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Vicki L. Hansen
Brian K. Banks
Rebecca R. Ghent
Department of Geological Sciences, Southern Methodist University, Dallas, Texas 75275, USA

ABSTRACT

Many workers assume that tessera terrain—marked by multiple tectonic lineaments and exposed in crustal plateaus—comprises a global "onion skin" on Venus. A growing body of structural, mechanical, magmatic, gravitational-topographic, and geologic evidence indicates that tesserae record the local interaction of individual deep-mantle plumes with an ancient, globally thin Venusian lithosphere, resulting in local regions of thickened crust.

INTRODUCTION

Tessera terrain constitutes the oldest known crust on Venus. Tesserae are characterized by intersecting sets of tectonic lineaments, high relief relative to the surroundings, and high surface roughness (Basilevsky et al., 1986), and compose 8%–10% of Venus's surface. Many workers assume that tessera terrain forms a global layer that extends almost everywhere beneath the regional plains (e.g., Solomon, 1993; Turcotte, 1993; Grimm, 1994; Herrick, 1994; Ivanov and Head, 1996; Nimmo and McKenzie, 1998; Head and Basilevsky, 1998). A global "onion skin" of tesserae would facilitate global stratigraphic studies and provide important constraints for geodynamic models; however, such an assumption may be grossly unwarranted. Crustal plateaus (Fortuna and Tellus Tesserae and Alpha, eastern Ovda, western Ovda, Phoebe, and Thetis Regiones)—steep-sided, flat-topped, quasicircular regions ~1600–2500 km in diameter—host most tesserae (Fig. 1). Scientists agree that thickened crust supports crustal plateaus, as evidenced by small gravity anomalies, low gravity to topography ratios, shallow apparent depths of compensation, and consistent admittance spectra (Smrekar and Phillips, 1991; Bindschadler et al., 1992; Grimm, 1994; Simons et al., 1997), but disagree about the thickening mechanism. Two popular models are debated: (1) thickening by subsolidus flow and horizontal lithospheric accretion associated with mantle downwelling and (2) thickening by magmatic underplating and vertical accretion associated with mantle upwelling (Phillips and Hansen, 1994; Bindschadler, 1995; Hansen et al., 1997). Because tesserae are exposed mostly within crustal plateaus, issues of tessera distribution and the mechanism of crustal-plateau formation are intimately related through the question: Did tesserae formation predate, postdate, or accompany crustal thickening associated with crustal-plateau formation? The tectonic and magmatic history of tessera across crustal plateaus provides critical constraints for these questions. Spatial correlation of tectonic patterns with thick crust favors local tesserae formation. Strain patterns within tesserae provide constraints on the crustal-thickening mechanism. Geodynamic models of crustal thickening by downwelling predict strong surface contraction (Bindschadler and Parmentier, 1990), whereas upwelling-plume models predict initial surface extension (Phillips, 1990; Bindschadler et al., 1992). We discuss a growing body of evidence that tesserae record local tectonism and magmatism above individual mantle hotspots.

GEOLOGIC RELATIONSHIPS

Tessera terrain was originally defined by using Soviet Venera radar data; high-resolution (~100 m/pixel) Magellan SAR (synthetic aperture radar) data allow better delineation of tessera terrains and better characterization of discrete tectonic elements within tesserae. The SAR data allow geologic, structural, kinematic, and mechanical analyses not previously possible, and these data are well suited to unravel plateau-strain histories. Each crustal plateau except Phoebe hosts at least three distinct tectonic ele-

ments: ribbons (extensional troughs, described in the following), folds, and complex grabens (Figs. 2 and 3; Hansen and Willis, 1996, 1998). Ribbons occur across crustal plateaus, and fold and graben relationships define two structural domains (Fig. 4): (1) the marginal-fold domain and (2) the interior basin and dome domain dominated by either folds or grabens.

