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

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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 Solomonov, 1998; O'Neill et al., 2007), recent studies suggest that weak asthenosphere also exerts a significant control on generation of plate tectonics (Hönk 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 Magellan 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; Stefan 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
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-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 topography and gravity (i.e., geoid) for Venus. The models employ the extended-Bousinesq 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):

\[ \nabla \cdot \mathbf{u} = 0 \]  \hspace{1cm} (1)

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

\[ \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) \left[ 1 + \sum_{k=1}^{2} \left( \frac{\partial R_k d\Gamma_k}{Ra} \right) D_k(T + T_\gamma) \right] = \nabla^2 T \]  \hspace{1cm} (3)

where \( \mathbf{u} \), \( P \), \( \eta \), \( T \), \( T_\gamma \), \( D_k \), \( \epsilon_r \), \( \Gamma_k \), and \( \eta_k \) 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. \( \Gamma_k \), \( \gamma_k \), and \( \eta_k \) are phase function, Clapeyron slope, and excess pressure for phase \( k \) \((k = 1, 2, 2, 2)\) for olivine-spinel and spinel-perovskite phase changes, 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 change \( k \) \((k = 1, 2, 2, 2)\), respectively.

Characteristic scales for the above equations are: length \( R_0 \), time \( R_0^2/C_19 \), \( \kappa \) \((k \text{ is the thermal diffusivity})\), and temperature \( \Delta T \). \( Ra \), \( Ra_k \), \( D_i \) and \( H \) are defined as

\[ Ra = \frac{\rho g z z \Delta T \theta^3}{\kappa \eta_0} \] \hspace{1cm} (4)

\[ Ra_k = \frac{\delta \rho_k}{\rho g z z} Ra \] \hspace{1cm} (5)

\[ D_i = g z R_0 / C_p \] \hspace{1cm} (6)

\[ H = \frac{Q R_0^2}{C_p \eta_0 \Delta T \kappa} \] \hspace{1cm} (7)

where \( \rho_0 \) and \( \eta_0 \) are the reference values for density and viscosity. \( x \), \( g \), and \( C_p \) are coefficient of thermal expansion, gravitational acceleration and specific heat, respectively. \( \delta \rho_k \) is the density change for phase change \( k \) \((k = 1, 2, 2, 2)\) 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

\[ \eta = \eta_0 \exp \left[ \frac{E + V(1-r)}{T + T_\gamma} - \frac{E + V(1-r)_{\text{core}}}{1 + T_\gamma} \right] \] \hspace{1cm} (8)

where \( \eta \) is a pre-exponential factor, which is used to specify the viscosity contrast between the upper and lower mantle, and \( r \) 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, \( r_{\text{core}} \) 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

\[ E = \frac{E^*}{R \Delta T}, V = \frac{D_i g z R_0}{R \Delta T} \] \hspace{1cm} (9)

where \( R \) is the gas constant. The viscosity is cut off with a maximum non-dimensional value of \( 2 \times 10^8 \) 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, \( \eta_0 \), and Clapeyron slope of the phase changes, \( \gamma \), 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 \( \sim 10^{21} \) to \( 10^{22} \) 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.
convection (Smrekar and Sotin, 2012). Our calculations are the current Earth and also with a recent study on Venus’ mantle into 12 approximately equal size caps and each cap is further code CitcomS (Zhong et al., 2008). The spherical shell is divided performed with the 3-D spherical finite element and parallel Newtonian rheology to approximate a non-Newtonian rheology relatively small activation energy was used because we used a activation volume and did not vary them in our models. The fluxes are in a quasi-steady state. Table 2 lists all the 15 cases temperature for all cases. All cases are run until convective heat boundary layers. We run a test case to the resolution is sufficient to resolve the heat flux across the boundary layers such that there are at least three elements to surface and bottom heat flux (qs). For each case, we compute its averaged temperature- and pressure-dependent viscosity with a unit pre-exponential factor (i.e., no asthenosphere), Rayleigh number Rd=7.3×105, and no phase changes (Table 2). The horizontal-tempered and pressure-dependent viscosity and viscosity indicate ~240 km thick top thermal boundary layer or lithosphere with a viscosity that is five orders of magnitude larger than the mantle below (Fig. 2), characterizing a stagnant-lid convection. The pressure-dependent viscosity leads to six times of gradual increase in viscosity with depth from the base of lithosphere to the core–mantle boundary (CMB) region. The time-averaged internal heating ratio is 78% (Table 2), suggesting largely internally heated convection with small heat flux at the CMB. A representative convective thermal structure (Fig. 3a) shows dominant short-wavelengths with 78 upwelling plumes (Table 2), typical of stagnant-lid convection (Orth and Solomatov, 2011; Smrekar and Sotin, 2012; Zhong et al., 2008). While the surface topography and geoid are highly correlated (Figs. 1c and 3a), the powers of topography and geoid spectra are significantly reduced at long-wavelengths, compared with the observed (Fig. 1a and b).

