D2-91

N86-31121

RIFTING ON VENUS: IMPLICATIONS FOR LITHOSPHERIC STRUCTURE

W. B. Banerdt and M. P. Golombek, Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109

Introduction

Several approaches have been used to estimate the elastic lithosphere thickness (L) on Venus. In particular, Solomon and Head [1] have used a number of compressional (elastic and viscous folding) and tensional (wedge subsidence, imbricate normal faulting, and plastic necking) models to predict L based on the hypothesis that the linear bands of greater and lesser backscatter observed in earth-based radar images of Ishtar Terra are of tectonic origin. All of these models predict very thin lithospheres, on the order of 0.3 to 8 km. These values are in agreement with those obtained from strength calculations based on laboratory measurements of crustal rocks ([1]; see below) which give negligible yield stresses at shallow depths due to the high surface temperature and a probable earth-like thermal gradient [2].

However, rift-like features which are at least superficially similar in size, morphology, and association with volcanism and doming to continental rifts on the earth [3] also exist on Venus. They typically have widths on the order of 75-100 km. Estimates of L derived from the widths of these features in Aphrodite Terra and Beta Regio using a wedge-subsidence model are on the order of 50-70 km [4], which is inconsistent with our knowledge of rock rheology at probable Venus temperatures. Plastic necking models also require a relatively thick, strong high viscosity zone (about 30 km thick [5]), again too thick for expected crustal temperatures.

How can these large rift structures which appear to require a thick, strong lithosphere be reconciled with the hot, apparently thin lithosphere implied by rock rheology at venusian conditions and the spacing of the banded terrain? In this abstract we briefly review lithospheric strength envelopes and explore their implications for large scale rifting on Venus. We then use these results to constrain possible crustal thicknesses and thermal gradients.

Lithospheric Strength Envelopes

The maximum stress levels found in the earth's crust are accurately predicted by Byerlee's law [6,7]. This relation is based on laboratory measurements of the frictional resistance to sliding on pre-existing fractures, which occurs at stresses less than those required to break intact rock. Byerlee's law is of the form $\sigma_1=\mu\sigma_3+b$, where σ_1 and σ_3 are, respectively, the maximum and minimum principle effective stresses (stress minus pore pressure), μ is the coefficient of friction, and b is a constant. Laboratory friction measurements show that μ and b are virtually independent of stress (except for a slight change at 135 MPa), rock type, displacement, surface conditions, and temperature. Because the vertical stress is generally quite close to the lithostatic load, this relation predicts a linear increase in yield stress with depth.

With increasing temperature, rock deformation occurs predominantly by ductile flow. Flow laws for many rocks and minerals have been experimentally determined for stresses up to 1-2 GPa and strain rates down to 10^{-8}/sec. These results can be extrapolated to geological strain rates via creep equations, which generally are of the form $\dot{\varepsilon}=A(\sigma_1-\sigma_3)^n\exp(-Q/RT)$, where $\dot{\varepsilon}$ is the strain rate, R is the gas constant, T is absolute temperature, and A, Q (the activation energy), and n are experimentally determined constants. As a result, the ductile strength is negligible at depths where T is high and increases exponentially with decreasing depth. Flow laws are also highly dependent on rock composition, with a silicic crust much weaker than a mafic mantle.

The failure criterion for a given depth in the lithosphere is determined by the weaker of the frictional or ductile strength at that depth (Figure 1). The yield stress increases with depth according to Byerlee's law until it intersects the crustal flow law. It then decreases exponentially until it reaches the moho, where the mantle flow law causes an abrupt increase in yield stress followed by another exponential decrease. In actuality, semibrittle and low temperature ductile processes tend to round off the intersection points between the brittle and ductile curves in Figure 1 [8]. The lithosphere will behave elastically for stresses less than the yield stress.

This type of analysis has been used by many investigators to study lithospheric strength on the earth and the terrestrial planets (e.g. [9-12]). However determining the geometry and characteristics of faults at failure for this type of lithospheric model has not received much attention. Based on an analysis of two phases of extension in the Rio Grande rift, Morgan and Golombek [13] suggested that during the early phase a shallow brittle-ductile transition along with the absence of a zone of upper mantle strength (due to an elevated geotherm) allowed a shallow decollement to develop with significant strain between the brittle upper crust and the ductilely extending material below. Thus, large strains resulted in the strongly rotated blocks bounded by numerous listric (curved) or planar normal faults which are characteristic of this period (see also Smith and Bruhn [10]). During the late phase of extension that formed the present horst and graben physiography the geotherm had cooled enough to allow a significant zone of ductile strength in the upper mantle (although most of the strength was still in the crust) which prevented large scale intracrustal decoupling and steep rotation of upper crustal blocks.

