GLOBAL DYNAMIC STRESS MODELLING ON VENUS; W. B. Banerdt, Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109

Patterns of tectonic deformation on planetary surfaces provide direct information on the lithospheric state of stress. This information can be compared with theoretical calculations of the global stress field in order to provide insight into the general structure of the lithosphere and (sub-lithospheric) mantle and into the mechanisms of tectonic deformation and topographic support. I have previously published global thin shell stress calculations for Venus's lithosphere in which a static model was assumed [1]. In the static case support for long-wavelength topography and gravity anomalies is provided by a combination of Pratt compensation in the mantle and Airy compensation in the crust. Those results showed that single-layer compensation cannot produce stress fields consistent with observed tectonic features, whereas two-layer compensation models, in which the topography is either isostatic or is due entirely to vertical displacement of the lithosphere, predict stress patterns which agree quite well with stress directions inferred from rifting in the equatorial highlands.

However two major problems exist with this type of model. Although the computed stress directions appear consistent with observed surface deformation, the magnitude of these stresses is too small. Maximum model stress differences are about 40 MPa (400 bars), whereas experimental flow laws indicate that a stress difference of nearly 100 Mpa is required to fracture a basaltic crust on Venus and up to 400 MPa are needed to cause failure of the near-surface dunite layer which is inferred from the presence of large rifts in Aphrodite Terra and Beta Regio and 300 km wide fold belts in Atalanta Planitia and the northern plains [2,3]. The second, more fundamental, problem is that the Pratt density anomaly in the model must extend to depths of 300-400 km. As the mechanical lithosphere of Venus is unlikely to be more than about 30 km thick [2], this density anomaly in the upper mantle cannot be statically maintained. These results led to the conclusion that whereas a static model can provide a useful first-order approximation to the stress field, motions in the mantle (presumably caused by solid state thermal convection) must have a major, and perhaps dominant, effect on both long-wavelength topography and subsurface density structure and thus on the lithospheric stress field [see also 4,5].

Preliminary results regarding the effects of a dynamic mantle have been obtained by Phillips [6] for the planar two-dimensional case in which a thick-plate elastic or viscous lithosphere overlies a newtonian viscous half space with an exponential depth dependence of the viscosity and with a specified density distribution driving the flow in the fluid. These theoretical solutions, combined with information on the depth and strength of density anomalies in the venusian interior inferred from analyses of long wavelength gravity data and topography, suggest that stresses in excess of 100 MPa would be generated in a 10 km thick elastic lid.

In order to address this problem more fully, a global spherical coordinate formulation of an elastic lithosphere coupled to a viscous mantle has been developed. The dynamic model is extended to spherical coordinates in three dimensions by combining the viscous mantle model of Richards and Hager [7] with a spherical thin elastic lithosphere [1]. The velocity, stress, and displacement boundary conditions are matched at the elastic-viscous boundary, and the resulting system of equations are solved for a piecewise layered rheological structure using propagator matrix techniques [8]. Spherical harmonic representations of the topography [9] and the gravity [10] to degree

and order 18 are utilized as external boundary conditions, and two laterally varying density anomalies (one in the viscous mantle, corresponding to chemical or thermal variations which drive the flow, and the other at the crust-mantle boundary, corresponding to variations in the crustal thickness) are included in the formulation. This results in an underdetermined linear system of twelve variables and eleven independent equations. If an additional condition is supplied, one can simultaneously solve for the state of stress at the surface (which can then be compared to observed tectonic deformational patterns) and the distribution of crustal thickness and mantle density variation over the planet. The approach taken here is to specify the flexural state of the lithosphere (uplifted, flexurally loaded, isostatic) as the remaining constraint. This formulation is superior to previous global dynamic stress models [11] in that the surface displacement boundary condition (*i.e.* the topography) is handled in a self-consistent manner and the effects of near surface density variations (such as crustal thickness variations) are taken into account.

As a first step toward understanding the effects of mantle dynamics on Venus's lithosphere, a vertically homogeneous density anomaly within the mantle is assumed. The simplifying assumption is also made that the topography is due solely to displacement of the lithosphere, as opposed to surface construction. This corresponds to the approach taken by Banerdt [1] and Kiefer *et al.* [5], and allows a first order comparison between stress fields due to static and dynamic support of topography and gravity anomalies. Preliminary results indicate that the maximum stress differences generated within the lithosphere for this dynamic model are as much as an order of magnitude higher than those for the static model, on the order of 400 MPa. Thus stresses generated in the lithosphere by long-wavelength density-driven flow in the mantle which is consistent with the observed gravity field are sufficiently large to cause tectonic disruption at the surface.

This formulation can also be thought of in terms of the elastic shell at the surface imposing a complex normal stress boundary condition on the viscous mantle which is a function of both the radial and tangential stresses as well as the normal displacement at the upper surface. Extensional horizontal tractions on the lithosphere tend to "flatten out" long-wavelength topography, whereas compressional tractions tend to increase its magnitude. This produces a deviation from the simple ρgh boundary condition used in [5] and [7] which varies from about 15% at the lowest harmonic degrees to a few percent at the higher degrees. As the potential calculated from these models results from a small difference between the two much larger contributions from the mantle density anomaly and the surface boundary deformation, even a small discrepancy in the surface displacement may have a large effect on the potential and therefore on the calculated spectral admittance. Thus the inclusion of an elastic boundary condition at the surface may be necessary for an accurate calculation of the low frequency spectral admittance.

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