Constraints on the Lithospheric Structure of Venus from Mechanical Models and Tectonic Surface Features

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Radar images of the surface of Venus show numerous structures that are interpreted as having formed due to horizontal compression and extension of the lithosphere. Many of these features exhibit characteristic scales (widths or spacings) of deformation, one of 10–20 km and another of 100–300 km. For a range of simple models, we test the hypothesis that these length scales are controlled by dominant wavelengths arising from unstable compression or extension of the Venus lithosphere. Results show that the existence of tectonic features that exhibit both length scales can be explained if, at the time of deformation, the lithosphere consisted of a crust that was relatively strong near the surface and weak at its base and an upper mantle that was stronger than or of nearly comparable strength to the upper crust. The spacings of these features imply crustal thicknesses in the approximate range 5–30 km and a thermal gradient not greater than 25 K km⁻¹. Features that exhibit only the smaller scale of deformation can be explained by either a lithosphere with a thick crust that overlies a weak mantle or a lithosphere with a strong mantle but with small internal strength contrasts. For a broad range of parameters, the models require not only that smaller scale compressional and extensional features have similar spacings, but also that the widths and spacings of larger scale compressional features be greater than those formed in extension. This is consistent with observed differences in the characteristic length scales of larger scale surface features, as well as the similarities in spacings of observed smaller scale features.

INTRODUCTION

Radar images of the surface of Venus obtained by Pioneer-Venus [Masursky et al., 1980; Pettengill et al., 1980], earth-based observations [Campbell et al., 1983, 1984], and the Venera 15/16 spacecraft [Barsukov et al., 1986; Basilevsky et al., 1986] reveal a variety of features of presumed tectonic origin. Although, as shown in Figure 1, these structures are globally distributed over the surface of the planet, many exhibit distinct morphological similarities that suggest a common mechanism of formation. Specifically, numerous features display linear trends and characteristic widths and/or spacings that are suggestive of deformation controlled by dominant wavelengths arising from compressional and/or extensional instabilities in the lithosphere. Previous analyses have shown how deformation at preferred wavelengths can arise when a model lithosphere that contains one or more rheologically competent layers is horizontally stressed [Fletcher and Hallet, 1983; Zuber et al., 1986; Zuber and Parmentier, 1986; Ricard and Froidevaux, 1986; Zuber, 1986a]. Since these wavelengths are primarily controlled by the thicknesses of the competent layers and the strength and density stratification of the lithosphere, observations of the geometries of tectonic features arising from unstable deformation provide a direct indication of the underlying lithospheric structure. Unstable deformation has been invoked to explain large-scale extensional tectonic features in the Basin and Range Province [Fletcher and Hallet, 1983; Froidevaux, 1986; Zuber et al., 1986; Ricard and Froidevaux, 1986] and rift zones [Zuber and Parmentier, 1986] and intraplate compressional deformation structures on the seafloor [Zuber, 1986b]. This mechanism may also explain periodic anomalies in the earth's gravity field related to long wavelength deformation of the upper mantle [Froidevaux, 1986; Zuber et al., 1986; Ricard et al., 1986; Zuber, 1986b]. If the regular development of tectonic features on Venus is a consequence of unstable deformation, then the widespread occurrence of these structures provides direct evidence for the lithospheric structure of this planet on a global scale.

In this study, we investigate the stability of the Venus lithosphere in extension and compression. The purpose is to assess the conditions for which unstable compressional and extensional deformation can develop, and to relate the predicted dominant wavelengths to the geometries of observed tectonic features and the rheological structure of the lithosphere. We begin with a discussion of evidence for the extensional or compressional origin of some prominent surface features, followed by a description of morphological characteristics of features that are consistent with formation due to unstable deformation. Subsequently, we show how variations in temperature, pressure, and composition with depth in the lithosphere can result in strength contrasts that may be of sufficient magnitude to allow unstable deformation to develop. Finally, we discuss implications for the compositional and rheological structure of the Venus lithosphere in terms of the relationships between the characteristic widths and spacings of observed features and various model parameters. We conclude that unstable compression and extension constitute viable mechanisms for the formation of tectonic features on Venus. On the basis of observations and model results, we suggest that the Venus lithosphere, at least in some regions, contains two mechanically competent layers that we infer to correspond to the upper crust and upper mantle.

TECTONIC FEATURES

Beta Regio Rift

One of the most conspicuous tectonic features on Venus is the Beta Regio Rift (see Figure 2). First interpreted as an extensional feature by Masursky et al. [1980], the rift trends approximately north-south and runs for over 1000 km. As observed in 2-km-resolution Arecibo radar images [Campbell et al., 1984], the rift consists of a central depression with a depth of over 1 km flanked by topographic uplifts. The average
Fig. 1. Sketch map of Venus showing the locations of some prominent surface features of presumed tectonic origin that exhibit one or two length scales of deformation. Features described in this study that exhibit smaller scale (10-20 km) spacings are located in Beta Regio, Akna and Freyja Montes, and in the ridge belts. Features that exhibit larger scales of deformation are located in Beta Regio, Aphrodite Terra, and the ridge belts.

flank-to-flank width of the rift is in the range 100-200 km. McGill et al. [1981] have noted that this feature is morphologically similar to the earth’s East African Rift. Within the central depression of Beta Regio are alternating radar-bright and -dark lineations, which have characteristic spacings of 10-20 km and are continuous along strike for up to several hundreds of kilometers. On the basis of the Arecibo radar data, which are sensitive to variations in surface roughness at centimeter-to-meter scales, the lineations have been interpreted as faults by Campbell et al. [1984]. Characterization and mapping of combined Arecibo and Venera data sets led Stofan et al. [1986] to suggest that these faults were produced by extension related to the formation of the central rift depression.

