

Venus Interior Structure and Dynamics

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Abstract No two rocky bodies offer a better laboratory for exploring the conditions controlling interior dynamics than Venus and Earth. Their similarities in size, density, distance from the sun, and young surfaces would suggest comparable interior dynamics. Although the two planets exhibit some of the same processes, Venus lacks Earth's dominant process for losing heat and cycling volatiles between the interior and the surface and atmosphere: plate tectonics. One commonality is the size and number of mantle plume features which are inferred to be active today and arise at the core mantle boundary. Such mantle plumes require heat loss from the core, yet Venus lacks a measurable interior dynamo. There is evidence for plume-induced subduction on Venus, but no apparent mosaic of moving plates. Absent plate tectonics, one essential question for interior dynamics is how did Venus obtain its young resurfacing age? Via catastrophic or equilibrium processes? Related questions are how does it lose heat via past periods of plate tectonics, has it always had a stagnant lid, or might it have an entirely different mode of heat loss? Although there has been no mission dedicated to surface and interior processes since the Magellan mission in 1990, near infrared surface emissivity data that provides information on the iron content of the surface mineralogy was obtained fortuitously from Venus Express. These data imply both the presence of continental-like crust, and thus formation in the presence of water, and recent volcanism at mantle hotspots. In addition, the study of interior dynamics for both Earth and exoplanets has led to new insights on the conditions required to initiate subduction and develop plate tectonics, including the possible role of high temperature lithosphere, and a renewed drive to reveal why Venus and Earth differ. Here we review current data that constrains the interior dynamics of Venus, new insights into its interior dynamics, and the data needed to resolve key questions.

Venus III

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1 Introduction

Venus is Earth's slightly smaller twin in terms of size and bulk composition. However, it differs in many respects, including its very dense CO₂-rich atmosphere, its very high surface temperature, its slow spin rate of 243 Earth days, and the apparent absence of plate tectonics. It is also closer to the Sun and one may wonder whether this fact alone is responsible for the very different evolution that Venus has undergone after its accretion. This question has become increasingly important with the discovery of thousands of exoplanet candidates and the search for the so-called Earth 2.0 that would look like the Earth with the presence of some kind of life at its surface. One objective of Venus exploration is to understand why Earth and Venus have followed different evolutionary paths and to determine what factors result in an Earth-like planet. In particular, it is crucial to understand why plate tectonics is present on Earth and absent on Venus. Plate tectonics is a very efficient process to cool down a planet and to recycle volatiles in the interior. Determining the interior structure and simulating its interior dynamics contributes to the global understanding of how a planet evolves, including its long-term habitability (Franck et al. 2000; Driscoll and Bercovici 2013; Foley and Driscoll 2016; Hoening and Spohn 2016).

The information collected by the Venus Express mission that directly relates to the interior structure and dynamics is limited because the science payload was mostly devoted to determining the composition and dynamics of Venus' atmosphere. Magnetic field measurements are consistent with the lower limit set by Pioneer Venus (Phillips and Russell 1987). Doppler tracking data from Venus Express, which had a nominal periapsis of 250 km (Svedhem et al. 2007), is currently being incorporated into a revised gravity field model (Goossens et al. 2017). Although the Magellan gravity cycles were acquired at 200 km, the viewing geometry also contributes to resolution, and the Venus Express data may locally improve the resolution of the field.

Venus Express made an unexpected contribution to our understanding of the interior of Venus through the discovery of possibly active volcanism via a thermal signature (Shalygin et al. 2012) and recent volcanism from surface emissivity near 1 micron that distinguishes fresh from weathered basalt (Helbert et al. 2008; Smrekar et al. 2010; Davaille et al. 2017). All four of the geologic regions with apparently recent basaltic volcanism occur in regions previously classified as sites of mantle upwelling, or hotspots, based on their geology (e.g. McGill 1994; Ivanov and Head 2010). The gravity and topography data are consistent with the presence of active mantle plumes at depth (Kiefer and Hager 1991; Smrekar and Phillips 1991; Simons et al. 1997). The presence of apparently recent flows corroborates this interpretation. Another six hotspots in the northern hemisphere, which are not covered in the 1 micron emissivity data, have been proposed to have plumes at depth (Kiefer and Hager 1991; Smrekar and Phillips 1991; Smrekar and Stofan 1997; Simons et al. 1997). The new evidence for recent volcanism at hotspots spurred several studies of interior convection aimed at reproducing a similar number of active plumes (Smrekar and Sotin 2012; Huang et al. 2013). The presence of active mantle plumes, which implies significant heat extraction from the core, together with the absence of a dynamo, constitutes a key constraint for interior evolution models. Additionally, VIRTIS data support the hypothesis that Venus' large plateaus are analogous to Earth's continents (Gilmore et al. 2015, 2017). On

Earth, continents form when basalt remelts in the presence of water, commonly at subduction zones.

Further, over the lifetime of Venus Express, our understanding of Venus' interior has advanced due to both a new understanding of planetary volatiles and rocky planet interior evolution. A longstanding hypothesis is that Venus evolved differently from Earth because its interior is dry. A planet's volatiles strongly affect not only climate and habitability, but also interior viscosity, melting temperature, and surface deformation. New measurement of cometary volatiles and solar system dynamics suggest a new paradigm of planetary volatile origins. Recent measurements of D/H ratio (Altwegg et al. 2015) and noble gases including Xenon isotopes (Marty et al. 2017) suggest that the cometary contribution to the Earth's ocean is much smaller than thought previously. Instead the majority of terrestrial planet's original volatiles are believed to come from the accretion of carbonaceous chondrites (Altwegg et al. 2015). Volatiles are then released to the atmosphere via the outgassing from magma ocean and volcanic processes. The discovery of greater concentrations of water in Earth's mantle (Palot et al. 2016) than previously known is consistent with this understanding of planetary volatile origins. Additionally, the discovery of many low density exoplanets close to their suns challenge the idea that a planet's volatile concentration as a function of distance from its star is fixed in time.

Models of the interior dynamics of a terrestrial-type planet have made significant progress. This paper reviews the models that have been developed in order to understand why the two sister planets Earth and Venus are so dramatically different. The relationship between mantle convection and plate tectonics deserves particular attention. This question is triggered by the exponentially growing list of exoplanet discoveries that raises the issue of whether an Earth-sized exoplanet is Earth-like or Venus-like. As of April 2018, over 3700 exoplanet candidates have been confirmed (<http://exoplanet.eu/>).

