Earth's Oldest Rocks
Edited by Martin J. Van Kranendonk, R. Hugh Smithies and Vickie C. Bennett
Developments in Precambrian Geology, Vol. 15 (K.C. Condie, Series Editor)

DOI: 10.1016/S0166-2635(07)15081-7

Chapter 8.1

VENUS: A THIN-LITHOSPHERE ANALOG FOR EARLY EARTH?

VICKI L. HANSEN

Department of Geological Sciences, University of Minnesota Duluth, 231 Heller Hall, 1114 Kirby Drive, Duluth, MN 55812, USA

8.1-1. INTRODUCTION

Plate tectonics requires a specific global-scale rheology. The Hadean to Eoarchaean Earth likely lacked this rheology locally, regionally, or perhaps even globally. And yet heat was transferred to the surface, and the planet cooled. It can be difficult to envision regional or global processes other than plate tectonics, in part due to the elegance and comprehensiveness of the plate tectonic model. Yet early Earth was significantly different that modern Earth: bolides must have impacted the surface; magma oceans or seas likely existed locally or globally; the 'lithosphere' may have been weak, marked by a ductile-solid rather than a brittle-solid rheology. All of these factors would influence the tectonic processes, as well as the preserved record of operative processes. Venus – Earth's sister planet – of similar age, density, size, inferred composition, and inferred heat budget as Earth, might be expected to cool through similar, terrestrial plate tectonic processes. However, Venus lacks any evidence of plate tectonics (e.g., Solomon et al., 1991; Phillips and Hansen, 1994; Nimmo and McKenzie, 1998). Thus, Venus might provide a rich arena in which to stretch ones' tectonic imagination with respect to non-plate tectonic processes of heat transfer for an Earth-like planet, providing for means to test geologic histories against multiple hypotheses, aimed at understanding possible early Earth processes (e.g., Gilbert, 1886; Chamberlin, 1897).

Contemporary Earth differs from the Hadean to Eoarchaean Earth; contemporary Venus likely also differs from ancient Venus, and certainly differs from contemporary Earth. Although there is much debate about Venus' evolution, it might be that at least a part of the history recorded on Venus' surface provides clues to processes on very early Earth. Venusian conditions now, and in the past, are perhaps more akin to environmental conditions of the early Earth (Lecuyer et al., 2000). Venus' atmosphere, ∼95 bars of supercritical CO_2, forms a strong blanket of insulation with a current surface temperature of ∼475 °C and higher temperatures likely in the past (Bullock and Grinspoon, 1996, 2001; Phillips et al., 2001). Venus is homogeneously hot, resulting in an ultra-dry environment with little sediment formation, transport, or deposition, although it may have experienced a wetter past (Donahue and Russell, 1997; Donahue et al., 1997; Donahue, 1999; Lecuyer et al., 2000; Hunten, 2002). Venus' ~970 recognized impact craters represent some of the most pristine impact features in the solar system. The pristine nature of Venus' surface and these craters are testament to a lack of sediment. The current surface shows only minimal amounts of weathering and erosion, with typical weathering rates estimated at $<10^{-3}$ mm/yr (Campbell et al., 1997). With essentially no difference in diurnal, or equatorial-polar temperatures, Venus lacks regionally organized surface winds.

Venus' crust is believed to be homogeneously basaltic (e.g., Grimm and Hess, 1997), due in part to a lack of water, which plays a critical role in the formation of granitic magma. Venus' dense atmosphere also likely affects volcanic, and presumably tectonic processes in that supercritical CO_2 acts more like a conducting layer than a convective layer with regard to heat transport (Snyder, 2002). Venus' atmosphere may have inhibited crystallization and solidification of lava, contributing to low lava viscosity, and perhaps volcanotectonic styles significantly different that of uniformitarian Earth. Venus preserves a record of an ancient era in which the lithosphere (or crust) was globally thin, followed by contemporary Venus with a thick immobile lithosphere. The interplay of atmosphere and lithospheric processes deserves attention in Venus tectonic-volcanic investigations, and may also play an important role in Hadean to Eoarchaean terrestrial processes.

Venus' atmosphere also blocks optical light, completely veiling the surface prior to development of radar technology. The NASA *Magellan* mission, returned incredible views through Venus' clouds providing detailed, and globally comprehensive, images of the surface – and with this astounding data set, glimpses of a whole new world of tectonic processes. These data are digital in form, global in coverage, and accessible via the world-wide-web.

It is likely that Venus and Earth were most similar at birth. Earth's more contemporary plate tectonic processes destroyed much of the surface record of its early history. Venus' lack of plate tectonics means that Venus might preserve a better record of its formative years, and the tectonic processes that shaped it. Thus, although we cannot travel back in time to the Hadean to Archean on Earth, perhaps we can travel through space to consider possible global-scale tectonic processes in an environment possibly akin to that of early Earth.

In this contribution, I briefly discuss four different types of Venusian tectonomagmatic features: (1) radial coronae, (2) Artemis, (3) crustal plateau fabrics, and (4) deformation belts. Each of these types of features likely formed on thin lithosphere, and in some cases weak lithosphere, and none record plate tectonic-related processes (although their modes of formation may be debated, quite enthusiastically in some cases). Radial coronae and Artemis might represent lithospheric signatures of diapirs, though resulting from compositional and thermal buoyancy, respectively. Hypotheses proposed for crustal plateau evolution include mantle downwelling, mantle plumes, and solidification of huge lava ponds. The latter is favored herein, and may provide clues to the evolution of terrestrial magma ocean surfaces. Crustal plateaux (or the lava ponds they represent) may owe their origin to ancient large impact events. Thus Venus provides a reminder that exogenic processes likely play critical roles on all terrestrial planets, particularly in early planet evolution. Deformation belts form linear high strain zones separated by low strain domains, and may record density inversion of the crust, similar to terrestrial granite-greenstone terrains. These belts provide a caution for interpreting orogen-scale linearity as compelling evidence for plate tectonic processes.

8.1-2. VENUS OVERVIEW

Venus and Earth share many similarities, yet they also have profound differences. Venus, 0.72 AU from the Sun, is 95% Earth's size and 81.5% Earth's mass. 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). Data from Soviet Venera and Vega landers indicate surface element abundance consistent with basaltic composition, although the limited data could accommodate other compositions (Grimm and Hess, 1997). Slow retrograde motion makes a Venus day longer than its year (243 and 225 Earth days, respectively), a factor that may contribute to Venus' lack of a magnetic field (Yoder, 1997). Atmospheric composition (96% CO_2, 3.5% N_2 and 0.5% H_20, H_2SO_4, HCl and HF), surface pressure ($\sim$95 bars) and temperature ($\sim$475 °C) might be similar to Earth's early atmosphere (Lecuyer et al., 2000).

Venus' surface conditions are intimately related to its dense caustic atmosphere, which includes three cloud layers 48–70 km above the surface. The upper atmosphere rotates at a rate of $\sim$300 km/hr, circulating in four Earth days. The clouds reflect visible light and block optical observation. The dense atmosphere results in negligible diurnal temperature variations and an enhanced global greenhouse that makes a terrestrial-style water cycle impossible. Given high surface pressure and temperature, CO_2 exists as a supercritical fluid. Venus lacks obvious evidence of weathering, erosion, and sediment transport and deposition processes, or extensive sedimentary layers clearly deposited by wind or water. 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 $\geqslant$1 billion years ago (Donahue and Russell, 1997; Donahue et al., 1997; Donahue, 1999; Lecuyer et al., 2000; Hunten, 2002). A lack of water renders Venusian (current) crustal rock orders of magnitude stronger than terrestrial counterparts, even given Venus' elevated surface temperature (Mackwell et al., 1998), a factor critical to topographic support. For example, Maxwell Montes, Venus' highest point (11 km above mean planetary radius, MPR, $\sim$6052 km) could only be 5 m.y. old under Venus' current surface conditions if the rock was typical 'wet' terrestrial basalt (Grimm and Solomon, 1988).

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 that of 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^{22} Pa s, Kaula (1990); $10^{22–24}$ Pa s, Turcotte et al. (1999)). Volatiles are of course important in understanding Venus' interior – particularly with regard to viscosity. However, interior volatile values and compositions are currently unconstrained. Lack of volatiles will increase strength and increase the mantle solidus. In contrast, the presence of volatiles will decrease strength and decrease the mantle solidus. Large viscosity contrasts are likely across thermal boundary layers: notably across the lithosphere and core-mantle boundaries. Furthermore, it is possible, and perhaps likely, that mantle viscosity structure has changed through time, and there is no guarantee that contemporary mantle viscosity represents the viscosity structure that accompanied the formation of surface features, and particularly not the surface features discussed herein. The same caution is of course true for the Hadean to Archaean Earth. It does seem clear, however, that, in contrast to Earth, Venus currently lacks a low viscosity asthenosphere, and most likely did so for the duration of its recorded surface history. The lack of an asthenosphere is no doubt critical to Venus' lack of plate tectonics.

Venus' lack of plate tectonics and terrestrial surficial processes (glaciation, erosion, and deposition), results in preservation of a unique surface record of tectonomagmatic processes. Large portions (perhaps all) of the lithosphere have not been completely recycled to the mantle. (In contrast to most workers, Turcotte et al. (1999) propose for episodic lithospheric overturn driven by turbulent mantle flow.) Nor has Venus' surface been extensively dissected, carved or buried as is common on Earth and Mars. Although Venus preserves $\sim$1000 impact craters, its dense atmosphere has shielded its surface from extensive cratering, associated 'gardening', and the development of a thick impact regolith like the Moon. Thus, Venus' surface provides a unique record of non-plate tectonomagmatic processes. Until recently, Venus' atmosphere has veiled this surface from view, but with radar technology the veil has fallen away, allowing us to study this unique surface in incredible global detail. In this contribution, I focus on information gleaned from the surface through the NASA *Magellan* mission.

