Tessera is characterized by troughs, which define ~110–370-km-long and ~15–50-km-wide east-west-oriented segments of CRT. These elongated areas have some of the highest topography in the Meni Tessera. One of the segments forms the previously mentioned eastern extension of the CRT ("G" in Fig. 1). Although these elongated parts of the Meni Tessera do not have a similar strike as the ridgelike typography and morphology, and they are not composed of clear individual ridges, they do have a strike similar to the ridgelike components in northeastern Tellus Regio.

Structurally Meni Tessera CRT is more complex than northeastern Tellus Regio. The oldest underlying structures are curvilinear ridges, but they are wider and shorter (2–8 km wide and 10–30 km long) and more widely spaced than in northeastern Tellus. There does appear to be very fine ridgelike structures superposed on these ridges, but they can not always be distinguished from scarps and normal faults. The dominant direction of the ridges is to be north-south/northeast-southwest, but this orientation may be due to more later deformation and the original directions may not be observable any more. Near the eastern edge of central Meni Tessera the ridges follow the curving tessera border. The central parts of Meni Tessera are characterized by areas of orthogonal terrain of intersecting northeast-southwest and northwest-southeast graben ("H" in Fig. 1). This terrain has been partly covered by lavas of intratessera plains in the areas where it is visible. There are also places where graben cut across the border between the CRT and the plain, especially around western and northern edges of Meni Tessera.

The relationships between graben with different orientations is complex: North-south striking graben and individual scarps (probably normal faults) cut other features extensively in the eastern and northern parts of the central Meni Tessera. There are also graben oriented in the northeast-southwest direction, especially near the northwestern and southeastern borders, which cut ridges and northwest-southeast graben. In the central parts of the CRT there are northeast-southwest-oriented graben that cut other features, but these graben are frequently covered by lavas. There are also small areas where graben are not widespread or at least can not be distinguished from small-scale ridges or closely spaced faults. Deformation seems to have followed the same kind of basic sequence as in northeastern Tellus Regio except that there have been several different episodes of graben formation with both spatially and chronologically more complex relationships. Also, differences in orientation and morphology of ridges in Meni Tessera and northeastern Tellus Regio may reflect different original stress regimes. Although no major strike-slip faults were identified in Meni Tessera, there is evidence of probable shear deformation in nearby plains areas.

Discussion and Conclusions: Similarities in the topographic trends, especially the similar types of linear ridgelike features in northeastern Tellus Regio and corresponding elongated segments in northern Meni Tessera, which together form a roughly south-concave arc of topographically higher CRT, as well as some similarities with structures of the CRT of the easternmost Meni Tessera and western edge of northeastern Tellus Regio, indicate that these areas of CRT were probably earlier interconnected. The troughlike plain area between Meni Tessera and Tellus Regio is probably underlain by CRT, which has been disrupted and covered by lavas. The adjacent northern Leda Planitia is deformed by complex intersecting systems of fractures and ridges. Some of this deformation may reflect a presence of a covered basement of CRT. The arclike pattern of tesserae between Kamari Dorsa and northeastern Tellus Regio may also reflect an earlier larger area of tessera. Similar conclusions were earlier presented on the basis of analysis of Venera data [14,15] and more recently by a comprehensive analysis of distribution and characteristics of tessera from Magellan images [16]. Based on this work, however, it is very hard to define exactly the original extent of the CRT in this region.

Tessera are proposed to form by hot-spot-related volcanism and tectonism [17,18] or by convection-driven tectonics above mantle downwellings [1,12,19]. The results of this work do not conclusively rule out either model. Analysis of structures and deformation shows that the earliest distinguishable deformation was compression, which was followed by widespread extension and volcanism (formation of intratessera plains). This result is in agreement with other studies [e.g., 1,3,11,13] and similar results have been used to support the mantle downwelling model [1,12], but in our opinion they do not leave out other possibilities.

The arclike arrangement of topographically higher ridgelike features in northeastern Tellus Regio and northern Meni Tessera is roughly similar in planform, but smaller than the Dekla Tessera–northeastern Tellus Regio arch in the north. These arcuate patterns of tessera are typical to the area between longitudes 0° and 150°E [16] and could tell us about the scales of deformation of the crust in these areas. Observed complex deformational sequences in the northeastern Tellus Regio–Meni Tessera region do support the idea that the CRT is probably a result of repeated deformation through different mechanisms [20]. We are currently analyzing in more detail structures in Meni Tessera and northern Tellus Regio and their relationships with topography, intratessera volcanism, and the deformation and volcanism on the adjacent plains.

