crustal growth (crustal spreading). If this were the case, then the vast majority of the crustal volume would be produced in the subsurface below rise crests and moved laterally during crustal spreading. Surface volcanism would reflect the emplacement of an extrusive veneer on this basic 15 km thick crust. Measurements of the mean volcanic flux of 2 km^3/yr from the density of preserved impact craters (Grimm and Solomon, 1987), or about 1 km^3/yr using SO_2 reaction rates (Fegley and Prinn, 1989), would only represent this veneer and would considerably underestimate the total crustal production rate, perhaps by as much as a factor of 5–10 (Figure 1c).

Another approach to estimating the significance of vertical thickening and recycling is to assess what rates of thickening are required to reach thicknesses in which recycling due to melting or density inversion occurs. Whether melting or density inversion occurs is dependent on the conductive geotherm (Anderson, 1980; Hess and Head, 1989,1990b). For a conductive geotherm of 25°C/km, the basalt liquidus is crossed at about 40 km depth (Figure 1c) (Hess and Head, 1990b). If a thickness of crust equivalent to this depth was produced over the history of the planet, then the crustal production rate would be 4.1 km^3/yr and the thickness production rate would be about 8.9 km/b.y. (Table IV). In this case, any vertical recycling by melting would require crustal thickness production rates in excess of 8.9 km/b.y. Crust forming in a conductive geotherm of 15°C/km would undergo melting before significant phase changes could occur (Hess and Head, 1989, 1990b). In this case, the liquidus would be approached at about 80 km (Figure 1c), implying crustal production rates of 8.2 km^3/yr and crustal thickness production rates of about

17.8 km/b.y. to reach the liquidus, and rates in excess of this to initiate crustal loss by basal melting (Table IV). If one of these two cases exists, that is, vertical accumulation, basal melting, and recycling, the question of the fate of the melted crust and the nature of processes of continued vertical recycling remain to be determined.

Basalt crust forming in a conductive geotherm of 10°C/km is likely to encounter phase changes before melting and to undergo density changes to granulite at about 40 km and to eclogite at about 75 km (Hess and Head, 1989, 1990b). Such density changes would contribute to negative buoyancy and to processes of crustal loss (e.g., negative diapirism, delamination). In this case, thicknesses in excess of about 75 km (Figure 1c) are required to initiate crustal loss due to density inversion, and this implies a crustal thickness production rate of at least 16.7 km/b.y. to initiate this process (Table IV). In summary, crustal thickness values for the range of conductive geotherms and for crustal loss by basal melting (>40–80 km) or density inversion (>75 km) are well in excess of the presently favored average crustal thickness values (10–20 km) discussed above (Figure 1c). Although these processes of vertical recycling are apparently not favored for the average crust because of the requirement for a very thick global crust, they may well be operating on a more local scale in areas of local convergence and crustal thickening as in Ishtar Terra (Vorder Bruegge and Head, 1989, 1990a, b; Burt and Head, 1989).

Other processes of crustal thickening are associated with horizontal crustal convergence (Figure 1b–3). On Earth, underthrusting, subduction and crustal thickening commonly occur at convergent plate margins and the degree and style of thickening is linked to the type of convergent boundary (Uyeda, 1982), the type of crust, and its thermal structure. Crustal thickening can involve underthrusting, crustal imbrication, ductile deformation, melting and remobilization of crustal material, and combinations of these. On Earth, significant thickening takes place in linear orogenic belts, such as the Andes (Megard, 1987; Kono *et al.*, 1989), or in more distributed regions, such as the Tibetan Plateau (Molnar and Tapponier, 1975). Crustal thickening linked to horizontal convergence can take place through ductile deformation and underthrusting, and does not require subduction (Molnar, 1988). At any one time only a very small percentage of the surface of the Earth is undergoing these types of crustal thickening processes. These processes can produce localized increases in the thickness of the crust but must operate over a long period of time, and change location from one place to another in order to increase the average crustal thickness. On the Earth, continental crustal formation processes (many of which are linked to horizontal crustal convergence) have operated over geologic time to produce a crust of 40 km average thickness covering about 41 percent of the planet (Taylor and McClennan, 1985), producing a total continental crustal volume of 0.84×10^{10} km^3 (Table I).

In summary (Table IV), observed and estimated rates of recent volcanic resurfacing on Venus (2–4 km thickness/b.y.) could produce a global crust in the thickness range of 10–20 km over the history of the planet, but would operate with an average global crustal production rate far below that of the Earth at present. In order to produce crustal thickness sufficient to begin crustal recycling by melting (>40–80 km) or density inversion (>75 km), crustal formation and thickening processes associated with vertical differentiation require rates of crustal addition (about 9–18 km thickness/b.y.) that are very high relative to observed rates of volcanic resurfacing (2–4 km/b.y.) (Tables I, IV). In addition, such rates and thicknesses are far in excess of the presently estimated crustal thickness discussed above (10–20 km). Lateral crustal spreading processes on Venus could produce crustal thicknesses (about 15 km; Sotin *et al.*, 1989) in the range of that presently thought to characterize Venus (10–20 km). Crustal spreading could also produce a global crust of that thickness in about 500 m.y., assuming 1.5 cm/year spreading rate and about 60×10^3 km ridge length (Table III). A crustal spreading process requires mechanisms for significant crustal recycling (e.g., subduction, delamination) if it is operating over the history of the planet. Another process that provides many mechanisms for crustal thickening is horizontal convergence, but these tend to be local (orogenic belts) or regional (distributed ductile thickening), rather than global. In order to distinguish between these various possibilities, it is important to analyze evidence for crustal thickness variations on the surface of Venus and to assess the geologic processes operating in those areas.