8.1-2.1. Venus Data

The NASA *Magellan* mission collected four global remote data sets: emissivity, high-resolution gravity, altimetry, and synthetic aperture radar (SAR) images (Ford and Pettengill, 1992; Ford et al., 1993). These data, together with early data from Soviet Venera missions, Pioneer Venus and Arecibo, provide views of Venus' surface. The Soviet Venera Landers provided visible glimpses of the surface and compositional information at a few locations (Barsukov et al., 1986; Surkov et al., 1986). The *Magellan* data, ancillary documentation and software are available through the Planetary Data System [PDS, http://pds.jpl.nasa.gov/]. Emissivity, not discussed herein, is chiefly controlled by dielectric permittivity and surface roughness (Pettengill et al., 1992). Gravity data can resolve features >400 km, and provides clues to subsurface architecture, although interpretations are nonunique. Altimetry data (spatial resolution of $\sim$8 km by $\sim$20 km, along- and across-track; vertical resolution $\sim$50 m) resolves long-wavelength features and morphology, but most topographic features related to primary and secondary structures are only resolvable using SAR data. SAR data ($\sim$100 m/pixel), which covers 98% of the surface with local overlap among mapping cycles, allows for geomorphic and geological interpretations, including geologic surface histories.

8.1-2.1.1. SAR data

SAR data (available at http://pdsmaps.wr.usgs.gov/maps.html) was collected in three cycles: left- (cycle 1) and right-look (cycle 2), and stereo (cycle 3). The effective resolution (Zimbelman, 2001) of SAR images depends, in part, on the features of interest; as a general rule, features, other than lineaments, should be >300 m (Fig. 8.1-1). SAR data from cycles 1 and 3 can be combined to provide true stereo (3D) views (Plaut, 1993), although cycle 3 data is limited. A combination of SAR and altimetry data results in synthetic stereo, 3D, views (Kirk et al., 1992), with near global coverage. Cautions for interpretation of geologic features and histories are outlined in a variety of contributions (Wilhelms, 1990; Ford et al., 1993; Tanaka et al., 1994; Hansen, 2000; Zimbelman, 2001). Global geologic mapping is underway as part of the NASA-USGS VMap (1:5,000,000) program (http://astrogeology.usgs.gov/Projects/PlanetaryMapping/PGM_home.html).

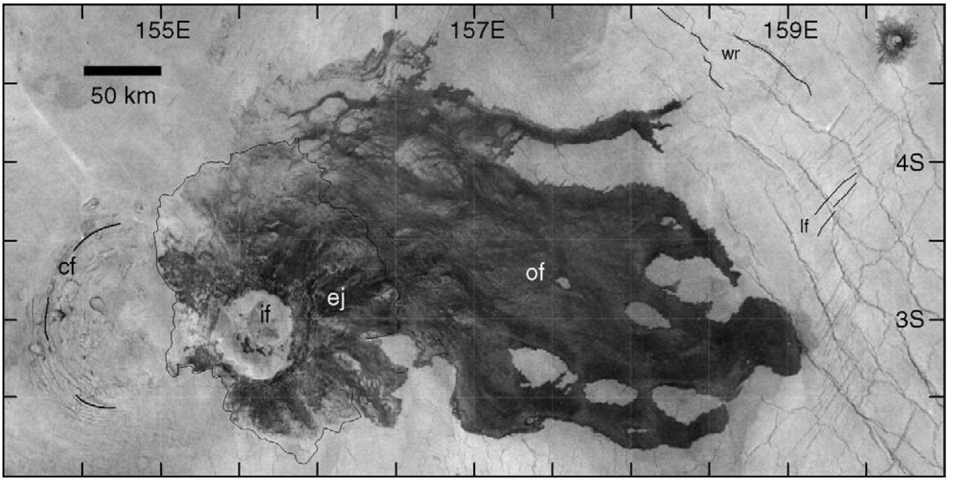
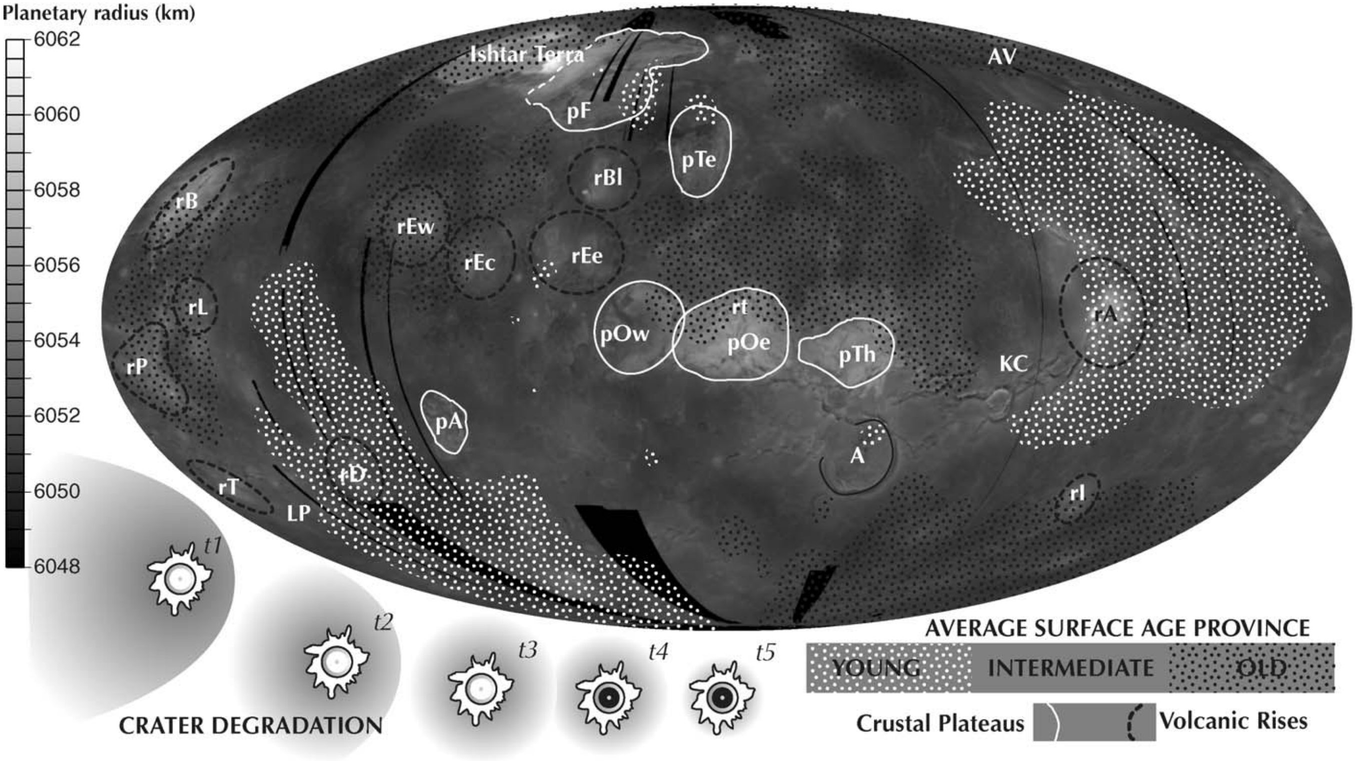


Fig. 8.1-1. Inverted, left-illumination SAR image of Markham Crater, Venus, with extensive outflow material. Flows might be related to the impact event, or they represent subsurface magma that escaped to the surface as a result of bolide impact. Note the central peak, and radar-smooth (bright) flooded interior (if). The crater rim and ejecta (ej) and outflow material (of) appear mostly radar-rough (dark), whereas the basal material is generally radar smooth (bright). Both the ejecta and the outflow material preserve various radar-backscatter facies indicating a range of backscatter properties (likely roughness) within individual geologic material units. The black line marks the limit of the ejecta deposit. Basal material is cut by concentric fractures (cf) to the west of Markham, and by linear (though somewhat sinuous) NW-trending wrinkle ridges (wr, topographic ridges) and NE-trending linear (straight) fractures (lf, narrow troughs) to the east. A small impact crater with a flooded interior and radar-rough ejecta occurs in the NE corner of image.


Planetary radius (km)
6062
6060
6058
6056
6054
6052
6050
6048
Ishtar Terra
AV
pF
pTe
rBl
rB
rEw
rEc
rEe
rL
rA
rP
pOw
rt
pOe
pTh
KC
pA
A
rD
rT
LP
rI
t1
t2
t3
t4
t5
CRATER DEGRADATION
AVERAGE SURFACE AGE PROVINCE
YOUNG
INTERMEDIATE
OLD
Crustal Plateaus
Volcanic Rises

8.1-2.2. Venus Features

Magellan data permit first order characterization of Venus' surface, which is divisible into the lowlands (80%), mesolands (10%), and highlands (10%), based on altimetry. The lowlands, which lie at or below MPR, include relatively smooth low-strain surfaces called plains, or planitiae, and linear deformation belts (Banerdt et al., 1997). The mesolands lie at intermediate elevations and host many coronae, quasi-circular tectonomagmatic features, and chasmata – regional scale linear troughs decorated with tectonic lineaments. Highland regions include volcanic rises, crustal plateaux, and the unique feature Ishtar Terra (Hansen et al., 1997) (Fig. 8.1-2). Volcanic rises are large (1500–2500 km diameter) domical regions, 1–3 km high, marked by local radial volcanic flows. They are widely accepted as contemporary (that is, currently thermally supported) surface expressions of deep mantle plumes on thick lithosphere (e.g., Phillips et al., 1981, 1991; McGill, 1994; Phillips and Hansen, 1994; Smrekar et al., 1997; Nimmo and McKenzie, 1998). Crustal plateaux, similar in planform to rises, but steep sided and flat topped, host unique tectonic fabrics called ribbon-tessera terrain. Plateaux lie 0.5–4 km above their surroundings, the result of shallow isostatic support (Hansen et al., 1997, and references therein). A wide variety of volcanic landforms, preserved at a range of scales, occur across the surface, generally independent of elevation (Head et al., 1992; Crumpler et al., 1997). Hundreds of thousands of volcanic shields, 1–20 km diameter (Guest et al., 1992; Addington, 2001), occur in shield fields ($<$300 km diameter regions) and as 'shield terrain' (Aubele, 1996; Hansen, 2005) distributed across millions of km^2; lava flows up to hundreds of km long are commonly associated with volcanoes, coronae, and fractures (Crumpler et al., 1997). Volcanic forms are generally consistent with basaltic compositions (e.g., Bridges, 1995, 1997; Stofan et al., 2000). Venus also displays unique narrow channels (1–3 km wide) that trace across the lowlands for tens or hundreds of km (up to the $\sim$6900 km long Baltis) (Baker et al., 1997). Although all scientists agree that the channels are fluid cut, many questions remain