Ribbons (Hansen and Willis, 1996) form a distinctive tectonic fabric composed of alternating long (~50–100 km), narrow (~1–3 km), shallow (<0.5 km) extensional troughs (e.g., steep troughs of Bindschadler et al., 1992; narrow troughs of Head, 1995) separated by ridges (~1–5 km wide). Two types of ribbons are recognized; the more common shear-fracture ribbons have parallel trough-wall terminations indicating a graben-like geometry, whereas tensile-fracture ribbon troughs have V-shaped terminations reflecting a tensile-fracture origin (Hansen and Willis, 1998). Complex grabens also record extension; they typically cut fold crests in a lens-shaped map pattern, indicating that they formed after the folds. Despite common extensional origin, ribbons and grabens differ morphologically (Figs. 2 and 3; Hansen and Willis, 1996, 1998; Ghent and Hansen, 1999). Ribbons describe a relatively simple periodic ~3–5-km-wavelength fabric with a large length:width aspect ratio; opposite trough walls remain generally parallel as ribbon troughs track across several folds. In contrast, complex grabens are wider, shorter, and deeper than ribbon troughs and have walls defined by multiple synthetically and antithetically dipping normal-fault scarps. Ribbon ridge-trough morphology formed by extension of a thin, brittle layer above a weaker, ductile substrate, analogous to a chocolate layer over caramel and to grabens in Utah's Canyonlands (McGill and Stromquist, 1979). Wavelength instability and Mohr-Griffith fracture analyses constrain brittle-layer thickness and thus the depth to the brittle-ductile tran-
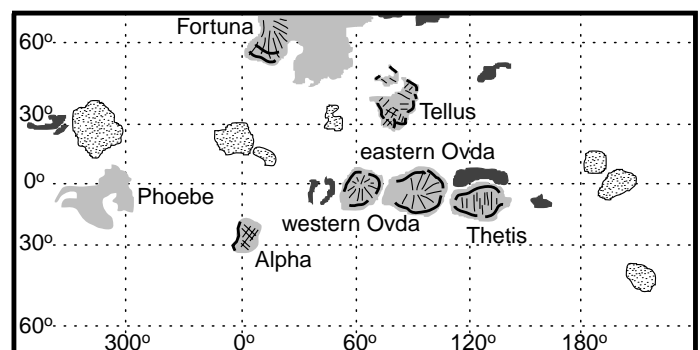

Figure 1. Distribution of various features on Venus. Patterns: fine dots—volcanic rises, dark areas—large tessera inliers, light gray areas—crustal plateaus with fold (thick lines) and ribbon (thin lines) trends (trends from Bindschadler et al., 1992; Pritchard et al., 1997; Ghent and Hansen, 1999).
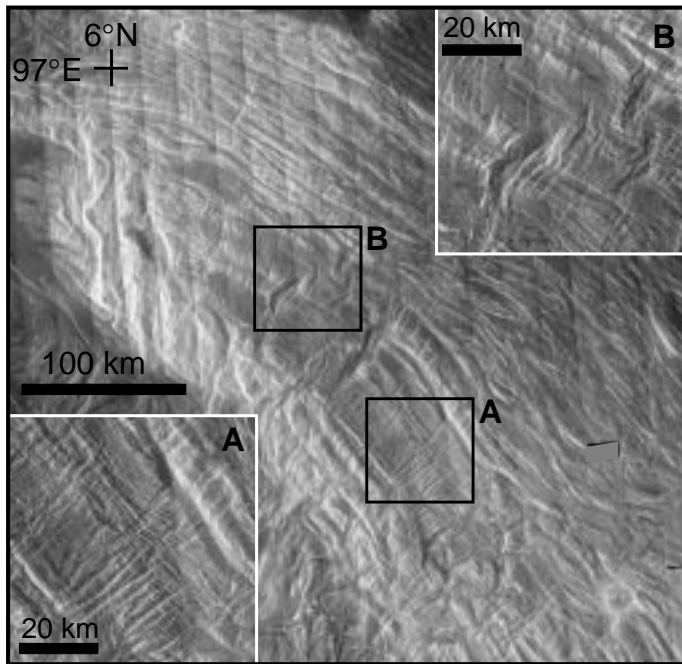
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Figure 2. Synthetic aperture radar image of northeastern Ovda showing northwest-trending folds cut by northeast-trending (A) ribbons and (B) grabens.

sition: during ribbon formation the transition was at ~1 km depth (Hansen and Willis, 1998; Ghent and Hansen, 1999).

Long-wavelength folds (15–30 km) include (1) upright, gentle cylindrical folds as long as 400 km with axes generally parallel to and coincident with plateau margins and (2) short (~50 km) upright, gentle folds with varying orientation, preserved within plateau interiors (Fig. 4; Ghent and Hansen, 1999). These folds reflect minor (<<10%) shortening and formed with a brittle-ductile transition at ~6 km depth (Brown and Grimm, 1997; Ghent and Hansen, 1999). Lineaments at the limits of SAR resolution parallel fold ridges. Lineament morphology is uncertain and may not represent the same structure everywhere; some lineaments likely reflect minor fold crests with <1–2 km wavelength. In eastern Ovda Regio, Ghent and Hansen (1999) postulated synchronous formation of small- and long-wavelength folds on the basis of their parallel trends. However, geologic relationships in Tellus Tessera indicate that small-wavelength folds locally predate uplift associated with long-wavelength folds (Fig. 5; Banks and Hansen, 1999). Complex grabens widen across and cut fold crests and thus postdate folds (Figs. 2 and 3; Bindschadler et al., 1992; Head, 1995; Hansen and Willis, 1996, 1998; Ghent and Hansen, 1999).

