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LIPs on Venus

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Abstract

Venus, a planet similar to Earth in heat budget lacks plate tectonics, yet displays ample evidence of extensive volcanic and tectonomagmatic processes, including regions large enough to be considered LIPs. Thus Venus provides an excellent opportunity to examine large-scale magmatic processes outside a plate tectonic framework. I discuss four groups of Venus' largest tectonomagmatic provinces: volcanic rises, large coronae (Artemis, Heng-O, Quetzalpetlatl, Atahensik), crustal plateaus, and 'plains with wrinkle ridges', unit *pwr*. Unit pwr, widely interpreted as representing catastrophic resurfacing, has been suggested to be the solar system's largest LIP. I argue herein that these features, each covering ~1 million km^2 or more, and recording extensive volcanic activity, record different evolutionary processes, including both endogenic and exogenic processes. Volcanic rises represent surface manifestations of deep-mantle plumes on thick lithosphere. Large coronae may record different evolutionary paths. Artemis represents a plume signature on thin lithosphere. More mapping is required to determine if Heng-O formed by endogenic or exogenic processes. Preliminary mapping suggests Quetzalpetlatl represents a young mantle plume; Atahensik marks a diapiric structure, though buoyancy mode is unconstrained. Crustal plateaus record solidification of huge lava ponds formed by massive partial mantle melting caused by large bolide impact on thin lithosphere. The status of unit pwr as the largest LIP within the solar system is challenged based on geological considerations, and results of a growing body of geologic mapping.

Keywords: Large igneous province; Venus; Volcanic risel crustal plateau; Corona; Catastrophic resurfacing

1. Introduction

Currently vigorous discussion regarding the origin of large igneous provinces, LIPs, centers on whether LIPs result from plate-related processes (e.g., Anderson and Natland, 2005), or deep-mantle plumes that likely track to the core-mantle boundary (e.g., Campbell and Griffiths, 1990; 1992; Eldholm and Coffin, 2000; Ernst et al., 2005). One way to contribute to this discussion is to study a planet similar to Earth that displays large tectonomagmatic provinces, but lacks plate tectonic processes. Venus fits the bill. The value of Venus–Earth comparison results from similarities in gross planet character, and yet critical differences. Venus is expected to have a heat budget similar to Earth given its similarity in size, age, density and presumed composition; thus Venus should require mechanisms through which heat is transferred from core (or mantle) to crust. Venus lacks plate tectonic processes (Solomon et al., 1992; Phillips and Hansen, 1994; Simons et al., 1994, 1997); therefore plate-related hypotheses for LIPs are not viable on Venus. In addition, a lack of plate tectonics may mean that rich geologic histories might be preserved across Venus. In a similar vein, Venus affords a view of primary structural morphology of dynamical processes given that

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it lacks a hydrologic cycle (being much too hot for surface water to survive), and thus the surface is free of oceans, sediments, or erosion. Despite Venus' remote field location, and harsh surface environment, both of which inhibit field study, widely available NASA Magellan Mission data provide incredible global data sets which allow first-order geologic analysis, coupled with subsurface constraints based on gravity–topography data. Venus 'fieldwork' only requires computer access, making data collection, data analysis, and hypothesis testing widely accessible.

Venus displays ample evidence of extensive volcanic and tectonomagmatic processes (e.g., Head et al., 1992; Crumpler et al., 1997; Hansen et al., 1997; Smrekar et al., 1997), including provinces large enough to be considered LIPs (e.g., Head and Coffin, 1997; Ernst and Desnoyers, 2004). Venus, a planet of circles, displays circular features ranging from <1–2500 km diameters. Venus also displays several large regions (>1 million km^2) that could be considered LIPs. These large features include: 1) volcanic rises, 2) large coronae, and 3) crustal plateaus. Volcanic rises (domical forms) and crustal plateaus (plateaus-shaped) comprise large (~1000–2500 km diameter) quasi-circular topographically defined regions marked by extensive, but spatially focused, volcanic activity. Rises variably display rift or fracture belts, large volcanoes, or clusters of larger-than-average coronae. Crustal plateaus differ from rises in topographic form, as well as unique tectonic deformation fabric, called ribbon tessera terrain. Large coronae (2600–700 km diameter) include four features, Artemis, Heng-O, Quetzalpetlatl, and Atahenski. In addition to these features, many workers propose that Venus hosts a regionally extensive geologic unit variably referred to as: 'wrinkle ridge plains', 'plains with wrinkle ridges', or 'regional plains', which represents extensive volcanic activity. This unit, herein called wrinkle ridge plains, or *pwr*, following Basilevsky and Head (1996, 1998), is proposed to represent a ~2.5 km thick flood lava unit that was emplaced across ~80% of the surface in a very short time (10–100 m.y.). Unit pwr comprises the unit responsible for postulated catastrophic resurfacing of Venus (e.g., Strom et al. 1994; Herrick, 1994; Basilevsky and Head, 1996; Head and Coffin, 1997; Basilevsky et al., 1997; Basilevsky and Head, 1998, 2002; Head and Basilevsky, 2002), and the solar system's largest LIP (e.g., Head and Coffin, 1997; Ernst and Desnoyers, 2004).

In this contribution I present arguments that these LIPs could have formed through a range of processes including: deep-mantle plume–lithosphere interactions (volcanic rises and some large coronae), solidification of huge lava ponds—likely the result of massive partial mantle melting caused by large bolide impact on thin lithosphere (crustal plateaus). Unit pwr, although by far the globally most extensive, likely represents extremely low volumes of lava taken globally. Although the idea of a catastrophically resurfaced Venus permeates textbooks and popular science accounts, a growing number of studies call catastrophic resurfacing into question (e.g., Guest and Stofan, 1999; Deshon et al., 2000; Brian and Stofan, 2003; Stofan et al. 2005; Hansen, 2005; Hansen and Young, in press).

2. Background

Venus and Earth share similarities, yet they also have profound differences. Venus, 0.72 AU from the Sun, is 95 and 81.5% Earth's size and mass, respectively. Solar distance, similar mean density, and cosmo-chemical models for solar system evolution lead to the inference that Venus and Earth share similar bulk composition and heat producing elements (Wetherill, 1990). Soviet Venera and Vega landers indicate surface element abundance consistent with basaltic composition, as supported by geomorphic and geochemical arguments (Bridges, 1995, 1997; Grimm and Hess, 1997). Slow retrograde motion makes a Venus day (243 Earth days) longer than its year (~225 Earth days), a factor that may contribute to Venus' lack of a magnetic field (Yoder, 1997). Venus' atmosphere (96% CO_2, 3.5% N_2 and 0.5% H_20, H_2SO_4, HCl and HF), surface pressure (~95 bars) and temperature (~475 °C) vary significantly from Earth.

Venus' surface conditions are intimately related to its atmospheric properties. Its atmosphere includes three cloud layers ~48–70 km above the surface, which reflect visible light and block optical observation; the upper atmosphere rotates ~300 km/hr. The thick atmosphere results in negligible diurnal temperature variations, and an enhanced greenhouse prohibiting a terrestrial water cycle. Venus lacks obvious evidence of extensive sedimentary layers clearly deposited by wind or water. Venus currently lacks weathering, erosion, sediment transport and deposition processes that play dominate roles in shaping Earth's surface. Although Venus is presently ultra-dry, the past role of water is unknown. Isotopic data are consistent with, but do not require, extensive reservoirs of water ≥1 billion years ago Donahue and Russell, 1997; Donahue et al., 1997; Donahue, 1999; Lecuyer et al., 2000; Hunten, 2002). The present lack of water renders contemporary crustal rock orders of magnitude stronger than terrestrial counterparts, even given Venus' elevated surface temperature (Mackwell et al., 1998).

Most workers assume that Venus' mantle is similar in composition and temperature to Earth's mantle. A reasonable working hypothesis is that Venus' effective mantle viscosity is similar to Earth, and similarly has a strong temperature-dependent viscosity profile. However, some workers consider Venus' mantle to be stiffer than Earth's due to presumed drier conditions (e.g., 10^{20} Pa, Nimmo and McKenzie, 1998). Volatiles are of course important in understanding Venus' interior—particularly with regard to viscosity; however, interior volatile values and compositions are currently unconstrained. Lack (presence) of volatiles will increase (decrease) strength and increase (decrease) the mantle solidus. Large viscosity contrasts are likely across thermal boundary layers: notably across the lithosphere and core–mantle boundaries. In contrast to Earth however Venus lacks a low viscosity asthenosphere, likely a contributing factor to Venus' lack of plate tectonic processes. In this contribution I focus on information gleaned from the surface, although it is useful to be mindful of the global context, and numerous unconstrained values.

Current study focuses on four remote global data sets from the NASA Magellan mission: altimetry, synthetic aperture radar (SAR) imagery, emissivity, and gravity data (Ford and Pettengill, 1992; Ford et al., 1993). Emissivity is chiefly controlled by dielectric permittivity and surface roughness (Pettengill et al., 1992), and not discussed herein. Gravity data can resolve features >400 km, and provides clues to subsurface architecture, although interpretations are notably non-unique. Altimetry has a spatial resolution of 8 km (along-track) by 20 km (across-track), and vertical resolution ~30 m (Ford and Pettengill, 1992; Ford et al., 1993). SAR data (~120 m/pixel available via the web at http://pdsmaps.wr.usgs.gov/maps.html), provides the highest resolution view of Venus, and allows for geomorphic and geological interpretations, including geologic surface histories. SAR and altimetry can be digitally combined to construct synthetic stereo, 3D, views of the surface (Kirk et al., 1992). Cautions for interpretation of geologic features and histories are outlined in a variety of references (Wilhelms, 1990; Ford et al., 1993; Tanaka et al., 1994; Hansen, 2000; Zimbelman, 2001).

