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Studying lower crustal flow beneath Mead basin: Implications for the thermal history and rheology of Venus

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ABSTRACT

Mead, the largest crater on Venus, has low topographic relief at the surface and at the crust-mantle boundary. Due to high surface temperatures, viscous deformation could play an important role in crustal structure. Using the finite element method, we simulate the long-term viscoelastic deformation of Mead crater and investigate the role of lower crustal flow in the evolution of the surface and subsurface topography. We examine the thermal states that allow this evolution to occur and determine the background heat flux. Our study constrains the background heat flux in the vicinity of Mead basin to 55–90 mW m^{-2}. This surface heat flow is generally higher than the average Venusian global values suggested by recent thermal models. In addition by applying hydrous and anhydrous creep rheological parameters, we demonstrate that the Venus's interior is rheologically dry and that the crust near Mead could be relatively high in plagioclase.

1. Introduction

Studying the deformation of crusts and lithospheres leads to a better understanding of the internal structure and thermal history of planetary bodies. Previous studies on the thermal evolution of Mars and the Moon have investigated the viscoelastic relaxation of large craters primarily driven by lower crustal flow, and have constrained the background heat flux of these planetary bodies (e.g., Mohit and Phillips, 2006, 2007; Karimi et al., 2016). Unlike the Moon and Mars, the surface of Venus has a relatively sparse crater record. There are about 1000 craters on the surface of Venus, among which Mead, with the size of 270 km in diameter, is the largest (e.g., Herrick and Sharpton, 1996; Hauck et al., 1998) and the only one to be resolved in gravity models, a fact likely to not be rectified by future orbiting platforms because of Venus's thick atmosphere. For example, the proposed VERITAS mission would generate a globally more uniform gravity field with an improved accuracy of 3 mGal, but with a spatial resolution of 145 km (Smrekar et al., 2016), larger than the size of all other craters on Venus.

Mead basin is of particular interest due to it being isostatically undercompensated (Banerdt et al., 1994), perhaps indicating a geologic process within the lithosphere and upper mantle. Using the MGNP60FSAAP gravity model, Banerdt et al. (1994) demonstrated that unlike large lunar basins with uplifted mantle underneath the crater depression (e.g., Neumann et al., 1996), there is not prominent mantle uplift beneath Mead basin. One viable explanation for the lack of definitive topographic relief at the crust-mantle boundary could be viscoelastic processes, notably lower crustal flow. The very high surface temperatures of Venus combined with a relatively thick crust lead to very high temperatures at the base of the crust and consequently lower viscosities. A lateral pressure gradient, generated by crustal thickness variations, then can induce lateral movement of lower crustal material. A large amount of lower crustal flow is likely for a large impact crater on Venus, thus reducing the topography on the crust-mantle boundary (e.g., Grimm and Solomon, 1988; Karimi et al., 2016).

In this study, we investigate the possible viscoelastic deformation of Mead basin primarily within the subsurface, and determine the role of lower crustal flow in the relaxation of mantle topography. Using the Finite Element Method (FEM), we simulate the crustal deformation and use it as a probe of the heat flux of Venus. Furthermore, by testing various viscous creep parameters, we aim to determine the appropriate viscous rheology for the crust and mantle of Venus.

2. Mead basin

Mead, the largest exposed crater on the surface of Venus at 12.5° N 57.2° E, was first revealed by the Magellan mission (Herrick and Sharpton, 1996). The age of Mead likely falls within the estimated average surface age of Venus, ranging from 300 to 750 Myr (McKinnon et al., 1997). This crater is a shallow basin with a

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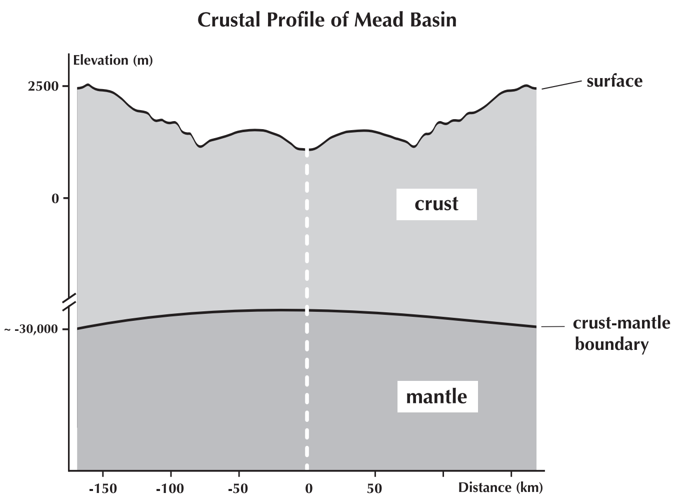

Fig. 1. A schematic that shows the crustal profile of Mead basin. Azimuthally average surface topography is plotted according to Herrick and Sharpton (1996). The mantle topography is notional and not based on crustal thickness models.

