Evolution of a previously thickened thermochemical lithosphere: Application to Venus

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Abstract. Magellan images have revealed areas with high topography on Venus like Maxwell Montes, interpreted as the result of a lithospheric thickening, associated to a global compressional stress regime. In addition to compressional structures, Maxwell Montes displays younger extensional structures and magmatism at its summit. We present a numerical model to explore the main factors that may explain the chronology of compressional and magmatic events observed on Maxwell Montes. In particular, this study aims to quantify the process of destabilization of a thickened lithosphere by taking into account thermal and chemical instabilities. Indeed, the contribution of basaltic/eclogitic rocks and depleted mantle rocks is not negligible in mantle dynamical models. The initial situation in the upper mantle is represented in a two-dimensional cartesian domain by a stationary state of thermal convection. The thermal lithosphere is then instantaneously thickened and buoyant tracers of depleted mantle and basalt are placed in the lithospheric mantle and crust, respectively. This material distribution controls both the thermal evolution of the lithosphere and its recycling after thickening. Assuming a continuous and slow basalt/eclogite transformation, the lithosphere acts as an insulating layer during 100 Myr before an effective recycling can occur. For an instantaneous basalt/eclogite transformation, combination of basaltic crust and depleted upper mantle prevents recycling of the lithosphere during the first few millions years. The recycling is all the more enhanced as the basalt transforms to eclogite because it is an ongoing process. It is twice as fast (20 Myr) as the recycling of a purely thermal lithosphere (40 Myr) but the recycled rocks of a thermochemical lithosphere represent less than the recycled rocks of a thermal one because of the great stability of the depleted layer. This results in a thermochemical lithosphere thicker by several tens of kilometers. At an upper mantle scale, this chemical diversity favors the mixing of chemical heterogeneities and the perturbation of internal dynamics. This involves the impingement of a hot depleted diapir at the base of the lithospheric root. For a sufficiently hot mantle, a partial melting zone is trapped inside the lithospheric root and induces magmatism about 80-100 Myr after the lithospheric recycling. This could explain the late volcanic flows observed in Maxwell Montes. Detailed modeling of a thermochemical lithosphere also could explain the formation and evolution of Maxwell Montes as evidenced by volcanic and tectonic features.

1. Introduction

Magellan altimetry data have revealed three major topographic highs on Venus, i.e., Beta Regio, Aphrodite Terra, and Ishtar Terra. From the interpretation of radar images, different scenarios of formation have been proposed. Beta Regio, where topography reaches 3 to 5 km and where volcanoes and extensional tectonic structures are present, is usually interpreted as a rift valley [Schaber, 1982; Senske et al., 1991, 1992; Solomon et al., 1992]. Two models have been proposed for the formation of the 2000-km circular features (Ovda, Artemis, Atia, etc.) including coronae, volcanic rises, and tesserae which compose Aphrodite Terra (5-km mean altitude over more than 15,000 km in length). On the one hand, a hotspot model explains the high topography associated with complex tectonics and volcanism [Herrick and Phillips, 1990; Kiefer and Hager, 1991; Grimm and Phillips, 1992; Head et al., 1992; Janes et al., 1992; Squyres et al., 1992; Stofan et al., 1992; Ansan and Blondel, 1996]. In this model the high topography is the result of magmatic processes. On the other hand, a coldspot model is based on the peripheral trenches of the coronae being interpreted as surface expressions of mantle downwelling [Bindschadler et al., 1992; Sandwell and Schubert, 1992].

Similar to both equatorial regions of Beta Regio and Aphrodite Terra, the polar region of Ishtar Terra (altitude >6 km) presents high topography. This relief is interpreted as mountain belts (Akna, Danu, Freyja, and Maxwell Montes) surrounding the high plateau of Lakshmi Planum. The plateau may be due to a downwelling [Bindschadler and Parmentier, 1990] or a plume rising underneath [Pronin, 1986], the overturn of which in the mantle produces mountain belts at the periphery [Bindschadler and Parmentier, 1990; Grimm and Phillips, 1990]. As the elevation induced by such a thermal plume cannot exceed 4 km [Kiefer and Hager, 1991], such mountain belts may result from thickening related to regional compressional stresses [Morgan and Phillips, 1983; Crumpler et al., 1986; Vorder Bruegge et al., 1989, 1990; Head, 1990; Kiefer and Hager, 1991; Solomon et al., 1992]. Considering

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the basaltic composition of the Venustian crust with a density of 3 g/cm³, a crustal root of 110 km would be necessary to support isostatically the 11 km elevation of Maxwell Montes [Surkov, 1983; Surkov et al., 1986]. However, at depths greater than 50-70 km, basalt transforms into dense eclogite [Hess and Head, 1990]. Therefore a 110-km basaltic crust is highly unlikely on Venus. In consequence, such mountain belts can only result from a global thickening of the lithosphere [Kaula et al., 1992; Ansan et al., 1994; Ansan and Vergely, 1995].