Among those, 782 have a known radius of less than three times Earth's radius. If we restrict the statistics to the number of planets with known mass and radius (mass less than 30-MEarth and radius less than 2-REarth), the number drops to 51.

One key parameter controlling the tectonic style, i.e. plate tectonics or stagnant lid or episodic plate tectonics, may be the surface temperature (Landuyt and Bercovici 2009; Lenardic and Crowley 2012; Noack et al. 2012; Bercovici and Ricard 2014; Gillmann and Tackley 2014) as discussed in Sect. 3. The effective temperature (i.e. surface temperature of a rapidly spinning blackbody sphere simulating the planet) of most exoplanets is larger than the equilibrium surface temperature of Mars, Earth, and Venus (Fig. 1). The effective temperature is calculated assuming an albedo of 0 (for simplicity) for exoplanets and albedos of 0.67, 0.367, and 0.17 for Venus, Earth, and Mars, respectively. We assume a fast rotator for the exoplanets. The equation is:

$$T_{eff} = \left[\frac{L(1-a)}{16\pi\sigma d^2} \right]^{1/4}$$

where L is the star luminosity, a is the albedo, σ is the Stephan–Boltzmann constant, and d is the semi-major axis of the planet. Note that for a slow rotator like Venus or a planet in spin-orbit resonance, the coefficient would be 8 instead of 16. Because of the power 1/4, the differences are not that large and due to our ignorance of those parameters for exoplanets, we have chosen the simplest parameters. The actual temperature varies around this equilibrium temperature for Mars and the Earth whereas it is much larger for Venus due to its greenhouse. This demonstrates the importance of understanding the coupling between interior dynamics and atmospheric composition. It also highlights the importance of studying the tectonic

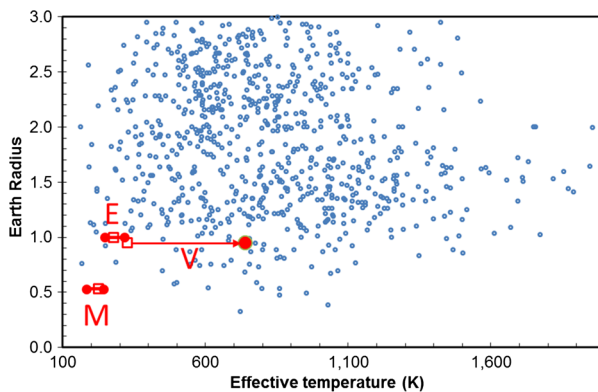


Fig. 1 A large number of exoplanets have the size and effective temperature close to those of the solar system rocky planets (M = Mars; E = Earth; V = Venus). The red circles represent the maximum and minimum temperatures for the Mars and the Earth and the average surface temperature for Venus. The effective temperature falls between the max and the min temperatures for both Mars and the Earth. For Venus, the effective temperature is much lower than the actual surface temperature due to the very strong greenhouse effect. Data are from ‘The Extrasolar Planets Encyclopedia’ (<http://exoplanet.eu/>)

evolution of Venus for providing insights on the interior dynamics and history of habitability of planets that have high equilibrium temperatures.

After describing our present knowledge of Venus’ interior structure (Sect. 2), this paper provides a review of our present perception of the processes at work in the mantle (Sect. 3) and the core (Sect. 4) based on laboratory experiments and numerical experiments, and how these processes differ between the Earth and Venus. Finally, Sect. 5 discusses the kind of measurements that would help constrain the interior structure of Venus.

2 Constraints on the Interior Structure and Dynamics

Observations of Venus constrain both the geologic history of the planet and its interior dynamics. Some fundamental features include a small area of highlands relative to low relief plains, a lack of interconnected plate boundaries, only ~ 1000 impact craters, and a strong correlation between gravity and topography. Venus’ surface is dominated by volcanism, with roughly 80% of the surface covered in low relief volcanic plains. Some highlands are large volcanic rises, which have been interpreted as sites of active mantle plumes, like Hawaii. These large volcanic rises, or hotspots, have broad topographic rises (> 1000 km diameter), large positive gravity anomalies, large volcanoes, and sometimes rifts. The six tessera plateau highlands are also ~ 1000 – 1500 km in diameter, and have multiple, intersecting tectonic deformation trends. Analysis of the gravity and topography of these features shows that they are isostatically compensated (Smrekar and Phillips 1991; Jiménez-Díaz et al. 2015; Yang et al. 2016). Only one tessera plateau has spectral data related to composition. The VIRTIS 1 micron data at Alpha Regio indicates that the plateau is more felsic than the surrounding plains (Gilmore et al. 2015, 2017), further supporting the hypothesis that these features may be analogous to terrestrial continents.

2.1 Resurfacing History and Recent Activity

The surface age of a planet is a first order constraint of its level of geologic and interior activity. Venus is the only rocky body in our solar system with an average resurfacing age comparable to that of Earth. This value is derived from its population of ~ 1000 craters, and was initially estimated to be 300 m.y. to 1 b.y. (e.g. Phillips et al. 1992; Strom et al. 1994; McKinnon et al. 1997). The uncertainty is large as the population of impactors in the inner solar system during the last 1 b.y. is not well constrained. Recent surveys have improved the characterization of this population, leading to updated resurfacing ages for Venus of 150–250 m.y. (Le Feuvre and Wieczorek 2011; Bottke et al. 2016). Bottke et al. (2016) proposes that the population of near-Earth objects is responsible for the formation of the impact population on Venus, which yields an average resurfacing age of 120–250 m.y. This assumption is controversial as it requires that projectiles create larger impact craters than predicted by current models. Additionally, Herrick and Rumpf (2011) interpret the dark floored craters, which comprise 80% of impact craters, as being volcanically flooded, giving an average resurfacing age of ~ 150 m.y. Although the exact resurfacing age is debated, clearly Venus has remained geologically vigorous compared to the smaller terrestrial planets.

In addition to the relatively young surface age, the distribution and modification of the impact crater population constraint the geologic evolution. The crater distribution cannot be distinguished from a random one, and only a small fraction are unequivocally modified by volcanism or tectonism. These two observations lead Strom et al. (1994) to propose that Venus completely resurfaced within a short time, or ‘catastrophically’, followed by a relatively quiescent period (Strom et al. 1994; Phillips et al. 1992). As discussed below, numerous geodynamic models have been proposed to satisfy this ‘catastrophic resurfacing’ hypothesis, including episodic lithospheric foundering (Parmentier and Hess 1992; Turcotte 1993) and episodic plate tectonics (or active lid) convection (Moresi and Solomatov 1998; Armann and Tackley 2012; Rozel 2012).