4. Evidence for Variations in Crustal Thickness on Venus

There are several methods by which crustal thickness variations can be detected and estimated. The range and frequency distribution of topographic elevations (hypsometry) can be analyzed on the assumption that the topography of Venus has no thermal component but is completely linked to Airy isostasy in the form of crustal thickness variations. In this case, assuming a basaltic crust that is 15 km thick at 0 elevation, the 13 km range in topography observed on Venus (Masursky *et al.*, 1980) would suggest that crustal thicknesses range from less than 5 km to about 110 km (Figure 2). Obviously, the greater the initial crustal thickness, the less significant are the variations. Assuming a basaltic crust that is 50 km thick at 0 km elevation, the 13 km range suggests crustal thicknesses ranging from 40 to 145 km.

The frequency distribution of elevations (hypsometry) (Figure 3a) shows that although the range of elevations on Venus is comparable to the Earth, the distribution is not. On the Earth (Figure 3b), the bimodal distribution reflects variations in crustal thickness and composition between the continents and ocean basins. The thickness of the continental crust, not its composition, is the major factor in its topographic distinctiveness from the ocean basins. A continental crust with an average thickness of 40 km and a 'sialic' composition (density of 2.8 g/cc^3) would only stand about 1.24 km above the isostatic elevation of a 40 km thick continental crust of basaltic composition (density of 2.9 g/cc^3). This is only 22% of the total difference in elevation between 5.5 km thick oceanic (simatic) crust and 40 km

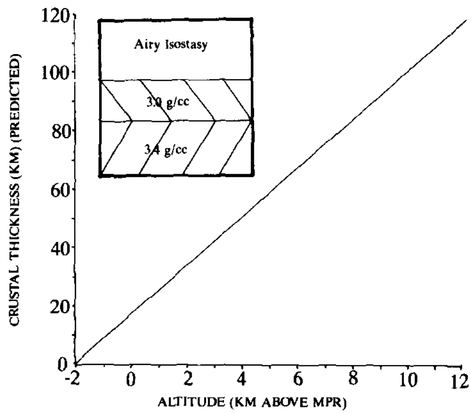

Fig. 2. Crustal thickness and isostatic topography based on an Airy isostatic model. Crust at mean planetary radius is assumed to be 15 km thick. Modified from Vorder Bruegge and Head (1989).

thick continental (sialic) crust in Airy isostatic equilibrium. Therefore, a continental crust of the same basaltic composition as the oceanic crust would still produce a bimodal hypsometric curve, simply because of the pronounced difference in crustal thickness between the two provinces and the corresponding isostatic elevation differences. Observed variations in the terrestrial ocean basin peak (Figure 3b) primarily reflect the influence of the evolving oceanic thermal boundary layer on the small and relatively uniform crustal thickness. For example, if the oceanic crust ceased spreading and came to thermal equilibrium, the peak should be much narrower, reflecting a uniform crust of 5–6 km thickness, skewed slightly to higher elevations by the presence of oceanic plateaus. Variations in the terrestrial continental peak reflect the planation effect of erosion which produces a peak centered near sea level, and crustal thickness variations ranging up to about 70 km primarily at active orogenic belts (Andes, Himalayas), which produce skewness toward higher elevations.

Although the unimodal nature of the Venus hypsometric curve says nothing per se about composition and/or major processes (e.g., presence or absence of plate tectonics), it can be used to test the feasibility of various mechanisms for the formation and support of topography. For example, Morgan and Phillips (1983) showed that the majority (>93%) of the Venus hypsometric curve could be explained by areal variations in heat flow, and that only about 7% (primarily Ishtar Terra) could not be explained by this mechanism and thus required crustal thickness variations. If we use the same end-member approach to test the feasibility of crustal thickness variations, it is obvious that crustal thickness variations can in principle account for the whole spectrum of topography. The question then becomes whether or not such an interpretation is reasonable and consistent with the geologic observations. If we assume that the crust is basaltic in composition (Surkov *et al.*, 1984, 1987) and that the mean planetary radius is equivalent to a crustal thickness of 15 km, then the observed topographic range is equivalent to