---

Fig. 8.1-2. (*Previous page*.) Mollwiede projection of *Magellan* altimetry with average model surface age (AMSA) provinces (data from Phillips and Izenberg (1995)) and major geologic features including crustal plateaux (Alpha (pA), Fortuna (pF), eastern and western Ovda (pOe, pOw), Phoebe (pP), Tellus (pTe), and Thetis (pTh)) and volcanic rises (Alta (rA), Beta (rB), Bell (rBl), Dione (rD), western, central and Eastern Eistla (rEw, rEc, rEe), Imdr (rI), and Themis (rT)). Phoebe (pP) is transitional between a plateau and a rise (Grimm, 1994; Simons et al., 1997; Hansen and Willis, 1998; Phillips and Hansen, 1998). Crater degradation stages show youngest (t1) to oldest (t5) changes in crater morphology; with time and degradation, an crater loses its halo and its interior fills with lava (Izenberg et al., 1994). Three relative AMSA provinces – old, intermediate and young – are defined based on impact crater density and impact crater degradation stage (Phillips and Izenberg, 1995). Figure locations include: Khabuchi Corona, KC (Fig. 8.1-3); Artemis, A (Fig. 8.1-4), Alpha Region, pA (Fig. 8.1-5), Ovda ribbon-tessera terrain, rt (Fig. 8.1-6), Atlanta-Vinmara deformation belts, AV (Fig. 8.1-7), and Lavinia Planitia deformation belts, LP (Fig. 8.1-8). Modified from Hansen and Young (2007).

debated: Are channels erosional or constructional? Do they represent thermal or mechanical processes? What was the nature of the fluid? What is the substrate? Were channels constructional, down cut, or formed by subsurface stoping? (e.g., Baker et al., 1992, 1997; Komatsu and Baker, 1994; Gregg and Greeley, 1993; Bussey et al., 1995; Williams-Jones et al., 1998; Jones and Pickering, 2003; Lang and Hansen, 2006).

8.1-2.3. Venus' Surface, Time and Cautions

Any discussion of Venus geology is not complete without a brief discussion about time. SAR images provide high-resolution views of the surface, which allow determination of cross cutting relations and relative history. However, global, and even regional, correlation of geologic units, or interpreted events, commonly involve circular reasoning given the fundamental 2D nature of remote sensing data (Hansen, 2000). Furthermore, absolute geologic time cannot currently be constrained on Venus. To date, impact crater density provides the only hope of constraining absolute time on planet surfaces other than Earth. Impact crater 'dating' might be viable on Moon, Mars and Mercury due to the extremely high number of total craters and the wide range in surface crater density. Impact crater dating is ultimately a statistical exercise, and includes several geological challenges (Hartmann, 1998). Venus lacks small craters due to screening by the dense atmosphere. 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). In addition, Venus' craters are distributed in near random fashion (Schaber et al., 1992; Phillips et al., 1992; Hauck et al., 1998). The low number of craters and near random spatial distribution prohibit robust temporal constraints for individual geomorphic features or geologic units (Campbell, 1999). The minimum size area that can be dated statistically by crater density alone is 20×10^6 km^2, or 4.5% of the surface (Phillips et al., 1992). Some workers propose age constraints based on combining morphologically similar features/units into large composite regions for crater density dating (e.g., Namiki and Solomon, 1994; Price and Suppe, 1994; Price et al., 1996). Because these works implicitly assume the combined features formed synchronously, the analyses are circular and lack temporally robust conclusions. Furthermore, dating such large surfaces, even if contiguous, requires assumptions that severely limit the uniqueness of any temporal interpretation (Campbell, 1999). In short, even large surfaces (20×10^6 km^2) are effectively 'undatable'.

Impact crater density analysis of Venus results, at best, in determination of average model surface age (AMSA) provinces – the integrated age of a huge region. Venus records a *global* AMSA (that is, an AMSA for the entire surface) of $\sim 750 + 350/{-400}$ Ma, based on total impact craters and impactor flux (McKinnon et al., 1997). This global AMSA could be accommodated by a wide range of possible surface histories – conceptually similar to a terrestrial $\varepsilon_{\mathrm{Nd}}$ average mantle model age (e.g., Farmer and DePaolo, 1983). The global AMSA must be met by any hypothesis, but it provides few unique requirements.

Strom et al. (1994) explored catastrophic (Schaber et al., 1992) versus equilibrium (Phillips, 1993) resurfacing models though Monte Carlo modeling. They varied the areal coverage and iterations of resurfacing from 50%, 25%, 10%, 0.03% and 0.01% of the surface, and considered the final crater distribution (random or not) and the number of embayed craters. The first three experiments yielded low crater embayment, as observed (e.g., Phillips et al., 1992; Schaber et al., 1992; Herrick et al., 1997), but not random crater distribution. In contrast the last two experiments met the random distribution criteria, but predicted high crater embayment. Thus, Strom et al. (1994) called for catastrophic volcanic resurfacing of Venus, ~500 Ma, with ~3 km thick flood lava emplaced globally over a 10–100 m.y. event. The longer the 'catastrophic' event, the more embayed craters, and thus the more at odds with the data. In keeping with catastrophic resurfacing, Basilevsky and Head (1996, 1998, 2002) proposed that Venus displays a coherent global stratigraphy with basal tessera terrain buried by 1–3 km thick flood lava, emplaced quickly and recently.

Although crater density alone cannot delineate statistically distinct, temporally defined, regions, Phillips and Izenberg (1995) subdivided the surface into three AMSA provinces using impact crater density *and* crater morphology (Fig. 8.1-2). Izenberg et al. (1994) recognized a temporal sequence of impact crater degradation allowing the division of craters into relative age groups. Young craters display haloes and radar-rough interiors; old craters lack haloes and show radar-smooth (presumably flooded) interiors. The AMSA provinces – which represent relative rather than absolute age provinces – cannot constrain the age of individual geologic features or units, but rather they represent an average age of an integrated history of these surfaces, reflecting geologic processes that would lead to formation, modification, or destruction of impact craters. Because crater formation is global, and because craters are mostly pristine, the critical factor would seem to be process(es) of crater destruction. Although no individual geologic units or features are robustly temporally constrained, individual features, or groups of features, might show spatial patterns with respect to the three AMSA provinces, and such patterns might provide clues to the relative temporal evolution. But such spatial correlation should never be accepted as a robust age. The presence of three AMSA provinces does, however, provide strong evidence against the hypotheses of global catastrophic volcanic resurfacing of Venus (e.g., Schaber et al., 1992; Strom et al., 1994) and global episodic lithospheric overturn (Turcotte, 1993; Turcotte et al., 1999), which each require a single global AMSA (Phillips and Izenberg, 1995). Hansen and Young (2007) evaluate resurfacing hypotheses with implications for Venus evolution, a topic outside the limits of the current contribution. The important point for the current discussion is that absolute age is unconstrained across Venus, although most workers agree that Venus likely experienced an early Era marked by globally thin lithosphere (<30 km), followed by contemporary Venus marked by thick (100–300 km) lithosphere (e.g., Solomon, 1993; Grimm, 1994; Solomatov and Moresi, 1996; Phillips et al., 1997; Schubert et al., 1997; Hansen and Willis, 1998; Brown and Grimm, 1999; Phillips and Hansen, 1998). The timing of the global transition from thin to thick lithosphere is unconstrained. Venus currently lacks a sharp asthenosphere boundary (Phillips and Hansen, 1994; Phillips et al., 1997; Schubert et al., 1997), perhaps the most important feature of terrestrial plate mechanics.

8.1-3. THIN LITHOSPHERE TECTONOMAGMATIC FEATURES

Terrestrial provinces that preserve views into early Earth typically provide a record of crustal depth, but a regional plan-view record is extremely limited. Venus, on the other hand, provides essentially only a plan-view. Venus' plan-view is continuous and available at an amazing scale of observation, such that geologic histories (and hence temporal dimension – albeit only relative time) might be interpreted, with appropriate cautions.

8.1-3.1. Radial Coronae: Surface Expression of Compositional Diapirs?

Coronae (Barsukov et al., 1984), commonly considered unique to Venus, are circular to quasi-circular features typically marked by a raised rim or annulus that displays concentric annular structures (fractures, faults or folds), and variable tectonic and volcanic features, including radial fractures and extensive lava flow deposits (Fig. 8.1-3). Coronae range in size from 60–1050 km diameter (200 km median), and number about 500 (Stofan et al., 1992, 2001). Most coronae occur in chains (68%) or clusters (21%) spatially associated with mesoland chasmata and volcanic rises, respectively; limited coronae (11%) occur as isolated features in the lowlands (Stofan et al. 1992, 1997, 2001; DeLaughter and Jurdy, 1999). 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 forming tectonomagmatic 'blisters' in/on the lithosphere (e.g., Stofan et al., 1992, 1997; Squyres et al., 1992a; Janes et al., 1992; Janes and Squyres, 1995; Koch and Manga, 1996; Smrekar and Stofan, 1997). Diapiric models propose evolution characterized by: central doming, radial fracturing, volcanism, eventual reduction of interior topography, production of an annular ring, and possible late subsidence. The wide range of coronae characteristics might represent stages of corona evolution, or they might indicate that features collectively referred to as coronae include genetically unrelated features. For example, some or all coronae could represent volcanic calderas, impact craters, or Rayleigh–Taylor instabilities in a density stratified lithosphere (e.g., Squyres et al., 1992a; Nikolayeva, 1993; Hamilton 1993, 2005; Schultz, 1993; McDaniel and Hansen, 2005; Vita-Finzi et al., 2005; Hoogenboom and Houseman, 2006). The spatial association of corona chains and clusters with chasmata and rises, respectively, favors endogenic (over exogenic) formation for these types of coronae. In short, all coronae might not have formed by the same processes. Discussion here focuses on coronae marked by radial fractures.