**Dali and Diana Chasmata**

Two prominent ridge-and-trough systems within the equatorial highlands of Aphrodite Terra, Dali and Diana Chasmata, were interpreted as rift structures by Pettengill et al. [1979] and Masursky et al. [1980]. On the basis of Pioneer-Venus altimetry, these features have lengths of over 1000 km [Ehmann and Head, 1983], widths of 75–100 km, depths greater than 2 km, and raised rims with heights of 0.5-2.5 km [Schaber, 1982]. McGill et al. [1983] noted that the terrain within the central troughs of the chasmata is of a complex nature, but agreed with the interpretation of a tectonic origin. Smaller scale lineations such as those found within the Beta Regio Rift are not observed in the chasmata; however, at Pioneer-Venus resolution such features, if they are present, are not resolvable.

**Banded Terrain**

Banded terrain is located in the mountains surrounding Lakshmi Planum in the Ishtar Terra region of Venus. As revealed in 3-km-resolution Arecibo radar images [Campbell et al., 1983], these features consist of linear bands of alternating greater and lesser backscatter that are generally aligned in the directions of the topographic contours that define the mountain ranges. Bands are continuous for several hundreds of kilometers and have characteristic spacings of 10-20 km. On the basis of their parallel and continuous nature and their relationship to the topography, Campbell et al. [1983] suggested that the bands are most likely of tectonic origin. Solomon and Head [1984] showed that a number of simple extensional and compressional models are compatible with the spacings of bands for plausible ranges of physical properties of the Venus lithosphere. However, these authors suggest that the linearity and continuity of the bands combined with evidence for band closure in Maxwell
Fig. 3. Sketch map showing the distribution of ridge belts (shaded) in part of northern hemisphere of Venus as determined from Venera 15 and 16 radar images [after Barsukov et al., 1986]. Within the ridge belts, smaller scale ridges and grooves strike approximately parallel to the belts. The spacings of the ridge belts and the ridges and grooves define two length scales of apparent compressional deformation.

Montes are most consistent with folding or faulting in response to regional-scale compression.

**Ridge Belts**

Figure 3 shows a sketch map of a group of prominent features termed ridge belts that were revealed by Venera 15 and 16 imaging radar [Barsukov et al., 1986; Basilevsky et al., 1986]. The Venera radar system is sensitive to surface slopes and therefore topography. On the basis of the Venera data, the ridge belts have been described as north-south trending linear ridges that are continuous along strike for distances of up to 1000 km [Basilevsky et al., 1986]. Direct measurements from maps derived from the Venera data show the ridge belts to have a regular spacing perpendicular to strike of approximately 300 km [Zuber, 1986c]. Within the ridge belts, systems of subparallel ridges and grooves trend generally parallel to the belts. The ridges and grooves are continuous for distances of 100–200 km and have regular spacings of 10–20 km [Basilevsky et al., 1986]. On the basis of their similarity to the banded terrain in Ishtar Terra, the ridge belts have been interpreted as a product of compressional deformation by Basilevsky et al. [1986].

**Relationships of Tectonic Features to Lithospheric Structure**

All of these features have linear and often parallel strikes, regular spacings, and/or characteristic widths. Other tectonic features that exhibit these characteristics are discussed by Basilevsky et al. [1986]. Table 1 shows that the widths and spacings of these features fall broadly into two ranges, one of 10–20 km and another of approximately 100–300 km. The Beta Regio Rift and the ridge belts exhibit both scales of deformation, while the banded terrain shows only the smaller scale. For the chasmata only the larger scale of deformation is resolvable with present data, so it is unknown whether a smaller scale exists.

Solomon and Head [1984] examined a range of tectonic models for the banded terrain and concluded that the regular spacing of the bands was controlled by a brittle or high-viscosity surface layer that they interpreted as the Venus elastic lithosphere. In studies applied to the Basin and Range Province of the western U.S., Zuber et al. [1986] and Ricard and Froidevaux [1986] showed that two scales of periodic deformation can arise in an extending medium that contains two strong layers separated by a weaker layer. For the earth's continental lithosphere, the strong layers correspond to the upper crust and that part of the mantle that makes a major contribution to lithospheric strength, while the intermediate layer corresponds to the weak part of the ductile lower crust [Froidevaux, 1986; Zuber et al., 1986; Ricard and Froidevaux, 1986]. By analogy, it is suggested that the two scales of tectonic features on Venus can also be explained by a lithosphere that, as illustrated in the following section, consists of a crust that is weaker at its base than the underlying mantle.

