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Constraints of the topography, gravity and volcanism on Venusian mantle dynamics and generation of plate tectonics

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ABSTRACT

Venus's mantle convection model was studied in a three-dimensional spherical shell domain with depth- and temperature-dependent viscosity. Numerical results show that key observations of Venus including the number of major "hotspot" volcanic systems, spectral patterns of the surface topography and geoid at long- and intermediate-wavelengths can be explained in models that have a spinel-to-post-spinel endothermic phase change of -3.5 MPa/K Clapeyron slope and averaged mantle viscosity of 2×10^{21} Pa s (i.e., convective Rayleigh number of 1.8×10^{7}). Our models with the endothermic phase change show relatively weak time-dependence, suggesting that the phase change may not be the primary cause for "catastrophic" resurfacing on Venus. Our calculations also show that Venus cannot have a weak asthenosphere that is similar to that on the Earth, in order to match the observations, thus supporting a key role of asthenosphere in producing plate tectonics.

1. Introduction

Despite of their similar sizes and compositions, Venus and the Earth show distinctly different surface tectonics and dynamic evolution. Venus is characterized by one-plate, stagnant-lid mantle convection, while the Earth is controlled by mobile-lid, plate tectonics type of convection (Kaula and Phillips, 1981; Nimmo and Mckenzie, 1998; Smrekar et al., 2007). What controls the style of mantle convection, i.e., stagnant-lid versus plate tectonic convection, remains one of the most important unresolved questions in geodynamics. While lithospheric deformation including faulting plays important roles (Moresi and Solomatov, 1998; O'Neill et al., 2007), recent studies suggest that weak asthenosphere also exerts a significant control on generation of plate tectonics (Höink et al., 2012; Richards et al., 2001).

Venus is a geologically active planet with significant young volcanism, as observed recently by the Venus Express spacecraft (Smrekar et al., 2010). These surface features of volcanism and tectonics, together with satellite observations of surface topography and gravity anomalies by the Pioneer Venus Orbiter and Megellan spacecraft (Konopliv et al., 1999; Konopliv and Sjogren, 1994; Rappaport et al., 1999; Sjogren et al., 1997), provide important constraints on the dynamics of Venus. The gravity and topography anomalies on Venus are highly correlated and show a relatively large ratio or admittance at lower degrees (Fig. 1), suggesting a dynamic origin for these anomalies (Pauer et al., 2006; Simons et al., 1997; Smrekar and Phillips, 1991). Mantle dynamic modeling of large topographic rises with volcanic features showed that their gravity and topography can be explained as a result of mantle upwelling plumes (Kiefer and Hager, 1991; Nimmo and McKenzie, 1996; Smrekar and Parmentier, 1996). Such features are known as "hotspot" on Earth, with Hawaii being the classic example. Nine "hotspots" have been identified based on observations of geologic features, gravity anomalies and topographic rises, and several of them are supposed to be with geologically recent volcanism based on recent data from Venus Express (Smrekar et al., 2010; Stofan et al., 1995). Therefore, these nine "hotspots" or mantle plumes represent the characteristic convective wavelength for Venus (Smrekar and Sotin, 2012).

However, little effort has been made to investigate the relationship between the spectra of the topography and gravity and mantle convective structure in global models of stagnant-lid mantle convection. Such studies are necessary for two reasons. First, the topography and gravity spectra are inherently related to mantle convective structure at intermediate- and long-wavelengths. While convective structure wavelength is often prescribed in regional models for individual plumes, only global models of mantle convection yield dynamically self-consistent convective structure (e.g., the number of plumes). Second, convective structure including its dominant wavelength is affected significantly by

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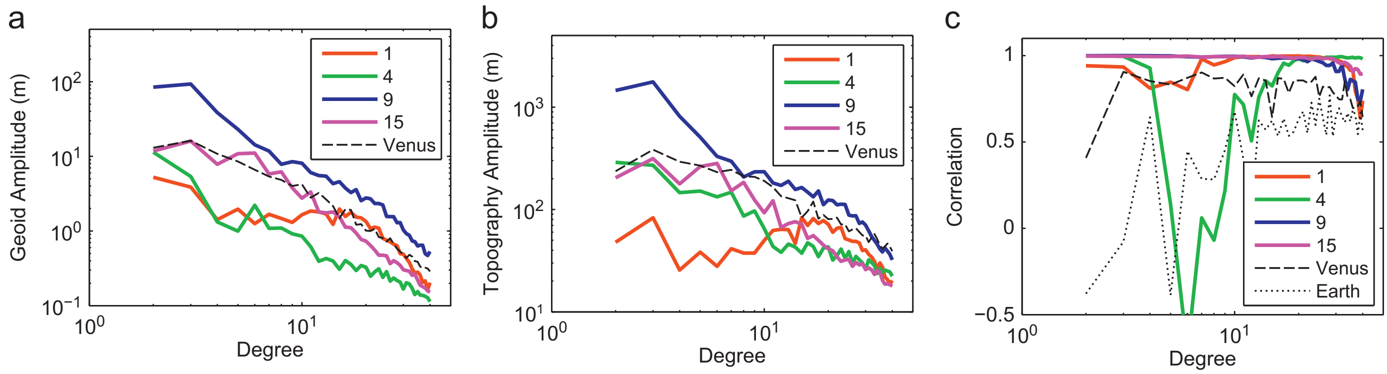

Fig. 1. Power spectra of geoid (a), topography (b), and correlation between geoid and topography (c) for Venus and Cases 1, 4, 9 and 15. Also included in (c) are results for the Earth.

mantle viscosity structure and mantle phase changes (Roberts and Zhong, 2006; Tackley, 1996). For example, stagnant-lid convection with relatively uniform mantle viscosity under the lid that is preferred by regional models of individual plumes (Kiefer and Hager, 1991), typically contains a large number of mantle plumes (Reese et al., 1999; Smrekar and Sotin, 2012) that may be inconsistent with the inferred nine mantle plumes for Venus. However, both endothermic spinel-to-post-spinel phase change and asthenosphere may increase the dominant convective wavelength and reduce the number of plumes (Roberts and Zhong, 2006).