However, the impact crater population is equally consistent with equilibrium resurfacing, as shown by Phillips et al. (1992). Several recent studies have used Monte Carlo modeling to examine the time evolution of the impact crater population and specific models of the size and shape distributions of volcanic flows. This work is consistent with both catastrophic (Romeo and Turcotte 2010) and more equilibrium (but regionally variable) resurfacing histories, and requires resurfacing on a scale of 100 s–1000 km (Bjornes et al. 2012; O’Rourke and Korenaga 2014). Analysis of the density of the extended impact ejecta blanket and crater density shows that there are both relatively young, volcanically resurfaced, and old regions (Phillips and Izenberg 1995). Thus, the impact crater population is consistent with, but does not require, episodic plate tectonics.

Data from Venus Express has also shown evidence for geologically recent and even active volcanism. The concentration of atmospheric SiO_2 increased significantly over the duration of the Venus Express mission, which can be interpreted as either long period oscillations in atmospheric circulation or evidence for volcanic outgassing (Marcq et al. 2013). As discussed more fully in the Gilmore et al. chapter, variations in surface emissivity at 1.02 microns derived from Venus Express VIRTIS data provide evidence for several locations of recent volcanism (Helbert et al. 2008; Smrekar et al. 2010). All of these locations occur above sites identified as having topography, gravity, and deformation structures consistent with mantle plumes (Smrekar et al. 2010; Ivanov and Head 2010; Davaille et al. 2017). This dataset covers much of the southern hemisphere, although the signal to noise provides good quality precision over only approximately 25% of the planet (Mueller et al. 2008). The other mantle plumes not covered by this dataset may well be active. Other types of

volcanic features that were either not well observed or are too small to be observed at the ~ 90 km resolution of the VIRTIS emissivity may also have been active on geologically recent time scales. Rapidly eroding pyroclastic flows identified in Earth-based, polarized radar also provide evidence of recent activity (Campbell et al. 2017). These flows are very rough, suggesting collapse of a very volatile (up to $\sim 5\%$ CO_2 or H_2O) rich plumes. They are found at one of the mantle plume areas with a high emissivity anomaly, as well as at some that are not covered in VIRTIS emissivity. Possible active volcanism was identified in a rift zone using data from the Venus Monitoring Camera on Venus Express (Shalygin et al. 2012). Collectively these observations do not rule out catastrophic resurfacing or provide a useful constraint on resurfacing rate, but do confirm that Venus is geologically active today.

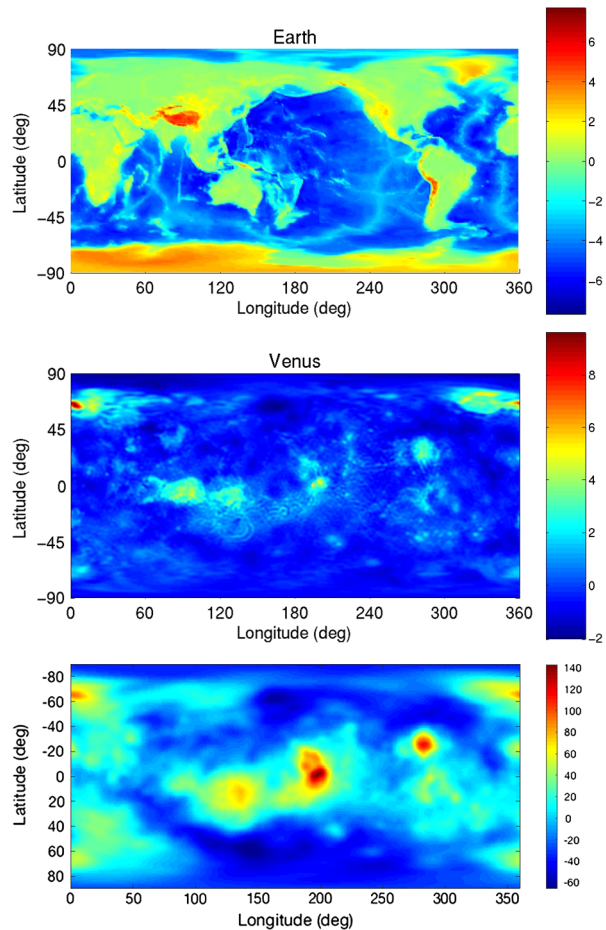
2.2 Gravity and Topography

Gravity and topography data currently provide the best available constraints on Venus' tectonic framework, interior structure and viscosity. The Magellan topography has a horizontal resolution of 10–25 km, and a nominal vertical resolution of 80 m (Pettengill et al. 1992). Vertical resolution is much worse in regions of steep slope or small-scale variations; steep slopes sometimes produce topographic artifacts. Comparing the topography of Venus and Earth at the same resolution reveals that Venus does not have a global pattern of plate boundaries (Fig. 2). On Earth, oceanic spreading centers are conspicuous at the resolution available for Venus. Although not as contiguous as spreading centers, subduction zones also produce topographic features both at the edge of continents and at ocean–ocean subduction zones that extend for 1000s of kilometers.

Hypsometry Earth's bimodal elevation distribution is affected by the combined forces of plate tectonics, isostatic equilibrium differences due to the densities of continental and oceanic crust, and erosion due largely to surface water (Fig. 3). The low elevation slope is due to the formation of new oceanic crust at volcanic ridges that cools and thickens as it spreads away, typically until it reaches a subduction zone and is recycled back into the mantle. Overall Earth's altimetry is modulated by glaciation at the highest elevations and sedimentation in all but deepest ocean basins and trenches Earth's very deep oceanic elevations correspond to subduction trenches far from sediment sources. Continents are higher elevation due to their lower density composition, which also makes them difficult to subduct. As plates collide, continental crust piles up to form mountains, counter-balanced by the forces of erosion and gravitational relaxation. However, plate tectonics with oceans is not the only process capable of creating a binomial hypsometric curve. Mars' hemispheric dichotomy, which has been attributed to degree 1 convection processes (Zhong and Zuber 2001; Roberts and Zhong 2006), or a giant impact in the North (Andrews-Hanna et al. 2008) or in the South (e.g. Reese and Solomatov 2006; Golabek et al. 2011; Leone et al. 2014), also produces a bimodal distribution (Keller and Tackley 2009).

