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TECTONIC AND MAGMATIC EVOLUTION OF VENUS

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1. INTRODUCTION

Modern Venus exploration dates from the first geochemical measurements of the surface in the early 1970s by Soviet landers (Vinogradov et al 1973) and from the acquisition by the *Pioneer Venus Orbiter*, beginning in 1978, of nearly-global topographic and gravity data (Pettengill et al 1980, Sjogren et al 1980). Venus was revealed geologically as a planet with both strong similarities to and strong differences from Earth. Over the past decade, a wealth of new data has become available from continued surface exploration by the Soviet Union, from improved Earth-based radar astronomy by the United States (e.g. Campbell et al 1991), and from orbiting radar missions by both countries (Barsukov et al 1986, Saunders et al 1992). The tectonic and magmatic style of Venus has come into sharper focus and numerous, conflicting models for the evolution and present state of the planet have been advocated.

Knowledge of Venus increased dramatically with the high-resolution radar mapping of the surface by the *Magellan* spacecraft, commencing in September of 1990. Image resolution to order 100 m and altimetry with vertical and horizontal resolutions of about 80 m and 10 km, respectively,

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have allowed a mapping that enables us to consider both local detail and regional to global scale features in constructing evolutionary models.

The purpose of this paper is to provide a review of models of mantle and crustal dynamics and to attempt to tie those models to hypotheses about the magmatic and tectonic evolution and present state of the planet, including the nature of major surface features. We ask the following questions, among others:

1. What are the origins of highlands on Venus and are the different kinds of upland terrain related through evolution or are they manifestations of distinct processes?
2. What are the origins of the forces required for tectonic deformation?
3. What is the importance of lower crustal flow in crustal dynamics?
4. What is the role of melt residuum in the tectonic evolution of the planet?
5. What controls magmatic access to the surface? How much is magmatic style simply a passive response to lithospheric strain?
6. What is the overall style of global tectonics and magmatism?
7. What is the origin of coronae?
8. What role do upgoing mantle plumes play in lithospheric tectonics?
9. What is the behavior of tectonism and magmatism over the past billion years ? Has it been steady, has it undergone a secular decline, or have there been episodes of increased activity?

We begin with a brief review of Venus geology. Our attempt here is to acquaint the reader with the physiography of the planet and provide a limited description of its major tectonic and magmatic features. For an in-depth survey of the planet, the reader is referred to papers in two dedicated issues of the *Journal of Geophysical Research* published in 1992 (Issues E8 and E10, Volume 97, 1992).

2. REVIEW OF VENUS GEOLOGY

2.1 *A Snapshot of Venus*

The hypsometry of Venus is unimodal, in contrast to the bimodal hypsometry of Earth. We distinguish three topographic domains (Figure 1). *Lowlands*, or *plains*, generally lie below the mean planetary radius (MPR); *uplands*, or *highlands*, are greater than ~2 km above MPR; and the region in between (~0 to 2 km above MPR) is here termed the *mesolands* [which correspond approximately to the "rolling plains province" of Masursky et al (1980)]. The highlands comprise <20% of the surface, and the plains and mesolands represent, roughly, equal parts of the remainder. Areas assigned to the mesolands include the chasmata region lying between Thetis and Atla regiones, much of the region included in the triangle whose

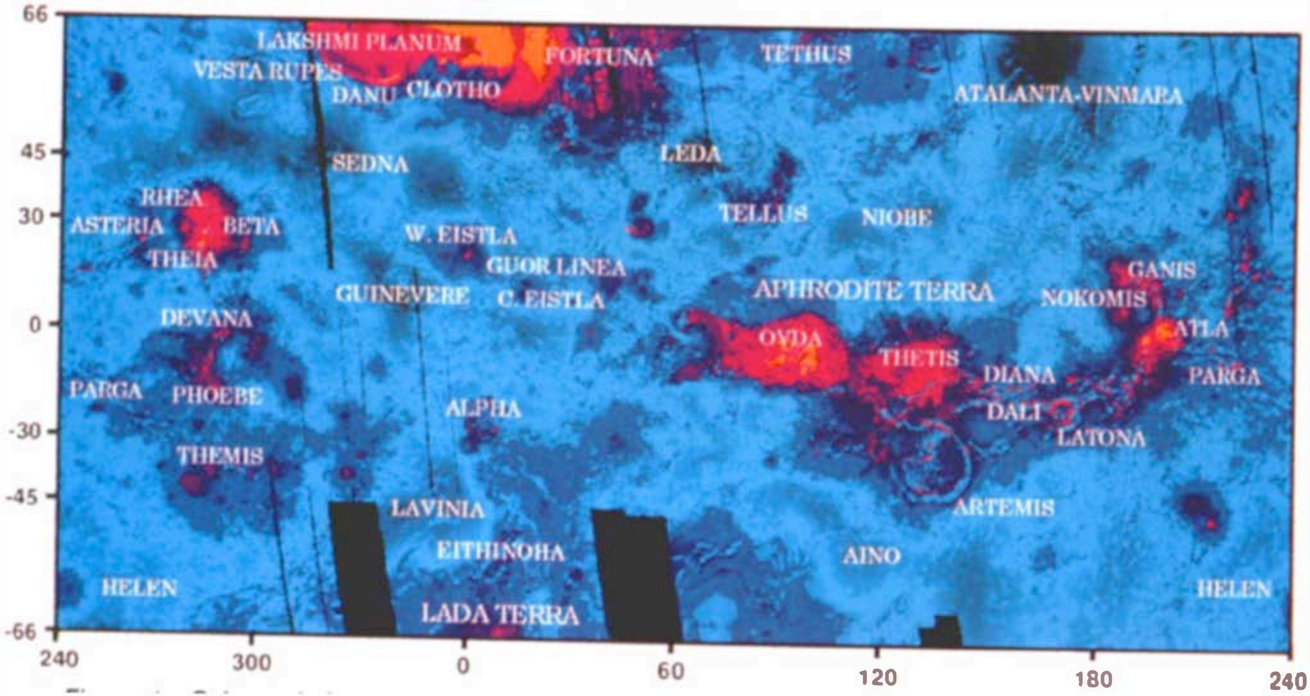
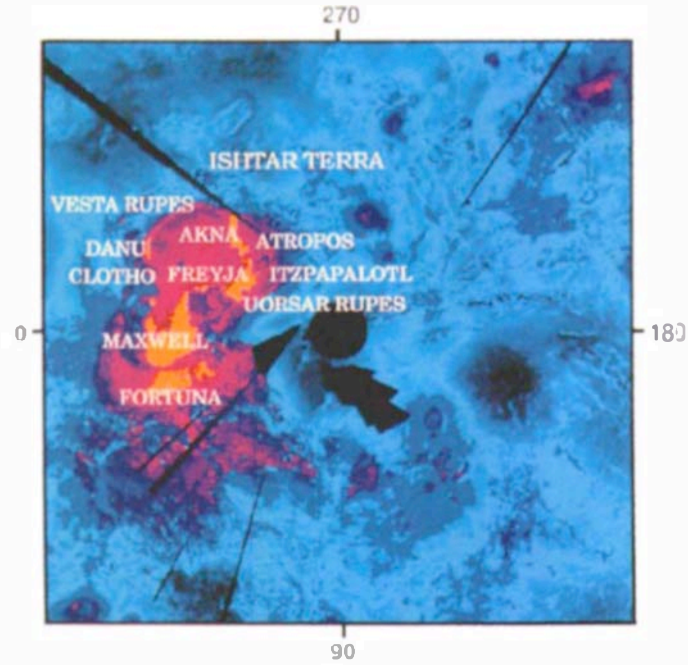


Figure 1 Color-coded representation of Venus surface topography (shown with place names) in Mercator (from 66°N to 66°S) and polar stereographic (from 47° to 90°N and S) projections (see flip side), derived from the final Global Topography Data Record (GTDR) product produced by Massachusetts Institute of Technology and Jet Propulsion Laboratory. Pixel size is $\sim$ 40 x 40 km and the mean planetary radius (MPR) is 6051.84 km. The black shades and lighter blues (lowlands/plains) correspond to elevations generally < 2 km below MPR and between 0 and 2 km below MPR, respectively. Regions of medium and dark blue ($\sim$ mesolands) lie between approximately 0 and 2 km above MPR. The magenta and orange-yellow regions (highlands) correspond to elevations between 2 and 4 km above MPR, and > 4 km above MPR, respectively.


270
ISHTAR TERRA
VESTA RUPES
AKNA
ATROPOS
DANU
CLOTHO
FREYJA
ITZPAPALOTL
UORSAR RUPES
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MAXWELL
FORTUNA
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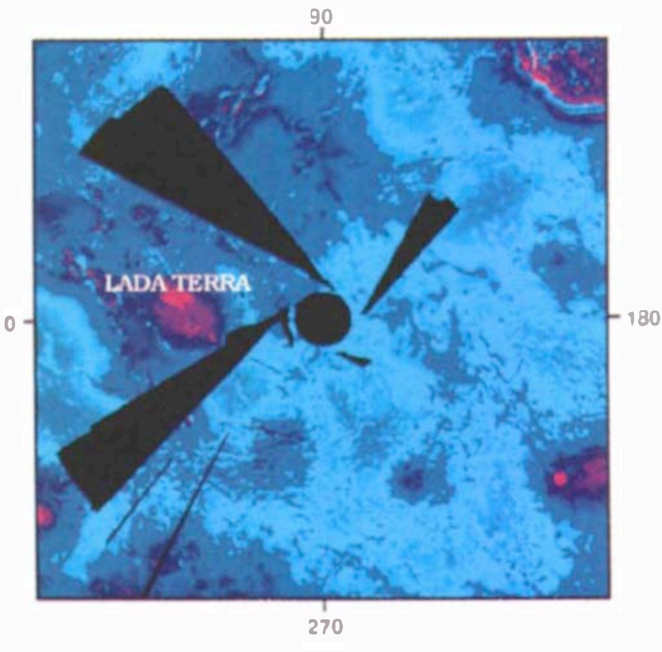
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vertices are defined by Beta, Atla, and Themis regiones (called the "BAT region" by Head et al 1992), the area including Artemis Corona and elevated regions to the west, and parts of the "Southern Highlands," including Lada Terra. Geographically, the uplands include Ishtar Terra, parts of Aphrodite Terra that lie along the equator from $\sim 80°E$ to 140°E (Ovda and Thetis regiones), Atla Regio, and the higher portions of Beta, Phoebe, and Themis regiones, which lie along the 285° meridian (Figure 1). Highstanding topographic areas in the equatorial regions, including Aphrodite Terra, and Beta and Eistla regiones are also known collectively as the "Equatorial Highlands" (Phillips et al 1981). Basins surrounding the highlands include: Atalanta, Vinmara, Sedna, Niobe, Leda, and Guinevere planitiae in the north, and Aino, Helen, and Lavinia planitiae in the south. Small highlands, Alpha, Eistla, Bell, and Tellus regiones, lie within the lowlands. Ishtar Terra, the highest region on Venus, is comprised of an interior plateau, Lakshmi Planum (3–4 km above MPR), surrounded by interior mountain belts ranging from 4–11 km above MPR, and outlying tesserae (1–5.5 km above MPR). Coronae (quasi-circular features) and chasmata (deep troughs) lie mainly within the mesolands; chasmata are further restricted to low latitudes. Tessera terrain [now also burdened with the name "complex ridged terrain" (CRT)] was originally described as terrain that ". . . is cut by two or more sets of ridges and grooves . . ." (Barsukov et al 1986); it occurs as a dominant tectonic fabric in some highland regions as well as in isolated single features and isolated patches largely in the plains.

2.2 *Structure*

Plains, or planitiae, lie below MPR, and are defined largely by radar-smooth (dark) fields, interpreted as regions characterized by flood or plains volcanism with locally preserved lava channels. Geologically, the highlands are divisible into Ishtar Terra, crustal plateaus, and volcanic rises. Crustal plateaus include Alpha, Tellus, Ovda, Thetis, and Phoebe regiones, which consist of steep-sided regions dominated by complex tectonic fabrics. These features correlate with small gravity/geoid anomalies, implying small (<100 km) apparent depths of compensation (ADC). Volcanic rises, which include Atla, Beta, Bell, and Eistla regiones, form domical, circular to elongate regions characterized by shield volcanoes, flows, and rifts. Large gravity anomalies, implying large ADCs (100–300 km), coincide with these features. The mesolands host topographic highs of crustal plateaus and volcanic rises and are characterized by linear to arcuate chasmata and quasi-circular coronae. For a summary of Venusian tectonism, see Solomon et al (1991, 1992).

2.2.1 PLAINS The plains record several episodes of volcanism and deformation; superposition demonstrates a continual interplay in time and space. Structural features are dominantly linear in plan view, and include, from largest to smallest: ridge and fracture belts, wrinkle-ridges, and extensional fractures, as well as "parallel lineations." Coherent patterns of these features over large areas (thousands of km^2) imply that they record a crustal response to mantle dynamics. Wrinkle ridges are probably the most common structure on Venus. They are similar to features preserved on the Moon and other terrestrial planets including Earth and Mars (Plesica & Golombek 1986, Watters 1988). Wrinkle ridges are long and sinuous, 1–2 km wide, and tens to hundreds of km long. Their continuity in orientation and spacing (tens of km) over broad regions indicates small amounts (1–5%) of regional crustal shortening normal to their trend (e.g. Suppe & Connors 1992, Bilotti & Suppe 1992, McGill 1992).

Extensional fractures—straighter than their wrinkle ridge counterparts—are also distributed over broad regions, and record crustal extension normal to their trend. Ridge and fracture belts comprise elevated zones of intense deformation characterized by parallel, closely spaced (5–10 km) ridges interpreted as contractional folds, and closely spaced normal faults, locally paired forming graben, respectively. These belts, best developed in Lavinia, and Atalanta-Vinmara planitiae, range in width from tens to a few hundred km and reach over 1000 km length. Ridge and fracture belts commonly occur together, either as distinct domains with different trends (e.g. Lavinia) or within single composite belts (e.g. Atalanta-Vinmara). These belts rise 500 m above the adjacent undeformed plains, and individual belts lie 100–300 km apart. The width of the deformed belts is comparable to the width of the relatively undeformed plain domains between belts. In Atalanta-Vinmara Planitiae, deformation belts trend north merging and bifurcating along strike to form an anastomosing pattern: ridge belts become hybrid fracture belts locally along strike; hybrid belts preserve fold ridges, although fractures, faults, and graben dominate the morphology (Solomon et al 1992). In Lavinia Planitia, fold belts trend NNE and fracture belts trend SSE, together expressing an orthogonal regional fabric. Locally, orthogonal contractional folds and extensional faults are present in a single deformed belt. Deformed belts surround relatively undeformed blocks of plains. Wrinkle-ridges and extensional fractures in the undeformed blocks parallel the NNE- and SSE-trending belts, respectively. The pattern of wrinkle ridges and extension fractures suggests that ridge belts and fracture belts record crustal shortening and extension, respectively, normal to their trends (e.g. Squyres et al 1992b). The coherence of strain patterns over thousands of km^2, with parallelism of principle strain axes inferred for both the deformed belts and adjacent relatively undeformed plains, indicates that the deformed belts record coherent strain over hundreds of thousands of km^2. Coherence in surface strain over such great distances may require that deformation was transmitted to the crust from the mantle below (e.g. Bilotti & Suppe 1992, McGill 1992, Squyres et al 1992b). This contrasts sharply with Earth, where most deformation is concentrated at plate boundaries.

"Parallel lineations," newly recognized in *Magellan* images, are sets of short, thin, straight, parallel fractures whose microwave reflectivity does not depend on the direction of radar illumination (Banerdt & Sammis 1992). These fractures may occur over hundreds of km^2 with little change in their spacing (1–2.5 km $\pm$ 1/3 km) despite diverse settings; therefore, the fractures must either always extend to a depth of 1 km, or the spacing must be independent of layer thickness. Banerdt & Sammis (1992) propose that fracture spacing relates to mechanical properties of the surface basalt layer (<1 km thick) above a horizontal detachment; during lithospheric extension, the horizontal detachment allows frictionally resisted lateral displacement of the surface layer resulting in a fracture spacing that is independent of layer thickness, and dependent on mechanical properties of the layer. Study of parallel lineations may, therefore, provide clues to mechanical properties of the surface materials, and they may also serve as relative temporal markers for deformational sequences.