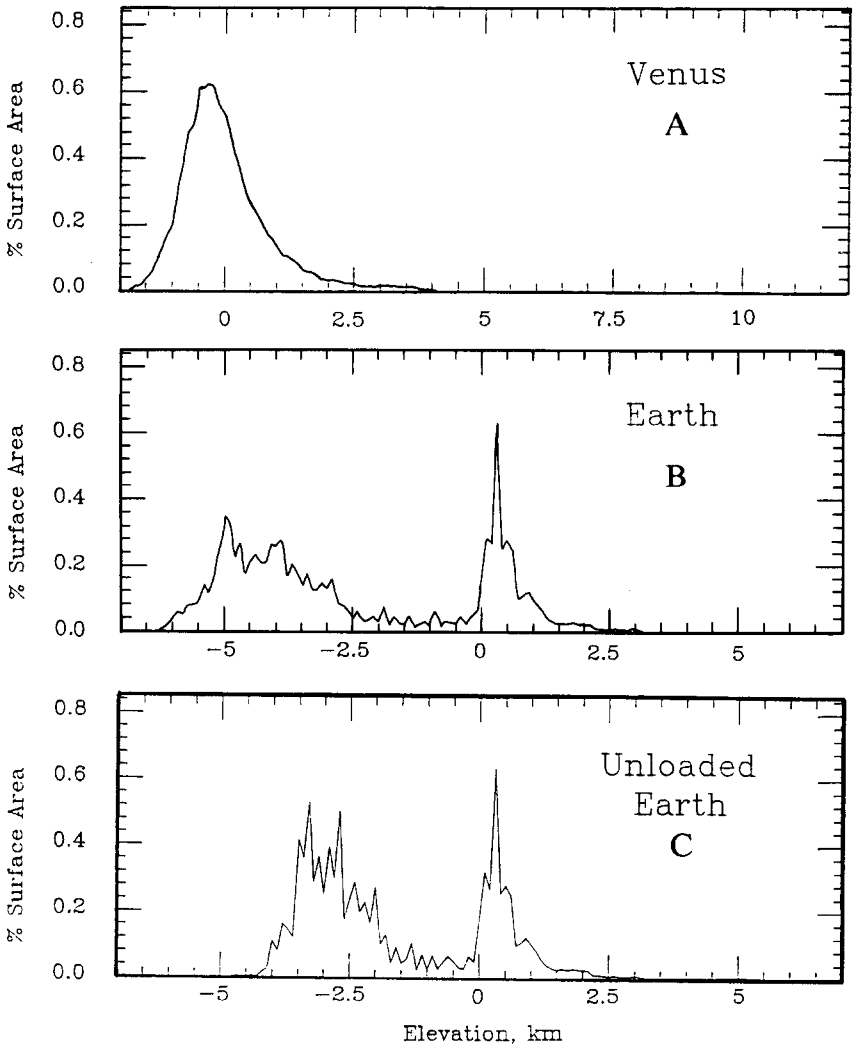

Fig. 3. Frequency distribution of topography on Earth and Venus. (a). Altitude frequency distribution for Venus. (b). Altitude frequency distribution for Earth. (c). Altitude frequency distribution for Earth, with the oceanic load removed. Modified from Sharpton and Head (1985, 1986).

a crustal thickness of 5 to about 110 km, with over 80% of the crust lying in the range of 10–20 km thickness. If we assume that the mean planetary radius is equivalent to a crustal thickness of 50 km, then the observed topographic range is equivalent to a crustal thickness of 40 to 145 km, with most of the crust lying in the range of 45–55 km thickness. For comparison, about 59% of the Earth's crust lies within the range of 5–6 km thickness (oceanic), and 41% averages 40 km in thickness (continental). In summary, crustal thickness variations can in principle explain much of the frequency distribution and character of the hypsometric curves of Earth and Venus (Head, 1990c,d). In order to assess specific models for the distribution of crustal thicknesses it is important to examine the geologic observations.

The hypsometric curve interpreted in terms of crustal thickness provides one approach to assessing crustal thickness variations as a mechanism for explaining the topography of Venus, but it portrays the topography in a global and statistical sense. Another approach is to assess potential crustal thickness differences in terms of the regional distribution of topography and then use additional data (slopes, geologic characteristics) to analyze the plausibility of crustal thickness variations as the sole cause of the topography. One indication of the location of changes in crustal thicknesses is distinctive variations in regional slopes (such as at the continental crust/oceanic crust boundary on Earth). Regional topographic slopes are a measure of the rate of change in topographic elevation over a several hundred km baseline, and thus are useful for locating areas where various processes (thermal, crustal thickness variations, etc.) are producing anomalous topography. Maps of regional slope for Venus and Earth (Sharpton and Head, 1985, 1986; Moore and Mark, 1986) show that the most distinctive and steepest slopes on Earth are at passive and active continental margins (linked to crustal thickness differences between the continents and ocean basins) and at subduction zones. On Venus, the steepest regional slopes are located around the margins of Western Ishtar Terra, and in the tessera of Eastern Ishtar Terra. Linear trends in slope changes are also found flanking western Aphrodite Terra, Beta Regio, Alpha Regio, and several tessera regions (e.g., Tellus Regio). These regions correspond to the distinctive uplands and highlands of Venus (Figure 4a) (Aphrodite, Beta, Ishtar, and the tessera regions).

Another approach is to use the areal distribution of topography to assess the end-member model of topography linked solely to crustal thickness variations (Airy isostasy). An Airy isostasy crustal thickness map of Venus (assuming a basaltic crust of 15 km thickness at mean planetary radius) (Figure 4b) shows that the vast majority of the lowlands and upland rolling plains (>75% of the surface) would be characterized by relatively uniform, thin (<20 km) crust. Regions of 20–30 km thick crust (<15% of the surface) correspond to the flanks of Aphrodite and Ishtar Terrae, Beta, Phoebe, and Atla Regiones, and to regions of tessera (Tellus, Tethus, Laima, Alpha, etc.). Regions of 30–50 km thick crust (<10% of the planet) correspond to central Beta, Atla, Ishtar and Aphrodite. Regions of crust in excess of 50 km thickness comprise only a very small percentage of the surface (<1%) and are concentrated in the linear mountain belts of western Ishtar Terra, and in a few spots in western Aphrodite Terra. We now proceed to analyze these regions of different hypothesized crustal thickness to see if the geologic evidence is consistent with crustal thickness variations, or whether other mechanisms for the formation of the topography are required.