In contrast, Venus has a unimodal topographic distribution. Although tessera plateaus may be analogs of continents, with a lower density crust, they cover only $\sim 8\%$ of the Venus' surface versus 30% continental crust on Earth. The basaltic crust is not depressed to a lower elevation by oceans, and surface water does not erode the topography. The dense atmosphere creates low velocity surface winds and minimal erosion. Lorenz et al. (2011) note that there is much more variation in the magnitude of high elevation on Venus than on Earth (Fig. 3, right), likely due to a lack of erosion on Venus. Conversely the high surface temperature means that gravitational relaxation should be important (Smrekar and Phillips 1991), depending on the water content of the crust (Mackwell et al. 1998).

Fig. 2 Topography of Earth (a) shown at the same resolution as is available for Venus (b). The geoid for Venus is shown in (c)



Gravity Resolution The gravity field reflects interior density variations, although interpretation is non-unique. The gravity field is derived from changes in the Doppler measurement of the communications signal with an orbiting spacecraft. Topography and interior density variations cause the spacecraft motion to speed up or slow down. The gravity field (Fig. 2) derived from tracking of the Magellan spacecraft has a resolution that varies from degree and order 40 to 100, with a mean value 70 (Konopliv et al. 1999). This corresponds to a half-wavelength resolution of 475 to 190 km, with a mean of 272 km. The spherical harmonic field is expanded to degree and order 180 to prevent aliasing, but the global error spectrum shows that on average the error is as large as the gravity field power at degree and order 70. This field is currently being updated using modern computational power and Venus Express data, which may be able to improve resolution locally (Goossens et al. 2017).

Gravity and Topography Correlation The global gravity spectrum is highly correlated with the global topography spectrum. Venus lacks the large impact basins that dominate Mars, Mercury, and the Moon, and the erosion and sedimentation processes that affect Earth, Mars, and Titan. In this regard, its topography and gravity may provide a more direct link to interior processes than on other rocky bodies. Modeling of the gravity and

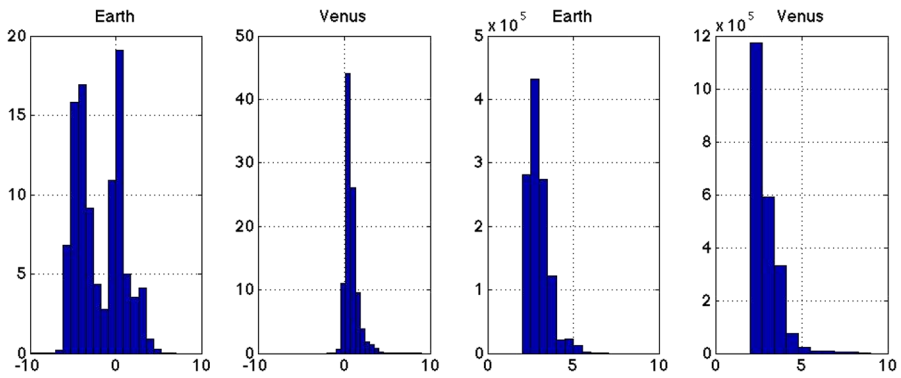


Fig. 3 Comparison of the hypsometry of Earth and Venus, showing the full range of elevations (left), and comparison of the hypsometry just above 2 km elevation, which is less influenced by oceanic erosion

topography constrain first order mantle viscosity and interior convection models at both the local and global scales. Geoid-to-topography ratios for Venusian hotspots are much larger than terrestrial oceanic hotspots. This was interpreted to indicate that unlike terrestrial oceanic lithosphere, there is no underlying low viscosity zone (Kiefer and Hager 1991; Smrekar and Phillips 1991). This result has been used to infer that the interior of Venus is dehydrated.

However, recent studies suggest that water is not the only means of creating Earth's asthenosphere, or low viscosity zone. CO_2 can also produce melting and thus a low viscosity zone (Sifre et al. 2014; Chantel et al. 2016). Other studies have shown that dry rheology can produce a low viscosity zone due to the activation volume (Armann and Tackley 2012). Although the interpretation of large GTRs at hotspots in terms of mantle hydration may be uncertain, it is clear that their signatures are different from hotspots on terrestrial oceanic lithosphere. However, the inference for mantle viscosity at hotspots doesn't necessarily carry over to the rest of the mantle. In the absence of plate motion, melting above a plume will likely change the underlying lower lithosphere and upper mantle, creating mantle residuum and desiccation (e.g. Smrekar and Parmentier 1996). Some studies have attempted to use gravity and topography to constrain mantle convection with partial success (Pauer et al. 2006; Steinberger et al. 2010; Orth and Solomatov 2011; Armann and Tackley 2012; Huang et al. 2013).

Gravity and Topography of Highlands The major topographic highs on Venus all have associated positive geoid anomalies (Fig. 2). The large volcanic rises, or hotspots, have broad topographic rises (>1000 km diameter), large positive gravity anomalies, one or more large volcanoes, and sometimes rifts. The gravity and topography at the 10 hotspots regions have been interpreted to indicate the presence of active mantle plumes at depths of ~ 100 – 200 km (Kiefer and Hager 1991; Smrekar and Phillips 1991). Although the number of active plumes on Earth is debated, many authors find a similar number (e.g. Courtillot et al. 2003). The presence of plumes rising from the core-mantle boundary provides constraints on mantle dynamics (e.g. Smrekar and Sotin 2012).

Venus' tessera plateaus have been proposed to form above mantle upwellings (Grimm and Phillips 1991; Phillips and Hansen 1994), downwellings (Bindschadler et al. 1992), as well as to be analogs of terrestrial continents (e.g. Gilmore et al. 2017). The six tessera plateaus are ~ 1000 – 1500 km in diameter, and have multiple, intersecting tectonic deformation trends. Analysis of the gravity and topography of these features shows that they

have shallow compensation depths, interpreted as isostatic compensation by crustal roots (Smrekar and Phillips 1991). Crustal thickening can be accomplished via voluminous partial melting above a plume, as in the upwelling hypothesis, or via crustal shortening, in the downwelling hypothesis. Both models make predictions for surface deformation, which cannot be fully distinguished with current data. If the crust of Venus is dehydrated and strong, the downwelling model is implausible (Lenardic et al. 1995). In an alternative model of tesserae as continental analogs, hydrated, basaltic crust remelts as it is recycled into the interior via downwelling, delamination, or subduction (Campbell and Taylor 1983; Hoening and Spohn 2016). Although felsic rocks can be produced without water, the production of the huge volume of felsic crust found in Earth's continents, and potentially in tesserae on Venus, requires melting of basaltic crust in the presence of water. Thus tesserae have been proposed to be remnants of a past wetter era on Venus (Gilmore et al. 2017).