Moreover, the internal regions of the mountain belts present extensional structures and magmatic rocks embaying valleys [Kaula et al., 1992]. The lack of water on Venus [Von Zahn et al., 1983] implies that magmatism results from the development of mantelic plumes, which is consistent with the associated extension. The study of the relative chronology of geological structures suggests that Maxwell Montes results from a compressional phase leading to lithospheric thickening, followed by an extensional tectonic phase and magmatism [Ansan et al., 1994]. Such a scenario suggests that the destabilization of the thickened lithosphere has been strong enough to reverse the associated convective field from downwelling to upwelling.

In this study we investigate the modifications of convective structures under a thickened lithosphere and, more precisely, the effects on internal dynamics induced by lithospheric thickening. Indeed, lithospheric thickening creates an unstable cold root which can be recycled into the deeper mantle if the thickening occurs rapidly compared to the heating of the root from underneath [Housenman et al., 1981; Molnar et al., 1993]. The replacement of the cold lithospheric root by hot asthenosphere would provide some magmatism [Housenman et al., 1981] and produce uplift of the above rocks [England and Houseman, 1988, 1989].

Hansen and Phillips [1995] have proposed a mechanism for the formation of Ishtar Terra which includes thickening of mantle residuum and lower crust, in agreement with Magellan gravity and topography data. Chemical composition of the lithosphere also controls its stability [Christensen, 1991; Christensen and Hofmann, 1994; Dupeyrat et al., 1995]. Indeed, a depleted mantle layer stabilizes the lithosphere, whereas basalt/eclogite transformation enhances its recycling in the mantle [Dupeyrat and Sotin, 1995]. The chemical effects of depleted mantle, basalt and eclogite have been investigated by Jull and Arkani-Hamed [1995] in a one-dimensional conductive model. The aim of our study is the quantification of the evolution of a thickened lithosphere, taking into account both thermal and compositional aspects in a dynamical model. The stability of a thickened thermochemical lithosphere has been investigated by Lenardic and Kaula [1995] for the case of Tibet, considering two chemical components, light crust, and dense mantle. In our model, we have considered a more realistic lithospheric composition and structure, with an upper basaltic layer (crust) and a depleted mantle layer (lithosphere) over a primitive mantle layer (asthenosphere). Moreover, transformation from light basalt to denser eclogite is taken into account.

We will first present the model and the initial state of an instantaneously thickened lithosphere. Then we will do a comparative study of the evolution of purely thermal lithosphere and thermochemical lithosphere. We will explore the conditions required to obtain postthickening magmatism.

2. Model

2.1. Boundary Conditions and Venus Parameters

Discussions of numerical models of thermal convection in the Venustian mantle show that the style of convection in the Venustian mantle is still controversial. First, the presence of an upper rigid conductive lid decreases the degree of layering according to Steinbach and Yuen [1992] whereas Schubert et al. [1997] found that the degree of layering is greater with a rigid upper boundary than with a free-slip upper boundary. Venustian models also take into account the greater depths of phase transitions in Venus compared to Earth (440 km instead of 410 km and 740 km instead of 670 km, for the exothermic and endothermic phase changes, respectively). Three-dimensional spherical models of convection with phase changes [Tackley et al., 1993, 1994; Schubert et al., 1997] show that the system is never completely layered or completely whole mantle. The Venustian mantle may be partially layered both spatially and temporally.

Therefore, because Earth and Venus present comparable sizes (with average radii of 6378 km and 6052 km respectively), similar basaltic surface rocks [Surkov, 1983; Surkov et al., 1986] and a common origin, we arbitrary choose to model a Venustian upper mantle 670 km thick in a 2-D Cartesian box of aspect ratio 4 with 65 x 257 grid points, providing a spatial resolution of 10 km.

All boundaries are free-slip boundaries, except the top boundary, which presents a no-slip condition in order to model the immobile surface of Venus. The top and bottom boundaries are isothermal, at temperatures of 700 K and 2000 K respectively, whereas the lateral boundaries are reflective (i.e., \( \frac{dT}{dx} = 0 \)). Typical Earth values are used for the coefficient of thermal expansion (\( \alpha = 3.14 \times 10^{-5} \text{K}^{-1} \)),

a) before instantaneous thickening

b) after instantaneous thickening (f=1.5)

Plate 1. The Venus upper mantle is modeled in a 2-D Cartesian box (670 km depth x 2680 km width), with a spatial resolution of 10 km. The temperature varies from 700 K at the surface (purple) to 2000 K at the base of the upper mantle (red). Plate 1a shows the thermal field before and Plate 2a after instantaneous thickening of the lithosphere with a factor \( f = 1.5 \). The average surface heat flux is reduced from 66 to 44 mW/m². The partial melting zones (in white) are reduced by about 50%, whereas lithospheric roots have increased in depth and in width.
Plate 2. Case of purely thermal convection. Evolution of the internal dynamics from the initial state of an instantaneously thickened lithosphere (Plate 1b). At each time (23 Myr, 54 Myr, 85 Myr), the thermal field and partial melting zones (upper box), the location of the passive tracers of depleted mantle (middle box), and the location of the crustal tracers (basalt or eclogite) (bottom box), are represented. The color scale for crust and depleted mantle tracers allows one to distinguish close tracers.
thermal diffusivity (K=10^{-6} m^2/s) and the viscosity (\eta=10^{21} Pas), whereas the gravity field g is assumed to be 8.8 ms^{-2}. Such values provide a thermal Rayleigh number equal to 3.6x10^5.