**MODEL DEVELOPMENT**

**The Venus Lithosphere**

Figure 4 shows a possible range of maximum principal stress differences that can be supported by the Venus lithosphere in

<table>
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<td>Lineations</td>
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<tr>
<td>Dali and Diana Chasmata Rift Zone Width</td>
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<td>Not resolvable if present</td>
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<td>Banded Terrain Bands</td>
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<td>Ridge Belts Belt spacings</td>
<td>300</td>
<td>Ridges-and-grooves</td>
<td>Compressional</td>
<td>Venera 15/16</td>
<td>B</td>
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*C1 = Campbell et al., [1984]; S = Schaber [1982]; C2 = Campbell et al. [1983]; B = Basilevsky et al. [1986].
correspond to the upper crust and the upper mantle just beneath layer. As for the earth's continental lithosphere, the strong layers upper mantle is considerably stronger than the lower crust. A could contain two relatively strong layers separated by a weaker in Figure 4, the lower crust will be weaker than the upper mantle crust that is thicker than that shown would be even weaker [Surkov et al., 1984]. A comparison of relative strengths at the in chemistry and mineralogy to tholeiitic and alkaline basalts in a number of rolling plains and lowlands sites to be similar in bulk composition of Venus and the earth [Basaltic Volcanism Study Project, 1981, pp. 682-685]. A diabase crust is assumed on the basis of Venera lander results that show surface rocks that overlies a halfspace. Two different strength stratifications are considered for the mantle. In the strength jump (J) model, strength is discontinuous at the base of the layer and falls to a lower uniform value in the substrate. In the continuous strength (C) model, strength is continuous at the base of the layer and decreases exponentially with depth in the substrate. Both of these strength stratifications are considerably simpler than that which is likely to exist on Venus (cf. Figure 4). However, the nearly analytical solutions for the flow obtained in each case permit insight into the physical nature of the deformation that cannot be gained from fully numerical solutions. While the growth rates of the instabilities predicted by the models vary somewhat, the dominant wavelengths are in good agreement to the degree that they can be compared [Zuber et al., 1986].

Idealized rheologies are chosen to approximate the response of the Venus lithosphere to imposed deviatoric stresses. Experiments have shown that a nonlinear viscous material with a steady-state constitutive law of the form $\dot{\epsilon} = \sigma^n$ (where $\dot{\epsilon}$ is the strain rate, $\sigma$ is stress, and $n \approx 3$ is the stress exponent) represents a medium in which deformation occurs primarily by ductile creep [e.g., Weertman and Weertman, 1975]. A perfectly plastic rheology, which is a continuum representation of a medium that undergoes deformation by faulting, is approximated by a stress exponent of $n = \infty$ in the steady-state creep relationship. Both perfectly plastic and nonlinear viscous rheologies are considered for the upper crustal and upper mantle layers, and nonlinear viscous behavior is assumed for the lower crustal layer and mantle substrate.

**Lithosphere Models**

The Venus lithosphere is assumed to consist of a crust and mantle with density contrasts at the surface and at the base of the crust. As shown in Figure 5, the crust is modeled as a strong layer overlying a weaker layer, each with uniform strength, and the mantle consists of a uniformly strong layer that overlies a halfspace. Two different strength stratifications are considered for the mantle. In the strength jump (J) model, strength is discontinuous at the base of the layer and falls to a lower uniform value in the substrate. In the continuous strength (C) model, strength is continuous at the base of the layer and decreases exponentially with depth in the substrate. Both of these strength stratifications are considerably simpler than that which is likely to exist on Venus (cf. Figure 4). However, the nearly analytical solutions for the flow obtained in each case permit insight into the physical nature of the deformation that cannot be gained from fully numerical solutions. While the growth rates of the instabilities predicted by the models vary somewhat, the dominant wavelengths are in good agreement to the degree that they can be compared [Zuber et al., 1986].

**Styles of Unstable Deformation**

In a layered medium that is extended or compressed at a mean horizontal strain rate, $\dot{\epsilon}_{xx}$, deformation develops under conditions for which small amplitude (much less than layer thicknesses) perturbations along the free surface or layer interfaces amplify with time [cf. Biot, 1957, 1960]. The nature of these perturbations in some instances determines the style of deformation that develops. In an extending or compressing medium in which initial perturbations are distributed randomly, deformation develops periodically at the dominant wavelength, $\lambda_d$, in a direction normal to the applied stress [e.g., Fletcher, 1974; Smith, 1975]. For a compressing medium with a single competent layer the asymmetric or folding mode of deformation is preferred, while for an extending medium the symmetric or pinch-and-swell mode is most likely. These are illustrated in Figure 6 for a medium consisting of a single strong layer overlying the Moho, and the weaker layer corresponds to the lower crust. Later results will show that in an extending or compressing model lithosphere with two strong layers separated by a weaker layer two wavelengths of deformation may develop. While the models do not require that the layers and substrate correspond to the crustal and mantle regions discussed above, this rheological stratification is appealing in its simplicity and plausibility. In the absence of better data from which to estimate Venus' near-surface composition and rheology, we will proceed under the assumption that the lithosphere is stratified in this manner.

![Figure 4](image_url)

**Figure 4.** Strength, defined as differential stress ($\sigma_1 - \sigma_3$), as a function of depth in the Venus lithosphere for horizontal compression (left) and extension (right). The strength envelope was constructed for a surface temperature of 700 K, a thermal gradient of 15 K km$^{-1}$, a strain rate of $10^{-6}$ s$^{-1}$, and zero pore pressure. The lithosphere consists of a diabase crust and a dry olivine mantle. An arbitrary crustal thickness of 10 km is assumed. The brittle strength of the crust and mantle was determined from Byerlee [1968], and the ductile strengths of diabase and olivine were taken from Shelton and Tullis [1981] and Brace and Kohlstedt [1980], respectively. Note for both compression and extension that the lower crust is weaker than the upper crust and upper mantle. Such a strength stratification may lead to unstable deformation in a compressing or extending lithosphere.