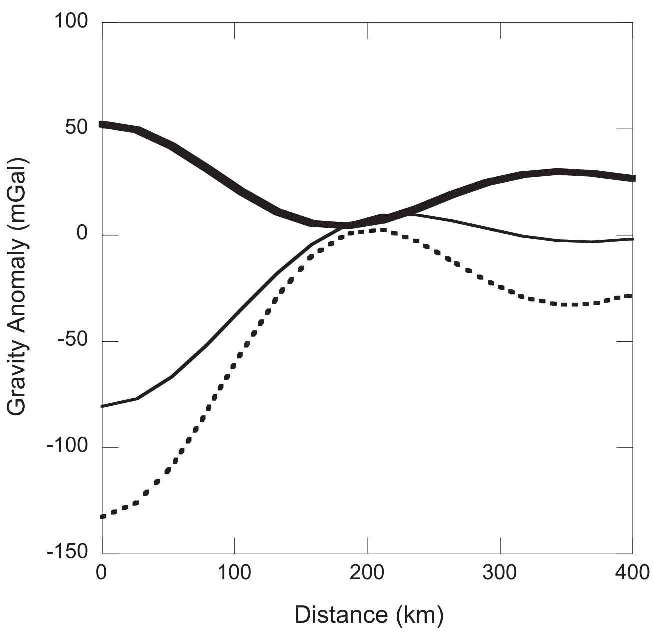

Fig. 2. The azimuthally averaged free-air (thin solid), topographic (dotted), and Bouguer (thick solid) gravity anomalies over Mead basin, from a degree-and-order 15–150 spherical harmonic expansion.

total (rim to floor) depth of about 1100 m; in contrast, a fresh crater of this diameter should have a minimum depth of 1350 m (Herrick and Sharpton, 1996; McKinnon et al., 1997). The central part of the basin (Fig. 1) has topographic variations and is not flat (Herrick and Sharpton, 1996).

We plot the azimuthally averaged free-air gravity signal from Mead in Fig. 2, created from an expansion of the MGNP180U 180-degree spherical harmonic potential model of Venus (archived in NASA's Planetary Data System [http://pds.nasa.gov]). This gravity model uses Magellan mission data and does not possess a uniform spatial resolution across the surface. Mead, however, is located in an area with relatively high resolution (Konopliv et al. 1999). We omit degrees below 15 from the expansion, which detrends regional signals with wavelengths of ∼2500 km and longer. We also truncate the expansion at degree 150, in order to avoid possible errors occurring at higher degrees (Wieczorek, 2007). An expansion to degree 150 has a spatial resolution of ∼125 km. Consequently, Mead, with a diameter of 270 km, is the only crater on the surface of Venus that can be resolved (though just barely) in existing gravity models.

Previous studies have determined the current topographic structure of Mead basin (e.g., Herrick and Sharpton, 1996). Using a spherical harmonic model of the topography (also archived in NASA's Planetary Data System) expanded like the gravity and assuming a crustal density of 2900 kg m^{-3}, the predicted free air gravity anomaly due to the surface topography alone would be ∼− 130 mGal (Fig. 2), which reflects the mass deficiency of the excavated crater (Fig. 2). Thus similar to Banerdt et al. (1994), we find that the gravity anomaly is dominated heavily by the surface topography, and the contribution of the subsurface topography to the gravity signal is small. Assuming a density contrast across the crust-mantle boundary of 400 kg m^{-3} (Dombard et al., 2007; James et al., 2013), this Bouguer central high translates to a maximum mantle uplift of ∼3 km (and possibly less, depending on the average value of the background terrain).

The mantle uplift beneath Mead was undoubtedly much higher initially. Studies of lunar and Martian crustal thickness show that the mantle is uplifted beneath large impact craters (e.g., Neumann et al., 1996; Neumann et al., 2004; Neumann et al., 2008). This phenomenon is likely due to the collapse of the transient crater, which is narrower and deeper than the final crater (e.g., Melosh, 1989; Wieczorek and Phillips, 1999). The inward and upward collapse of the transient crater leads to widening and shallowing of the crater at the surface and to uplifting the crus-mantle boundary. Namiki et al. (2009) suggested a nearly isostatically compensated structure for large craters at the central point, while Melosh et al. (2013) suggested this uplift could even be slightly superisostatic. Applying the same concept to Mead with an expected initial depth of 1350 m, the initial mantle uplift should have been up to 10 km.

Clearly, the shape of Mead has evolved, and the very high surface temperatures of Venus (740 K) implicate viscous processes, specifically lower crustal flow. Like for many large craters on Mars (Karimi et al., 2016), high temperatures in the lower crust and underlying mantle, and thus low viscosities, result in relaxation of the mantle topography. This viscous response also serves to decouple mechanically the surface topography from the compensating mantle topography. The surface basin, in turn, elastically flexes upwards, the remaining topography having virtually lost any isostatic support and now having to be supported by the strength of a lithosphere thinner than the crust. The mechanics just described are more complex than has been previously modeled for Venus (e.g., Grimm and Solomon, 1988); consequently, we re-examine the potential for crater relaxation on Venus by looking at the case of Mead basin here. In particular, we explore the thermal states and rheological conditions that allow the relaxation of the subsurface topography to the current state.

3. Methodology

3.1. Initial shape of the basin

We use the same methodology as described in Karimi et al. (2016), simulating 1 radial plane beneath an axisymmetric crater with the bottom and side boundaries sufficiently far to not affect the solutions. In order to model the viscoelastic deformation of Mead basin, the shape of the fresh crater is required. Since we are not certain about the initial depth of Mead crater at the time of formation, we consider two cases of (1) shallow and (2) deep structures that bracket the likely initial depth. This approach permits constraint of the upper and lower limits of the background heat flux. For the shallow structure, we use the depth-diameter curve of McKinnon et al. (1997), which indicates that the initial depth of Mead basin at the time of formation was ∼1350 m. For

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