Venus is divisible into three topographic provinces: lowlands, highlands and mesolands. The lowlands (70–80% of the surface) include long-wavelength (thousands of km) low amplitude basins (<1 km), marked by relatively smooth low strain regions with local belts of concentrated deformation (Banerdt et al., 1997). The highlands (8–10% of the surface) host volcanic rises, crustal plateaus, and the unique feature Ishtar Terra (Hansen and Phillips, 1995; Hansen et al., 1997). Mesolands lie between the lowlands and highlands, and host many of Venus' coronae, (60–2600 km diameter quasi-circular tectonomagmatic features) and spatially associated chasmata (troughs). About 970 impact craters (1–270 km diameter) pepper the surface with a spatial distribution indistinguishable from random (Phillips et al., 1992; Schaber et al., 1992, Herrick et al., 1997; Hauck et al., 1998). A variety of volcanic landforms ranging in size from km to 100s of km occur across the surface. Lava flows, with areal extent 100s to 1000s of km, are commonly associated with volcanoes, coronae, and fractures (Head et al., 1992; Crumpler et al. 1997). Volcanic shields (1–20 km diameter), occur in shield fields (<300 km diameter regions) and as 'shield terrain' distributed across millions of km^2 (Guest et al., 1992; Crumpler et al., 1997; Hansen, 2005). Fluid cut channels, or canali, extending 10s or 1000s of km trace across the surface (Baker et al., 1997). Volcanic forms are consistent with basaltic compositions (e.g., Sakimoto and Zuber, 1995; Bridges, 1995, 1997; Stofan et al., 2000).

The highlands, mesolands and lowlands each host features and/or deposits variably interpreted as LIPs. Volcanic rises and crustal plateaus reside in the highlands; large coronae lie within all three regions, and unit pwr—the postulated ~2.5 km thick unit deformed by wrinkle ridges characterizes the lowland. Each of these features is discussed in turn below.

3. Venus and time

Before turning attention to LIPs on Venus it is important to make clear that although some definitions of LIPs include short emplacement histories, absolute geologic time cannot currently be constrained on Venus. To date, the surface density of impact craters provide the only means to constrain absolute time on planet surfaces, other than Earth. Although Venus' ~970 impact craters represent the most pristine impact craters in the solar system their low number taken together with their near random spatial distribution prohibits robust temporal constraints for individual geomorphic features or geologic units (Campbell, 1999). Impact crater density analysis results, at best, in determination of an *average model surface age* (AMSA), that is, the integrated age of a surface under consideration. Whether that surface in turn represents a single geologic unit, or feature, and as such represent the time of a geologic event, is a separate question that must be addressed independent of crater density. Venus records a *global* AMSA of ~750+350/−400 Ma, based on total impact craters and impactor flux (McKinnon et al., 1997). Because this value is a *global* AMSA it represents the average (integrated) model age of a planet's entire surface. The interpretation of this value is non-unique, similar in concept to an average mantle model age (e.g., Farmer and DePaolo, 1983); this single AMSA could be accommodated with a wide range of possible surface histories.

Impact crater dating is ultimately a statistical exercise, and includes several geological challenges. Part of the problem results from Venus' lack of small craters, due to atmospheric screening. Small craters typically comprise the largest number of carters on a planetary surface, with crater density ages dependent on binning across a range of crater diameters—a technique not possible on Venus (McKinnon et al., 1997). Venus' *global* AMSA represents the average (integrated) model age of a planet's entire surface, but could Venus could preserve smaller domains with distinct AMSA provinces? The size, location, correlation with other factors, of individual AMSA provinces could provide critical clues to the range of processes that contributed to a planet's surface evolution. Individual AMSA provinces must be statistically robust (a function of both local crater density and total crater population (Phillips 1993; Hauck et al, 1998; Campbell 1999)). The minimum size area that can be dated statistically on Venus by crater density alone is 20×10^6 km^2, or ~4.5% of the planet surface (Phillips et al., 1992). Furthermore, an area this large would require assumptions with regard to surface formation that severely limit the uniqueness of any temporal interpretation (Campbell, 1999). Some workers have attempted to date geological units by combining morphologically similar units into large composite regions for crater density dating (e.g., Namiki and Solomon, 1994; Price and Suppe, 1994; Price et al., 1996). However these studies lack statistical validity (Campbell, 1999), and they require the implicit assumption that similar appearing units formed at the same time, even if spatially separated.

No statistically distinct areas of crater density occur across Venus based on impact crater density alone.

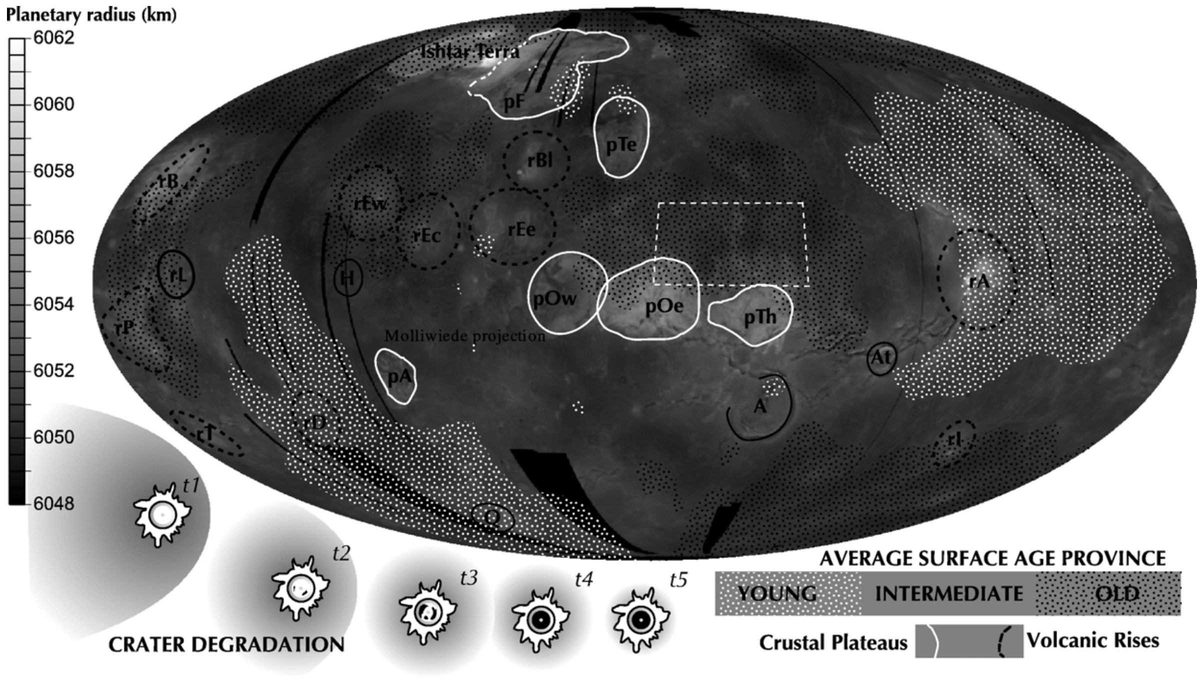

Fig. 1. Mollwiede projection of Magellan Venus altimetry with average model surface age (AMSA) provinces (data from Phillips and Izenberg, 1995) and major geologic features including: crustal plateaus (Alpha (pA), Fortuna (pF), eastern Ovda (pOe), western Ovda (pOe), Tellus (Te), and Thetis (pTh); large coronae, Artemis (A), Atahensik (At), Hengo (H), and Quetzalpetlatl (Q); and volcanic rises (Alta (rA), Beta (rB), Bell (rBl), Dione (rD), western, central and Eastern Eistla (rEw, rEc, rEe), Imdr (rI), Laudry (rL), Phoebe (rP), and Themis (rT)). Dash-line box indicates region of mapping in northern Aphrodite lowlands (see text). Crater degradation stages illustrated in the cartoon show youngest (t1) to oldest (t5) changes in crater morphology; with time and degradation an impact crater loses its halo and its interior becomes radar-smooth, presumably as a result of lava fill (Izenberg et al., 1994). AMSA provinces are defined based on impact craters density and impact crater degradation stage. Old-AMSA has high impact crater density ($>2.35\times10^6$/km^3) and a deficiency in craters with halos. Young-ASA has low impact crater density ($<1.85\times10^6$/km^3) and a deficiency in craters with halos. Intermediate-AMSA has intermediate impact crater density and no deficiency in craters with halos (Phillips and Izenberg, 1995). Figure modified from Hansen and Young (in press).

Therefore, any subdivision of the global AMSA, and resulting the determination of robust individual AMSA provinces, requires geological criteria in addition to statistical data (crater density). Impact crater morphology provides such criteria. Impact craters preserve morphological characteristics that record a temporal sequence of degradation in which crater halos are lost and crater troughs and interiors become progressively filled with radar-smooth material over time (Fig. 1) (Izenberg et al., 1994). These relations suggest that impact craters are divisible into broad relative age groups. Young craters display haloes and radar-rough interiors. Old craters lack haloes and show radar smooth interiors.

Impact crater density taken together with impact crater morphology allows delineation of three separate AMSA provinces (Phillips and Izenberg, 1995). Phillips and Izenberg (1995) reasoned that if Venus' surface only experienced weathering processes (e.g., Arvidson et al., 1992), then impact crater haloes, but not the associated craters, would disappear with time. However, if volcanic flows bury a surface, then the impact crater and the associated halo would be lost from view. Therefore, old surfaces should have an abundance of impact craters but should be statistically deficient in craters with halos; in contrast, young *volcanically resurfaced* regions should be statistically deficient in both craters and crater with halos. Fig. 1 illustrates the spatial distribution of the three AMSA provinces globally (Hansen and Young, in press). Low crater density (<1.5 craters/10^6 km^2) and a deficiency in craters with halos defines the young-AMSA province; intermediate crater density (2.5–1.5 craters/10^6 km^2) without a halo deficiency defines the intermediate-AMSA province; high crater density (>2.5 craters/10^6 km^2), and a deficiency in impact craters with halos defines the old-AMSA province. Spatial correlation of geologic time-dependent criteria (crater morphology) with impact crater density suggests that these AMSA provinces reflect true temporal domains and are not simply the result of stochastic fluctuations in a random distribution (e.g., Campbell, 1999). However, these AMSA provinces do not constrain the age of individual geologic features or units, but rather represent and average age of an integrated geologic history of a surface. The recorded surface history specifically reflects those geologic processes that would lead to formation, modification, or destruction of impact craters. The three AMSA provinces are relative, not absolute, age provinces. Although no individual geologic units or features can be robustly constrained in time, individual features, or groups of features, might show spatial patterns with respect to these three average model surface age provinces that might provides clues to their formation.

