Hot-Spot Evolution and the Global Tectonics of Venus

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The global tectonics of Venus may be dominated by plumes rising from the mantle and impinging on the lithosphere, giving rise to hot spots. Global sea-floor spreading does not take place, but direct convective coupling of mantle flow fields to the lithosphere leads to regional-scale deformation and may allow lithospheric transport on a limited scale. A hot-spot evolutionary sequence comprises (i) a broad domal uplift resulting from a rising mantle plume, (ii) massive partial melting in the plume head and generation of a thickened crust or crustal plateau, (iii) collapse of dynamic topography, and (iv) creep spreading of the crustal plateau. Crust on Venus is produced by gradual vertical differentiation with little recycling rather than by the rapid horizontal creation and consumption characteristic of terrestrial sea-floor spreading.

THE GLOBAL TECTONICS OF A TERRESTRIAL PLANET REFLECT the means by which its outer rigid shell or lithosphere is able to dispose of internal heat (*1*). The thermal history of a terrestrial planet depends, in turn, on its size. To first order we would expect that two planets of the same size and mean density would experience the same thermal and tectonic history if they have equivalent energy sources (*2*). These qualifications appear to be satisfied for Earth and Venus (*3*), so that with no further information we might expect these two planets to share the same tectonic style. The major questions for studies of Venus are the degree to which this paradigm is correct and the roles of seemingly second-order differences such as surface temperature and volatile abundance (*4, 5*). Our major tools to address these questions are radar images and limited geophysical data gathered in U.S. and Soviet space missions and an appreciation of the various tectonic processes recognized on Earth, with an understanding of how they might change in importance when the parameters are adjusted for Venus.

One of the principal hypotheses concerning the nature of Venusian tectonics is that it is dominated by the effects of convective plumes rising in the mantle, impinging on the lithosphere, and producing hot spots—areas of volcanism and uplift created by the magmatic, thermal, and dynamic effects of the plumes (*3, 5, 6*). Two major elements of this model are (i) that large-scale horizontal motion of the lithosphere does not take place or is, at most, a minor process (*6, 7*) and (ii) that Venus loses most of its heat through lithospheric conduction, which is enhanced at the thin lithosphere associated with hot spots (*6, 8*).

The hot-spot hypothesis stands in contrast to the plate tectonics and sea-floor spreading concepts for Earth, in which the terrestrial lithosphere is broken up into a number of rigid plates that are in horizontal motion and interact at their boundaries, new oceanic lithosphere (and crust) is formed at mid-ocean ridges and consumed at oceanic trenches, and most heat is removed from the interior by cooling and thickening of the new lithosphere as it diverges and moves away from the ridges (sea-floor spreading). On Earth the hot-spot mechanism (*9*) does explain many isolated intraplate volcanic provinces such as Hawaii and Bermuda, but in terms of both volume of new crust generated and quantity of heat removed from the interior this process is insignificant compared to sea-floor spreading (*8, 10*).

Head and Crumpler (*11*) have proposed a tectonic style for Venus that has elements of both the terrestrial mechanism and the Venusian hot-spot hypothesis. In their crustal spreading–mantle plume model, the lithosphere is diverging in the Aphrodite Terra region, which has also been the site of mantle plumes and hot spots that generate magmatically thickened crust. A good analogy would be the mid-Atlantic ridge if it had several Icelands along its axis. The regions of thickened crust are described as plume plateaus and are split by crustal spreading and carried off into the Venusian plains. There they are found as tessera characterized by a trough-and-ridge structure similar to the fabric of the terrestrial sea floor (*12*). In this model, lithospheric convergence is taking place at Ishtar Terra, with accretion of buoyant plateau material (*13*) or underthrusting of normal thickness crust (*14*), but rigid-plate tectonics is not required to operate. The major difference between the crustal spreading–mantle plume model and the hot-spot model that we discuss in this article is that in the latter crustal spreading is at most a limited, regional-scale process accompanied by significant deformation of the lithosphere.

This article was revised well into the first mapping cycle of the Magellan mission to Venus. We do not attempt to incorporate these new, high-resolution radar imaging data into the discussions here. This is partly because only a small fraction of the planet has been mosaicked into standard data products that could be interpreted and placed in a global context. More important, we wish to describe a model for the hot-spot hypothesis that can be tested with the global Magellan imaging data set, attempting to constrain this model with the information already available from previous space missions.

Any model attempting to describe the global tectonics of Venus must explain the following major observations:

1) Elevations on Venus belong to a unimodal population, with most of the surface lying within 1 km of the level of the vast Venusian plains (*15*). High-standing areas occupy a small fraction of

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the surface area (*16*) and are for the most part associated with the equatorial highlands (*5*) (Fig. 1). Here the regional morphology is best described as broadly aligned topographic rises. Those portions of the equatorial highlands that have been imaged from Earth display large-scale rifting and shield volcanism (*17*). The other major region of elevated topography is Ishtar Terra, which is centered about the prime meridian at high northern latitudes. The western portion of this region, Lakshmi Planum, is surrounded by linear mountain belts. There are also a number of basins lying about 1 km beneath the mean planetary elevation. Some (such as Sedna and Guinevere planitiae) are elongate, extending for thousands of kilometers along strike.

2) Long-wavelength free-air gravity anomalies correlate strongly with large-scale topographic features, indicating large depths of apparent isostatic compensation (*18*). The compensation depths are quite variable (*19*), however, indicating that more than one mechanism may be involved or that a single mechanism has a complex behavior.

3) Images acquired by Venera 15 and 16 of the Venusian northern hemisphere poleward of 30° at a resolution of several kilometers and by the Arecibo Observatory reveal a variety of tectonic features (*20–22*). Widespread faulting, ridge belts, the intricate structure of mountain ranges on Ishtar Terra, major elevated regions of intense cross-cutting faults (tessera), and enigmatic circular structures up to 1000 km in diameter (coronae) are observed. These same data sets showed that volcanism operates on different scales and styles and pervasively modifies the surface of Venus. The inventory of impact craters mapped also holds implications for the resurfacing history of the planet.

The Venera and Arecibo radar images and a better appreciation of the complexity of topographic compensation mechanisms allow us to formulate a specific model for the hot-spot hypothesis. Further, a number of recent reports have dealt with the relationship of terrestrial hot-spot plumes to magmatism (for example, *23, 24*), leading to a better understanding of the initial stage of plume behavior, partial melting stages associated with plume evolution, and the relationship of plumes to the overall scheme of mantle convection. The efficacy of hot-spot tectonism on Venus must be evaluated within this new framework of terrestrial analysis.

In this article, we evaluate the hot-spot hypothesis on the basis of the present understanding of lithospheric mechanics and plume dynamics and from the observations described above. The model is taken to an extreme to determine whether it can account for all the major observations; thus it draws an "end-member" picture of the way Venus might work. Other less severe applications can moderate the view, as appropriate, after its limitations are more clearly established and explored (*25*).

High stresses in the lithosphere of Venus, both extensional and compressional, are suggested by the observed tectonic deformation seen in Venera 15 and 16 and in Arecibo radar images; the magnitudes of these stresses depend on the rheology and composition of both the crust and the mantle, the thickness of the crust, and the lithospheric temperature gradient. Knowledge of the effects of rheology and stresses acting on the terrestrial lithosphere as they relate to sea-floor spreading, subduction, and hot spots can be applied to Venus with an appropriate adjustment of parameters in an attempt to understand the horizontal length scale over which stresses might act and lithosphere can be transported coherently. We first present an overview of horizontal stress mechanisms and discuss how they might apply to both crustal spreading and hot-spot tectonics.