2.2. Fluid Mechanics Equations and Equations of State

The temperature T and the velocity \mathbf{v} (v_x, v_y) are determined by solving the 2-D equations of conservation of mass (1), energy (2), and momentum (3):

\[ \nabla \cdot \mathbf{v} = 0 \tag{1} \]

\[ \frac{dT}{dt} + \mathbf{v} \cdot \nabla T = K \nabla^2 T \tag{2} \]

\[ \nabla^2 \psi = -\frac{g}{\eta} \frac{dp}{dx} \tag{3} \]

and assuming the stream function \psi is simply related to the velocity (v_x=-\partial \psi/\partial z, v_y=+\partial \psi/\partial x). The production of internal heat is neglected (2) in the upper mantle, since most of the heat is produced in the lower mantle. The upper mantle is assumed to be incompressible and isoviscous (3).

The Boussinesq approximation allows us to neglect density variations except in the expression of buoyancy (3). In an isoviscous case, the buoyancy force appears to be controlled by the lateral density gradient (3). In order to take into account both thermal and chemical buoyancies, the density is expressed as a function of temperature and nature of the rock. For a mantle rock the density is expressed as a function of its degree of depletion (4),

\[ \rho_{mantle}(T,deple) = [\rho_{mantle}(T_{ref}) - 250 \cdot deple][1-\alpha(T-T_{ref})] \tag{4} \]

whereas for a crustal rock, it depends on its mineralogy, basalt or eclogite (5),

\[ \rho_{crust}(T,deple) = [\rho_{crust}(T_{ref})(1-deple) + \rho_{eclogite}(T_{ref})(1-deple)][1-\alpha(T-T_{ref})] \tag{5} \]

where deple is the degree of depletion of the mantle rock (0<deple<0.3), \alpha is the coefficient of thermal expansion, \rho_{mantle}(T_{ref}) (=3340 kg/m^3), \rho_{crust}(T_{ref}) (=2800 kg/m^3), and \rho_{eclogite}(T_{ref}) (=3500 kg/m^3) are the densities at the reference temperature T_{ref} (=1400 K) for an undepleted mantle rock, a basaltic rock and an eclogitic rock, respectively, and \rho_{mantle} is equal to 1 and 0 when the crustal rock is in its basaltic and eclogitic phase, respectively. The transformation of basalt into eclogite is formulated from the phase diagram of the basalt-eclogite system proposed by Wyllie [1971] and modified by Anderson [1980]. Then basalt is stable until 25 km in depth, and eclogite is stable below 50 km in depth. At depths between 25 km and 50 km, the basalt is assumed to transform into eclogite when it reaches the temperature T_{nw}(z) (6):

\[ T_{nw}(z) = 48.6z - 942 \tag{6} \]

where the temperature and depth are in K and km, respectively.

2.3. Initial State of a Thickened Thermochemical Lithosphere

2.3.1. The thermal lithosphere. For the range of parameters considered above, a steady state of thermal convection is obtained and can be characterized by its average thermal profile (Figure 1). The thermal lithosphere usually presents an important vertical thermal gradient, compared to the nearly isothermal asthenosphere. Indeed, the upper thermal boundary layer presents a vertical gradient of 14 K/km, consistent with a value <25 K/km deduced by Zuber [1987]. Therefore the limit of this thermal boundary layer is defined by the node at which the vertical derivative of the horizontally averaged vertical geotherm approaches zero [Lenardic and Kaula, 1995]. Once this criterion has provided a thermal limit, which in our case is 1655 K (Figure 1), the lithospheric thickness is then defined by the corresponding isotherm. Such a criterion provides an average lithospheric thickness of 118 km.

Thickening is supposed to occur instantaneously and to affect the whole lithosphere. During thickening, the initial thermal gradient is divided by the thickening factor f. Then the lithospheric isotherms are simply displaced downward according to T_{nw}(z)=T_{nw}(z/f). The base of the thermal lithosphere is defined by the isotherm 1655 K. This provides an average thickness of the thermal lithosphere equal to 118 km and 177 km, before and after thickening, respectively.

\[ \text{Figure 1. Average vertical thermal profiles before and after thickening (f=1.5). During instantaneous thickening, the lithospheric isotherms are displaced downward according to } T_{nw}(z)=T_{nw}(z/f). \text{ The base of the thermal lithosphere is defined by the isotherm 1655 K. This provides an average thickness of the thermal lithosphere equal to 118 km and 177 km, before and after thickening, respectively.} \]
(equal to 1655 K), has been displaced from 118 km to 177 km in depth (Figure 1).

2.3.2. The chemical lithosphere. The thickness of the chemical and mechanical lithosphere is supposed to be 40% of the thickness of the thermal lithosphere [Jordan et al., 1989]. This provides a thickened chemical lithosphere (h=1.5) with an average depth of 71 km (40% at 177 km) and an associated temperature equal to 1332 K. This reference temperature defines the base of the initial chemical lithosphere.

