Pulsating continents on Venus: An explanation for crustal plateaus and tessera terrains

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A B S T R A C T

We propose that tessera terrains on Venus represent continental crust that does not participate in the periodic recycling of the lithosphere through global subduction events. We have studied the force balance on the boundary of a continental area that survives a global subduction event using an analytical model. In the proposed model, the ratio between the crustal and lithospheric mantle thicknesses controls the force balance. If the crust thickness is less than ~2/5 of the lithospheric mantle thickness, the continental area will be compressed, but if the crustal thickness is higher than ~2/5 of the lithospheric mantle thickness, the continental area will spread out and collapse. Consequently, if the lithospheric mantle beneath a continental region is delaminated during a global subduction event, the continent will collapse generating tessera inliers dominated by extensional tectonics. But if a significant portion of lithospheric mantle remains, then the continental area will be compressed generating a plateau by crustal shortening. The observed plateau heights can be explained by this model, a ≈2 km height plateau can be generated by a lithospheric mantle thickness of 40 km while a ≈4 km height plateau can be generated by a 90 km thick lithospheric mantle. We have modelled this crustal thickening of a continental area by tectonic contraction using a thin viscous sheet approach with a Newtonian viscosity for the crust. The force from a hot mantle elevated during a global subduction event is enough to build up a plateau by compression in ~50 Ma using a viscosity for the continental crust of $\eta = 10^{21}$ Pa s and ~200 Ma for $\eta = 5 \cdot 10^{21}$ Pa s. During this compressional stage concentric fold and thrust belts are generated in the plateau-continent, erasing any impact craters that were present. The subsequent stabilization of a new crust and lithosphere in the surrounding mantle changes the force balance allowing a moderate gravitational collapse of the plateau-continent accommodated by radial grabens. The pulsating continent model links for the first time the generation of crustal plateaus and the origin of the volcanic plains predicting the observed equivalent effective crater density for both terrains.

Keywords:
crustal plateau
continental crust
global subduction event
tessera terrain
Venus

1. Introduction

The origin and evolution of crustal plateaus on Venus have been controversial topics since high-resolution radar images were obtained by the Magellan mission. Crustal plateaus are subcircular areas with diameters in the range of 1500 to 2500 km, and elevations of 0.5 to 4 km above the surrounding plains. They are made up of intensely deformed terrain, known as tessera, characterized by different cross-cutting sets of structures indicating a complex tectonic history (Bindschadler and Head, 1991; Hansen and Willis, 1996; Hansen et al., 1999, 2000). The origin and evolution of crustal plateaus have been a focus of attention because they are among the oldest terrains on Venus and therefore are keys to the understanding of the geodynamic evolution of the planet.

Radar images obtained by the Magellan mission showed that impact craters on Venus are nearly randomly distributed (Schaber et al., 1992; Strom et al., 1994). In order to explain this observation Turcotte (1993, 1995, 1996) and Turcotte et al. (1999) proposed that Venus loses heat by episodic global subduction events. Following this hypothesis, the volcanic plains were formed during the stabilization of a new crust after the last global subduction event about 500 Ma ago. Crustal plateaus are terrains that suffered a complex tectonic evolution before emplacement of the regional volcanic plains; therefore, under the geodynamic hypothesis of a planet losing heat through global episodic subduction events, they are terrains that survived the last of these events. The tectonic evolution of crustal plateaus probably continued after the emplacement of the surrounding volcanic plains.

Crustal plateaus show small gravity anomalies, low gravity to topography ratios, and shallow apparent depths of compensation (ADC), all indicating a thickened crust (Smrekar and Phillips, 1991; Bindschadler et al., 1992a; Kucinskas and Turcotte, 1994; Grimm, 1994; Simons et al., 1997). The good spatial correlation of crustal plateau elevation and highly deformed tessera terrains clearly indicates that deformation plays a major role during crustal thickening (Bindschadler and Head, 1991; Bindschadler et al., 1992a, b; Hansen and Willis, 1996,

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1998; Ghent and Hansen, 1999; Hansen et al., 1999, 2000; Hansen, 2006).

Different models for crustal plateau formation have been proposed, but no consensus has been reached. The two principle models consider crustal plateaus as either the surface expression of downwelling or upwelling flows. The downwelling model involves a tectonic crustal thickening due to concentric compression caused by a subsolidus flow and horizontal accretion of an ancient thin lithosphere on a cold mantle downwelling flow (e.g., Bindschadler and Head, 1991; Bindschadler et al., 1992a, b; Bindschadler, 1995; Gilmore and Head, 2000). The upwelling (plume) model accomplishes crustal thickening by magmatic underplating and volcanism due to interaction of an ancient thin lithosphere with a large thermal mantle plume (Hansen and Willis, 1998; Phillips and Hansen, 1998; Ghentand Hansen, 1999; Hansen et al., 1999, 2000).

The tectonic patterns of crustal plateaus are complex. Both extensional and compressional structures appear with a wide range of wavelengths. Compressional structures are typically concentric and oriented parallel to the plateau margins, while extensional structures are mainly perpendicular to the margins and radially distributed (Ghent and Hansen, 1999). While there is agreement that the final tectonic stage is characterized by radial extension which generates grabens, the initial tectonic stages are debated. The supporters of the plume model distinguish between two families of extensional structures, long-narrow grabens with small wavelengths, so called "ribbons", and wide short multiple-scarp grabens (Ghent and Hansen, 1999). These authors consider ribbons the first structure to be formed, later the concentric folds, and finally the wide complex graben. This interpretation is based on the assumption that the wavelength of each of these types of structures is indicative of the thickness of the brittle crust, being generated with a brittle–ductile transition deepened with time during the cooling of the plateau. However, Gilmore et al. (1997, 1998) and Romeo et al. (2005) concluded that the deformation phases are first compressional and finally extensional. In this interpretation the extensional structures, including the long-narrow grabens, postdate or are contemporary with the generation of the compressional structures.

Both hypotheses have problems in explaining all the characteristics of plateaus. On the one hand, the weak points of the downwelling model are: (1) a predicted domical shape instead of the flat-topped plateau geometry (Kidder and Phillips, 1996); (2) too much time is required for the thickening by crustal flow (1–4 billion years) (Kidder and Phillips, 1996). On the other hand, the main challenges of the plume hypothesis are: (1) there is no explanation for the extensive contractional tectonics observed (Ghent et al., 2005; Hansen, 2006); (2) the predicted timing of the extensional tectonics contradicts cross-cutting relationships in different locations (Gilmore et al., 1998; Romeo et al., 2005). Although Gilmore et al. (1998) argued that the formation of ribbon-tessera terrain requires an excessive geothermal gradient, Ruiz (2007) indicated that the heat flow needed for generating ribbons is reasonable for a plume environment.

Hansen (2006) rejected the hot spot and downwelling models and proposed a new catastrophic model where crustal plateaus were formed by huge lava ponds generated by massive mantle melting due to large bolide impacts on a thin ancient lithosphere. According to this model the mantle beneath the lava pond would be a depleted residuum compositionally buoyant in respect to the adjacent undepleted mantle, causing the plateau uplift by isostasy. The main challenges of this impact model are: (1) there is a wide discussion about whether large bolides are able to melt a significant portion of the mantle for generating such an amount of magma (Ivanov and Melosh, 2003); (2) except for the final extensional tectonics, this model considers that the initial tectonic deformations, both extensional and compressional, observed in crustal plateaus are caused by the convective forces generated during the lava pond cooling, in this context, the significant amounts of shortening observed in the plateau margins are very difficult to explain; (3) The generation of the large wavelength folds requires stresses enough to deform a brittle layer that is several km thick, how can the underlying liquid magma in convection transmit such forces to the upper brittle layer?; (4) no numerical calculation of the isostasy forces involved in the plateau raising has been provided yet, in other words, what elevation can be supported by this method?

Finally, the process of crustal plateau formation is time-transgressive for all the proposed models. In all the cases plateaus are considered to be elevated and the compressive structures to be formed before the regional volcanic plains emplacement. The density of craters located in crustal plateaus is equivalent to that of the regional volcanic plains. The surface of crustal plateaus is not sufficiently large to determine their age from crater statistics even considering the possibility that all the plateaus have the same age. But they cannot be too much older than the regional volcanic plains considering that they have approximately the same crater density. The crater density of crustal plateaus indicates that they were probably formed around 500 ma, although, considering that their area is not statistically representative, they could be as much as 1000 ma old (Nunes et al., 2004). If crustal plateaus were formed and deformed diachronously since the early history of Venus they should record different crater densities always higher than the average planetary crater density, which does not occur.

Tessera inliers dominated by extensional tectonics outcropping across the Venusian regional volcanic plains are generally considered remnants of collapsed crustal plateaus (e.g., Bindschadler et al., 1992b; Bindschadler, 1995; Ivanov and Head, 1996; Hansen, 2006; Hansen and Willis 1996, 1998; Nunes et al., 2004; Nunes and Phillips, 2007). Hansen and Willis (1996) distinguished between tessera inliers dominated by graben, and tessera inliers dominated by fractures, proposing that the former are remnants of collapsed crustal plateaus and the latter flooded ancient corona-chasmata chains. Nevertheless, the analytical and finite element models performed by Nunes et al. (2004) indicate that the range of preserved crustal plateau morphologies and tessera inliers is not possible to achieve through lower crustal viscous flow during collapse of a compensated plateau. The analytical and finite element models for the evolution of uncompensated plateaus of Nunes and Phillips (2007) better address the morphologies of crustal plateaux and tessera terrains although these models do not explain all the extensional structures observed.