Horizontal Forces in the Terrestrial and Venusian Lithospheres

The role of an asthenosphere in plate tectonics. Terrestrial plate motion is driven by body forces within the plates (*26*), so-called ridge push and slab pull. Ridge push is associated with the excess pressure exerted by the oceanic lithosphere as a result of cooling and isostatic adjustment. Although this force may be incapable of breaking the lithosphere (*27*) or initiating subduction (*28*), it places oceanic lithosphere (except near the ridges and trench–outer rise complexes) in a general state of deviatoric compression, as evidenced by the paucity of extensional earthquakes seaward of oceanic outer rises (*29*) and the observed stress orientations in the plates (*30*). Slab pull, the other body force available for driving plate motion, is due to the negative buoyancy associated with the sinking, subducted lithosphere. Plate tectonics are driven chiefly by the large negative buoyancy forces stored in subduction zones; these forces control long-term average plate velocity. Strain diffusion allows finite displacements in subduction zones to be transferred to the accompanying plate (*31*).

The asthenosphere or low viscosity zone (LVZ) beneath the oceanic plates controls the rate of strain diffusion and thus perhaps the length scale over which a coherent plate exists. The presence of an LVZ beneath the oceanic lithosphere with a viscosity perhaps two orders of magnitude less than that of normal upper mantle has

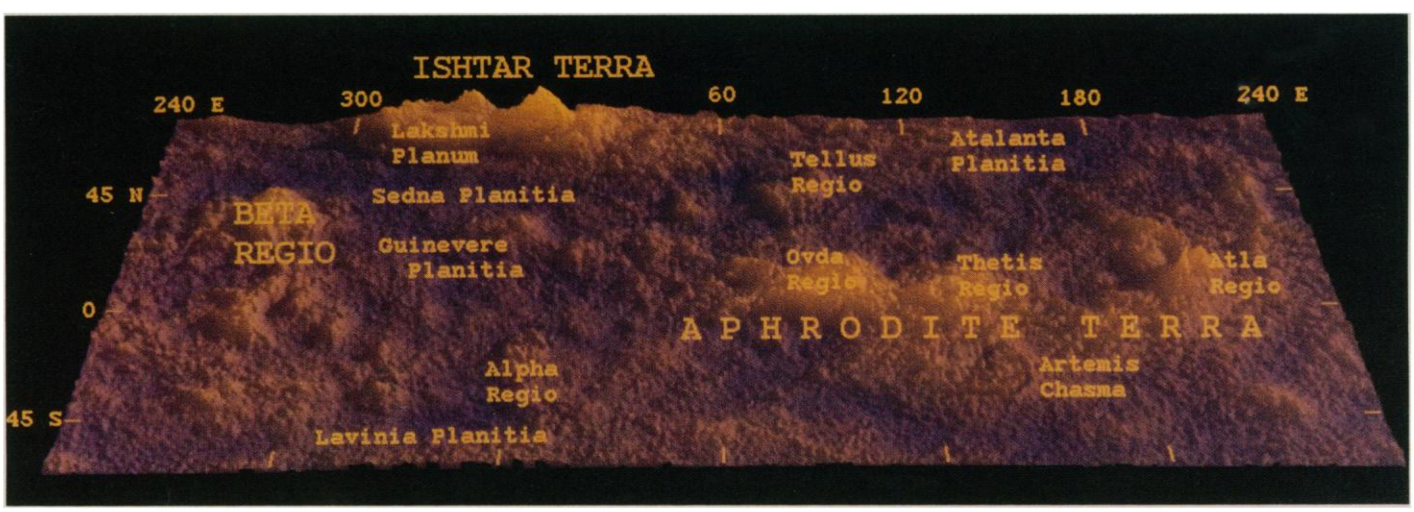

Fig. 1. Shaded relief map of the topography (*15*) and nomenclature of Venus, simple cylindrical projection.

been substantiated from a number of lines of inquiry (32, 33), including studies of stress diffusion after large earthquakes (34). The distance to which significant strain diffuses in a time that is small compared to that for significant plate movement varies inversely, in a complex way, with viscosity (31). Thus we would expect the average size of plates to decrease in the absence of an LVZ.

The role of asthenospheric viscosity in controlling the rate of strain diffusion in the overlying plate can be viewed equivalently in terms of drag at the base of the plate: Compressive stresses associated with ridge push must exceed the opposing force due to drag. In fact, the observed deviatoric compression in the lithosphere can be used directly to constrain the viscosity of the LVZ to values less than 10^{18} to 10^{19} Pa s (32).

We focus on the role of an LVZ in plate tectonics because Venus does not appear to have one. On Venus, the observed large apparent depth of compensation (ADC) values (Table 1) require implausibly deep density sources to support overlying topography if an LVZ is present. The lack of an LVZ on Venus is expected because the upper mantle of Venus is thought to be drier than Earth's, so that temperatures do not so nearly approach the solidus (35). If sea-floor spreading and subduction exist on Venus, then their style may be quite different from that on Earth because of the lack of an LVZ. In particular, it appears that an inability to transmit stress over great distances would preclude the existence of large-scale plates. Also, because of a thinner, less dense lithosphere (6) ridge-push forces on Venus should be about 40% of terrestrial values (36).

Convective stress coupling. Lack of an LVZ inhibits plate motion and also allows convective stresses to couple directly into the lithospheric lid overlying the mantle flow system. Could such shear tractions provide the driving force for lithospheric motion on Venus (11)? If direct coupling is occurring, the free-air gravity field can be used to estimate the magnitude of in-plane forces induced by shear stresses from the convective flow. These data indicate that compressional and extensional in-plane forces could be several times 10^{12} N m^{-1} (37). If the lithosphere is broken and end boundaries are free to move, then lithospheric motion on the regional scale of the convection geometry may be possible.

Whether or not convectively coupled stress fields are able to generate major horizontal motion, they could generate significant tectonic deformation and induce large changes in topography, such as uplift over rising plumes. If gravity anomaly magnitude is used as a guide to stress level (37), dynamically maintained topography may be subject to large-scale failure; strength envelope analysis (38) suggests that the upper crust of Venus fails in a brittle manner whereas the lower crust fails by ductile flow. Large-scale motion would display a distributed deformation (39) instead of narrowly focused failure at plate boundaries.

Table 1. ADC values for Venusian highlands. Values in column 2 were determined from simulation modeling with a single orbit of Pioneer Venus data [from (19), except Gula Mons]; values in column 3 were determined by various modeling techniques, generally with the use of multiple orbits.