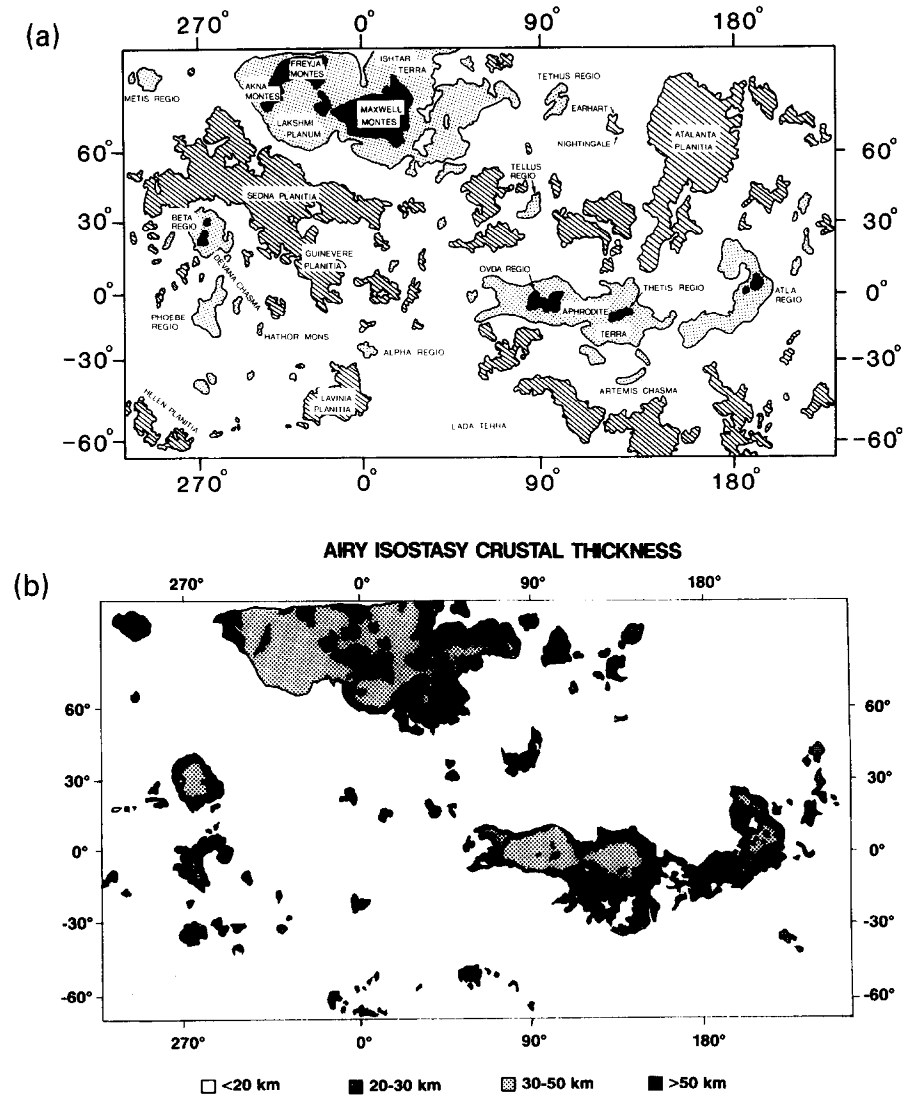

Fig. 4. Venus altimetry and crustal thickness. (a) Mercator map of topography and major features on Venus. Mean planetary radius is 6051. 9 km. Patterns are: obliquely ruled, −0.4 km below mean planetary radius (MPR) and below; white, −0.4 below to +1.6 km above MPR; +1.6 km to 3. 6 km above MPR, and black, >3. 6 km above MPR. Modified from Head *et al.*, 1985. (b) Venus crustal thickness map compiled using assumption of Airy isostasy and 15 km thick crust at mean planetary radius (6051.5 km). Crustal thickness of <20 km (white); 20–30 km (gray); 30–50 km (stippled); in excess of 50 km (black).

5. Analysis of Regions of Different Crustal Thickness

A. Ishtar Terra. Ishtar Terra is a distinctive region of apparent enhanced crustal thickness (Figure 4b) and is the location of the 7% of Venus topography that Morgan and Phillips (1983) could not account for by variations in thermal structure and attributed to crustal thickening processes. The regions in excess of 50 km modelled crustal thickness (MCT) (Figure 4b) are the linear mountain belts Akna, Freyja, and Maxwell in Western Ishtar Terra, structures interpreted to be orogenic belts (Crumpler *et al.*, 1986) and the loci of regional compressional deformation forces (Basilevsky *et al.*, 1986; Head, 1990a; Vorder Bruegge and Head, 1989, 1990a). On the basis of the geologic features indicative of compressional deformation mapped in these mountain belts there is ample evidence to indicate tectonic crustal thickening processes such as underthrusting, crustal imbrication, and ductile thickening. There is no known evidence (e.g., volcanic edifices, etc.) that thermal processes are directly responsible for the formation of the elevations associated with these mountain belts. Based solely on Airy isostasy assumptions, crustal thickness would be up to 65 km for Akna, up to 75 km for Freyja, and up to 110 km for Maxwell Montes. The values for Akna and Freyja are comparable to maximum values typical of terrestrial orogenic belts (see summary in Meissner, 1986). Although the topography of Western Ishtar Terra can thus reasonably be interpreted in terms of crustal thickness variations, gravity data indicate an apparent depth of compensation of 130 km (Sjogren *et al.*, 1984; Grimm and Phillips, 1990), and these data, together with the presence of volcanic edifices and deposits in Lakshmi Planum (Roberts and Head, 1990a, b), suggest that other thermal compensation factors are also involved.