Lithospheric Strength on Venus

Figure 1 illustrates a typical lithospheric strength curve for Venus with a crustal thickness (c) of 10 km. For these calculations we assume a crustal density (ρ_c) of 2.8 Mg/m^3, a mantle density (ρ_m) of 3.3 Mg/m^3, a surface temperature of

$740^{\circ}C$, a thermal gradient of $12^{\circ}/km$ [2], and $\dot{\varepsilon}=10^{-15}/sec$ (equivalent to about 3% extension per million years). Based on surface geochemical measurements which indicate a basaltic composition, we use a dry diabase flow law [14] for the crustal layer. A dry olivine flow law [15] is assumed for the mantle. In this abstract we are interested in horizontal tensional stresses, which correspond to the left-hand branch of the curves in Figure 1.

Implications for Rift Width and Crustal Thickness

One consequence of this model is that the mechanical structure of the lithosphere depends strongly on the crustal thickness. For crusts thicker than about 30 km, there is no appreciable strength in the mantle and the brittle extension of the upper crustal layer should be effectively decoupled from the mantle. This was the implicit assumption of Solomon and Head [1]. However if the crust is thin, the lithospheric strength will be dominated by the upper mantle, and it is possible that this layer will control the style of deformation. In this case the effective lithosphere which controls rift width would be much thicker than the few km expected from crustal rock rheology. Instead, the elastic lithosphere thickness would be on the order of 20-30 km, based on an olivine flow law.

Here we will consider the implications for one theory for rift formation, the wedge-subsidence hypothesis of Vening Meinesz [16,17]. This theory predicts that the graben width w will be given by $w=\pi\alpha/4$, where $\alpha=(EL^3/3g\rho(1-\nu^2))^{1/4}$, E is Young's modulus, g is gravitational acceleration, ρ is the difference in density between the layers above and below the faulted layer, and ν is Poisson's ratio. For rifting of a crustal layer, ρ is just ρ_m. However for a layer at depth with an effectively fluid layer above it, $\rho=\rho_m-\rho_c$. Assuming $E=1.25\times10^{11}$ Pa, $\nu=.25$, $\rho_m-\rho_c=0.5$ Mg/m^3, and L=20 km gives the width for a graben formed in the mantle brittle zone of 98 km, in good agreement with observed rift widths on Venus' surface.

This result requires that the crust be no thinner than ~3 km (in order for there to be a viscous layer over the mantle strong zone) and no thicker than ~20 km (in order for there to be a brittle zone in the mantle). In addition, the thermal gradient could not be significantly greater or less than $10\text{-}15^{\circ}/km$, since L should be ~20 km. It should be noted that these values depend on flow laws which are somewhat uncertain due to their extrapolation from laboratory conditions.

This model is not inconsistent with the thin lithosphere implied by the banded terrain. One might expect smaller scale fault blocks whose widths are controlled by the thickness of the crustal strong layer to be superimposed upon the rift structure (see Figure 2). In Ishtar Terra, where no rifts are observed, the bands could have been formed by compressive stresses [1,18], or the crust in this region might be thicker than in rifted areas [2,18].

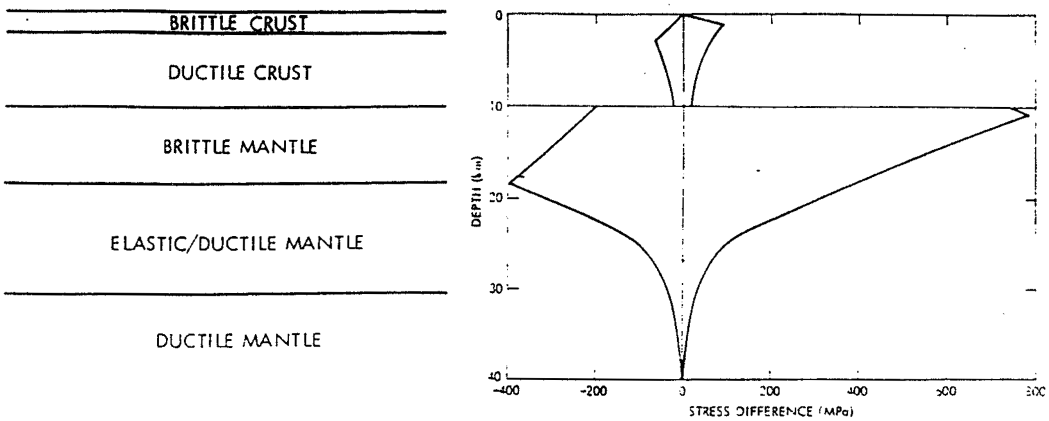

Figure 1. Lithospheric strength envelope for Venus. Stress difference is the vertical minus the maximum or minimum horizontal stress. Thus negative values denote tension and positive values correspond to compression. The horizontal line at 10 km shows the crust-mantle boundary assumed for this case. The flow laws used for the crust and mantle are dry diabase [14] and olivine [15], respectively; both assume a strain rate of 10^{-15}/sec, a surface temperature of 730^{0}C, and a thermal gradient of 12^{0}/km. The properties of the layers delineated by the tensional envelope are shown to the left. The elastic/ductile portion of the mantle deforms ductilely when its yield stress is exceeded, but behaves elastically at the stress levels required for ductile flow in the lower crust.