Feature	ADC (km)	Other ADC estimates (km)
Asteria Regio	200 ± 75	
Atla Regio	200 ± 23	200 to 250 (88)
Bell Regio	170 ± 61	110 (88)
Beta Regio	270 ± 26	300 to 400 (6, 88)
Gula Mons	190 ± 20	
Ishtar Terra (w.)		150 to 180 (71, 72)
Nokomis Montes	230 ± 45	
Ovda Regio	70 ± 7	75 (88)
Phoebe Regio	60 ± 39	
Sappho Patera	160 ± 41	
Tellus Regio	70 ± 21	30 (88)
Thetis Regio	90 ± 11	100 (88)
Ulfrun Regio	50 ± 41	

Initiation and maintenance of subduction. In comparison to Earth, resistance to subduction on Venus could be less because of a hotter, thinner lithosphere, and slab pull could be a resistive force because of positive, not negative, buoyancy. For Earth, resistance to the initiation and then maintenance of subduction appears to be supplied largely by the shear resistance of the lithosphere. The magnitude of this force depends on crustal thickness, temperature gradient, strain rate, and dip of the subduction zone in the lithosphere. There are plausible values of these parameters for which the lithospheric resisting force on Venus might be less than the shear coupling driving force (40).

Greater shear resistance forces can, of course, be overcome by slab pull. Unless the crust is no thicker than a few kilometers, however, the lithosphere will be positively buoyant (6) and will resist subduction until it reaches a depth where eclogite can be formed. For temperature gradients of 10° to 20°C km^{-1}, the basalt-eclogite transition is at depths of about 55 to 85 km (41). Thermal model calculations indicate that a lithosphere with a 20-km basaltic crust has a negative density contrast with the underlying mantle of about 0.06 Mg m^{-3} (6). By the time a 100-km-thick lithosphere has been subducted to a depth of 50 km, a resisting force of about 3×10^{12} N m^{-1} is attained. This force is comparable to the convective coupling force; therefore, it is not clear whether significant underthrusting can be achieved, and a subduction depth several tens of kilometers beneath the basalt-eclogite transition might be required to achieve even neutral buoyancy in the subducted portion of the lithosphere.

Unequivocal evidence for Venusian subduction zones is lacking. Although some underthrusting may be occurring beneath western Ishtar Terra (14, 42, 43) and the ridge belts (44), it is doubtful that zones such as these drive global motions.

Ductile décollements. Although an LVZ may be absent at the base of the lithosphere on Venus, some lithospheric motion could occur if ductile décollements develop in the lower part of the crust in response to the pressure supplied by elevated regions. For crustal thicknesses in excess of 15 km and temperature gradients exceeding 15°C km^{-1}, velocities in excess of 0.1 cm $year^{-1}$ are possible on even small topographic slopes (45). Development of ductile detachments would lead to a thin-skinned style of deformation that might mask the effects of motion of the mantle below (4). At zones of lithospheric compression, obduction or overthrusting might be expected to occur. The presence of décollements will make the establishment of subduction difficult because work must be expended both in thickening the crust and in overcoming the shear resistance at the base of the thrust sheet.

Under a lithostatic stress assumption, pushing forces in the crust of approximately 5×10^{11} N m^{-1} for each kilometer of elevated topography are possible with a detachment in the lower crust (46). The strength of a basaltic crust that is more than 10 km thick is about 7×10^{11} N m^{-1}, assuming a temperature gradient of 15°C km^{-1} (37). This suggests that significant tectonic deformation and crustal thickening could commence on the flanks of uplands that exceed 1 or 2 km in elevation.

Summary. Taken in total, our picture is that lithospheric motion on Venus probably does not extend over large distances. Large-scale thrust faults exist, but true subduction is not developed. The tectonic evolution of Venus can be couched in terms of a hot spot–plume model that is dominated by vertical tectonics, with the resulting horizontal forces being driven by gravity; that is, large-scale lithospheric motion is not significant in the overall scheme. The absence of an LVZ means that plumes can interact directly with the lithosphere; that is, their stress fields are not attenuated in an asthenosphere.

Hot Spots and Mantle Convection

The hot-spot hypothesis (*3, 5, 6*) was suggested from several lines of evidence drawn from both morphological and geophysical observations. Many of the large topographically elevated areas in the tropical and subtropical latitudes of Venus are quasi-circular and isolated (Fig. 1) and have large apparent depths of isostatic compensation (>100 km). If the heat source concentration in Venus is like that in Earth (*3*), these levels of apparent compensation are deeper than can reasonably be expected for passive support; they must be supported, at least in part, dynamically, that is, by convective upwellings in the mantle. Beta Regio, the "type" hot spot on Venus, also displays extensive volcanism and rifting. Because the hot-spot hypothesis is based on observations of the present-day gravity field, topography, and tectonics, it is evident that it applies to the way that Venus has been operating over recent geological times, perhaps a billion years or less as indicated by estimates of the surface age derived from cratering rate models (*47*). We do not comment on how Venus may have functioned in more ancient times.

Mechanisms of heat loss. Venera observations of the northern quarter of Venus have established in two ways that lithospheric recycling is not sufficient on Venus to account for much of the planet's heat loss. First, the impact crater frequency indicates that the surface is too old to be undergoing recycling at terrestrial rates (*48*); a recent estimate of the crater retention age of Venus (*47*) limits the heat loss to less than about 10% of the total budget. Second, no evidence was found in Venera images of the rolling plains regions for low fast-spreading ridges that were thought necessary to accommodate heat loss dominated by lithospheric recycling (*49*). Therefore, an earlier bound (7) of 15% on the heat loss due to lithospheric recycling, based on the assumption that the equatorial highlands act as divergent margins, remains valid (*50*). Such small upper bounds to the heat loss by lithospheric recycling are consistent with the difficulty of developing large-scale motions discussed above.

Because the crater retention age of the surface similarly excludes volcanism as an important mechanism of heat loss (*48*), lithospheric conduction must dominate. Nevertheless, there remains the question of how heat is delivered to that level. Numerical studies of convection in a spherical shell show that for a planetary mantle with contributions to heating both from radioactive elements within and from core cooling, some fraction of the total heat flux is delivered by major plumes arising from instabilities in the core-mantle boundary layer, and some fraction is delivered in a more diffuse manner from weak upwellings of the internally heated component (*51*).

Because the significance of plumes depends on the amount of heat removed from the core, we consider the nature of the Venusian magnetic field to obtain a crude assessment of this quantity. The magnetic dipole moment of Venus is less than 1/25,000th of the terrestrial moment (*52*); Venus does not possess a geodynamo-generated magnetic field. The magnetic moment is thought to scale linearly with planetary rotation rate (*53*), so that the 243-day rotational period of Venus does not appear to account for its lack of a dipole field. To generate a geodynamo, the liquid portion of a planetary core must have sufficient energy sources to be convecting. Heating by exothermic freezing of the liquid core combined with secular cooling is the most likely source of energy enabling core convection and driving Earth's present-day geodynamo (*54*). The absence of a magnetic field suggests that Venus does not have a freezing inner core, as plausible parameterized convection solutions demonstrate (*54*). The implication is that the heat flux from the Venusian core is less than that for Earth. In turn, we would expect that the thickness of the core-mantle thermal boundary is greater, and the effective Rayleigh number of the mantle lower, for Venus. In comparison to the case for Earth, Venusian plumes may have larger horizontal dimensions because the effective mantle viscosity may be higher, and plumes may be rarer. The horizontal dimension of a plume head scales linearly with the diameter of the induced topographic swell (*55*). If the Hawaiian swell is representative of terrestrial hot-spot features, then major plumes on Venus (for example, those associated with Beta, Ovda, and Thetis regiones) are about three times the radius of their terrestrial counterparts. We suggest that most heat escaping Venus may be internally generated in the mantle and delivered in spatially and temporally variable zones of weak mantle upwelling, which is manifested at the surface by broadly distributed volcanism in the rolling plains and as fields of small domes (*56*). In the absence of sea-floor spreading and subduction, however, the effects of major plumes should dominate the surface tectonics of the planet because of their large and concentrated buoyancy flux and because their effects are not muted by an LVZ. Dynamic compensation by large transient plumes from the core-mantle boundary can account for the major highland regions of Venus. Smaller plumes, whether from the upper or the lower mantle, may affect the surface as coronae (*57*).