In this paper we propose an alternative model for crustal plateau and tessera inliers origin and evolution. A model of crustal plateau formation has to give: (1) a reasonable mechanism for crustal thickening; (2) an explanation of the tectonic evolution deduced from structural analysis; (3) take into account the observed distribution of impact craters, and (4) should explain the current coexistence of different sizes and heights of crustal plateaus and tessera inliers on Venus. Assuming that Venus loses heat through global subduction events, the last of which led to the generation of the regional volcanic plains 500 Ma ago, and considering that tessera terrains are stratigraphically older than the regional plains in every location, it is clear that the tessera terrains that make up crustal plateaus and tessera inliers are remnants of an older crust that survived the subduction event. On Earth, the only crust that does not participate in the subduction process is the continental crust due to its buoyancy associated with its lower density. We propose that tessera terrains, both the lowland inliers and the crustal plateaus, represent Venusian continental crust, or buoyant areas of differentiated composition, that survived the proposed last global subduction event. In this context crustal plateaus would represent continents and the tessera inliers, collapsed continents. The change in the force balance after a global subduction event in the surrounding of an area of thin continental crust (for example a collapsed continent) with a significant lithosphere underneath is enough to build up by compression a crustal plateau in 50 Ma. However, if the subduction event delaminates a

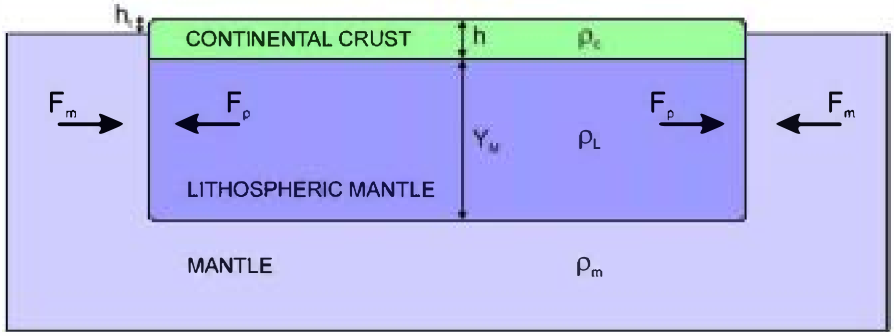


Fig. 1. Model geometry and parameters of a continental area after a subduction event.

significant portion of the lithosphere under a continent, a subsequent collapse will occur generating the observed tessera inliers. Thus, controlled by the proposed periodic subduction events, the Venusian continents can suffer cycles of thickening events driven by tectonic compression and also extensional collapse events. Therefore we propose the term pulsating continents to describe the cyclic behavior that links crustal plateaus and tessera inliers.

2. Model description

We will analyze the evolution of an area of thin low-density crust with a relatively thick lithosphere that survived a global subduction event. Our two-dimensional analytical model contains two blocks: one with a single layer representing the hot mantle raised to the surface during a global subduction event, and an adjacent three-layer block representing the future crustal plateau. The layers of this block are a thin low-density (ρ_c) continental crust, the lithospheric mantle beneath with a density controlled by a linear temperature profile, and the underlying hot mantle with a density ρ_m (Fig. 1). The temperature at the surface is T_0, the temperature of the mantle is T_1 and the temperature at the base of the crust is T_M. The initial temperature profile is assumed to be linear from T_0 to T_1. The temperature at the base of the crust is

$$T_M = \frac{Y_M T_0 + h_0 T_1}{h_0 + Y_M} \tag{1}$$

where Y_M is the thickness of the lithospheric mantle and h_0 is the initial thickness of the crust.

We assume that the topography is compensated so that

$$\rho_m(Y_M + h - h_t) = \rho_c h\left(1 + \frac{1}{2}\alpha(T_M - T_0)\right) + \rho_m Y_M\left(1 + \frac{1}{2}\alpha(T_1 - T_M)\right) \tag{2}$$

where h is the crustal thickness, h_t is the plateau elevation and $\alpha = 2.4 \cdot 10^{-5}$ K^{-1} is the thermal expansion coefficient with the same value for the crust and the mantle.

Integrating the pressure profile of a mantle column outside the plateau we obtain the force from the mantle surrounding the plateau, F_m,

$$F_m = \frac{1}{2}\rho_m g(Y_M + h - h_t) \tag{3}$$

where $g = 8.6$ ms^{-2} is the surface gravity on Venus. The pressure profile, P_0, beneath the plateau as a function of depth, y, is

$$P_0 = \rho_c g y + \frac{1}{2}\rho_c \alpha(T_M - T_0) g y \quad 0 \le y \le h$$
$$- \rho_c g h + \frac{1}{2}\rho_c \alpha(T_M - T_0) g h + \rho_m g(y - h) + \frac{1}{2}\rho_m \alpha(T_1 - T_M) g(y - h) \quad h \le y \le h + Y_M. \tag{4}$$

Integrating over depth we obtain the value of the force from the plateau, F_p,

$$F_p = \frac{1}{2}\rho_c g h^2 + \frac{1}{4}\rho_c g \alpha(T_M - T_0) h^2 + \rho_c g h Y_M + \frac{1}{2}\rho_c g \alpha(T_M - T_0) h Y_M$$
$$+ \frac{1}{2}\rho_m g Y_M^2 + \frac{1}{4}\rho_m g \alpha(T_1 - T_M) Y_M^2 \tag{5}$$

In equilibrium $F_m = F_p$, and assuming compensated topography we combine Eqs. (3), (5), (2) and (1) to obtain

$$Ah^2 + BhY_M + Ch^2Y_M + DhY_M^2 + EY_M^3 + FhY_M^3 + Gh^2Y_M^2 + HY_M^4 = 0 \tag{6}$$

where A, B, C, D, E, F, G and H are constants defined by:

$$A = \frac{1}{2}\rho_c h_0^2\left[\frac{\rho_c}{\rho_m}\left(\frac{1}{4}\alpha^2(T_1 - T_0)^2 + \alpha(T_1 - T_0) + 1\right) - \frac{1}{2}\alpha(T_1 - T_0) - 1\right] \tag{7}$$

$$B = \frac{1}{4}\rho_c \alpha^2 (T_1 - T_0)^2 h_0^2 \tag{8}$$

$$C = \rho_c h_0\left[\frac{1}{2}\left(\frac{\rho_c}{\rho_m} - \frac{1}{2}\right)\alpha(T_1 - T_0) + \frac{\rho_c}{\rho_m} - 1\right] \tag{9}$$

$$D = \frac{1}{2}\rho_c \alpha(T_1 - T_0) h_0 \tag{10}$$

$$E = \frac{1}{4}\rho_m \alpha(T_1 - T_0) h_0 \tag{11}$$

$$F = \frac{1}{2}\rho_c \alpha(T_1 - T_0) \tag{12}$$

$$G = \frac{1}{2}\rho_c\left(\frac{\rho_c}{\rho_m} - 1\right) \tag{13}$$

$$H = \frac{1}{4}\rho_m \alpha(T_1 - T_0)\left(1 + \frac{1}{2}\alpha(T_1 - T_0)\right) \tag{14}$$

The results for equilibrium that relate the crustal thickness (h) to the lithospheric mantle thickness (Y_M) for different crustal densities (ρ_c) are plotted in Fig. 2a using Eq. (6) with an initial crustal thickness $h_0 = 7$ km. The change of the plateau elevation (h_t) with the lithospheric mantle thickness (Y_M) in equilibrium is also plotted in Fig. 2b using Eqs. (6) and (2) for an initial crustal thickness $h_0 = 7$ km.

The results for equilibrium shown in Fig. 2 indicate that when the crustal thickness represents ~2/5 of the lithospheric mantle thickness, then $F_m = F_p$. From this we can conclude that if the crustal thickness is less than ~2/5 of the lithospheric mantle thickness then the $F_m > F_p$ and the plateau will be compressed by the surrounding hot mantle, but if the crustal thickness is greater than ~2/5 of the lithospheric mantle thickness then $F_m < F_p$ and the plateau will spread out and collapse, these fields of extension and compression are shown in