4. Possible LIPs on Venus

Venus includes several examples of possible LIPs: volcanic rises, large coronae, crustal plateaus, and unit pwr. Several first order observations emerge with regard to the relative location of possible Venus LIPs with the three AMSA provinces (Fig. 1). 1) Volcanic rises show no preferred spatial correlation with old, intermediate, or young AMSA provinces. 2) Artemis, Heng-O and Atahensik lie within the intermediate AMSA province, whereas Quetzalpetlatal lies within the young AMSA province that corresponds to Lada Terra. 3) Crustal plateaus generally lie within the intermediate-AMSA province, and completely outside the young AMSA province. 4) Large tracts of the lowlands (hosting pwr) lie within the old-AMSA province. 5) The Beta-Atla-Themis (BAT) and southern Lada volcanic provinces, regions of documented tectonic activity and constructional volcanism (Head et al., 1992; Crumpler et al., 1997), lie in the young AMSA province.

5. Volcanic rises

Venus' volcanic rises—each marked by broad domical regions (~1300–2300 km diameter; ~1–3 km high), local radial volcanic flows, and deep apparent depths of compensation (ADC; interpreted as evidence of thermal support within the mantle)—are widely accepted as contemporary surface expressions of deep mantle plumes on thick (i.e., ~100 km) Venusian lithosphere (e.g., Phillips et al., 1981, 1991; McGill, 1994; Phillips and Hansen, 1994; Smrekar et al., 1997; Nimmo and McKenzie, 1998; Stofan and Smrekar, 2005). A plume interpretation for volcanic rises is notable in its lack of disagreement and general consensus across a scientific community otherwise rich in diverse opinions. The consensus no doubt results, at least in part, from the different data sets that contribute to a plume interpretation, including both geological and geophysical relations such as: size, topographic form, gravity–topography ratios, deep ADCs, and wide spread evidence of volcanic flows across their surfaces.

Volcanic rises are divisible into three groups based on surface morphology: rift-dominated (Beta, Atla), volcano dominated (Dione, Western Eistla, Bell, Imdr), and coronae-dominated (Central Eistla, Eastern Eistla, Themis) (Stofan et al., 1995; Smrekar et al., 1997; Smrekar and Stofan, 1999). Rift-dominated rises show the highest swells (2.5 and 2.1 km) whereas coronae-dominated rises generally show the lowest swells (1–1.5 km). Rift- and volcano-dominate rises show the highest ADCs ~175–225 km and 125–260 km, respectively, as compared to coronae-dominated rises with lower ADCs (65–120 km) (Schubert et al., 1994; Stofan et al., 1995; Simons et al., 1997). Smrekar et al. (1997) provide an excellent overview of Venus' volcanic rises. General characteristics are summarized here.

Atla (9.2/200E; Fig. 2) and Beta (25/284E), marked by triple junction-like geometry rift zones display extensive associated volcanic flows, which overlapped in time (Senske et al., 1992). Atla and Beta also host the largest volcanic edifices on Venus, including large shield volcanoes and intermediate volcanic edifices. Beta Regio is underlain by a thermal anomaly in the mantle, indicative of a contemporary mantle plume (Kiefer and Hager, 1991; Simons et al., 1997; Kiefer and Peterson, 2003), as is Atla (Smrekar, 1994).

Phoebe Regio (−6/283E), commonly considered crustal plateau-like as a result of its tectonic fabric, is here considered more akin to volcanic rises. Let me explain. Phoebe, which lies between rises Beta and Themis, is similar in size to rises and plateaus; however, it topographically forms a rise rather than a plateau. Its tectonic fabric, marked by extensional faults and fractures with a wide range of orientations is unique, and structurally unlike the tectonic fabric that characterize all crustal plateaus (Hansen and Willis, 1996, 1998). Additionally, Phoebe's geophysical signature is hybrid

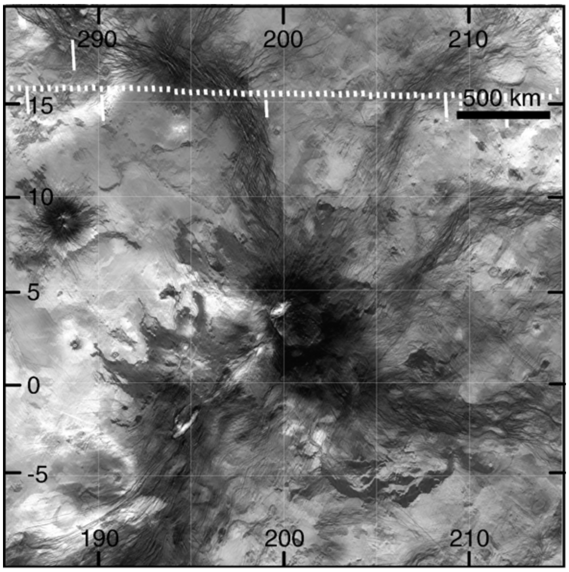

Fig. 2. Inverted SAR image of Atla Regio; radar-smooth regions are bight, radar-rough regions are dark, in contrast to normal SAR images. Inverted images show primary and secondary structures in better detail. Five wide fracture/rift zones come together and formed synchronously with two large shield volcanoes marked by extensive lobate flows. A small double peaked volcano is located in the adjacent lowlands at ~11°N along the western margin of the image.

between plateaus and rises (Grimm, 1994b; Simons et al., 1997), and recent analysis suggests that a mantle thermal anomaly underlies Phoebe (Kiefer and Peterson, 2003). Phoebe also lies along the intersection of two chasmata, or rift/fracture zones, and hosts a large shield volcano marked by radial fractures and extensive radial lava flows that extend over 1000 km. Fracture zones and extensive flows clearly post-date the formation of Phoebe's distinctive tectonic fabric. Thus, Phoebe might be more akin to Atla and Beta, as a rift-dominated volcanic rise.

Beta, like Phoebe, displays extensive tracts of basal tessera terrain (Senske et al., 1992). Beta tessera terrain is interpreted as indicative of thickened crust (Ivanov and Head, 1996). The tessera terrain fabric at Beta has not been mapped in detail and it is unclear if it is more similar to Phoebe's unique fabric, or to that of crustal plateaus, or if it represents a unique tectonic style. Similar to Phoebe however, geologic relations clearly indicate that the tessera terrain fabric pre-dates the formation of rift structures and volcanic flows, and as such, tessera terrain and rift-volcanism are likely genetically unrelated.

Venus' volcano-dominated rises are also home to large volcanic edifices, though they lack extensive rift/fracture zones. Volcanism at Bell (33/51.5E) evolved from low relief volcanic centers to steep-sided edifices (Campbell and Roger, 1994; Campbell and Rogers, 2002). Dione (−31.5/328E)—marked by a subtle topographic swell (1000 km, ~0.5 km high), three large volcanoes, and limited rift/fracture zones—might represent a decaying plume (Stofan et al., 1995; Smrekar et al. 1997). Imdr (−43/212E) displays the most modest surface structure (one major volcanic edifice, and a minor rift structure) of any volcanic rise, matching its modest topographic signature. Western Eistla hosts two large shield volcanoes cut by a zone of minor rifting interpreted as coeval with volcanic activity (Senske et al., 1992); analysis of Pioneer gravity data suggests that an active plume underlies the region (Grimm and Phillips, 1992). Laufey Regio (7/315E), a 0.5 km high, elongate topographic rise (1000 by 2000 km), marks the newest addition to the volcano-dominated rise family (Brian et al., 2004). Two large volcanoes and numerous coronae with associated flow deposits dominate the rise, marked by modest topography and a relatively shallow ADC (32–74 km). Laufey is interpreted as a late-stage rise, underlain by a waning plume (Brian et al., 2004).

Coronae-dominated rises, as their name implies, host numerous larger than average coronae (Smrekar and Stofan, 1999). Eastern Eistla includes five coronae (diameters 300–600 km). Central Eistla, which hosts a shield volcano and three coronae (McGill, 1994), overlies a postulated waning mantle plume (Grimm and Phillips, 1992). Themis Regio (−37/284) hosts eight coronae (diameters 300–675 km); geophysical (Johnson and Richards, 2003) and geological (Smrekar and Stofan, 1999) evidence indicate that coronae formed broadly contemporaneously with volcanic rises, and as such are likely genetically related. The size and interpreted deep thermal mantle anomalies (although not unique) might indicate that the rises result from mantle plumes derived from the core-mantle boundary, whereas the origin of the coronae is less clear. These coronae could represent: thermal break up of a large plume head (Griffiths and Campbell, 1990; McGill, 1994; Stofan et al., 1995), compositional diapirs spawned by a deep thermal anomaly (Hansen, 2003), focused transient small-scale thermals focused by large deep plumes (Johnson and Richards, 2003), relatively shallowly derived thermals that cluster together to form a local rise (Smrekar and Stofan, 1999), or of course, something else entirely. The challenge, and no doubt the wide range of proposed origins, results from the inability to uniquely constrain the mode of buoyancy, particular over the scale ranges involved in coronae-dominated rises, and in an environment in which both thermal and compositional buoyancy likely contribute. It is possible too that coronae-dominated volcanic rises form on relatively thin regional lithosphere, perhaps representing plumes associated with a cylindrical mantle upwelling.

As a group, volcanic rises represent broadly uplifted topographic domes; they display clear evidence of extensive associated magmatism, local rifting or fracture zones, and large ADCs. All of these characteristics taken together are consistent with the interpretation that they represent surface manifestations of contemporary mantle plumes. These features cannot be associated with plate tectonics, nor is there any evidence that they represent thermal insulation beneath large-tracts of continental-crust, which do not exist on Venus. Although the characteristics do not provide indisputable proof of mantle plumes, equally clearly a plume interpretation is certainly reasonable. Although Hamilton (2005) calls for every circular and/or volcanic feature on Venus to represent a bolide impact, he does not outline any testable hypothesis for how volcanic rises and the characteristics outlined herein (and in much more detail in the literature) could be accommodated within the context of such a hypothesis. Additionally, there is no evidence for or against a heterogeneous mantle on Venus (e.g., Meibom and Anderson, 2004; Anderson and Natland, 2005), and as such it seems most prudent to consider a homogeneous mantle as a first order assumption until the specific nature of proposed mantle heterogeneity is outlined within the context of a specific hypothesis or data set. Fundamentally, calling for the formation of volcanic rises by widespread melting of a heterogeneous mantle (and the inherent assumption of a heterogeneous mantle) seem equally untestable (if not more so), than a plume hypothesis. Perhaps the most interesting observation is that volcanic rises reside between 45N and 45S latitudes—reminiscent of Courtillot et al. (2003).