The thickened chemical lithosphere is supposed to be composed of a basaltic and a depleted peridotite layer over an asthenosphere composed of primitive mantle. From the average degree of depletion of the mantle, defined as the amount of melt extracted, we compute the depth of the crustal layer. In this example, an average degree of depletion of 30% provides a crustal thickness of 30% of the lithosphere thickness, 15 km and 22.5 km before and after thickening, respectively. These values are consistent with the 10-20 km crustal thickness deduced from the spacing of tectonic features and the depth of impact craters [Zuber, 1987, Grimm and Solomon, 1988, Zuber and Parmentier, 1990]. In order to represent basaltic crust and depleted mantle in the lithosphere, crustal and mantle tracers with a size of 2.6 km are initially placed in the thickened lithosphere.

2.4. Evolution of the Internal Dynamics of the Coupled Lithosphere/Asthenosphere

At each time step the chemical effects are computed through three successive stages (see Dupuyrat et al. [1995] for a detailed description for the method). First, the displacement of each tracer is computed with a fourth-order Runge-Kutta method. Second, the chemical state of depleted mantle tracers is kept unchanged (cf. discussion), whereas crustal tracers eventually transform from basalt to eclogite and vice versa. Finally, the chemical buoyancy at each grid point is computed from the average density of tracers located in the vicinity of the grid point, taking into account crustal tracers, depleted mantle tracers, and also “invisible” tracers of undepleted mantle.

Three cases are investigated, for which different categories of tracers are activated. In the first experiment, only thermal dynamics is studied, and passive tracers provide information only about localization of lithospheric rocks. In the second experiment, mantle and basaltic tracers are active, but the transformation from basalt to eclogite is not taken into account. In the last experiment, the transformation from basalt to eclogite is added to the previous case. The results of these experiments are compared, in order to evaluate separately the thermal and chemical effects of the different components in the recycling of the thickened lithosphere.

3. Results

3.1. Postthickening Dynamics and Evolution of the Thermal Field

The initial stationary state of thermal convection (Plate 1a) is characterized by a Rayleigh number of 3.6x10^5 and an average surface heat flux of 66 mW/m^2. Such parameters provide convective cells with a 1:1 aspect ratio, and allow some partial melting in zones about 200 km wide. The instantaneous thickening of the thermal lithosphere with a thickening factor f of 1.5 (Plate 1b) provides wider and deeper lithospheric roots, partial melting zones reduced to about 50%, and a decrease of the average surface heat flux to 44 mW/m^2. Such an evolution illustrates the resulting cooling of the lithosphere.

In the first case, only purely thermal convection is investigated (Plate 2). During the 100 Myr following the thickening, the convective pattern remains unperturbed. The recycled lithosphere remains along the boundaries of convective cells. Fluctuations of the sizes of the partial melting zones (Plate 2) illustrate fluctuations of the intensity of the convection.

In a second case a thermochemical lithosphere is considered, composed of mantle rocks, depleted by 30% and basalt. The evolution of the thermochemical lithosphere after thickening (Plate 3) differs significantly from the evolution of the purely thermal lithosphere. The negative chemical buoyancy associated to the less dense depleted mantle and basalt slows down the downwelling of the root (Plate 3a). This provides a more efficient heating of the lithospheric root (Plate 3a) than in the purely thermal case (Plate 2a). Simultaneously, the lithospheric root becomes wider (compare Plates 2a and 3a), since most of the depleted mantle material concentrates in the root while basalt remains near the surface. Then, after some less dense depleted mantle has reached the base of the upper mantle, both thermal and chemical forces act upward, and the intensity of convection increases (Plate 3b). The downwelling becomes colder whereas the upwelling seems to remain unperturbed (Plates 3b and 3c).

After 92 Myr, no basaltic recycling has occurred, and only a small amount of depleted mantle has been recycled, forming lobate patterns (Plate 3d).

The last experiment builds upon the previous one, by adding the transformation from basalt to eclogite. When downwellng basalt transforms to denser eclogite, it provides a negative chemical buoyancy which adds to the thermal one and favors recycling of the lithospheric root (Plate 4a). Therefore both the transformation from basalt to eclogite and destabilization of the lithospheric root are kept going. This transformation to eclogite also provides important perturbations of the internal dynamics. For instance, some new areas of downwelling and upwelling initiate (Plates 4b and 4c) and the major upwelling is affected, inducing fluctuations of the partial melting zones. At 115 Myr (Plate 4e), some warm depleted mantle rises and penetrates the cold lithospheric root. This interesting feature could be at the origin of a melting process affecting either mantle rocks carried by the chemical plume or basaltic rocks. The fluctuations of internal dynamics favor the mixing of heterogeneities. After 115 Myr (Plate 4c), depleted mantle and eclogite are dispersed in the whole mantle even if most of the depleted mantle and crust remain localized near the surface and in the root.