Where is the return flow? Return flow in a convective system is associated with instabilities in the cold upper thermal boundary layer and consists mostly of thin, linear, connected sheets and occasional cylinders (*51*). If long-wavelength positive topography (that is, the regiones) is supported dynamically, there is no reason to suspect that long-wavelength negative topography is not also supported in the same way; that is, the linear planitiae (for example, Guinevere, Sedna, and Niobe) mark the site of active mantle downwellings (*58, 59*). Using an orbit simulation procedure (*60*), we have estimated an ADC of 100 to 150 km beneath Guinevere and Sedna and of 180 km beneath Niobe. These compensation depths are comparable to those of highland regions and suggest a common mechanism. Bindschadler *et al.* (*58*) have also pointed out the possibility of cylindrical mantle downwellings and that large circular basins such as Atalanta Planitia and Lavinia Planitia are associated with such structures.

One problem in this interpretation is that the linear planitiae show evidence of extensive volcanism, which would not be expected of a region where downwelling is occurring. The volcanism is of uncertain age and may be older than the downwellings, however, and the boundaries between broadly diffusive upwellings (expected to be responsible for volcanism) and linear downwellings are not well defined. Also, some remelting associated with downwellings might be possible.

Plume Evolution

Gravity, topography, and magmatism. When a convective instability breaks upward from a thermal boundary layer, a plume head will form to accommodate the mass flux if material is fed into the plume faster than the plume can rise through the mantle (*23, 24*). As the large, relatively hot plume head reaches the uppermost part of the mantle, it can undergo partial melting, leading to large-scale magmatic activity. Richards *et al.* (*24*) suggested that the quasi-steady-state hot spots observed on Earth are associated with the feeder conduit that follows the plume head. Hot-spot tracks are observed on Earth because plate motion moves the massive magmatism associated with the plume head (for example, the Deccan Traps) away from the trailing conduit. If there is no plate motion on Venus, then the magmatic effects of the conduit may be masked.

Originally, Phillips and Malin (*6*) interpreted all the equatorial highlands of Venus (Beta Regio, Eistla Regio, and Aphrodite Terra) as hot-spot features, although possibly in different stages of evolution. Herrick and Phillips (*61*) suggested that various features represent a specific evolutionary sequence in the interaction of a plume with the lithosphere, the development path being exhibited by differences observed from Beta Regio (youngest), to Thetis Regio, to Ovda Regio, to the Artemis Chasma region, to the semicircular region west of Ovda Regio (including Hestia Rupes). In this sequence, the topography and style of volcanism and magmatism are functions of the depth of the ascending plume head, which can be traced through the ADC (Table 1). Early in the sequence (Beta Regio), the ADC is large, the lithosphere is rifted and otherwise deformed in response to the normal and shear stresses associated with uplift (*37*), and volcanism is limited because the plume head is subject to high lithostatic pressures. As the plume head ascends (Thetis Regio and then Ovda Regio), the ADC decreases, and a decrease in pressure leads to large volumes of partial melt, substantial addition of crust, and flood-basalt volcanism (*61*). For example, if the plume head temperature is elevated an additional 200°C above the ambient mantle temperature, then about 20 km of new crust could be generated if the plume head ascends to within about 60 km of the surface (*62*).

As the plume head penetrates the base of the lithosphere, magmatic activity will eventually decrease as the equilibrium fraction of partial melt is attained. Flattening of the plume head will lead to a decrease in buoyancy pressure, diminishing dynamic support and leading to partial collapse of topography. Magmatism and buoyancy flux associated with the plume head will dominate over material rising in the trailing conduit or plume tail in this model. Additional topography will be supported by isostatically thickened crust (*63*) and thermally thinned lithosphere (*8*). Thus topography will evolve from a broad swell supported dynamically by the plume to the development of a crustal plateau compensated by lithospheric mechanisms and associated with an abating component of dynamic uplift (Fig. 2). Ultimately, the former site of mantle upwelling will be marked by a plateau alone. Such plateaus will have a finite lifetime because continued slow creep will eventually diminish the plateau topography to the point where the former plateau is indistinguishable from surrounding rolling plains and deformational units are broken up and buried by plains volcanism (*64*). Simple viscous relaxation calculations (*65*) indicate that crustal plateau lifetimes are of order 0.5×10^9 to 1.0×10^9 years. Thus a plume or hot-spot cycle (Fig. 3) will include (i) creation of dynamic topography and associated deformation due to uplift, (ii) generation of a crustal plateau, (iii) collapse of dynamic support, and (iv) slow creep relaxation of the plateau.

Tectonics. Several sources of stress are present in the lithosphere during the plume cycle. These include (i) largely extensional deformation associated with uplift (*37*); (ii) forces associated with ductile detachments, which may lead to thickened crust residing on the slopes of uplifts (*45*); (iii) membrane compressional forces acting on the new crustal block as dynamic topography subsides; (iv) extensional and compressional forces associated with spreading of the crustal block (*66*); and (v) thermoelastic stresses as the hot spot cools, particularly in the new crustal cap. The superposition of several sources of stress will lead to complex deformation of a crustal plateau; because such deformation is continuous, these features are not likely to preserve impact craters over long periods.

Complexly deformed regions on Venus, typically less than 1 km in average elevation and 1 to 2 km in peak elevation, are known as tessera and consist of "sets of ridges and grooves, commonly intersecting each other" (*21,* p. D399). Tessera may be the surface signature of tectonically deformed crustal plateaus (*67*) and mark the sites of former plumes (for example, Laima tessera and Tellus and Alpha regiones but also Ishtar Terra, which, as shall be argued below, has been subject to disruption by a fresh plume).