Khabuchi Corona (Fig. 8.1-3) represents a typical 'radial corona', within an equatorial coronae chain. Radial coronae fit the predictions of diapiric models with radial fractures and flows that formed broadly synchronously, followed by temporally overlapping concentric fracture formation, with emergence of additional surface flows. Radial coronae seem to lie dominantly within corona-chasmata chains and in clusters associated with volcanic rises, and record rich histories involving broadly contemporaneous fracturing, folding, volcanism, and presumably subsurface magmatism (Hamilton and Stofan, 1996; Stofan et al., 1997; Copp et al., 1998; Chapman, 1999; Hansen and DeShon, 2002). Radial coronae with

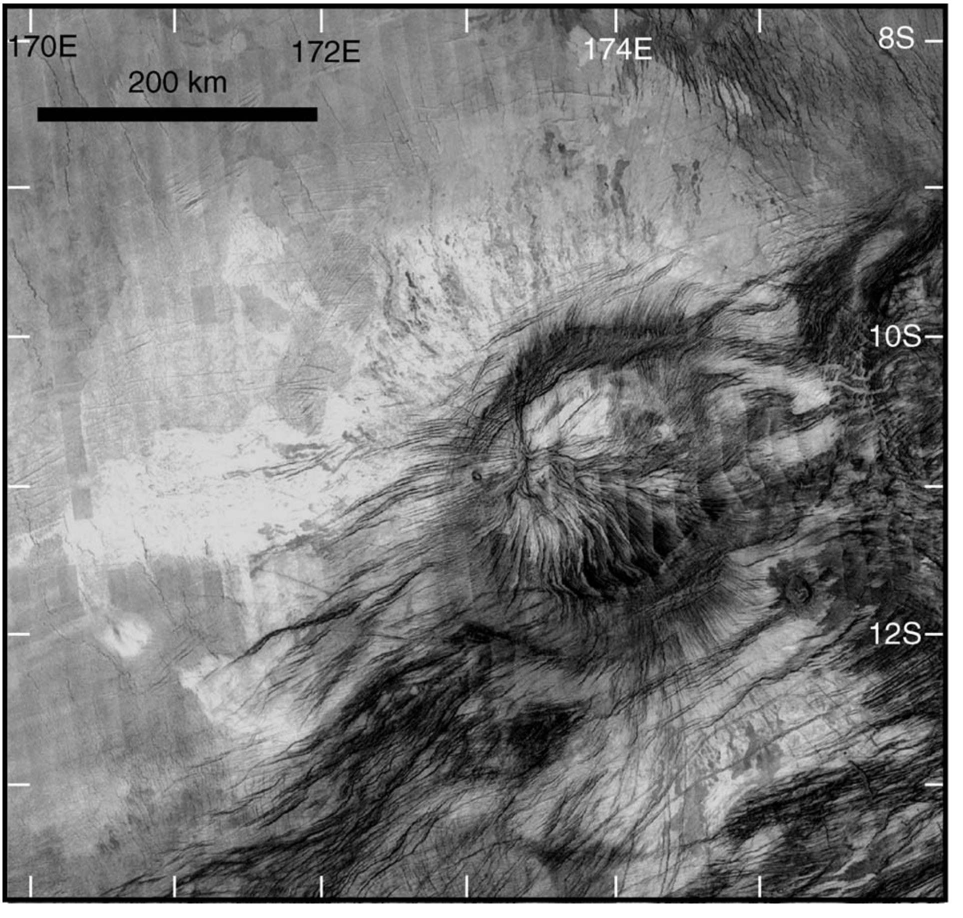


Fig. 8.1-3. Inverted, left-illumination SAR image of Khabuchi Corona, marked by radial and concentric fractures and radial volcanic flows. Interior region sits topographically high relative to the surroundings.

diameters <350 km likely formed on relative thin crust/lithosphere (5–10 km; Janes and Squyres, 1995; Koch and Manga, 1996), and result from diapirs driven by compositional buoyancy (Hansen, 2003). The periodic spacing of chained coronae may reflect the depth (150–250 km) to a layer of mantle instability that served as the source layer (Hamilton and Stofan, 1996).

Radial corona, likely related to diapiric rise and associated deformation of cover material in the form of radial fracturing, concentric shedding of cover (forming concentric folds or fractures), and synchronous volcanism might be broadly analogous to the formation of some terrestrial granite-greenstone belts (e.g., Rey et al., 2003). Indeed, clustered coronae associated with volcanic rises are proposed to result from a deep mantle plume, similar to hypotheses proposed for granitoid doming in the Pilbara Craton (e.g., Pawley et al., 2004; Van Kranendonk et al., 2004; Smithies et al., 2005b) and the northeast Superior Province (e.g., Bédard et al., 2003). Hoogenboom and Houseman (2006) propose that coronae (they do not differentiate corona type) result from lithospheric density inversions (density inversions are discussed further in the section on deformation belts). Presumably, topography would subside with time, being thermally supported, but the tectonomagmatic signature would remain. Similarly, chained coronae – postulated to form above cylindrical mantle upwellings – might be analogous to Archean granite-greenstone terrains, which might represent just a fragment of an originally much more expansive terrain. If coronae existed in the early Earth, they might have played a critical role in early tectonic processes contributing to planet cooling and heat transfer, as well as perhaps crustal differentiation. If coronae formed in a subaqueous environment, they could have harbored early life forms with interaction of aqueous systems (e.g., Van Kranendonk, 2006). Numerous questions with regard to coronae evolution on Venus remain unanswered and controversial.

Radial fracture patterns also form giant radial dike swarms across Venus (e.g., Ernst et al., 2001, 2003). Some overlap exists between features mapped as radial coronae and as giant radial dike swarms; the two types of features could be genetically related, or the overlap may be serendipitous. Radial dike swarms typically have radii that far exceed that of coronae annuli. For example, the radial dike swarm that centers on Heng-O Corona (1010 km diameter annulus) has a radius of >1000 km. Although giant radial dike swarms occur on Venus and Earth, the oldest known terrestrial giant radial dike swarm is Early Proterozic – far younger than Hadean to Eoarchaean. The formation of giant radial dike swarms require huge expanses of strong lithosphere, or plates. Therefore, the occurrence of giant radial dike swarms might place a minimum temporal limit on the existence of global scale plates, and provide robust temporal constraints on lithosphere rheology. Indeed Venus' giant radial dike swarms cross cut, and are therefore younger than, both ribbon tessera terrain and deformation belts (Ernst et al., 2003), discussed below. Although giant radial dike swarms exist on both planets, they likely formed relatively late in planet evolution and therefore are not discussed further herein. On Earth the formation of such features might reflect global conditions ripe for modern plate tectonic processes.

8.1-3.2. Artemis: Surface Expression of a Large Mantle Plume?

The formation of Artemis, the largest circular feature on Venus, and perhaps the largest circular feature in the solar system, remains a puzzle. Artemis comprises a huge topographic welt, 2600 km in diameter, that includes a paired circular trough (150–200 km wide; $\sim$1–1.5 km-deep) and outer rise (200 km wide) (Fig. 8.1-4). Artemis defies geomorphic classification: it is similar in size to crustal plateaux and volcanic rises, yet topographically more akin to many coronae. Artemis has been classified as a corona (Stofan et al., 1992), but given its large size, this classification is questionable (Stofan et al., 1997, Hansen, 2002). Herein the feature is simply referred to as 'Artemis', following Hansen (2002). Artemis' trough describes a partial circle that extends clockwise from $\sim$12:00 to 10:30 in

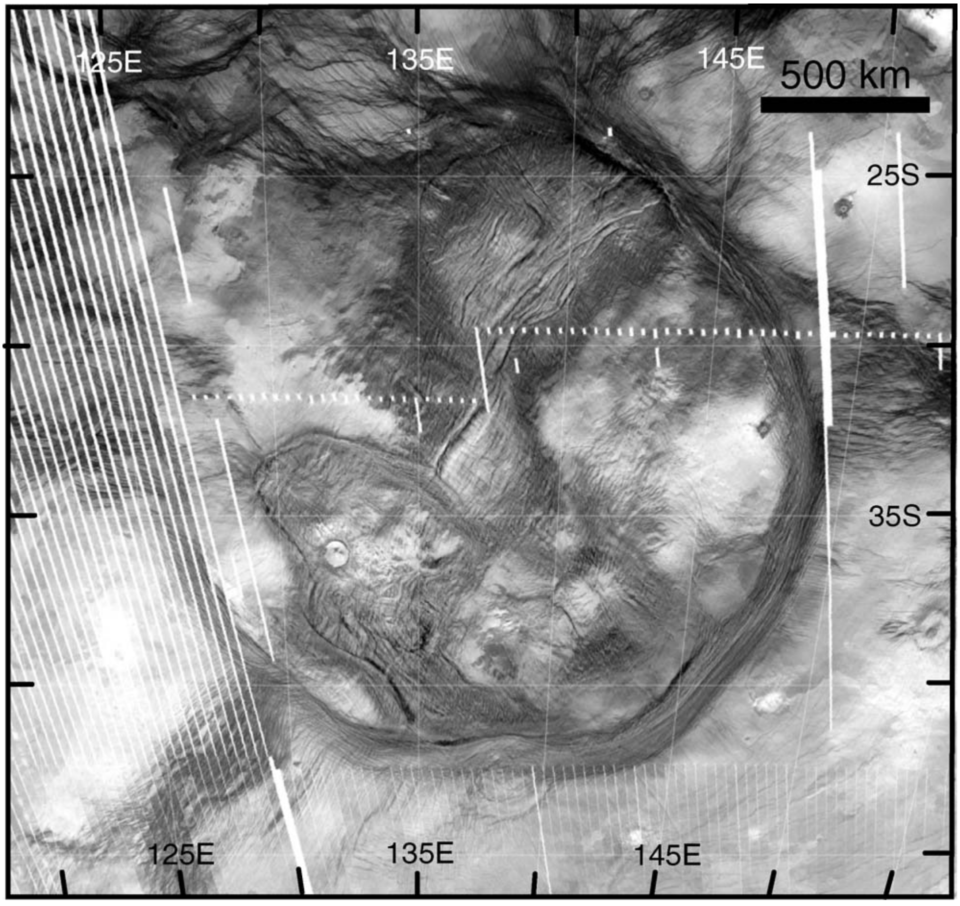


Fig. 8.1-4. Inverted SAR image of Artemis. The white strips, most obvious on the western side of the image, mark gaps in the SAR data. Artemis is defined by the circular feature (dark), which forms a 150–200 km wide topographic trough marked by closely-space (1–2 km) linear structures that parallel the associated portion of the trough. The interior hosts four tectonomagmatic centers marked by radial lineaments and flows, and preserves a penetratively developed linear fabric that generally trends northeast, becoming radial near tectonomagmatic centers.