2.2.2 MESOLANDS Topographic troughs and quasi-circular features dominate the mesolands. Circular to quasi-circular features include: coronae, ranging from 100–2600 km in diameter with a strong mode at 200–300 km; and novae, which have diameters less than 100 km (Stofan et al 1992). Chasmata, marked by topographic troughs and parallel lineaments, are divisible into symmetric and asymmetric varieties.

2.2.2.1 *Coronae* Coronae are unique to Venus, and are divisible into features that display radial, and concentric fractures, or both (Stofan et al 1992, Squyres et al 1992a). Novae generally display radial fractures and may represent an early stage of corona formation (Stofan et al 1991, Janes et al 1992). Coronae typically display a central region higher than surroundings, a raised rim, and a moat. Concentric fractures parallel the rim-moat transition. Radial fractures may be preserved only within the coronae, or may extend beyond the outermost concentric fracture set that marks the typically narrow (generally <100 km) tectonic annuli. Annuli commonly parallel the rim of the coronae just inboard of the moat. Moderate to large amounts of volcanism accompany corona formation. Lava reaches the surface locally through both radial and concentric fractures. Double-ring concentric coronae show a sequence of inner concentric faulting followed by volcanism, followed in turn by faulting and volcanism within an outer ring, indicating that coronae can grow outward by deformation associated with magmatic pulses (Hansen & Phillips 1993a). Extensional faults characterize the annuli, although contractional ridge belts and small-scale strike-slip structures are present locally along parts of individual corona and formed in response to local stresses that added constructionally or destructively with stresses involved in corona formation (e.g. Cyr & Melosh 1993). Radial fractures likely result from crustal doming early in corona evolution (Janes et al 1992, Squyres et al 1992a) which may continue with further development (Hansen & Phillips 1993a). Concentric structures form as the coronae expands, and during crustal relaxation. Intrusive magma is probably responsible for the thickened crust of the interior plateau.

Coronae are common on Venus: Over 360 are recorded across the planet (Stofan et al 1992). Stofan et al (1992) stated that coronae are "randomly" distributed in the plains, display an affinity with low latitudes, and cluster along Parga and Hecate chasmata, and within the Beta-Atla-Themis region. We suggest, however, that the distribution of coronae is sensitive to elevation. Coronae avoid both the plains and the uplands, and are concentrated in the mesolands; they are commonly located near chasmata or rifts. Herrick & Phillips (1992) showed that coronae cluster about MPR, and that they are uncorrelated with both strong positive and strong negative geoid anomalies. Their study suggests to us that if coronae are related to mantle convection, they are not directly linked to either upwellings or downwellings.

2.2.2.2 *Chasmata* Chasmata are long linear to arcuate troughs marked by lineaments that are probably fractures and faults. Chasmata are divisible into two groups—symmetric and asymmetric—depending on their topographic profiles normal to their trend (Hansen 1993). Asymmetric chasmata, the deepest chasmata, are bounded on one side by a ridge that is as high as the trough is deep. The trough-to-ridge differential can be as much as 7 km, over a 30 km distance. Fractures generally parallel the trough. Asymmetric chasmata are spatially associated with coronae, and overlap the coronal troughs with coronae residing on the ridge side (topographic high). Symmetric chasmata are not spatially associated with individual coronae, although they may be associated with corona clusters. These chasmata are significantly shallower than asymmetric chasmata, and trend outward from volcanic rises. They are marked by single, linear topographic troughs, or a chain of troughs, that range in width from 50 to 150 km, and extend for hundreds to thousands of km. On the basis of their structures and the presence of extensive syn-tectonic volcanism, symmetric chasmata are interpreted as rifts (Schaber 1982, McGill et al 1981). The amount of crustal extension across each rift zone is small.

2.2.3 HIGHLANDS Geologically, the uplands are divisible into volcanic rises, crustal plateaus, and Ishtar Terra. We discuss each in turn.

2.2.3.1 *Volcanic rises* Volcanic rises, defined both topographically and geologically, form a class of structures interpreted as the surface expressions of mantle upwellings on the basis of geological and geophysical relations (e.g. Phillips & Malin 1984). The upwelling hypothesis for volcanic rises was originally based on the strong analogy of volcanic rises to terrestrial hotspot terrains (McGill et al 1981), quasi-radial rifting suggesting buoyant uplifting from below, and large ADC values (e.g. Smrekar & Phillips 1991). Volcanic rises include Atla, Beta, Western Eistla, and Bell regiones. We describe these regions below, and then summarize the characteristics that must be accommodated by evolutionary models.

Atla Regio—a broad highland dome marked by numerous volcanic centers and rifts, shield volcanoes, and fractured terrains—occupies easternmost Aphrodite Terra. Atla comprises several large shield volcanoes. Ozza Mons lies at the intersection of five chasmata, each interpreted as a rift (Bindschadler et al 1992b, Senske et al 1992). Volcanic flows fill graben, and are in turn cut by faults, demonstrating temporal and spatial interplay between tectonism and volcanism. Local tesserae comprise the oldest unit. Atla has an ADC of 200 km.

Beta Regio is composed of Rhea and Theia Montes defined by tesserae and constructional volcanoes, respectively (Senske et al 1991, 1992). The three arms of Devana Chasma, marked by numerous normal faults and interpreted as rifts (McGill et al 1981, Schaber, 1982, Campbell et al 1984), intersect at Theia. Rift structures clearly cut across the tesserae of Rhea Mons; rifting and volcanism occurred concurrently in Theia Mons (Senske et al 1992). Beta has an ADC of 270–400 km.

Western and Central Eistla define broad rises with large volcanic constructs and rift zones correlated with strong positive gravity anomalies and ADCs of 100–200 km (Grimm & Phillips 1992). Twin peaks, the volcanoes Sif and Gula Montes, dominate Western Eistla, and Sappho Patera dominates Central Eistla. Radial fractures mark the topographic rise and the individual volcanoes. The largest rift, Guor Linea, extends 1000 km southeast of Gula Mons and reaches a width of 50–75 km; it may record minor clockwise rotation and bulk NNE-SSW crustal extension normal to its trend.

Bell Regio comprises local tesserae and the shield volcano Tepev Mons and several smaller volcanic centers. A north-trending fracture zone, once interpreted as a rift (Janle et al 1987), cuts the region, but unlike the rift zones of Beta and Eistla, there is little evidence of extension in *Magellan* data (Solomon et al 1992). Tesserae form the oldest terrain type. Bell Regio correlates with a 100–200 km ADC (Smrekar & Phillips 1991).

These volcanic rises share the following characteristics: (*a*) they have broad topographic rises; (*b*) they are generally composed of several shield volcanoes; (*c*) they are located at the convergence of rifts; (*d*) they exhibit widespread volcanism synchronous with rift formation; (*e*) their tesserae, when present, constitute the oldest terrain type; and (*f*) they coincide with large ADCs, which taken together with the high topography, probably indicates that a significant component of the topography is compensated by mantle convection. These features are all consistent with the interpretation that volcanic rises represent surface manifestations of mantle upwellings.

2.2.3.2 *Crustal plateaus* Crustal plateaus include Ovda, Thetis, Phoebe, Alpha, and Tellus regiones (Figure 1). Ovda (3000 × 2000 km) and Thetis (1500 × 1500 km) are topographically higher than Phoebe (1700 × 1000 km), Alpha (1500 × 1300 km in extent), and Tellus (2000 × 1600 km). They all share the following characteristics:

1. They are steep-sided regions reaching elevations of 2–4 km above the surrounding plains.
2. The highest elevations are at the margins.
3. They display rough topography at a horizontal scale of tens of km.
4. They are dominated by tesserae or CRT that exhibit structures best described as basin and dome morphology, suggestive of polyphase strain histories. We term this fabric, which is best preserved within the interior of the plateaus, "amorphous terrain" because of its lack of structural coherence. Amorphous terrain commonly hosts small (20–50 km wide and 50–150 km long) elliptical radar-dark and radar-smooth basins, which have been called "intertesserated plains" (Bindschadler & Head 1991); we use the term intratesseral plains, which more appropriately defines the basins as within a single tessera rather than between tesserae. Intratesseral plains show no preferred orientation or spatial distribution within the plateau interior.
5. The boundary of deformation with undeformed plains is extremely irregular.
6. The most coherent structural fabrics are fold belts preserved locally along parts of the plateau margins. Although Bindschadler et al (1992b) state that marginal fold belts parallel their respective boundaries, fold belts do not totally encircle plateaus; fold belts may parallel, or they may trend at a high angle into the surrounding regions. In addition, the boundaries of some plateaus are locally dominated by extensional structures representing the youngest deformation on these plateaus (Solomon et al 1992).
7. Contractional and extensional fabrics are variably developed through-

out plateaus; crosscutting relations and local lava-flooded regions indicate that contractions and extensions took place at different times in different areas.

8. Plateaus have small gravity/geoid anomalies, and ADCs arc consistently < 100 km, and often < 50 km.

2.2.3.3 *Ishtar Terra* Ishtar Terra includes Lakshmi Planum (4 km above MPR) and the surrounding deformed belts which include the interior mountain belts—Maxwell, Danu, Akna, and Freyja Montes (3.5–11 km above MPR), as well as the outlying tesserae—Fortuna, Clotho, Atropos, and Itzpapalotl (1–4.5 km above MPR) (Figure 1). Uorsar Rupes marks the steep slope along the northern boundary of Itzpapalotl, and Vesta Rupes marks the steep slope that bounds Danu to the south. Maxwell Montes hosts the highest point on Venus, and Fortuna Tessera comprises the largest tessera terrain. We use Fortuna Tessera for only that part of Fortuna east of Maxwell and west of 30°E; farther east the region has not been imaged by *Magellan* because of a superior conjunction data acquisition gap. Ishtar Terra is notably free of coronae.

Lakshmi Planum is relatively undeformed. Colette and Sacajawea calderas are preserved within it, and flooded tesserae are exposed at the highest elevations (Roberts & Head 1990a,b; Kaula et al 1992). The high plain of Lakshmi is relatively structureless, except along the periphery where wrinkle ridges deform lava flows near the mountains, indicating that some deformation postdates plateau volcanism.

Danu Montes and outboard Clotho Tessera are different from their Ishtar counterparts. Danu is a long, narrow mountain belt that reaches only 1.5 km above Lakshmi; Clotho is lower than other Ishtar tesserae (1–1.5 km above MPR) and is dominated by linear troughs of probable extensional origin.

In contrast, the structural pattern of the other mountain belts and tesserae consists of tightly spaced (< 3 km) "penetrative" lineaments ["penetrative fabric" of Keep & Hansen (1993b), "fine-scale structures" of Bindschadler et al (1992a)] that are deformed by the linear fold ridges and valleys, which dominate the structural fabric. The penetrative lineaments may represent parallel lineations (Banerdt & Sammis 1992) contracted during crustal shortening. Fold ridges—the result of crustal contraction (Campbell et al 1983; Barsukov et al 1986; Crumpler et al 1986; Kaula et al 1992; Solomon et al 1991, 1992)—extend for hundreds of km. Small-scale extensional fractures trend normal to the fold ridges, indicating that local extension parallel to ridge trends accompanied contraction (Hansen & Phillips 1993b,c; Keep & Hansen 1993a,b). Locally, structural valleys are filled with lava floods that formed late in the deformational history (e.g. Akna and Atropos); few vents and channels are identified. In Maxwell and Fortuna, and in Akna and Atropos, parallelism of structures defines a coherent tectonic pattern that extends from Lakshmi wrinkle ridges through peripheral mountain belts and well out (> 500 km) into adjacent tesserae; spacing of fold ridges (3–15 km range, ~6 km average) does not change from mountain belt to tessera (Hansen & Phillips 1993b, Keep & Hansen 1993b). Topography, and not structure, marks the major difference between mountain belts and neighboring tesserae. Extensional structures are identified locally in Freyja and Danu, and within the northern and southern slopes of Maxwell (Kaula et al 1992, Smrekar & Solomon 1992), but the highest elevations of the mountain belts and the tesserae display no evidence of gravitational collapse.

2.2.3.4 *Summary* The highlands are divisible into volcanic rises, crustal platcaus, and Ishtar Tcrra. Although most workers agree that volcanic rises represent hotspots, crustal plateaus and Ishtar Terra are variably interpreted as surface expressions of major subsolidus crustal flow in response to mantle downwelling, and crustal thickening by magmatism related to late-stage evolution of mantle upwelling. Models for the formation of crustal plateaus and Ishtar Terra must account for these important differences: (*a*) Ishtar is structurally much more coherent; dominant fold ridges parallel the orientation of their host mountain belt with respect to Lakshmi; (*b*) elevated regions (mountain belts) are located inside Ishtar Terra, against Lakshmi, rather than against the exterior plains boundary, as is the case for the crustal plateaus; (*c*) the interior of Ishtar, Lakshmi Planum, is a high radar-smooth and radar-dark plateau that is essentially free of deformation; in contrast, crustal plateau interiors record spatially complex strain histories; (*d*) Ishtar tesserae are structurally more coherent than the amorphous terrain that dominates crustal plateaus; and (*e*) the spatial distribution of polyphase versus coherent single-phase deformation is the opposite for Ishtar and crustal plateaus.

2.3 *Volcanism*

Evidence for volcanism is globally widespread on Venus (for a review see Head et al 1992). Volcanic features include shield volcanoes and calderas (20 to > 100 km diameter), and are not concentrated in linear zones as they are on Earth where they mark convergent and divergent plate boundaries. On Venus volcanoes are distributed throughout the mesolands and occur locally in the uplands with a concentration of volcanic centers (2–4 times the global average) within the Beta-Atla-Themis region (Head et al 1992). Evidence for volcanism is ubiquitous on the planet. The plains generally host flood volcanism and sinuous channels; obvious volcanic centers are mostly lacking; volcanic rises are dominated by shield volcanoes and rifts; crustal plateaus display evidence of probable early volcanism; and lava-flooded structural valleys are present throughout the Ishtar deformed belts. The majority of volcanic land forms are consistent with a basaltic composition with possible exceptions including steep-sided domes that may represent more siliceous compositions and sinuous rilles possibly indicative of ultramafic lava (Head et al 1991, 1992; Pavri et al 1992; McKenzie et al 1992b; Baker et al 1992). The scale of individual features is typically small—much less than 125,000 km^2 (Head et al 1992).

2.4 *What Impact Craters Tell Us about Resurfacing History*

The distribution and degradation states of impact craters on planetary surfaces have long been used as a guide to both resurfacing history and surface age. Venus has provided a special challenge amongst the terrestrial planets in understanding the spatial disposition and preservation state of impact craters; interpretations of the cratering record have led to wildly different scenarios for the resurfacing history of the planet. In part this is because there are only about 900 impact craters on the surface, so spatial statistics are subject to large uncertainty on a regional basis.

The surface age of Venus has been estimated to be about 400–500 Ma. This value is obtained by dividing the observed number of craters larger than ~ 30 km diameter by an estimate of impact crater formation rate for that size range. Formal uncertainties in regression fits to the crater size–frequency distribution, as well as uncertainties in the meteoroid flux rate, lead to a 90% confidence interval no better than 300–600 Ma (Phillips et al 1992). An age estimate thus obtained has a variety of interpretations because: (*a*) it is a global average, and says nothing about age variations from place to place; (*b*) it gives the "production age" of a surface (or surfaces) under an assumption that once a fresh surface is formed, impact craters subsequently accumulate without disturbance; and (*c*) it is subject to an alternative interpretation that does not give the age of a surface but rather the average lifetime of a crater before it is destroyed (the "retention age").