The ADC values (Table 1) for highland features appear to be resolved into two broad groups associated with opposite phases of the hot-spot cycle: (i) highland features (Nokomis Montes, Gula Mons, and Atla, Asteria, Bell, Beta, and Sappho regiones) that have a large component of dynamic compensation and only a modest crustal component, and (ii) highland features (Ovda, Phoebe, Tellus, Thetis, and Ulfrun regiones) that are tesserated crustal

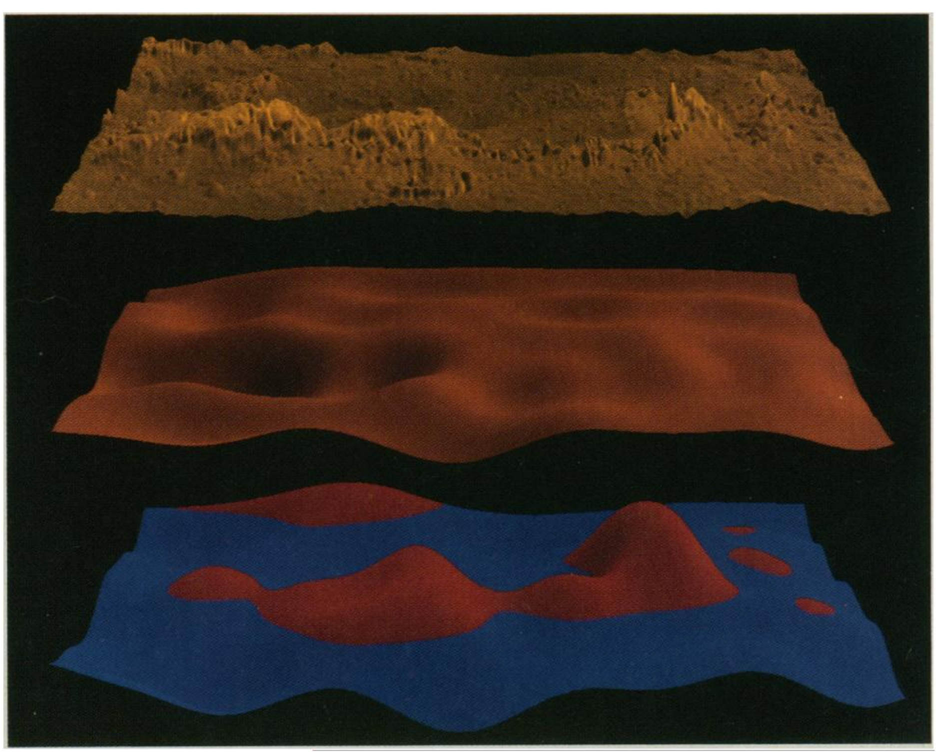

Fig. 2. Global dynamic model for Venus (*59*), simple cylindrical projection of the eastern hemisphere viewed in perspective from the south. (**Top**) Surface topography. (**Middle**) Relief on crust-mantle boundary. (**Bottom**) Distribution and relative strength of plumes. A strong plume component is evident for Atla, and a crustal component exists for Ovda and Thetis.

plateaus and are supported in the lithosphere by crustal and thermal compensation mechanisms (*19*). Four of the seven features in the first group can be described morphologically as broad domes capped with steeper peaks or small plateaus; four of the five features in the second group plus Asteria Regio in the first group can be described as plateaus (*19*). In the second group, Tellus Regio is a known tessera, and the other three plateaulike features in this group (unmapped by Venera) are predicted to be tessera on the basis of radar reflectivity and roughness properties mapped by Pioneer Venus (*68*).

Other Features

Western Ishtar Terra. A major anomaly in the hot-spot model is western Ishtar Terra, specifically Lakshmi Planum and its surrounding mountain belts. If locally compensated, these mountain belts attest to crustal shortening of at least 100 to 200 km. Terrestrial analogy suggests that such strain patterns are more plausibly the result of horizontal convergence (*14*) and mantle downwelling (*58, 69*) than of plume upwelling (*70*). A cylindrical mantle downwelling, which could lead to convergent tectonics at the surface, is not inconsistent with the global model of Venus given here (Fig. 4A). Although gravity highs are observed over terrestrial subduction zones, the positive free-air gravity signature of western Ishtar Terra is difficult to account for quantitatively as the result of regional convergence or mantle downwelling (*71*). The ADC of 150 to 180 km defined by the gravity and topography data (*71, 72*) is, however, quite consistent with global ADC values of 100 to 200 km (*73*), and western Ishtar Terra may be understood within the framework of a plume-tectonics model if it represents the site of a second-generation plume (*70*). A hot mantle upwelling beneath a tessera (Fig. 4B) of thickened crust formed by an earlier plume could result in a different tectonic signature than that seen in first-generation plumes. In particular, lithospheric strength can be strongly heterogeneous laterally because a thickened crustal pile would create a central weak zone. Shear tractions exerted by an impinging mantle plume may thin crust in the center of the weakened zone and transport it outward. The margins of the tessera would act as a strength barrier where the crust returns to normal thickness and the lithosphere is no longer weakened. Here outward flow is attenuated, and crust is accumulated into mountain belts. Melting associated with the rising plume also generates volcanism, giving rise to the relatively smooth plains of Lakshmi Planum covering the older tessera (*74*). This model, although quantitatively untested, provides a specific mechanism to accommodate earlier suggestions (*64, 75*) of mountain belt formation around a plume upwelling.

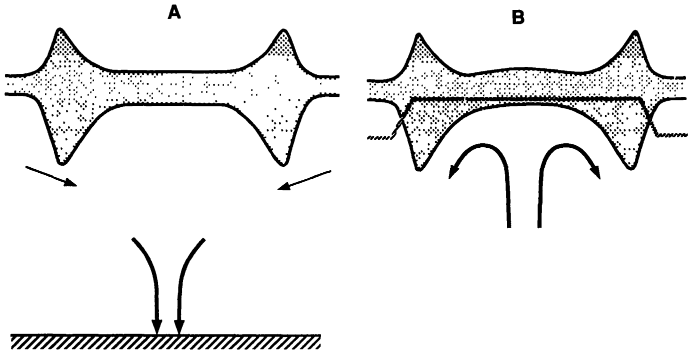

Fig. 4. Schematic illustrations of alternative models for western Ishtar Terra that are consistent with regional gravity and topography (*70*). (**A**) Compact, deep mantle downwelling sinks onto the stronger lower mantle; near-surface flow shortens the crust. (**B**) Broad, shallow plume drags the thick, weak crust outward, forming mountain belts.

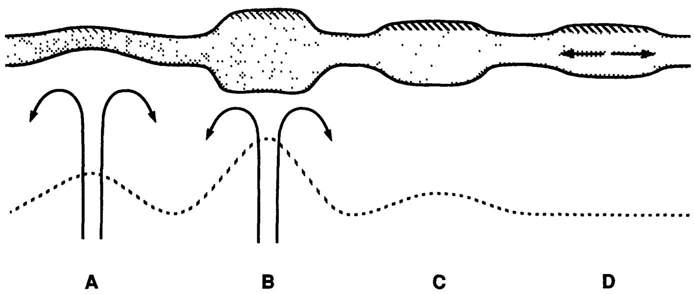

Fig. 3. Plume cycle and the evolution of topography on Venus. (**A**) Upwelling plume (arrows) begins to thin the thermal lithosphere (dashed line), causing limited volcanism, and domes the surface. Uplift stresses cause brittle failure in the upper crust and ductile failure in the lower crust (*37*), forming tessera. (**B**) Massive partial melting of the plume head forms a crustal plateau about 20 km thick. (**C**) Plume abates and dynamic topography collapses, causing continued surface deformation; thermoelastic stresses due to cooling of the lithosphere may also contribute. (**D**) Residual plume plateau relaxes viscously in 0.5×10^9 to 1×10^9 years. Late deformation is caused by this crustal creep.