an analog clock framework, with ends that gradually disappear both topographically and structurally. Short wavelength structures (<1 km) within the trough parallel the trend of the trough. Brown and Grimm (1996) mapped Artemis' trough and rise structures in detail, and Hansen (2002) mapped the interior, trough, and adjacent region in reconnaissance fashion. A 1:5,000,000 scale geologic map (V-48, Artemis) is under review with the U.S. Geological Survey (Bannister and Hansen, 2006). The interior, which sits 2–4 km above the adjacent lowlands, hosts four tectonomagmatic centers – three marked by radial fractures and flows. A fifth possible center, marked by radial fractures but lacking obvious flows, overlaps a portion of the southern trough. A penetratively developed, ∼500 m wavelength, fabric occurs across much of Artemis, trending generally northeast, but taking on a radial character near three of the tectonomagmatic centers (Bannister, 2006; Bannister and Hansen, 2006). The trough hosts trough-parallel structures, likely a combination of folds, faults and scarps. Radial extension fractures and trough-concentric wrinkle ridges dominate the rise outboard from the trough, with concentric wrinkle ridges continuing outward for hundreds of kilometers.

Four different hypotheses have been proposed for the formation of Artemis: (a) and (b) related to a subduction interpretation (Brown and Grimm, 1995, 1996; Spencer, 2001); (c) as Venus' largest impact structure (Hamilton, 2005); (d) as the surface expression of a large mantle plume on thin lithosphere (Griffiths and Campbell, 1991; Smerkar and Stofan, 1997; Hansen, 2002). Part of the challenge of understanding Artemis' formation is tied to how Artemis is defined. Is Artemis composed of interior, rim and outer rise that are genetically related; or did each of these regions form separate from one another, with the interior representing a sort of 'captured' real estate?

The subduction hypothesis 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 apparent depth of compensation 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 underplated material or melt residuum within a plume context. Gravity analysis, fraught with assumptions including assumed single depths of compensation, results in non-unique interpretations. Brown and Grimm (1995, 1996) proposed that Artemis Chasma resulted from northwest-directed subduction beneath Artemis' interior; they further suggested that Artemis Chasma includes three 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; 12:00 to 2:30 represents an associated trough dominated by left-lateral displacement; and 6:30–10:30 represented an older feature, genetically unrelated to the other two segments. Spencer (2001) interpreted a part of Artemis' interior as a region of major crustal extension similar to a terrestrial metamorphic core complex. Although, Spencer (2001) did not place the study within a regional context, he inferred that the proposed extension related to regional subduction.

Compelling arguments against the subduction hypothesis include: (1) the angle of subduction required by the tight curvature of Artemis trough is not geometrically viable on a Venus-sized planet; (2) documented continuity of structures along the entire trough in a trough parallel fashion clockwise from 12:00 to 10:30, and a shared central location of trough topography, trough structures, radial fractures and wrinkle ridges, support the interpretation that the various features of Artemis are genetically related; (3) kinematic arguments would require right-lateral displacement along the southwestern part of the trough to accompany subduction, yet interior flows traverse the southwestern trough margin; and (4) within this same region, interior graben extend across the trough to the exterior, providing further evidence that this portion of the trough did not experience right-lateral displacement. These observations collectively argue for the evolution of Artemis through a single coherent process, rather than serendipitous alignment of two or more unrelated events as required within the context of the subduction hypothesis (Hansen, 2002; Bannister, 2006; Bannister and Hansen, 2006).

Hamilton (2005) asserted that Artemis records the impact of a huge bolide on a cold solid Venus at ∼4–3.5 Ga. Unfortunately the 'hypothesis' lacks details, or even clarifying statements or predictions. The impact hypothesis for Artemis formation does not consider many first-order aspects of Artemis, including topography and geologic relations. Artemis' topographic form, with a narrow (100–150 km) circular trough surrounding a raised interior, is opposite to that of large impact basins on Mars and the Moon, with circular rims surrounding interior basins. For example, Mars' Hellas Crater, widely accepted as impact in origin, forms a 2000 km diameter, 6–8 km deep, circular basin surrounded by a greatly modified, but still present, outer rim. Hamilton (2005) infers that early Venus would have been rheologically similar to Mars during the formation of Hellas and therefore, within the context of the impact hypothesis, the two huge impact features should show similar first-order character. Large impact basins also commonly show multiple ring morphology (Hartmann, 1998), features Artemis clearly lacks. Contrary to the assertion by Hamilton (2005), there is no evidence that the northwest margin of Artemis (the arc between 9:30–12:00 in the analogue clock model) is buried beneath other constructs (Brown and Grimm, 1996; Hansen, 2002; Bannister, 2006; Bannister and Hansen, 2006); yet such a large impact basin would be expect to show a complete circular structure. Finally, the impact hypothesis does not address the formation of documented interior tectonomagmatic features, or penetrative fabric, despite the inference that Artemis represents a coherent set of features formed within a geological instant of time.

Currently the most viable hypothesis for Artemis formation seems to be the surface manifestation of a mantle plume on thin lithosphere, consistent with its large size and circular planform. Gravity-topography analysis, though non-unique, is consistent with at least partial dynamical support for Artemis (Simons et al., 1997). As a deep mantle plume rises toward the lithosphere, the lithosphere will be uplifted, and, if the strength of the lithosphere is exceeded, radial fractures could form above the plume head. Alternatively, if the lithosphere were sufficient heated, it might 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 Griffiths and Campbell's (1991) experiments, as a plume head approached the rigid horizontal boundary, it collapsed and spread laterally. A layer of surrounding 'mantle', squeezed between the plume and the surface, resulted in a gravitationally trapped asymmetric instability and led to the formation of an axisymmetric trough. In addition, the interior squeeze layer might 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 become manifested as interior tectonomagmatic centers. It was on the basis of these experiments that a plume model for Artemis formation was proposed following initial 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). In this case, the trough results from lithospheric delamination. Delamination might contribute to a 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 may have resulted, at some first-order level, from the interaction of a deep mantle plume and relatively thin lithosphere, and as such it may provide valuable clues to the possible formation of Archaean terrestrial plumes and the structures generated therein. It is possible that Artemis could hold clues for processes transitional between plume-dominated and plate-dominated (e.g., Bédard et al., 2003; Bédard, 2006).

8.1-3.3. Crustal Plateaux: Analog for Ancient Magma Ocean Surfaces?

Crustal plateaux (Figs. 8.1-2 and 8.1-5) host distinctive deformation fabrics (Fig. 8.1-6), herein called ribbon-tessera terrain following terminology of Hansen and Willis (1996, 1998). Scientists generally agree that crustal plateaux are isostatically supported in the shallow crust or mantle, as evidenced by small gravity anomalies, low gravity to topography ratios, shallow apparent depths of compensation, 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, 1992b; Bindschadler, 1995; Hansen et al., 1999; Ghent and Hansen, 1999). Researchers also widely accept that arcuate-shaped inliers of characteristic ribbon-tessera terrain within the lowland represent ancient collapsed crustal plateaux remnants (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).

Two basic questions emerge with respect to plateau formation.

1. How were plateau surfaces deformed and concurrently uplifted?
2. How did plateaux 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 and Parmentier, 1990; Bindschadler et al., 1992a, 1992b; Bindschadler, 1995). The plume hypothesis accommodates thickening and deformation via magmatic underplating 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). Both hypotheses call for time-transgressive deformation of ancient *thin* lithosphere above individual spatially localized regions, and both embrace the suggestions that a root of thickened crust supports each plateau and that 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 through lower crustal flow at