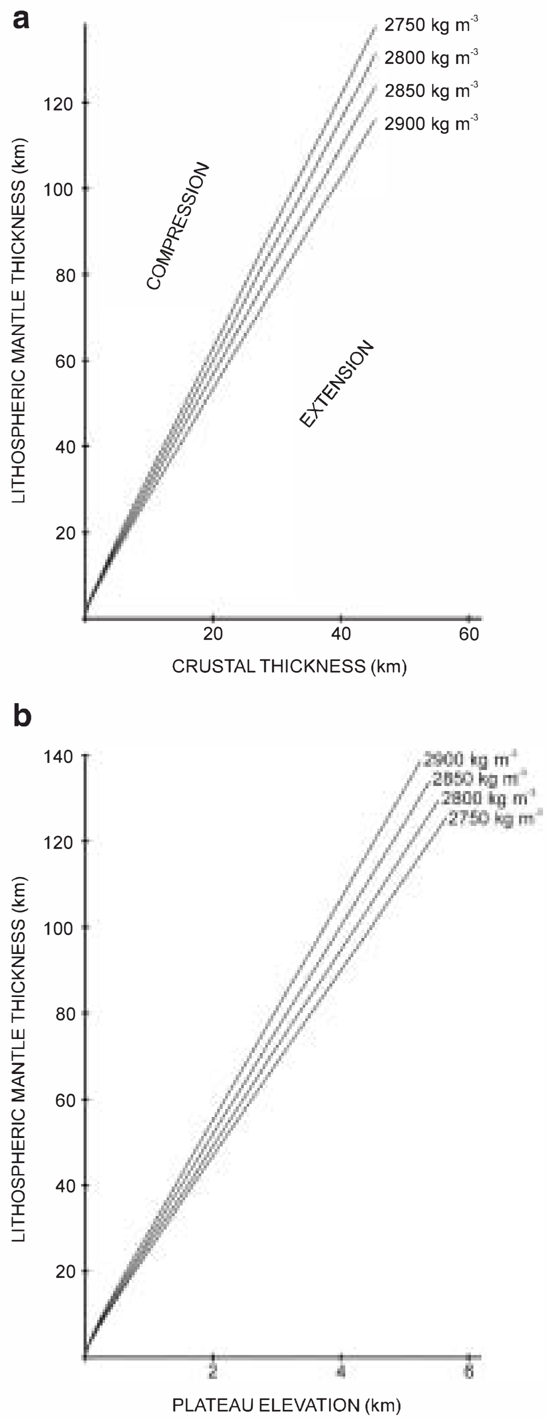


Fig. 2. Results for an equilibrium force balance ($F_p=F_m$). a) Dependence of the lithospheric mantle thickness on the crustal thickness of the continental area. The fields that will undergo compression or extension are shown. b) Dependence of the lithospheric mantle thickness on plateau-continent elevations. Results are plotted for different densities of the continental crust and a density for the mantle of $\rho_m=3300$ kg m^{-3}.

Fig. 2a.We have studied in detail the case of the tectonic thickening of the crust during compression of the plateau driven by a force balance where $F_m>F_p$ due to an initial crustal thickness less than ~2/5 of the lithospheric mantle thickness.

Several studies of plateau deformation on the Earth have been carried out with particular emphasis on the Tibetan Plateau. Lithospheric deformation is a complex problem involving both brittle and ductile deformation and has been studied by a number of workers (Molnar and Lyon-Caen, 1988; Dewey, 1988; Gaudemer et al., 1988; Zoback and Townend, 2001; Zoback et al., 2002; Jackson, 2002; Sleep, 2007). One approach to modelling gravitational collapse is to treat the lithosphere as a highly viscous material in which a lateral gradient in gravitational potential energy, due to the elevated topography, provides a driving mechanism for the lateral flow of the lithosphere form high pressure (high elevations) to low pressure (lower elevations) (England and McKenzie, 1982; Fleitout and Froidevaux, 1982, 1983; England, 1987; Fleitout, 1991; Rey et al., 2001). This is the approach that we use.

Models for buoyancy driven flows vary in the treatment of rheologic layers into which the lithosphere is divided as well as the rheology of these layers (Bird, 1991). In addition, Zhou and Sandiford (1992) have carried out a detailed study of buoyancy driven flows using a stress envelope for the lithospheric rheology that includes friction constraints at shallow depths and thermally activated creep at greater depths. They found that the collapse rate is sensitive to the Moho temperature.

Clearly, the deformation of the brittle upper lithosphere is dominated by fracture and faulting, however, this deformation takes place on a wide range of scales, and therefore it is often appropriate to model it as a continuum deformation. This problem has been treated by Nanjo and Turcotte (2005) and by Nanjo et al. (2005) using concepts of damage mechanics. These authors show that it is appropriate to treat continuum brittle deformation using a nonlinear viscous rheology with a yield stress.

In order to approach this time-dependent problem we have introduced a Newtonian viscous rheology for the crust giving a viscous force opposite to the compression of the plateau during deformation. If during compression both the crust and the lithosphere are deformed and thickened, then the ratio of the thicknesses of the layers will remain constant and consequently F_m will always be bigger than F_p. With this approach, equilibrium would never be reached. Therefore we have considered that the crust thickens during the compressional event of the plateau but the thickness of the lithospheric mantle (Y_M) remains constant. We assume that the excess lithosphere is removed during the compression by a downwelling flow of the cold lithospheric mantle due to a gravitational instability (Houseman and Molnar, 1997) beneath the center of the plateau. Houseman et al. (2000) modelled a lithospheric shortening event that produced crustal thickening and obtained a lithospheric mantle thickness that remained approximately constant during the compressional event. The thickness of the lithospheric mantle remains unchanged due to a downwelling flow of the cold lithospheric mantle driven by the development of a gravitational instability when the lithosphere starts to thicken. Consequently, following the results of Houseman et al. (2000), we have considered that the lithospheric mantle thickness (Y_M) remains constant to a first approximation.

Since we assume that the crust has a constant specified viscosity, the temperature of the crust does not influence its rheology. However, the temperature does change the density of the crust through thermal expansion. For this purpose we assume that the temperatures at the top of the crust, T_0, and the base of the crust, T_M, remain constant so that the crustal temperature gradient decreases $\sim h_0^{-1}$ as the crust thickens. We further assume that the linear temperature profile of the lithospheric mantle from T_1 to T_M remains unchanged during plateau deformation.

We note that the characteristic time for thermal equilibration in this section is about 50 million years, so our assumptions should be reasonable. Also this choice has a relatively small influence on our calculation since the change of continental thickness dominates in the calculation of forces.

We apply the thin viscous sheet approach and assume the lubrication approximation is valid neglecting inertial forces (England and McKenzie, 1982; Houseman and England, 1986; Sonder and

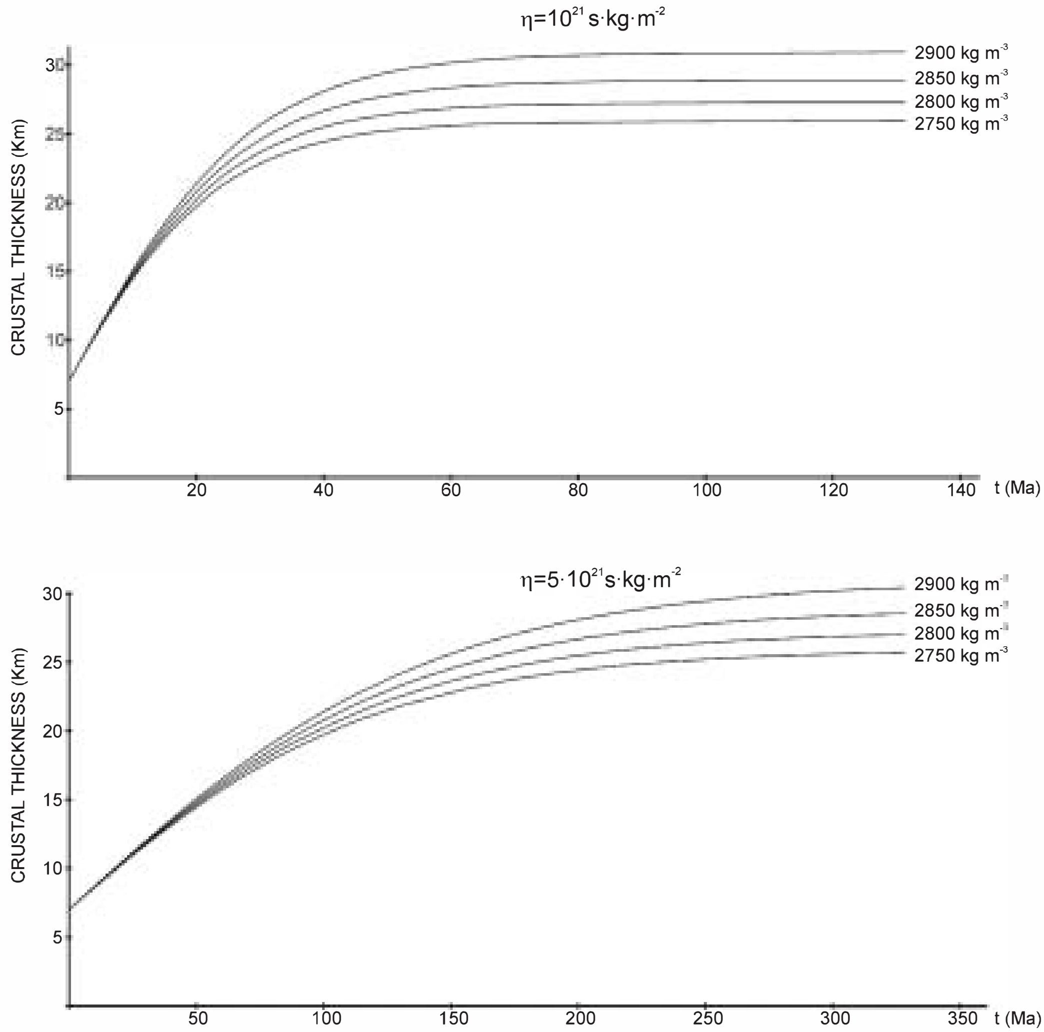


Fig. 3. Values of the thickness of the continental crust, h, as a function of time, after a subduction event, for effective crustal viscosities of $\eta = 10^{21}$ Pa s and $\eta = 5\cdot 10^{21}$ Pa s. The model parameters are: T_0=750 K, T_1=1600 K, h_0=7 km, Y_M=80 km, ρ_m=3300 kg m^{-3}. Results are plotted for different densities of the continental crust.