Compression and extension as predicted by laboratory experiments on rock rheology. Where stress increases approximately linearly with depth, deformation occurs in a brittle manner. Brittle strength is determined by the frictional resistance to sliding at fracture surfaces and is essentially insensitive to strain rate, temperature, and mineralogy [Byerlee, 1968]. At greater depths deformation occurs by ductile flow and strength is dependent on strain rate, composition, and most critically on temperature [e.g., Weertman and Weertman, 1975]. The mode of deformation at a given depth is determined by the lesser of the brittle and ductile strengths.

In Figure 4 it is assumed that mantle flow is governed by a dry olivine rheology on the basis of the hypothesized similarity in bulk composition of Venus and the earth [Basaltic Volcanism Study Project, 1981, pp. 682-685]. A diabase crust is assumed on the basis of Venera lander results that show surface rocks in a number of rolling plains and lowlands sites to be similar in chemistry and mineralogy to tholeiitic and alkaline basalts [Surkov et al., 1984]. A comparison of relative strengths at the Moho shows that diabase is weaker than olivine at the same P-T conditions for both compression and extension; thus the upper mantle is considerably stronger than the lower crust. A crust that is thicker than that shown would be even weaker at its base as the ductile strength of the lower crust progressively decreases with increasing depth. For the conditions assumed in Figure 4, the lower crust will be weaker than the upper mantle at the Moho for crustal thicknesses up to about 30 km.

Figure 4 illustrates a scenario for which the Venus lithosphere could contain two relatively strong layers separated by a weaker layer. As for the earth's continental lithosphere, the strong layers correspond to the upper crust and the upper mantle just beneath...
Fig. 5. Models of a Venus lithosphere containing a strong upper crust and upper mantle separated by a weak lower crust. The wider and narrower shading indicate the crust and mantle, respectively. The diagram at the left shows the parameters that describe the layers and substrate. Parameters $\rho$, $n$, and $\tau$ are the density, stress exponent in the stress-strain rate relationship, and strength, respectively. In the relationship between strain rate, $\dot{\gamma}$, and stress, $\sigma$, $Q$ is the activation energy, $R$ is the gas constant, $T$ is temperature, and $A$ is the frequency factor. The strength is defined by the product of the viscosity and the horizontal strain rate in the basic state of uniform extension or compression. The diagrams to the center and right schematically show the strength stratifications examined, where the length of each section represents the relative strength of the medium. In the J model strength in the mantle is discontinuous, while in the C model strength is continuous and decreases exponentially with depth in the substrate. Both models predict similar dominant wavelengths for compressional and extensional instabilities.

For a power-law viscous layer in which viscosity $\mu$ decreases with depth $z$ as

$$\mu = \mu_e \xi^{2/3}$$

where $\xi$ is the viscosity decay depth and $\mu_e$ is a reference viscosity, the governing equation for the perturbing flow is

$$D^2W + 2\zeta D^3W + \left[\xi^2 - 2k^2(2/n-1)\right]D^4W = 0$$

$$D^2W - 2k^2\dot{\gamma}^2(2/n-1)DW + k^2(2k^2 + \xi^2)W = 0$$

where $D = d/dz$, $k(=2\pi/\lambda)$ is the wave number, and $W$ is the stream function. Solutions of (2) in the form shown are given by Fletcher and Hallet [1983], while those for a layer of uniform viscosity or strength in which $\xi \rightarrow \infty$ are given by Zuber et al. [1986]. These solutions yield the perturbing velocities and stresses in the layers and substrate. The amplitude of a perturbation at the $i$th interface at time $t$, $\Delta_i(k,t)$, for a medium in compression can be written

$$\Delta_i(k,t) = \Delta_i(k,0) \exp \left[\left(Q_{\text{comp}} + 1\right)\dot{\gamma} \xi \Delta t\right]$$

and for a medium in extension

Theory

Fig. 6. Styles of unstable deformation in a medium consisting of a layer that is more competent than the underlying substrate and for which initial disturbances along interfaces are distributed randomly. For compression an antisymmetric or folding mode of deformation develops while for extension deformation occurs in a symmetric or pinch-and-swell mode. The parameter $\Delta$ refers to the amplitude of the perturbing flow at an interface. The total flow in the medium consists of the perturbing flow plus a pure shear component resulting from the basic state of uniform compression or extension illustrated by the arrows.
Fig. 7. Growth rate spectra for model lithospheres in compression and extension. The dominant wavelength occurs at the maximum of the growth rate, \( q \). The growth rate is nondimensionalized by the mean horizontal strain rate, \( \dot{e}_{xx} \), while the wave number and wavelength are nondimensionalized by the thickness of the surface layer, \( h_1 \). The upper curves, constructed for a medium with a single strong layer (shaded), have one peak and indicate deformation characterized by one dominant wavelength. The lower curves, constructed for a medium with two strong layers separated by a weaker layer, have two peaks and indicate deformation with two dominant wavelengths. Extensional and compressional instabilities with one and two dominant wavelengths may explain the length scales of many tectonic surface features on Venus. For both single and multiple layer models \( S_1 = 1, S_2 = 0, \alpha = 0.2 \) and 0.6 for compression and extension, respectively, \( n_1 = 10^4 \), and \( n_2 = 3 \). For the multilayer model \( S_3 = 0.01, S_2 = S_4 = 0, R_1 = 100, R_2 = 0.5, \) and \( n_2 = n_3 = 3 \).