3.2. Evolution of a Few Parameters After Thickening

The instantaneous thickening leads to a rapid cooling of the lithosphere, marked by a heat flux drop of 20 mW/m^2 (from 66 mW/m^2 to 44 mW/m^2) (Figure 2) and a decrease in...
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(1) thermal convection
(2) = (1) + chemical convection of depleted mantle and basalt
(3) = (2) + transformation basalt/eclogite

Figure 2. Evolution of the average surface heat flux during and after thickening for the three cases investigated. The vertical arrow indicates the decrease due to instantaneous thickening (from 65 mW/m² to 44 mW/m²). See text for description.

size of the partial melting zone of 70% (Figure 3a). After thickening, the evolution of the overall intensity of convection can be illustrated by the evolution of the average surface heat flux (Figure 2). Conversely, the evolution of the local intensity of upwelling and downwelling can be represented by the evolution of the volume of partial melting (Figure 3a) and the recycling ratio (Figure 3b), respectively. A similar evolution is observed for these parameters, which show a first phase of increase followed by a relative stabilization phase. The lithospheric recycling is indeed a phase during which convection is enhanced by the sinking of the thickened lithosphere, which ends when the lithospheric root reaches the bottom of the convective fluid. For Houseman et al. [1981], it is marked by the maximum peak of kinetic energy. Buck and Toksöz [1983] quantify the vigor of the flow by the average dissipation and associate the maximum dissipation value to the first removal of the boundary layer. For Lenardic and Kaula [1995] and Lenardic et al. [1995], the decrease in thickness of the thermal boundary layer marks the end of recycling. The recycling time can also be calculated as an increased function of the thickening factor, viscosity and thickness of the thermal boundary layer [Houseman et al., 1981; Molnar et al., 1993]. In the present study the end of recycling is characterized by the maximum value of the average surface heat flux, the partial melting zone sizes, the decrease of the recycling rate (recycling ratio divided by time), and a relative stabilization of the average thickness of the lithosphere after a previous decrease of its thickness. The duration of this first phase is identical for all these parameters and provides the duration of effective recycling.

3.2.1. Average surface heat flux. For a thickened thermal lithosphere, the average surface heat flux increases and reaches around 55 mW/m² after 40 Myr (Figure 2). The increase of the average surface heat flux corresponds to reheating of the lithosphere previously cooled by thickening. When the chemical buoyancy of depleted mantle and basalt are included, the average surface heat flux does not evolve significantly. This suggests that the chemical buoyancy related to the depleted and basaltic composition of the thickened lithosphere balances the thermal buoyancy related to thickening. The addition of the basalt-eclogite transformation generates a small increase in average surface heat flux, which may be due to the convective recycling of cold lithosphere, replaced by hot asthenosphere. Nevertheless, this effect does not compensate totally the stability generated by the depleted and basaltic composition of the lithosphere. Indeed, in that third case, the average surface heat flux reaches a smaller maximum value (45 mW/m²) than in the purely thermal case (55 mW/m²). This lower flux is the result of the insulating effect of the stable basaltic and depleted lithosphere. Finally, the recycling occurs more rapidly for a thermochronical (20 Myr) than for a thermal lithosphere (40 Myr).

3.2.2. Partial melting zone size and recycling ratio. The local-scale behavior of upwelling (Figure 3a) is similar to the average convective behavior (Figure 2). After thickening, the first phase of increase in size of the partial melting zones illustrates the convective reheating, whereas later fluctuations in melting zones illustrate modifications of the convective pattern.

Because our main concern is the evolution of a thickened lithosphere, we focus on the downwelling zones (Figure 3b). We define a recycling ratio corresponding to the percentage of lithosphere tracers which have traversed a given depth, arbitrarily chosen as 450 km.

In the case of a purely thermal lithosphere, the recycling ratio increases linearly to 40% after 40 Myr. In the case of a depleted and basaltic lithosphere, the very low recycling ratio

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Figure 3a. Evolution of the number of grid points (nfus) located in a partial melting zone. nfus illustrates the cumulated size of partial melting zones in arbitrary units. The vertical arrow indicates the instantaneous decrease in size (from 680 to 200 grid points) of the partial melting zone due to instantaneous thickening.
Plate 3. Case of thermochemical convection, taking into account the chemical buoyancy of depleted mantle-basalt. For details, see Plate 2.
When the transformation of basalt to eclogite is added, the lithosphere is more stable than a thermal lithosphere during the first few millions years after thickening. However, when the crustal basalt reaches the stability domain of eclogite, there is a phase of very effective recycling during which the recycling of the thermochemical lithosphere is more efficient than the recycling of a purely thermal lithosphere. Indeed, between 12 and 20 Myr, the recycling occurs at rates of 1%/Myr and 1.6%/Myr for a purely thermal and for a thermochemical lithosphere, respectively. Finally, after this phase of effective recycling (20 Myr), the stabilizing effect of depleted mantle and basalt results in a lower recycling ratio for a thermochemical lithosphere than for a purely thermal one.

3.2.3. Average thickness of the thermal lithosphere. The average thickness of the thermal lithosphere is calculated from the local thicknesses of the thermal lithosphere localized by the thermal limit of 1655 K (cf. Figure 1).