England, 1986; Jadamec et al., 2007). Including the viscous term in our force balance equation we get

$$4\eta\frac{dh}{dt} = ah^2 + bh + c, \tag{15}$$

where η is the crustal viscosity and a, b, c are constants defined by

$$a = \frac{1}{2}\rho_c\left[\frac{\rho_c}{\rho_m} - 1 + \alpha(T_M - T_0)\left(\frac{1}{4}\frac{\rho_c}{\rho_m}\alpha(T_M - T_0) + \frac{\rho_c}{\rho_m} - \frac{1}{2}\right)\right], \tag{16}$$

$$b = \frac{1}{2}\rho_c Y_M\left[\alpha(T_1 - T_M) + \frac{1}{2}\alpha^2(T_M - T_0)^2\right], \tag{17}$$

$$c = \frac{1}{4}\rho_m\alpha(T_1 - T_M)Y_M^2\left[\frac{1}{2}\alpha(T_1 - T_M) + 1\right]. \tag{18}$$

Integration of Eq. (15) with the initial conditions ($h = h_0$ when $t = 0$), and solving for the crustal thickness (h) as a function of time (t) gives

$$h = \frac{1}{2a}\left[-b + \sqrt{4ac - b^2}\tan\left(\frac{t\sqrt{4ac - b^2}}{8\eta} + \arctan\left(\frac{b + 2ah_0}{\sqrt{4ac - b^2}}\right)\right)\right]. \tag{19}$$

The evolution of the crustal thickness with time has been plotted in Fig. 3 using Eq. (19), and the change of elevation with time has been given in Fig. 4 substituting the crustal thickness, h, by an expression function of the plateau elevation, h_t, obtained from Eq. (2) into Eq. (19). We have used T_0=750 K, T_1=1600 K, h_0=7 km, Y_M=80 km, ρ_m=3300 kg m^{-3}, four values of crustal density, ρ_c=2750, 2800, 2850, and 2900 kg m^{-3} and two values of crustal viscosity, $\eta = 10^{21}$ Pa s and $\eta = 5\cdot 10^{21}$ Pa s, for the solutions represented in Figs. 3 and 4. The main thickening of the continental crust of the plateau by compression can be accommodated in 50 Ma with a low viscosity, $\eta = 10^{21}$ Pa s, and will take 200 Ma using a higher viscosity $\eta = 5\cdot 10^{21}$ Pa s. The final plateau elevation at equilibrium is controlled by the lithospheric mantle thickness, Y_M. The plateau elevations obtained for different values of Y_M are given by the equilibrium relation in Fig. 2b.

The temperature profile of the lithospheric mantle during compression remains unaltered because of the constant lithospheric mantle thickness. But, for the crust, we have included the change of the linear temperature profile as the crust thickens, being always linear from T_M to T_0.

During the construction of a crustal plateau by tectonic crustal thickening the volume of continental crust remains constant. This can be expressed as

$$w_0^2 h_0 = w(t)^2 h(t) \tag{20}$$

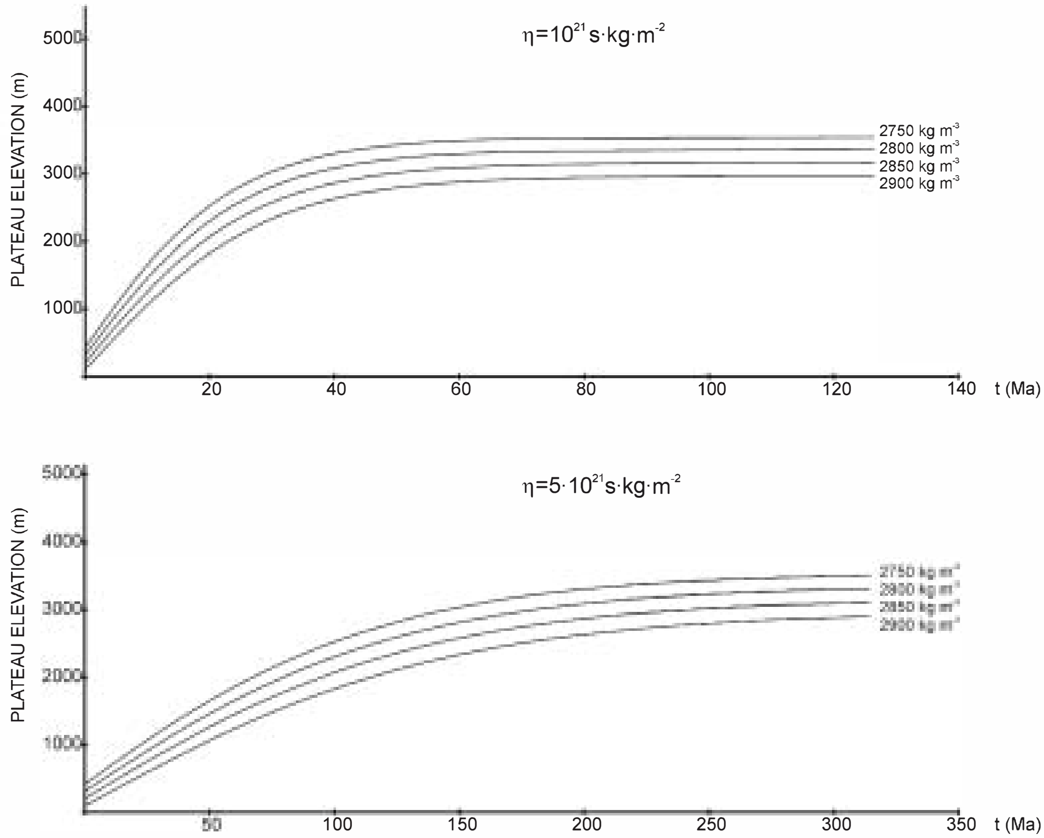


Fig. 4. Values of plateau elevation, h_t, as a function of time, after a subduction event, for effective crustal viscosities of $\eta=10^{21}$ Pa s and $\eta=5{\cdot}10^{21}$ Pa s. The model parameters are: $T_0=750$ K, $T_1=1600$ K, $h_0=7$ km, $Y_M=80$ km, $\rho_m=3300$ kg m^{-3}. Results are plotted for different densities of the continental crust.

where w_0 is the initial diameter of the plateau and $w(t)$ is the diameter of the plateau as function of time. The linear shortening, e, in a section of the plateau during the tectonic compression is given by

$$e=\frac{w_0-w(t)}{w_0}. \tag{21}$$

An expression of the shortening, e, as a function of the initial crustal thickness, h_0, and the crustal thickness, $h(t)$, can be obtained combining Eqs. (20) and (21),

$$e=1-\sqrt{\frac{h_0}{h(t)}}. \tag{22}$$

The strain rate, $\dot{e}$, of the linear deformation can be obtained by

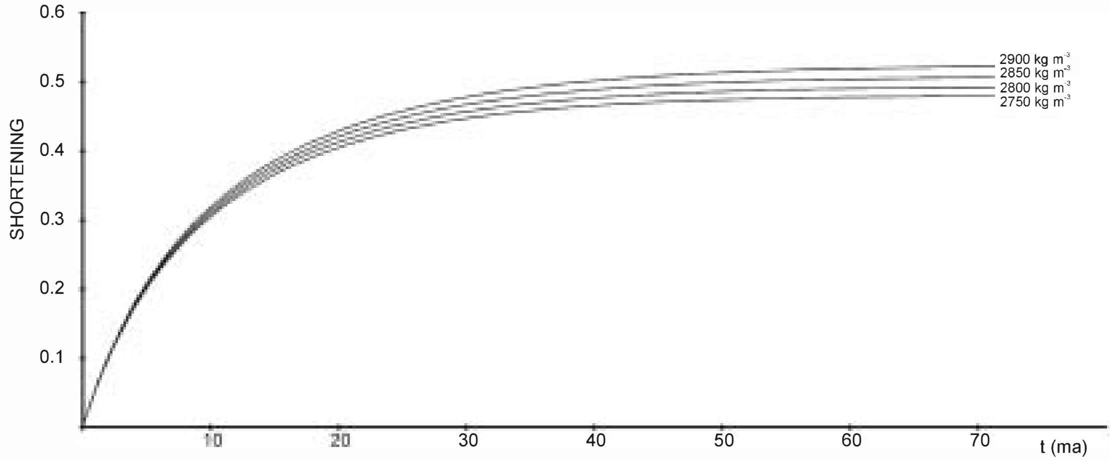
$$\dot{e}=\frac{1-\sqrt{\frac{h_0}{h(t)}}}{t}. \tag{23}$$

The results for the evolution of shortening, e, and strain rate, $\dot{e}$, with time obtained respectively from Eqs. (22) and (23) combined with Eq. (19) are given in Fig. 5.