\[
\Delta(k,t) = \Delta(k,0) \exp \left[ (q_{\text{ext}} - 1)\dot{e}_{xx} t \right]
\]

where \( \Delta(k,0) \) is the initial amplitude and \( q_{\text{comp}} \) and \( q_{\text{ext}} \) are the compressional and extensional growth rate factors. The value of unity in each of the exponential terms represents the kinematic distortion of the medium due to uniform extension or compression. Instabilities will amplify with time for conditions in which the exponentials are greater than unity. If this condition is not met, initial perturbations will decay and an extending or compressing medium will thin or shorten uniformly.

The growth rate factor reflects the relative contributions to dynamic instability growth of the driving force, which depends on the slope of the perturbed interface, and the viscous resistance of the medium to deformation. In an unstable medium, the wave number at which \( q \) is a maximum defines the wavelength at which a disturbance will grow most rapidly and eventually dominate the flow. This is the dominant wavelength [Biot, 1961]. In this formulation, the dominant wavelength is determined from the eigenvalue solution of the system of equations for the perturbing flow in each layer, the boundary conditions requiring the continuity of stresses and velocities at each interface, and the rates of perturbation growth at each interface. The method of solution for a multilayered medium is discussed by Zuber et al. [1986].

**RESULTS**

*Wavelength Selection*

Figure 7 shows examples of growth rate spectra for the C model in compression and extension. The growth rate is shown as a function of wave number, \( k' \), on the bottom axis and wavelength, \( \lambda/h_1 = 2\pi/k' \), on the top. Both are normalized to the thickness of the strong surface layer. The upper curve in each case, which corresponds to a medium with one strong layer, contains a single peak. This indicates deformation characterized by a single dominant wavelength, which is about four layer thicknesses for both the compressional and extensional cases shown. The lower curve in each case corresponds to a medium that contains two strong layers separated by a weaker layer. These growth rate spectra exhibit two maxima, which indicate that deformation develops with two dominant wavelengths.

In an unstable medium with two strong layers, the positions and amplitudes of the shorter and longer wavelength peaks in the growth rate spectrum are controlled to the greatest extent by the properties of these layers. However, because the strong layers are coupled to the weak layer and the substrate, the growth rates of the instabilities and the dominant wavelengths are determined by the mechanical properties of the entire medium. Consider an extending model lithosphere with a plastic upper crustal layer and a nonlinear (\( n = 3 \)) viscous lower crustal layer, mantle layer, and mantle substrate. Although, as illustrated in Figure 7, two wavelengths of deformation can develop, the mantle layer is not itself unstable, but deforms passively in response to the unstable deformation of the upper crustal layer. The longer wavelength arises because the mantle layer resists deformation, and suppresses over a range of wave numbers the instability induced by the upper crustal layer. The maximum suppression occurs at the relative minimum between the peaks in the growth rate spectrum [Zuber et al., 1986]. If the upper mantle contains a region in which deformation occurs predominantly by brittle (see Figure 4) or ductile [cf. Chapple and Forsyth, 1979] faulting rather than ductile flow, then a
plastic rheology may be appropriate for the upper mantle layer. An extending model lithosphere, in which both the strong upper crustal and mantle layers are plastic and the weak lower crustal and mantle substrate are nonlinear viscous, deforms in response to instabilities induced by each of the strong layers [Ricard and Froidevaux, 1986]. This illustrates that two wavelengths of deformation can develop if the lithosphere contains two competent layers separated by an incompetent layer; it is not necessary for both of the competent layers to be unstable.

Effects of Buoyancy Forces and Strength Stratification

If the stresses required for deformation are large, then buoyancy forces arising from density contrasts within the lithosphere will have a negligible effect on the pattern of unstable flow. However, buoyancy forces will dominate when deviatoric stresses in the lithosphere are insufficient to dynamically support the topography arising from unstable deformation. In this formulation, the parameters $S_i$, listed in Table 2, relate the buoyancy forces arising from density contrasts at each interface to the vertically averaged strength of the layers and substrate. The limiting case of a very strong layer in which buoyancy forces have no effect on the characteristics of deformation corresponds to $S = 0$. Large $S$ corresponds to a weak layer in which buoyancy forces play a dominant role in determining the flow.

The parameter $S_i$, which is the ratio of the buoyancy force due to the density contrast at the free surface to the strength of the surface layer, has the greatest effect of the $S$ parameters on the dominant wavelengths. The relationships between the dominant wavelengths and $S_i$ for the multilayered C model are represented in Figure 8. For compression, the effect of a stabilizing density contrast ($\rho_0 < \rho_t$) is to decrease the longer dominant wavelength for a range of $S_i$, while for extension the opposite holds. For compression, the longer dominant wavelength is independent of the influences of buoyancy and strength for $S_i < 0.1$, then decreases rapidly in the range $0.1 < S_i < 10$. For extension, $\lambda_d/h$ markedly increases for $0.1 < S_i < 10$ and is relatively independent of this parameter otherwise.