Data from Venus Express corroborate the presence of active plumes at hotspots and felsic crust at tesserae. Surface brightness temperatures from the VIRTIS spectrometer corrected for atmospheric effects and elevation-dependent temperature changes provide a map of near-infrared surface emissivity over much of the southern hemisphere (Mueller et al. 2008). The difference between the high emissivity anomalies and moderate background emissivity is consistent with the expected compositional anomalies between unweathered and weathered basalts. Thus these areas can be interpreted as geologically recent volcanism, supporting the interpretation of hotspots as sites of active mantle plumes (Smrekar et al. 2010; Davaille et al. 2017). Additionally, the one tesserae plateau in the southern hemisphere has an emissivity signature consistent with felsic crust (Gilmore et al. 2015). See Gilmore et al. (2017) for further discussion of emissivity data interpretation.

2.3 Tectonics

Strain Estimates and Tectonic Surface Features Even though plate tectonics is not operative on Venus today, its surface shows a wide range of types and scales of deformation features, including lineaments, graben, rifts, polygonal terrain, wrinkle ridges, and deformation belts (see Solomon et al. 1992 and McGill et al. 2009 for reviews). The intensely deformed tesserae terrains are described below. Strain can be difficult to estimate from topography and imaging data alone, but it is possible to estimate lower bounds. Sandwell et al. (1997) estimated the present-day global strain field, inverting the geoid height and assuming that Venus lithosphere behaved as a uniform thickness thin elastic shell over an inviscid sphere. Globally, strain amplitude remains lower than 0.5%, and the model strain field map compare quite well with Magellan maps of wrinkle ridges (i.e. compression) and rift zones (i.e. extension), the former corresponding to areas where the geoid amplitude is lower than 20 m. So much of the observed deformation matches the present-day model strain orientations. This suggests that most of the rifts and wrinkle ridges formed in a stress field similar to the present one. At smaller scale, observations of strike-slip faults with up to 450 km of displacement (e.g. Harris and Bedard 2015) and a 10-km wide rift that bisects the Somerville impact crater (Solomon et al. 1992) provide direct evidence of strain. Strain estimates are also derived from modeling of deformation features, such as the 1–7% strain evaluated at radially fractured centers (Grinrod et al. 2005) and the 0.5–5% regional shortening at wrinkle ridges (Bilotti and Suppe 1999; Banerdt et al. 1997).

Upwelling and Delamination A number of other geologic features can also be directly linked to interior processes. In addition to hotspots, Venus also has about 500 quasicircular features proposed to form via small scale upwelling, delamination, melting, or some

combination of these processes (Schubert and Sandwell 1995; Koch and Manga 1996; Smrekar and Stofan 1997; Hoogenboom and Houseman 2006; Piskorz et al. 2014). They are called “coronae” (Latin for crown), and are defined by full or partial annuli of fractures (Stofan et al. 1991). The mean diameter of the ~500 coronae is 250 km (Glaze et al. 2002). The outlier is Artemis Corona, with a diameter of 2600 km. Coronae always have associated volcanism. Their very diverse topographic shapes fall into nine topographic classes (Smrekar and Stofan 1999). They are approximately evenly split into those with interior topographic highs and lows, and sometimes have surrounding rims and troughs.

Another 65 “novae” have been identified (Krassilnikov and Head 2003). About 100–300 km in diameter, they are characterized by dense radial fracturing, usually issuing from a topographic dome. Some of them also have an annulus of fractures, probably emplaced at a later stage. It has been proposed that dike swarm emplacement was the dominant process in novae fracturing formation, whereby a magma chamber in the lower ductile crust inflates and generates radial dikes below the neutral buoyancy level (Krassilnikov and Head 2003). 3D numerical modeling further suggested that novae could result from plume-induced convection of the thick and soft Venusian crust, and may constitute the first stage of coronae formation (Gerya 2014), although only a subset of coronae have radial fractures.

Subduction One of the key questions for Venus is why does it lack plate tectonics, and if it always lacked plate tectonics. Although current data shows no evidence of distinct plates, subduction is proposed to occur on Venus. There are ~10 locations that have the topographic morphology, fracture patterns, and estimated mechanical properties observed at terrestrial subduction zones (McKenzie et al. 1992; Sandwell and Schubert 1992b). However, almost all of the regions also have the characteristics of an upwelling mantle plume, which has been used to argue against the subduction interpretation (Hansen and Phillips 1993). Specifically, these regions have a quasi-circular planform defined by a topographic high, an annulus of fractures, volcanism, and, in some cases, a gravity high consistent with a mantle plume. The hypothesis of plume-induced subduction (Sandwell and Schubert 1992b, Schubert and Sandwell 1995) reconciles the characteristics of subduction and plumes. Recent experimental and numerical modeling (Ueda et al. 2008; Davaille and Smrekar 2014; Gerya et al. 2015; Cramer and Tackley 2016; Davaille et al. 2017) of plume-induced subduction further explains the presence of these features on Venus today and the potential for this mechanism to operate on early Earth.

2.4 Crust and Elastic Lithosphere

Crustal Composition and Thickness The majority of the crust on Venus is believed to be basaltic, which provides a reasonable constraint on crustal density. The transformation of basalt to eclogite may limit the thickness of the crust at a depth of roughly 50–70 km (e.g. Turcotte 1989), although the transition to eclogite may be limited by temperature and the specific composition of the crust. The specific depth is a function of the thermal gradient. The rate at which the transformation occurs is poorly constrained, and thus this constraint cannot be considered a firm upper bound on crustal thickness. As noted above, 8% of the surface is covered by tessera terrain, which may be felsic in composition and may imply remelting of basalt in the presence of water. Gilmore et al. (2017) review available constraints on crustal composition.

Modeling of gravity and topography also provides an estimate of global crustal thickness, an important lower bound on the amount of melting over the planet’s history. James et al. (2013) performed a global analysis aimed at putting bounds on mean crustal thickness.

