GEOPHYSICAL RESEARCH LETTERS, VOL. 19, NO. 21, PAGES 2111-2114, NOVEMBER 3, 1992

GLOBAL DECOUPLING OF CRUST AND MANTLE: IMPLICATIONS FOR TOPOGRAPHY, GEOID AND MANTLE VISCOSITY ON VENUS

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Abstract. The surface of Venus is so hot that the lower crust may be weak enough to allow decoupling of mantle and crust. An analytic model of such decoupling assumes that the shallow mantle forms the top boundary layers of large scale mantle convection cells. Crustal flow is driven by the motion of the mantle and by topographically induced pressure gradients. The model predicts that the lowest lowlands are sites of mantle upwelling and thinner than average crust. Highlands are places where mantle downwells and the crust is thick. Surface heat flow is inversely correlated with elevation, consistent with recent estimates of brittle layer thickness variations on Venus. If the average crustal thickness is about 20 km then the average lower crustal viscosity must be close to 10^{18} Pa s to allow decoupling. The observed amplitude of geoid highs over highlands requires an Earth-like increase in mantle viscosity with depth.

Introduction

One of the most surprising differences between Venus and Earth is that the long wavelength geoid shows a strong positive correlation with topography on Venus, unlike on Earth (Sjogren et al., 1983). The accepted interpretation of this observation is that the convecting mantle of Venus has a constant viscosity with depth (Phillips and Malin, 1984, Kiefer et al, 1986; Kaula; 1990; Kiefer and Hager, 1991a; Bindschadler et al., 1992). According to these workers, topography results from vertical normal stresses caused by mantle convection, and highlands occur where mantle upwells and lowlands where mantle downwells.

This view of Venus is in marked contrast to the accepted view of Earth, where most topography is not a result of vertical tectonics. Much of geoid variations on Earth are probably related to mantle dynamics, but viscosity must increase by about two orders of magnitude through the mantle to explain these anomalies (e.g. Hager, 1984).

The weakness of hot Venusian crust may provide an alternative explanation of the observed correlation of geoid and topography. If crust is very weak then topography can result from horizontal mantle motions. For this to occur, topographic gradients must drive crustal flow as fast as flow caused by mantle shearing. This situation is termed decoupling. Decoupling requires that crust is thickened where mantle downwells and thinned where mantle upwells (Figure 1). To explain the correlation of geoid and topography, the viscosity of the mantle must increase with depth.

Local decoupling of crust and mantle is inferred to occur on Earth in areas of anomalously thick and hot crust such as Tibet or the Basin and Range Province of the Western U.S. (e.g. Burchfiel et al., 1989). Many authors have noted that the high temperature of the surface of Venus may cause the crust to be quite weak at relatively shallow depth (Weertman, 1979, Grimm & Solomon, 1988) and the effects of local decoupling have been modeled by several workers (e.g. Smrekar & Phillips, 1988, Bindshadler & Parmentier, 1990, Kiefer & Hager, 1991b). Decoupling may occur everywhere on Venus because its surface is 450°C on average.

In this paper I discuss the requirements for and consequences of global decoupling of crust and mantle on Venus. The first step in this process is to estimate the velocity of mantle motions.

Paper number 92GL02462
0094-8534/92/92GL-02462$03.00

Thermal and Rheologic Model

By treating the cooling of a mantle plate in terms of the cooling of a halfspace of constant initial temperature we can estimate the heat flux out of the plate. Figure 2 shows the boundary conditions assumed here. The average heat flux out of the top of such a plate of length L moving at velocity u_p is:

$$q_{ave} = 2\,K\,\Delta T\,\left(\frac{u_p}{\pi\,\kappa\,L}\right)^{1/2} \qquad (1)$$

where K is the conductivity, ΔT is the temperature drop across the plate and κ is the thermal diffusivity of the plate (Turcotte and Schubert, 1982).