3. Tectonic evolution of crustal plateaus

The tectonic evolution of a crustal plateau is illustrated in Fig. 6. Our model shows that initially the tectonic stress in a low density crustal area overlaying a significant lithosphere that survived a global subduction event is compressional. Compression continues until the equilibrium between the forces from the plateau and from the surrounding hot mantle is reached. The amount of shortening needed to build up a plateau by tectonic compression depends on the initial geometry of the low density crustal area (h_0 and Y_M). Starting with an initial crustal thickness $h_0=7$ km and a lithospheric mantle thickness $Y_M=80$ km the linear shortening in the two horizontal dimensions is about 40–50% (Fig. 5). This large amount of shortening is accommodated by concentric thrusting and folding in the margins of the plateau and by a high-angle cross-cutting thrust and fold interference in the center yielding the observed basin and dome tessera terrain of this area. The basin and dome interference patterns of the central areas of crustal plateaus were previously described by Hansen and Willis (1996) and Ghent and Hansen (1999). This shortening is accommodated by compressional structures with a wide range of wavelengths from 0.1 to 150 km, probably due to deformation of a complex layered crust. This initial compressional stage is represented in Fig. 6 a b c.

After a period of intense magmatism, new crust surrounding the crustal plateau is stabilized and a new lithosphere grows by heat conduction. This will change the force balance reducing the force that compressed the plateau leading to a period of extension by gravitational collapse. This gravitational extension does not lead to a total collapse because the new rigid lithosphere thickening in the region surrounding the plateau will block the spreading process, stabilizing the crustal plateau as it is observed today. Radial grabens and fractures are expected to be generated in the margins of the plateau, during this aborted collapse process. The extension in the center generates a basin and dome interference pattern formed by grabens with different orientations. During this period, a tectonic inversion occurs in the majority of the initially compressional structures. The extensional grabens and fractures observed in crustal plateaus show a wide range of spacing from 0.5 to 50 km which is again indicative of the deformation of a multilayered complex crust. This extensional stage makes the plateau wider by compressing the crust of the surrounding plains where concentric folds and thrusts are formed and the plains adjacent to the plateau are raised by moderate



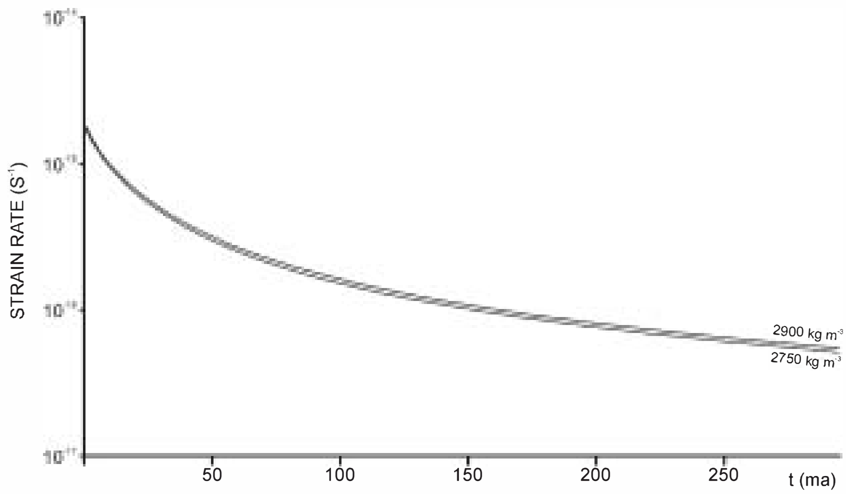


Fig. 5. Dependence of the one-dimensional shortening of the plateau on time, and the dependence of the strain rate on time. The results are plotted for an effective crustal viscosity of $\eta = 10^{21}$ Pa s. The model parameters are: $T_0 = 750$ K, $T_1 = 1600$ K, $h_0 = 7$ km, $Y_M = 80$ km, $\rho_m = 3300$ kg m^{-3}. Results are plotted for different densities of the continental crust.

crustal thickening. This final extensional stage of our tectonic model for crustal plateaus evolution is represented in Fig. 6d.

4. Pulsating continents: a cyclic behavior of crustal plateaus and tessera inliers

The cyclic evolution of the Venusian geodynamics outlined by the model of periodic global subduction events of Turcotte (1993, 1995) and Turcotte et al. (1999) strongly suggests a cyclic behavior of the continental crust that does not participate in the subduction process. If we consider the possibility that tessera terrains represent differentiated low-density crust, both crustal plateaus and tessera inliers could represent different stages of the same cycle. Our proposal of cyclic behavior for the continental crust is outlined in Fig. 7. The stages b, c, d and e of Fig. 7 are equivalent to the tectonic evolution of a crustal plateau described in Fig. 6. During the deformation and thickening of the plateau from b to c the crust becomes partially decoupled with respect to the lithospheric mantle. This decoupling allows the delamination of the lithospheric mantle during the next global subduction event. Then the buoyant continental crust remains surrounded by hot mantle (Fig. 7f). At this moment there is no lithospheric mantle beneath that would undergo plateau compression, thus the crustal plateau collapse gravitationally by extension in a very hot mantle environment (Fig. 7g). The tectonic structures expected in such a hot extensional environment are short-wavelength narrow grabens (ribbons). Contemporary with plateau collapse, the extensive volcanism generated by the hot mantle produces a partial flooding of the collapsed plateau-continent generating the observed tessera inliers mainly characterized by extensional tectonics. A new crust is stabilized surrounding the completely collapsed plateau-continent and a new lithosphere grows by heat conduction (Fig. 7h), blocking the collapse process. This new lithosphere generated by cooling under the collapsed plateau is mechanically coupled to the continental crust because the crust is thin and therefore the upper lithospheric mantle is relatively cold as new lithosphere gets thicker (Fig. 7a). Thus, the next global resurfacing process does not delaminate the lithosphere under the continent, yielding the initial conditions for plateau growth (Fig. 7b).

Obviously this is an idealized model that links crustal plateaus and tessera inliers, based on the cyclic geodynamic model of subduction events. The main key for the evolution of a continental area after a global subduction event is the ratio between the crustal and lithospheric mantle thicknesses that survive that subduction event. If the ratio is greater than the equilibrium value it will undergo gravitational collapse and if it is smaller the continental crust will be compressed (Fig. 2a).

5. Discussion

5.1. Structural geology

Fig. 8 shows different radar images of Ovda Regio, the largest crustal plateau of the planet that has been the main location for