6. Large coronae

Venus hosts ∼515 quasi-circular tectonomagmatic features called coronae (Stofan et al., 1992, 1997, 2001). Coronae dominantly form chains, but also occur as clusters associated with volcanic rises, and, most rarely, as isolated lowland features. Coronae, meaning crown, was initially a descriptive term, but it has evolved into a term which commonly carries genetic connotations. Coronae are widely accepted as representing the surface manifestation of mantle diapirs (e.g., Squyres et al., 1992; Janes et al., 1992; Stofan et al., 1992, 1997; Koch and Manga, 1996). Features collectively referred to as coronae show a wide range of characteristics however, and might represent more than one group of genetically unrelated features. This topic is outside the limits of the current contribution. Coronae range in size from 60–2600 km, but are strongly bias toward the lower values with a median diameter of ∼220 km. Artemis 'Corona' (2600 km) is an order of magnitude larger than median coronae, and over twice the diameter of the next largest corona, Heng-O (1060 km). Quetzalpetlatal (780 km), and Atahensik (∼700 km; 'Latona' in early publications) complete the family of large coronae. Although a few other large coronae are noted in some data bases with diameters ≥600 km (Stofan et al. 1992, 2001), these coronae are: a) highly non-circular and thus their stated diameter is not a true diameter but rather a long axis (e.g., Ninmah Corona), b) are double coronae with two foci, or c) are likely not coronae but instead represent several genetically unrelated features (e.g., 850 km diameter Zisa 'corona'). The discussion here focuses on Venus' four large coronae (Fig. 3). These large coronae share some characteristics, yet each also defies clear classification with other features, or one another. Each are described briefly below in turn.

Artemis (2600 km; −35/135E; $\sim 4*10^6$ km^2; Fig. 3a) comprises a large topographic welt that includes a paired circular trough and outer rise. Artemis defies

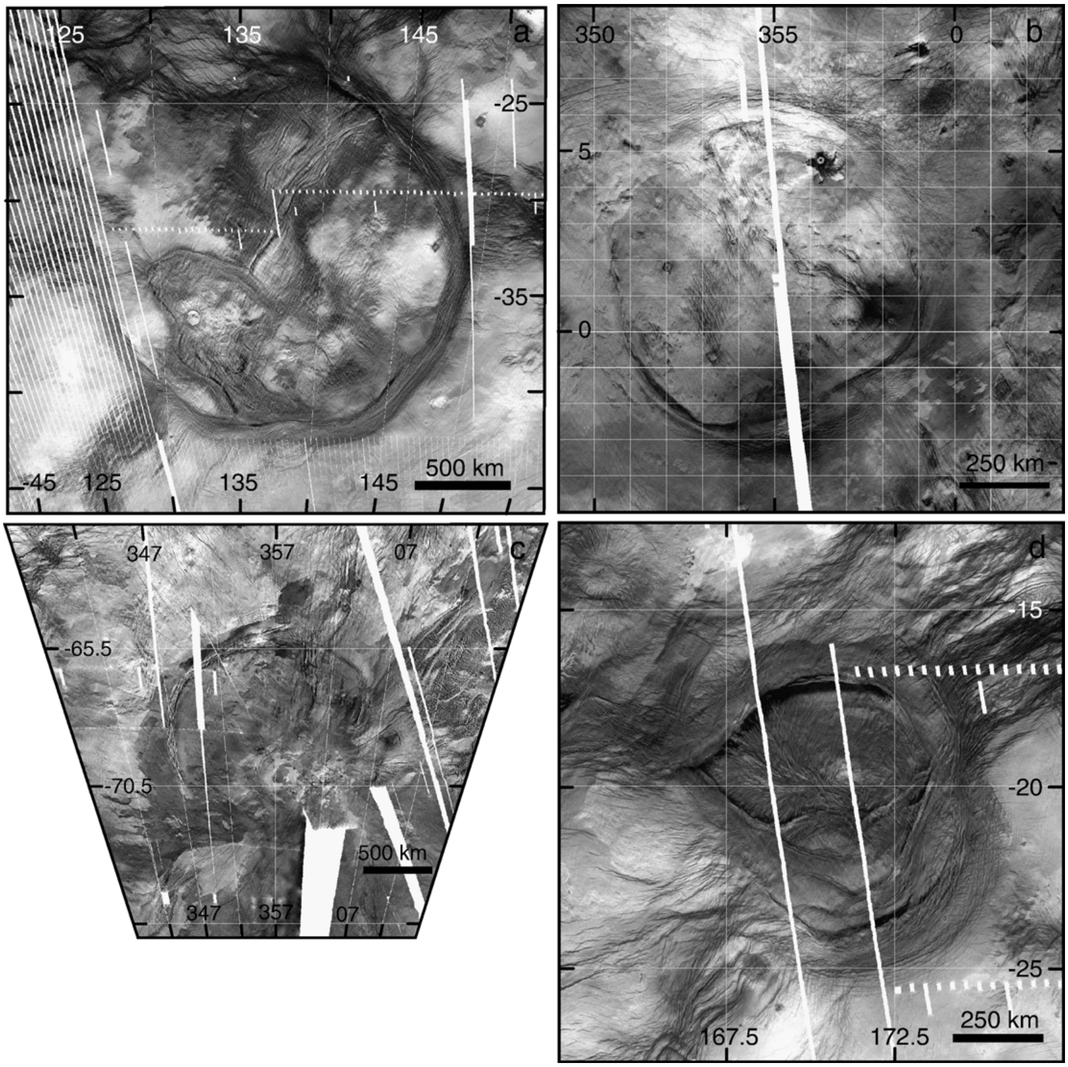

Fig. 3. Inverted SAR images of large coronae: Artemis (a), Heng-O (b), Quetzalpetlatl (c), and Atahensik (d). Note similarities and differences of each feature. See text for discussion. White regions indicate data gaps in the SAR imagery.

geomorphic classification; it is similar in size to plateaus and rises, yet topographically more akin to corona. The ~2200 km diameter trough (Artemis Chasma) is 150–200 km wide and ~1–1.5 km-deep. The outer rise (200 km wide) parallels the trough, completing the structure. Artemis' trough does not describe a complete circle, extending from ~12:00 to 10:30 in an analog clock framework; the trough gradually disappears on either end, both topographically and structurally. Artemis has been classified as a corona (Stofan et al., 1992), but given its size I simply consider it 'Artemis'. Brown and Grimm (1996) mapped the trough and rise structures in detail, and Hansen (2002) mapped the interior, trough, and adjacent region. (Geologic mapping as part of the NASA-USGS Venus mapping program is underway (Bannister and Hansen, 2005). Geologic mapping by both groups shows broad agreement of structural elements, though respective interpretations vary. The interior, which sits 2–4 km above the adjacent lowlands, hosts four coronae—three marked by radial fractures and flows. A fifth corona, marked by radial fractures but lacking obvious flows, overlaps with the southern trough. A penetratively developed, 1–2 km wavelength, fabric occurs across much of Artemis, trending generally northeast, but taking on a radial character near three interior coronae. The trough hosts trough-parallel normal faults and folds; faults occur along the inner trough slope, whereas folds dominate the outer trough slope. Radial extension fractures and trough-concentric wrinkle ridges dominate the rise, with concentric wrinkle ridges continuing outward for hundreds of kilometers.

Four different hypotheses have been proposed for the formation of Artemis: two related to a subduction interpretation, a plume hypothesis, and as Venus' largest impact structure. Spencer (2001) interpreted a part of Artemis's interior as a region of major crustal extension similar to a terrestrial metamorphic core complex; although, he did not place the study within a regional context Spencer inferred that the extension related to regional subduction. Brown and Grimm (1995, 1996) proposed that Artemis Chasma resulted from northwest-directed subduction beneath Artemis' interior. These workers suggested that Artemis Chasma represents two distinct trough segments. The trough from ~2:30 to 6:30 represents a subduction zone marked by ~250 km of under-thrusting of lowlands to the southeast under Artemis' interior, and from 12:00 to 2:30 represents an associated trough dominated by left-lateral displacement. The trough from 6:30–10:30 represented an older feature, genetically unrelated to the proposed subduction zone. The subduction interpretation stems from the topographic asymmetry from the outer high, across the trough and into the interior, similar in profile to terrestrial subduction zones (McKenzie et al., 1992; Schubert and Sandwell, 1995). Artemis has an ADC of ~200 km, which has been interpreted as evidence of a subducted slab (Schubert et al., 1994; Brown and Grimm, 1995), but it might also represent under-plated material or melt residuum within a plume context. Gravity analysis is fraught with assumptions including assumed single depths of compensation, and can result in incredibly non-unique interpretations.

Hansen (2002) outlines several problems with a subduction zone interpretation. The most compelling arguments against subduction are continuity of structures along the entire trough in a trough parallel fashion from 12:00 to 10:30, and the shared central location of trough topography, trough structures, radial fractures and wrinkle ridges. These observations taken together argue for Artemis to have formed by a single coherent process, rather than unrelated geologic events. A subduction interpretation would also require left-lateral displacement within the trough along the northeast margin (12:00–2:30) and right-lateral displacement within the trough along the southwestern margin (7:30–10:00). (Hansen (2002) argued against the evidence presented for left-lateral strike-slip displacement along the northeastern margin (Brown and Grimm (1996)). Furthermore, volcanic flows, associated with one of the interior corona, cross the trough along the southwestern margin, yet these flows are also involved in trough structure deformation. These relations indicate that at least some interior magmatism and trough formation overlapped in time, and provide clear evidence against the strike-slip displacement required by the subduction hypothesis. Additionally, it is unclear why strike-slip displacement should result in the formation of a trough along the northeast or southwestern–western margins of Artemis.

Hamilton (2005) asserts that Artemis records the impact of a huge bolide ~3.5–4.0 Ga. Unfortunately the 'hypothesis' lacks any details, or even the most general qualifying statements or predictions; as such it is not testable. Additionally, Artemis' topographic form is opposite to what might be expected in the case of a large bolide impact. Large impact basins generally show multiple ring morphology, features Artemis clearly lacks. Contrary to the assertion by Hamilton (2005), there is no evidence that a northwest margin of Artemis (19:30–12:00) is buried beneath other constructs; yet a large impact basin would be expect to show a complete circular structure. As noted by Brown and Grimm (1996) and Hansen (2002) a northwestern margin of Artemis likely never existed. Perhaps a carefully constructed impact hypothesis for Artemis formation will emerge; such a hypothesis must address concurrent formation of interior coronae and trough, the topographic trough, rather than classic rim, and the lack of concentric ring structures, and it must account for the formation of the geologic features on published geologic maps including extremely organized folds and normal faults within the trough.