If the heat budget of Venus is similar to that for Earth then the average surface heat flux of Venus is about 70 mW/m^2 (Solomon and Head; 1982). Plate recycling might account for about 50 mW/m^2 of this flux. The shallow mantle temperature based on parameterized convection models for Venus are somewhat higher than for the Earth (e.g. Phillips and Malin, 1984). Here the base of the mantle plate is set at 1450°C. Crust forming rocks can begin to partially melt at relatively low temperatures, and melting should buffer the temperature of the crust. The temperature of the base of the crust T_M is set at 850°C, and this gives $\Delta T = 600\,°C$. Taking $K = 4\ Wm^{-2}\,°C^{-1}$ and $\kappa = 10^{-6}\ m^2\ s^{-1}$, the plate velocity would have to be just over 5 cm/yr to give an average heat flux of 50 mW/m^2 for $L = 5000$ *km*.

Neglecting heat producing elements in the crust, and assuming the crust is in thermal equilibrium with the heat flux coming out of the plate, the temperature at depth z in the crust will be:

$$T(z) = T_S + \frac{q(x)}{K_c}\,z \qquad (2)$$

where T_S is the temperature of the surface, K_C is the crustal conductivity assumed to be half that of the mantle. A linear temperature gradient is assumed between the depth where $T=750\,°C$ and the base of the crust, as shown in Figure 3. The base of the crust may be cooler than 850°C if the heat flux from the mantle plate is low enough. The temperatures in the mantle

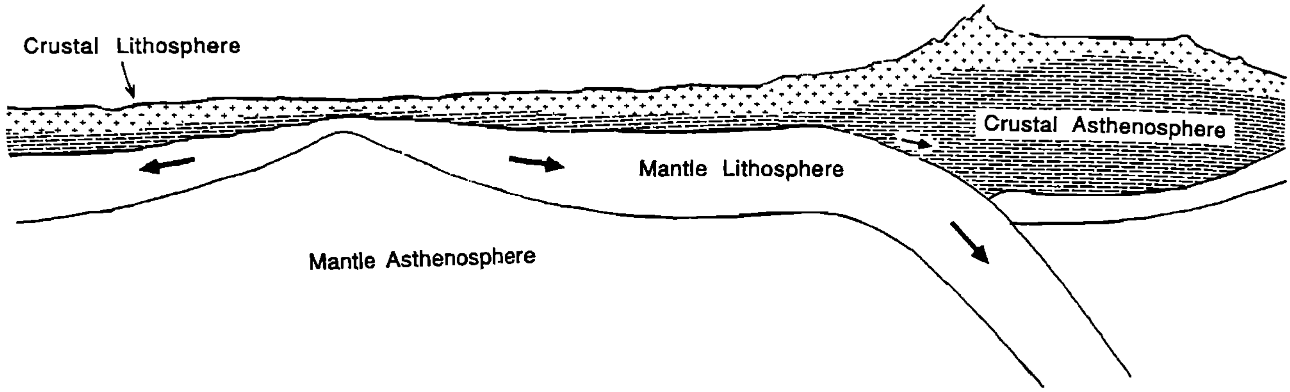

Fig. 1. Cross-section illustrating the model of decoupling of relatively static upper crustal lithosphere from moving mantle lithosphere.

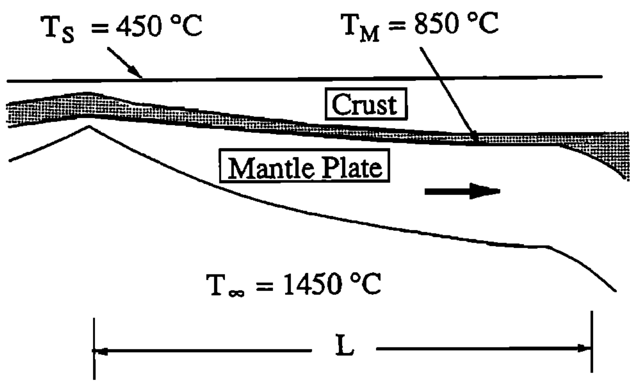

Fig. 2. Boundary conditions for subcrustal mantle plate thermal model. The shaded area is the weak lower crustal channel.