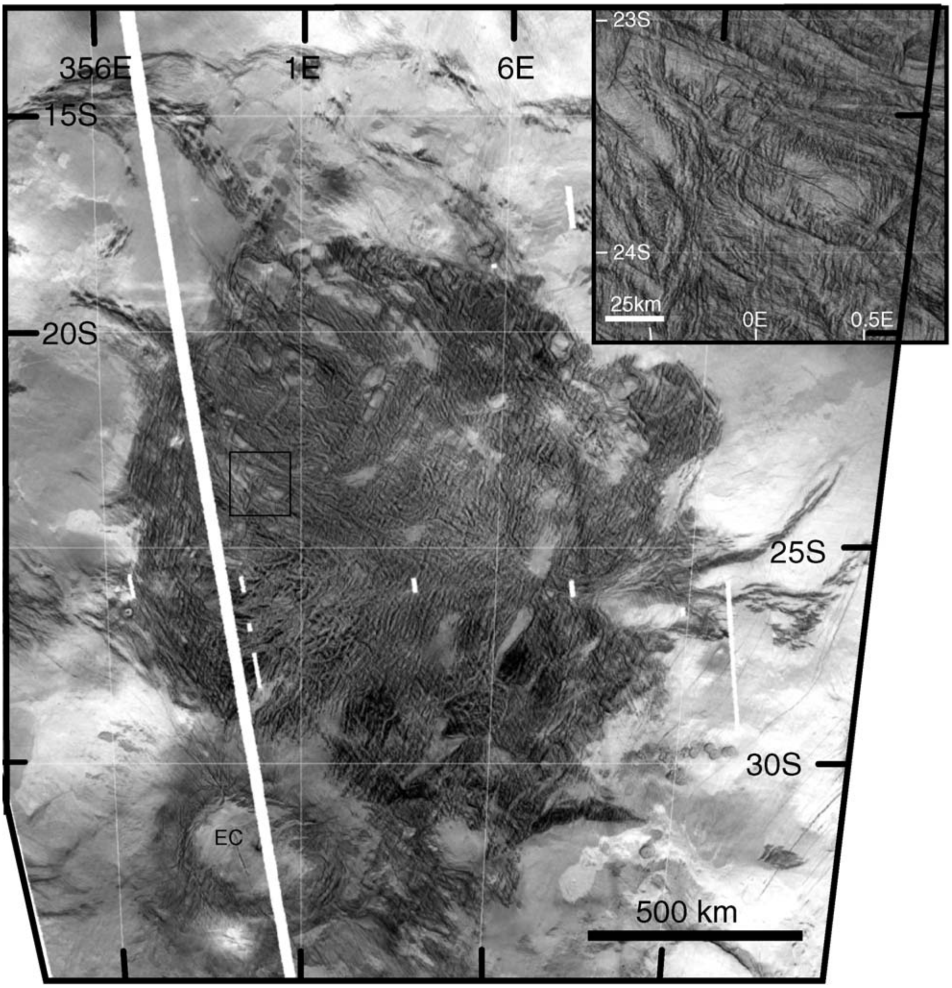
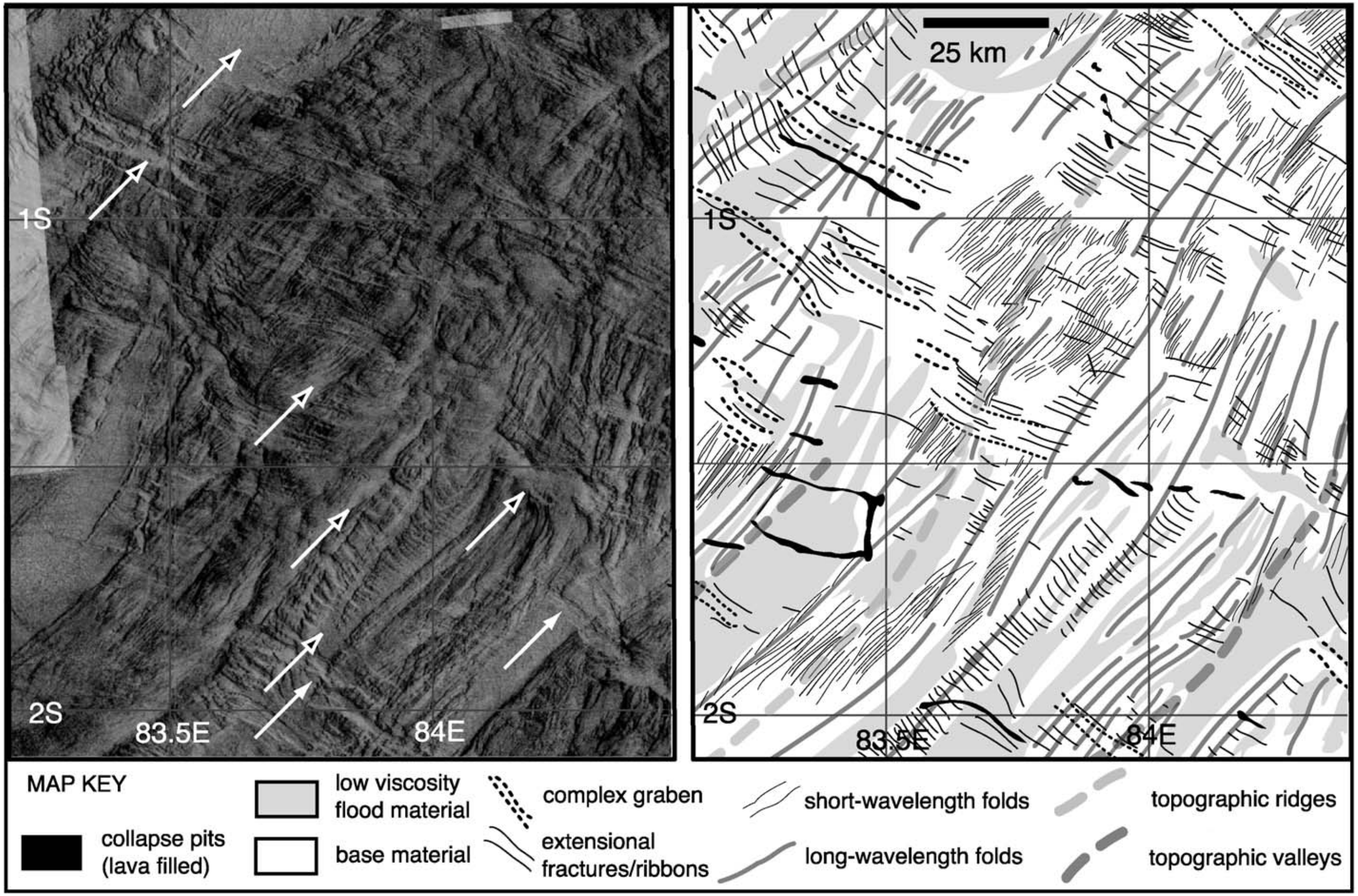


Fig. 8.1-5. Inverted SAR image of Alpha Regio, a typical crustal plateau with distinctive radar-rough (dark) terrain residing in a elevated plateau above the adjacent radar-smooth lowlands (bright); the circular feature that overlaps Alpha along its southwest margin is younger Eve Corona, EC. Inset (box) shows detail of the ribbon-terrain fabric: periodically spaced parallel ridges and troughs, trend north; fold ridges trend west-northwest. Short wavelength folds ($\sim$1 km) parallel longer-wavelength folds, but occur below image resolution here. Smooth, light-colored regions represent radar-smooth surfaces, interpreted as areas covered by low viscosity lava flows. White bold lines mark data gaps and indicate the spacecraft track.


1S
2S
83.5E
84E
25 km
MAP KEY
collapse pits (lava filled)
low viscosity flood material
base material
complex graben
extensional fractures/ribbons
short-wavelength folds
long-wavelength folds
topographic ridges
topographic valleys

geologically reasonable time scales (Nunes et al., 2004). Thus, neither the range in plateaux elevations, nor the inliers of ribbon tessera-terrain are addressed by either the downwelling or the plume hypotheses.

Additionally, neither of these hypotheses address all characteristics of crustal plateaux, and each carries specific burdens. Challenges for downwelling include: (a) a predicted domical form (Bindschadler and Parmentier, 1990) rather than the observed plateau shape; (b) lower crustal flow called upon for crustal thickening requires 1–4 billion years, well outside reasonable time constraints (Kidder and Phillips, 1996); and (c) formation of documented short-wavelength extensional structures (ribbon fabrics) requires a high geothermal gradient (Hansen and Willis, 1998; Gilmore et al., 1998), which is difficult to justify in a relatively cold downwelling environment (Hansen et al., 1999). The plume hypothesis can accommodate formation of a plateau shape and extensional features. However, both extensive contractional strain, and formation of short-wavelength folds are difficult to accommodate (Ghent at al., 2005). Although the plume hypothesis addresses formation of late long-wavelength folds (or warps), which record very little shortening ($<1\%$), 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 well above that expect within the environment of a plume-lithosphere interaction.

Despite deep divides within the crustal plateau debate, SAR image mapping on both sides leads to four, mutually agreed upon, observations: (1) plateaux host both contractional structures (folds) and extensional structures (ribbons, extensional troughs, graben), which are generally mutually orthogonal; (2) there are multiple suites of folds, defined by wavelength; (3) there are multiple suites of extensional structures, defined by spacing; and (4) low viscosity fluid, presumably lava, fills local to regional topographic lows. Despite these agreements, controversy exists as to the relative timing of flooding and deformation, and until recently, the amount of shortening has been unconstrained.

Detailed SAR image mapping aimed at addressing the timing of deformation and flooding, and placing limits on shortening strain, yielded new observations and refined geologic histories for plateau surfaces (Fig. 8.1-6), resulting in the proposal of a third hypothesis – the lava-pond hypothesis (Hansen, 2006). Geologic relations call for progressive deformation of an initially very thin layer (10s to 100 m) developed across individual plateaux. The layer shortened, forming ductile folds, and extended in an orthogonal direction along brittle structures (ribbons). With additional shortening, earlier formed short-wavelength structures

---

Fig. 8.1-6. (*Previous page.*) Inverted left-illumination SAR image and interpretive map of a region within crustal plateau eastern Ovda Regio, illustrating the nature of the ribbon-tessera terrain fabric along the crest of a long-wavelength ($\sim$100 km) fold. Note local flooding of medium-wavelength fold troughs preserved in crests, limbs and trough of long-wavelength folds. Also note late collapse pits and associated lava deposits. Arrows with black heads indicate locations where flooding postdated local deformation; arrows with white heads indicate locations where deformation postdated local flooding. See Hansen (2006) for details.

were carried piggyback 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. Early flooded lows were carried piggyback on younger, longer-wavelength structures (Fig. 8.1-6). Subsurface liquid (magma) formed a sharp decrease in viscosity with depth, required by structural constraints, and served as the source of flood material. Early terrestrial magma oceans may have followed crystallization processes akin to plateaux surface evolution, although testing this might require identifying large tracts of ancient surfaces, rather than subsurface exposures.