Mechanical considerations constrain ribbon-fold timing. Because ribbons require a shallower brittle-ductile transition than long-wavelength folds, ribbons likely formed first in areas hosting both ribbons and folds (Hansen and Willis, 1998). If ribbons postdated long-wavelength folds, the shallow transition required for ribbon formation would result in relaxation of earlier folds. However, if ribbons formed first with a shallow brittle-ductile transition, broad folds could form later as the transition deepened, thus preserving geometric characteristics of both ribbons and folds. Regardless of timing, the mechanics of ribbon and fold formation require a shallow brittle-ductile transition across crustal plateaus.

Volcanism and Stratigraphic Constraints

Volcanic relationships constrain local timing among tessera tectonic elements. Evidence for crustal-plateau volcanism is most apparent in intratessera basins (Figs. 4 and 5). Throughout crustal plateaus are irregularly shaped embayment basins, characterized by digitate margins resulting from

passive flooding of low-viscosity lava into tectonic topographic lows; quasi-circular to elongate, fault-bounded structural basins occur preferentially in marginal-fold domains (Banks and Hansen, 1998). Many intratessera basins exhibit characteristics of both embayment and structural basins. Late flood lava commonly postdated basin deformation, as evidenced by the smooth character of basin fill. Because two-dimensional SAR images only record the youngest surface flows, it is difficult to decipher evidence of earlier volcanism. However, detailed examination of an embayment basin perched along the crest of a broad anticline in Tellus indicates that flood lava postdated orthogonal ribbons and minor folds, but predated long-wavelength folds (Fig. 5; Banks and Hansen, 1999). The crest of the host anticline widens around the basin, apparently the result of strain partitioning during folding, analogous to tectonic fabric wrapping around a porphyroclast, suggesting that basin flooding predated anticline uplift.

The orthogonality of ribbons and minor folds, the similarity of ribbon and minor fold wavelengths, and the broad fold-lava temporal constraints are consistent with the following sequence: (1) broadly synchronous deformation of a thin, competent layer above a ductile substrate; (2) lava flooding; (3) uplift of the broad anticline; and (4) local, late lava flooding in broad synclinal valleys. Thus, volcanism and tectonism were intimately related throughout tessera evolution. Gilmore et al. (1998) argued on the basis of stratigraphic relationships that long-wavelength folds predated ribbons. However, Gilmore et al. did not differentiate ribbons from late grabens (late graben and fold relationships do not constrain ribbon and fold timing), and the material units proposed by Gilmore et al. to flood fold valleys and predate ribbons are simply fold limbs facing away from radar, not material units (Fig. 3), and thus provide no temporal constraints. Thus, geologic relationships indicate that crustal-plateau tesserae record widespread early extension and local, minor perpendicular contraction of a thin, competent layer above a ductile substrate. This distinctive crustal rheology extended across individual crustal plateaus. As the depth to the brittle-ductile transition increased with time, broad, gentle folds formed along plateau margins and short, variably oriented folds formed in the interior; local extension formed late complex grabens. Volcanism accompanied tectonic evolution of tesserae.

GRAVITY, TOPOGRAPHY, AND TECTONIC PATTERNS

Tesserae tectonic patterns preserved within individual crustal plateaus correlate with plateau topography and gravity (Figs. 1 and 4; Ghent and Hansen, 1999). If tessera terrain formed globally and was later locally uplifted at crustal-plateau sites, one would not expect folds to coincide with and parallel plateau margins (Ghent and Hansen, 1999). In addition, if deformation postdated plateau uplift, collapse structures should have formed along the plateau margins, yet none has been identified. The correlation of patterns in topography and gravity data with tessera tectonic patterns strongly favors local over global formation.

CRUSTAL PLATEAU FORMATION

Although preliminary SAR analysis lent support to crustal thickening as a result of downwelling (Bindschadler et al., 1992), several geologic relationships cannot be accommodated by downwelling: (1) an extremely shallow brittle-ductile transition across thousands of square kilometers, (2) early widespread pervasive extension, (3) an increase in depth to the brittle-ductile transition with time, (4) minor rather than major shortening associated with broad folds, and (5) volcanism associated with all stages of tesserae evolution. Furthermore, downwelling would require diapiric sinking and associated crustal transport inward toward individual crustal plateaus, but evidence for such structures is lacking (Phillips and Hansen, 1994). In contrast, these five relationships can be accommodated within a magmatic-accretion model in which local, deep-mantle plumes impinge on globally thin lithosphere (Phillips and Hansen, 1998).