Phillips et al (1992) have found that: (*a*) The crater distribution cannot be distinguished from a spatially random population. (*b*) Most craters ($\sim 80\%$) appear "fresh" in that the crater and its continuous ejecta blanket are not modified by faulting ("tectonization") or volcanic embayment. (*c*) Modified (embayed and/or tectonized) craters have a statistically significant negative spatial association with unmodified craters. That is, modified craters occur in or near the borders of regions of low overall spatial crater density. (*d*) There is an inverse correlation between radar cross section, σ_o, and spatial crater density.

A useful way to interpret the distribution of surface ages on Venus is to consider an age spectrum, defined as the fractional surface area as a function of age (Phillips 1993). In this regard, points (*a*) and (*b*) have been used to argue that Venus underwent a catastrophic resurfacing (Schaber et al 1992), i.e. the age spectrum is a "spike" at ~300–500 Ma ago. Points (*b*), (*c*), and (*d*) have been used to argue that the age spectrum is much broader, i.e. surface modification is much more widely distributed in time (Phillips et al 1992). Point (*c*) has been used to argue that craters have been removed from the surface, i.e. many low crater density regions exist because craters have been removed by geological processes and are not low density by Poissonian chance as is implicit in the catastrophic resurfacing model. N. Izenberg (personal communication) has shown, by analyzing the degradation states of the complete set of crater ejecta deposits, that crater degradation and removal has proceeded more rapidly in areas of low crater density. Phillips (1993) proposed that the Venus surface can be divided into at least three distinct regions by crater age: 1. parts of the planet (~25%) that have not been resurfaced for ~0.6–1 Ga, termed "High Crater Density" (HCD) areas, and which are largely associated with volcanic plains; 2. parts of the planet (~25%) with younger than average surface ages ($\lesssim$0.2 Ga), termed "Low Crater Density" (LCD) areas, and which are correlated strongly with the mesolands; and 3. the remainder of the planet (~50%), associated with surface retention ages nearer to the planetary mean.

Additional crater observations bear on the tectonic and magmatic history of the planet (Phillips et al 1992). In general, the oldest 90% of impact craters have dark (low values of specific radar cross section, σ_0) floors and the youngest 10% do not. Dark floor material is widely interpreted as volcanic in origin, and the implication is that magma has been widely available in space and time in the interior. Secondly, many of the regions of low crater density follow distinct extensional tectonic trends, and craters that have been volcanically altered have an unusually high association with craters that have been tectonically altered, more than three times the number expected if the processes act independently. The implication is that volcanism follows tectonism in a passive manner. That is, combining both observations, magmas, or at least incipient partial melts, have existed commonly in the subsurface, and where tensional environments occurred in the lithosphere, volcanism resulted.

Ivanov & Basilevsky (1993) measured crater density on tesserae and compared this to nontessera regions. While the overall crater density for the two classes is the same, tesserae display a deficiency (relative to the planetary mean) in craters with diameters of < 16 km and an excess of craters for diameters greater than this. There is apparently an observational bias for craters of <16 km diameter that tends to underestimate the number of these craters in tesserae. The excess of larger craters implies that the average age of tessera surfaces exceeds the mean surface age of the planet. These results do not mean that all tesserae are older than all nontessera areas (e.g. Thetis Regio is relatively young), but the overall results indicate that, on average, tesserae are amongst the oldest units on the surface, approximately equivalent in age to HCD areas. This view is supported by the observation that 73% of tessera borders are characterized by onlapping (and hence younger) volcanic units (Ivanov & Head 1993).

3. MANTLE DYNAMICS

On Earth, mantle convection controls tectonism and magmatism, which are largely confined to the vicinity of plate boundaries. The bulk of magma is produced at the midocean spreading centers by pressure release partial melting. Lesser amounts are produced near subduction zones and as intraplate magmatism associated with hotspots. Tectonism is largely associated with the interaction of plates (e.g. plate collision, transform boundaries) and in a secondary way by gravity-driven mechanisms related to topography resulting from plate interaction. The key point for our purposes here is that on Earth the styles of tectonism and volcanism relate directly to the fact that the oceanic lithosphere participates in mantle convection. It is the notion that this phenomenon does not take place on Venus that presages a very different style of magmatism and tectonism for that planet. A very brief review of mantle convection will set the stage for this consideration.

3.1 *Mantle Convection*

The heat passed to a planetary mantle from a core and that generated within the mantle by the decay of radiogenic isotopes must escape across the outer surface of the planet. If the forces associated with thermal buoyancy exceed the viscous resistance of mantle rock, then the heat will be passed outward dominantly by convection. Formally, the ratio of these two effects is the thermal Rayleigh number, which must exceed some critical value for convection to proceed. Interfaces that are a barrier to convection form thermal boundary layers in which heat is transferred conductively. The outer boundary of mantle convection in a terrestrial planet is always a thermal boundary layer marking the transition from mantle convective heat transfer to lithospheric conductive heat transfer.

If horizontal velocity is not impeded at the surface of a planet (a "free" boundary condition in which shear stress vanishes), then convection can be very efficient because cold temperatures in the thermal boundary layer can be advected downward and large quantities of negative buoyancy are available to drive the flow. This is the situation for Earth's ocean basins in which seafloor spreading is the top of the mantle convection system. If, on the other hand, horizontal velocity is zero at the surface (a "rigid" boundary condition), then a velocity boundary layer will form. At least the upper part of the thermal boundary layer will not be easily advected downward and convection will be less efficient. Thus the critical Rayleigh number for a rigid boundary condition will be higher than that for a free boundary condition—all other factors being equal—and the convecting region will be hotter and cool more slowly. This is easily demonstrated in parameterized convection models, and is a basic result of calculations using more sophisticated two- and three-dimensional finite element convection codes (Schubert et al 1990, Leitch & Yuen 1991). That portion of the thermal boundary that is not advected in the main mantle flow is still buoyantly unstable and will spawn cold downgoing plumes that will be superimposed on the general circulation pattern of the mantle.

3.2 *Application to Venus*

3.2.1 LONG-WAVELENGTH GRAVITY AND TOPOGRAPHY CONSTRAINTS

Because of their high correlation, the relationship between long-wavelength gravity and topography on Venus is remarkably different from that on Earth (Phillips & Lambeck 1980, Sjogren et al 1983). On Earth, long-wavelength gravity anomalies arise from convective flow in the mantle and from subducting slabs and have little to do with long-wavelength topography (Hager & Clayton 1989). On Venus, the conventional wisdom is that because there is no asthenosphere (see below), long-wavelength topography is a first-order reflection of mantle convection. The idea that long-wavelength topography is "dynamically supported" ("code words" for support by some type of convective flow in the mantle) is rooted in early estimates of the ADCs of large-scale topographic features (Phillips et al 1981).

A simplified version of the argument that large ADC values imply dynamic support of topography goes like this (Phillips et al 1981, Phillips & Malin 1983): If topography is not supported dynamically, it is supported isostatically. The very term "isostatic" means that the compensating mass stays put, i.e. it is static. The most common picture we have of isostasy is the compensation of mountains by crustal roots (depression of Moho) in Earth's lithosphere. At some depth in any terrestrial planet, the idea of static support becomes implausible simply because the temperature is high enough for lateral density contrasts inherent in a compensation mechanism to be unstable against dispersal by creep on geological time scales. The demonstration of implausibility comes from finding an upper bound on temperature to maintain in place a compensating mass at the given depth, and then showing why such a low temperature is unrealistic. The magical depth on Venus has been around 100 km, but this is not a rigorous bound on the maximum depth of static support.

More recent work (Smrekar & Phillips 1991, Simons & Solomon 1993) has shown that large-scale features range over nearly an order of magnitude in ADC (~30–300 km) and that multiple compensation mechanisms are probably operative under many features. However, geoid-to-topography ratios (GTR) cluster into two statistically significant groups (Smrekar & Phillips 1991). Generally, the larger group is associated with volcanic rises and the smaller group with crustal plateaus, as discussed in Section 2. This has reinforced the idea that the former features are linked to mantle plumes, whereas the latter have a large component of crustal and lithospheric thermal compensation. Inversion of long-wavelength gravity and topography data for the relative contributions of compensation mechanisms shows a dominant dynamic component beneath such rises as Atla and Beta regiones and a dominant lithospheric component beneath such crustal plateaus as Ovda and Thetis regiones (Phillips et al 1991, Herrick & Phillips 1992). These gravity analyses have underscored the inherent differences between volcanic rises and crustal plateaus. High resolution gravity data from the *Magellan* mission have shown, however, the possible importance of flexurally supported volcanic constructs to the gravity signals associated with volcanic rises, and this may diminish the contribution of the modeled dynamic components to their associated free-air gravity anomalies.

3.2.2 MANTLE VISCOSITY STRUCTURE Kaula (1990a) has proposed that a relative lack of volatiles in the mantle of Venus compared to Earth may lead to a higher viscosity for the Venusian mantle. However, because of the self-regulation of convection, an increase in activation energy (higher solidus temperature) will be offset partially by an increase in mantle temperature. Given also the effect of a rigid upper surface boundary condition (see below) and higher surface temperature, we expect that overall the viscosity of the convecting Venusian mantle will be little different than that of Earth (see also Turcotte 1993).

Phillips (1986) argued that the large ADC values for some topographic features provide evidence against the presence of a low-viscosity zone or asthenosphere beneath the Venusian lithosphere. If an asthenosphere existed, then ADC values would become quite small, the exact value depending on the magnitude of the viscosity contrast between asthenosphere and the rest of the mantle (Robinson et al 1987). Conversely, the presence of an asthenosphere would place actual compensation depths implausibly deep. Phillips (1990) showed with analytical models how observed ADCs on Venus could not be satisfied in the presence of a low-viscosity zone. The lack of an asthenosphere of Venus might be anticipated because of the lack of water or possibly other volatiles in the interior (Phillips 1986). A good review of the Venus interior water issue can be found in Kiefer & Hager (1991a).

Kiefer et al (1986) argued for a uniform-viscosity Venusian mantle based on matching the predictions of viscous flow models to the observed spectral admittances—the spherical harmonic spectral ratios of gravitational potential to topography. Kiefer & Hager (1991a) showed that topographic features in the Equatorial Highlands (Atla, Beta, Ovda, Thetis regiones) could be explained by models with no asthenosphere and upper mantle/lower mantle viscosity ratios ranging from 0.1 to 1.0. These results depend on the assumption that the topography and geoid anomalies of all of these features are caused solely by upwelling mantle plumes. These suppositions are controversial, as we shall see later. Smrekar & Phillips (1991) argued, however, that the GTRs for volcanic rises alone require the absence of an asthenosphere.

A potentially more robust result is that the observed admittance spectrum is positive up to spherical harmonic degree 18 (Bills et al 1987; see also McNamee et al 1993). The viscosity structure that best reproduces Earth's long-wavelength geoid has the following depth range/relative viscosity structure: 0–100 km/1.0; 100–400 km/0.032; 400–670 km/1; 670–2900 km/10.0 (Hager & Clayton 1989). The depth range from 100 to 400 km represents an asthenosphere. The result of a slightly different model (relative viscosity in the 0 to 100 km range is 10.0) produces a negative admittance spectrum for degrees 4–9 (Kiefer et al 1986, Kiefer & Hager 1991a). Because the low-viscosity region tends to uncouple flow stresses from the surface, over this degree range the geoid contribution from the interior outweighs the contribution from the dynamically deformed surface. The admittance calculations were for a perturbing density distribution that is constant throughout the mantle. This cannot be correct, but it is expected that more realistic density distributions will not change these results qualitatively.

For Earth, it is not clear that this exercise would achieve the expected result—a negative portion of the observed admittance spectrum—owing to the presence of an asthenosphere. Variations in lithospheric and crustal structure are mostly responsible for topographic fluctuations (Hager & Clayton 1989), so even in the presence of an asthenosphere, negative admittance values may not be produced. Thus it could be argued that the presence of a non-negative admittance spectrum for Venus is not necessarily an argument against the presence of a low-viscosity zone or asthenosphere.

However, because the long-wavelength topography and geoid are so well correlated on Venus, this likelihood is diminished. The positive correlation itself is a reasonable piece of evidence that Venus does not possess an asthenosphere.

3.2.3 BOUNDARY LAYER PLUMES We consider both the vigor of boundary layer plumes and their transit time through the mantle.

3.2.3.1 *Vigor of plumes* We expect that the ratio of heat passing across the core-mantle boundary to that generated by isotopic decay within the mantle is less for Venus than for Earth (Phillips et al 1991). Venus does not have an internal dipole magnetic field and this is the basis of the argument; the magnetic dipole moment of Venus is no more than 1% of that of Earth even after adjusting for the different rotation rates of the two planets. The most likely explanations for Earth's geodynamo are necessarily accompanied by an outward heat flux, the bulk of which is associated with the freezing of an inner core. The absence of a magnetic field suggests that a freezing inner core does not exist on Venus, and this is borne out by parameterized convection models (Stevenson et al 1983). A lower core heat flux for Venus would imply more infrequent plumes formed by Rayleigh-Taylor instabilities that develop in the core-mantle boundary layer (Phillips et al 1991).

Interior phase transitions and chemical discontinuities may also form barriers to convection and generate plumes. Steinbach & Yuen (1992) considered the relative abilities of terrestrial and Venusian phase transitions to resist penetration by convection. They found that the greater depths for equivalent phase transitions and the probable presence of a conducting lid increases the likelihood that Venusian convection can flow through phase boundaries. Nevertheless, there is no way to decide whether or not interior phase boundaries in Venus are a barrier, or if they are, whether there is still episodic flow through these boundaries. The issue of phase (and chemical) boundaries is important because it bears on the importance of plumes both in removing heat from the interior and in the tectonic and magmatic history of the lithosphere.

3.2.3.2 *Mantle transit times* How long will it take a plume to traverse the mantle? This question has important implications for the origin of coronae. Stofan et al (1992) suppose that those plumes that form coronae come from the core-mantle boundary. The majority of coronae fall in the diameter range 200–300 km; conservatively, we take the diameter of the proposed causative plume as onc-half the corona diameter, so as plumes reach the base of the lithosphere, they must have diameters in the range 100–150 km. The trick is to launch a plume from the core-mantle boundary and have it arrive at the base of the lithosphere with the right diameter—and then see how long it takes for the plume to form and traverse the mantle.

A formulation of this problem is given by Whitehead & Luther (1975), who describe a growing spherical plume trailed by a cylindrical conduit. To produce plumes of the correct size, the transit times over the viscosity interval 10^{21} to 10^{22} Pa s range from a few hundred million years up to nearly four billion years. This result is based on a dry olivine flow law (Goetze 1978), a strain rate of $10^{-14}\ s^{-1}$, and a plume temperature contrast, ΔT, of 300 K. It would seem implausible that such small plumes could traverse the entire mantle, given also that the mantle will reorganize its flow regime on a time scale of a few 100 Ma or less (e.g. Steinbach & Yuen 1992). Increasing the plume-mantle temperature contrast, ΔT, which decreases the plume viscosity and density, has only a modest effect on the transit time. Reasonable transit times (< 100 Ma) are obtained only for large plumes, the type that might be associated with volcanic rises. This same time interval can also be obtained for the smaller plumes if they are launched from the upper mantle/lower mantle boundary (spinel-perovskite phase transition), assuming a thermal boundary layer exists there.

3.2.4 UPPER BOUNDARY OF CONVECTION *Magellan* radar and altimetry confirmed earlier views that Earth-like plate tectonics does not exist presently on Venus [see Solomon et al (1992) for a thorough discussion]. There are only a few examples of large-scale strike-slip faulting, and there is no evidence of lithospheric spreading such as is found at oceanic spreading centers. By topographic analogy to terrestrial subduction zones, there is evidence for subduction at the edges of some large coronae (McKenzie et al 1992a, Sandwell & Schubert 1992a,b). This hypothesis, discussed elsewhere here, is controversial. Our view is that horizontal lithospheric motion operates presently on a restricted, regional scale (Phillips et al 1991) of order 100 km (Solomon et al 1992).

The absence of large-scale lithospheric motion on Venus implies that a rigid boundary condition at the top of the convecting mantle may be appropriate as long as there is not a weak region (a low-viscosity zone or asthenosphere) separating the convecting region from the lithosphere. We presented several arguments above as to why Venus does not have such a weak zone. However, a rigid boundary condition can only be an approximation for the behavior of flow in the uppermost portion of the mantle.