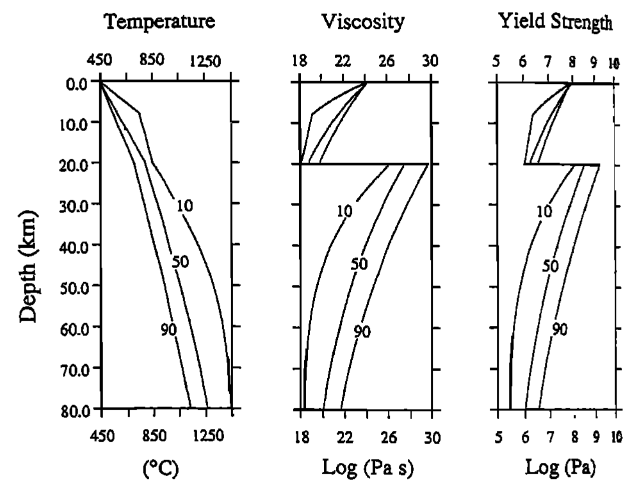

Fig. 3. Profiles of temperature, effective viscosity and yield strength for the crust and mantle for three plate cooling ages given in units of millions of years on the plots. The viscosity is calcualted for a deviatoric stress of 1 MPa the yield strength for deformation at a strain rate of 10^{-14} s^{-1}.

plate where the crust is cooler are then given by the solution for half-space cooling with a lower top boundary temperature. The small error due to this approximation should not significantly affect the results.

When minerals are hot enough for ductile flow to occur, results of laboratory deformation experiments can be expressed in terms of the strain rate $\dot{e}$ as a function of the applied deviatoric stress σ and temperature T as:

$$\dot{e} = A\,\sigma^n \exp\left(\frac{-E}{RT}\right) \tag{3}$$

where n is the power law exponent, E is the activation energy and R is the universal gas constant. In the model calculations the values of A, E and n for the crust and mantle are given by parameters determined for anorthisite (Koch, 1983) and olivine (Kirby and Kronenberg; 1987), respectively. The strength of a material is the deviatoric stress that can be maintained for a given strain rate and temperature.

The effective Newtonian viscosity of the crust can be expressed as:

$$\mu = \mu_0 \exp\left[\frac{E}{R}\left(\frac{1}{T} - \frac{1}{T_M}\right)\right] \tag{4}$$

where μ_0 is the viscosity of the crust at $T=T_M$. The value of μ_0 is adjusted in these calculations to match observations. Figure 3 shows yield strength and effective viscosity profiles used in the present calculations.

Crustal Flow

Crustal flow can be driven in two ways: by the relative motion between mantle and crust, and by pressure gradients related to topography. Consider the lower crust to consist of a channel of viscosity μ and thickness H. The flux F_s of crust moved laterally due to mantle shear is $u_s H/2$, where u_s is the relative velocity between crust and mantle. The flux Fp of crust driven by pressure gradients is $(H^3/12\mu)\ \partial P/\partial x$, where P is pressure (see Turcotte and Schubert, 1982). The pressure gradient $\partial P/\partial x$ is related to isostatic topography as $\partial P/\partial x = g\rho_c \partial w/\partial x$, where g is the acceleration of gravity, w is topography, h is crustal thickness, and ρ_c is the density of the crust.

For a constant shear velocity u_s equal to the plate velocity, topography can be maintained in steady state as long as $F_p = F_s$ as noted by Kiefer and Hager (1991b). For this case the equilibrium topographic gradient is:

$$\frac{\partial w}{\partial x} = 6u_p \frac{\mu}{g\rho_c H^2} \quad . \qquad (5)$$

Calculations show that most of the flow driven by pressure gradients in a layer with a depth dependent viscosity occurs in the region where the viscosity is within ten times the layer minimum (Buck, 1991). Thus, the lower crustal channel is taken to be the region with a temperature within 50°C of T_M. The average viscosity of the channel is taken to be the viscosity of the base of the crust.