The lava-pond hypothesis calls for progressive solidification and deformation of the surface of huge individual lava ponds, each with areal extent marked by individual plateaux. Ribbon-tessera terrain represents lava pond 'scum'. Individual lava ponds resulted from massive partial melting in the shallow mantle caused by large bolide (20–30 km diameter) impact on thin lithosphere (Hansen, 2006). Melt rose to the surface leaving behind a lens of low-density mantle residuum (e.g., Jordon, 1975, 1978). This hypothesis follows the recent suggestion that the terrestrial greater Ontong-Java Plateau formed as a result of large bolide impact on thin lithosphere (i.e., Ingle and Coffin, 2004; Jones et al., 2005), following earlier suggestions (Rogers, 1982; Price, 2001). Isostatic adjustment in the mantle, resulting from the low-density residuum lens, 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 individual plateaux. Remnants of distinctive ribbon-tessera terrain fabrics could survive as a record of an ancient lava pond. Thin surface deposits could partially or completely cover the fabrics, obscuring or erasing, respectively, evidence for individual lava ponds. The lava pond hypothesis addresses the detailed ribbon-tessera history of orthogonal folding and extension at a wide range of wavelengths from 0.1 km to tens of km, as well as the formation and subsequent collapse of ancient crustal plateaux.

Massive partial melting within the shallow mantle could result from: (a) a large bolide impact with ancient thin lithosphere, (b) rise of an extremely hot deep mantle plume beneath ancient thin lithosphere, or (c) a plume spawned by large bolide impact on thin lithosphere. In any case, crustal plateaux require thin lithosphere (as with the downwelling and plume hypotheses), and they owe their topographic stature to a low-density mantle residuum lens, rather than thickened crust. A bolide impact mechanism for melt-generation is favored because the formation of a lava-pond necessitates a large volume of magma at the surface at one time. 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 a large 'hole', but the lithosphere across a several thousand-km scale could retain its strength – although it might likely be riddled with fractures (Jones et al., 2005). Ivanov and Melosh (2003) state that large bolide impact cannot generate huge volumes of melt, yet others present convincing counter arguments, particularly if a large bolide impacts hot, thin lithosphere (Jones et al., 2005; Elkins-Tanton and Hager, 2005). Clearly such lines of inquiry are in nascent stages of investigation. Hot thin lithosphere, critical to formation of huge melt volumes, might be easily accommodated on ancient Venus, or on early Earth. In addition, a huge body of lava might cool slowly because Venus' dense CO_2 atmosphere acts more like a conductive layer than a convection layer in terms of heat transfer (Snyder, 2002). This brings to mind how early Earth's atmosphere might also affect heat transfer processes, and lava solidification.

The bolide impact and lava pond hypotheses also provide a mechanism to concentrate radiogenic elements in early-formed crust, with possible further differentiation into a subsurface felsic layer beneath a more mafic surface 'scum'. Crustal scale lithologic/density/radiogenic stratification at a map-scale similar to Venusian crustal plateaux is proposed for terrestrial granite-greenstone terrains (e.g., West and Mareschal, 1979; Mareschal and West 1980; Collins et al., 1998; Chardon et al., 2002; Rey et al., 2003; Sandiford et al., 2004), although a lava pond mechanism has not been considered to date. Surely early Earth was bombarded by bolides, which likely affected the early lithosphere. Bolides could have contributed to early mantle differentiation processes, including residuum formation, which could in turn lead to cratonization (e.g., Jordon, 1975, 1978; Bédard, 2006). Large bolide impacts may have contributed to the formation, and preservation of early crust.

8.1-3.4. Deformation Belts

Although circular features dominate Venus' surface, it also preserves large-scale linear features, including: wrinkle ridges, distributed across huge tracts of the surface; extensive fracture belts, thousands of km long and hundreds of km wide; and zones of focused strain, called deformation belts. Deformation belts, first recognized in Venera data (Basilevsky and Head, 1988), rise $\sim$1 km above their surroundings in the lowlands, and host 1-km wide ridges that mark folds or graben (Frank and Head, 1990; Kryuchkov, 1990). Deformation belts commonly occur in groups separated by inter-belt regions. Belts are 100–250 km wide and 100 to $>$1000 km long; inter-belt regions are $\sim$100–400 km wide and are elongate to equant (Solomon et al., 1992; Squyres et al., 1992b). The periodic nature of deformation belts is particularly apparent in the Atlanta-Vinmera region (Fig. 8.1-7). In early works, the periodicity of deformation belts was proposed as resulting from either harmonic buckling instabilities driven by regional compression, likely the result of the coupling of large-scale mantle convection with the lithosphere (Zuber, 1987, 1990), or widespread contraction of the crust, with deformation belts elevated by focused thrusting (Frank and Head, 1990). Both models addressed constraints derived from low resolution Venera data, and called for region crustal shortening with post-deformational flooding of the inter-belt regions. Analysis of Magellan data revealed several challenges to these early models, including: superposed contraction and extension structures, along strike changes from contraction to extension structures, syntectonic volcanism, orthogonal deformation belts, and evidence for strain localization within the deformation belts, as opposed to evidence for burial of inter-belt deformation (e.g., Squyres et al., 1992b; Phillips and Hansen, 1994; Addington, 2001; Rosenberg and McGill, 2001; Young and Hansen, 2005). The origin of deformation belts remains enigmatic. I briefly review geologic relations within

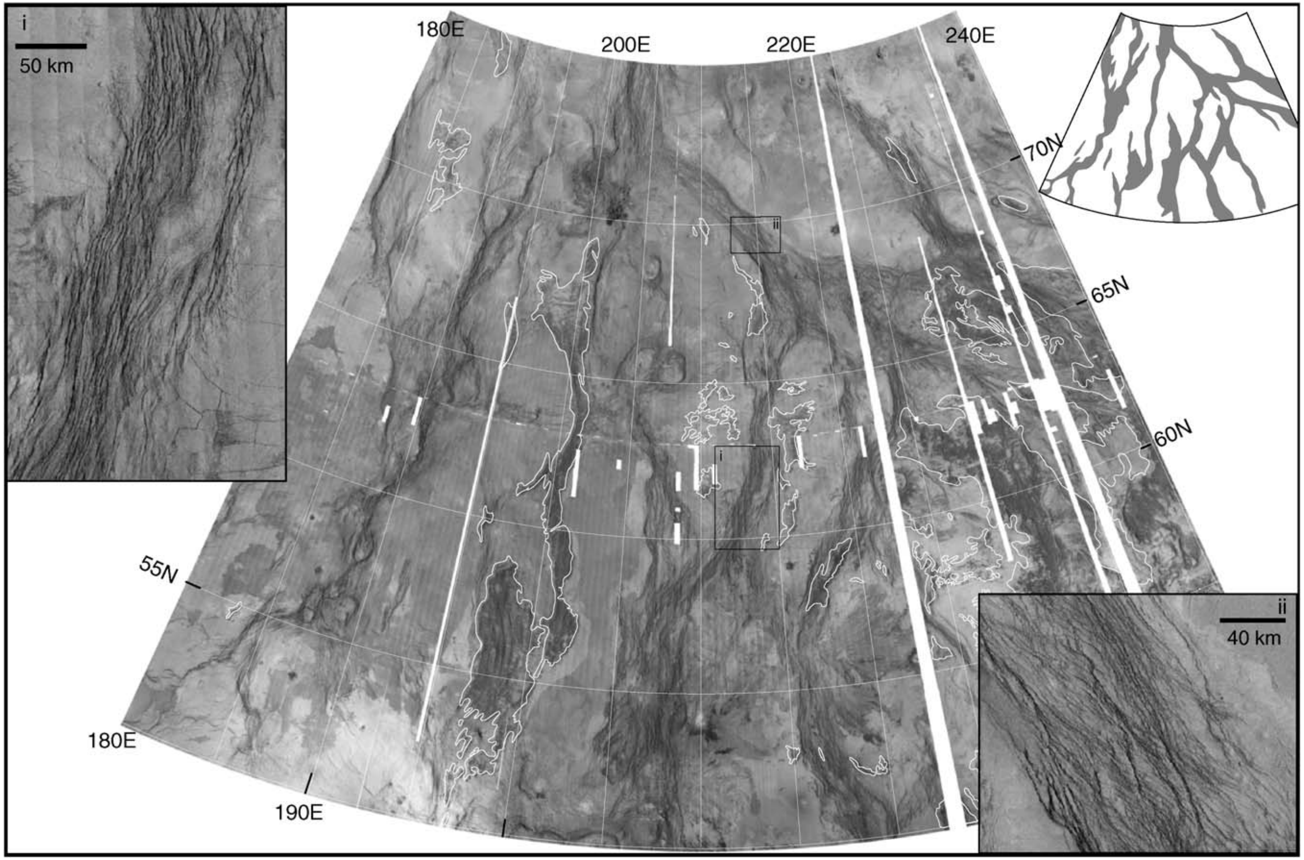

i
50 km
180E
200E
220E
240E
70N
65N
60N
55N
180E
190E
ii
40 km

Fig. 8.1-7. (*Previous page*.) Inverted SAR image of the Atlanta-Vinmara deformation belts; close-up images (i and ii) illustrate local detail; and a sketch of anastomosing deformation belt patterns (upper right corner). Dark regions (radar-rough) characterize the deformation belts, whereas homogeneous gray areas (radar-smooth) represent intervening regions of low strain. White lines indicate exposures of ribbon-tessera terrain. The large image covers ~2500 km in a north-south direction.

two deformation belt provinces; the broadly parallel, but anastomosing Atalanta-Vinmara belts of the northern hemisphere, and the orthogonal belts of Lavinia Planitia preserved in the southern hemisphere. I compare these provinces with terrestrial granite-greenstone terrains.

The Atalanta-Vinmara belts form an impressive array of anastomosing deformation (Fig. 8.1-7), generally interpreted as resulting from global-scale shortening normal to their trends. Within the belts, smooth ~1 km wide ridges generally define folds, although fractures and graben occur locally, and even locally dominate. Deformation features grade outward from each belt, whether marked by contractional or extensional structures. Both the belts and inter-belt regions preserve outcrops of ribbon-tessera terrain, which clearly predated belt formation. The belts and the inter-belt regions also preserve evidence of localized volcanic activity in the form of small shields. Volcanism predated, accompanied, and postdated deformation. Collectively, these relationships illustrate that the belts represent high strain zones, or strain localization, compared to low strain inter-belt domains.