Cases 2–4 differ from Case 1 in having an asthenosphere that is realized by reducing the pre-exponential factor of the viscosity equation for the upper mantle by a factor of 3, 10 and 30, respectively (Fig. 2b and Table 2). Because the asthenosphere leads to more vigorous convection and larger surface heat flux, internal heat generation rates are increased to maintain similarly large internal heating ratios (Table 2). Cases 2, 3 and 4 have on average 19, 12 and 10 plumes, respectively, compared with 78 plumes in Case 1 (Table 2), reflecting the effect of asthenosphere on promoting long-wavelength convective structures (Roberts and Zhong, 2006). The increased convective wavelengths for Cases 2–4 are also evident in the thermal structure, surface topography and geoid (Figs. 1 and 3b for Case 4 and Figs. 4 and 5a). However, the powers of topography and geoid for Cases 2–4
endothermic phase change depend on $Ra$ and Clapeyron slope, which is characteristic of Earth's mantle convection with a weak asthenosphere (Hager and Richards, 1989), but inconsistent with Venus. Since Case 2 has significantly more plumes than the observed, this leaves Case 3 with a factor of 10 viscosity reduction in the upper mantle as more viable. However, additional calculations of Cases 5 and 6 that differ from Case 3 in having different $Ra$s do show that the powers of topography and geoid for these cases are always significantly smaller than the observed (Fig. 4).

We now present models with olivine-to-spinel exothermic and spinel-to-post-spinel endothermic phase changes that should exist in Venus at depths of 490 km and 690 km, respectively (Schubert et al., 1997; Steinbach and Yuen, 1992). An endothermic phase change is known to promote long-wavelength convective structures (Tackley et al., 1993, 1994). Such effects of the endothermic phase change depend on $Ra$ and Clapeyron slope, and are stronger for larger $Ra$ and Clapeyron slope (Christensen and Yuen, 1985; Tackley et al., 1993, 1994). Cases 7, 8 and 9 include these phase changes with Clapeyron slopes $\gamma$ of $\pm 3$, $\pm 4$, and $\pm 5$ MPa/K, respectively, but these cases are otherwise identical to Case 1 (Table 2). That is, these cases also include a stagnant-lid and strongly temperature-dependent viscosity that were not included in a previous global mantle convection model with the phase changes for Venus (Schubert et al., 1997). Here we assume that the two phase changes have the same magnitude of Clapeyron slopes. However, the dynamics is mainly controlled by the endothermic phase change (Zhong and Gurnis, 1994). The dominant convective wavelength increases and the number of plumes decreases with the magnitude of Clapeyron slopes $|\gamma|$ (Figs. 5b and 3c for Cases 8 and 9, respectively). The number of plumes is 76, 24 and 3, as $|\gamma|$ increases from 3, 4 to 5 MPa/K for Cases 7, 8 and 9, respectively (Table 2). The powers of the topography and geoid spectra also increase with $|\gamma|$ (Fig. 6). Case 8 with $|\gamma|=4$ MPa/K provides the best fit to the observed spectra among these three cases (Fig. 6), but with 24 plumes this case significantly over-predicts the number of plumes.

Cases 10, 11 and 12 with the same $|\gamma|=3$ MPa/K as Case 7 but higher $Ra$ at $1.8 \times 10^7$, $2.7 \times 10^7$ and $3.6 \times 10^7$, respectively, explore the effects of $Ra$. Internal heat generation rate for Cases 10–12 is increased accordingly to maintain similar internal heating ratios (Table 2). Cases 10–12 show that for a given Clapeyron slope $|\gamma|$, increasing $Ra$ leads to increase in convective wavelengths. The number of plumes decreases to 15, 12 and 10 for Cases 10, 11 to 12, respectively (Table 2). Consequently, the