In this model the highlands are supported by stress transmitted laterally by the crust and mantle. The magnitude of this stress can be estimated by integrating the shear stress at the base of the crust with horizontal distance. For a mantle lithospheric plate moving at a velocity u_p relative to crust which has a low viscosity layer of thickness H and viscosity μ, the shear stress τ_c is $\mu u_p/H$. The maximum tectonic force G per unit length maintained by a mantle plate and the upper crust due to the shear between them is just the integral of τ_c over horizontal distance from where the plate originates (i.e. at an upwelling). This interaction will cause the crust to be in compression and the mantle in extension.

Results

Results from two model calculations for a 5000 km plate moving at 5 cm/yr are presented here. The crustal thickness on Venus is generally estimated to be less than 30 km (Zuber and Parmentier, 1990). Results are given for crustal thicknesses of both 20 and 30 km. The bottom panel of Figure 4 shows that surface heat flow varies from over 100 mW/m^2 within 200 km of a center of plate spreading down to less than 25 mW/m^2.

The most important result of these calculations concerns the equilibrium topographic slope as a function of position (see top of Figure 4). In the area where the plate is old the slopes are large and in the area of mantle divergence the slopes are small. This is a result of the topographic gradient depending on the lower crustal thickness and viscosity, as given by Equation 6. In an area of mantle plate divergence the lower crust is thicker, hotter and weaker than in an area of mantle downwelling. The long wavelength slope of the surface in the lowlands of Venus is generally less than 10^{-3} m/m. The value of μ_0 in equation 4 had to be set to 3×10^{17} Pa s to give about this slope. In the coldest area for a 20 km thick crust, the lower crustal viscosity is about 10^{19} Pa s.

The next to the bottom panel shows the calculated values of the tectonic force $G(x)$ for these calculations. The shear stress is greater where the heat flow is small, thus most of the tectonic force builds up over the cold part of the mantle plate. A value of $G = 5 \times 10^{12}\ N\ m^{-1}$ is about the force required to support the highest elevations on Venus if the average crustal thickness is 20 km. This is the amount of tectonic force resulting from an average stress τ_c of 1 MPa integrated over a 5000 km plate, or 5 MPa over the last 1000 km of the plate.

The force that can be transmitted by the mantle plate without large magnitude deformation, the lithospheric strength, is the integral of the yield strength over depth. The plots in Figure 4 show that the mantle lithospheric strength is greater than the forces due to coupling between crust and mantle. Therefore, the plate should not be disrupted.

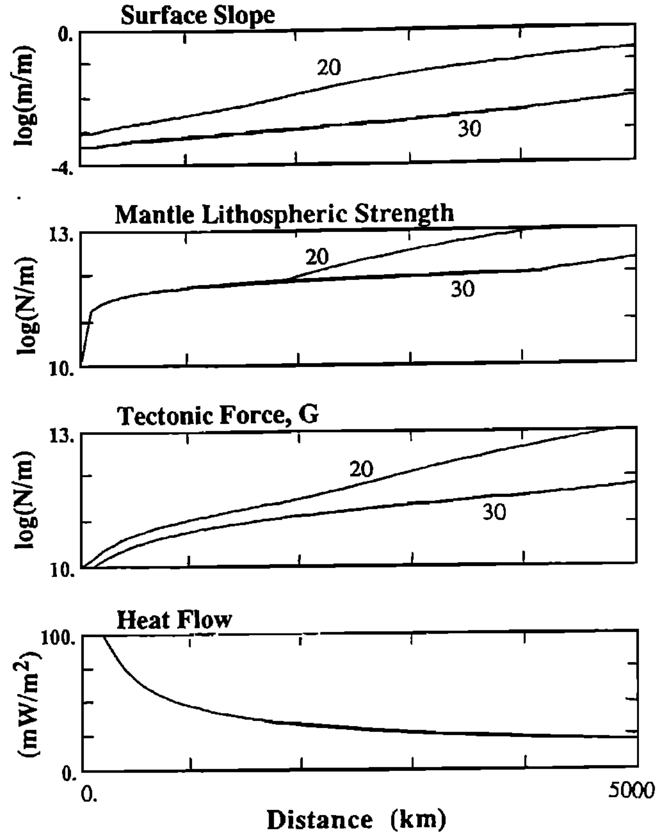

Fig. 4. Calculation results for two crustal thickness, 20 and 30 km, as labeled. See text for discussion.