We have formulated three-dimensional global models of mantle convection to simultaneously explain the number of plumes and the spectra of surface topography and gravity (i.e., geoid) for Venus. The models employ the extended-Boussinesq approximation and realistic temperature- and depth-dependent viscosity, similar to those used for Mars (Roberts and Zhong, 2006). In total, 15 cases are computed. By comparing with observations we seek constraints on mantle dynamics including the mantle viscosity structure, convective vigor, and phase changes.

2. Model setup

The Venus' mantle is treated as an infinite Prandtl number fluid in a three dimensional spherical shell under the extended Boussinesq approximation. The non-dimensional governing equations of mantle convection are (Zhong, 2006; Zhong et al., 2008):

$$\nabla\cdot\mathbf{u}=0 \tag{1}$$

$$-\nabla P+\nabla\cdot[\eta(\nabla\mathbf{u}+\nabla^T\mathbf{u})]+\left(\frac{R_0}{D}\right)^3\left[RaT-\sum_{k=1}^{2}(Ra_k\Gamma_k)\right]\mathbf{e}_r=0 \tag{2}$$

$$\begin{aligned}\left(\frac{\partial T}{\partial t}+\mathbf{u}\cdot\nabla T\right)\left[1+\sum_{k=1}^{2}\left(\gamma_k^2\frac{Ra_k}{Ra}\frac{d\Gamma_k}{d\pi_k}\right)D_i(T+T_s)\right]&=\nabla^2T\\ -\left(1+\sum_{k=1}^{2}\gamma_k\frac{Ra_k}{Ra}\frac{d\Gamma_k}{d\pi_k}\right)D_i(T+T_s)u_r&+\left(\frac{D}{R_0}\right)^3\frac{D_i}{Ra}\sigma_{ij}\frac{\partial u_i}{\partial x_j}+H\end{aligned} \tag{3}$$

where $\mathbf{u}$, P, η, T, T_s, D_i, u_r, σ_{ij} and H are the velocity vector, pressure, viscosity, temperature, surface temperature, dissipation number, radial velocity, deviatoric stress, and heat production rate, respectively. R_0 is the radius of Venus and D is the Venusian mantle thickness. Γ_k, γ_k, and π_k are phase function, Clapeyron slope, and excess pressure for phase k (k=1 and 2 for olivine–spinel and spinel–perovskite phase changes, respectively), respectively (Christensen and Yuen, 1985). $\mathbf{e}_r$ is the unit vector in radial direction. Ra and Ra_k are a Rayleigh number and phase change Rayleigh number for phase k (k=1 and 2), respectively. Characteristic scales for the above equations are: length R_0, time R_0^2/κ (κ is the thermal diffusivity), and temperature ΔT. Ra, Ra_k, D_i and H are defined as

$$Ra=\frac{\rho_0 g\alpha\Delta TD^3}{\kappa\eta_0} \tag{4}$$

$$Ra_k=\frac{\delta\rho_k}{\rho_0\alpha\Delta T}Ra \tag{5}$$

$$D_i=\alpha gR_0/C_p \tag{6}$$

$$H=\frac{QR_0^2}{C_p\rho_0\Delta T\kappa} \tag{7}$$

where ρ_0 and η_0 are the reference values for density and viscosity. α, g and C_p are coefficient of thermal expansion, gravitational acceleration and specific heat, respectively. $\delta\rho_k$ is the density change for phase change k (k=1 and 2) and Q is the volumetric internal heat generation rate.

The viscosity of the mantle is assumed to be temperature- and pressure-dependent (Karato and Jung, 2003). The non-dimensional viscosity in our model is given by

$$\eta=\eta_r\exp\left[\frac{E+V(1-r)}{T+T_s}-\frac{E+V(1-r_{core})}{1+T_s}\right] \tag{8}$$

where η_r is a pre-exponential factor, which is used to specify the viscosity contrast between the upper and lower mantle, and η_r is reduced for the upper mantle (*i.e.*, above the 690 km depth) to model the weak asthenosphere. r is the non-dimensional radial position, r_{core} is the Venusian non-dimensional core radius, and E and V are the non-dimensional values of activation energy, E^*, and activation volume, V^*, which are given by

$$E=\frac{E^*}{R\Delta T},V=\frac{\rho_0 gDV^*}{R\Delta T} \tag{9}$$

where R is the gas constant. The viscosity is cut off with a maximum non-dimensional value of 2×10^4 at the surface. Table 1 lists model parameter values.

Among the model parameters, we mainly consider three controlling parameters in our models, i.e., Rayleigh number, Ra, viscosity pre-factor, η_r, and Clapeyron slope of the phase changes, γ, with the goal to search and find these parameters that could reproduce the observations on Venus. With less constraint for Venus, most of the parameters used here are based on those for the Earth. The viscosity in the Earth's mantle is on average $\sim10^{21}-10^{22}$ Pa s with the lower mantle that may be a factor of 30 stronger than the upper mantle (e.g. Hager and Richards, 1989). Therefore, in our models here to test the influences of a weak upper mantle (or asthenosphere), the upper mantle viscosity is set to be 3–30 times weaker