IMPACT CRATER REBOUND AND LITHOSPHERIC STRUCTURE ON VENUS

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Introduction. This study takes advantage of the relative prevalence and uniform distribution of impact craters on Venus [1,2] to attempt to place constraints on crustal thickness (H) and temperature gradient (dT/dz) using models of viscous relaxation and elastic flexure. The lithosphere responds to the instantaneously imposed load of a crater by isostatically rebounding: first by rapid flexural bending, and, over time, by viscous relaxation of relief. Earlier work using such models estimated upper bounds of $H \leq 20$ km and $dT/dz \leq 20$ K/km [3]. Magellan observations show that craters are unexpectedly deep [1,4], suggesting that little modification has occurred. Also, recent gravity modeling favors a thicker crust [5]. We therefore reevaluate the earlier work [3] by representing crustal viscosities using an improved flow law for anhydrous diabase [6], and drawing from the larger sample of craters available in Magellan altimetry. We assume that all craters are initially uncompensated based on the limit of 30% compensation inferred from Mead's gravity anomaly [7].

Table I lists data on ten of the largest Venus craters. The outermost ring diameter (D) was measured from radar images. Gridded topography was azimuthally averaged about the crater center to arrive at depths from rim to floor center. Meitner and Klenova are located in Magellan altimetry data gaps; the Klenova depth is from Venera 15/16 measurements [3]. The presence of floor fractures (ff) implies rebound [8]; their absence, however, might be explained by later volcanic burial. Here we neglect any diameter dependence on depth, and calculate a mean and standard deviation of eight depths, excluding Cleopatra, of 1.05±0.40 km. We choose an initial, synthetic crater 100 km in diameter and 1.45 km deep (one standard deviation greater than the mean observed depth), and seek models that produce shallowing to one standard deviation less than the mean depth, or 0.65 km.

Table I. Properties of large Venus craters.

name	latitude	longitude	D / km	depth / km	ff
Mead	12.52°N	57.16°E	270	1.07±0.16	✓
Isabella	29.79°S	204.23°E	175	1.41±0.22	✓
Meitner	55.61°S	321.61°E	150	—	✓
Klenova	78.18°N	104.32°E	140	~0.78	✓
Stanton	23.30°S	199.26°E	105	1.02±0.10	
—	62.55°N	40.21°E	105	1.85±1.00	
Cochran	51.84°N	143.34°E	100	0.86±0.08	
Bonheur	9.71°N	288.78°E	100	0.62±0.15	
Cleopatra	65.92°N	6.94°E	90	~2.5	
Joliet-Curie	1.64°S	62.42°E	85	0.79±0.52	

Viscous relaxation. We employ a viscous relaxation model which allows an arbitrary viscosity profile with depth [3]. Effective viscosities are parameterized from laboratory flow laws for dry Columbia diabase [6] and dunite [9] as $\eta = \Delta\sigma/3\dot{\varepsilon}$ where $\Delta\sigma$ is differential stress and $\dot{\varepsilon}$ is principal strain rate. A "brittle–ductile" transition depth is chosen from yield stress envelopes; above this depth the viscosity is linearly decreased to the surface. We cast the solution in a manner that places upper bounds on H and dT/dz: (1) Viscosity is maximized by choosing a low characteristic stress, $\Delta\sigma_c = 3$ MPa. (Utilizing a constant strain rate results in comparatively lower viscosities.) A minimum viscosity of 10^{20} Pa s is enforced, but we apply no upper bound on viscosity. (2) Elapsed relaxation time is minimized at 50 Ma; the oldest, large craters are probably 100s of millions of years old. (3) We choose the starting and ending crater depths to maximize the viscous relaxation. (4) Initial compensation state has a strong effect on relaxation, and should be maximized to promote upper limits on H and dT/dz. We, however, assume zero initial compensation.

Due to the substantially greater strength of dry diabase compared to the flow law used earlier [3], the crustal thickness is *unbounded* by the viscous relaxation approach. Topographic relief is reduced by no more than 5% for $H \leq 150$ km and $dT/dz \leq 35$ K/km, even after 500 Ma. As others have found [10], H has little control on relaxation since dry diabase and dunite viscosities are nearly identical at low $\Delta\sigma_c$. With so little relaxation, dT/dz is also essentially unbounded. These results have two interpretations: there has been virtually no viscous relaxation, or we have excessively overestimated viscosity. The rims of the largest craters appear subdued, and there has apparently been rebound at

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Mead since the volcanic flooding of the floor [11], so we suspect that viscous relaxation has occurred. Choosing a larger $\Delta\sigma_c$, in particular one that more realistically varies with depth, would produce significant relaxation.

Flexural rebound. Our alternative model of crater modification is flexural bending caused by the unloading of the crater excavation [3]. We use an elastic, axisymmetric model [12] that treats the lithosphere as a single plate, and does not account for separate mechanical behavior of the crust and mantle. Free parameters of interest include the elastic plate thickness (T_e), Young's modulus (E), and density (ρ). To express T_e in terms of dT/dz we apply the principle that the maximum elastic bending moment and topographic curvature lie on the corresponding inelastic moment–curvature curve [13,14]. To place upper bounds on dT/dz we assume: (1) The lithosphere consists entirely of mantle of $\rho = 3300$ kg m^{-3}. (2) The Young's modulus is large, $E = 10^{11}$ Pa. (3) The strain rate is high, $\dot{\varepsilon} = 10^{-15}$ s^{-1}. (4) The crater size is at the lower end of those under consideration (Table I), $D = 100$ km (Fig. 1). (5) The initial cavity is shallow, 1.45 km rim-to-floor (Fig. 1, solid line). (6) The floor center rebounds to 0.65 km. This approach assumes that the observed variation in floor depths reflects different degrees of impact excavation, lava infilling, contributions from gravitational relaxation, and/or variations in local lithospheric strength.

Our model places a lower bound on the elastic plate thickness of $T_e = 7.9$ km, and a corresponding upper bound on temperature gradient of $dT/dz = 26$ K/km. Figure 1 shows a fit of the rebounded synthetic crater (dashed line) to the azimuthally averaged Bonheur topography (dotted line). Studies of flexural topography at the margins of large coronae [15,16] and gravity anomalies over large volcanic edifices [17] have indicated a much thicker lithosphere and lower thermal gradient. The presence of crater floor fractures also has been used to infer a thick plate, $T_e \approx 50$–70 km [8]. Our results are inclusive of lower thermal gradients, and can be made consistent with previously reported values by relaxing any of the above conservative conditions. The mechanical contribution of the weaker crust, a slower strain rate [18], and a reduced extent of flexural rebound would all serve to decrease dT/dz.

Conclusions. Using a conservative approach, we have applied models of viscous relaxation and elastic flexural rebound of large impact craters to place upper limits on the lithospheric thermal gradient. Viscous relaxation is currently unable to constrain crustal thickness or temperature gradient because of the stiffness of anhydrous diabase and the large crater depths [cf. 3]. Current work is directed towards estimating a more reasonable depth profile of characteristic stress by which to evaluate viscosity; resulting models should be admissive of more significant relaxation. Flexural rebound limits temperature gradients to ≤26 K/km, a value that might be reduced with independent constraints on other free parameters.

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