Currently the most viable hypothesis seems to be the surface manifestation of a mantle plume on thin lithosphere, consistent with its large size and circular planform. Gravity–topography analysis is consistent with at least partial dynamical support for Artemis (Simons et al., 1997). As a deep mantle plume rises toward a lithosphere, the lithosphere will be uplifted, and, if the strength of the lithosphere is exceeded, radial fractures could form above the plume head, or if the lithosphere were sufficient heated it could develop a penetrative tectonic fabric. A circular trough could also form, as illustrated in laboratory experiments aimed at modeling the interaction of thermal plumes with the lithosphere (Griffiths and Campbell, 1991). In these experiments, as a plume head approaches a rigid horizontal boundary, it collapses and spreads laterally; a layer of surrounding mantle, squeezed between the plume and the surface could result in a gravitationally trapped asymmetric instability, and lead to formation of an axisymmetric trough. In addition, the interior squeeze layer can lead to convection on a scale much smaller than that of the original plume. These smaller-scale instabilities could interact with the lithosphere inside the axisymmetric trough, and be manifested as interior and trough corona. It was on the basis of these experiments that a plume model for Artemis formation was proposed following release of Magellan SAR data (Griffiths and Campbell, 1991). Finite-element models of the interaction of a large thermal plume with lithosphere, aimed at modeling corona topography, also show development of an axisymmetric trough above large thermal mantle plumes (Smrekar and Stofan, 1997). The size feature modeled by Smrekar and Stofan (~1200 km diameter trough) is perhaps better applied to Artemis than to median-sized coronae an order of magnitude smaller. Smrekar and Stofan's (1997) modeling predicts lithosphere delamination. Delamination might be a way to develop a sort of hybrid model that incorporates aspects of plume–lithosphere interactions with signatures that some workers propose might be better addressed through subduction. Fundamentally it seems that Artemis' formation requires, at some first order level, the interaction of a deep mantle plume and relatively thin lithosphere. Perhaps future detailed mapping coupled with iterative modeling aimed at testing detailed hypotheses would be fruitful. Clearly current plume–lithosphere models are too simple for nature. Artemis might be akin to coronae-dominate volcanic rises, though formed on thinner lithosphere which allowed for trough formation as a result of plume–lithosphere interaction and/or delamination processes.

Heng-O (1060 km; 6.6/355E; $\sim 1*10^6$ km^2; Fig. 3b) lies within the confluence of several planitiae, and forms a low circular rise surrounded by a rim (~100 km wide, ~1–1.5 km high) that rises above the surrounding terrain, both interior and exterior. The rim, which is not equally developed all the way around being subdued along the eastern and northwestern margins, is variably decorated with rim-parallel ridges. It is locally paralleled, or interrupted, by rim-parallel lows locally forming a moat up to 400 m deep. A northwest-trending suite of fractures and shield fields decorate Heng-O's relatively featureless interior, in strong contrast with that of Artemis. Heng-O has been variably interpreted as the surface expression of a deep-mantle plume (Ivanov and Head, 2001) and an impact feature (Hamilton, 2005; Vita-Finzi et al., 2005). To date no published studies include geologic mapping and analysis, hence both endogenic and exogenic hypotheses seem equally viable. Harris et al. (2002) noted that a giant radiating graben system converges on Heng-O; the swarm apparently fans outward to the northwest and southeast, extending over 3000 km. The radial dike system is not obvious in Magellan imagery, likely due to post-radial graben surface activity. Future geologic mapping will be required to determine if the radial graben system is genetically related to the circular rim feature, and to constrain the mechanism of Heng-O formation. Although surface flows clearly postdate the rim structures and associated moat, it is not clear if Heng-O formation involved extensive magmatic activity, and as such, it is unclear whether Heng-O should be considered a LIP, formed by whatever mechanism. Heng-O shows an ADC of ~150 km (Schubert et al., 1994), but interpretation of ADC is non-unique, and certainly should not be interpreted outside the context of geologic mapping.

Quetzalpetlatal (780 km; −68/357E; $0.5*10^6$ km^2; Fig. 3c), the most southern feature discussed herein, is marked by an ~180° topographic arc-shaped scarp. Geologic mapping is currently in progress as part of the NASA-USGS Venus mapping program (Ivanov and Head, 2001, 2005). Quetzalpetlatal dominates Lada Terra, a region marked by apparently young volcanic feature corresponding to a young AMSA province (Fig. 1). The 180° arcuate scarp (~1 km high) that marks Quetzalpetlatal exists from ~8:00 to 2:00. An ~100–150 km wide 200–300 deep moat parallels the scarp. A dense swarm of ridges, ~50 km wide, parallel the scarp for about 850 km. The southern half of Quetzalpetlatal lacks topographic expression, either never having formed, or bury by subsequent surface flows. Massive volcanic flows emanate from Boala Corona, which resides within Quetzalpetlatal's plateau-shaped interior; flows extend radially covering much of the interior and spilling across the scarp, into the outer moat, and locally outward to the surrounding lowlands. Quetzalpetlatal lies within an ~1700 km diameter dome-shaped high and the flows locally extend to the limits of the topographic dome. Flow fronts are lobate and extremely well preserved providing clear evidence of their interior source. Quetzalpetlatal clearly classifies as a LIP. Quetzalpetlatal provides the source region for much of Lada Terra, and therefore is likely quite young. Genetic relations between the arcuate scarp, scarp structures, moat and extensive flows are currently unconstrained, but should emerge from geologic mapping in progress. Based on preliminary mapping Ivanov and Head (2001) suggest that Quetzalpetlatal represents the surface signature of a large, deeply rooted mantle plume. Hamilton (2005) asserts that Quetzalpetlatal represents a large impact structure. But, as is the case of his other suggestions, Hamilton presents very little data to support his proposal. According to Hamilton (2005) an impact origin is the only viable means to form circular or quasi-circular geologic structures.

Atahensik (~700 km; −19/170E; $0.4*10^6$ km^2; Fig. 3d) lies along a chain of coronae, in contrast to the other large coronae. Atahensik also displays the largest range in topography and the steepest slopes—some of the largest slopes on Venus. Although Atahensik is typically noted as ~700 km, its topographic signature is ~1000 km in the N–S direction, and associated flows extend across a 1200–1400 km diameter region. Atahenski, although quasi-circular in form, is decidedly not circular (Fig. 3d). Deep topographic troughs (chasmata) bound Atahensik describing a sort of eye-shaped feature. The northern trough, which is concave to the south, spans ~75 km, and displays topographic asymmetry with the northern trough wall plunging ~2 km below mean planetary radius and the south (north-facing) slope of the trough reaching up to 5 km above mean planetary radius; thus the trough has a steep north-facing slope that drops ~7–8 km. Three nested, sub-parallel, troughs that are concave to the north, mark Atahensik's southern margin. Each of these troughs shows similar asymmetric topography with steep inner (south-facing) slopes and more gentle outer (north-facing) slopes. Atahensik' highest elevations lie along the inner scarps; the interior region resides ~3 km above mean planetary radius.

Geologic maps provide clues to Atahensik's character and formation (Hansen and Phillips, 1993; Hansen and DeShon, 2002). Atahensik displays an incredibly penetrative (i.e. closely spaced) delicate tectonic fabric that radiates outward from a generally centralized region. The lineaments, representing fractures and graben, can be traced from the interior across the topographic troughs to the surrounding region. Atahensik's penetrative fabric is similar to that of Artemis, although in the case of Atahensik the fabric displays a more singular radial pattern, and many display a slightly wider spacing. Flows follow the penetrative fabric, locally filling lows and burying the fabric. Troughs cut the fabric and flows indicating relatively late trough formation. Local trough deposits likely represent slope collapse rather than magmatic flows. Flows extend as much as 750 km outward from the center of Atahensik, and cover an area over 1.4 million km^2; the flows are extensive, but they are likely <1 km thick.

Atahensik's steep asymmetric troughs were initially interpreted as evidence of subduction, and hence plate tectonics, on Venus (e.g., McKenzie et al., 1992; Schubert and Sandwell, 1995), but the lack of mismatch of the radial pattern in the penetrative fabric provides clear evidence against a subduction (Hansen and Phillips, 1993). Atahensik is one of several coronae in the chain extending to the southwest from Atla. Although Atahensik is the largest corona in the chain, its geologic character and history are similar to other chain coronae: radial lineaments, concentric structures, topographically elevated interiors, associated topographic troughs, extensive flows, and rich tectonomagmatic histories (e.g., Hamilton and Stofan, 1996; Stofan et al., 1997; Hansen and DeShon, 2002). The geologic history that emerges from mapping of Atahensik is similar to that of many coronae, and in particular coronae that host suites of radial lineaments that extend beyond their topographic forms.

Detailed discussion of coronae, a rich field of investigation, is outside the limits of the current contribution. Coronae studies include global volcanic, tectonic, and topographic classifications, regional mapping, gravity signature investigations, and evolution models. The general consensus is that many coronae result from buoyant mantle material impinging on the lithosphere (see Stofan et al., 1997). Coronae could result from compositional diapirs, thermal diapirs, or a combination, and also certainly coronae have contributed significantly to heat loss on Venus (Smrekar and Stofan, 1997; Hansen, 2003; Johnson and Richards, 2003). Indeed, Atahensik represents the surface manifestation of a diapir. The mere size of Atahensik favors a thermal diapir (e.g., Hansen, 2003), although large compositional diapirs could result from the coalescing of small compositional diapirs (Kelly and Bercovici, 1997; Manga, 1997), and therefore Atahensik could result from the coalescing of two or more diapirs at depth, in keeping with its location along an entire chain of similar features.