Plateau topographic profiles provide further evidence for upwelling. Although mantle downwelling could account for plateau planform shape and
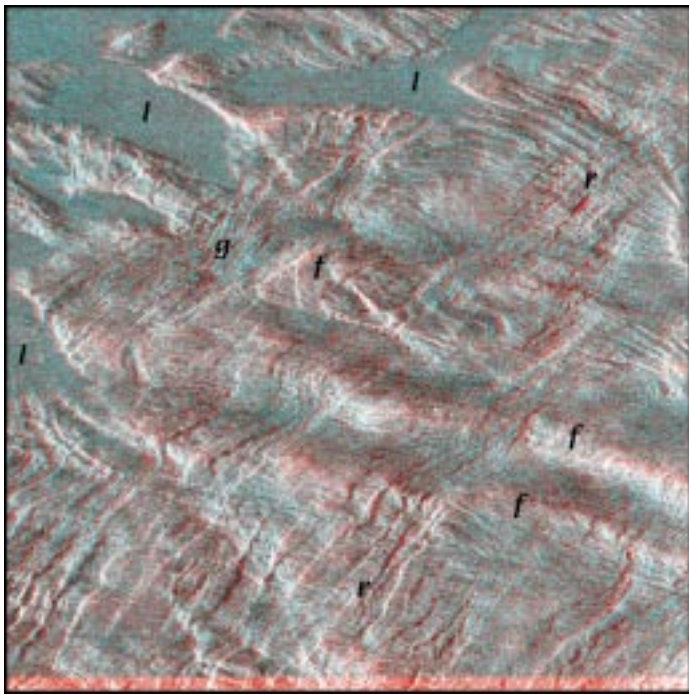
Figure 3. Three-dimensional synthetic aperture radar image (75 × 75 km) of north-central Ovda centered at lat 3.8°S, long 77.1°E (viewed through red [left eye] and blue glasses); local late flows (*l*) postdate structures including northeast-trending ribbons (*r*), west-northwest–trending folds (*f*), and northeast-trending complex graben (*g*). North-northeast–facing fold limbs, interpreted by Gilmore et al. (1998) as post-folding flood lava, are not material units as fine-scale west-northwest–trending lineaments can be traced across fold valleys.

Figure 4. Schematic structure map of typical crustal plateau showing marginal-fold domain and interior basin and dome domain, and intra-tessera basins.

Figure 5. Synthetic aperture radar and generalized geologic map of intratessera basin, Tellus, showing ribbons (red), minor fold crests (short dark blue), intermediate fold troughs (light blue), broad folds (blue), and flood lava (grays and black). Large contiguous patch of black is intratessera embayment basin.

size, downwelling results in broad domical topography rather than steep-sided plateau topography (Bindschadler and Parmentier, 1990, Fig. 13a; Kidder and Phillips, 1996, Fig. 7). Plateau topography likely did not result from late collapse because collapse tends to decrease rather than increase slopes and collapse structures are absent. In contrast, upwelling can accommodate plateau shape, size, and a sharp transition from thick to thin crust. For example, the leading edge of the Hawaiian hotspot marks a sharp transition from thick to thin crust (Watts and ten Brink, 1989, Fig. 15), and high-resolution bathymetry illustrates the steep-sided character of terrestrial oceanic plateaus (Smith and Sandwell, 1997), similar in size to crustal plateaus and postulated as signatures of deep-mantle plumes (Coffin and Eldholm, 1994). Furthermore, given realistic flow laws and thermal budgets, crustal-plateau formation by downwelling requires more than 1 b.y. (Kidder and Phillips, 1996), whereas mantle plumes can thicken the crust in less than 50 m.y. (Coffin and Eldholm, 1994; Farnetani and Richards, 1994).