Flow velocities will decrease upward toward the crust-mantle boundary (Moho) if the surface is immobile and also because the temperature will decrease in the thermal boundary layer. We expect that the upper mantle will flow at ever-decreasing velocity upward, so the imposition of a rigid boundary necessarily must be accompanied by a time scale of interest because over a sufficiently long time interval, even the uppermost mantle will have undergone significant motion. It is possible that the lower crust itself is the zone of decoupling of convective flow from the surface (Buck 1992), and that the lower crust then acts as a velocity boundary layer. Regardless of whether or not the uppermost mantle participates in large-scale flow on geologically interesting time scales, it is buoyantly unstable (unless possibly it contains a magmatically depleted layer; see below) and will initiate cold, downgoing plumes.

The existence of a velocity boundary layer implies a vertical gradient in horizontal velocity, and thus a shear stress acting on the lithosphere in addition to the normal forces associated with buoyancy in the mantle flow regime. In general, normal forces will create dynamic topography and the resulting stresses in the lithosphere relate directly to the gravitational body forces associated with topographic variations. The transfer of shear forces (as well as normal forces) to the overlying lithosphere is the basis of the concept of "convective stress coupling," which is considered to have been a significant element in the tectonic deformation of the Venusian lithosphere (Phillips 1986, 1990). While the flow shear stresses themselves may not be large, they give rise to strong normal stresses acting on vertical planes in the overlying lithosphere. However, flow will be more effectively driven in lithospheric regions, such as the lower crust, if velocity, in addition to stress, is coupled (Ramberg 1968). In the presence of a strong viscosity contrast at the Moho (see Section 4.2.2), neither shear stress nor velocity will be well coupled into the crust. Because viscosity is non-Newtonian, shear stresses will in fact act to maintain crust-mantle decoupling.

3.2.5 THE ROLE OF A MELT RESIDUUM IN THE UPPER MANTLE Phillips & Grimm (1990) proposed that the relative thinness of the Venusian crust could be attributed to the existence of a layer of buoyant melt residuum that has accumulated beneath the lithosphere. If upwelling mantle was unable to penetrate this residuum, then pressure-release partial melting would be inhibited, eventually shutting down completely if the residuum reached a critical thickness. Kaula (1990b) argued that mantle convection should easily penetrate the melt residuum, so that a thick layer would never build up. Work now in progress (Parmentier et al 1993, Smrekar & Parmentier 1993) is examining the conditions for plume penetration of a residuum layer (see also Ogawa 1993). The stability of the residuum depends on the ratio, R, of the chemical density contrast of the residuum (to normal mantle) to the thermal density contrast of the plume (to normal mantle), as well as the ratio, μ, of residuum viscosity to that of underlying mantle. There is a minimum value of R, $R_{\min}$, depending on μ, below which entrainment and mixing of residuum will occur (Parmentier et al 1993). R_{min} will decrease with increasing μ. Because the residuum is colder and more magnesium-rich than the underlying mantle, μ will be greater than unity. It may be that the lithosphere is stable over geological time scales and that only very large plumes, which may be rare, cause disruption. However, Tackley & Stevenson (1993b) find through numerical convection simulations that large-scale convection is able to entrain residuum in downwellings, and they suggest that an equilibrium may exist between generation of residuum by partial melting and its return to the deeper mantle.

The existence of a stable residuum layer has other implications besides limiting crustal growth. Large ADC values for some topographic features could be indicative of plumes trapped at the base of a residuum layer (Smrekar & Parmentier 1993). Also, the residuum layer may become buoyantly unstable and its overturn may have tectonic and magmatic consequences for the lithosphere. Under a very general set of conditions, Parmentier & Hess (1992) showed that a residuum layer in the upper mantle would become buoyantly unstable and that a fraction of it at temperatures above a critical temperature for flow, T_f, would remix with the mantle. In an evolutionary calculation, the planet eventually cools to the point that a stable residuum layer will accumulate. The initial mixing takes place when that part of the layer with temperatures greater than T_f becomes unstable because of the decrease in compositional buoyancy that results when melting is pushed to greater depths as the residuum layer and crust thicken. Episodicity (near periodicity) is a fairly robust feature of the solutions (~300–500 Ma), but the time of onset depends strongly on initial temperature, viscosity, and heat production rate and exceeds the age of the planet in some cases. Subsequently, we will present a model for the formation of Ishtar Terra that uses this instability mechanism.

3.2.6 OTHER SOURCES OF TIME-DEPENDENT BEHAVIOR We separate our discussion of the time-dependence of mantle flow into episodic/chaotic behavior and secular variation.

3.2.6.1 *Episodic/chaotic behavior* Time-dependent, chaotic behavior is a basic feature of high Rayleigh number convection (e.g. Schubert et al 1990, Leitch & Yuen 1991) and would lead to variable conditions for tectonism and magmatism at several temporal scales over the history of Venus. We have concluded that the viscosity of the Venusian mantle is no higher, and therefore the Rayleigh number is no lower, than that of Earth. We expect time-dependent behavior to have been an important feature of mantle convection over the lifetime of Venus. Arkani-Hamed et al (1993) have argued that the proposed global resurfacing event ~500 Ma ago marks the transition from high Rayleigh number chaotic convection to low Rayleigh number steady, sluggish convection. The argument is based on the supposition that Venus has cooled more rapidly than Earth because the high surface temperature of Venus had allowed the lithosphere prior to 500 Ma ago to fully participate in mantle convection. This assumes, of course, that the atmospheric greenhouse has been in place during most of Venus history. This free boundary condition allowed Venus to cool more rapidly and drop below the transitional Rayleigh number for chaotic behavior. Curiously, these authors argue that despite seafloor spreading and subduction, Earth has operated with a rigid boundary condition and thus has not cooled sufficiently to remove chaotic, time-dependent convection.

Other sources of episodicity include transitions between layered convection controlled by phase boundaries and whole mantle convection. Steinbach & Yuen (1992) speculate that a transition to whole mantle convection, which would be accompanied by a large vertical mass flux in the mantle, could lead to (significant) partial melting and large-scale resurfacing.

3.2.6.2 *Secular behavior* Is Venus "running down" in any sort of secular fashion? Certainly the rate of heat production in the interior is decreasing, as it is for Earth (and all planets). Solomon (1993) proposed that in the lower crust of Venus, the exponential dependence of strain rate on temperature would lead to a rapid cessation of tectonism as the planet went through gradual secular cooling. This shutdown in tectonism would be delayed in the highlands, which are hotter and/or have a thicker crust. This hypothesis was formulated to explain the putative pristine nature of craters in the plains and supposes that prior to approximately 500 Ma ago tectonic strain removed craters very rapidly. It is in contradistinction to models that require episodic events (e.g. Parmentier & Hess 1992, Arkani-Hamed et al 1993) to provide a catastrophe, the latest (or only) one occurring 500 Ma ago. The Solomon model proposes that the critical temperature in the crust required to rapidly decrease the strain rate was reached $\sim$500 Ma ago beneath the plains. Subsequently, the tectonically deformed plains were covered by volcanism to hide the evidence. This volcanism must have been fairly rapid ($\sim$50 Ma) to match the relatively small number of embayed craters; thus the model does not seem to completely avoid a volcanic catastrophe that might still require some profound event in the mantle.

Grimm (1993), following the wisdom of Solomon's basic idea, performed a specific calculation using the secular decrease in strain rate in a convecting mantle. The reciprocal of the present time scale for surface retention ($\sim$500 Ma) provides an upper bound of the presently operative strain rate, and is $O(10^{-16}\ \mathrm{s}^{-1})$. Because only a fraction of the craters are even partially faulted, the actual strain rate is at least an order of magnitude less than this. If craters are to have been removed rapidly during an earlier episode, then the strain rate must be of order one-tenth the retention age (see Section 2.4) or $10^{-15}\ \mathrm{s}^{-1}$. Therefore the strain rate must have decreased by a factor of 100 or more over some characteristic interval, say 1 Ga. Using parameterized convection techniques, Grimm shows how the net cooling of the mantle attributable to a decline in heat production leads to a corresponding decrease in strain rate by more than a factor of 100 over 1 Ga. However, the crustal strain rate change is diminished sharply, relative to the mantle, owing to its lower creep activation energy, and a mechanism must be available to transmit upper mantle strain rate to the upper crust.

Parameterized convection calculations based on the approach of Phillips & Malin (1983) yield a mantle strain rate change of about a factor of 50 at the crust-mantle boundary over the past 1 Ga. This particular solution produces the Earth-scaled value of surface heat flow of 74 mW m^{-2} (Solomon & Head 1982a). Here, the Moho is within the thermal lithosphere, whose base is defined by the 1500 K isotherm. Over 1 Ga, the Moho temperature decreases from 1176 K to 1061 K. By equating resurfacing rate to strain rate, it can be shown that the present number of craters is consistent with a model in which the resurfacing rate has been declining exponentially, and prior to 1 Ga ago craters were rapidly destroyed.

Grimm (1993) proposes that the corresponding crustal tectonic effect does not take place on Earth because of the uncoupling of mantle convection from the lithosphere resulting from the presence of a low-viscosity zone or asthenosphere (see Phillips et al 1991). Conversely, we argue that convective flow is strongly coupled to the lithosphere on Venus (Phillips 1986, 1990) and its effects would have been greater in the past because of greater heat production in the mantle. Indeed, the inference that tesserae could underlie much of the plains (Ivanov & Head 1993) is consistent with the concept of a decline in surface tectonic activity directly associated with a secular decrease in strain rate, as proposed by Solomon (1993). A decline in tectonism also implies a decline in volcanism because of a decrease in the number of extensional pathways for magma access to the surface.

3.3 *Summary*

We summarize an interior model for Venus. Our confidence level varies for different aspects of the model. We state with reasonable conviction that:

- If both Earth and Venus have the same heat source concentrations, then the interior convecting temperatures and mantle viscosities for these planets are about the same.

- Venus lacks an asthenosphere beneath its lithosphere and shear coupling of mantle convection to the overlying lithosphere is important, which is not the case for Earth.
- Like Earth, convection in the Venusian mantle is both time-dependent and chaotic.
- The strain rate in the mantle has decreased with time; this is reflected at the surface as a secular decrease in tectonism and volcanism.

With relatively less certainty we state that:

- A thermal boundary exists at the core-mantle boundary that initiates large, hot plumes infrequently.
- A residuum layer has existed from time to time in the upper mantle and has had a strong effect on the tectonic and magmatic history of the planet.

At the level of speculation we suggest that:

- Interior phase boundaries act as thermal boundary layers and can initiate plumes, which at times have carried upward a significant amount of the interior heat of the planet. These phase boundaries at times may have stratified flow in the mantle and at other times may have given way to whole-mantle convection.

4. LITHOSPHERIC DYNAMICS

4.1 *Rheological Considerations*

Models for the mechanical behavior of the interior structure of Venus rely on topography and gravity as boundary conditions and plausible postulates about rheology and temperature (unless the latter is considered a free variable). For lithospheric studies, rock strength is assumed to be controlled approximately by the weaker of frictional sliding or viscous (ductile) flow mechanisms (Brace & Kohlstedt 1980); strength as a function of depth can be characterized by a yield strength envelope (YSE). The ductile portion of the YSE defines the stress at which the rock will flow at the adopted strain rate (for a given temperature). The YSE extends to the base of the mechanical lithosphere (BML), which is defined somewhat arbitrarily (e.g. when the yield stress reaches 50 MPa). While frictional strength on preexisting fractures is largely independent of composition (Byerlee 1978), viscosity is not. Under an assumption of a basaltic crustal composition, the viscosity contrast between the crust and mantle (assumed olivine rheology) portions of the lithosphere is large, so lithospheric strength is strongly dependent on estimates of crustal thickness, as well as temperature gradient.

Our understanding of the tectonics of Venus depends strongly on our estimates of lithospheric structure and strength. Hypotheses regarding the support of large mountains, the detachment of the upper crust, subduction, and episodic plate tectonics depend implicitly, if not explicitly, on assumptions regarding the mechanical state of the lithosphere.

4.1.1 CRUSTAL THICKNESS AND TEMPERATURE GRADIENT Estimates of crustal thicknesses on Venus have come primarily from two sources: (*a*) the spacing of tectonic features, and (*b*) the depth of impact craters. These methods yield crustal thicknesses in the range of 20 km or less. Grimm & Solomon (1988) found an upper bound for crustal thickness of 10 to 20 km based on the application of linear viscous relaxation models to the depths of impact craters measured altimetrically by the *Venera* 15/16 spacecrafts. Because crust is most likely weaker than the underlying mantle and because impact craters show little evidence of viscous relaxation, a relatively thin crust is required by the modeling. However, consideration of non-Newtonian rheology and a viscoelastic lithosphere may enable thicker crusts to accommodate the crater depth observations (e.g. Hillgren & Melosh 1989).

Zuber & Parmentier (1990; see also Zuber 1987) applied models of uniform horizontal compression and tension of rheologically stratified lithospheres to examine instability growth as a function of wavelength. Wavelengths with the fastest growth times are related to spacing of tectonic features to constrain model parameters. Rift zones and ridge belts display two horizontal scales of deformation (e.g. for ridge belts, the wavelengths are defined by belt spacing together with the wavelength of structures within the deformed belts) indicating that both a strong upper crust and a strong upper mantle control deformation (as required to produce two maxima in the growth-rate spectra). Results for these features indicate crustal thicknesses of less than 20 km and temperature gradients, dT/dz, of less than 25 K/km. In Ishtar Terra, the mountain belts surrounding Lakshmi Planum have only a single, short (<20 km) wavelength of deformation indicating a crust greater than approximately 20 km in thickness and, because a long-wavelength instability has not developed, a mantle that is weak. This follows logically because the presence of a thick crust elevates the temperature at the Moho. Banerdt & Golombek (1988) also examined rift zone and ridge belt horizontal scales of deformation using classical elastic plate theory modified by more realistic considerations of rheology as described by strength envelopes. They find crustal thicknesses ranging from 5 to 15 km and dT/dz ranging from 5 to 15 K/km. In the Ishtar Terra mountain belts, they also find that the crust may be "considerably thicker" than 15 km.

These last two studies by necessity did not cover all regions of Venus, but converge to similar answers for crustal thickness: less than 20 km. Further, they support the notion that a "jelly sandwich" rheology exists in places, consisting of a strong upper crust overlying a weak lower crust overlying a strong upper mantle. The analyses also suggest that Ishtar Terra may be a region of thicker crust than the plains. Because in general there is a single wavelength of deformation in both the mountain belts (e.g. Akna, Freyja, Maxwell Montes) and the outboard tesserae (e.g. Atropos, Itzpapalotl, Fortuna tesserae) of Ishtar Terra, instability analyses (Zuber & Parmentier 1990) are not able to resolve crustal thickness variations across this region.

The upper bound on temperature gradients found from these studies is not as interesting as the crustal thickness estimates because it includes the average temperature gradient for Earth and, as we shall see below, a more interesting result would have been the lower bound for this parameter.

4.1.2 LITHOSPHERIC THICKNESS AND TEMPERATURE GRADIENT Considerations of both elastic and elastic-plastic rheology of the lithosphere yield estimates of lithospheric thickness and temperature gradient.

4.1.2.1 *Elastic thickness estimates* A planetary lithosphere loaded from above, below, or within will deform in response to the applied load. The resulting topographic profile can be used to constrain the rheological properties of the lithosphere. In the most basic of studies, the lithosphere is treated as a bending elastic plate of thickness T_e and parameters are adjusted until the modeled plate flexure matches the observed topography. Estimates are obtained of the magnitude of the applied loads and of the flexural rigidity of the elastic lithosphere given by $D = ET_e^3/[12(1-\nu^2)]$, where E and ν are Young's modulus and Poisson's ratio, respectively.