First, we notice that this thickness of the thermal lithosphere is different from that calculated from the average geotherm (Figure 1). Instead of 118 km and 177 km of Figure 1, we have 160 km and 200 km before and after thickening, respectively. Indeed, in both cases, we use the same criterion of \( dT/dz \) going down to zero to define the base of the thermal lithosphere. But to determine the initial lithosphere, we apply this criterion on the average geotherm, whereas during the time evolution, we apply it locally, in order to define the local thickness of the thermal lithosphere from which we compute the average thickness.

We can set apart the case of chemical convection of depleted mantle and basalt because it gives extreme results, with an important increase of the lithospheric thickness due to the great stability of the chemical components. For the two other cases, as we saw previously for the heat flux, the partial melting zone and the recycling ratio, a first phase of effective recycling during which the average depth of the thermal lithosphere decreases, is followed by a phase of relative stabilization (Figure 4a) with the same recycling time as previously pointed out (40 Myr and 20 Myr for the thermal and thermochemical lithosphere, respectively). The thermal lithosphere, thickened to 200 km, nearly recovers its initial thickness of 160 km after 40 Myr. During a few million years, chemical buoyancy of depleted mantle and basalt increases the lithospheric thickness, but when the basalt/eclogite transformation occurs after a few million years, it enhances the recycling until 20 Myr, but the chemical lithosphere stabilizes at a greater depth (175 km) than a purely thermal one (160 km). The addition of 15 km in thickness for the thermochemical lithosphere can be correlated with a lower surface heat flux (-15 mW/m²) and illustrates the chemical effects of depleted mantle and basalt.

3.2.4. Average thickness of the chemical lithosphere. Similarly, the average thickness of the chemical lithosphere (Figure 4b) is calculated from the local thicknesses of the chemical lithosphere. The transition from the chemical lithosphere to the primitive mantle underneath is supposed to be reached when the number of lithospheric tracers per grid unit decreases below a critical number. For any critical number between 0 and 16 tracers per grid unit, the same average chemical lithosphere thickness is obtained. In the case of thermal convection, the chemical lithosphere has no physical meaning, as it simply represents the localization of passive tracers initially placed above the 1332 K isotherm. Except during the first 10 Myr after thickening, the relative position of the two curves is the same for the thermal (Figure 4a) and chemical (Figure 4b) lithospheres. This shows the strong thermochemical coupling effect. In the case of thermochemical convection, the thickness of the chemical lithosphere recovers its initial value (68 km).
### 3.3. Postthickening Magmatism

One issue of this paper is to determine if a correlation exists between the thickening of the lithosphere during the formation of Maxwell Montes and the volcanism observed inside the mountain belt, interpreted as a late magmatism [Ansan et al., 1994]. Magmatism related to a local temperature increase can be induced by thickening through different ways. First, the local temperature may increase because of an increase in local internal heating related to the addition of new radioactive elements due to crustal thickening. This magmatism supposes a relatively slow thickening unlikely to occur on Maxwell Montes [Vorder Bruegge and Head, 1991] or a high temperature gradient. Another mechanism would be that after recycling the lithospheric root is replaced by hot asthenosphere. This could eventually raise the local temperature above the solidus and provide some melting [Houseman et al., 1981; Kiefer and Hager, 1991]. In the particular case presented here, our model does not show any magmatism under the lithospheric roots during the first 100 Myr. However, our model predicts that some hot depleted mantle plumes are able to penetrate the cold lithospheric root when the basalt/eclogite transformation is taken into account.

If we consider the case of a hotter Venusian mantle by imposing a greater temperature at the base of the upper mantle (Ti=2500 K instead of 2000 K), we obtain some postthickening magmatism. This hotter mantle provides the conditions for large-scale melting and can therefore generate some global resurfacing, except in the zones of mountain formation by lithospheric thickening (Plate 5a). Both chemical effects will then allow some magmatism at the emplacement of the mountain. After 20 Myr, as the recycled lithospheric root will progressively reheat, both thermal and chemical buoyancy of warm, depleted, recycled mantle are positive (Plates 5a and 5b). This resulting upwelling goes transversally and penetrates the lithospheric root, heating it (Plates 5a and 5b). The lithospheric thickening also enhances the recycling of basalt to eclogite and creates some new crustal instabilities (Plate 5a). In particular, a new downwelling appears at about 100 km distance from the cold root. This cold instability grows and cuts the preexisting melting zone, trapping a melting zone between the two downwellings. This short-lived partial melting zone is heated further by the depleted mantle diapir. This trapped melting zone persists only a few tens of million of years, from 90 Myr to 110 Myr after the thickening. In order to correlate thermal predictions to the mechanical state, we also present the difference of the normal stresses which are exerted on the fluid layer constituting the overriding part of the mantle [Flettout and Froidevaux, 1982]. The corresponding state of stress is extensional or compressional for a positive or negative value of $(\tau_{xx} - \tau_{zz})$. In Plate 5, we verify that the initial downwelling is associated to a compressional state of stress (20 Myr). However, after 80-100 Myr, a small extensional zone appears in the initial compressional one from 80 Myr to 100 Myr, which precedes and remains contemporary with the development of the trapped melting zone observed on the thermal pattern.