Coronae. Circular structures 150 to 1000 km in size defined by an annulus of concentric ridges around a mildly raised interior are abundant on Venus (*20*). These features, called coronae, are probably also manifestations of a sequence of transient mantle upwelling involving uplift, tectonic deformation, volcanism, and collapse (*57*). The relation between this type of mantle plume and the types that form the principal highlands is not clear, however. Western Ishtar Terra (*64, 70*) and Artemis Chasma (*61, 76*) have been described in the context of both major hot-spot highlands and megacoronae. The largest coronae are still somewhat smaller than the highlands, suggesting a smaller plume. Because the thermal convection arguments above suggest that plumes from the core-mantle boundary greatly exceed the mean 400-km diameter for coronae, the latter features may therefore be related to a different regime or scale of mantle upwelling. One hypothesis is that coronae form above locally concentrated flows within the broader background of diffuse upwelling that expels the large part of the mantle heat of Venus. Alternatively, if the mantle of Venus is stratified, coronae could be related to a different scale of convection arising from boundary layers within the mantle.

Ridge belts. Despite their large areal extent and probable tectonic origin (*77–79*), ridge belts remain an unexplained element in Venusian global tectonic models. The same morphologic and geometric features have been supported as evidence of extension (*78, 79*) and compression (*21*). On the basis of pre-Magellan observations of morphology alone, neither form of strain can be proved or disproved.

Zuber (*77*) argued on the basis of instability models that compression accompanying downwelling mantle best explains the large-scale form of the ridge belts (Fig. 5), citing as evidence primarily the swell-and-swale spacing of 300 to 400 km. The compressional stresses would arise largely from shear coupling associated with downgoing flow (*37*). The apparent youth of the ridge belts (*78*) and the lack of surface uplift (resulting from thickening of the crust) may also indicate that the downwelling is incipient (*77*). As it presently stands, however, this model is insufficiently detailed to provide the range of morphological features observed because both extension and compression may have occurred at scales smaller than several hundreds of kilometers.

We have argued above that linear planitiae could be the sites of mantle downwelling, and we concur that ridge belts could mark the first stage of this process (77). If this model is correct, then ridge belts should be found in linear planitiae, possibly parallel to topographic strike. This does not appear to be the case, and plains volcanism must be invoked to bury the evidence.

Alternatively, ridge belts may mark the sites of former downwellings. As a downwelling becomes extinct or migrates substantially (*58*), the linear planitia would rebound (Fig. 5). The subsequent deformation would depend on the stress state at the initiation of the process, but compressional membrane stresses should dominate if any earlier deviatoric stresses had relaxed. In either model, ridge belts could be geologically young (*78*).

A potential tectonic response to downwelling is thickening of the crust resulting from inward flow toward the dynamically maintained low (*58*). Linear planitiae would then evolve to linear highlands with relatively shallow depths of compensation instead of to ridge belts. A candidate for this process is Ulfrun Regio.

Crustal Evolution

Without large-scale divergence of the lithosphere, globally interconnected access of pressure-release melting to the surface is lost (*80*). If this is so for Venus, then generation of crust would be restricted to hot spots, whether at large plumes rising from the core-mantle boundary or at widely distributed sites of weak upwelling from an internally heated mantle. Bounds on the rate of crustal generation on Venus may be established from the volume and lifetime of magmatic uplands and from the impact crater frequency. One can estimate the mean rate at which crust is produced by plumes by dividing the present excess volume (topography plus root) V of crustal plateaus by their average lifetime t. Because calculations based on individual crustal plateaus would be subject to considerable uncertainty, an upper bound $V = 3.3 \times 10^8$ km^3 may be found simply by integrating the volume under all elevations (*15*) above the mean planetary radius under the assumption of Airy isostasy. If the lifetime of relief is limited by creep, then a minimum viscous relaxation time $t = 0.5 \times 10^9$ years may be used to ensure a maximum in crustal production rate. These values yield a maximum mean rate of crustal generation of 0.7 km^3 $year^{-1}$, or 1.4 km of crustal thickness per 10^9 years if distributed planetwide. The global mean rate of volcanism may be estimated from the rate at which impact craters are obliterated (*48*). The effective crater retention age for plains in the northern quarter of Venus is about 0.4×10^9 years using obvious impact craters and 2×10^9 years if a larger group of structures that includes potential degraded impact craters is considered (*47*). Vertical resurfacing rates of 1.3 and 0.3 km per 10^9 years follow (*47, 48*) for the obvious and degraded groups, respectively. If intrusive rocks make up 80 to 90% of the total magma volume (*81*), then the corresponding vertical rates of crustal generation are 9 and 2 km per 10^9 years, respectively. Because vertical obliteration produces a continuous spectrum of crater preservation states, use of the extended group is more appropriate, although subject to greater uncertainty as a result of ambiguities in crater identification. These uncertainties, however, as well as geographical variations in crater density (*82*), are unlikely to alter these figures substantially. Therefore, the net rate of crustal generation on Venus appears to be less than 10 km per 10^9 years over the past 0.4×10^9 to 2.0×10^9 years.

The typical thickness of the Venusian crust has been estimated to be less than 10 to 30 km (*83, 84*), which suggests that little or no crustal recycling is required to accommodate the maximum rates of crustal generation of several kilometers per 10^9 years derived above. Higher crustal production rates in the past are plausible from the perspective of decreasing secular heat loss of the planet. Recycling would take place when crust is depressed far enough to undergo transitions to denser phases or to remelt, which can occur at depths greater than 30 to 80 km on Venus (*85*). Because the crust apparently does not fill the entire stability volume, earlier vertical recycling (*86*) does not appear to be necessary. Higher crustal production in the past may have been accommodated by lateral recycling (underthrusting and subduction). Such processes may be occurring locally today (*14, 44*), but there is little need to invoke this mechanism given the low crustal production rate. On Earth, new crust is created principally at mid-ocean ridges at a rate of 18 km^3 $year^{-1}$ (*87*). Island arc and intraplate volcanism contribute less than 2 km^3 $year^{-1}$ (*10*). The net crustal production rate on Venus, less than about 1 km^3 $year^{-1}$, appears to be comparable to that at terrestrial hot spots.

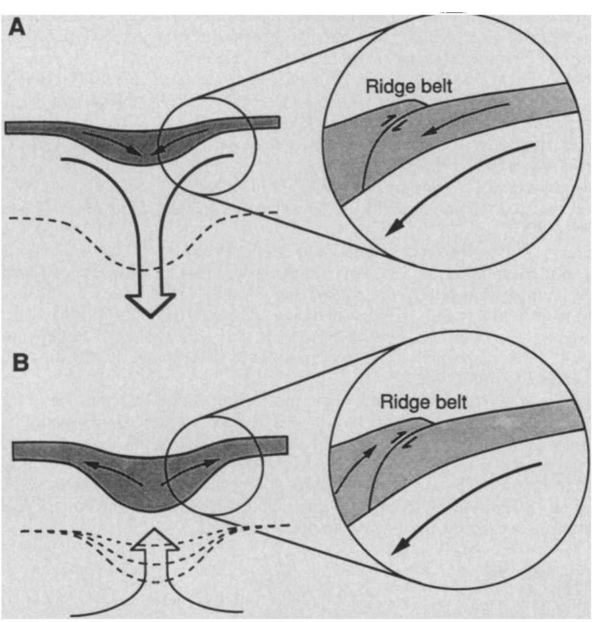

Fig. 5. Schematic illustrations of alternative models for ridge belts, whose deformation is assumed to be largely contractional. (**A**) Downwelling drag; (**B**) rebound thrusting.