Head (1990) proposed that the Freyja Montes mountain belt and the region to its north is a site of ". . . large-scale crustal convergence and imbrication involving the underthrusting of the Northern Polar Plains beneath Ishtar Terra." In this region, Solomon & Head (1990) interpreted the topographic rise north of the linear depression flanking Uorsar Rupes as flexural in origin. Fitting topographic profiles with a broken-plate model, they obtained values of T_e in the range 11–18 km by adopting values of E and ν of 60 GPa and 0.25, respectively. In another study, Sandwell & Schubert (1992b) fit flexural profiles to the outboard topographic highs of four coronal trenches and obtained elastic thickness estimates of 10 to 60 km.

Cyr & Melosh (1993) predicted the deformation patterns around several coronae by combining regional, uniform stress fields with the local stress fields of early- and late-stage models for coronae. They obtained elastic thickness estimates of 10 ± 5 km (and 2 km in one case) and regional stress field magnitudes of approximately 10 to 60 MPa. This thickness result is consistent with the flexural solution obtained in northern Ishtar Terra, but less by factors of two to four than thickness estimates for Artemis, Latona, and Heng-O coronae.

4.1.2.2 *Elastic-plastic models* The meaning of an elastic thickness is not easily understood in terms of a stratified rheology, even in terms of an elastic core. A preferable alternative to simple elastic formulations is to solve the bending problem by treating the lithosphere as an elastic-plastic plate with plastic yielding defined by a yield strength envelope (Bodine et al 1981, Phillips 1990); the solutions then depend on dT/dz. This more complicated approach is possibly circumvented by equating the elastic and elastic-plastic bending moments for a given plate curvature (McNutt 1984). The latter moment is obtained by integrating the stress-distance product from the neutral plane using the lesser magnitude of either the yield stress or the stresses in the elastic cores of the crust and mantle.

Both of the flexure studies cited above ignored the crustal contribution and adopted the moment-matching technique (at the first zero-crossing "seaward" of the trench axis) to obtain temperature gradients. For the Freyja region, Solomon & Head (1990) and Sandwell & Schubert (1992b) estimated gradients of 14–23 K/km and 9.5–26 K/km, respectively. For the coronae Eithinoha, Heng-O, Artemis, and Latona, ranges of 6.8–24 K/km, 5.8–8.9 K/km, 3.2–4.8 K/km, and 3.5–7.7 K/km, respectively, were found by Sandwell & Schubert (1992b).

Plate curvatures for Eithinoha, Latona, and particularly Artemis are large, indicating that lithospheric bending is close to or at plastic saturation; here the moment-matching approximation breaks down. Fully elastic-plastic solutions for Artemis and Latona yield values of about 2 and 4 K/km, respectively. These are upper bounds neglecting the weak crust. Small temperature gradients (i.e. thick mechanical lithospheres) are required to support the large bending moments of these features in the face of large plate curvatures.

Johnson & Sandwell (1993) updated the earlier work of Sandwell & Schubert (1992b), expanding the number of coronae studied to twelve. Seven coronae yielded mechanical thickness estimates of less than 30 km using the McNutt technique. This is in good agreement with the mechanical thickness estimate obtained by Phillips (1990) for a dry olivine rheology, strain rate of 10^{-14} s^{-1}, and linear temperature gradient of 15 K/km (consistent with Earth-scaled expectations of dT/dz). Four of the remaining five coronae are too close to plastic saturation to obtain estimates using the moment-matching scheme.

What should we make of estimates of elastic lithosphere thicknesses (T_c) and dT/dz? First, we see a wide range in estimates of temperature gradients. While we should expect spatial variation in dT/dz, values as low as 2 K/km are certainly suspect (if not guilty of being obviously wrong). For those coronae where curvature indicates a large fraction of saturation in an elastic-plastic rheology, flexural interpretations should be questioned, and it seems worthwhile to consider alternative mechanisms. Relaxation of uncompensated coronal loads is one possibility (Janes et al 1992, Smrekar & Solomon 1992), and the formation of topography by mantle plume–head waves is another (Bercovici & Lin 1993). Generally, temperature gradients inferred at other locales appear to fall within our expectations of Earth-scaled values. However, given models for the origin of coronae (see Section 5.2.1), such regions of the lithosphere must have had anomalously high temperature gradients over all or part of their history.

4.1.3 ABSOLUTE CRUSTAL VISCOSITY There are several models for Venusian topography and tectonics that assume the lower crust can flow large distances on geologically reasonable time scales (e.g. Buck 1992, Bindschadler & Parmentier 1990). Our ability to understand the mechanical behavior of the crust is frustrated by the absence of a reliable flow law for proposed Venusian crustal material. Compositional measurements in the plains of Venus by *Venera* and *Vega* landings indicate a basaltic composition (Surkov et al 1984, 1987) and for that reason the Maryland diabase stress-strain rate measurements of Caristan (1982) have been used often for a flow law for the Venusian crust. There is reason to believe that this flow law provides, at best, a lower bound on the ductile strength of the presumably dry crust (Smrekar & Solomon 1992). For Maryland diabase, an increase in activation energy of 60% would remove the ductile regime in a 20-km-thick crust (temperature gradient 15 K/km; strain rate 10^{-14} s^{-1}). Zuber's work shows, however, that in the plains the lower crust can be weak (relative to the upper crust), and thus it's deformation must be controlled by its ductile strength and not its brittle strength. This constrains a tradeoff (for a given crustal thickness and strain rate) between the allowable bounds of activation energy and temperature gradient, two pairs of which are given above for Maryland diabase.

4.1.4 SUMMARY Our understanding of the rheological and thermal properties of the Venusian lithosphere is meager. For constructing models of tectonics we adopt the most likely set of parameters based on analyses to date:

1. Crustal thickness: $\lesssim 20$ km in the plains.
2. Layered rheology: Strong upper crust, weak lower crust, strong upper mantle.
3. Rheology of Ishtar mountain belts (and outboard tesserae): Thick crust overlying weak upper mantle.
4. Nominal temperature gradient: 10 K/km $\lesssim dT/dz \lesssim$ 20 K/km.
5. The flow law applicable to the Venusian crust might be much stronger than Maryland diabase.

4.2 *The Role of Lower Crustal Flow*

Lower crustal flow can be induced in two ways: (*a*) by the pressures associated with topographic gradients (Weertman 1979, Smrekar & Phillips 1988, Smrekar & Solomon 1992), and (*b*) by direct coupling of velocity or shear stress associated with mantle convection (Bindschadler & Parmentier 1990; Bindschadler et al 1990, 1992b; Kiefer & Hager 1991b; Buck 1992).

4.2.1 TOPOGRAPHICALLY DRIVEN FLOW The most straightforward model of crustal dynamics is flow driven by topographic variations. An application of this problem to Venus was carried out by Smrekar & Solomon (1992) in their study of gravitational spreading of high terrain in Ishtar Terra. They used a finite element code ("TECTON," developed by H.J. Melosh and A. Raefsky) that allows analysis of deformation with a visco-elastic non-Newtonian rheology. Relaxation of a crustal plateau was considered by parameterizing the flow law, plateau boundary slope, crustal base boundary condition, crustal thickness, and temperature gradient. Their models indicate the importance of plateau relaxation by lower crustal flow, accompanied by only minor strain at the surface. Using a Maryland diabase flow law and a 15 K/km temperature gradient, Smrekar & Solomon found that for crustal thickness greater than 10 km, failure in a plateau (3° boundary slope, rigid basal boundary condition) higher than 3 km occurred in less one million years; relaxation of the plateau by 25% occurred in less than a few hundred million years. Relaxation times decreased markedly with increasing crustal thickness because the lower part of thc crust became hotter. The results obtained are broadly consistent with the observed relaxation of some parts of the Ishtar mountain belts (e.g. Freyja), as inferred from the development of normal faults. More interestingly, they offer the choice that orogeny is recent, or that the crust is a great deal stronger than Maryland diabase, or both. For example, with a websterite flow law (Ave Lallement 1978), 25% relaxation had not been achieved by 1 Ga for the model parameters described above.

4.2.2 HOW WELL DOES CONVECTIVE FLOW COUPLE INTO THE LOWER

CRUST? Bindschadler & Parmentier (1990) analyzed the deformation of a rheologically stratified lithosphere subjected to flow stresses imposed by a density-driven flow representing the effects of sublithospheric convection. The formalism for this problem had been developed earlier (e.g. Ramberg 1968). The results show the importance of an interplay of buoyancy-driven stresses related to dynamic topography with forces introduced by flow and/or shear stress related to convective coupling. In particular, emphasis was given to the concept of a two-stage process for lithospheric response to convective stresses. Considering a downwelling or downgoing cold plume, the stages are: 1. A relatively rapid depression of the surface takes place until the lithosphere and the mantle flow are mutually compensated; this takes place with a characteristic time constant τ_2. 2. The crust then thickens by inward flow with a characteristic time constant τ_1. In isostatic calculations, thc time constants τ_2 and τ_1 correspond, respectively, to the characteristic time it takes a topographic disturbance to become compensated and the characteristic time it takes for the compensated system to relax gravitationally. The net result is positive topography over a downwelling (see also Grimm & Phillips 1991). This has led to the idea that some (Bindschadler et al 1992b) to all (Buck 1992) Venusian highlands are over mantle downwellings. Bindschadler in particular has promoted the origin of crustal plateaus by crustal thickening over downwellings, emphasizing that this process is consistent with interpretations of early contractional features observed at crustal plateaus. Earlier we argued that crustal plateaus are altered by complex, polyphase deformation recording many extensional and contractional events.

The question is not whether crust is able to thicken over downwellings (or thin over upwellings), but rather how long it will take. In lithospheric response to perturbations from below, the crust acts as a filter, and most of the topographic energy lies in wavelengths of order 100 times the crustal thickness or greater. For a crustal thickness of 20 km, Bindschadler & Parmentier (1990) show that the long-wavelength response time (one-half steady-state amplitude) for crustal thickcning ovcr a downwclling cxcccds one billion years for a constant viscosity model of 10^{21} Pa s. These authors also considered more realistic stratified rheologies; for a strong upper crust—weak lower crust—strong upper mantle, they found that thickening occurs only at very long wavelengths and at very long times. For example, for a 20-km-thick crust and a mantle viscosity beneath the strong mantle layer of 10^{21} Pa s, the wavelength that gives one-half the amplitude of the long wavelength limit is 1.26×10^5 km (obviously not very realistic!) and corresponds to a characteristic time τ_1 in excess of two billion years. These large values result because the upper mantle acts as a quasi-rigid boundary and flow is not efficiently coupled into the crust.

4.2.3 LARGE-SCALE FLOW The work of Zuber (1987) demonstrated that the lower crust of Venus has behaved in a ductile manner, deforming locally into boudinage structures. A number of models (e.g. Bindschadler et al 1992b), however, call for large-scale crustal flow driven by mantle convection. By large-scale, we mean that flow is able to take place over a horizontal scale that is many times the thickness of the crust. We summarize arguments against such a mechanism.

1. Arguments for crustal thickening invoking observations of radially-limited azimuthal contractional features associated with crustal plateaus or Ishtar Terra must also account for the full space-time history of strain, which will lead to many styles of faulting.
2. Building crustal plateaus by lower crustal flow requires transport distances that are exceedingly large. For example, material required to thicken the crust at Ovda Regio must travel in the lower crust (~ 10 km thick) at least 4000 km on either side of the Ovda structure. This must happen without surface evidence of the process.
3. An extremely weak lower crust would act as an asthenosphere; buoyant stresses from beneath the crust would be attenuated, leading to weakened surface uplift. By the same arguments we presented above regarding the ADCs estimated from gravity data, it is unlikely that a crustal asthenosphere exists. The lithosphere responds as an essentially coherent unit to buoyancy forces from below.
4. Models that require second-phase thickening or thinning of crust must have surface elevations that pass through zero during this stage. The corresponding ADC (or GTR) will pass through a singularity and then be negative for up to a billion years or more, eventually recovering to a small positive value (Bindschadler & Parmentier 1990, Simons et al 1992). Negative ADC or GTR values are not observed anywhere on Venus (Smrekar & Phillips 1991; M. Simons, personal communication).
5. The persistence of crustal plateaus over long periods of geological time, as inferred from the tessera study of Ivanov & Basilevsky (1993), argues that crustal compensation also persists for long periods, i.e. τ_1 is large.

4.3 *Is Crust Recycled?*

Is crust recycled on Venus? If it is not, why is the typical crustal thickness on Venus no greater than about 20 km? There would seem to be several ways for Venus to recycle crust. A tried-and-true method is to subduct the entire lithosphere, which works quite well for Earth but only in the case of oceanic crust. Continental crust seems quite happy to stay put for billions of years, although the upper part is removed by erosion and eventually recycled and some of the lower crust may be removed in places by delamination. For Venus, lithospheric subduction has been proposed around large coronae, as discussed earlier; we present arguments against this hypothesis in Section 5.2.2. If the crust is thick enough it will enter the garnet granulite or eclogite stability fields and become buoyantly unstable; however, 20 km is not thick enough under a plausible range of temperature gradients. The base of crustal plateaus, if they are compensated by deep roots, may contain these higher density phases. This would not seem to be an efficient way to recycle crust, given the limited number of crustal plateaus in existence today. Convection models have shown that the sinking of overthickened boundary layers beneath crustal plateaus could possibly entrain crust (Lenardic et al 1991, 1993). This phenomenon could take place independently of the origin of crustal plateaus (subsolidus or supersolidus flow); all that is required to overthicken the boundary layer is the thermal blanketing effect of thicker crust and co-operating lithospheric viscosities (which may be unlikely).

In Section 3 we reviewed several models that have been put forward to provide an explanation for the catastrophic resurfacing hypothesis (Schaber et al 1992). While we doubt that a catastrophic resurfacing event took place 500 Ma ago (see Section 2.4), the models are in and of themselves interesting as potential initiators of episodic tectonic and magmatic phenomena. Most of these models have implicit if not explicit recycling of crust. The alternative, of course, is that Venus does not make very much crust and does not return very much crust to the mantle.

5. ORIGIN OF TECTONIC AND MAGMATIC FEATURES

5.1 *Origin of Highlands*

Ishtar, crustal plateaus, and volcanic rises are the main highland types on Venus. Here we review ideas for their origins, and possible linkage. Ishtar Terra is unique on Venus, and possibly requires that some unique aspects come into play during its evolution [the "special pleading" of Solomon et al (1992)].

5.1.1 VOLCANIC RISES As noted in Section 2.2.3.1, it has long been argued that volcanic rises on Venus mark the sites of upwelling mantle plumes, i.e. volcanic rises are hotspots. We see nothing in particular that would replace this notion with a stronger, alternative hypothesis (e.g. all highlands on Venus are associated with downwelling and crustal thickening (Buck 1992)).

Despite the name, volcanism is limited for the most part in these regions and is expressed largely as shield volcanoes (and their attendant flows) associated with rifting (Senske et al 1992). Beta Regio is particularly striking in that it is perhaps the only extensively high area on the planet with a greater than average impact crater density (Phillips et al 1992). Along with the presence of tesserae, the distinct impression is of uplift and rifting of an ancient surface. Other volcanic rises display relatively less tesserae and more shield volcanism (Senske et al 1992), and have crater densities at or less than the planetary average. All in all, the implication is that buoyant uplift from below is responsible for volcanic rises but that partial melting is taking place at relatively great depths, producing large volcanic constructs but otherwise being relatively feeble. Herrick & Phillips (1990) and Phillips et al (1991) proposed that the plumes beneath volcanic rises had not reached the base of the lithosphere to enter a stage of massive partial melting. The most volcano-free of the rises, Beta Regio, also has the largest ADC.

In support of the plume hypothesis, Smrekar & Phillips (1991) showed that the most likely mode of compensation for volcanic rises is dynamic. A detailed study yielded the same result for western and central Eistla Regio, and was also consistent with observed tectonics (Grimm & Phillips 1992). Kiefer & Hager (1991a) showed that even the very large ADC value for Beta Regio (~ 300 km) could be fit by a convection model with a high viscosity lid.