Discussion

These simplified model calculations show that the lower crustal viscosity must average 10^{18} Pa s or less to allow decoupling of crust and mantle. In the flattest lowland regions the viscosity of the lower crust must be at least several times smaller than this. At temperatures that allow the mantle to retain considerable strength, flow laws for crustal rocks predict higher viscosities than assumed here. Partial melting or the formation of shear zones might lead to a local reduction in viscosity at the base of the crust.

Numerical calculation of equilibrium topographic profiles are beyond the scope of this paper. However, it should be clear that highlands are predicted to occur over areas where mantle plates converge and downwell. The lowest lowlands occur where mantle upwells. Thermal buoyancy should elevate the mantle at a site of plate spreading, but the crust should be thinned there so that the elevation is lower than the average (Figure 1). The main thickening of the crust in highlands would be accomplished by the transport of lower crust, in much the same way that is discussed by Kiefer and Hager (1991b). The spacing of lowlands and highlands is on the order of 1000's of km which is the reason for taking L=5000 km as a representative size in the example shown in Figure 4.

Decoupling on a global scale can explain the low rate of surface deformation inferred for Venus (e.g. Solomon et al., 1991). The strongest crustal deformation should be in the same regions where large topographic slopes could be maintained: namely, where the plate is cold and so the lower crust is strong. However, simple predictions of patterns of deformation may be difficult to make since stress can be transmitted both by the upper crust and the mantle lithosphere. The tessera terrain indicate that many parts of the upper crust have failed and deformed.

Variability in plate size and speed may explain the variability in topography and deformation patterns seen in different highlands. The crust above a larger plate would be cooler, higher viscosity and so better coupled to the plate. Better coupling should lead to strong compressional deformation as is seen in plateau-shaped highlands such as Ishtar Terra and Thetis Regio. A small fast-moving plate might not result in significant deformation of the surface of the crust. This might produce domal highlands such as Beta and Atla Regio which show more evidence of extension than compression (Bindschadler et al., 1992). The extension would be driven by the collapse of the high topography.

Predicting the regional styles of volcanism associated with this model requires assumptions regarding the composition of the mantle and crust. Without making those assumptions I can say that melting of the upwelling mantle should be greatest in the regions of mantle plate divergence in the lowlands. Many topographically low regions are covered with relatively undeformed volcanics, though the most obvious sources are at higher elevations (Head et al., 1992). Crustal melting might occur in the highlands where thick crust could be very hot.

The highest heat flow is predicted to occur in the lowest lowlands of Venus and the lowest heat flow should occur in topographically elevated areas. This is qualitatively consistent with a recent estimate of the variation of heat flow based on tectonic features. Suppe and Connors (1992) show that the amplitude of the topographic step across compressional mountain belts scales with the elevation of the lower side of the belt. They argue that the topographic step is roughly equivalent to the thickness of the brittle layer of upper crust. The data cannot be explained solely by the lapse rate in the atmosphere, but requires that heat flow be inversely correlated with elevation (Suppe and Connors, 1992). Whether the present model can quantitatively match these observations can only be addressed with numerical models of decoupling.

Acknowledgments. Thanks to John Suppe and Maria Zuber for helpful discussions and to Walter Kiefer, John Hopper and two anonymous reviewers for comments on the manuscript. Support came from NSF grant OCE-91-04076. Lamont-Doherty Geological Observatory contribution 4999.

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(Received: June 9, 1992;
revised: August 24, 1992;
accepted: September 9, 1992)