Conclusions

Our viewpoint is strongly influenced by the long-wavelength gravity and topography and the inferences that follow from these data, namely, that Venus does not possess an LVZ beneath its lithosphere and that variations in the gravity signature may be interpreted in terms of transient mantle plumes rather than steady-state convection. Although there is not agreement among geophysicists on the role of an LVZ in terrestrial sea-floor spreading, we presume that it is necessary to achieve at least the large ridge-to-trench length scales that exist. On the contrary, we believe that the lack of an LVZ on Venus leads to regional deformation that is driven by direct coupling of mantle flow into the lithosphere. Horizontal motion at the surface is probably limited because shear resistance to such motion and buoyant resistance to subduction are stronger than the shear coupling driving force. The major geological features of the planet may be accounted for in two evolutionary sequences: upwellings and downwellings. In the first sequence, a mantle plume produces a broad domal uplift, and partial melting leads to a massive crustal block that outlives the plume. The sequence for downwellings is less certain but results in formation of linear (and possibly circular) planitiae and ridge belts.

REFERENCES AND NOTES

1. Unless otherwise stated, we are referring to the thermal lithosphere, the outer shell of the planet that transports heat predominantly by conduction, not convection.
2. W. M. Kaula, *Icarus* **26**, 1 (1975).
3. R. J. Phillips and M. C. Malin, *Annu. Rev. Earth Planet. Sci.* **12**, 411 (1984).
4. W. M. Kaula, *Science* **247**, 1191 (1990).
5. R. J. Phillips, W. M. Kaula, G. E. McGill, M. C. Malin, *ibid.* **212**, 879 (1981).
6. R. J. Phillips and M. C. Malin, in *Venus*, D. M. Hunten, L. Colin, T. M. Donahue, V. I. Moroz, Eds. (Univ. of Arizona Press, Tucson, 1983), pp. 159–214.
7. W. M. Kaula and R. J. Phillips, *Geophys. Res. Lett.* **8**, 1187 (1981).
8. P. Morgan and R. J. Phillips, *J. Geophys. Res.* **88**, 8305 (1983).
9. J. T. Wilson, *Sci. Am.* **28**, 86 (April 1963); W. J. Morgan, *Geol. Soc. Am. Bull.* **132**, 7 (1972); in *The Sea*, C. Emiliani, Ed. (Wiley, New York, 1981), pp. 443–487.
10. A. Reymer and G. Schubert, *Tectonics* **3**, 63 (1984).
11. J. W. Head and L. S. Crumpler, *Nature* **346**, 525 (1990).
12. J. W. Head, *J. Geophys. Res.* **95**, 7119 (1990).
13. K. M. Roberts and J. W. Head, *Geophys. Res. Lett.* **17**, 1341 (1990).
14. J. W. Head, *Geology* **18**, 99 (1990).
15. G. H. Pettengill *et al.*, *J. Geophys. Res.* **85**, 8261 (1980); P. Ford, *Pioneer Venus Hypsometry* (MIT Center for Space Research, Massachusetts Institute of Technology, Cambridge, 1986).
16. H. Masursky *et al.*, *J. Geophys. Res.* **85**, 8232 (1980).
17. G. E. McGill, S. J. Steenstrup, C. Barton, P. G. Ford, *Geophys. Res. Lett.* **8**, 737 (1981); B. A. Campbell and D. B. Campbell, *ibid.* **17**, 1353 (1990).
18. W. L. Sjogren *et al.*, *J. Geophys. Res.* **88**, 1119 (1983).
19. S. E. Smrekar and R. J. Phillips, in preparation; *Lunar Planet. Sci.* **21**, 1176 (1990).
20. V. L. Barsukov *et al.*, *J. Geophys. Res.* **91**, D378 (1986).
21. A. T. Basilevsky *et al.*, *ibid.*, p. D399.
22. D. B. Campbell, J. W. Head, J. K. Harmon, A. A. Hine, *Science* **221**, 644 (1983); *ibid.* **226**, 167 (1984); D. B. Campbell *et al.*, *ibid.* **246**, 373 (1989).
23. R. S. White and D. P. McKenzie, *J. Geophys. Res.* **94**, 7685 (1989); I. H. Campbell and R. W. Griffiths, *Earth Planet. Sci. Lett.* **99**, 79 (1990).
24. M. A. Richards, R. A. Duncan, V. E. Courtillot, *Science* **246**, 103 (1989).
25. Other overviews of the way Venus might work are available in the recent literature, including the crustal spreading/mantle plume model discussed in the text (*4, 11, 35*). See also D. R. Williams and V. Pan, *Geophys. Res. Lett.* **17**, 1397 (1990).
26. B. H. Hager and R. J. O'Connell, *J. Geophys. Res.* **86**, 4843 (1981).
27. S. Cloetingh, R. Wortel, N. J. Vlaar, *Pure Appl. Geophys.* **129**, 7 (1989).
28. S. Mueller and R. J. Phillips, *J. Geophys. Res.* **96**, 651 (1991).
29. D. A. Wiens and S. Stein, *ibid.* **89**, 11442 (1984).
30. R. M. Richardson, S. C. Solomon, N. H. Sleep, *Rev. Geophys. Space Phys.* **17**, 981 (1979).
31. H. J. Melosh, *J. Geophys. Res.* **81**, 5621 (1976).
32. D. E. Wiens and S. Stein, *Tectonophysics* **116**, 143 (1985).
33. C. H. Craig and D. McKenzie, *Earth Planet. Sci. Lett.* **78**, 420 (1986).
34. J. Melosh, *Pure Appl. Geophys.* **115**, 429 (1977).
35. W. M. Kaula, *Geophys. Res. Lett.* **17**, 1401 (1990).
36. This can be determined by use of the appropriate parameters given in (7) and the simple formula for ridge push given by D. P. McKenzie [in *Island Arcs, Deep Sea Trenches, and Back-Arc Basins*, M. Talwani and W. C. Pitman III, Eds. (American Geophysical Union, Washington, DC, 1977), pp. 57–61].
37. R. J. Phillips, *J. Geophys. Res.* **95**, 1301 (1990).
38. W. F. Brace and D. L. Kohlstedt, *ibid.* **85**, 6248 (1980).
39. R. E. Grimm and S. C. Solomon, *ibid.* **94**, 12103 (1989).
40. A set of plausible values is as follows: crustal thickness, 30 km; temperature gradient, 10°C km^{-1}; strain rate, 10^{-15} s^{-1}; and thrust fault dip, 10°. The lithospheric resisting force calculation is based on the method described in (*28*) applied to Venus conditions.
41. H. S. Yoder, Jr., *Generation of Basaltic Magma* (National Academy of Sciences, Washington, DC, 1976).
42. L. S. Crumpler, J. W. Head, D. B. Campbell, *Geology* **14**, 1301 (1986).
43. S. C. Solomon and J. W. Head, *Geophys. Res. Lett.* **17**, 1393 (1990).
44. S. L. Frank and J. W. Head, *Earth Moon Planets* **50/51**, 421 (1990).
45. S. Smrekar and R. J. Phillips, *Geophys. Res. Lett.* **15**, 693 (1988).
46. The crustal pushing force is calculated from $F_{cp} = \rho_c g h T$, where ρ_c is crustal density, h is the height of maximum topography above the point of reference, and T is crustal thickness above the detachment, taken here as 20 km.