Eastern Ishtar Terra lies at slightly lower elevations with the majority of it (Fortuna Tessera) in the range of 20–30 and 30–50 km MCT. Structural and tectonic patterns mapped in this region show trends from simple tessera terrain at low elevations in the east, through increasingly complex deformation patterns and intermediate elevations toward the west, culminating in the 11 km high Maxwell Montes orogenic belt on the western margin of Fortuna Tessera. These morphologic and tectonic trends (Figure 5) have been interpreted to be due to east to west convergence and compression, and the increasing elevations are interpreted to be due to crustal thickening processes associated with the convergent deformational environment (Vorder Bruegge and Head, 1989, 1990a, b). In this interpretation, plateaus of tessera terrain have converged with the edge of Western Ishtar Terra, producing a broad region of distributed deformation and crustal thickening in Western Fortuna Tessera (Figure 5). Although volcanic plains are seen in the adjacent lowlands, there is no evidence for extensive volcanism associated with Fortuna Tessera, nor evidence for a thermal origin for the major topographic and tectonic elements. The observed distributed deformation and crustal thickening is in contrast to the linear orogenic belts of Western Ishtar Terra, but is thought to involve similar crustal thickening processes. Lakshmi Planum, in the center of Western Ishtar Terra, shows evidence for central volcanic structures and resurfacing by volcanic flows. These observations led some earlier workers (Pronin, 1986) to propose a hot-spot origin for the plateau as a whole, but other analyses have interpreted the planum to be due to crustal thickness variations (Morgan and Phillips, 1983) and have shown that it is likely to be a block of tessera capped

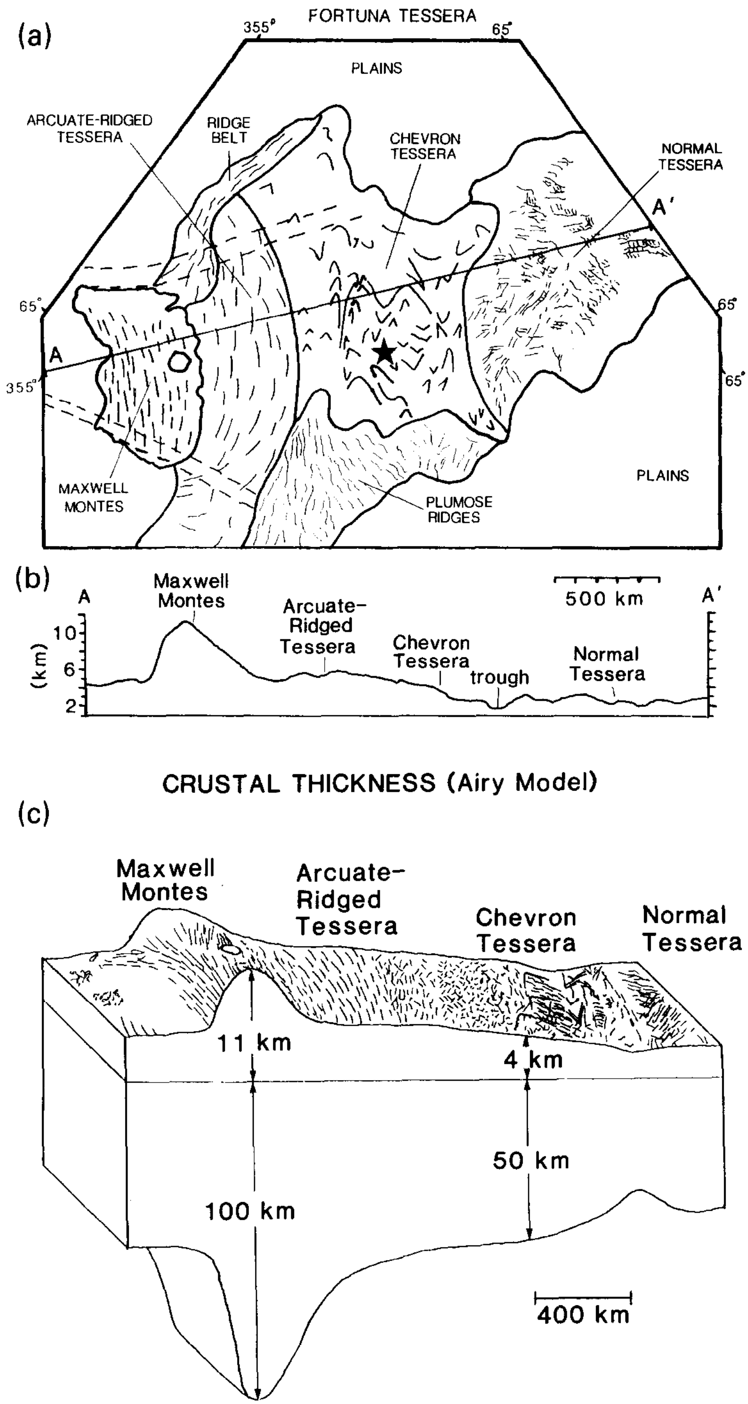

Fig. 5. Geomorphic map of Fortuna Tessera in Eastern Ishtar Terra showing increase in topography and structural complexity from west to east (a, b), and the interpreted increase in crustal thickness (c) (see also Figure 4b). Horizontal line is mean planetary radius.

by a thin veneer of volcanic deposits (Roberts and Head, 1990). Gravity data (Sjogren *et al.*, 1984; Grimm and Phillips, 1990) yield an apparent depth of compensation that requires at least some thermal component to the topography. In summary, there is abundant evidence for orogenic belts and convergence of plateaus of thickened crust (tessera terrain) and associated crustal thickening processes. These factors are interpreted to be important in producing the topography and large modelled crustal thicknesses typical of Ishtar Terra, although other factors are also contributing to the topography.