They calculated Geoid to Topography Ratios (GTRs) using regions with three different diameters and a model with a crustal layer and dynamic compensation at depth. This approach requires a number of assumptions that are applied globally: the elastic thickness is 20 km, the thermal gradient is 5–10 °C/km, and the mantle is isoviscous. They also assume that the crustal thickness is non-zero everywhere, and that the basalt eclogite transition limits the crustal thickness locally to 70 km. With these constraints, they estimate a mean crustal thickness in the range of 8–25 km. If the eclogite constraint the upper bound is removed, the mean can be up to 45 km. As shown by James et al. (2013) if the thermal gradient is 20 °C/km, no eclogite transition is expected. Jiménez-Díaz et al. (2015) follow the method of Forsyth (1985) to assess global crustal and thickness over broad regions. They estimate crustal thickness to be 20–25 km over much of the planet, exclusive of highlands. Anderson and Smrekar (2006) also estimated crustal thickness via a global analysis of gravity and topography using localized, spectral admittance methods. They find a wide range of values, up to 100 km, with a majority of the planet having values less than 50 km. One issue is that the gravity field resolution is below degree and order 70 for ~50% of the planet. In these regions, the crustal thickness models may have significant errors. Nonetheless, these different methods produce broadly similar estimates of crustal thickness in most regions. In those area with significant different estimates, such as Beta and Eistla Regiones, there are likely to be processes such as dynamic support above a plume head, mantle residuum, and variable elastic thickness. Given uncertainties in the thermal gradient and crustal layering, the thickness of the crust can be considered to be known to within a factor of 2–3.

Elastic Thickness Global estimates of elastic thickness from gravity and topography find a wide range of values, from 0–100 km (Anderson and Smrekar 2006; Jiménez-Díaz et al. 2015) Large values occur where models of bottom-loading provide the best fit; bottom-loading fits are known to be biased towards larger values. Over half of the planet has best fit values of 20 km or less, which cannot be distinguished from isostatic compensation. Thus either these regions have quite high heat flow, or, if they are isostatically compensated, the elastic thickness estimate is invalid. Given the variable geologic settings with apparently thin elastic thickness, it is likely that different interpretations apply in different areas. Elastic thicknesses derived from modeling of topographic flexure also show a range of values from effectively 0 km up to 70 km (e.g., Brown and Grimm 1996; Johnson and Sandwell 1994; Sandwell and Schubert 1992a; O'Rourke and Smrekar 2018).

2.5 Magnetic Field

The diameter of Venus is 5% smaller than that of Earth. Thus for the same mantle composition the Venusian core radius would be 0.51 of its planetary radius as compared with 0.55 of its planetary radius for Earth. When corrected for pressure, Venus' density is only 3% less than Earth's (Phillips and Malin 1983). Venus is the only terrestrial planet that lacks evidence for a magnetic field. The upper bound provided by the Pioneer Venus spacecraft is 10^5 lower than Earth's current field, or ~0.3 nTat 150 km (Phillips and Russell 1987). Analysis of Venus Express data suggest that Venus could have a weak dipole up to an order of magnitude larger than the Pioneer Venus derived estimate, though other interpretations are possible (Luhmann et al. 2015). But the Mariner 2 spacecraft made it clear that Venus does not have an Earth-like magnetosphere. The apparent lack of a dynamo is puzzling given evidence of a metallic core. Unlike Earth, Venus lacks a significant hydrostatic bulge. Thus the available gravity field provides only modest constraints on the core size and state. Konopliv and Yoder (1996) estimated the k_2 love number to be 0.295 ± 0.066 . Using an elastic

model of tidal deformation, they find that this value is consistent with a completely or at least partially liquid core. Dumoulin et al. (2017) computed a new estimate of the Venusian k_2 using thermodynamically derived densities and a viscoelastic model. They find that the current k_2 estimate cannot completely rule out a fully solid metal core.

3 Mantle Dynamics

Given the size of Venus, the comparison with the Earth strongly suggests that convection processes are present to extract the primordial heat that was stored during accretion and the internal heat produced by the decay of the long-lived radioactive elements. One indication of the current activity and planform of convection on Venus comes from evidence for surface deformation above upwelling mantle plumes, based on analysis of Magellan imaging, gravity and topography data (Sect. 2). New evidence for very recent volcanism at several of these hotspots (Smrekar et al. 2010) implies pressure release melting is occurring. Such features are therefore analogous to areas like Hawaii on Earth. On the other hand, although Venus surface seems uniformly young, no system of Earth's like Plate Tectonics exists today. This limits the amount of heat that can be transferred.

So how has Venus extracted its internal heat? Is the young age of its surface due to episodic catastrophic resurfacing or to a more continuous process? As we are waiting for new observations to constrain models of Venus evolution, the next sections describe how theoretical studies, laboratory experiments and numerical simulations can help us understand these processes.

3.1 Convective Patterns

The vigor of convection can be characterized by the Rayleigh Number Ra . It is the ratio between the buoyancy force generated by the deep hot mantle being less dense than the shallow cold mantle, and the friction due to the material viscosity. It is expressed by

$$Ra = \frac{\alpha \rho g \Delta T d^3}{\kappa \eta} \quad (1)$$

where α is the coefficient of thermal expansion, ρ is the mean mantle density, g is the gravitational constant, ΔT is the temperature difference across the entire mantle, d is the thickness of the convective mantle, η is the viscosity, and κ is the coefficient of thermal diffusivity. Additional key factors in accurately predicting the planform and number of upwellings are the mantle geometry, the phase transitions, the amount of internal heating, the presence of density heterogeneities, and the viscosity range (which is dominated by the temperature dependence).

Both laboratory studies (Davaille and Jaupart 1993) and numerical studies (Moresi and Solomatov 1995) pointed out the role of strongly variable viscosity in reducing the efficiency of convection in the absence of plate tectonics because a conductive stagnant lid lays on top of the convective mantle. The viscosity of the material that constitutes the mantle is very much variable, with temperature being one of the major parameters responsible for its large variations. Typically, the viscosity of materials depends on temperature (T), shear stress (τ), and grain size (d). It also depends on the amount of water contained in the silicates: dry silicates have higher values of viscosity than wet silicates, although recent studies show a variation limited to 30% at most (i.e. Demouchy and Bolfan-Casanova 2016). Viscosity is a

parameter that links shear stress (τ) to strain rate ($\partial\varepsilon/\partial t$). The strain rate of solids has been studied experimentally and a general expression of the strain rate is:

$$\dot{\varepsilon} = \frac{\partial\varepsilon}{\partial t} = \frac{A\tau^n}{d^m} e^{-\frac{Q}{RT}} \quad (2)$$

where Q is the enthalpy that contains a pressure term, and n and m are stress and grain size exponents, respectively. The values of these three parameters depend on the processes involved in the deformation of silicates. Those processes include diffusion, migrations of dislocations, grain boundary sliding. The viscosity can then be expressed as:

$$\eta = \frac{\tau}{\dot{\varepsilon}} = \frac{\tau^{1-n} d^m}{A} e^{\frac{Q}{RT}} \quad (3)$$