References: [1] Bindschadler D. L. et al. (1992) *JGR*, submitted. [2] Barsukov V. L. et al. (1986) *Proc. LPSC 17th*, in *JGR*, *91*, D378–D398. [3] Bindschadler D. L. and Head J. W. (1991) *JGR*, *96*, 5889–5907. [4] Smrekar S. and Phillips R. J. (1990) *LPSC XXI*, 1176–1177. [5] Smrekar S. and Phillips R. J. (1992) *EPSL*, *107*, 582–597. [6] Williams D. R. and Gaddis L. (1991) *JGR*, *96*, 18841–18859. [7] Baker V. R. et al. (1992) *JGR*, submitted. [8] Komatsu G. and Baker V. R. (1992) *LPSC XXIII*, 715–716. [9] Gulick V. C. et al. (1992) *LPSC XXIII*, 465–466. [10] Komatsu G. et al. (1992) *LPSC XXIII*, 719–720. [11] Bindschadler D. L. and Tatsumura M. J. (1992) *LPSC XXIII*, 103–104. [12] Bindschadler D. L. et al. (1992) *JGR*, submitted. [13] DeCharon A. V. and Stofan E. R. (1992) *LPSC XXIII*, 289–290. [14] Sukhanov A. L. (1986) *Geotectonics*, *20*, 294–305. [15] Nikishin A. M. (1990) *Earth Moon Planets*, *50/51*, 101–125. [16] Ivanov M. A. et al. (1992) *LPSC XXIII*, 581–581. [17] Herrick R. R. and Phillips R. J. (1990) *GRL*, *17*, 2129–2132. [18] Phillips R. J. et al. (1991) *Science*, *252*, 651–658. [19] Bindschadler D. L. and Parmentier E. M. (1990) *JGR*, *95*, 21329–21344. [20] Solomon S. C. et al. (1992) *JGR*, submitted.

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EPISODIC PLATE TECTONICS ON VENUS. Donald Turcotte, Department of Geological Sciences, Cornell University, Ithaca NY 14853, USA.

Studies of impact craters on Venus from the Magellan images have placed important constraints on surface volcanism. Some 840 impact craters have been identified with diameters ranging from 2 to 280 km. Correlations of this impact flux with craters on the Moon, Earth, and Mars indicate a mean surface age of 0.5 ± 0.3 Ga. Another important observation is that 52% of the craters are slightly fractured and only 4.5% are embayed by lava flows. These observations led Schaber et al. [7] to hypothesize that a pervasive resurfacing event occurred about 500 m.y. ago and that relatively little surface volcanism has occurred since. An alternative hypothesis has been proposed by Izenber et al. [3]. They suggested that the observations can be explained by a random distribution of small volcanic events.

Strom et al. [8] have pointed out that a global resurfacing event that ceased about 500 MYBP is consistent with the results given by Arkani-Hamed and Toksoz [1]. These authors carried out a series of numerical calculations of mantle convection in Venus yielding thermal evolution results. Their model 4 considered crustal recycling and gave rapid planetary cooling. They, in fact, suggested that prior to 500 MYBP plate tectonics was active in Venus and since 500 MYBP the lithosphere has stabilized and only hot-spot volcanism has reached the surface. Thus they suggested that the transition to a near pristine lithosphere was the result of the secular cooling of the planet.

In this abstract we propose an alternative hypothesis for the inferred cessation of surface volcanism on Venus. We hypothesize that plate tectonics on Venus is episodic. Periods of rapid plate tectonics result in high rates of subduction that cool the interior resulting in more sluggish mantle convection. With a cool viscous interior the surface lithospheric plate stabilizes and plate tectonics cease. The stable lithosphere thickness increases with time reducing the surface heat flow. As a result the interior temperature increases leading to an increase in the plume flux. Eventually the lithosphere is sufficiently thick and its gravitational instability initiates an episode of global subduction.

This hypothesis is illustrated qualitatively in Fig. 1. During a period of about 500 m.y. the lithosphere is globally stable and no plate tectonics occurs. During this period the lithosphere thickens, the surface heat flow decreases, and the mantle temperature increases. As the lithosphere thickens it becomes gravitationally unstable. Eventually this instability leads to a catastrophic global subduction event. In the model illustrated in Fig. 1 this event occurred 600 MYBP. At the time of the global lithospheric instability the mantle on Venus is expected to be considerably hotter and less viscous than on the Earth so that rapid subduction would occur. Without a lithosphere, vigorous mantle convection would lead to extensive volcanism, vigorous plate tectonics, a high surface heat flow, and a rapid cooling of the mantle. As the mantle cools the mantle convection and its surface manifestation, plate tectonics, become less vigorous. Eventually the global lithosphere stabilizes and plate tectonics ceases. In the model illustrated in Fig. 1 this occurs at 500 MYBP. The lithosphere thickens, the surface heat flow decreases, and the cycle repeats.