### Table 2
Model input parameters and outputs.

<table>
<thead>
<tr>
<th>Case no.</th>
<th>$Ra$ (C2)</th>
<th>$H$</th>
<th>$1/\eta$</th>
<th>$\gamma$ (MPa/K)</th>
<th>$\langle T \rangle$</th>
<th>$Q_0$</th>
<th>$f_\phi (%)$</th>
<th>$N_{plume}^a$</th>
<th>$V_{rms}$</th>
<th>$\delta$ (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$7.3 \times 10^6$</td>
<td>30</td>
<td>1</td>
<td>No</td>
<td>0.618</td>
<td>10.1</td>
<td>78.0</td>
<td>78.1 (±17.6)</td>
<td>698.5</td>
<td>245</td>
</tr>
<tr>
<td>2</td>
<td>$7.3 \times 10^6$</td>
<td>40</td>
<td>3</td>
<td>No</td>
<td>0.621</td>
<td>12.9</td>
<td>80.8</td>
<td>19.2 (±5.0)</td>
<td>1258</td>
<td>162</td>
</tr>
<tr>
<td>3</td>
<td>$7.3 \times 10^6$</td>
<td>60</td>
<td>10</td>
<td>No</td>
<td>0.613</td>
<td>17.3</td>
<td>82.0</td>
<td>11.7 (±2.5)</td>
<td>2231</td>
<td>134</td>
</tr>
<tr>
<td>4</td>
<td>$7.3 \times 10^6$</td>
<td>90</td>
<td>30</td>
<td>No</td>
<td>0.596</td>
<td>23.2</td>
<td>82.3</td>
<td>9.9 (±0.3)</td>
<td>3656</td>
<td>103</td>
</tr>
<tr>
<td>5</td>
<td>$3.7 \times 10^6$</td>
<td>45</td>
<td>10</td>
<td>No</td>
<td>0.607</td>
<td>14.0</td>
<td>81.1</td>
<td>14.6 (±3.3)</td>
<td>1290</td>
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</tr>
<tr>
<td>6</td>
<td>$1.5 \times 10^7$</td>
<td>75</td>
<td>10</td>
<td>No</td>
<td>0.607</td>
<td>20.9</td>
<td>80.0</td>
<td>9.1 (±2.3)</td>
<td>3599</td>
<td>109</td>
</tr>
<tr>
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<td>$7.3 \times 10^6$</td>
<td>30</td>
<td>1</td>
<td>$\pm 3$</td>
<td>0.617</td>
<td>10.5</td>
<td>73.2</td>
<td>76.2 (±4.4)</td>
<td>702.5</td>
<td>226</td>
</tr>
<tr>
<td>8</td>
<td>$7.3 \times 10^6$</td>
<td>30</td>
<td>1</td>
<td>$\pm 4$</td>
<td>0.637</td>
<td>10.6</td>
<td>72.4</td>
<td>24.4 (±4.7)</td>
<td>718.6</td>
<td>245</td>
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<tr>
<td>9</td>
<td>$7.3 \times 10^6$</td>
<td>30</td>
<td>1</td>
<td>$\pm 5$</td>
<td>0.646</td>
<td>9.72</td>
<td>79.5</td>
<td>3.1 (±0.3)</td>
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<td>1</td>
<td>$\pm 3$</td>
<td>0.636</td>
<td>13.1</td>
<td>75.5</td>
<td>14.6 (±3.8)</td>
<td>1353</td>
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</tr>
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<td>$2.7 \times 10^7$</td>
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<td>1</td>
<td>$\pm 3$</td>
<td>0.637</td>
<td>14.6</td>
<td>75.2</td>
<td>11.7 (±3.0)</td>
<td>1777</td>
<td>176</td>
</tr>
<tr>
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<td>45</td>
<td>1</td>
<td>$\pm 3$</td>
<td>0.629</td>
<td>15.1</td>
<td>72.0</td>
<td>9.8 (±3.9)</td>
<td>2098</td>
<td>170</td>
</tr>
<tr>
<td>13</td>
<td>$1.8 \times 10^7$</td>
<td>40</td>
<td>1</td>
<td>$\pm 4$</td>
<td>0.648</td>
<td>12.7</td>
<td>74.4</td>
<td>3.0 (±1.1)</td>
<td>1259</td>
<td>212</td>
</tr>
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<td>14</td>
<td>$3.6 \times 10^7$</td>
<td>45</td>
<td>1</td>
<td>$\pm 4$</td>
<td>0.634</td>
<td>15.2</td>
<td>69.9</td>
<td>3.3 (±0.5)</td>
<td>1886</td>
<td>203</td>
</tr>
<tr>
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<td>40</td>
<td>1</td>
<td>$\pm 3.5$</td>
<td>0.643</td>
<td>13.1</td>
<td>75.4</td>
<td>7.8 (±3.4)</td>
<td>1356</td>
<td>200</td>
</tr>
</tbody>
</table>

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$a$ The numbers in parentheses are the standard deviations.