B. Aphrodite terra

Aphrodite Terra, a 16,000 km long highland forming part of the Equatorial Highlands (Phillips and Malin, 1983), shows distinct contrasts to Ishtar Terra in terms of topography and geological structure. Aphrodite Terra, comprising <10% of the planet, is characterized by 20–30 km MCT (primarily Eastern Aphrodite Terra) and 30–50 km MCT (primarily Western Aphrodite Terra). It does not have the linear orogenic belts, nor evidence for the extensive compressional deformation seen in Ishtar Terra. Instead, Aphrodite Terra has a system of regional troughs parallel to the linear rise which are interpreted to be rifts of extensional origin (Schaber, 1982). A series of linear cross-strike discontinuities (Crumpler *et al.*, 1987), together with topographic symmetry across the rise (Crumpler and Head, 1988) and split and separated topography, are evidence of crustal spreading. Head and Crumpler (1987) predict spreading rates of 1–3 cm/yr. They further interpreted the topography of Eastern Aphrodite Terra to represent a normal spreading center, and the plateaus of Western Aphrodite Terra to represent a mantle plume or hot spot superposed on normal spreading to produce an Iceland-like plateau of thickened crust. Other workers have proposed on the basis of gravity and topography data that Aphrodite Terra represents the location of upwelling mantle plumes or hot spots (Kiefer *et al.*, 1986; Herrick *et al.*, 1989), with no crustal spreading.

In summary, there is abundant evidence for a thermal contribution to the topography in Aphrodite Terra, and in fact thermal sources can account for the majority of the topography in the region (Morgan and Phillips, 1983). Is there any evidence then for crustal thickness variations? Additions to the crust could come from extrusion, intrusion, and underplating (Figure 1a) associated with thermal uplift. In addition, if crustal spreading dominates Aphrodite Terra, crustal formation and modification processes can be modelled and their influence and contribution to topography can be assessed. Sotin *et al.* (1989) have mapped typical crustal spreading environments to Venus conditions and have shown that the Venus crust in a normal spreading environment would be about 15 km thick due to the elevated upper mantle temperatures. They then examined the gravity and topography of Ovda Regio in Western Aphrodite Terra and showed that the data were consistent with a model of crustal spreading producing a 15 km thick crust, with a superposed region of elevated upper mantle temperatures (about 100°C) causing enhanced melting and producing an Iceland-like plateau of elevated crustal thickness of about 30 km. In this context, the areas of 20–30 km MCT in Eastern Aphrodite Terra would be interpreted as thermal topography associated with the rise crest region, and the flanking lowlands would be crust less than 20 km thick formed by the spreading process and moved laterally into regions of thermal equilibrium so that their topography only represents the crustal thickness component. The plateau-like areas of 30–50 km MCT in Western Aphrodite Terra (Ovda and Thetis Regio) would be interpreted as crust of about 30 km thickness produced at the rise crest and spread laterally as Iceland-like plateaus, with an additional topographic component coming from the thermal structure of the rise crest. The flanking areas of crust less than 20 km thick would have derived from earlier spreading that is now near thermal equilibrium. In summary, Aphrodite Terra topography appears to be due to thermal contributions, and to crustal thickness variations linked to these thermal sources. A crustal spreading model (Head and Crumpler, 1987; Sotin *et al.*, 1989) predicts that the crustal thicknesses in normal spreading areas should be about 15 km, and in the plateau regions, 30 km, with the remainder of the rise topography due to thermal contributions. Beta and Atla Regiones, located along the Equatorial Highlands (Figure 4), show abundant evidence for thermal contributions on the basis of their great apparent depths of compensation, and the presence at Beta of flanking plateaus of tessera terrain (Campbell *et al.*, 1990; Senske *et al.*, 1990) suggests possible contributions from crustal thickness variations.

C. Other regions

The remainder of the regions of enhanced modelled crustal thickness (<15% of the surface) are distributed as patches of 20–30 km thick crust ranging in area from several hundreds of km^2 to tens of thousands of km^2. The geologic characteristics of these plateau-like areas show that they are comprised of: (1) tessera (e.g., Tellus, Tethus, Alpha), (2) domal rises (Bell, Metis), which often have associated rifting and volcanism and appear to be due to contributions from thermal anomalies (Janle *et al.*, 1987), and (3) regions of unknown affinity, the majority of which have radar properties similar to the tessera (Bindschadler *et al.*, 1990). Thus, tessera terrain makes up the vast majority of regions in this range of crustal thickness. A number of modes of origin have been proposed for the tessera regions (reviewed in Bindschadler and Head, 1988,1989) including horizontal compression and crustal thickening, vertical uplift and associated deformation, gravity sliding, and sea-floor spreading. Recently, Head (1990b) has presented evidence that one of the distinctive tessera types (trough-and-ridge) has many similarities to the orthogonal fabric of fracture zones and abyssal hills formed at sea-floor spreading centers on Earth. These observations, and the similarity of the radar characteristics of the Western Aphrodite plateaus to known tessera regions (Bindschadler *et al.*, 1990) has led to the hypothesis that tessera terrain represents regions of enhanced crustal thickness produced at rise crests as Iceland-like hot spot plateaus. These

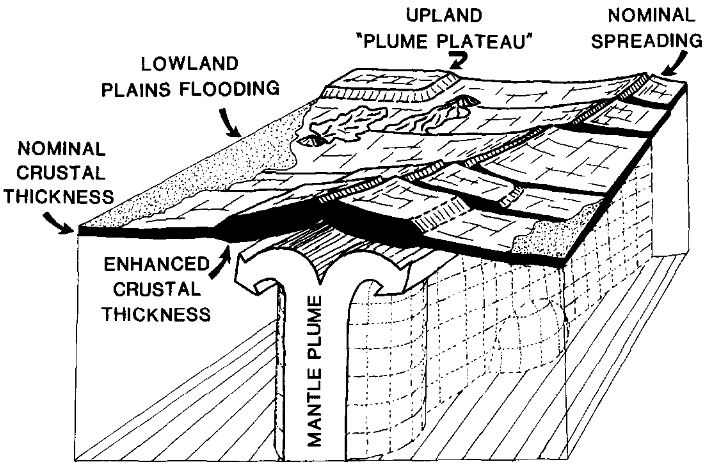

Fig. 6. Crustal spreading/mantle plume model. Variations in upwelling plumes in space and time in a crustal spreading environment cause production of plume plateaus, and their rifting and transport into surrounding lowlands. From Head and Crumpler (1990).