Lavinia Planitia preserves generally orthogonal deformation belts ~50–200 km wide, and up to ~800 km long that form topographic highs and record concentrated strain (Fig. 8.1-8). Strain corresponds to belt orientation. NE-trending belts display folds, whereas NW-trending belts exhibit fractures and graben (Fig. 8.1-8). Belts that trend between these orientations host folds and fractures in patterns that reflect plan-view non-coaxial shear. ENE-trending belts record right-lateral shear whereas NNW (to N)-trending belts record left-lateral shear (Koenig and Aydin, 1998; Hansen, 2006, unpublished mapping). Inter-belt regions record relatively low strain, with contractional (wrinkle ridges) and extensional (fractures) structures parallel in trend to their respective counterparts within the belts. A strikingly simple pattern represented by a single regional bulk strain ellipse emerges across Lavinia, suggesting that deformation within and between belts occurred broadly synchronously. As in the case of Atalanta-Vinmara, ribbon-tessera terrain and shields occur in both the belt and inter-belt domains; shields broadly predated, accompanied and post-dated deformation. Coronae-sourced flows locally embay and bury eastern deformation belts.

8.1-4. EARLY EARTH ANALOGUES

The Atalanta-Vinmara and Lavinia regions share first-order characteristics and histories, although the shapes of their low strain regions differ: elongate versus equant, respectively.

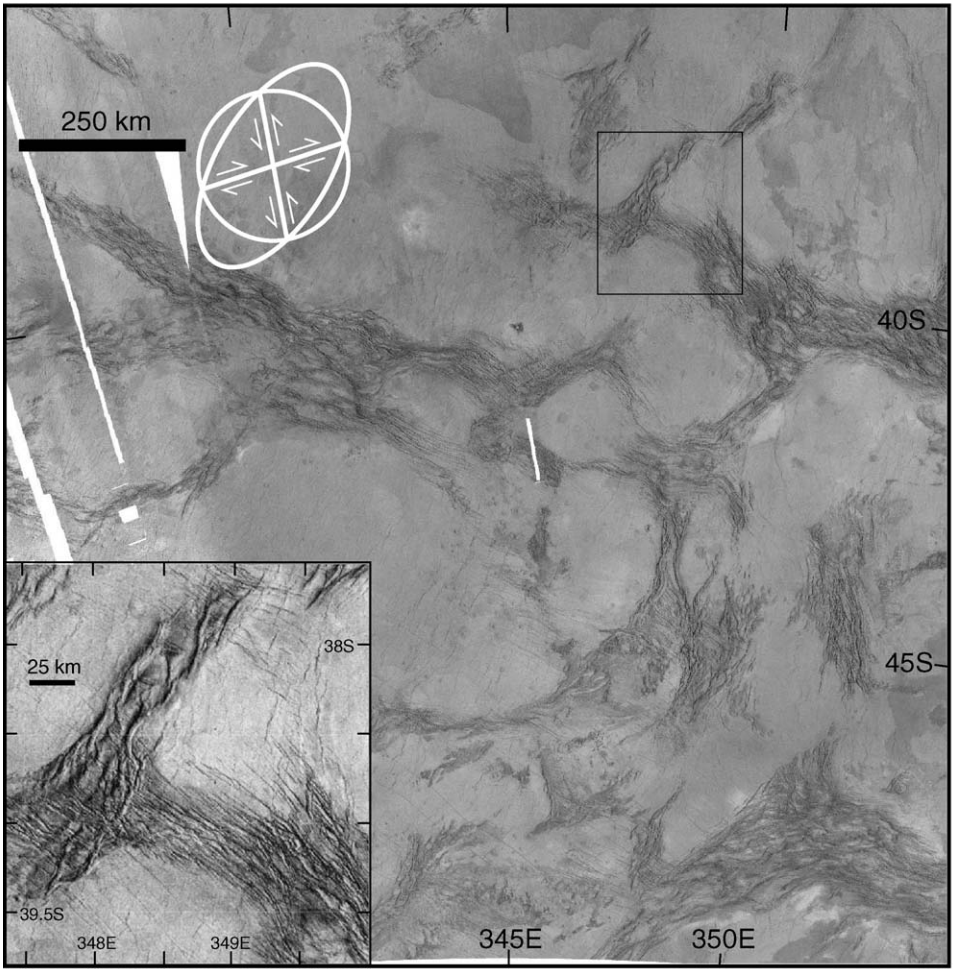


Fig. 8.1-8. Inverted SAR image of Lavinia Planitia deformation belts. Belt-parallel folds characterize NE-trending belts whereas belt-parallel extension fractures characterize NW-trending belts. Inset shows detail of NE-trending belt marked by fold ridges and an orthogonal NW-trending belt marked by extensional structures. Low inter-belt regions display low strain; wrinkle ridges (NE-trending) and extension fractures (NW-trending) parallel deformation belt folds and fractures, respectively. Box indicates location of inset SAR detail. Regional bulk strain ellipse (orientation, not magnitude) matches strain in individual deformation belts as a function of belt orientation shown. See text.

Their map-view stain patterns are similar in plan-view to terrestrial granite-greenstone terrains, which display crustal views. In this analogy, low strain inter-belt domains correspond to granite domes, and deformation belts correspond to greenstone belts. The Atalanta-Vinmara region is more akin to the Yilgarn and Superior Provinces – each displaying elongated granite-greenstone patterns; whereas Lavinia mimics the Pilbara and Dharwar cratons, displaying more equant patterns. The Yilgarn and Superior Provinces have been widely interpreted as evidence for terrestrial Archaean plate tectonic processes, due in large part to the linear structural pattern that extends for thousands of km, and taken to record progressive accretion of distinct arc terranes similar to the modern North American Cordillera (e.g., Card, 1990; Van Kranendonk, 2003c, 2004a, and references therein). However, the linear pattern across millions of km^2 preserved in the Atalanta-Vinmara province did not result from plate tectonic processes.

Terrestrial granite-greenstone terrains are variably interpreted as the result of two end-member models: diapirism and sagduction, versus the accretion of arc terranes within a plate tectonic framework (e.g., Mareschal and West, 1980; Choukroune et al., 1997; Card, 1990; de Wit, 1998; Lin, 2005). This debate has raged for decades and shows little sign of subsiding (e.g., Van Kranendonk, 2004; Van Kranendonk et al., 2004; Cawood et al., 2006). Although terrestrial belts preserve moderately deep crustal views, Venus' deformation belts provide a surface plan-view, and it is more extensive than the view presented by Archaean cratons. In addition, although the role of plate tectonics can be debated in terrestrial cases, it is highly unlikely that plate tectonic processes operated on Venus. Thus the Venusian examples might provide clues for crustal scale density inversion processes. The low strain regions in Venusian deformation belts could mark sites of subsurface diapirism, or low density crust which has moved upward, and the deformation belts could represent sagduction, with local high strain and thickening of a surface cover layer as it sheds off the low strain regions. The Venusian belts, like terrestrial granite-greenstone terrains, show broadly synchronous deformation and volcanism. In the case of the Venusian provinces, synchronous volcanism and contractional deformation has been difficult to explain to date, but both processes might be predicted and addressed within the context of a partial convective overturn model (cf. Collins et al., 1998).

First order similarities of deformation-belt terrains and granite-greenstone terrains beg for future comparative study with fundamental first-order benefits for Venusian and terrestrial studies. Do deformation belts provide evidence of an ancient density-layered crust on Venus? Venus is currently believed to boast a basaltic undifferentiated crust based on Venera composition data and hypsometric relations. A compositionally layered crust with surface basalt and a felsic subsurface could accommodate current constraints, yet could also provide a mechanism for deformation belt formation. Perhaps a density/composition/isotopic-layered crust marks a common stage of terrestrial planet formation. An early-formed felsic (and thus also radiogenic) rich layer could serve to insulate the underlying mantle, leading to partial melting and subsequent formation of mafic melt, which could in turn make its way to the surface where it could form a high-density layer, and perhaps a thermal blanket. Could the higher geothermal gradient expected in early terrestrial planet evolution, together with blanketing and insulation (e.g., Rey et al., 2003; Sandiford et al., 2004) lead to rheological softening of the layered crust, leading in turn to ductile flow and subsequent density inversion? Perhaps a subsurface felsic layer rich in radiogenic elements, could insulate the underlying mantle, leading to partial melting without requiring a plume (e.g., Rey et al., 2003; Bédard, 2006). Crustal density inversion processes would be variably arrested with cooling, or with loss of water – each of which would inhibit crustal ductility. Crustal inversion would not require regional scale horizontal shortening or extension, although such processes could accompany crustal inversion. The Atalanta-Vinmara and Lavinia regions might preserve evidence of a regionally extensive weak crustal rheology – that is, a regionally extensive crust as a ductile solid, as opposed to a brittle-solid crust, following suggestions for a weak terrestrial Archaean crust (e.g., Choukroune et al., 1997; Chardon et al., 2003; Sandiford et al., 2004; Bédard, 2006; Cagnard et al., 2006). Clearly the ideas presented here require further study, but perhaps comparison of Venus' deformation belts and Archaean granite-greenstone terrains could lead to new understanding of both geological provinces, and even early processes of crust formation. Conceivably, granite-greenstone terrains are a natural evolutionary process in the formation of terrestrial planet crust; they may not require mantle plumes, but rather a precursor stratified crust that insulated the underlying mantle and led to partial melting.

8.1-5. SUMMARY

Heat transfer processes drive terrestrial planet evolution. The heat budget, structure and rheology likely change throughout the evolution of terrestrial planets. Similarities and differences between Venus and Earth provide a valuable tectonic experiment that might provide critical clues to understanding the earlier history of our own planet, which may have experienced early tectonic processes quite different than its current trademark plate tectonics. Within the last 15 years, Venus' surface has become visible and accessible to anyone with world-wide-web connection, making Venus 'field work' accessible and inexpensive. Venus' environmental boundary conditions are certainly different than contemporary Earth, and may be more similar in many ways to Earth's Hadean to Archean Era. Any understanding of Venus tectonic processes and planet evolution are surely still in a nascent stage, and yet Venus serves as a rich tectonic playground to stretch one's imagination, and to challenge one to think beyond the elegance of plate tectonics in constructing hypotheses of ancient terrestrial planet processes.