What are the remaining issues in the hypothesis that plumes are responsible for volcanic rises? First, we must still assume that there is a hot thermal boundary layer in the mantle to generate plumes. Volcanic rises, of order 2000 km in diameter, number less than a dozen and are of uncertain age. This is consistent with our earlier contention (Section 3.2.3) that core-mantle boundary layer plumes are large and infrequent. Transit time through a 10^{23} Pa s viscosity mantle for a 1000-km-diameter plume (to make a 2000-km-diameter volcanic rise) is 65 Ma. The interpretation that plumes are generated at the upper mantle–lower mantle phase boundary seems less likely because—coronae notwithstanding (see Section 5.2)—hotspots would be much more numerous (and smaller).

Secondly, the contention that volcanic rises are (ironically) relatively magmatically dry and have large ADCs begs for an interpretation beyond the idea that all attendant plumes have been caught in various stages of not quite reaching the base of the lithosphere. While convection modeling shows (nonuniquely) that steady state plumes can account for large ADC values (Kiefer & Hager 1991a), the lack of magmatism is unexplained. The possibility that plumes are trapped at the base of a residuum layer (Phillips & Grimm 1990, Smrekar & Parmentier 1993) might provide a reason for this observation.

Finally, we note that the higher resolution and higher signal-to-noise ratio of *Magellan* gravity data ovcr *Pioneer Venus* gravity data should introduce some caution into earlier interpretations of free-air anomalies in terms of large depths of compensation. In particular, it is becoming increasingly clear that the gravity signals over volcanic rises could contain a near surface component from flexurally supported volcanic constructs; this could compromise simple interpretations of large ADC values.

5.1.2 CRUSTAL PLATEAUS Here we consider how crustal plateaus are formed, and how they are deformed.

5.1.2.1 *Building crustal plateaus* Crustal plateaus are variably interpreted as surface expressions of major subsolidus flow in the crust in response to mantle downwellings (e.g. Bindschadler & Parmentier 1990, Bindschadler et al 1992b) or crustal thickening by magmatism related to late-stage evolution of mantle upwellings (Herrick & Phillips 1990, Phillips et al 1991). This disagreement has developed into the "hotspot-coldspot controversy" for the origin of crustal plateaus, which is a variation of the argument regarding the time scale of significant lower crust flow. Each of these end-member models has serious shortcomings.

Our objections to the downwelling or coldspot model have essentially been given in Section 4.2. The sine qua non of the model requires extremely large amounts of crustal flow inward over a cold mantle downwelling (hence coldspot) to create a plateau. Our main arguments against this are: (*a*) the extremely long time scales required, (*b*) the observed polyphase deformation that neither coldspot or hotspot models explain, (*c*) the lack of negative GTRs, (*d*) the extremely long distances that lower crust must bc transportcd to build the larger crustal plateaus, and (*e*) gravitational evidence against the presence of a crustal asthenosphere.

In the hotspot model, volcanic rises evolve into crustal plateaus because of massive partial melting of the plume head—a process proposed for the formation of basaltic plateaus (oceanic and continental) on Earth (e.g. Richards et al 1989). Objections to this model include: (*a*) the aforementioned problem of generating polyphase deformation, (*b*) the need to do away with large shield volcanoes (because thcy are generally not associated with crustal plateaus), and (*c*) the lack of evidence for transitional forms between volcanic rises and crustal plateaus. This last point is not only evident from geological observations but also from the clustering of GTRs into two statistically distinct families associated with the two feature types (see Section 3.2.1).

The hotspot and coldspot models do agree that crustal plateaus are indeed regions of thickened crust but formed by supersolidus and subsolidus flow, respectively. The plateau-shaped physiography suggests that the plateau crust is thicker and/or less dense than plains crust, and that plateaus stand above the Venusian plains much as continents stand above ocean basins on Earth. ADC/GTR values for crustal plateaus suggest a large component of lithospheric compensation (Smrekar & Phillips 1991, Simons & Solomon 1993). Additionally, both models also propose that smaller crustal plateaus such as Alpha, Tellus, and Phoebe regiones, which lie 1–2 km above adjacent plains, evolve from larger crustal plateaus such as Ovda and Thetis regiones, which lie 4 km above adjacent plains. The smaller features represent a state of quiescence in which the plateau is no longer being actively built and so is collapsing gravitationally into the plains. Indeed, patches of tesserae observed in the plains would be the death rattle of crustal plateaus, which ultimately become entombed in plains volcanism.

It is not clear that this view is correct. The conclusions of Ivanov & Basilevsky (1993) that crustal plateaus are old and have undergone little gravitational relaxation over the retention age of their crater population of ~700 Ma would suggest that crustal plateaus stay put. If much of the surface of Venus was actively deforming prior to ~1 Ga ago, as outlined in Section 3.2.6.2, then exposed, low-lying patches of tesserae may simply have been low to begin with. Under this view, there is nothing special about the tectonic deformation of crustal plateaus compared to the plains other than their height, which makes it more likely that they would not be covered by volcanism today. However, patches of tessera in the plains often strongly resemble the high-standing, arcuate boundaries of crustal plateaus (Head & Ivanov 1993). Either these tessera patches represent relaxed crustal plateaus, or large-scale dynamical lowering of the surface in places has allowed the embayment of these features. Two other mechanisms could lead to the lowering of crustal plateaus: Simple cooling (following a magmatic origin) which would decrease elevation isostatically in analogy to Earth's oceanic lithosphere, or the development of denser phases in the root of a crustal plateau which would also lead to an isostatic decrease in elevation.

Our hypothesis for crustal plateaus is that 1. they are built magmatically, and 2. that the described sequence from large to small plateaus to tessera patches is essentially correct, but the time scale for this process can be exceedingly long. Our adoption of the former point is the natural outcome of our rejection of large-scale subsolidus crustal flow. However, we do not subscribe to the view that crustal plateaus are always associated with strong upwelling plumes that at first lead to volcanic rises, especially because of objection (*c*) to the hotspot model. Rather, we suggest that volcanic rises and crustal plateaus result from distinctly different magmatic styles.

In the simplest view, we propose that crustal plateaus result from the massive partial melting that can take place in a hot Venusian mantle at relatively shallow depths. McKenzie & Bickle (1988, Figure 7) show how a column of melt ranging from several to many tens of km thick could be produced (at the convecting temperatures of the Venusian mantle) by melting garnet peridotite in a mantle up to a minimum depth ranging from 100 km to 0 km, respectively.

There are two mechanisms that might lead to shallow upwelling of mantle, and thus extensive partial melting. First, crustal plateaus could be caused by plumes, but the difference could be that residuum blocks the plume at depth in the case of a volcanic rise and limits the amount of partial melting. Why should there be a difference? We speculate that either residuum is not pervasive in the upper mantle and volcanic rises, and crustal plateaus occur where residuum is and is not, respectively, or that crustal plateaus are associated with only those plumes hot enough to penetrate residuum.

The second mechanism recognizes that most of the outward convective heat transfer in the Venusian mantle is caused by internal heating and thus takes the form of diffuse upwellings whose temperature modestly exceeds the average convecting temperature of the mantle. Lithospheric stretching following buoyant uplift allows diffuse upwellings to rise to relatively shallow depths. The resulting partial melting adds to the thickness of crust, and the accompanying edifice stresses maintain the lithosphere in tension, which continues the upwelling and melting processes. This is essentially the hypothesis proposed by Solomon & Head (1982b) for the creation of the mammoth Tharsis plateau on Mars by magmatic construction. Once established, this is a passive phenomenon that allows melting to continue. The occurrence of extensional structures throughout Aphrodite Terra (and forming the southern boundaries of Ovda and Thetis regiones) supports the view that crustal plateaus are formed by the mechanism described here and may be linked to mesolands formation.

5.1.2.2 *Deforming crustal plateaus* The complex, polyphase deformation observed for crustal plateaus suggests a long history of deformation resulting from distinct episodes of contractional and extensional strain in the lithosphere. Solomon et al (1991, 1992) proposed that crustal plateaus may undergo deformation from strain supplied exogenically. Part of the evidence lies in the observation that for some crustal plateaus it is clear that boundary deformation can be traced into the surrounding plains. Once a crustal plateau is created, by whatever means, it may represent a relatively weaker portion of the lithosphere because it has a thicker than average crust (as implied by small ADC values). Because of their inherent strength boundaries, crustal plateaus may deform significantly in response to large-scale regional strain fields that may wax and wane on Venus over geological time. A crustal plateau may act as a "strain magnet" (Grimm & Phillips 1990), recording the complex history of strain in any given region. Whatever the level of deformation in the original formation of a crustal plateau, it has been overprinted significantly by other events under this hypothesis. Thus the tectonics of a crustal plateau would not be a guide to its origin. This exogenic deformation may not have been constant over time for the presently observed population of crustal plateaus. Rather, most of the deformation we see may have taken place early, marking the end of a prolonged period of high strain rate in the crust (e.g. Solomon 1993, Grimm 1993, Ivanov & Basilevsky 1993).

Under the concept of a strain magnet, the boundaries of crustal plateaus should exhibit the most recent episodes of deformation because they mark the transition in lithospheric strength. Observationally, this is borne out. The polyphase deformed interiors (amorphous terrain) and the more structurally coherent folds belts located along portions of the margins of these regions suggest that crustal plateaus have deformed progressively outward in time (including modest accretion of surrounding crust). The highest topography marking the outer limits of the crustal platcaus may represent the youngest deformation, or the most recently deformed crust. Intratesseral plains may preserve undisturbed regions along the boundary that were trapped between successive deformed belts. Minor episodes of volcanism are likely associated with extensional strain related to episodes of gravitational relaxation.

5.1.3 ISHTAR TERRA

5.1.3.1 *Existing models* Ishtar Terra is unique on Venus. Any model for its formation must acknowledge and address: (*a*) its irregular planform and complex topographic profile including high interior plateau, peripheral mountain belts, and surrounding high tesserae; (*b*) coherent contractional strain patterns across thousands of kilometers with no strain gradient despite the several kilometer change in elevation; (*c*) dominant fold ridges which parallel the orientation of their host mountain belt (and tesserae) with Lakshmi, which itself is generally undeformed at the surface; (*d*) a lack of evidence for extensional collapse at the highest elevations; (*e*) a mean crater retention age near the planetary mean; and (*f*) volcanic plains contemporaneous with tectonism locally. Previous models fall short in addressing these observations.

Pre-*Magellan* convergent plate-tectonic models (Crumpler et al 1986, Head 1990, Roberts & Head 1990b) set out to explain the general contractional nature of Ishtar structures and high topography; however the topographic and structural "details" observed in *Magellan* data (*a*)–(*d*) create structural and kinematic complications. These models imply subduction all around Lakshmi, yet, Uorsar Rupes and Vesta Rupes form the only slopes similar to terrestrial subduction trenches. The distribution and orientation of Ishtar structures are not consistent with subduction along either of these boundaries; southward subduction along Uorsar Rupes (Head 1990, Solomon & Head 1990) requires that Akna and Maxwell are left-lateral and right-lateral shear zones, respectively; northward subduction along Vesta Rupes would require the opposite shear sense in each belt. Subduction along the other Ishtar margins would require radial contraction in the subducting lower plate, or radial extension in the upper plate; evidence for the structural and kinematic framework implied by these models has not been identified, and the observations are counter to model implications.

Regional crustal shortening preserved in Ishtar deformed belts is variably interpreted as the result of mantle downwelling (Bindschadler et al 1992b) or mantle upwelling (Grimm & Phillips 1991). Neither model is sufficiently precise to predict detailed surface strain or topography. We discuss downwelling and upwelling models, and then present a model we believe addresses the observations summarized above.

The mantle downwelling model proposes that sinking mantle beneath Lakshmi Planum pulls crustal material inward and downward; subsequently, the crust thickens resulting in high surface elevation. When downwelling ceases, the thickened crust spreads gravitationally, decreasing in elevation. Many of the problems with this model are the same as those presented for crustal plateau formation by downwelling. Theoretical time scales for crustal flow are not geologically reasonable, and gravitational evidence argues against the presence of an inferred crustal asthenosphere. In addition, the model calls on huge volumes of lower crust to thicken Ishtar Terra, and therefore requires means by which to produce and transport large volumes of lower crust; evidence for these processes is not identified despite available high-resolution *Magellan* imagery, and the observed surface strains are inconsistent with thickening by lower crustal flow. Furthermore, if Ishtar elevation results from thickened crust, the highest regions would be thickest, and therefore these regions should display evidence of extensional collapse; yet the main part of Maxwell Montes, residing greater than 7.5 km above MPR, is void of extensional structures.

The upwelling model proposes that Ishtar Terra comprises a surface expression of a mantle plume that produces topography dynamically and by volcanic construction, and holds that the mountain belts result from incipient mantle return flow (Grimm & Phillips 1991, Phillips et al 1991). This model supposes that an upwelling plume impinged on a preexisting crustal plateau, which would make the lithosphere in this region weaker than its surroundings (Phillips et al 1991). Following earlier suggestions (Pronin 1986, Basilevsky 1986), the hypothesis is that outward crustal flow is attenuated as the strength barrier is reached at the plateau margin; this results in the accumulation of crust to form mountain belts. Because this is a region of enhanced temperature gradients and thick crust, lower crustal flow presumably takes place on geologically reasonable time scales [$O(10^8)$ Ma or less]. Thus, as in downwelling, mountain belts result from thickened crust; therefore the highest elevations, underlain by the thickest crust, should be loci of gravitation collapse, which is not observed. In addition, this model does not account for outboard tesserae with structures parallel to those of adjacent mountain belts; further, it is not clear that attenuation of outward crustal flow would result in steep slopes leading away from a plateau free of evidence of synchronous deformation.

5.1.3.2 *A mantle imbrication model* We discuss a model for the formation of Ishtar Terra (Hansen & Phillips 1993b,c) that begins with geologic and structural observations and is developed within the context of the theoretical and geophysical constraints outlined previously. Observations (*a*)–(*d*) require a means to deform huge expanses of crust without developing strain gradients, a means to support topography without greatly thickening crust, and specific relations between surface strains and topography. These observations can be accounted for by a model in which Lakshmi Planum represents an old crustal plateau that acts as a buttress to upper mantle residuum at depth, causing residuum to stack up locally around its perimeter during lithospheric contraction resulting from mantle downwelling (Figure 2). The greatly thickened upper mantle and mantle downwelling could be the result of an instability of the type described by Parmentier & Hess (1992). The upper part of the residuum, at temperatures less than T_f, is buoyant, and imbricates in a quasi-brittle fashion in response to shear forces associated with the instability developed in the lower, negatively buoyant, part of the residuum layer. This lower part flows downward in a ductile fashion, eventually becoming entrained in the general mantle circulation pattern. As uppermost upper mantle (buoyant residuum) translates at depth, a décollement forms between the upper crust and upper mantle. Shear stress is transmitted to the base of the upper crust, which responds by forming coherent ridge and valley folds with axes generally perpendicular to the direction of relative displacement; these structures form everywhere that the upper crust and upper mantle are displaced relative to one another. As the uppermost mantle is translated, it thickens locally, deforming in a brittle fashion because of its relative strength. The increase in residuum thickness supports surface topography.

Ishtar's irregular planform and topography reflect the shape and thickness, respectively, of residuum stacked at depth.

Lakshmi Planum forms a crustal nucleus for Ishtar deformation. The buoyant, thickened crust of Lakshmi cannot be drawn downward by mantle downwelling, and it acts as a buttress to the uppermost mantle at depth. We postulate that the "keel" of upper mantle that extends below Ishtar Terra is similar to the mantle keel than extends beneath continents on Earth (Jordon 1975, 1981). Topography of Ishtar deformed belts are supported by variably thickened residuum, and ridge and valleys within deformed belts result from folded, modestly thickened crust. The crust of Ishtar deformed belts, supported by residuum, not by thickened crust, is mechanically stable even at the highest elevations. Ishtar Terra would also be free of coronae, assuming coronae form as a result of melt instabilities within regions of tension. Flood-type volcanism can occur in regions of local lithospheric extension that allow upward movement of residuum and renewed, but modest, pressure-release partial melting. The crater density at Ishtar, close to the planetary average, attests to the tectonic and volcanic stability of the region. Quantitative tests of this model would include verifying that the high-resolution gravity constraints provided by the *Magellan* near-circular orbit can be satisfied, and examining the stability of the thickened residuum and the longevity of the downwelling.