Note that the viscosity is independent of stress (Newtonian behavior) when the stress exponent is equal to 1. The value of the activation energy is in the range 420–540 kJ/mol (Karato and Wu 1993), which leads to a viscosity increase between the interior (temperatures on the order of 1600 K) and the surface ($T_s = 735$ K for Venus) of more than 10 orders of magnitude. In a Newtonian, strongly-temperature-dependent viscosity fluid, the surface-interior viscosity contrast is one the two key-parameters to describe the different convective regimes (Fig. 4).

Depending on the Rayleigh number and viscosity contrast, three different convective regimes are encountered (Fig. 4): a nearly constant viscosity case where the viscosity contrast is so small that hot and cold convective instabilities have nearly the same characteristics, a “mobile lid” regime for intermediate viscosity contrasts where the upper thermal boundary layer becomes stiffer but can still move laterally and sink, and the “stagnant lid” regime, where convective motions occur below a sublayer that remains stagnant (because it is too viscous). Heat extraction at the surface occurs via conduction only in the stagnant lid regime, and is therefore much less efficient than for the other two regimes. So for a given surface temperature, increasing the viscosity ratio leads to increasing the mantle temperature and diminishing the viscosity. Taking into account the sphericity of the planet or the amount of internal heating does not change the overall regime diagram (Fig. 4), but both ingredients increase the mantle temperature compared to the Cartesian case without internal heating (Ratcliff et al. 1996, 1997; Sotin and Labrosse 1999; Zhong et al. 2008; Limare et al. 2015). Note that the “episodic lid” regime is not recovered in Fig. 4. It would correspond to a system that oscillates between the mobile and stagnant modes. This intermediate regime appears if the lithosphere rheology presents a yield stress (Fig. 5; Sect. 3.3.ii).

In the framework of Fig. 4, the high viscosity ratio between the surface and interior of Venus and Earth place both mantles within the dash box. In this “stagnant lid regime” the conductive lid is contained within the thermal lithosphere. However, either plate tectonics or a global resurfacing event involve a mobile lithosphere and should plot (even momentarily) in the “mobile lid” regime (pink area in Fig. 4) with intermediate viscosity ratios. So an ingredient is missing from this simple convective modeling. We shall come back to this point in Sect. 3.3.

3.2 Existence, Number and Characteristics of Venusian Hot Mantle Plumes

Hot plumes develop from a hot boundary layer, and therefore indicate that a substantial part of the mantle is heated from below. Either the core is hotter than the mantle, or there is an interface in the mantle hosting a strong temperature gradient. If the mantle temperature becomes too high, there will be no bottom thermal boundary layer. The existence and

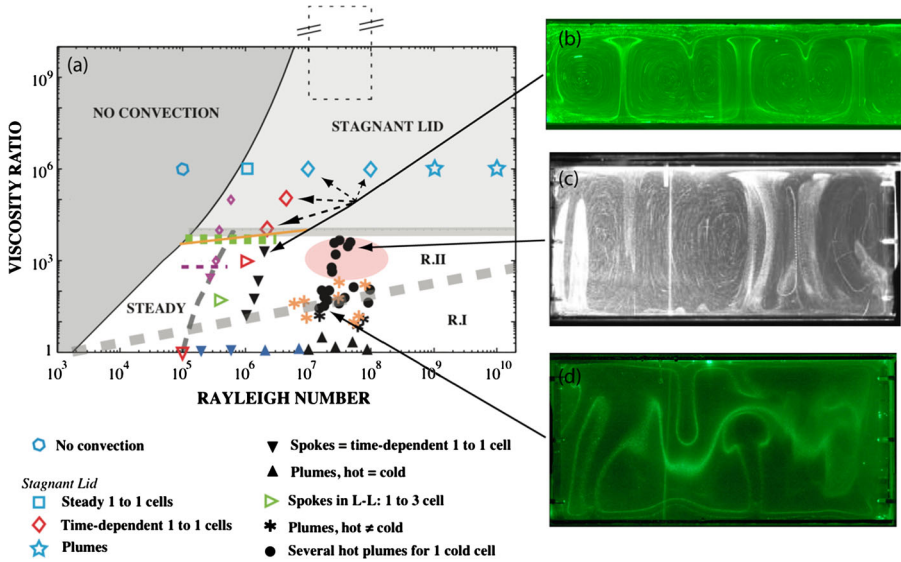


Fig. 4 Convective pattern as a function of the Rayleigh number at the bottom boundary and the viscosity ratio for 3D experiments in a Cartesian box. (a) Regime diagram. Open symbols designate experiments run with free surfaces (mostly numerical simulations) and solid symbols designate experiments run with fixed boundaries. Black: Androvandi et al. (2011); light blue: Trompert and Hansen (1998); red: Tackley et al. (1994), Tackley (1996) and Ratcliff et al. (1997); green: Weinstein and Christensen (1991); dark blue: Guillou and Jaupart (1995); orange: Schaeffer and Manga (2001). The dashed grey border between the steady and time-dependent whole layer regimes is after White (1988). The thick dashed grey line shows the theoretical limit of Solomatov (1995) between the low-viscosity ratio regime ('R.I') and the 'sluggish-lid' regime ('R.II'). The thick solid grey line between the whole layer ('R.II') and the stagnant lid regimes is after Solomatov (1995); the orange line is from the steady state calculations of Ogawa et al. (1991); the dashed thick purple line from Nataf and Richter (1982) and Richter et al. (1983); and the green thick line after Giannandrea and Christensen (1993). The limit between no convection and convection is drawn after Stengel et al. (1982). (b), (c) and (d) show the convective patterns encountered in the different regimes as seen in the laboratory. The bright lines are isotherms. Thermal boundary layers are the zones with high temperature gradients near the hot and cold horizontal boundaries. In (a), the pink area designates the parameter range where the two scales of convection (cold cells encasing several hot plumes) are reminiscent of the present-day Earth's mantle convective pattern, while the thin black dashed line shows the actual parameter range for the Earth's mantle

number of hot plumes on Venus today therefore puts some bounds on the amount of internal heating (Smrekar and Sotin 2012). Depth-dependent viscosity and an endothermic phase transition in the mantle also influence the wavelength of convection and therefore the number of plumes (Huang et al. 2013).

