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GEOPHYSICAL MODELS OF WESTERN APHRODITE-NIOBE REGION: VENUS, K.I. Marchenkov, Institute of Physics of the Earth, Moscow, Russia, R.S. Saunders, and W.B. Banerdt, Jet Propulsion Laboratory, Calif. Institute of Technology, Pasadena, CA 91109

In terms of its mechanical parameters (such as mass, mean radius, and mean density) Venus is very similar to the Earth. But the tectonic regimes of Venus and Earth are quite different. Unlike the Earth, where the effects of mantle convection are manifest on the surface in mid-oceanic ridges (spreading zones) and trenches (subduction zones), a global system of ridges rifts and subduction zones is not observed on Venus. There is no evidence of Earth-like plate tectonics and Venus may be a one-plate (or no-plate) planet. Magellan radar data confirm this point of view. One possible consequence of this conclusion is that Venus's convection is confined beneath a thick, buoyant lithosphere. In thermal models of Venus [1] a thick Venusian crust overlies a convecting mantle. The style of convection in Venus is an open issue (e.g. two layer or whole mantle convection. But two layer convection provides high temperature at the crust-mantle boundary. Additionally, the geochemical data show that the upper mantle of Venus is apparently depleted [1].

This could explain why convection is only partly revealed in the long-wave part of the gravitational field. It is reasonable to suppose that at least the long-wavelength part of gravitational field of Venus, for spherical harmonics greater than two (which correlate with corresponding harmonics of the relief), is due to crustal thickness variations [2] or possibly also to variations of crustal density (for example, hot mantle plumes, intruding into the base of the crust).

The new topography and gravitational field data for Venus expressed in spherical harmonics of degree and order up to 50 allow us to analyze the crust-mantle boundary relief and stress state of the Venusian lithosphere. In these models, we consider models in which convection is confined beneath a thick, buoyant lithosphere. We divide the convection regime into an upper mantle and lower mantle component. The lateral scales are smaller than on the Earth. In these models, relative to Earth, convection is reflected in higher order terms of the gravitational field. On Venus geoid height and topography are highly correlated, although the topography appears to be largely compensated. We hypothesize that Venus topography for those wavelengths that correlate well with the geoid is partly compensated at the crust-mantle boundary, while for the others compensation may be distributed over the whole mantle. In turn the strong sensitivity of the stresses to parameters of the models of the external layers of Venus together with geological mapping allow us to begin investigations of the tectonics and geodynamics of the planet. For stress calculations we use a new technique of space-and time-dependent Green's response functions using Venus models with rheologically stratified lithosphere and mantle and a ductile lower crust. In the basic model of Venus the mean crust is 50-70 km thick, the density contrast across the crust-mantle boundary is in the range from 0.3 to 0.4 g/cm³. The thickness of a weak mantle zone may be from 350 to 1000 km. Strong sensitivity of calculated stress to various parameters of the layered model of Venus together with geological mapping and analysis of surface tectonic patterns allow us to investigate the tectonics and geodynamics of the planet. The results are presented in the form of maps of compression-extension and maximum shear stresses in the lithosphere and maps of crust-mantle boundary relief, which can be presented as a function of time.

The technique of using Green's response functions for the distribution of stresses and

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deformation in one-plate planets with elastic mantles including a low viscosity zone was developed earlier [1,2]. In our present Venus model the rheology of the interior layers is more complicated but possibly more realistic. The main features of the rheology are: 1) In the mantle, dislocation climb is the dominant process, although diffusion creep is also included; 2) the rheology of the lithosphere includes a so-called unstable beta-creep and stable gamma-creep. The basic, parametrically simple model of Venus, is constructed using a new geochemical model of Dreibus and Wanke [3], the mean crustal thickness is varied from 50 to 70 km, and the density contrast across the crust-mantle boundary is in the range 0.3 to 0.6 g/cm³.

It is instructive to calculate the time-history of stresses and deformation in the interior during the formation of major topographic features, Beta Regio, Ishtar Terra, Aphrodite Terra and others. First we model the rheology then find a corresponding transformation of our equations to Laplace transform space. Thus we will have in Laplace space the equivalent elastic problem, which can be solved. The next step is to find the inverse transformation of these solutions back to the time domain. This step is the most difficult. The rheology of the lithosphere and mantle is modeled by a generalized Bingham-Maxwell Body Law, including the brittle-elastic-ductile transition in the crust.

To perform the calculations we need to know the effective viscosity variations with depth and with time, as well as gravity and topography at high resolution.

With this approach it is possible to simultaneously satisfy the observed topography and gravity with spherical harmonic models using reasonable strength and viscosity parameters for the Venusian crust and mantle. We have modeled the region of Western Aphrodite and the Niobe plains to get reasonable depths of compensation. These results are based on Magellan topography and Pioneer Venus gravity fit to spherical harmonic models of order and degree 50. Continuing work uses the higher resolution Magellan data as they become available, both as local spherical harmonic models and the highest resolution line-of-sight gravity data. Crust mantle boundary relief is calculated for Western Aphrodite - Niobe relative to a mean crustal thickness of 50 km. The calculations include the consequences of simple crust models and more complicated models with a weak, ductile lower crust, a strong upper mantle and a weak lower mantle layer. We use a mean crustal thickness of 50 km, of which the upper 20 km is elastic. The mantle between 50 km and 200 km is strong, and the mantle at depths between 200 km and 481 km is weak and acts like an asthenosphere. The calculated crust-mantle boundary relief is similar for simple models and the more realistic layered models, but the stress distributions are markedly different. This can be explained by the role of lithospheric bending in the more complicated model.

References: [1] Solomotov and Zharkov, *Icarus*, 1990.; [2] Marchenkov, K.I. et al.: 1990. *Earth, Moon, and Planets*. V.50/51. P.81.; [3] Dreibus, G. and Wanke, H., *Adv. Space Res.* V. 10, No. 3-4, p (3) 7.

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