'plume plateaus' are then moved laterally off the rise, are split and separated as plumes decay, and then migrate laterally into the lowlands to produce isolated plateaus of enhanced crustal thickness (Figure 6) (Head and Crumpler, 1990). These isolated 'plume plateaus' are eventually the locus of peripheral deformation (e.g., Tellus Regio) or distributed deformation (e.g., Fortuna Tessera) as they converge and collide with each other (Head and Crumpler, 1990). The 20–30 km modelled crustal thickness for the tessera regions is consistent with this origin. The percentage of tessera that was formed by this mechanism, and that formed by other mechanisms such as compressional deformation or gravitational relaxation of high topography (see Bindschadler and Head, 1988, 1989) has not been established.

On the basis of the geological evidence for the origin of the topography associated with areas of enhanced modelled crustal thickness, it is concluded that crustal thickness variations alone can account for the majority of Ishtar Terra topography (although contributions from other sources are likely), that crustal thickness variations make a significant contribution to the topography of plateaus associated with the broad thermal rise in Aphrodite Terra, and that tessera plateaus may represent regions of crust thickened by a variety of processes. The topography of the Equatorial Highlands can be accounted for by variations in thermal structure (the broad rise) and crustal thickness (the plateaus), although the exact proportion of contributions to each has not been established.

6. Summary and Conclusions

Morgan and Phillips (1983) tested the hypothesis that conductive heat loss is an efficient heat loss mechanism and that most of the topography of Venus (about

93%) could result from spatially varying thermal expansion in response to spatially varying heat flow (hot spots). They proposed that the remaining topography (at high elevations) could be accounted for by crustal thickness variations. In this article the complementary end-member hypothesis, that the topography of Venus could result solely from crustal thickness variations, has been examined. Existing data for Venus (composition, hypsometry, regional topography, regional slopes, geology) are consistent with a basaltic crust with a wide range of thickness for the planet as a whole, but a narrow range in terms of crustal thickness frequency distribution, with the vast majority of the planet (>80%) having crustal thicknesses within ±5 km of the mean value of about 15 km.

Geologic evidence, such as the presence of orogenic belts and compressional deformation in Ishtar Terra, is consistent with most of the highest topography being due to variations in crustal thickness. Geologic evidence leading to reasonable hypotheses for the origin of other regions of intermediate to high topography (the tessera terrain) are also consistent with crustal thickness variations, although not uniquely so. Geologic evidence indicates that a portion of Venus (primarily the Equatorial Highlands) is the site of extensional deformation and that the topography can be explained by a combination of thermal sources (broad rise) and crustal thickness variations (plateaus). Although crustal thickness is not uniquely determined, various lines of evidence point to a crustal thickness for the Venus lowlands of about 10–20 km. On the basis of these observations and interpretations, a simple end-member model of Airy isostasy relative to global Venus topography has been used to assess the significance of crustal thickness variations in explaining the topography of Venus. The distinctive unimodal hypsometric curve can be explained by: (1) a crust of relatively uniform thickness (less than 20 km thick) comprising over 75% of the surface, (2) local plateaus (tessera) of thickened crust (about 20–30 km) forming <15% of the surface, (3) regions (Beta, Ovda, Thetis, Atla Regiones and Western Ishtar Terra) forming <10% of the surface showing 30–50 km modelled crustal thickness, which can be interpreted on the basis of geologic observations to be combinations of crustal thickening (plateaus) and thermal effects, and (4) areas in which Airy isostasy predicts crustal thicknesses in excess of 50 km (the linear orogenic belts of Western Ishtar Terra). It is concluded that crustal thickness variations linked to crustal formational and modificational processes can account for the vast majority of the observed topography. Regional variations in heat flux (lithospheric thickness variations) are important locally and regionally, as along the rise crest of the Equatorial Highlands. More subtle thermal variations [e.g., very broad low thermal rises analogous to the Pacific superswell (McNutt and Fischer, 1987), and the flanks of spreading centers between rise crest and thermal equilibrium] can also obviously contribute to variations in topography. Gravity and topography data from MAGELLAN will permit the further delineation of the contributions of thermal effects and crustal thickness variations for local regions.

As yet not uniquely determined is the mechanism or mechanisms for crustal formation and evolution. Impact-related early crustal differentiation seems unlikely for the surface of Venus observed thus far because of its young age. A crust of impact origin buried by subsequent vertical differentiation and volcanic flooding requires at least a 6–8 km thickness of lava to obscure the early record of impact craters and basins. (Head, 1981). The size-frequency distribution of observed circular to oval features that could potentially be modified impact craters, does not support an impact origin for these features (Nikolaeva *et al.*, 1986). In any case, such a process would require that the present crust is primarily the result of vertical differentiation processes. Presently observed rates of resurfacing (4 km/b.y.; Grimm and Solomon, 1987) are probably sufficient to obscure such an early crust, but the early impact-derived crust must be very thin (<5–10 km) if the total average crustal thickness is less than 20 km.