47. R. E. Arvidson, R. E. Grimm, R. J. Phillips, G. G. Schaber, E. M. Shoemaker, *Geophys. Res. Lett.* **17**, 1385 (1990).
48. R. E. Grimm and S. C. Solomon, *ibid.* **14**, 538 (1987).
49. S. C. Solomon and J. W. Head, *J. Geophys. Res.* **87**, 9236 (1982).
50. For comparison, lithospheric recycling accounts for about 65% of the terrestrial heat loss. See J. G. Sclater, C. Jaupart, D. Galston, *Rev. Geophys. Space. Phys.* **18**, 269 (1980).
51. D. Bercovici, G. Schubert, G. A. Glatzmaier, *Science* **244**, 950 (1989).
52. C. T. Russell and O. Vaisberg, in *Venus,* D. M. Hunten, L. Colin, T. M. Donahue, V. I. Moroz, Eds. (Univ. of Arizona Press, Tucson, 1983), pp. 873–940.
53. F. H. Busse, *Phys. Earth Planet. Inter.* **12**, 350 (1976).
54. D. J. Stevenson, T. Spohn, G. Schubert, *Icarus* **54**, 466 (1983).
55. P. Olson and I. S. Nam, *J. Geophys. Res.* **91**, 7181 (1986).
56. J. C. Aubele and E. N. Slyuta, *Earth Moon Planets* **50/51**, 493 (1990).
57. E. R. Stofan and J. W. Head, *Icarus* **83**, 216 (1989).
58. D. L. Bindschadler, G. Schubert, W. M. Kaula, *Geophys. Res. Lett.* **17**, 1345 (1990).
59. R. R. Herrick and R. J. Phillips, *Lunar Planet. Sci.* **21**, 499 (1990).
60. R. J. Phillips, W. L. Sjogren, E. A. Abbott, S. H. Zisk, *J. Geophys. Res.* **83**, 5455 (1978).
61. R. R. Herrick and R. J. Phillips, *Geophys. Res. Lett.* **17**, 2129 (1990).
62. This is based on the partial melting model of D. McKenzie and M. J. Bickle [*J. Petrol.* **29**, 625 (1988)] used in conjunction with a parameterized convection calculation.
63. The excess elevation will be given by $T(\rho_m - \rho_c)/\rho_m$, where T is the excess crustal thickness and ρ_c and ρ_m are the crust and mantle densities, respectively. Topographic elevation of about 10% of T will be produced.
64. A. Nikishin, *Earth Moon Planets* **50/51**, 101 (1990).
65. Plume-plateau lifetimes are taken to be one to two times the long-wavelength limit for the $1/e$ relaxation time of locally compensated relief on an isoviscous layer overlying an inviscid half-space; see S. C. Solomon, R. P. Comer, J. W. Head, *J. Geophys. Res.* **87**, 3975 (1982). An average lithospheric viscosity of 3×10^{23} Pa s was used, following from crustal thicknesses H of 10 to 20 km at thermal gradients dT/dz of 10° to 20°C per kilometer. These bounds on H and dT/dz were established from the relaxation of impact craters greater than 100 km in diameter (*84*). The adopted effective viscosity is the mean value to a depth of 30 km calculated from this range using an effective stress difference of 50 MPa.
66. The initial spreading rate of a hot crustal block is probably large but should decrease as the crustal cap cools.
67. The idea that tessera are sited on crustal plateaus was originally advanced by J. W. Head [*J. Geophys. Res.* **95**, 7119 (1990)] and was developed in (*11*), where the term *plume plateau* was used. Note, however, that the specific mechanisms for generation of tessera given in this article differ from those in that model, in which primary structures are considered analogous to the pattern of abyssal hills and fracture zones formed at terrestrial divergent margins and secondary structures are produced by collision of plume plateaus [R. W. Vorder Bruegge and J. W. Head, *Earth Moon Planets* **50/51**, 251 (1990)].
68. D. L. Bindschadler *et al.*, *Geophys. Res. Lett.* **17**, 171 (1990).
69. W. S. Kiefer and B. H. Hager, in preparation.
70. R. E. Grimm and R. J. Phillips, *Geophys. Res. Lett.* **17**, 1349 (1990).
71. ———, *J. Geophys. Res.*, in press; *Lunar Planet. Sci.* **21**, 437 (1990).
72. W. L. Sjogren, B. G. Bills, N. A. Mottinger, *Geophys. Res. Lett.* **11**, 489 (1984).
73. W. S. Kiefer, M. A. Richards, B. H. Hager, B. G. Bills, *ibid.* **13**, 14 (1986).
74. K. M. Roberts and J. W. Head, *Earth Moon Planets* **50/51**, 193 (1990).
75. A. A. Pronin, *Geotectonics* **20**, 271 (1986); A. T. Basilevsky, *ibid.*, p. 282.
76. E. R. Stofan and R. S. Saunders, *Geophys. Res. Lett.* **17**, 1377 (1990).
77. M. T. Zuber, *ibid.*, p. 1369.
78. A. L. Sukhanov and A. Pronin, *Proc. Lunar Planet. Sci. Conf.* **19**, 335 (1989).
79. R. C. Kozak and G. G. Schaber, *Geophys. Res. Lett.* **16**, 175 (1989).
80. This is dramatically illustrated when the volumes of melt generated by the Hawaiian and Iceland hot spots are compared [N. H. Sleep, *J. Geophys. Res.* **95**, 6715 (1990)]. Despite the fact that the plume excess temperature (over normal mantle) is probably higher for the Hawaiian hot spot, "the average fraction of plume material which melts . . . is about 1/10 that of Iceland. Most of the plume material beneath Hawaii remains at lower lithospheric depths where extensive melting does not occur, while plume material may ascend to shallow depths along the [mid-Atlantic] ridge axis and melt extensively beneath Iceland" (*ibid.*, p. 6719).
81. J. A. Crisp, *J. Volcanol. Geotherm. Res.* **20**, 177 (1984).
82. D. B. Campbell, N. J. S. Stacy, A. A. Hine, *Geophys. Res. Lett.* **17**, 1389 (1990); B. A. Ivanov, *Earth Moon Planets* **50/51**, 159 (1990); J. W. Head and L. S. Crumpler, *Earth Moon Planets* **44**, 219 (1989).
83. W. B. Banerdt and M. P. Golombek, *J. Geophys. Res.* **93**, 4759 (1988); M. T. Zuber, *ibid* **92**, E541 (1987).
84. R. E. Grimm and S. C. Solomon, *ibid.* **93**, 11911 (1988).
85. D. L. Turcotte, *ibid.* **94**, 2779 (1989); P. Hess and J. W. Head, *Earth Moon Planets* **50/51**, 57 (1990).
86. J. L. Warner, *J. Geophys. Res.* (suppl.) **88**, A495 (1983).
87. Mean oceanic crustal thickness of 6 km multiplied by a plate creation rate of 3 km^2 year^{-1} [B. Parsons, *Geophys. J. R. Astron. Soc.* **67**, 437 (1981)].
88. Other sources of ADC values are the following: R. R. Herrick, B. G. Bills, S. A. Hall, *Geophys. Res. Lett.* **16**, 543 (1989); S. E. Smrekar and R. J. Phillips, in preparation; P. B. Esposito, W. L. Sjogren, N. A. Mottinger, B. G. Bills, E. Abbott, *Icarus* **51**, 448 (1982).

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