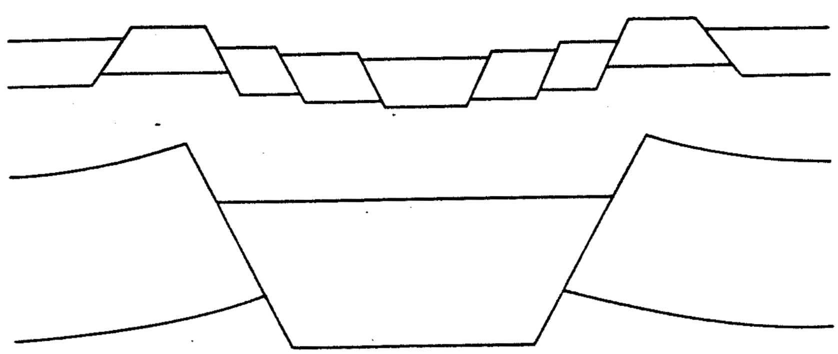

Figure 2. Schematic diagram of a modified Vening Meinesz model for rift formation on Venus, with the strong mantle layer controlling rift width and the brittle crustal layer controlling fault spacing.

References

[1] Solomon, S.C. and J.W. Head (1984) Venus banded terrain: Tectonic models for band formation and their relationship to lithospheric thermal structure, JGR, 89, 6885-6897.

[2] Morgan, P. and R.J. Phillips (1983) Hot spot heat transfer: Its application to Venus and implications to Venus and Earth, JGR, 88, 8305-8317.

[3] McGill, G.E., S.J. Steenstrup, C. Barton and P.G. Ford (1981) Continental rifting and the origin of Beta Regio, Venus, GRL, 8, 737-740.

[4] Schaber, G.C. (1981) Venus: Limited extension and volcanism along zones of lithospheric weakness, GRL, 9, 499-502.

[5] Zuber, M.T., E.M. Parmentier and J.W. Head (1985) Ductile lithosphere extension: Implications for rifting on the earth and Venus (abs.), Lunar Planet. Sci. XVI, 948-949.

[6] Byerlee, J. (1978) Friction of rocks, Pageoph, 116, 615-626.

[7] Brace, W.F. and D.L. Kohlstedt (1980) Limits on lithospheric stress imposed by laboratory experiments, JGR, 85, 6248-6352.

[8] Kirby, S.H. (1980) Tectonic stresses in the lithosphere: Constraints provided by the experimental deformation of rocks, JGR, 85, 6353-6363.

[9] Vink, G.E., W.J. Morgan and W.-L. Zhao (1984) Preferential rifting of continents: Asource of displaced terranes, JGR, 89, 10072-10076.

[10] Smith, R.B. and R.L. Bruhn (1984) Intraplate extensional tectonics of the eastern basin-range: Inferences on structural style from seismic reflection data, regional tectonics, and thermal-mechanical models of brittle-ductile deformation, JGR, 89, 5733-5762.

[11] Banerdt, W.B. and M.P. Golombek (1985) Lithospheric strengths of the terrestrial planets (abs.), Lunar Planet. Sci. XVI, 23-24.

[12] Solomon, S.C. (1985) The elastic lithosphere: Some relationships among flexure, depth of faulting, lithosphere thickness, and thermal gradient (abs.), Lunar Planet. Sci. XVI, 799-800.

[13] Morgan, P. and M.P. Golombek (1984) Factors controlling the phases and styles of extension in the northern Rio Grande rift, N.M. Geol. Soc. Guide. 35th Field Conf., 13-19.

[14] Shelton, G. and J. Tullis (1981) Experimental flow laws for crustal rocks (abs.), EOS Trans. AGU, 62, 396.

[15] Goetze, C. (1978) The mechanisms of creep in olivine, Phil. Trans. R. Soc. Lond., A288, 99-119.

[16] Vening Meinesz, F.A. (1950) Les graben africains, résultat de compression ou de tension dans la croûte terrestre?, Bull. Inst. R. Colon. Belge, 21, 539-552.

[17] Bott, M.H.P. (1976) Formation of sedimentary basins of graben type by extension of the continental crust, Tectonophys., 36, 77-86.

[18] Banerdt, W.B. (1985) Support of long wavelength loads on Venus and implications for internal structure, submitted to JGR.