Although many workers accept that coronae represent the surface expression of endogenic diapirs that impinged on the crust or lithosphere other hypotheses have been proposed, including formation by exogenic impactors (e.g., Schultz, 1993; Hamilton, 2005; Vita-Finzi et al., 2005; McDaniel and Hansen, 2005). It is important to note that given the wide range of morphological features included as coronae, it is possible, and perhaps most likely, that as a group, not all coronae formed by the same mechanism. It is equally important to note that given the occurrence of Atahensik along a localized chain of similar style features, and the rich geologic history recorded, an impact mechanism for the formation of Atahensik is extremely difficult to justify. Hamilton (2005) suggests that all circular features on Venus (over 5000 in number) represent impact craters. Other than Atahensik's quasi-circular planform, there is nothing about this feature that indicates an impact origin.

In summary, Atahensik represents one of Venus' largest coronae features clearly akin to many (but likely not all) features described as coronae. It occurs within a chain of similar features and geologic relations support a diapiric origin, although the nature of the diapir, whether compositionally, or thermally driven (or both), and the source of the diapir(s) (deep, middle, or shallow mantle) is unconstrained.

Brief examination of Venus' four largest coronae suggests that they likely form by a range of mechanisms. Artemis most likely represents the surface manifestation of a large, and hence likely deep, thermal diapir or plume on thin lithosphere. Heng-O's origin will require further geologic mapping, but viable hypotheses include both endogenic and exogenic processes: plume and bolide impact. It is also possible that Heng-O does not record an extensive magmatic history, and might not represent a LIP. Testing hypotheses of Quetzalpetlatal's formation will also depend on the result of ongoing geologic mapping, although an endogenic, and specifically plume, origin is currently favored (Ivanov and Head, 2001). Geologic relations are most consistent with a diapiric origin for Atahensik, though diapiric details are unconstrained. Any hypothesis that calls for formation of Artemis, Heng-O, Quetzalpetlatal, and Atahensik by bolide impact must account not only for the detailed geologic patterns and histories of each feature, but also address their marked differences.

7. Crustal plateaus

Crustal plateaus (Fig. 4) are similar to volcanic rises in term of planform shape and size, but they differ in topographic signature and geologic evolution. Crustal plateaus lack rifts, large volcanoes, and coronae. Crustal plateaus host distinctive deformation fabrics (Fig. 5), herein called ribbon-tessera terrain following terminology of Hansen and Willis (1998). Scientists generally agree that thickened crust supports crustal plateaus, as evidenced by small gravity anomalies, low gravity to topography ratios, shallow ADCs, and consistent admittance spectra (see citations in Phillips and Hansen (1994) and Hansen et al. (1997)). Spatial correlation of plateau topography and tectonic fabrics strongly suggests that the thickening (uplift) mechanism and surface deformation are genetically related (Bindschadler et al., 1992a,b; Bindschadler, 1995; Hansen and Willis, 1996; Ghent and Hansen, 1999). Researchers also widely accept that arcuate-shaped inliers of characteristic ribbon-tessera terrain that outcrop across expanses of Venus' lowland represent remnants of collapsed crustal plateaus (e.g., Bindschadler et al., 1992b, Phillips and Hansen, 1994; Bindschadler, 1995; Ivanov and Head, 1996; Hansen et al., 1997; Hansen and Willis, 1998; Ghent and Tibuleac, 2000).

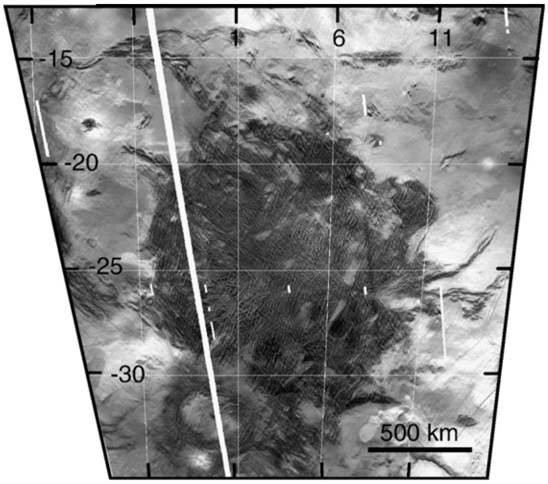

Fig. 4. Inverted SAR image of crustal plateau Alpha Regio, a typical crustal plateau with distinctive radar-rough terrain residing in a elevated plateau above the adjacent radar-smooth lowlands; the circular feature that overlaps Alpha along its southwest margin is a younger corona.

Two basic questions emerge with respect to plateau formation. 1. How were plateau surfaces deformed and concurrently uplifted? 2. How did plateaus collapse? Initially two end-member hypotheses emerged in response to the first question—the downwelling and plume hypotheses. The downwelling hypothesis involves concurrent crustal thickening and surface deformation due to subsolidus flow and horizontal lithospheric accretion associated with a cold mantle diapir beneath ancient thin lithosphere (e.g., Bindschadler et al., 1990, 1992a,b; Bindschadler, 1995). Following this hypothesis, crustal plateaus would not be considered LIPs by any definition. The plume hypothesis accommodates thickening and deformation via magmatic under-plating and vertical accretion due to interaction of a large deep-rooted mantle plume with ancient thin lithosphere (Hansen et al., 1997; Hansen and Willis, 1998; Phillips and Hansen, 1998; Hansen et al., 1999, 2000). Following the plume hypothesis, crustal plateaus would represent LIPs.

Though calling on quite different mechanisms of plateau formation, both hypotheses call for time-transgressive deformation of ancient thin lithosphere above individual spatially localized regions, that of each individual plateau. In addition, both hypotheses embrace the premise that a root of thickened crust supports each plateau and plateau collapse results from lower crustal flow. Recently published finite element modeling illustrates, however, that the range of preserved crustal plateau morphologies and arcuate ribbon-tessera terrain inliers is difficult to achieve

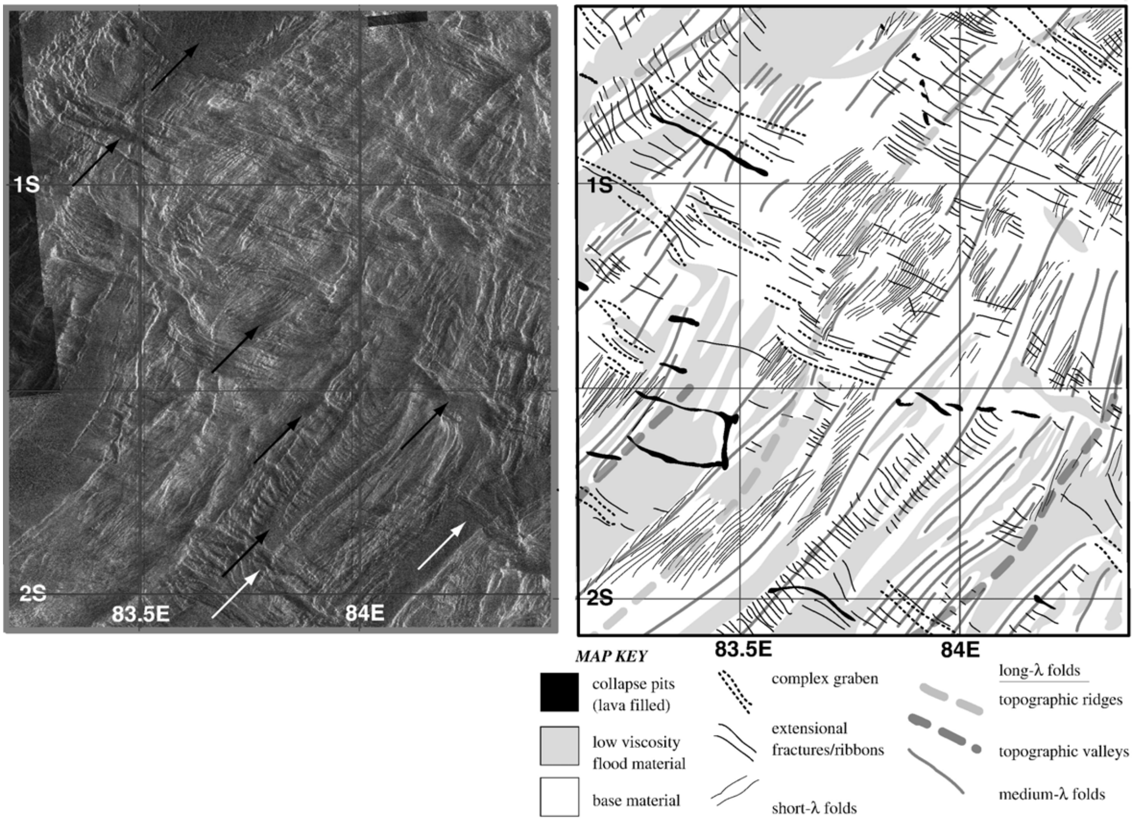

Fig. 5. Normal SAR image (radar-smooth areas are dark; radar-rough areas are bright) and geologic map of ribbon tessera terrain, eastern Ovda Regio (from Hansen, 2006). The area encompasses the crest (~6055 km) and trough (~6054.2 km) of a prominent fold. Medium- and short-wavelength folds trend parallel to the long-wavelength fold crest. Orthogonal ribbon fabrics locally disrupt short- to medium-wavelength folds. Ribbon-parallel complex graben cut long-wavelength folds. All folds trend broadly parallel to one another, whereas extensional structures (ribbons and complex graben) parallel one another, and consistently trend orthogonal to fold trends. Locally, fold troughs of both short- and medium-wavelength folds are radar smooth, presumably the result of local flooding by low-backscatter, low viscosity material. NW-trending extensional structures post-date the formation of the shortest-wavelength folds, locally cutting fold crests and troughs, including locally flooded troughs near the center of the image (black arrow). In the southeast corner of the image, periodically spaced ribbon troughs are locally filled by low-backscatter flood material (black arrows), yet other extensional structures also cut trough fill (white arrows) indicating that flooding locally both predated and post-dated formation of various extensional structures.

through lower crustal flow at geologically reasonable time scales (Nunes et al., 2004). Thus, although both the downwelling and the plume hypotheses call for lower crustal flow to accommodate the range in plateau and inlier topography, collapse by lower crust flow might not be viable, or it may be that the parameters employed by the collapse model are not appropriate. Thus, the range in elevations of individual plateaus and arcuate inliers of ribbon tessera terrain are not addressed by either hypothesis.