Figure 8 exemplifies the effects of the coupling of the strong layers by illustrating how the longer dominant wavelength, which arises due to the presence of the strong mantle layer, is affected by the physical properties of the near-surface. In contrast, the shorter dominant wavelength, which is to the greatest extent controlled by the surface layer, is not significantly affected by variations in $S_i$.

If, as on the earth, the density contrast at the crust-mantle boundary of Venus is stabilizing, then for compression and extension increasing $S_i$ will decrease and increase, respectively, the longer dominant wavelength. However, over a broad range of $S_i$ values the changes in the longer dominant wavelengths are not significant. As density variations within the crust and mantle would be expected to be less than that at the Moho, the effects of variations of $S_2$ and $S_4$ should be even less than those of $S_3$. Thus the wavelengths of tectonic features observed at the surface cannot provide meaningful constraints on the subsurface density structure.

The dominant wavelengths and growth rates can also be considered in terms of the internal strength stratification of the lithosphere. The parameters $R_i$, which are defined in Table 2,
represents the strengths of the subsurface layers and substrate in terms of the strength of the surface layer. In the C model the strength in the mantle substrate varies with depth. For this model the substrate rheology is described by $\alpha$, the ratio of the decay depth of the substrate strength to the thickness of the surface layer. Zuber et al. [1986] and Zuber and Parmentier [1986] have investigated the influence of these parameters on the unstable flow in layered media. These studies showed that if the medium has a plastic surface layer and is nonlinear viscous otherwise, changes in the internal strength in most cases do not significantly affect the dominant wavelengths. Therefore uncertainties in the strength stratification do not markedly influence estimations of the dominant wavelengths. However, since instability growth is driven by differences in strength across interfaces in the medium, the growth rate factors do depend on the strength stratification. The effect of increasing the strength contrast at an interface is to increase the growth rate factor and vice-versa. In other words, increasing the relative strength of a competent layer with respect to an incompetent layer increases the degree of instability. For models that exhibit two dominant wavelengths, increasing the relative strength of the strong subsurface layer with respect to the strong surface layer results in an enhanced longer wavelength of instability. Similarly, increasing the relative strength of the strong surface layer with respect to the strong subsurface layer increases the growth rate of the shorter wavelength of instability. If the relative strength of the surface layer is more than a few times greater than that of the strong subsurface layer, the longer wavelength of instability can be suppressed. The longer wavelength can also be suppressed if strength contrasts elsewhere in the lithosphere are small because of the coupling of strong layers described previously. In contrast, if the relative strength of the mantle layer is many times greater than that of the surface layer, both wavelengths of instability will still occur. The shorter wavelength is not suppressed in this case because the growth rate factor varies directly with the power-law exponents of the strong layers. The plastic surface layer can be unstable even if it is markedly weaker than the strong subsurface layer, as long as both are stronger than the intermediate layer and substrate.

**Compressional and Extensional Growth Rates**

The model lithosphere in Figure 7 is more unstable in compression than extension, as evidenced by the fact that the peak in the growth rate function of the former is greater than that of the latter. This is a general characteristic of hydrodynamic instability growth [Smith, 1975, 1977]. For the range of cases examined in this study, a model lithosphere with at least one strong layer is always unstable (i.e., always exhibits at least one dominant wavelength) in compression, while a model lithosphere in extension can be stable for large $S_i$ and/or small internal strength contrasts. Extensional instability can also be suppressed if the surface layer is characterized by a viscous ($1 \leq n_v \leq 3$) rather than plastic ($n_v = \infty$) rheology. The predominance of viscous behavior near the surface would be expected for a lithosphere with a sufficiently high thermal gradient.

Whether a theoretically unstable medium is unstable in practice depends on the dominant growth rate, the amplitude of an initial perturbation, $\Delta(k,0)$, and the mean horizontal strain, $\bar{\varepsilon}_{x,y}$. As illustrated by equations (3) and (4), if any of the above are sufficiently small, the amplification of the instability will be negligible.

**Application to Venus Lithosphere**

The previous section illustrates how various model parameters affect the dominant wavelengths and growth rates and describes in general terms how unstable deformation can occur for a range of conditions in an extending or compressing density- and strength-stratified medium. By considering the results in the context of the wavelengths and characteristic widths of observed surface features on Venus, we can place broad constraints on the structure of the lithosphere.

**Surface Layer Thickness and Strength**

The widths and spacings of tectonic features summarized in Table 1 and the relationships between $\lambda_d/h$ and $S_i$ in Figure 8 allow constraints on the thickness of the strong surface layer. In intervals where $\lambda_d/h$ varies markedly with $S_i$, the thickness of this layer is constrained by its strength. This criterion is met for a range of $S_i$ for the longer wavelength of instability.

To estimate the thickness of the surface layer from the longer wavelength features, first observe from Table 1 that these features have spacings or widths of approximately 300 and 150 km, respectively, for compression and extension. For the conditions assumed in Figure 4 it can be seen that the strength of the upper crust does not exceed a few hundred MPa. Compressional and extensional strengths of 200 and 100 MPa, respectively, are assumed, bearing in mind the considerable uncertainties associated with these values. If parameter values in the numerator of $S_i$ of $\rho_i = 3.0$ g cm$^{-3}$, $\rho_o = 0.0$ g cm$^{-3}$, and $g = 887$ cm s$^{-2}$ are appropriate for Venus, then the compressional and extensional long wavelength relationships imply surface layer thickness estimates of about 16 and 8 km, respectively.

