topography. In contrast, tessera terrain is commonly characterized by a steep-sided plateau shape. Domal uplifts are commonly characterized by large rift zones (e.g., Beta and Atla regiones), and all domal uplifts contain at least two large (~200 km diameter) shield volcanoes [*McGill et al.*, 1981; *Janle et al.*, 1987; *Stofan et al.*, 1989; *Senske and Head*, 1989]. The only shield-type volcanic structure observed in the tessera terrain is located in Tellus Regio (Figure 5*a*) and is no more than 50 km in diameter. Moreover, deposits related to the volcano clearly embay tessera structures, indicating that the feature postdates deformation.

Some of the differences between domal uplifts and tesserae might be explained if tesserae represented older, more evolved mantle upwellings, in which the mantle source has died out or moved away. However, no structures are apparent in the tessera terrain that represent good candidates for altered or degraded versions of rifts or large shield volcanoes. Regions of tessera are present on the flanks of both Bell [*Barsukov et al.*, 1986; *Janle et al.*, 1987] and Beta regiones [*Bindschadler et al.*, 1990*a*; *Campbell et al.*, 1989]. In both cases, the tessera terrain appear to be relatively old, embayed by plains volcanism. While it is possible that the formation of these few small regions of tessera is related to mantle upwelling, available data strongly suggest that most regions of tessera did not form in this way.

Crustal Underplating

Emplacement of a large volume of low-density (crustal) intrusive material at or near the crust-mantle boundary should result in surface uplift and deformation (Figure 8). On the basis of the distribution of elevations for the tessera terrain, *Nikolaeva et al.* [1988] suggest that tessera terrain is composed of highly feldspathic material, such as anorthosite. Rapid emplacement of low-density material such as anorthosite near the crust-mantle boundary can be modeled as a gravitational relaxation process, in which surface topography grows in order to balance the excess low-density root at depth. Intruded/underplated material could be of either basaltic or anorthositic composition; the only requirement is that it be less dense than underlying mantle materials. Underplating is predicted to result in uplift and extensional deformation of the surface [*Bindschadler*, 1990]. Unless underplating proceeds very rapidly and has occurred relatively recently, gravity anomalies should only reflect the thickened crust. There are several reasons why such a model is unlikely to explain the formation of any of the types of tessera terrain.

Surface deformation due to underplating is exclusively extensional in nature. Observations suggest a compressional origin for subparallel ridged terrain (T_{sr}), as well as much of the disrupted terrain (T_{ds}), which shares transitional boundaries with the T_{sr}. Trough and ridge terrain (T_{tr}) structures are consistent with extensional deformation but require that the underplating event create a highly organized orthogonal pattern of structures.

The volume, surface area, and heat input corresponding to such an intrusion are extremely large. For Tellus Regio alone, the intrusion would correspond to a surface area of ~1.5×10^6 km^2. Assuming a relatively large density contrast between intrusive material and mantle material (20%), a minimum of 7.5×10^6 km^3 of underplated material is required to support the topography of Tellus Regio through Airy isostasy. To differentiate this volume of material from the mantle of Venus would require a minimum of 10^{25} J, assuming a latent heat of fusion of 400 kJ kg^{-1}. This corresponds to the entire global heat loss for Venus over a period of ~10^5 years, based on a nominal rate of 3.4×10^{13} W [*Solomon and Head*, 1982]. Such a differentiation event might occur over a much longer period of time. However, relaxation of stresses due to underplating should occur quickly. The characteristic time for such relaxation is $\tau = {}^{4\pi\mu}/_{\rho g \lambda}$, where μ is mantle viscosity, ρ is density, and λ is the characteristic wavelength of a region of noncompensated topography [*Turcotte and Schubert*, 1982]. For λ =1000 km (e.g., Laima Tessera), $\mu = 10^{21}$ Pa s, and ρ = 3000 kg m^{-3}, we find $\tau = 1.6\times10^4$ years. Thus a single differentiation event of the suggested scale would require more energy than is available from the planet. The organized structural pattern of the T_{tr} requires that the geometry of the intrusion in plan view and thus the orientation of principal stresses remain approximately constant. It is highly unlikely that repetitive intrusion events, at least 10^6 years apart to satisfy heat flow constraints, could also repeat the same geometry time after time. Although underplating may occur on Venus, it does not appear to be feasible as a mechanism for the production of the observed high topography and deformation of tesserae.

Seafloor Spreading Analogy

This model is based upon the hypothesis that a process analogous to seafloor spreading occurs in western Aphrodite Terra [*Head and Crumpler*, 1987] and has been suggested to apply specifically to Laima Tessera [*Head*, 1990*b*]. As modeled by *Sotin et al.* [1989], spreading on Venus is expected to result in crustal thicknesses somewhat greater than on Earth (~15 km), due to higher average mantle temperature. To produce the relatively high elevations and thick crust of tessera terrain, a region of anomalously high mantle temperature (e.g., a hotspot) is required, similar to the situation postulated by *Sotin et al.* [1989] for Ovda Regio. On Earth, enhanced crustal thicknesses (up to ~9 km greater than average) are found along the Mid-Atlantic Ridge at its intersection with the Iceland hotspot [*Pálmason and Saemundsson*, 1974]. Thus a spreading-like process appears to be capable of producing a topographically high region characterized by relatively shallow compensation.

Structures created by the spreading process on Earth include fracture zones and abyssal hills in approximately orthogonal orientations. The exact mechanism(s) by which abyssal hills form on Earth is controversial, but both spreading-related extension and volcanism are thought to contribute. Tectonic processes associated with near-spreading center extension, thrusting, and block rotation are thought to dominate slow spreading ridges (~1-3 cm yr^{-1} full spreading rate) [e.g., *Harrison and Stieltjes*, 1977; *Macdonald*, 1986], while episodic volcanism occurring at the ridge crest becomes increasingly important along medium (~3-9 cm yr^{-1}) and fast spreading ridges (>9 cm yr^{-1}) [*Kappel and Ryan*, 1986; *Barone and Ryan*, 1988; *Pockalny et al.*, 1988]. Suggested rates of spreading on Venus [*Kaula and Phillips*, 1981; *Crumpler and Head*, 1988; *Sotin et al.*, 1989] are in the range of slow spreading ridges. On the basis of our understanding of the terrestrial spreading process, abyssal hills on Venus should be predominantly tectonic in origin, consistent with the tectonic origin of ridge and trough structures in the T_{tr}. However, the transition between tectonic-dominated and volcanic-dominated