Additionally, neither the downwelling nor the plume hypotheses address all characteristics of crustal plateaus, and each hypothesis carries specific burdens. Challenges for the downwelling hypothesis include: a) a predicted domical rather than the observed plateau shape; b) lower crustal flow called upon for crustal thickening requires 1–4 billion years, well outside the absolute time constraints allowed by geologic relations (Kidder and Phillips, 1996); and c) formation of documented short-wavelength extensional structures (ribbon fabrics) requires a high geothermal gradient which is difficult to justify in the cold downwelling environment (Hansen et al., 1999). The plume hypothesis can accommodate formation of a plateau shape and extensional features in a viable timeframe; however, extensive contractional strain, or the formation of short-wavelength folds is difficult to accommodate within the plume hypothesis (Ghent et al., 2005). Although the plume hypothesis addresses formation of late long-wavelength folds (or warps), which record very little

(<1%) shortening, early layer shortening, and/or large amounts of layer shortening would present a serious challenge for the plume hypothesis. In addition, Gilmore et al. (1998) argue that formation of ribbon fabrics requires a geothermal gradient much in excess of what might be achieved in a plume–lithosphere interaction environment.

Despite deep divides within the crustal plateau debate, geologic mapping on both sides leads to four, mutually agreed upon, observations: 1) plateaus host both contractional structures (folds) and extensional structures (ribbons, extensional troughs, graben), which are generally mutually orthogonal; 2) multiple suites of folds, defined by wavelength, occur; 3) multiple suites of extensional structures, defined by spacing, occur; and 4) low viscosity fluid, presumably lava, fills local to regional topographic lows (Fig. 5). Despite these agreements, controversy exists as to the timing of flooding relative to deformation. In addition, although limits exist for the amount of extensional strain, the amount of contractional strain, particularly that represented by short-wavelength features has been, until recent, unconstrained.

The flooding controversy can be summarized as: a) did flooding generally post-date deformation, and as such was likely genetically unrelated to crustal plateau formation (e.g., Gilmore et al., 1997, 1998; Basilevsky and Head, 1998); or b) did flooding accompany surface deformation, and was therefore genetically related to crustal plateau formation (e.g., Banks and Hansen, 2000)?

Recent detailed geologic mapping aimed at addressing these questions yielded new observations and refined geologic histories for plateau surfaces (Hansen, 2006; Nordberg et al., 2005), resulting in the proposal of a third hypothesis for plateau formation—the lava-pond hypothesis. Geologic relations observed at three crustal plateaus (eastern Ovda, Alpha, and Tellus) call for progressive deformation of an initially very thin layer (10s to 100 m) developed across individual plateau surfaces. The layer shortened forming ductile folds and extended in an orthogonal direction along brittle structures. With additional shortening earlier formed, short-wavelength structures were carried piggy-back on younger, progressively longer-wavelength folds. Local flooding accompanied progressive deformation of the increasingly thicker surface layer. Low viscosity flood material leaked from below into local structural lows. Structural lows, which were flooded relatively early, were also carried piggy-back on younger longer-wavelength structures, and are preserved atop long-wavelength structural ridges. Subsurface liquid (magma) formed the required sharp decrease in viscosity with depth, and served as the source of the flood material. The unique tectonic fabric and extensive local flooding, representing widespread magmatic activity that represents the ‘scum’ of a huge large pond.

The lava-pond hypothesis calls for progressive solidification and deformation of the surface of a lava pond with areal extent marked by individual crustal plateaus. Ribbon-tessera terrain, marked by orthogonal ductile fold and brittle ribbon fabrics, represents lava pond ‘scum’ developed through progressive deformation and solidification. Individual lava ponds, each representing a crustal plateau in the making, resulted from massive partial melting in the shallow mantle; melt rose to the surface to form a lava pond, leaving behind a lens of low-density mantle residuum (e.g., Jordan, 1975, 1978). Isostatic adjustment in the mantle resulting from the residuum, rather than thickened crust, raised a solidified lava pond to plateau stature. Later, local mantle convection patterns could variably strip away the low-density residuum root, resulting in subsidence and/or ultimate collapse of an individual plateau. Remnants of distinctive ribbon-tessera terrain fabrics, formed during lava pond solidification, could survive as a record of an ancient lava pond. Thin surface deposits could partially or completely cover the distinctive pond scum fabrics, obscuring or erasing, respectively, the record of an ancient lava pond. The lava pond hypothesis addresses the detailed geologic surface history that characterizes crustal plateaus, as well as the formation and subsequent collapse of ancient crustal plateaus, marked by arcuate tracts of lowland ribbon-tessera terrain.

Massive partial melting within the shallow mantle could result from either: a) a large bolide impact with ancient thin lithosphere, or b) rise of an extremely hot deep mantle plume beneath ancient thin lithosphere (Hansen, 2006). In either case, crustal plateaus require thin lithosphere to form, and they owe their topographic stature to a low-density mantle residuum lens, rather than thickened crust. A bolide impact mechanism is favored because the formation of a lava-pond necessitates a large volume of magma at the surface. Balancing formation of massive melt, yet preserving a local lithosphere able to support a large lava pond seems a challenge to address within the context of a plume hypothesis. In contrast, a 20–30 km bolide would simply punch through the lithosphere into the mantle forming an ~200–300 km diameter ‘hole’ but the lithosphere across a several thousand kilometer scale could retain its strength.

Although some workers state that large bolide impact cannot generate huge volumes of melt (e.g., Ivanov and

Melosh, 2003), others present convincing counter arguments, particularly if a large bolide impacts thin lithosphere (Jones et al., 2005; Elkins-Tanton and Hager, 2005). Thin hot lithosphere, critical to formation of huge volumes of melt as outlined by Jones et al. (2005), might be easily accommodated on ancient Venus—widely accepted to have had a globally thin lithosphere (e.g., Solomon, 1993a; Grimm, 1994a; Solomatov and Moresi, 1996; Schubert et al., 1997; Phillips and Hansen, 1998; Brown and Grimm, 1999). According to Jones et al. (2005) for impact generated melting on Earth (using the case of the Ontong Java Plateau), melt would be distributed predominantly as a giant sub-horizontal disc with a diameter in excess of 1000 km down to >150 km depth in the upper mantle within ~10 min, although most of the initial melt is shallower than ~100 km. The total volume of mostly ultramafic melt, would be $\sim 3\times10^6$ km^3 ranging from superheated liquid (100% melt, >500° above solidus) within 100 km of ground zero, to varying degrees of non-equilibrium partial melt with depth and distance. These workers suggest that it would take tens of thousands of years for the melt to solidify on Earth, and it would presumably take even longer on Venus, given its dense CO_2 atmosphere. Venus' atmosphere, ~95 bars of super-critical CO_2, acts more like a conductive layer, than a convection layer in terms of heat transfer, and as such serves to insulate, rather than cool, the crust (Snyder, 2002). It may be that the preservation of the pond scum fabric (ribbon tessera terrain) is a function of Venus' atmosphere. As noted by Jones et al. (2005), bolide impact could also spawn a thermal anomaly in the mantle, which could contribute to the thermal evolution from below, while Venus' dense atmosphere contributed to a unique crystallization environment above. In either case, melt would presumably rise to the surface along fractures, and would spread out, presumably as a function of surface topography. The range in individual crustal plateau planform and the specific patterns of ribbon-tessera terrain, to be determined through future mapping, might provide clues to first-order drivers of melt distribution.

Although there is a relatively strong spatial correlation of tessera terrain fabric with crustal plateau topography (e.g., Bindschadler et al., 1992b; Phillips and Hansen, 1994; Hansen et al., 1997; Phillips and Hansen, 1998; Ghent and Hansen, 1999), detailed mapping indicates that in some places coherent tessera fabrics extend into adjacent lowlands, as is the case of Tellus and Ovda regiones (e.g., Senske, 1999; Hansen, 2006; in press). In the context of the proposed hypothesis, the lava pond—responsible for ribbon terrain fabric—forms on the surface, whereas the residuum—responsible for elevation—forms at depth in the shallow ductile mantle. Although we might expect broad correlation of surface and subsurface features, one can also envision situations in which the correlation would not be exact, and lava pond scum—or ribbon-tessera terrain—could be variably uplifted due to mismatch of surface lava pond and subsurface residuum. It is important to understand that the timeframe of lava pond formation and solidification would differ from isostatic adjustment due to mantle melt-residuum (the plateau-uplift event), similar to the time lag of isostatic rebound and the disappearance of continental ice sheets. Thus a lava pond would form at the surface, solidify, and be uplifted later, and at a very different rate that lava pond solidification.

Mead Crater, Venus' largest impact crater, at ~270 km diameter provides evidence of a bolide-impact of the postulated size (~27 km diameter) in Venus' more recent past. In the case of Mead Crater massive melting did not occur, presumably because the lithosphere was thick at the time of impact. Thus a most critical parameter for impact-induced melting, thin lithosphere, was not met during the formation of Mead Crater, despite the presence of a large bolide impact.

8. The solar system's largest LIP?

Research were taken by surprise when it became apparent that Venus' surface preserved the most pristine impact craters in the solar system, and that these craters, ~970 in number, were displayed in near random fashion surface (Phillips et al., 1992; Schaber et al., 1992; Herrick et al., 1997; Hauck et al., 1998). Clearly plate tectonic processes, which would result in large tracts of young and old crust, were absent. How then did Venus work? How did interior heat escape? The observations of near random distribution together with the observation that few craters showed partial burial lead to proposal of the catastrophic resurfacing hypothesis. According to this hypothesis, Venus experienced catastrophic resurfacing involving a massive outpouring of flood lava to a depth of ~2.5 km, covering ~80% of the surface in 10–100 million years time ~500 Ma (e.g., Strom et al., 1994; Herrick, 1994; Basilevsky and Head, 1996, 1998). This postulated event is heralded as the solar system's largest LIP (Head and Coffin, 1997; Ernst and Desnoyers, 2004).

Catastrophic resurfacing of Venus seemed necessary to account for the near random distribution of craters, and their 'pristine' character, generally lacking evidence of partial crater burial, rim breaching, or modification.

Equilibrium-resurfacing hypotheses (e.g., Phillips, 1993; Solomon, 1993a,b), which could address the global AMSA but could not accommodate the observation that very few craters are partially buried or destroyed, were mostly abandoned. In contrast, the catastrophic resurfacing hypothesis predicted the lack of partially buried craters due to the postulated catastrophic outpouring of 2.5 km thick (up to 10 km thick proposed initially) lava emplaced quickly (10 m.y. favored, but 100 m.y. statistically acceptable; Strom et al., 1994). Thus, few, or no, partially buried craters should be preserved. Catastrophic resurfacing of Venus became widely accepted in the planetary community and made it way into numerous textbooks.