Given an upwelling model of crustal plateau formation, several other factors can be explained. (1) Impact craters preserved on tesserae are deformed only by late grabens (Solomon et al., 1991; Gilmore et al., 1997). Proponents of downwelling argued that folding led to crater destruction (e.g., Gilmore et al., 1997, 1998), but the minor-contraction-folds record is inconsistent with this hypothesis (Fig. 3). Craters could, however, have been eradicated prior to ribbon formation as a result of local crustal annealing of globally thin lithosphere above mantle plumes (Hansen and Willis, 1998). (2) The ribbon-fold patterns of individual crustal plateaus could record the interaction of plume-related and regional stresses (e.g., Withjack and Scheiner, 1982). (3) Large arcuate tessera inliers within the regional plains exhibit broad folds with perpendicular ribbons and may represent relict, mostly subsided crustal plateaus (Phillips and Hansen, 1994), whereas volcanic rises represent the lithospheric signature of contemporary mantle plumes on thick lit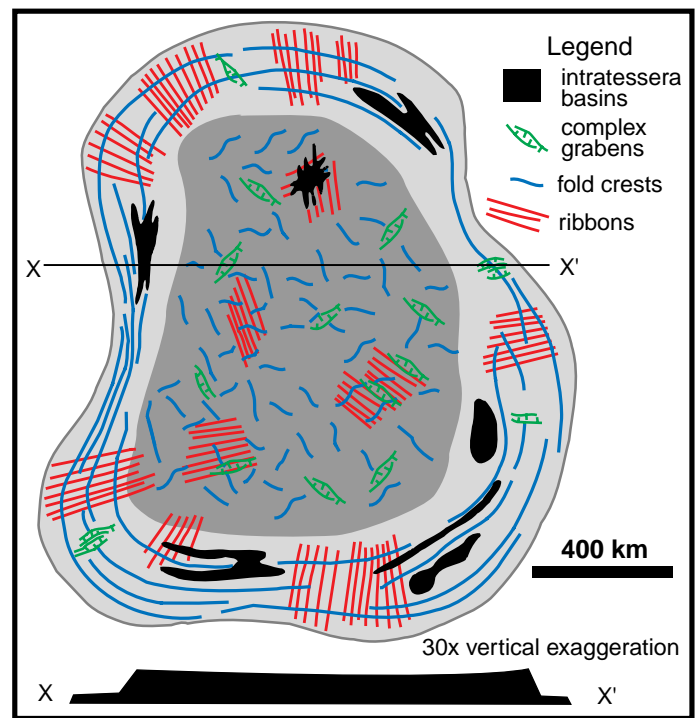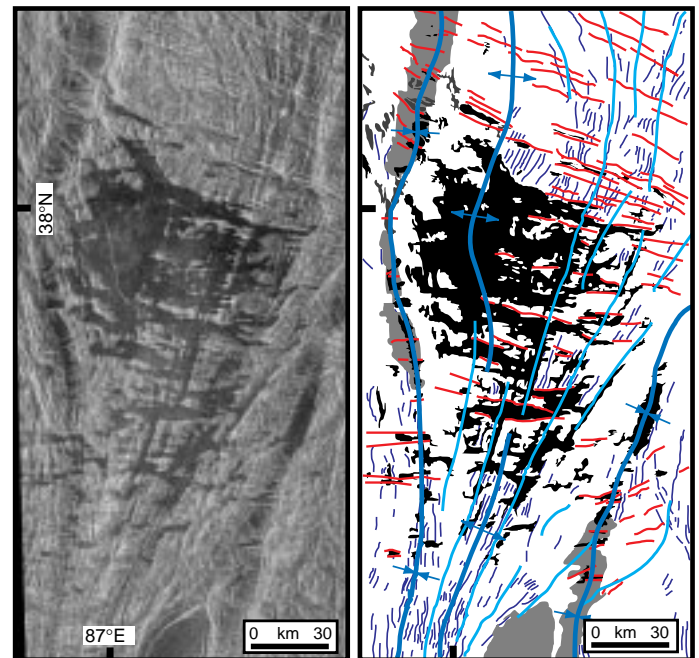hosphere (Smrekar et al., 1997). Phoebe Regio, which exhibits structures transitional between crustal plateaus and volcanic rises, records the transition from globally thin to globally thick lithosphere (Hansen and Willis, 1998; Phillips and Hansen, 1998). (4) Thermal modeling suggests that ribbons formed under conditions of higher surface temperature than at present (Phillips and Hansen, 1998). Thin lithosphere

would enhance regional pressure-release melting of the mantle leading to formation of the globally extensive volcanic plains; accompanying increases in volcanic greenhouses gases could then raise global surface temperature (Bullock and Grinspoon, 1996; Phillips and Hansen, 1998; Phillips and Bullock, 1999).

In summary, there is no evidence to suggest that tesserae form or formed a global onion skin on Venus, although tesserae reflect a time of globally thin lithosphere. Individual tessera terrains did not form synchronously, but rather punctuated time and space as individual deep-mantle plumes imparted a distinctive rheological and structural signature on ancient thin crust across discrete 1600–2500-km-diameter regions. Plume-related magmatic accretion led to crustal thickening at crustal plateaus.

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