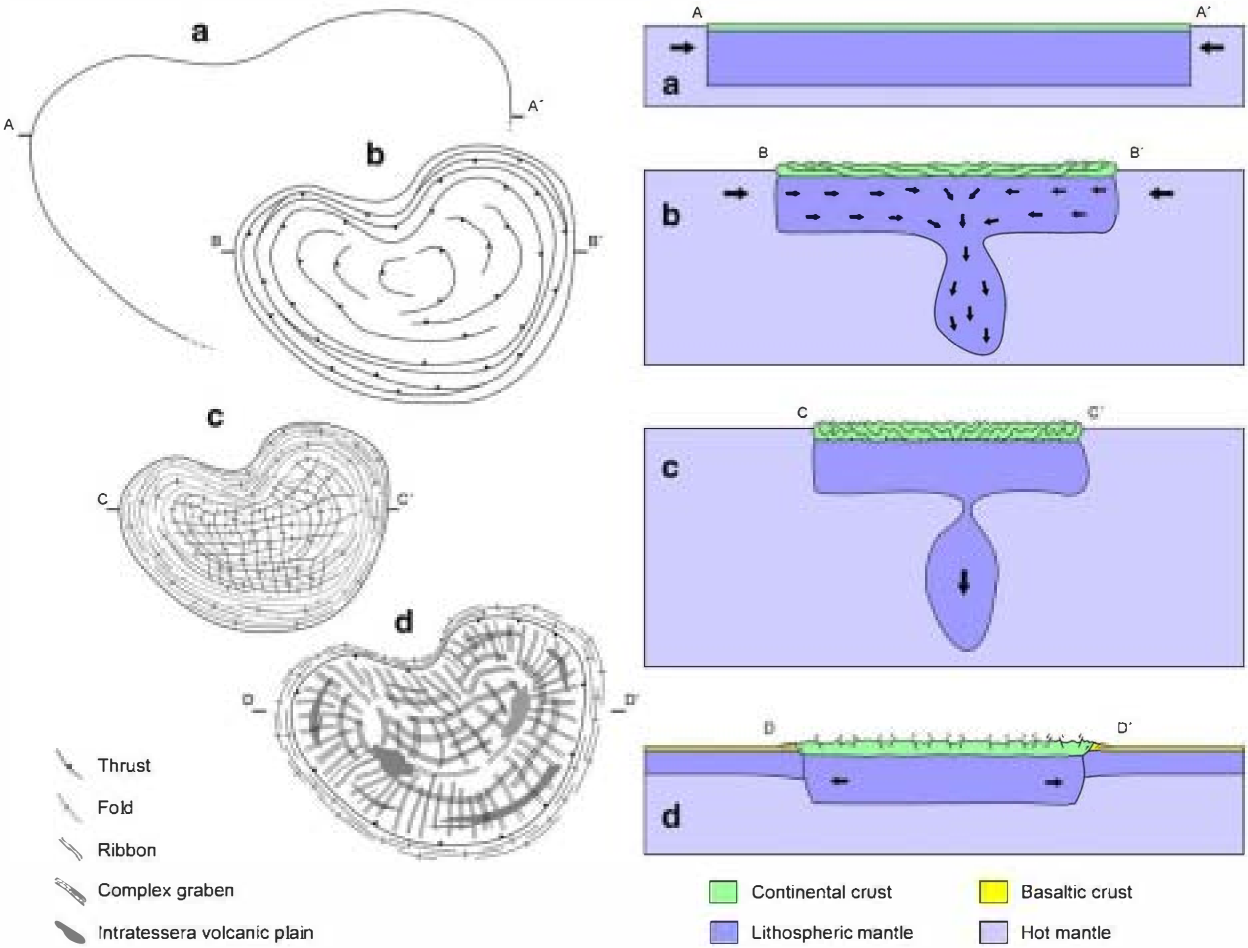


Fig. 6. Tectonic model for the evolution of a continental low density area after a subduction event. a) Collapsed continent consisting of flooded arcuate tessera inliers, featuring previous extensional tectonics (ribbons), that survived a global subduction event. b) The continental area is compressed by the hot surrounding mantle undergoing crustal thickening by concentric thrusting, and the development of a gravitational instability of the cold lithospheric mantle under the center of the plateau-continent. c) The last compressional stage refolds the initially generated fold and thrust belts and develop a basin and dome interference pattern in the center of the plateau. Compression ends when the crustal thickening is enough to reach the force balance equilibrium. d) The generation of a new lithosphere by heat conduction around the plateau-continent produces a change in the force balance yielding moderate gravitational collapse producing radial grabens in the plateau and concentric contractional structures in the adjacent plains. During this extensional stage partial melting of the plateau driven by decompression produces the observed intratessera volcanism.

testing the hypothesis of crustal plateau formation and evolution. Fig. 8a shows a left-looking radar image of Western Ovda. Fig. 8b shows an image of the topography of Western Ovda. The left-looking radar images shown in Fig. 8c d and e correspond to the northern margin, the basin and dome central area and the southern margin respectively. Fig. 8f shows a detailed area of the northern margin, where time relations between structures can be observed.

The amount of shortening accommodated by compressional structures in crustal plateaus has been considered very small by scientists that support the plume model (Ghent and Hansen, 1999) while the supporters of the downwelling model consider that it is the main deformation recorded (Bindschadler et al., 1992a,b). Even the authors that initially supported the plume model have accepted that their model does not explain the large shortening associated with the presence of the small-wavelength folds (Ghent et al., 2005; Hansen, 2006). This led Hansen (2006) to reject the plume model and propose the lava pond and giant impact hypotheses.

Although it is widely accepted that the short (<2 km) and medium (2–10 km) wavelength folds accomplish significant crustal shortening, we consider that the shortening associated with the long-wavelength (10–30 km) compressional structures has been underestimated. Ghent and Hansen (1999) interpret these long-wavelength structures to be gentle folds accommodating very small shortening. These long-wavelength compressional structures are characterized by very wide ridges with a gradual variation of the radar brightness alternating with thin depressions (Fig. 8e). We interpret this structure as the alternation of anticlines separated by thrusts. These concentric thrust belts would correspond to the main shortening and thickening of the continental crust of the plateau. The high compression suffered by the plateau has refolded and made vertical the initially low-dip angle thrusts (Fig. 6c), similarly to continental collisional areas on Earth. Most of the medium wavelength ridges could also correspond to thrusts contributing to a larger shortening than previously considered.

For instance, a large amount of compression associated with the northern border of Ovda Regio has been noted by several authors (King et al., 1998; Tuckwell and Ghail, 2002; Romeo et al., 2005), suggesting a collisional origin that was postulated by Tuckwell and Ghail (2002). Fig. 8c displays a left radar image of this highly compressed margin. The collisional origin of this highly deformed plateau margin yields an important question to answer: What collides against the crustal plateau? It is difficult to explain how the rigid lithosphere of the relatively young regional volcanic plains surrounding the plateau can collide concentrically against the plateau. This problem is solved by our model for crustal plateau formation because this concentric collision is driven by the hot mantle material raised in the last global subduction event. A collision against a soft fluid-like material allows the observed concentric shortening in different directions.

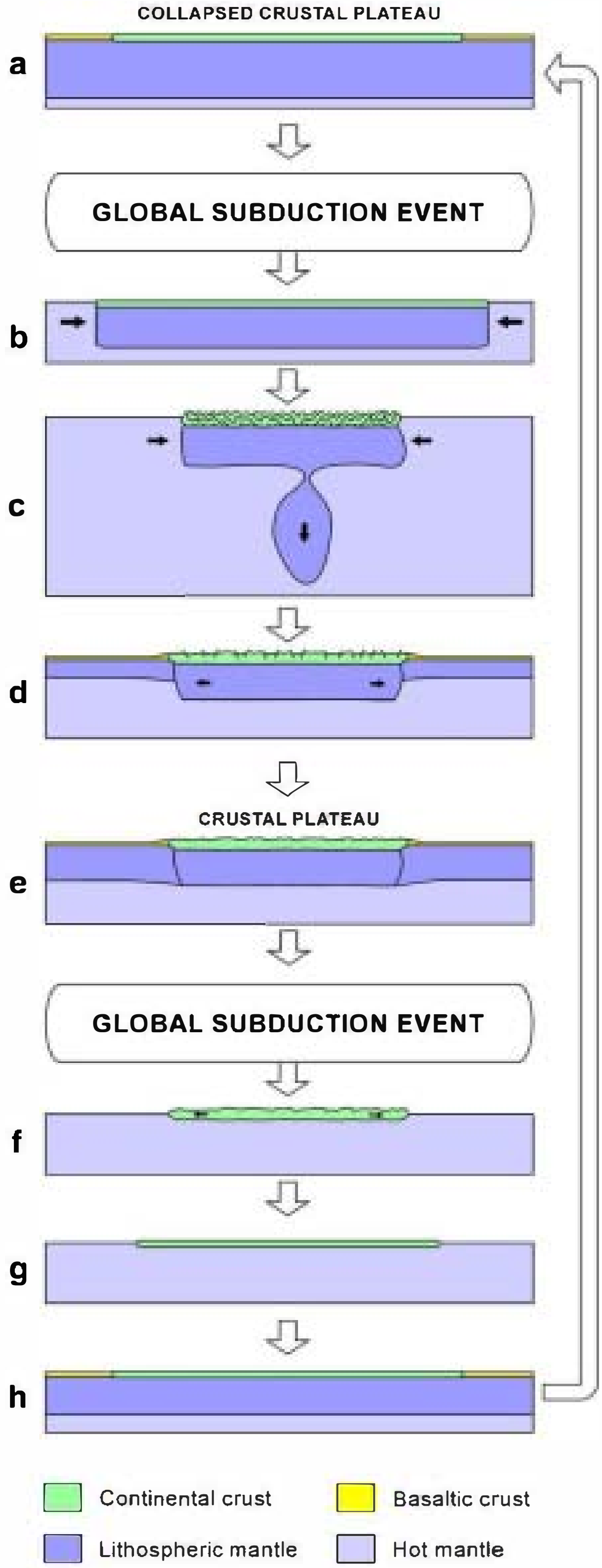


Fig. 7. Model of the cyclic evolution of the continental crust of Venus, relating crustal plateaus and tessera inliers (collapsed crustal plateaus).

The timing of the structures and volcanism in crustal plateaus can be deduced from the detailed structural study performed by Hansen (2006) in a small area of Ovda Regio. The folds and thrusts of different wavelengths appear flooded by the intratessera volcanic plains. This indicates that the observed volcanism is post-compressional because the folds and thrusts never deform the intratessera volcanic plains. Nevertheless the extensional tectonics characterized by small-wavelength long-narrow simple graben (ribbons) and long-wavelength wide complex graben are sometimes flooded by the intratessera volcanic plains and in other cases deform the already emplaced plains (Fig. 8f). Consequently, extensional tectonics is contemporary with volcanism and postdates compression. Both kinds of normal fault association, ribbons and complex wide graben, show the same cross-cutting relationship with the post-compressional intratessera volcanic plains, sometimes predating and sometimes postdating them. These cross-cutting relationships were noted by Romeo et al. (2005); and can also be deduced from the maps of the detailed structural analysis performed by Hansen (2006) in Ovda Regio.

The compressional or extensional character of the first deformation recorded in crustal plateaus has been subject of extensive discussions, because it is the main difference between the plume and downwelling models. The difficulties for reaching a consensus could be due to the possibility that extensional relict fabrics (ribbons) from the previous tessera inlier stage could be preserved in some crustal plateaus.

Both tectonic stages, the compression and the subsequent extension, are characterized by a wide range of wavelengths indicative of different thicknesses of the deformed layer for each wavelength value. The cross-cutting relationships between different structure sets do not provide a time trend in the formation of structures with different wavelengths, on the contrary structures with different wavelengths seem to form simultaneously indicating a complex layered crust. The models that associate each wavelength with a thickness of the brittle layer (Hansen et al., 1999, 2000; Ghent and Hansen, 1999; Hansen, 2006) have to assume that the rheology of the crust is homogeneous during deformation which is unrealistic. Most of the geological processes of rock formation and modification on Earth generally give layered compositions and rocks with multiple anisotropies. The probable volcanic origin of surface rocks on Venus provides a layered structure due to the accumulation of lava floods. Therefore on Venus, like on Earth, the deformation of a complex and probably layered crust is expected to yield structures at different scales with multiple wavelengths.