Although perhaps initially compelling, the catastrophic resurfacing hypothesis does not stand up to geologic considerations, or to post-Magellan data analysis (Hansen and Young, in press). First, the magma flux required to produce a 2.5 km thick lava over $\sim$368 Mkm^2 (80% of Venus' surface) in 10 and 100 m.y. would require 2.5 and 25 times, respectively, the magmatic flux of Io (e.g., Johnson, 2004). Io, the most volcanically active body in the solar system, gains its energy from tidal heating due to Io's orbital dance with massive Jupiter and sister moons, Europa and Ganymede. Io's incredible energy source is renewed daily. But how might Venus, whose maximum energy levels, due to accretion and radioactive decay, occurred $\sim$4 Ga, generate 2.5–25 times the current magmatic flux of Io, $\sim$750 Ma? Rather than attempt to address this question, it seems prudent to revisit the relations that lead to postulated catastrophic resurfacing, and to re-examine predictions (requirements) of the catastrophic hypothesis in light of geologic relations that have emerged from analysis of Magellan data. Catastrophic resurfacing, was after all, proposed before all the Magellan data was even collected, much less studied to a detail consistent with its high resolution.

The catastrophic resurfacing hypothesis requires: 1) The same average model surface age (AMSA) across the planet, and 2) at least 2.5 km thick lava flows sitting on local to regional 'basal' terrain. Recent studies indicate that neither one of these criteria are met as discussed below.

Venus records a *global* AMSA of $\sim$750+350/−400 Ma, but it also records three distinct AMSA age provinces, old, intermediate and young (Fig. 1). The occurrence of three distinct AMSA provinces violates predictions of global catastrophic resurfacing, which would require a single AMSA (Phillips and Izenberg, 1995). A related hypothesis, the global stratigraphy hypothesis (Basilevsky and Head, 1996, 1998, 2002), which includes catastrophic resurfacing, can also be tested against the AMSA provinces. The global stratigraphy hypothesis does not predict global resurfacing, but rather catastrophic resurfacing of lowland (80% of Venus), represented by the postulated unit pwr. The global stratigraphy hypothesis makes two specific predictions. The oldest material unit, marked by ribbon tessera terrain fabric, should represent an old surface whereas the unit pwr should correspond to significantly younger surfaces. These predictions are not supported by AMSA province distribution however. The largest tracts of ribbon-tessera terrain occur within highland crustal plateaus, yet these features lay within the intermediate, not the old, AMSA province. Unit pwr extends across $\sim$80% of the surface, avoiding the highest regions; unit pwr correlates widely with old AMSA provinces.

Geologic mapping across lowland regions allows testing of the second prediction required by catastrophic resurfacing and its sister global stratigraphy hypothesis, that is, $\sim$2.5 km thick deposits of unit pwr should lie stratigraphically above basal tessera terrain. Geologic mapping across a region $\sim$25,000 km by 60,000 km within the northern Aphrodite Terra lowland (spatially correlative with old AMSA) indicates that flows that overlie basal ribbon-tessera terrain are extremely thin, likely on the order of hundreds of meters at most, and commonly 10's to 100 m (Lang and Hansen, in press; Hansen, 2006, in press). Large tracts of lowland ribbon tessera terrain are locally covered by a thin veneer of shield-terrain material; pristine impact craters lie directly on basal tessera terrain and/or shield terrain. Numerous other studies also indicate that extensive lowland regions (10s of millions of km^2) are covered by thin (10s to 100 m), rather than 2.5 km thick flows (e.g., Guest et al., 1992; Aubele, 1996; Guest and Stofan, 1999; Deshon et al., 2000; Addington, 2001; Hansen and DeShon, 2002; Brian and Stofan, 2003; Stofan et al., 2005; Hansen, 2005, in press). Thus there is no evidence that unit pwr exists as a layer of $\sim$2.5 km thick flood lava above basal tessera terrain as required by the catastrophic resurfacing and global stratigraphy hypotheses. McGill (2004) also demonstrated that wrinkle ridges formed time transgressively, in strong contradiction with the global stratigraphy hypothesis.

Independent of these considerations, the time period for possible resurfacing has been expanded based on impact crater morphology with recent suggestions by Basilevsky and Head (2002), long time supporters of catastrophic resurfacing. These workers note that the preserved surface history of Venus could span nearly 2000 million years, with 80% of that involving the lowland surface area (Ernst and Desnoyers, 2004).

Proposal of a newly expanded history has critical implications for impact crater morphology. If one assumes that 2.5 km of lava (contrary to recent geologic mapping, but we will let that go for the sake of argument here) was emplaced over a time period of 750–2000 million years, then numerous examples of partially flooded impact craters should be preserved, including breached rims and clear partial burial. But the reason that the catastrophic resurfacing hypothesis was initially embraced over the geologically more reasonable equilibrium-resurfacing hypothesis (e.g., Phillips, 1993; Solomon, 1993a,b) was, in fact, due to the *lack of* partially flooded impact craters, including breached rims (Herrick and Sharpton, 2000). If proponents of the catastrophic resurfacing hypothesis now call for lava emplacement over several hundreds of millions—or billions—of years (e.g., Basilevsky and Head, 2002; Ernst and Desnoyers, 2004) then the (catastrophic) hypothesis, ceases too be 'catastrophic', and thus it no longer addresses the observation with regard to a lack of partially flooded craters. And as such, it does not accommodate the first order data relations, and should no longer be considered a viable hypothesis.

If shield terrain across much of the lowlands is on the order of ~100 m thick then the amount of required lava would be ~37 Mkm^2 (0.1 km over 80% of Venus' surface). This volume is slightly less than the entire suggested volume of lava to build the Ontong Java Plateau (44.4 Mkm^3), yet it would be distributed over an area ~185 times larger, and possibly erupted over a time period 100–200 times longer (Eldholm and Coffin, 2000). Such emplacement does not qualify as a LIP by any standard, and it cannot be labeled a catastrophic event. Thus the proposal that the unit pwr represents the solar system's largest LIP, seems questionable.

If Venus was not catastrophically resurfaced, and unit pwr does not represent the solar systems largest LIP, how then can the distribution and morphology of Venus' impact crater population be addressed? An alternate resurfacing hypothesis that builds on early suggests of equilibrium-resurfacing (e.g., Phillips, 1993; Solomon, 1993a,b) addresses the near global distribution of the crater population, as well as the apparent pristine nature of preserved impact craters (Hansen and Young, in press). The hypothesis, termed the SPITTER hypothesis (Spatially Isolated Time Transgressive Equilibrium Resurfacing), calls for near steady-state impact crater formation and destruction during a time of a globally thin lithosphere, followed by impact accumulation after a global transition to thick lithosphere. The SPITTER hypothesis postulates that localized complete crater destruction occurred through time-integrated formation of spatially isolated crustal plateaus, occurring in large local regions but punctuated in time and space. Although the hypothesis does not depend on a particular mechanism of crustal plateau formation (whether by downwelling, plume or impact-induced lava-pond, but focusing instead on a critical elements common to each of these crustal plateau hypotheses—the complete destruction of previous formed impact craters in the region of individual crustal plateau at the time of formation of that plateau), the lava-pond hypothesis is favored. With a secular change to thick lithosphere crustal plateaus (by any of the three mechanisms) could no longer form, and associated crater destruction would cease; at this time the surface would began to accumulate craters. Locally young surfaces would develop as a result of pronounced local volcanotectonic activity, such as the Beta-Atla-Themis and Lada Terra regions.

Thickening of the lithosphere would cause a marked change in surface evolution in addition to the impact crater preservation. Once the global lithosphere reached a critical thickness, localized focused mantle downwelling would not show surface effects and occurrence of such mantle dynamics would go completely unrecorded at the surface. Similarly, mantle plumes, which might form Artemis-like features on thin lithosphere, would lack the thermal energy to penetrate a thick lithosphere and impose a noticeable rheological imprint on the surface; a thermal plume might cause regional uplift, and localized volcanic activity, as in the case of volcanic rises, but it would not be able to cause destruction of pre-existing craters. The collision of a large bolide (20–30-km diameter) with a thick lithosphere would cause very little melt to form, and it would certainly not result in massive melting of the crust or mantle (e.g., Ivanov and Melosh, 2003). A large bolide would simply make a large crater like the 270-km diameter Mead Crater. Thus in any of these cases, preexisting impact craters would remain unaffected once the lithosphere thickened to some nominal value. The specific value might depend on the favored mechanism(s) of crustal plateau formation. Destruction of impact craters across large localized areas would cease, and the nature of LIPs, formed by almost any mechanisms, would change geologic expression, or even cease form.

9. Summary

Venus, Earth's sister planet lacks plate tectonics, yet hosts numerous examples of LIPs, which may provide clues to LIP formation processes on Earth. Volcanic rises represent the most likely candidates for contemporary plume signatures on thick lithosphere. Volcanic rises include three morphologically distinct groups, rift-, volcano- and coronae-dominated rises, each likely representing differences in either lithospheric properties, or relative stages of formation, or both. Four large coronae, distinguishable from other coronae based on size, are morphologically quite different, and may record vastly different processes of formation. Artemis likely represents the surface manifestation of a plume on thin lithosphere; Heng-O's evolution is unclear and could represent either endogenic or exogenic driven processes. Quetzalpetlatal, the southernmost, and at least in part, the youngest feature discussed, also requires more mapping. Preliminary work favors Quetzalpetlatal's formation above a plume, however Quetzalpetlatal may record several distinct events. Atahensik, the only large coronae to form part of a coronae chain represents a large diapiric structure, though the nature of the diapiric buoyancy is unconstrained. Crustal plateaus, previously variably interpreted forming above deep mantle downwellings, or plumes, likely represent neither. Their distinctive tectonic fabric with areally extensive but structurally localized lava flooding requires a massive amount of lava to have existed at one time. Crustal plateaus may represent the surface scum of huge lava ponds formed by massive partial melting in the mantle due to large bolide impact with ancient thin lithosphere. Finally, the status of unit pwr, previously proposed as the solar system's large LIP, is challenged. Continued study of Venus' LIPs in combination with terrestrial LIPs will benefit our understanding of the dynamics of both planets, and terrestrial planet processes in general.

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