It is also possible to estimate $h_i$ from the spacings of the shorter wavelength structures; however, because the dominant wavelength does not vary significantly with $S_i$ in either compression or extension, the layer strength provides no constraint. The ratio $\lambda_d/h$ for both compression and extension falls in the range 2.7-4, which for a feature with a 10-km spacing yields an upper crustal layer thickness in the range 2.5 < $h_i$ < 3.7 km. For a 20-km spacing, 5 < $h_i$ < 7.4 km. The best agreement between the short and long wavelength results occurs if for compression $S_i < 0.5$ and for extension $S_i > 1$, but the constraints provided by the layer strength are violated. Given the uncertainties in the strength, however, a broader range of stresses than considered above cannot be ruled out. For a wide range of $R_o$, $S_o$, and $\alpha$, the models predict $h_i$ to be between approximately 2 and 22 km. The upper part of this range, which corresponds to the limit of large $S_i$ for the longer wavelength compressional features, is inconsistent with the range determined from the shorter wavelength spacings. More refined estimates of $h_i$ will require better knowledge of the flow behavior of the crust and mantle, the regional strain rates and geothermal gradients, and more realistic models of the rheological structure of the lithosphere. For the broad range of parameters examined in this study, most of the range of $h_i$ is consistent with earlier results of Solomon and Head [1984]. In their models, which did not consider a strong subsurface layer, this parameter was found to be between 1 and 10 km.

Figure 8 illustrates another interesting outcome of the models. For the shorter wavelength instability, $\lambda_d/h$ is approximately the same for compression and extension over the range of $S_i$. Thus for similar surface layer thicknesses, the dominant
wavelengths for these extensional and compressional instabilities should be comparable. From Table 1 it can be seen that the short wavelength spacings of features interpreted as extensional and compressional fall in the same range. Alternatively, for $S_c < 1$ and a given surface layer thickness, the longer dominant wavelength for compression is considerably greater than that for extension. On Venus the ridge belts, which are thought to have formed in compression, have a spacing of approximately 300 km. In contrast, rifts, which formed in extension, have widths that do not exceed 200 km. If the thicknesses of the strong surface layers in regions in which these features are present are not significantly different, then the predicted dominant wavelengths for unstable extension and compression shown in Figure 8 may explain the differences in scales of the longer wavelength features and the similarities in the spacings of the shorter wavelength features. However, on the basis of the models alone, it is not possible to distinguish between an extensional and compressional origin for surface features that exhibit only the shorter wavelength spacing.

Mantle Rheology

For the arbitrarily chosen crustal thickness shown in Figure 4, deformation in the mantle just below the base of the crust occurs in a brittle manner. However, if the crust were sufficiently thick, deformation everywhere in the mantle would be dominated by ductile flow. For the parameter values assumed in Figure 4, this would occur for crustal thicknesses greater than 17 and 11 km for extension and compression, respectively. Previous studies have shown that the growth rates of extensional and compressional instabilities vary directly with the stress exponents of the strong layers [Smith, 1977; Fletcher and Hallet, 1983; Zuber et al., 1986]. Therefore a model lithosphere with a plastic (brittle) mantle layer is more unstable than that with a power-law viscous layer. The results shown thus far have been calculated assuming a nonlinear viscous ($n_1 = 3$) mantle layer; substitution of a plastic layer would slightly decrease the dominant wavelength of the longer wavelength of instability and increase the dominant growth rates of both long and short wavelength instabilities. The implications of this are that the lithosphere would be even more unstable than suggested by the results up to this point. To produce multiple wavelengths of lithospheric deformation in the present model, the upper crust and upper mantle must be stronger than the lower crust, but the dominant style of deformation is the strong part of the mantle can be either brittle or ductile. For the strain rate and compositions assumed in Figure 4, thermal gradients up to approximately 25 K km$^{-1}$ will accommodate a strong mantle region of sufficient thickness (only a few kilometers) to permit the development of a long wavelength of instability. This range of thermal gradients is consistent with estimates derived from thermal models that assume both purely conductive [Solomon and Head, 1982] and hot spot [Morgan and Phillips, 1983] heat loss on Venus.

The style of deformation in the Venus upper mantle as suggested by the geometries of tectonic features has also been addressed by Banerdt and Golombek [1986], who applied the wedge subsidence model of Vening-Meinesz [1950] to explain the widths of rifts. In their study, extension of a model Venus lithosphere containing a brittle mantle layer results in the formation of a simple graben in the mantle. The graben is defined by normal faults that bound a down-dropped section of the mantle layer; the spacing of these faults controls the width of the rift. In terrestrial rifts where good quality gravity, seismic, and heat flow data exist, evidence for a coherent down-dropped mantle block is not apparent [cf. Ramberg and Morgan, 1984]. If through-going normal faults as implied by the Vening-Meinesz model exist in the Venus mantle and control the widths of rift zones, then subsurface deformation associated with rifts on this planet must be fundamentally different than that characteristic of the best-studied terrestrial rift zones.

Crustal Rheology and Thickness

In the Results section, we noted that in a multilayered medium the longer wavelength of instability can be suppressed if the mantle layer is weaker than the upper crustal layer. If tectonic features that exhibit two dominant wavelengths formed in response to unstable deformation, then their presence in a given region implies that the upper crust in that area is either weaker or at least not significantly stronger than the upper mantle. This is consistent with the depth distribution of lithospheric strength shown in Figure 4. Unfortunately, this does not better constrain the composition of the Venus crust, as a range of plausible crustal materials satisfy this requirement.