The radial extensional structures cannot be found in the regional plains surrounding crustal plateaus, they are restricted to the plateau tessera terrain. Nevertheless, some compressional ridges that could correspond to folds or thrusts or both, appear in the area surrounding the plateau indicating moderate compression (Fig. 8c). These external folds are concentric and parallel to plateau margins and are typically formed in the closest area of the regional volcanic plains that are slightly elevated with respect to the external plains.

All the exposed structural observations in crustal plateaus can be explained by our proposed model of tectonic evolution: (1) first an initial compressional stage that thickens the crust raising the plateau by airy isostasy generating thrusts and folds, and (2) followed by an extensional stage with grabens and intratessera volcanism generated inside the plateau and moderate compression in the surrounding volcanic plains.

5.2. Origin of crustal plateau geometries

Both the plume and downwelling models relate the different plateau elevations observed to different degrees of gravitational collapse with similar initial elevations. We propose that the height of each individual crustal plateau, ranging from 0.5 to 4 km, is not due to different amounts of gravitational collapse, but indicate different initial conditions before plateau formation. In our model the main factor that controls the final plateau elevation is the thickness of the lithospheric mantle. From the equilibrium relation of Fig. 2b the plateau elevation at the end of the compressional stage is determined by the lithospheric mantle thickness. Also, the different elevations

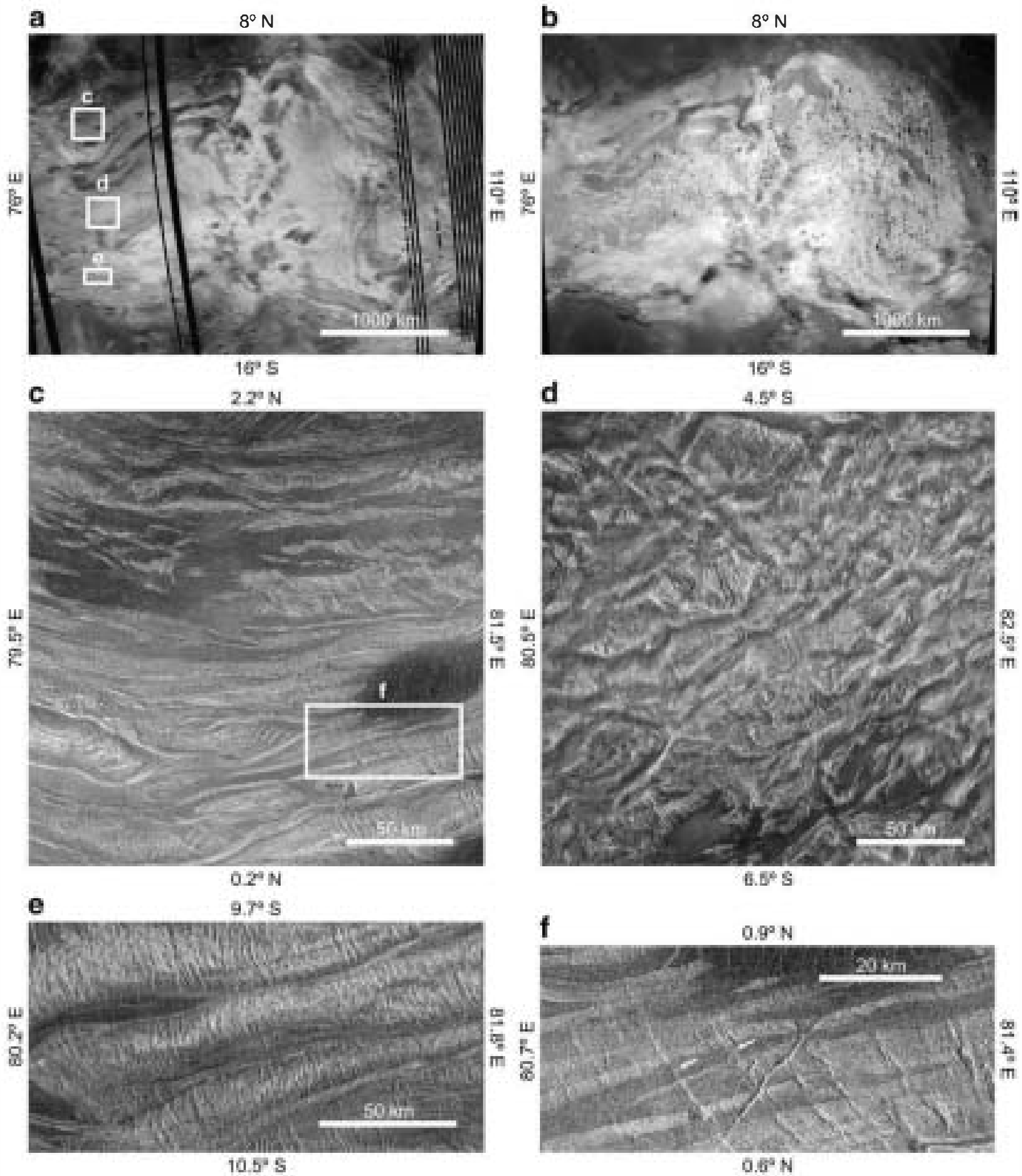


Fig. 8. Examples of different types of tessera terrains in Eastern Ovda, the largest crustal plateau on Venus. a) Left-looking radar image of Easter Ovda, the locations of images c, d and e are shown. b) Topography of Easter Ovda. c) Left-looking radar image of the northern margin, the location of image f is shown; the southern half of the image corresponds to a highly compressed plateau margin with moderate radial extension, while the northern half corresponds to the region surrounding the plateau characterized by moderate compression accommodated by folds without perpendicular extension. d) Basin and dome interference pattern of the central area; e) Left-looking radar image of the southern margin; the long-wavelength folds show narrow valleys between wide anticlines, suggesting the presence of thrust faults between anticlines rather than synclines. f) Detail showing the temporal relationship between narrow graben ("ribbons") and folds; the white arrows indicate locations where ribbons postdate folds because ribbons cross-cut an intratessera volcanic plain that was emplaced between anticlines.

found in the same crustal plateau can be achieved by different thicknesses of the lithospheric mantle that survived the global subduction event in different areas of the same plateau-continent region.

The width of the plateau, from our model, is determined by the initial size of the low-density continental area that will suffer compression. Consequently the observed variety of crustal widths and shapes is due to different initial sizes and shapes of the low density areas. The simple analytical model presented here gives an explanation for the compressive forces associated with plateau formation. Nevertheless, more detailed finite element models are needed to evaluate the suitability of pulsating continents for generating the observed rimmed plateau topography and the arcuate geometries of most of the tessera inliers. The observed rimmed morphologies of some crustal plateaus could be obtain when the downwelling flow associated to the gravitational instability of the lithospheric mantle during compression delaminates a significant portion of the lithosphere leading a thinner lithospheric mantle in the center of the plateau than in the margins.

5.3. The viscosity of the crust and duration of compression

The values of crustal viscosity needed to achieve plateau compression in a reasonable time (50–150 Ma) are in the range

$\eta = 10^{21}$ Pa s to $\eta = 5 \cdot 10^{21}$ Pa s. These low viscosity values are close to the range of effective viscosity of $5 \cdot 10^{21}$–$2 \cdot 10^{22}$ Pa s estimated by England and Molnar (1997) for the Tibetan plateau. More recently, Flesch et al. (2001) have considered the dynamics of the India–Asia collision zone and concluded that a vertically averaged viscosity in the range of $0.5 \cdot 10^{22}$–$5 \cdot 10^{22}$ Pa s is appropriate for Tibet. The collisional area of the Tibetan plateau built up by compressional tectonics has been used as an analog to the biggest Venusian crustal plateau, Ovda Regio, by Romeo et al. (2005). Houseman et al. (2000) calculate the effective viscosity for the continental orogeny of the Transverse Ranges of California, where a lithospheric downwelling flow driven by gravitational instability was suggested, to be at most about 10^{20} Pa s.

The very high viscosity values (10^{24}–10^{26} Pa s) used by Nunes et al. (2004) and Nunes and Phillips (2007), in their modelling of crustal plateaus collapse are not appropriate here, as they used the dry diabase rheology of Mackwell et al. (1998) for the plateau crust. Since we propose that tessera terrain represent low density differentiated crust, the diabase rheology is not a good approach for our model. Although we do not know the real composition of tessera, two facts indicate that we should use viscosities significantly lower that those of the models of Nunes et al. (2004) and Nunes and Phillips (2007): (1) The differentiated continental-like composition that we propose for tessera terrains will be softer than the diabase, and (2) the strong relation of continental crust origin and water content on Earth suggests that Venusian continents should not be fully dehydrated.

5.4. Crater distribution

The distribution of impact craters on the Venusian surface, including tessera terrains and volcanic plains, is indistinguishable from a random distribution (Strom et al., 1994). Although the area covered by all the crustal plateaus is not sufficient to statistically determine their age assuming a contemporary formation, the crater density observed in crustal plateaus is approximately equal to the crater density of the lowlands, indicating an apparently equivalent retention crater age for both provinces. The random fluctuations of the crater density can explain the small variations of crater densities between crustal plateaus. If a diachronous formation is considered the more craterized crustal plateaus could be as much as 1000 Ma old (Nunes et al., 2004) but not older. These observations strongly suggest that the formation of the volcanic plains of the lowlands and the crustal plateaus are linked.