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![Fig. 2](image-url)

**Fig. 2.** Radial dependence of horizontally averaged non-dimensional temperature (a) and viscosity (b) for Cases 1, 4, 9, and 15. In (a), the gray lines represent melting curves for dry peridotite (dashed line) and wet peridotite (dash-dotted line), and the magenta dashed curve and the dotted line are for the maximum temperature at different depths (i.e., within the plumes) and the bottom of thermal lithosphere in Case 15, respectively. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
power spectra of the topography and geoid increase at relatively long wavelengths as $Ra$ increases (Fig. 7). The results from these six phase change calculations (Cases 7–12) are further confirmed by Cases 13–15 with varying $Ra$ and Clapeyron slopes (Table 2). In particular, Case 15 with Clapeyron slopes of $\pm 3.5$ MPa/K and $Ra=1.8 \times 10^7$ (i.e., averaged mantle viscosity of $2 \times 10^{21}$ Pa s, if
scaled using parameters in Table 1) reproduces the number of plumes and the topography and geoid spectra the best (Figs. 1 and 3 and Table 2). All the cases with phase changes show positively correlated topography and geoid, consistent with the observed (Figs. 1, 6 and 7).

Our results show that phase changes may affect convective wavelengths and hence the number of plumes significantly (e.g. Tackley et al., 1993). It should be pointed out that phase changes may lead to different numbers of plumes in the upper and lower mantles. For example, Fig. 8 shows a zoom-in view of plume structures for case 15. Under each large plume in the upper mantle, there are several small separate plumes in the lower mantle.

4. Discussions

Our models represent the first three-dimensional spherical mantle convection calculations for Venus with realistic temperature- and pressure-dependent viscosity and phase changes, although previous studies considered separately either temperature-dependent viscosity (Orth and Solomatov, 2011) or phase changes (Schubert et al., 1997). We would like to make three remarks relevant to previous mantle convection modeling studies for Venus. (1) The episodic major mantle overturn or mantle “avalanche” associated with the endothermic phase change was suggested to cause the resurfacing of Venus, based on two-dimension models (Steinbach and Yuen, 1992). However, our models with the endothermic phase change show
relatively weak time-dependence, due to the combined effects of temperature-dependent viscosity and three-dimensional geometry in our models (Schubert et al., 1997; Zhong and Gurnis, 1994). (2) Stagnant-lid convection can have a variety of different dominant convective wavelengths, ranging from the traditional short-wavelength mushroom-type convection for relatively uniform mantle viscosity (Case 1) to long-wavelength convection when endothermic phase change or weak asthenosphere is present (Fig. 3). (3) Schubert et al. (1997) studied the effects of phase changes on the topography and geoid spectra in global models of mantle convection, but unlike our models their models ignored temperature-dependent viscosity and stagnant-lid convection which may affect convective structure significantly.

The absence of plate tectonics on Venus, a planet with a similar size and composition to the Earth, has prompted a number of proposals on controls for generation of plate tectonics (Kaula and Phillips, 1981; Nimmo and Mckenzie, 1998), and most proposals are related to lithospheric deformation (Landuyt and Bercovici, 2009; Lenardic and Kaula, 1994; Moresi and Solomatov, 1998). Recent studies suggest that weak asthenosphere may play an essential role in generating plate tectonics by increasing lithospheric stress to promote localized lithospheric deformation and by organizing long-wavelength convection (Höink et al., 2012). This proposal is corroborated by our findings and previous studies (Kiefer and Hager, 1991) that reject an Earth-like asthenosphere for Venus. This further raises the question on what causes asthenosphere and the differences in tectonics and climate between Venus and Earth. Water may play an essential role in forming asthenosphere on the Earth (Hirth and Kohlstedt, 1996), suggesting that Venus may be devoid of water in the mantle (Kiefer and Hager, 1991; Nimmo and Mckenzie, 1998) and that volatiles and water play an important role in controlling planetary evolution and plate tectonics (Smrekar et al., 2007).