The existence of two scales of deformation requires strong crustal and mantle layers that are separated by a weaker lower crust. For an upper limit of the thermal gradient of 25 K km$^{-1}$, $t_{sl} = 10^{15.5}$ s, and a diabase composition, the minimum crustal thickness for which a weak lower crust occurs is about 5 km. A lower strain rate or a dominant crustal mineral with a weaker ductile strength (e.g., feldspar) would permit a somewhat, though not significantly, smaller value. An approximate upper limit of the crustal thickness in a region that contains two wavelengths of deformation is 30 km, given an estimated lower limit for the thermal gradient of 10 K km$^{-1}$ and the conditions assumed above. For a thicker crust the lithosphere does not contain a region of upper mantle strength. Variations in thermal gradient, strain rate, and composition would alter this value in a manner similar to that described above. For example, for $dT/dz = 25$ K km$^{-1}$, the maximum crustal thickness is about 15 km. The overall range of crustal thicknesses in regions that exhibit two scales of deformation as determined from the present models is less than the thickness predicted on the basis of petrological arguments regarding the depth of partial melting in the Venus mantle [Anderson, 1980], but is in general agreement with estimates from thermal [Morgan and Phillips, 1983] and lithospheric stress [Banerdt, 1986] models.

The existence of a single, shorter wavelength of deformation in a region can be explained either by a lithosphere in which the crust is thick enough that the underlying mantle is too weak to allow the growth of a long wavelength instability, or a lithosphere that contains a strong upper mantle but in which the longer wavelength instability has been suppressed. The latter is possible if strength contrasts in the lithosphere are small. Of the tectonic features discussed in this study, only the banded terrain in Ishtar Terra exhibits a single, short wavelength of deformation; however, in radar maps of the Venus northern hemisphere produced from the Venera 15/16 data [Barsukov et al., 1986; Basilevsky et al., 1986], many ridge-and-groove patterns that display only short wavelength spacings are apparent. Morgan and Phillips [1983] determined that a crustal thickness of up to 60 km is required in Ishtar Terra on the basis of isostatic compensation models, but noted that in other
areas a thick crust is not required by the gravity data. A thick crust in Ishtar has also been suggested by Banerdt and Golombek [1986] in an application of the elastic-plastic buckling model of McAdoo and Sandwell [1985] to the banded terrain. In other regions of Venus that exhibit only the shorter wavelength of deformation, the models described in this study suggest that details of the lithospheric strength stratification can explain the absence of longer wavelength tectonic features. An underlying thick crust cannot be ruled out, but is not required.

CONCLUSIONS

We have invoked unstable deformation in a density- and strength-stratified lithosphere to explain the characteristic widths and spacings of many tectonic features on Venus. For compression and extension we have shown that two wavelengths of deformation can occur if the lithosphere contains two strong brittle or ductile layers, one at the surface and the other at depth, that are separated by a weaker ductile layer. We interpret the strong layers as the upper crust and upper mantle, and the intermediate layer as the lower crust in the Venus lithosphere. The upper crustal layer primarily controls the geometries of features with spacings in the range 10–20 km such as the banded and ridge-and-groove terrains, while the strong mantle region primarily controls the geometries of features with widths or spacings of 100–300 km, such as rift zones and ridge belts.

The models predict that the smaller scale compressional and extensional features should exhibit comparable wavelengths. This is consistent with the observation that surface features in various areas of the planet have been interpreted as having either compressional or extensional origins exhibit approximately the same range of smaller scale spacings. In contrast, for a given surface layer thickness, the dominant wavelength of the longer wavelength instability for compression may be considerably greater than that for extension. This may explain why ridge belts, which most likely formed in compression, have a length scale significantly greater than rifts, which formed in extension.

The lithosphere is always unstable in compression, but can be stable in extension if strength contrasts within the lithosphere are sufficiently small, or if deformation of both of the strong layers occurs in a ductile rather than brittle manner. Whether an instability actually results in deformation depends on the growth rate, which is controlled primarily by the strength stratification of the lithosphere and the power-law exponents that characterize the strong layers.

In areas of Venus where longer and shorter wavelengths of deformation are present, model results imply crustal thicknesses in the approximate range 5–30 km and a thermal gradient that was not greater than 25 K km⁻¹ at the time of deformation. In regions where only the smaller scale of deformation is observed, the results suggest either that there is a lithosphere with a relatively thick crust and no underlying region of upper mantle strength, or that the longer wavelength instability has been suppressed. The latter can occur if strength contrasts within the lithosphere are small.

The surface of Venus contains numerous other features with characteristics similar to those described in this study. Deformation is observed to occur in localized linear belts and over broad regions, but in both cases is believed to be the product of large-scale horizontal stresses [Basilievsky et al., 1986]. The origin of the stresses remains a matter of debate, and must be reconciled if the dominant mechanism of global heat transport and the style(s) of tectonics are to be discerned [e.g., Kaula and Phillips, 1981; Head et al., 1981]. Further understanding of the lithospheric structure of Venus through models such as those presented in this paper and future high resolution radar imaging of surface features will be required to help achieve these goals.

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