### 4. Discussion

#### 4.1. Model

**4.1.1. Rheology of the lithosphere.** In our constant viscosity model, the high viscosity of the lithosphere is not taken into account. In a model with a viscosity contrast smaller than $10^{5}$, a higher viscosity lithosphere behaves almost as a lithosphere included in a constant viscosity model [Solomatov and Moresi, 1996]. But a very viscous lithosphere moves so slowly that the cooling by convective recycling is less efficient than conduction [Solomatov and Moresi, 1996]. Therefore, the recycling time of the lithosphere is longer in a variable viscosity model than in an isoviscous model [Lenardic and Kaula, 1995]. In consequence, the recycling times obtained in our isoviscous model are underestimated.

**4.1.2. Modelling of the mantle differentiation.** The evolution of the thermal field controls the evolution of partial melting, which can produce new basalt and depleted mantle. In this study, we do not take into account this new production of crust and depleted mantle after thickening. One may argue that the production at upwelling zones of new depleted mantle and crustal rock should modify the dynamics of the whole mantle. But this effect is not significant for the temporal limits of this study because, during the maximum 120 Myr investigated, the quantity of new material produced is small and its influence remains located close to the upwelling. Moreover, this study focuses on the evolution of downwelling, which is not affected by differentiation processes in Venus. Whereas on Earth, water allows partial melting in the downwellings, the lack of water on Venus prevents such differentiation.

**4.1.3. Instantaneous or evolutive thickening.** In most cases, the lithospheric thickening results from global tectonic compressional stresses. Their effects on the lithosphere dynamics may be modeled by imposing the fact that velocity in the thermal equation depends on a strain rate related to compression [England, 1987; Vorder Bruegge and Head, 1991; Namiki and Solomon, 1993].

A typical time for Earth-like lithosphere shortening is about 20 Myr to 50 Myr [Molnar et al., 1993]. For Tibet,
Plate 4. The transformation to basalt to eclogite is added to the case in Plate 3. Case of thermochemical convection, taking into account the chemical buoyancy of depleted mantle, basalt, and phase transformation of basalt-eclogite. For details, see Plate 2.
Plate 5. Evolution of the temperature field, the compositional field, and the state of stress for a hotter Venusian mantle (temperature at the base of the upper mantle $T_1=2500$ K).
assuming a convergence velocity of 50 mm/yr [Molnar et al., 1993], the shortening time is at least 20 Myr for 1000 km of convergence. In an instantaneous thickening model applied to Tibet [Houseman et al., 1981], the recycling time of the lithospheric root is of the same order of magnitude as the shortening time.

For Venus, the shortening time is difficult to estimate because of the lack of relative age information of geologic features and of displacement velocities. However, the calculated shortening times and the calculated thermal evolution strongly depend on the strain rates included in the thermal equation. For example, Namiki and Solomon [1993] have estimated a timescale of 510 Myr and 51 Myr for the thickening of the basaltic crust under Maxwell Montes, taking into account strain rates of $10^{-4}$ s$^{-1}$ and $10^{-5}$ s$^{-1}$, respectively. On the other hand, considering strain rates of $10^{-5}$ s$^{-1}$ and $10^{-6}$ s$^{-1}$, respectively, the shortening of a thickened basaltic crust (110 km) under Maxwell Montes requires 65 Myr to 6.5 Myr [Vorder Bruegge and Head, 1991].

From a thermal point of view, considering an initial thermal gradient of 15 K/km, Vorder Bruegge and Head [1991] have shown that for a relatively small strain rate ($10^{-5}$ s$^{-1}$), the crustal heating by radioactive heat production is dominant, leading to a global crustal heating of the order of 300 K during the 65 Myr of shortening. On the other hand, for a strain rate equal to $10^{-4}$ s$^{-1}$, as the cooling by shortening is relatively fast, the radioactive heating is minor, resulting in a temperature decrease of about 300 K during the 6.5 Myr of shortening. At the lithospheric scale of our model, the radioactive heating can, at any rate, be neglected, as radioactive elements are essentially concentrated in the basaltic crust.

The approximation of an instantaneous lithospheric thickening is valid in the context of a rapid shortening of Maxwell Montes. Furthermore, the lack of deformed volcanic flows on Maxwell Montes [Schaber et al., 1987; Kaula et al., 1992; Ansan et al., 1994] suggests that there was no volcanism during the shortening, which favors a model of rapid thickening [Vorder Bruegge and Head, 1991; Namiki and Solomon, 1993]. In view of these results, a model can be used, consisting in instantaneous downward displacement of the lithosphere isotherms [Houseman et al., 1981].

A local crustal thickening is triggered by thermochemical thickening in the downwellings, related to the relatively small density of basalt [Lenardic and Kaula, 1995; Lenardic et al., 1991, 1993, 1995]. In our model, we integrate the two thickening processes by coupling effects of global compositional forces (global displacement of isotherms) and chemical effects of basalt. In addition, the chemical effects of depleted mantle and eclogite are taken into account.