Nevertheless, none of the models for plateau formation proposed so far (downwelling, plume, and impact hypotheses) link the genesis of the regional volcanic plains with the origin of crustal plateaus. On the contrary these models propose that crustal plateaus were diachronously formed over a thin lithosphere on ancient Venus, which is not rejected but is not strongly supported by crater observations. If that were the case, why do not the crustal plateaus show a significantly more craterized surface than the planet average? An explanation could be that post-formation tectonic processes have erased old craters, but even in this case, no genetic relationship is given between the main tectonic events in crustal plateaus and the formation of the volcanic plains.

Our model gives an explanation for the crater distribution observed both in volcanic plains and tessera terrains. The high shortening suffered by crustal plateaux immediately after a global subduction event tectonically erases all the previous craters present in the plateau-continent. Compression is expected to be terminated approximately at the same time that a new surface for the lowlands is produced by volcanism. From this moment, impact craters accumulate in both terrains, some of them being partially modified by extensional tectonics in the plateaus and by local volcanic activity in the lowlands.

The analysis of tectonically modified craters on tessera terrains performed by Gilmore et al. (1996, 1997) shows that the observed craters are never affected by compression but are sometimes affected by extension, suggesting a very rapid compressional event followed by a longer period of extensional tectonics. These observations are in very good agreement with our model. We predict a very rapid compression caused by a catastrophic change in the force balance on the margins of an area of low-density crust (collapsed continent) after a subduction event. The duration of this short compressional event is only controlled by the resistance of the continental crust to deformation ($\approx$50 Ma for $\eta = 10^{21}$ Pa s). This tectonic event is characterized by very large amounts of shortening that completely destroy the previous crater record, consequently no crater affected by compression remains. However, the subsequent extensional event is driven by a force balance change produced gradually on the margin of the plateau due to the growing of a new lithosphere in the lowlands; thus extensional tectonics occurred during a longer period than the initial short compressional event. This explains that some craters appear affected by grabens.

5.5. Continental crust on Venus?

Our model requires tessera terrain to be made up by low density compositions in order to explain why they survived the last subduction event. But the low density is not required by our model to achieve the compressional process that builds up crustal plateaus. Figs. 2–4 indicate that our model works with crust densities of the order of ρ_c=2900 kg m^{-3}.

The composition of the Venusian volcanic plains was analyzed by seven spacecraft that were successfully landed by the Soviet Union between 1972 and 1986: Venera 8,9,10,13,14 and Vega 1 and 2 (e.g., Vinogradov et al., 1973; Surkov et al., 1984). Although the uncertainties of the original rock compositions are important considering the significant degree of weathering and secondary transformations, the analyses indicate a basaltic affinity for the volcanic plains. However, the analyses obtained by Venera 8 and 13 yielded high potassium contents suggesting more differentiated alkaline compositions. Nikolayeva (1990) compared the geochemical data from Venera 8 and 13 with terrestrial analogues concluding that the high potassium content demonstrates the presence of continental crust on Venus. However the Venera 8 and 13 landers analyzed materials from the volcanic plains and there is no geochemical data available from tessera terrains.

Two volcanic morphologies have been proposed as evidence of differentiated materials, the steep-sided "pancake" domes and the large, steep-sided, ridged, radar-bright "festoon" flows (Pavri et al., 1992; Fink et al., 1993; Moore et al., 1992). Nevertheless these volcanic features can also be obtained for basaltic or intermediate compositions (Gregg and Fink, 1995). One of the scarce festoons is located in Ovda Regio (Ovda fluctus) suggesting the possibility of the generation of tertiary crust by partial melting (Head and Hess, 1996) at least in this crustal plateau.

The unimodal hypsometry of Venus shown by the Magellan topography data suggests a homogeneous crustal composition. However, the presence of two types of crust, represented by tessera terrains and volcanic plains, cannot be ruled out considering that: (1) The percentage of area occupied by tessera terrain (our proposal of continental crust) is small (8–9%), (2) the density difference between the tessera crust and the crust of the volcanic plains could be small, and more important (3) tessera terrains, including crustal plateaus and tessera inliers, show a wide range of elevations (~0–5 km) indicative of different Airy compensated crustal thicknesses. All these reasons can explain the lack of a maximum in the hypsometry caused by the Venusian continental crust. McGill et al. (1982) pointed out that Earth's hypsogram might be unimodal in the absence of oceans because they impose an erosional base level that homogenizes the thickness of the continental crust.

The high D/H ratio of the water found in the Venusian atmosphere is evidence for abundant water in the past. This ratio, indicative of the enrichment in deuterium due to the preferential loss of hydrogen from the atmosphere, is 150 times the D/H values of terrestrial water that is presumably equivalent to the primitive water on Venus (Donahue et al., 1997). The generation of protocontinents on ancient Venus with abundant water seems reasonable based on the role that water plays in the differentiation processes that yield continental crust on the Earth. The difficulty of recycling low density materials into the mantle due to their buoyancy might have preserved old continental crust on Venus until today. If Venus lacks continental crust a question remains: How did ancient Venus with abundant water avoid the formation of continents by crustal differentiation processes?

5.6. Ishtar Terra

The model of pulsating continents that explains the observed characteristics of crustal plateaus and tessera inliers also has implications for the origin of Ishtar Terra. Ishtar, a ~10,000 km of diameter highland close to the north pole of Venus, contains very different terrains, with a wide range of elevations, with different kinds of topographic compensation (Kucinskas et al., 1996) and distinct tectonic fabrics (Hansen and Phillips, 1995). In the western part of Ishtar is Lakshmi Planum, a plateau area with elevations of 3–4 km above the mean planetary radius (MPR) covered by smooth volcanic materials. It is surrounded by linear mountain belts (elevated from 5 to 10 km above MPR) featuring parallel ridges and troughs interpreted as contractional structures originated by crustal shortening. The eastern part of Ishtar is formed by Fortuna Tessera, a complex ridged terrain with different cross-cutting tectonic fabrics elevated 2 km above MPR. Other tessera terrains appear adjacent to the mountain belts, called Atropos Tessera to the north and Itzpapalotl Tessera to the west.

The admittance studies of Kucinskas et al. (1996) for Ishtar Terra constrains the kind of compensation of the different provinces. The elevation of the mountain belts is Airy compensated. The large crustal thickness (>60 km) obtained for the mountain belts of Ishtar together with a probable basalt–eclogite phase change strongly suggest a continental kind of crust for these terrains, provided they are older than ~25–50 Ma (Jull and Arkani-Hamed, 1995). Fortuna Tessera is also Airy compensated via thickened crust like other Venusian crustal plateaus. Nevertheless the elevation of Lakshmi Planum seems to be thermally supported with the thermal lithosphere thinned to ~ 100 km (Kucinskas et al., 1996).

Following our model of pulsating continents, both the mountain belts and the adjacent tessera terrains (Fortuna Tessera) would represent continental crust that survived the last global subduction event. The variety of crustal thicknesses can be explained by different initial thickness of the lithospheric mantle that remains after the global subduction event (Fig. 2). An alternative or complementary hypothesis is that a higher contractional strain was absorbed by the mountain belts with respect to the tessera provinces due to a rheological contrast. Though the pulsating continent hypothesis gives an explanation for the compressional stresses that generated the mountain belts and the tessera provinces of Ishtar, it does not account for the thinning of the thermal lithosphere of Lakshmi Planum, for which a recent (~100 Ma) delamination process of a thickened basaltic crust which may have transformed into eclogite was suggested by Kucinskas et al. (1996).

6. Conclusions

The catastrophic resurfacing of Venus (Schaber et al., 1992; Strom et al., 1994), together with the apparent thick thermal lithosphere with a mantle currently heating up (Kucinskas et al., 1996) seems to indicate a time-dependent thermal evolution with periodic subduction events (Turcotte 1993, 1995, 1996). Assuming that Venus loses heat through these periodic nearly-global subduction events and proposing that tessera terrains represent continental crust we propose a model that accounts for the characteristics of both crustal plateaus and tessera inliers, giving an explanation for: (1) The observed variety of elevations of tessera terrains, (2) their tectonic features and cross-cutting relationships, and (3) the observed crater distribution.

When a continental terrain survives a subduction event, it is surrounded by hot mantle material. We have analyzed the force balance on the boundary between the continental area and the adjacent hot mantle. Our results indicate that different ratios of the thickness of the continental crust to the thickness of the underlying lithospheric mantle can yield either compression or extension of the continental area. If the crustal thickness is less than ~2/5 of the lithospheric mantle thickness then the continental area will be compressed, but if the crustal thickness is greater than ~2/5 of the lithospheric mantle thickness it will spread out and collapse. Consequently, depending on the amount of lithospheric mantle that remains under a continental area after a subduction event, it could suffer compression and generate a plateau or it could gravitationally collapse generating tessera inliers. The periodic character of the subduction events could lead to a "pulsating behavior" of the continents of Venus generating alternatively crustal plateaus and tessera inliers. The proposal of tessera terrains as representing Venusian continental crust could be tested by spacecraft analysis of tessera terrains, which has been suggested as a main objective for future landing missions (Basilevsky et al., 2007). If tessera is made up of continental crust, analyzing their composition would also constrain the history of volatiles, in particular water, on Venus, which is critical to understanding terrestrial planet evolution.

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