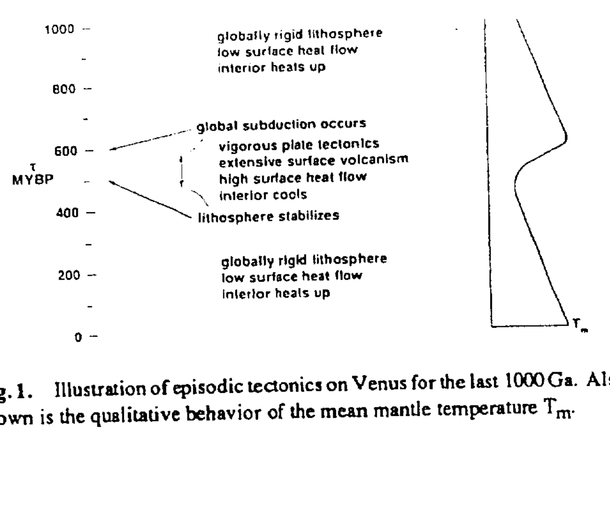

Fig. 1. Illustration of episodic tectonics on Venus for the last 1000 Ga. Also shown is the qualitative behavior of the mean mantle temperature T_m.

Assuming that the venusian lithosphere stabilized 500 MYBP, it is easy to determine its thermal structure, assuming no basal heating from mantle plumes and no partial delamination. After 500 m.y. the depth to the 1475-K isotherm is 290 km, the depth to the 1275-K isotherm is 180 km, and the depth to the 1125-K isotherm is 120 km. The corresponding depths for a venusian lithosphere with steady-state conductive heat transport are 34, 26, and 19 km respectively. The transient cooling of the lithosphere results in much greater thicknesses, almost an order of magnitude. Such a thick lithosphere is consistent with the large observed topographic and gravity anomalies.

McKenzie et al. [4] have argued that the perimeters of several large coronae on Venus, specifically Artemis, Latona, and Eithinoha, resemble terrestrial subduction zones in both platform and topography. Artemis chasma has a radius of curvature similar to that of the South Sandwich subduction zone on the Earth. Sandwell and Schubert [6] have shown that the morphologies of several coronae are in good agreement with the lithosphere flexure models that have been successful in explaining the sea floor morphology at ocean trenches on this planet. Their flexural profiles yield elastic lithosphere thicknesses of 37 km for Artemis, 35 km for Latona, 15 km for Eithinoha, 40 km for Heng-O, and 18 km for Freyja. These values are consistent with a thick conductive lithosphere. The presence of incipient subduction zones may be an indication of the onset of another episode of active plate tectonics.

References: [1] Arkani-Hamed J. and Toksoz M. N. (1984) *PEPI, 34,* 232–250. [2] Ghail R. and Wilson L. (1992) *LPSC XXIII,* 409–410. [3] Izenber N. K. et al. (1992) *LPSC XXIII,* 591–592. [4] McKenzie D. et al. (1992) *JGR,* submitted. [5] Parmentier E. M. and Hess P. C. (1992) *LPSC XXIII,* 1037–1038. [6] Sandwell D. T. and Schubert G. (1992) *LPSC XXIII,* 1209–1210. [7] Schaber G. G. et al. (1992) *LPSC XXIII,* 1213–1214. [8] Strom R. G. et al. (1992) *LPSC XXIII,* 1279–1380.

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SCATTERING PROPERTIES OF VENUS' SURFACE. G. L. Tyler, R. A. Simpson, M. J. Maurer, and E. Holmann, Center for Radar Astronomy, Stanford University, Stanford CA 94305-4055, USA.

Radar backscatter functions $\hat{\sigma}_0(\phi)$ for incidence angles between $0 \leq \phi \leq 4°–10°$ have been derived from Magellan altimetry radar echoes. The procedure includes constrained solution of a system of simultaneous equations for which the echo spectrum and echo time profile are inputs. A practical and workable set of constraints has been applied; optimization and improved results are expected as the analysis matures. The scattering functions yield information on small-scale surface structure (tens of centimeters to tens of meters) but averaged over hundreds of km^2. RMS surface slopes derived from fits of analytic functions to the $\hat{\sigma}_0(\phi)$ results have been converted to map form and show patterns similar to those reported using other techniques. While all three forms are found on Venus, fit residuals imply that an exponential scattering function matches data better than either the Hagfors or gaussian form in most areas, although the Hagfors function may be a better descriptor at some sites. Limited study of image data indicates that average backscatter cross section, and possibly its slope, can be derived at oblique angles ($17° \leq \phi \leq 45°$). Offsets of the echo peak in altimetry spectra are surprisingly common and are loosely correlated with Venus topography, but no cause for this phenomenon has yet been identified.