Given that recent studies show evidence of active volcanism on the Venusian surface (Smrekar et al., 2010), it is interesting to see if our models could predict the pressure release melting. The melting curves for dry and wet peridotite are shown in Fig. 2a. The equations for the melting curves are (Basaltic Volcanism Study Project, 1981; Smrekar and Sotin, 2012)

\[
T(K) = 1350 + 0.1P(MPa) \text{ for dry peridotite}
\]

\[
T(K) = 1240 + 49.8(\rho(GPa) + 0.323) \text{ for } P < 2.4 \text{ GPa and wet peridotite}
\]

\[
T(K) = 1266 - 11.8(\rho(GPa) + 3.5P^2) \text{ for } P > 2.4 \text{ GPa and wet peridotite}
\]

The averaged temperatures of our models are always smaller than the dry solidus, but are higher than the wet solidus under the lithosphere (Fig. 2a). However, the maximum mantle temperatures from mantle plumes are very close to that of the dry solidus (e.g., Fig. 2a for Case 15). This suggests that our models may explain the recent “hotspot” volcanism but not wide spread, large-scale volcanism (Smrekar et al., 2010), if the Venus’ mantle is devoid of water and volatiles.

5. Conclusions

Compared with the observations, our model calculations lead to three conclusions. First, our models, as the first attempt for global convection calculations for Venus with realistic mantle viscosity and phase changes, reproduce all the key observations for Venus including the number of plumes and the spectra of topography and geoid (Fig. 1 and Table 2). The model with Clapeyron slopes \(\gamma = 3.5 \text{ MPa/K} \) and averaged mantle viscosity of \(2 \times 10^{21} \text{ Pa s} \) (i.e., \(Ra = 1.8 \times 10^7\) (Case 15) provides the best fit to the observations, although other models with \(Ra\) ranging from \(7.3 \times 10^6\) to \(3.6 \times 10^7\) and \(\gamma\) between 3 and 4 MPa/K (Table 2 and Figs. 6 and 7) may also provide a reasonable fit. All these cases have lithospheric thickness ranging from \(~170\text{ km} \sim 240\text{ km}\) (Table 2), consistent with previous studies (Moore and Schubert, 1995). It should be pointed out that while our preferred model of Case 15 reproduces well the geoid spectra, its topography power remains slightly smaller than the observed. This is expected and reasonable, considering that our models ignore the crust and crustal compensation process that produces the topography but negligible geoid anomalies at intermediate- and long-wavelengths for Venus (Smrekar and Phillips, 1991).

Second, the Venusian mantle may not have a weak asthenosphere as that of the Earth’s mantle, because such an asthenosphere leads to negative topography–geoid correlations that are inconsistent with the observed (Case 4 in Fig. 1). This conclusion on mantle viscosity is consistent with previous studies of the
topography and geoid from regional mantle plume models for individual plume features (Kiefer and Hager, 1991; Smrekar and Phillips, 1991). Although the topography–geoid correlations are significantly smaller than the observed (Fig. 4). Furthermore, our results show that models with the relatively uniform mantle viscosity under lithosphere (i.e., Cases 1 and 2) as suggested by regional models of individual plumes (Kiefer and Hager, 1991; Smrekar and Phillips, 1991) tend to lead to too many mantle plumes that are inconsistent with the observed. Therefore, our results suggest that in order to explain the observed topography and geoid spectra and the number of mantle plumes, the endothermic phase change must play an important role in Venusian mantle dynamics.

Finally, our models with the endothermic phase change show relatively weak time-dependence in heat transfer across the mantle, significantly different from the 2-D convection models. This suggests that the endothermic phase change may only play a limited role in causing the 90-250 km. Our preferred model also predicts that partial melting could occur currently in upwelling plumes for a dry Venusian mantle, thus explaining the recently observed active “hotspot” volcanism on Venus.

Acknowledgments

The authors are thankful for the careful and constructive reviews by Drs. W. Kiefer and S. Smrekar and financial support by NSFC (91014005, 40774045), US-NSF (EAR-1015669 and 1135382), the Knowledge Innovation Program of the Chinese Academy of Sciences to JH and the CAS/SAFEA International Partnership Program for Creative Research Teams.

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