5.2 *Coronae and Mesolands*

Here we discuss the nature of plumes that give rise to coronae, the hypothesis for subduction at the margin of some large coronae, and the origin of the mesolands because they are strongly linked to coronae.

5.2.1 MODELS FOR CORONAE ORIGIN There is widespread agreement that coronae result from buoyant diapirs or plumes impinging on the base of the lithosphere (Basilevsky et al 1986; Schubert et al 1989, 1990; Pronin & Stofan 1990; Stofan & Head 1990; Stofan et al 1991, 1992; Squyres et al 1992a; Janes et al 1992; Hansen & Phillips 1993a). *Magellan* images provided the strongest observations in support of this notion—the oldest tectonic features of coronae are radial fractures; this is the deformation that results from pushing up an elastic plate from underneath (Squyres et al 1992a). However, there is disagreement as to whether the diapirs are thermally driven plumes arising from the core-mantle boundary, and are genetically related to volcanic rises (e.g. Stofan et al 1992, Janes et al 1992, Squyres et al 1992a), or if they are compositional diapirs resulting from melt instabilities within the upper mantle (e.g. Tackley & Stevenson 1991, Phillips & Hansen 1993). Any model for diapir genesis must explain (Stofan et al 1992): (*a*) the widespread global, but not spatially random, dis-

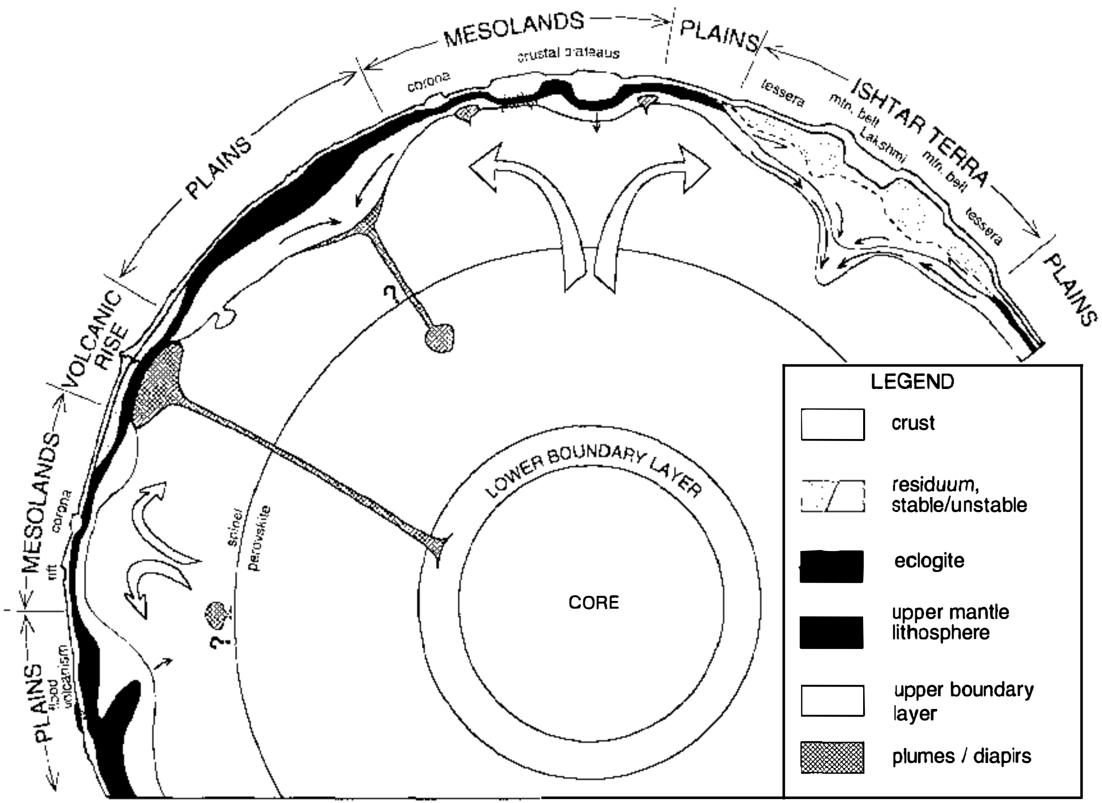

Figure 2 Cartoon cross section illustrating our model of Venus tectonics. Illustration is not to scale and is highly schematic.

tribution of coronae; (*b*) the local concentration of coronae within rift zones, or regions of crustal extension, and the negative association of coronae with both volcanic rises and planitae (Herrick & Phillips 1992); and (*c*) the wide range in corona sizes (100–2600 km) and mode in the size distribution of 200–300 km.

In Section 3.2.3.2 we argued that very long plume transit times through the mantle make it unlikely that the diapirs responsible for coronae origin are thermal plumes arising from the core-mantle boundary. Tackley & Stevenson (1993a) discuss a mechanism for spontaneous and self-perpetuating magmatism that could account for coronae (Tackley & Stevenson 1991, 1993b). In this model, regions of partial melt, or at least incipient partial melt, exist in the upper mantle where the geotherm approaches the solidus temperature. A Rayleigh-Taylor-like instability will develop if an infinitesimal upward velocity perturbation is applied to some portion of the partial melt zone. This is so because the perturbed element of rock will undergo an increment of pressure-release partial melting and thus become more buoyant. The buoyancy will increase the upward velocity leading to still more partial melting and a rising, growing compositional plume will develop until the growth rate is balanced by outward percolation of melt. Lithospheric extension could easily supply the required velocity perturbations, and the strong correlation of coronae with extensional tectonics (e.g. Parga Chasma) provides support for this view. Magee Roberts & Head (1993) have noted the tendency for those coronae with large volcanic flow fields to be associated with lithospheric extension. This is interpreted by them to be the result of increased partial melting of thermal plumes that reached shallower depths than normal because of lithospheric stretching. The Tackley-Stevenson mechanism would also yield a positive correlation between the amount of stretching and the amount of melt produced.

Since not all coronae are clearly associated with cxtensional tectonics, the Tackley & Stevenson mechanism would not necessarily explain the global distribution of coronae. It is possible that coronae are the result of thermal plumes generated at an interior thermal boundary layer associated with a phase transition—e.g. the spinel-perovskite phase boundary separating the upper and lower mantle at $\sim$700 km depth—where the plume transit times might be 100 Ma or less. If there is a barrier to convection in the interior, then much of the heat from beneath the barrier will be removed in the form of hot plumes. How many plumes would be required to remove this heat if the plumes are also required to make coronae? Following the analysis of Whitehead & Luther (1975), several thousand 300-km-diameter coronae are required to be active at any one time if we assume that plumes carry most of the basal heat (Bercovici et al 1989).

Compared to the observed population of ~400, coronae could not possibly represent the total heat output of the lower mantle and core. Either (*a*) such an internal thermal boundary layer does not exist on any permanent basis (e.g. Steinbach & Yuen 1992) and cannot be used to explain the origin of coronae, or (*b*) plumes associated with volcanic rises must represent the majority of heat removal from this interface, and corona–producing plumes play a minor role (with the possible exception of Artemis).

We cannot rule out completely the origin of coronae from thermal plumes associated with relatively shallow thermal boundary layers in the mantle, so it is possible that the global corona population is a mix of a spatially random population (perhaps the largest coronae) associated with this mechanism plus a second, regionally biased, population associated with extensional tectonism and caused by melt instabilities in the uppermost mantle. We note, however, that the absence of associated extensional tectonism with some coronae does not rule out a melt instability origin, and our conclusion is that this is the dominant mechanism in the formation of coronae. The intimate association of magmatism with coronae, including multiple phases of volcanism of apparently different composition (Hansen & Phillips 1993a), also suggests a compositional origin for the causative plume, although pressure release-partial melting associated with a purely thermal plume cannot be ruled out.

5.2.2 SUBDUCTION AT CORONA MARGINS? McKenzie et al (1992a) and Sandwell & Schubert (1992a,b) proposed that the peripheral troughs associated with some large coronae (Artemis, Latona, Eithinoha) and two chasmata (Dali and Diana) bordering corona-like structures are the sites of lithospheric subduction (see Figure 1 for locations). The strongest argument for this hypothesis rests in topographic analogy to terrestrial subduction zones. In particular, these troughs are arcuate in planform and bordered by parallel but asymmetric highstanding topography, the higher of the two border features on the concave side of the trough. Stofan & Bindschadler (1993) have noted that asymmetric topography is a common feature of all coronae with troughs regardless of size. The convex high is reminiscent of the flexural highs observed seaward of terrestrial trenches. Sandwell & Schubert (1992b) matched flexural models to the outboard topographic highs as discussed in Section 4.1.2; attempts to test the subduction hypothesis by comparing the bending moment supplied by inboard topographic loads to the model bending moments obtained by matching topographic profiles proved inconclusive (Sandwell & Schubert 1992a).

At Artemis, Latona, and Eithinoha, the troughs subtend an arc greater than 180°, so underthrusting would lead to strong contraction in the outboard lithosphere, which is not observed. Instead, Sandwell & Schubert

(1992a,b) propose that the outboard lithosphere is rolling back in response to the onslaught of a spreading corona. The subducted lithosphere would fail in tension; the most one might observe at the surface are radial fractures propagating outward from the trench site, for which there is some evidence (Stofan et al 1992).

Through study of high-resolution *Magellan* images, Hansen & Phillips (1993a) documented the presence of pre- to syn-trough tectonic fabrics that trend across a proposed convergent plate boundary at Latona Corona; these structures are not consistent with a subduction interpretation. Studies of this type at other proposed plate boundaries are less conclusive (G. Sharer, personal communication). Hansen & Phillips also documented crosscutting relations that indicate east- and northeast-trending fractures that parallel, and locally coincide with, Dali (and Diana) chasma postdate formation of radial and concentric fractures associated with proposed "upper-plate" coronae. These relations are inconsistent with a subduction interpretation, but they are consistent with Diana and Dali chasmata formation largely postdating spatially associated coronae.

Mechanically, subduction on Venus would appear to be difficult because of the expected large positive buoyancy of the lithosphere resulting mainly from the high surface temperature (Phillips & Malin 1983). Detachment of almost the entire crust would be required to impart negative buoyancy to the remaining lithosphere (Phillips et al 1991). Alternatively, if the lithosphere could be underthrust to depths of 50–100 km, the presumed basaltic crust would undergo a phase change to garnet granulite and eclogite, and resistance to subduction would be diminished (Phillips et al 1991, Sandwell & Schubert 1992a).

The subduction hypothesis remains as a tantalizing idea. The acquisition of new high-resolution gravity data in the latter part of 1993 and into 1994 resulting from the circularization of *Magellan*'s orbit, along with detailed structural mapping, should allow significant progress on this question.

5.2.3 MESOLANDS ORIGIN Morgan & Phillips (1983) proposed that topography on Venus less than about 1 km above the MPR could be supported solely by thermal compensation in the lithosphere, thc upper bound being dependent on the assumption of a mean thickness of thermal lithosphere of 100 km. Areas of excess heat flow (compared to the planetary average) will thin the lithosphere, replacing colder material with hotter, less dense material that will support elevated topography. Elevations in excess of the estimated upper bound would be supported by crustal thickness variations. We suggest that the mesolands fit the model put forward by Morgan & Phillips, and this view is supported by the generally positive but modest geoid anomalies associated with mesolands (Bills et al 1987, McNamee et al 1993, Nerem et al 1993). We associate the excess heat flow to broad, diffuse convective upwellings that characterize internally heated convec-tion. In addition to cooling linked to major downwellings, these upwellings must be the dominant heat transfer mechanism out of the mantle.

By virtue of their excess heat flux, mesolands would be characterized by large amounts of partial (or incipient partial) melt in the upper mantle and thus are fertile ground for generating coronae by the Tackley & Stevenson (1991, 1993a) mechanism. Further, rifting and the development of chas-mata are natural by-products of thermal uplift, and create a tensional environment to initiate partial melt diapirs that will form coronae. In addition, the crustal plateaus believed to be the youngest (Ovda and Thetis regiones) have a close spatial association with mesolands and may have formed from runaway partial melting events as described in Section 5.1.2.1

Anderson et al (1992) used the level of correlation of terrestrial hotspots with regionally broad, low-velocity anomalies in the upper mantle to argue that hotspots, as well as massive volcanism that forms basaltic plateaus, are the result of a juxtaposition of tensional portions of the lithosphere with broad areas of high temperature in the mantle. We believe this is a good analogy for the formation of mesolands and, in the extreme, the formation of crustal plateaus on Venus.

5.3 *Origins of Planitae*

The most likely interpretation of planitae on Venus is that, at least in their large-scale aspects, they are supported dynamically by sheet-like and possibly cylindrical downwellings associated with cold return flow in the mantle (Bindschadler et al 1990, 1992b; Phillips et al 1991). Large apparent depths of compensation for Guinevere/Sedna Planitae of 100–150 km and for Niobe Planitae of 180 km (Phillips et al 1991) support this view. Herrick & Phillips (1992) showed that inversions of long-wavelength gravity and topography data for a dynamic mantle component and a static lithospheric component produced sheet-like downwellings beneath the planitae. Zuber (1990) argued that the most likely origin of ridge belts is compressive stresses associated with large-scale convective downwellings. An alter-native view is that planitae mark the sites of hotspot upwellings where the crust has thinned substantially in response to initial dynamic uplift (Buck 1992).

The simple view that planitae and their attendant ridge belts represent a contractional response to convective mantle downwellings (e.g. Squyres et al 1992b) must be tempered with observations of heterogeneous tec-tonism and volcanism. An excellent review of this issue is given in Solomon et al (1992). It is clear that there have been a number of horizontal scales of deformation that have operated, indicating a range of lithospheric responses—from deformation of the uppermost few hundred meters of the crust, to the entire strong upper crust, to the crust and strong upper mantle. Three particularly challenging questions to the convective downwelling hypothesis are: 1. What is the origin of tensional tectonism? 2. What is the origin of volcanism? 3. Why aren't deformation belts distributed uniformly in the planitae? Unfortunately, regional-scale variations in lithospheric properties, crustal thickness, and heat flow will probably lead to spatial and temporal overprinting that will be difficult to decipher in light of models with extremely general predictive capabilities. Thus "proof" of one interpretation over another may prove exceedingly difficult. At a geological level, models for Venusian deformed belts in the planitae must explain: (*a*) the elevated nature (0.5–1 km) of deformed belts relative to adjacent plains; (*b*) the similar orientation of principle strain in deformed belts and adjacent "undeformed" plains; (*c*) local orthogonal contraction and extension in ridge and fracture belts (e.g. Lavinia Planitia); and (*d*) local along-strike morphological changes from ridge belts, marked by fold ridges, to hybrid fracture belts which preserve ridges, but are dominated by fractures, faults, and local paired graben (e.g. Atalanta-Vinmara Planitiae).

Underlying planitae may be the vestiges of a period of intense deformation associated with high crustal (Solomon 1993) and/or mantle (Grimm 1993) strain rates. Following an episode of lithospheric extension, there has obviously been a complex interplay of deformation and volcanism in the planitae, and in many places ridge belt formation is older than most or all of the volcanic plains units (Solomon et al 1992). Thus a model that assigns convective downwelling to planitae must allow partial melt to be generated in this environment. We suggest that the most likely possibility for this is that portions of the downwellings are manifested as delaminations of the upper part of the mantle, possibly including separation at the Moho (e.g. Bird 1988). These detachments are replaced by hot mantle material from below that will undergo partial melting because it is emplaced closer to the surface. Local extensional environments in the lithosphere then allow access of magma to the surface. The relatively shallow depth of the magma source would lead, for the most part, to the formation of volcanic plains rather than constructs, although other factors, affecting magma viscosity, may be involved. The emplacement of this hot mantle material should lead to elevated topography, but unlike our proposal for the mesolands, this is a transient event with smaller levels of partial melting.