Vertical differentiation processes are plausible mechanisms for the formation and evolution of the present crust. If such processes are the dominant mechanism, then some implied crustal growth rates are (in average thickness and volumes) 2–4 km/b.y. (1–2 km^3/yr) for a crust produced by surface volcanism, at least 9–18 km/b.y. (4.5–9 km^3/yr) for crustal thicknesses limited by melting, and at least 17 km/b.y. (8.5 km^3/yr) for a crustal thickness limited by negative buoyancy (Tables I, IV; Figure 1c). Vertical differentiation models require abundant local sources which vary in space and time, which produce a generally homogeneous crustal thickness, and which permit sufficient lateral crustal movement to yield the observed orogenic belts and regions of compressional deformation. If vertical differentiation processes are in fact the predominant mechanism for production of the crust on Venus, then the estimates of present average crustal thickness of 10–20 km discussed above constrain vertical differentiation models. Models which invoke crustal thicknesses limited by basal melting (about 40–80 km) or negative buoyancy (about 75 km) are not consistent with the 10–20 km estimated crustal thickness. Rates for vertical recycling models based solely on addition of surface volcanic deposits (1–2 km^3/yr) could produce the presently observed crust over the history of the planet. However, this would require that essentially no crust has been recycled, and that surface volcanic deposits are by far the major contributor to the crust. This seems unlikely because examples from Earth show that underplating and intrusion usually dominate volumetrically over extrusion (the ratio of intrusion to extrusion for Earth is about 5:1 for oceanic regions and about 10:1 for continental regions; Crisp, 1984). In order for this model to be viable, it would require that the average volcanic flux on Venus in the past be considerably less than 2 km^3/yr (Grimm and Solomon, 1988), or that some type of crustal recycling take place to account for the "missing crust". The average crustal thickness of about 10–20 km would seem to rule out global vertical crustal recycling by basal melting or negative buoyancy. Models of vertical differentiation require anomalously low resurfacing rates by terrestrial standards or an unspecified method of crustal recycling (Figures 1b, c).

Crustal spreading models provide a mechanism: (1) to produce a geologically young global crust of less than about 20 km thickness (about 15 km average thickness; Sotin *et al.*, 1989), (2) to produce Iceland-like plateaus of locally thickened crust (tessera-like regions), (3) to explain the generally extensional deformation of the regions where a thermal component to topography seems most likely (Equatorial Highlands), and (4) to move the crust and plateaus laterally to produce the observed areas of convergence, compressional deformation, orogenic belts, and crustal thickening. Crustal spreading models require that crustal recycling is taking place at zones of convergence and crustal thickening.

On the basis of this study and other analyses (Head, 1990c,d), the differences between the hypsometric curves of Venus and Earth (Figure 3) are interpreted to be related to crustal thickness variations and their distribution. Although thermal effects on topography are clearly present and important on both Venus and the Earth, the major difference between the hypsometric curves on Earth (bimodal) and Venus (unimodal) is attributed primarily to the contrast in relative average thickness of the crust between the two terrains on Earth (continental/oceanic; 40/5 km = 35 km, 8:1) and Venus (upland plateaus/lowlands; about 30/15 km = 15 km, 2:1) (35 km − 15 km = a difference of 20 km). The Venus unimodal distribution is thus attributed primarily to the large percentage of terrain with relatively uniform crustal thickness, with the skewness toward higher elevations due to the relatively small percentage of crust that is thickened by only about a factor of two. The Earth, in contrast, has a larger percentage of highlands (continents), whose crust is thicker by a factor of eight, on the average, leading to the distinctive bimodal hypsometric curve. The major differences between the long-baseline slope distributions for Venus and Earth (Sharpton and Head, 1985, 1986) also appear to be linked to the difference in contrast in crustal thickness variations (causing less distinctive slope changes at boundaries between two terrain types on Venus than Earth), the smaller abundance of highland terrain on Venus, and the larger number of highland regions on Earth causing an increased total length of continental boundaries with high slopes.

It is concluded that Venus hypsometry can be reasonably explained by a global crust of generally similar thickness with variations in hypsometry being related to (1) crustal thickening processes (orogenic belts and plateau formation), and (2) local and regional variations in the thermal structure. Vertical differentiation and crustal spreading processes are the most likely candidates for the formation and evolution of the crust. Crustal spreading processes are consistent with several observations derived from presently available data, but require crustal recycling to operate to remove the large volumes of crust that would be produced over the history of the planet. Vertical contributions to the crust of Venus from surface volcanism are presently observed, but are occurring at very low rates by terrestrial standards. If the present volume of crust is equivalent to the total volume produced over the history of the planet, then vertical differentiation mechanisms require very low rates of crustal growth, and if there is a significant intrusional component to the crust, rates must be even lower in earlier Venus history than they are now.

Determination of the exact proportion of crust created by these two mechanisms, and the exact proportion of the topography of Venus that is due to thermal effects and crustal thickness variations, must await new data and further study. In particular, global imaging data showing the age of the surface, the distribution and age of regions of high heat flux, and evidence for the global distribution of processes of crustal formation and crustal loss, would permit better estimates of these proportions. Together with geological data derived from images, high resolution global gravity and topography data would permit modelling of crustal thickness variations and thermal contributions, and testing of the various concepts of crustal growth, which would in turn lead to regional understanding of these processes, and an assessment of their contribution to the global values.

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