REVISITING THE ORIGIN OF TECTONIC SPACING ON VENUS: IMPORTANCE OF LOCALIZATION AND SURFACE TEMPERATURE. Laurent G. J. Montézi1 and Maria T. Zuber2, 1Woods Hole Oceanographic Institution, Department of Geology and Geophysics, Clark 241B, MS#22, Woods Hole MA 02543, montesi@whoi.edu, 2Massachusetts Institute of Technology, Cambridge, MA 02139, zuber@mit.edu.

Introduction: Since surface images of Venus have been available, it has been recognized that tectonic deformation often organizes with a characteristic lengthscale [1]. Several studies have linked the recognized tectonic wavelengths to the mechanical structure of the lithosphere, using a variety of models of large-scale tectonics [2-5]. The depth to the brittle-ductile transition (BDT) is commonly inferred from these models. However, most of this work was conducted before it was realized how much stronger dry diabase (thought to be a analogue of the venusian crust) is, compared to previously available flow laws [6]. In addition, recent theories that incorporate the dynamic process of localization, or formation of faults, have produced a new generation of models that can explain regularly-spaced faults in planetary lithosphere [7]. Hence, in this study, we re-evaluate models to explain the observed tectonic patterns.

Models for regularly-spaced tectonics:

Elastic folding. An elastic layer of thickness \( h \) develops folds if it is subjected to horizontal compression. The fold wavelength is roughly \( \lambda_E=2\pi(30 h^5)^{1/4} \), where the constant depends on the elastic properties of the lithosphere [3]. Although the stress required to fold a thick elastic lithospheric core exceed its strength, folding of thinner layers is possible. These layers can be bounded by intracrustal discontinuities or represent only an elastic core inside a stressed lithosphere.

Plastic buckling and necking. If the lithosphere is subjected to horizontal compression or extension, it may become unstable and develop regular folds (buckles) or necks [2,5]. The wavelength of this instability, \( \lambda_B \), is proportional to \( H \), the thickness of a strong layer, usually identified with the brittle upper crust of upper mantle. In compression we have \( 4H<\lambda_B<8H \), whereas in extension, we have \( 2H<\lambda_B<4H \). However, the buoyancy of the lithosphere and the depth-dependent strength profile reduce the growth of this instability.

Localization instability. The folding a buckling-necking instabilities predict broad folds in which deformation is distributed. The fold wavelength can control fault spacing if folds reach high-enough amplitude before the faults form. Alternatively, faults can form a pattern with a characteristic wavelength \( \lambda_L \) if the lithosphere weakens with deformation [7]. This process, the localization instability, is most likely to occur in the brittle upper crust and/or upper mantle, and \( \lambda_L \) is proportional to \( H \), the brittle layer thickness. The coefficient of proportionality depends on the effective stress exponent, \( n_e \), but for \(-50<n_e<-25\), as may be relevant for planetary lithospheres [7], we obtain \( 0.5H<\lambda_L<2H \).

Cracking. Joints and cracks may display a regular spacing \( \lambda_c \). Models of shear lag or stress shadow predict \( \lambda_c=H \), the thickness of the layer undergoing cracking [4,9]. This layer may be limited by the stratigraphy of the crust, or by the depth of the BDT. Cracks are different from the shear zones considered in the localization instability theory because they represent dilation rather than shearing of the rocks.

Figure 1: Depth to the brittle-ductile transition predicted for a dry diabase rheology as a function of the surface temperature and geotherm. The thick lines indicate the condition for which the solidus of dry gabbro ([10], red line), dry MORB basalt ([11], blue line) and alkali basalt ([12], green line) is reached at 30 km depth, representing the average thickness of the crust [13].

Brittle-ductile transition: The depth to the BDT may control the length scale of instability. We show in Figure 1 the BDT of the crust, using the flow law of dry diabase [6] with a strain rate of \( 10^{15} \text{ s}^{-1} \). We assume that the shape of the geotherm is an error function with asymptotic temperature of 1350°C, analogue to the Earth’s adiabatic mantle. However, we consider a wide range of geotherms because of the uncertain thermal history of the venusian lithosphere. We vary the surface temperature from 400 to 1000°C to reflect the possibility of higher surface temperature due to a climate/volcanism feedback [5]. However, the range of allowable thermal model is limited by melting in the
lower crust. We use the basalt solidus of [10] for dry gabbro, [11] for MORB (possible analogue for the Venera 9, 10, 12, and Vega 1 and 2 sites) and [12] for alkali basalt (possible analogue for the Venera 8 and 13 sites).

Origins of tectonic wavelengths: Long-wavelength deformation. Ridge belts often display a regular spacing, in excess of 100 km [1-3]. The width of rifts is at the same length scale. It has been proposed that this scale of deformation represents elastic or plastic buckling of the strong upper mantle [2, 3]. The BDT for this instability must in around 20 to 50 km, which requires a small geotherm and low surface temperature. Even if the controlling thickness is the BDT of the mantle, the crust must be mostly brittle, which indicate geotherms less than 5 K km$^{-1}$.

Figure 2: Summary of wavelengths predicted for the localization instability with a surface temperature of 470°C. The base models (black and green) consider only the crustal rheology and localization occuring to the brittle-ductile transition depth. Shorter wavelengths can be achieved if localization is instead limited by a given isotherm (magenta). If the crust has a finite thickness, the instability wavelength is limited upwards by the point at which the whole crust is brittle and the layer controlling the instability become the upper mantle (cyan).

Intermediate-wavelength deformation. Tectonic wavelengths of the order of 10 to 20 km were recognized in pre-Magellan data as ridges in ridge belts and individual faults and grabens in rift zones [2, 3]. Magellan data added in this category the distributed wrinkle ridges and ridges in tesserae [1]. These wavelengths were interpreted as the manifestation of folding of a brittle crust that is less than 5 km thick. With a dry rheology, this requires geotherms in excess of 30 K km$^{-1}$ or a high surface temperature (Figure 1). Alternatively, the localization instability can produce wavelengths of the right order for a wide range of geotherms (Figure 2). The localization instability may be favored as it treats directly the formation of faults, which are a manifestation of this wavelength.

Short-wavelength deformation. Magellan revealed a wealth of short-wavelength features, with spacing less than 1 km, and often of the order 100 m or less [1]. These include the pervasive lineations in the plains, and some tessera deformation such as ribbon terrain [14]. Buckling cannot explain the observed wavelength. Cracking, which was originally proposed to explain the spacing of plain lineations [8] and the localization instability have predicted wavelengths of order of $H$. Hence, the implied BDT is shallower than expected for a dry rheology with the current surface temperature. The very short instability wavelength is possible if the surface temperature was high at the time of formation on these features. However, crustal melting is a concern for geotherms higher than 10 K km$^{-1}$. The short wavelength deformation might indicate intracrustal layering rather than the brittle-ductile transition. Alternatively, some of these features might be a near-surface effect of dike intrusion, similar to radial graben observed in the plains [15].