4.2. Results

Heating of the lithosphere occurs simultaneously by thermal conduction and convective recycling. The characteristic time for conductive heating is about 130 Myr ($\eta/\eta T I K$, with $\eta$ as the thickness of the thickened plate (200 km) and $K=10^{10}$ m$^2$/s). In our purely thermal case, for typical values of Venusian parameters and under the condition of layered convection, a stable thermal state is reached in about 40 Myr after a 50% thickening ($f=1.5$). Such a value is comparable to those obtained by Houseman et al. [1981] for Earth’s thermal lithosphere. After Molnar et al. [1993], the thermal instability is attenuated very quickly, in about 0.6-6 Myr. If a temperature dependent rheology is included, this timescale is increased to 30-100 Myr [Buck and Toksoz, 1983]. Therefore, in the case of a purely thermal convection, this relatively short characteristic time favors a convective process which yields to lithospheric recycling. This recycling may be characterized not only by the recycling time but also by the quantity of removed material.

Houseman et al. [1981] have shown that the removal of the instability occurs more rapidly for larger values of $f$, $Ra$, and $a/d$, where $a$ and $d$ are the conductive and convective layer thicknesses, respectively. The role of the depleted layer in the stability of the lithosphere has been investigated by Parmentier and Hess [1992], Dupeyrat and Sotin [1995], and Head et al. [1994] in their isoviscous convective models. The lithosphere recycling is either continuous and slow [Dupeyrat and Sotin, 1995], or episodic and rapid [Parmentier and Hess, 1992; Head et al., 1994]. In the model of Dupeyrat and Sotin [1995], the low recycling ratio of a normal lithosphere is due to the competition between upward chemical buoyancy and downward thermal buoyancy. For a previously thickened lithosphere, its recycling is inhibited by the chemical buoyancy of depleted mantle and basalt. The episodic recycling time (500 Myr) is related to a first phase of thickening in a conductive context followed by a phase of instability when the depleted layer exceeds a critical depth [Parmentier and Hess, 1992; Head et al., 1994]. The differences between the two models may be due to the fact that in the latter model the basaltic and depleted layer does not participate in convection until a critical depth. However, this model simulates the strong rheology of the mechanical lithosphere. If the slow kinetics of the basalt-eclogite transformation was confirmed, this would prevent effective recycling during the 100 Myr following the thickening because of the great stability of the Venustian lithosphere.

Taking into account the two opposite chemical buoyancies of depleted mantle/basalt and eclogite, the recycling time (about 20 Myr) is about 50% smaller than in a purely thermal case. Whereas in a purely thermal case, the thermal lithosphere recovers its initial thickness (160 km), in the thermochemical case, the thermal lithosphere stabilizes at an average value of about 180 km because of the great stability of the depleted layer [Dupeyrat et al., 1995]. Considering a chemical lithosphere, in the case of a thermochemical convection, the lithosphere tends to stabilize at its thickness before thickening. In our model the basaltic/eclogite transformation occurs instantaneously. In fact, Ahrens and Schubert [1975] suggest this transformation needs at least 100 Myr to occur in a dry medium. Under such conditions, our results obtained for a basaltic and depleted lithosphere are valuable during the first 100 Myr of the evolution.

5. Conclusions

In this study, dealing with the evolution of a thickened lithosphere, it appears that the dynamical effects induced by the basaltic/eclogitic and depleted composition of the lithosphere cannot be neglected. On one hand, the stabilizing effects of chemical buoyancy are enhanced by the lithospheric thickening as the local quantity of basalt and depleted mantle increases. On the other hand, as crustal thickening favors the transition of basalt in the deeper eclogite domain, new crustal
instabilities are created, favoring root recycling. The competition between these light and dense components, added to the thermal effect controls recycling of lithospheric root. Our results show that under typical Venustian conditions, the root recycling may be twice as short for a thermochemical lithospheric root than for a purely thermal one, but there is less recycled material. This can be explained by the instability induced by the ongoing basalt/eglogite transformation and also by the stabilizing effect of the depleted lithosphere, which prevents recycling of a part of the root.

In addition to the control of lithospheric composition in root recycling, we have pointed out a possible scenario for Maxwell Montes to explain the succession of mountain formation in compressional state and magmatism. Because there is no water on Venus, both events would result from opposing dynamical features, namely mantle downwelling and upwelling. Such a transition may result from unsteady mantle dynamics. Our study shows that the magmatism may also be indirectly generated by the compressional events, considering both thermal and chemical aspects: when the depleted part of the recycled root reaches the base of the upper mantle, it becomes a diapir, which is deflected toward the mountain root and locally reheats it. In the case of a sufficiently hot Venusian mantle, this local reheating associated to the trapping of a partial melting zone between the mountain root and a new cold root created by eclogite instability may provide magmatism during some 20 Myr, about 100 Myr after the compressional event. This scenario may explain the formation and evolution of Venusian mountains like Maxwell Montes with a compressional phase followed by an extensional one, including volcanic events. Our model demonstrates that under Venustian conditions, lithosphere thickening could trigger late magmatism and extensional processes. We suggest that such a scenario could explain the relative chronology of compressional and extensional/magmatic events observed also on other planets.

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