It is clear that ridge belts are genetically related to widespread distributions of wrinkle ridges and fracture belts are similarly related to extensional grooves (Squyres et al 1992b). Thus the deformation belts appear to be one manifestation of lithospheric response to large-scale regional stress fields. Those cases where extensional features (fractures, graben, fracture belts) are orthogonal to contractional features (wrinkle ridges, ridge belts) are easily understood in terms of principal strain axes in a contractional environment—extensional features will form in accordance with the axis of least contractional strain. However, in many cases contractional and extensional features are not orthogonal, and in a number of instances contractional features change to extensional features along strike. Solomon et al (1992) offer two possible explanations for this: (*a*) Fracture belts are extensional structures that occur in response to local igneous intrusion or lithospheric heating, and (*b*) fracture belts occur as a result of stretching over the crest of contractional structures. These authors favor the former explanation because of a close association between fracture belts and volcanism, but there is a chicken-and-egg issue here in that the occurrence of a tensional environment may simply favor magmas reaching the surface.

Our model for deformed belts follows from our model of Ishtar Terra. Downwelling in the mantle is responsible for broad-scale regional lithospheric contraction normal to the trend of ridge belts. The uppermost mantle deforms in a heterogeneous fashion, as evidenced by the spacing between belts; however, because planitae lack a nucleating crustal plateau, the uppermost mantle is not buttressed at depth, and therefore does not imbricate as it does under Ishtar Terra. Orthogonal ridge and fracture belts may result from crustal shortening due to lithospheric contraction. The basic spacing of ridge belts and the deformational scale within individual belts result from instabilities developed in a jelly-sandwich rheology. Ridge belt elevation results chiefly from crust thickened by minor lower crustal flow. Along-strike changes from ridge belt to hybrid fracture belts record gravitational collapse following crustal thickening beyond the mechanical limit of the upper crust.

5.4 *What is Holding Topography Up (and Down)?*

The enigma of how large-scale topography is maintained on Venus in the face of high surface temperature remains and is punctuated by the presence of the circular Cleopatra impact crater sitting happily undeformed on the steep slope of Maxwell Montes.

Turcotte (1993) proposed that topography is supported by the strong lithosphere that results from a low temperature gradient. The low temperature gradient exists because heat flux from the convecting mantle is subdued while the mantle is rewarming following an episode of extremely active plate tectonics and vigorous subduction cooling ~500 Ma ago, as required by a model of catastrophic resurfacing (Schaber et al 1992). The attendant lithosphere supports topography by thermal isostasy, but is able to achieve a greater maximum elevation than that found by Morgan & Phillips (1983) because of the much thicker lithosphere: ~300 km vs 100 km. Additionally, the stronger crust is able to locally support mountain ranges, such as Maxwell Montes, by Airy compensation.

Thermal isostasy with a 300-km-thick lithosphere will also produce the large GTR values that are ascribed by others to dynamic support (e.g. Kiefer & Hager 1991a). Turcotte's proposal is interesting but speculative in that it requires the initiation of a global and essentially spontaneous subduction event to substantially cool the mantle. Further, Turcotte (1993) does not explain how the topography is created in the first place. If the topography is generated during the plate tectonics episode (presumably the most likely time), the lithosphere is thin and won't support topography by Turcotte's own arguments. If too much time has passed, then the lithosphere is thick and static and it would seem to be difficult to generate the observed highland features. There would have to be a window intermediate in the 500 Ma cycle in which there is enough tectonic activity in the lithosphere to create the highlands, their attendant mountain belts, and the planitae, but the lithosphere would at the same time have to be strong enough so that these features would stick around.

Our view is that topography is supported by a combination of dynamic, thermal, and Airy crustal compensation. At Ishtar Terra we have proposed that thickened mantle residuum is supporting topography. Because we have argued that the time scale for lower crustal flow most likely exceeds 1 Ga, once topography is compensated within the crust, it is relatively stable; although there may be an early episode of partial collapse, relaxation times are very long. The probability that the viscosity of the crust may be a good deal higher than has been assumed in modeling studies strengthens this argument. Finally, we point out that in those regions of the upper mantle in which stable partial melt residuum exists, the mantle layer that transfers heat by conduction can be substantially increased, leading to low temperature gradients and a strong lithosphere (Parmentier & Hess 1992).

5.5 *Is there Horizontal Lithospheric Motion?*

The mountain belts of Ishtar Terra and the northern ridge belts are regions of maximum crustal shortening. However, strong evidence for lithospheric-scale subduction and recycling has not yet been recognized in the geological record of Venus. In Section 5.2.2 we presented arguments against subduction at coronae boundaries. A subduction interpretation outboard of the Ishtar Terra mountain belts may be consistent with broken-plate elastic models representing underthrusting (Solomon & Head

1990), but it is not consistent with surface strain patterns (Section 5.1.3.1).

Earlier predictions that much of Aphrodite Terra is a site of lithospheric divergence (Head & Crumpler 1987, 1990) are invalidated by *Magellan* data (Solomon et al 1992). Evidence for lithospheric extension is limited to two classes of structures: quasi-circular coronae and associated chasmata of the mesolands, and volcanic rises with symmetric rift zones. Although both sites display evidence for volcanic flux, horizontal displacements in each case are probably limited to a few tens of km. In the case of eastern Aphrodite Terra, early-formed coronae, which are generally circular in planform, are cut by fractures of Diana and Dali chasmata, interpreted as rifts (e.g. Schaber 1982); however, the amount of crustal extension must be very small, less than 10% across the entire width of the zone as the coronae remain circular, and are not elongate in a direction normal to the trace of the rift (Hansen & Phillips 1993a). In addition, symmetric rifts that radiate outward from the broad topographic domes of volcanic rises show limited local extension (< 30%) normal to strike. Along strike away from the topographic high, the rift trough and associated structures broaden and splay outwards, recording progressively less crustal extension. It is evident that these rifts are simply a response to crustal doming. Not surprisingly, few large-offset strike-slip faults are recognized planetwide (Solomon et al 1992), although evidence for limited local horizontal shear is observed in regions of bulk crustal shortening and extension (e.g. Squyres et al 1992b).

6. THE REALLY BIG PICTURE—GLOBAL TECTONISM AND MAGMATISM

After a brief review of Venus geology, we provided an essentially theoretical view of mantle and lithospheric dynamics that was then used as a framework to provide models for the origin of specific physiographic features on Venus. We continue in this vein, now attempting to put together a model for global tectonism and magmatism. A cartoon of our Venus can be found in Figure 2.

Our view of Venus is simply this: Because Venusian mantle convection is driven almost entirely by internal heating, it is dominated by strong, cold downwellings from the upper thermal boundary layer and by broad, diffuse upwellings; these are manifested at the surface by the planitae and mesolands, respectively. Everything else is nothing more than minor blemishes superposed on this major system. There is but one Ishtar and but two crustal plateaus that by their size may have been the most recently active (Ovda and Thetis regiones). These are the result of unusual events caused, respectively, by (*a*) detachment of residuum and downwelling in the vicinity of a preexisting crustal plateau, and (*b*) runaway partial melting events associated with broad upwellings creating mesolands. There are but a handful of volcanic rises, which are probably linked to upwelling, hot plumes from a sluggish lower thermal boundary layer. We emphasize that this view is based on the Vcnus that we see today from images, altimetry, and gravity data, and does not necessarily apply to a Venus more than 1 Ga in the past.

While large-scale deformation in the lithosphere is still controlled by mantle convection, crustal tectonics linked directly to convective stress coupling passed through a critical strain rate perhaps 1 Ga ago, after which the rate of tessera production greatly diminished. Contemporary tessera formation may be taking place only at Ishtar Terra and in the mesolands.

This view of Venus is built on the following concepts developed earlier in this paper:

1. Mantle convection in Venus operates at about the same viscosity and Rayleigh number as on Earth.
2. The upper interface of convection acts as a rigid boundary, at least on a time scale of less than 1 Ga. Lithospheric recycling to the interior is a minor process with the possible exception of upper mantle detachment beneath planitae.
3. Venus lacks an asthenosphere; convective stresses couple directly into the lithosphere with a vigor that has decreased with time.
4. Only a modest amount of heat crosses the core-mantle boundary, leading to infrequent, large plumes.
5. The outermost part of the upper thermal boundary layer is relatively static, but it will initiate cold downgoing plumes and/or detach on occasion.
6. A residuum layer exists at various places in the upper mantle. At times it is disrupted, mixing with normal mantle, although a disruption may take thc form of a detachment and behave in a semi-rigid manner in the early stages of an event.
7. The crustal thickness is nominally $\lesssim 20$ km in the plains but has achieved greater thickness beneath crustal plateaus, beneath Ishtar Tcrra, and bencath the mesolands.
8. In the plains there is a stratified rheology of strong upper crust, weak lower crust, and strong upper mantle.
9. Venus has a nominal temperature gradient that lies between 10 K/km and 20 K/km; local variations may fall outside of this range.
10. The lower crust is able to deform in a ductile manner in response to horizontal forces. Tectonic features have resulted up to the scale of ridge and fracture belts.

11. Large-scale lower crustal flow has not taken place and thus does not offer an explanation for high-standing topography resulting from convective downwellings and subsequent crustal thickening.
12. Crustal recycling is minor and possibly occurs only in the deep roots of crustal plateaus that have transformed to denser phases such as eclogite. Large-scale underthrusting may also allow return of crust to the mantle below, but we find no convincing evidence for this.

Crust on Venus today is being generated dominantly in the mesolands. In this sense the mesolands are analogous to oceanic spreading centers in that crust is generated in response to diffuse mantle upwellings and partial melting. On Venus this is manifested at the surface as coronae and flood and constructional volcanism. The mesolands differ from terrestrial spreading centers in that the crust accumulates vertically rather than horizontally (as results from seafloor spreading). Crust may also be formed in the intermediate stages of plume evolution (Phillips et al 1991), but the evidence today is that most volcanic rises are relatively magmatically dry. More likely, crustal plateaus are an exuberant form of mesolands development. Over the history of Venus it would seem that significant crust could be developed. Either crustal development is inhibited by the presence of residuum in the upper mantle (Phillips & Grimm 1990), or significant amounts of crust are being recycled in contradiction to concept 12. above. It is possible that the crust thickens enough beneath mesolands to spawn phase changes to buoyantly unstable garnet granulite or eclogite, or that mesolands eventually evolve into crustal plateaus where such phase changes take place deep in plateau roots. These denser portions of crust may detach and eventually recycle in the mantle. The mesolands are presently the most tectonically/magmatically active areas of the planet, a conclusion strongly supported by crater data (Phillips et al 1992).

We suggest that a residuum layer has had a profound effect on the evolution of Venus. Jordan (1975, 1981) proposed, mainly on the basis of seismic evidence, that beneath the continents on Earth, a layer ("tectosphere") that is depleted in basalt and enriched in magnesium, i.e. a residuum, extends to several hundred km depth. This layer is held to be buoyantly neutral with respect to the surrounding mantle because positive chemical buoyancy is offset by colder temperatures. The tectosphere could be the fate of some of the residuum produced by partial melting in the oceanic mantle, being swept into the continents by processes associated with subduction (see Kaula 1990b). We suggest that residuum is also swept around in the mantle of Venus; lacking the strong, focused lithospheric subduction of Earth, its distribution may be more widespread on Venus. Accumulations may nucleate beneath crustal plateaus, where the mantle may be hotter because of thermal blanketing effects (e.g. Lenardic et al 1991). Buoyant residuum may form stable keels against mantle convection, may compensate large-scale topography, act in places as a barrier to mantle ascent (e.g. plumes), inhibit partial melting, and may occasionally undergo large-scale detachment with dramatic tectonic consequences.

The planitae mark the sites of mantle downwellings. These may be part of the normal mantle flow pattern, or may initiate in the upper part of the thermal boundary layer as plumes or semi-rigid detachments. Leitch & Yuen (1991) showed that formation of cold downgoing plumes is chaotic in variable-α convection (in which the coefficient of thermal expansion α decreases strongly with increasing pressure), and the imposition of a rigid boundary condition increased the number of instabilities in the upper thermal boundary layer. Because the upper thermal boundary layer in internally heated convection is able to store a great deal of potential energy, instability generation, including detachment, has provided the major source of convective stress coupling into the Venusian lithosphere. By the chaotic nature of this activity, we expect that strain in the overlying lithosphere to have been both spatially and temporally chaotic. The net result is that the lithosphere will be constantly subjected to a complex pattern of contractional and extensional strain. The level and rate of this strain have decreased over time as expected from the secular decrease of heat production, and the prolonged period of extensive tesserae formation ended about 1 Ga ago (Solomon 1993, Grimm 1993).

In Section 2.4 we argued from crater observations that at one time or another partial melts have been available on a regional basis everywhere on the planet; the coincidence of an extensional strain environment has allowed volcanism to occur as a passive phenomenon. The local style of volcanism will depend, inter alia, on the specific stress state of the lithosphere, the local state of the upper boundary layer, and the thickness of residuum in the upper mantle. These occurrences of volcanism will be maximized in those regions of hot upper mantle, i.e. the mesolands. Further, volcanism has secularly declined over the history of the planet and not just directly from the reduction in heat production. As the strain rate and corresponding deformation decreased in the crust, the opportunities for magma to reach the surface have also diminished; correspondingly, the ratio of intrusive to extrusive igneous activity has increased.

The chaotic nature of strain in the lithosphere also led to polyphase deformation and tessera formation in both crustal plateaus and lower-lying plains, as discussed in Section 5.1.2. As pointed out in that section, Ovda and Thetis regiones may be the most recently active crustal plateaus in the sense of their formation by crustal thickening. Once this active process ceases, crustal plateaus undergo a slow gravitational collapse governed by a time constant (τ_1) of 1 Ga or greater. These features slowly lose relief, but, acting as strain magnets, their areal extent may be increased by accretion of surrounding crust. Eventually these plateaus are reduced to patches of old tesserae embayed by volcanism. The widespread distribution of low relief tesserae in the plains attests to the importance of both crustal plateau generation and relaxation as well as in situ deformation of the plains themselves over time periods prior to 1 Ga ago.

Volcanic rises are but an energetically minor phase of heat transfer out of the interior, perhaps accounting for the relatively feeble heat produced in a cooling core. They are distinct in both their large gravity anomalies and styles of volcanism, both pointing to relatively deep sources in the upper mantle and possibly interaction of thermal plumes with a residuum layer.

The complex and chaotic nature of mantle downwellings, the spatial variability of residuum, and the relatively well-defined regions of broad upwellings suggest that at any given time the upper mantle of Venus can be divided into a finite number of domains, each characterized by one type of a limited number of characteristic behaviors. On Earth, domains might be defined by upper mantle beneath continents, beneath young oceans (e.g. the Atlantic) and beneath mature subducting oceans (e.g. the Pacific). Venus is different because it does not have plate tectonics and is the epitome of a planet with vertical tectonics. The crust is not recycled, but instead acts like a rug that locally rips and crumples as a result of relative displacement of domains within the mantle below; in a secularly decreasing manner, it has fractured and folded, domed upward and downward; it has been locally stretched by magmatic diapirs, and responded by fracturing, folding, and locally faulting when driven by convective stresses within these domains. The local rips and rumples in the crustal rug of Venus are never far from where they formed, and never far from where the rock that comprises it differentiated at depth. The surface is replenished or "repainted" with volcanism occurring today mostly in the mesolands and at volcanic rises.

Our view of Venus will not please everyone; indeed it may not please anyone. It is based on a number of theoretical arguments regarding mantle and lithospheric dynamics, and on an attempt to reconcile what we see on the surface and infer from gravity data with these theoretical considerations. It is a picture more complex than many earlier views (e.g. Phillips et al 1981, 1991; Phillips & Malin 1983, 1984; Kaula 1990a; Bindschadler et al 1992b), yet does not consider all possibilities [e.g. the tectonic and magmatic consequences of conversion from layered to whole-mantle convection (Steinbach & Yuen 1992)].

This paper is being written as the *Magellan* orbit is being circularized to acquire a high-resolution gravity field. This new data set, along with continued analyses of images and topographic information, will help a great deal in testing the model of Venus proposed here. As with previous attempts to describe Venus, parts of this model will be wrong, but hopefully other parts can be carried forward.

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