In addition, as convection under a stagnant lid generates only thin active boundary layers where the viscosity ratio is less than 10 (Davaille and Jaupart 1993; Solomatov 1995; Jellinek et al. 2002), hot instabilities are transient features with a roundish head but no thin and long-lived tails (Jellinek et al. 2002), contrary to the plumes (with big heads on a long-lived thin conduits) that develop when a viscosity ratio greater than 100 develops across the bottom hot thermal boundary layer (e.g. Whitehead and Luther 1975). Jellinek et al. (2002) therefore proposed that the absence of plate tectonics on Venus may lead to upwellings composed of thermals, the surface expression of which are coronae, as well as a smaller core heat flux than on Earth. Moreover, experiments run into the intermediate viscosity ratio range showed that several scales of motions could coexist and interact (Weeraratne

and Manga 1998; Manga et al. 2001; Lithgow-Bertelloni et al. 2001; Jellinek et al. 2003; Androvandi et al. 2011). Specifically, cold less viscous downwellings have spacing and characteristics periodicities systematically greater than hot less viscous upwellings. The latter also tend to be excluded from the downwelling regions and even pushed together to create the upwelling limb of a large-scale circulation forced by the downwellings (e.g. Jellinek et al. 2003). This could explain the relative paucity of coronae in the broad lowland plains if the latter are above mantle downwellings, and the concentration of upwellings and hence coronae in the adjacent Beta, Atla and Themis Regiones (Johnson and Richards 2003).

Last, it was shown that several scales of convection and several types of plumes could develop in a heterogeneous mantle and lead to the coexistence of large plumes coming from the core-mantle boundary together with smaller plumes developing in the upper mantle on top of thermochemical domes located in the lower mantle (Davaille 1999; Courtillot et al. 2003; Davaille et al. 2005). These different types of plumes could generate the large size distribution of coronae (50 to 2600 km) observed on Venus as they interact with the lithosphere (Stofan and Smrekar 2005).

3.3 Causes for Venus Resurfacing

We have seen (Sect. 2) that two end member models exist for Venus resurfacing: either a catastrophic resurfacing between 0.3 and 1 Gyr ago, or a more continuous resurfacing where random patches of Venus surface are covered by new material. Several mechanisms can be at play to produce resurfacing.

(i) *Mantle overturns*

Phase changes can modulate the pattern and wavelength of the convective motions. Associated with a viscosity increase, it will increase the wavelength of convection. Moreover, an endothermic phase transition could stratify mantle convection episodically or even permanently, especially at high Ra (e.g. Machel and Weber 1991). As the planet cools and Ra decreases, a transition from layered to whole mantle convection caused by the endothermic spinel to perovskite phase transition was suggested to have caused Venus global resurfacing event (Steinbach and Yuen 1992). The interaction of phase-change induced avalanches with non-Newtonian lithospheric rheology can also cause lithospheric overturn events (Weinstein 1996). However, these studies were done in 2D. Simulations in 3D spherical geometry suggest that the endothermic phase transition alone is not sufficient to produce catastrophic resurfacing (Tackley et al. 1994; Bunge et al. 1997; Huang et al. 2013; Yang et al. 2015).

On the other hand, Parmentier and Hess (1992) proposed that complete resurfacing of Venus could be due to episodic overturn in a layered mantle convection system due to competition between thermal and compositional buoyancy. This system was subsequently systematically studied, where initially two layers of miscible fluids with different densities, depths, and viscosities are subjected to a destabilizing temperature contrast (Davaille 1999; Le Bars and Davaille 2002, 2004a, 2004b). When the buoyancy number B (i.e., the ratio of the stabilizing chemical density anomaly to the destabilizing thermal density anomaly) is lower than 0.3–0.5, an episodic regime can develop where either (1) the interface between the two layers deforms in large domes moving up and down quasi-periodically, or (2) the two layers periodically exchange places. Global resurfacing of Venus could then correspond to an abrupt draining of the lower or of the upper reservoir. Scaling laws predict that for a less viscous (since hotter) lower reservoir, the period is controlled by the more viscous upper layer and a minimum periodicity of 420 m.y. is obtained; it can become much longer for a more viscous lower reservoir (Le Bars and Davaille 2004b).

(ii) Lithosphere breaking and foundering

Any lithospheric resurfacing requires breaking the lid, and therefore a rheology which strongly localizes deformation. This is obtained considering plasticity, where the lithosphere yields when the deviatoric stress locally exceeds the lithosphere strength σ_Y . In that case, Fowler and O'Brien (1996) presented a mathematical analysis showing that episodic lithospheric overturn could result. Such episodic overturn events were subsequently obtained in numerical convection models with a strongly temperature-dependent, yielding rheology (Fig. 5), in 2-D (Moresi and Solomatov 1998; Armann and Tackley 2012) and in 3-D (Trompert and Hansen 1998; Tackley 2000; Loddoch et al. 2006; Stein et al. 2004, 2010; Weller et al. 2015). However, numerical simulations show that instantaneous stresses arising from thermal convection alone are up to 100 MPa only, which is smaller than the laboratory measurements on natural samples, which give maximum lithospheric strength around 300 to 500 MPa.

Hence, one has to consider additional processes to decrease the lithosphere strength, such as shear-heating coupled to buckling (Crameri and Kaus 2010) or to grain-size reduction (Thielmann et al. 2015), long-time development of damage zones due to polycrystal dynamics (Bercovici and Ricard 2014), contraction cracks in the lithosphere (Korenaga 2007), and weak layer interconnection (Montesi 2013). Most mechanisms show a strong influence of surface temperature on the localization and the mobility of the lithosphere, a higher temperature usually does not favor the survival of plate boundaries (Crameri and Kaus 2010; Lenardic et al. 2008; Landuyt and Bercovici 2009; Noack et al. 2012; Lenardic and Crowley 2012; Bercovici and Ricard 2014). So the absence of continuous plate tectonics on Venus could result from its much higher surface temperature than Earth. This effect is also certainly enhanced by the lack of water on Venus' surface. On Earth, water has a key-role in forming weak hydrated minerals such as serpentine (e.g. Hilairet et al. 2007) and reducing fault friction through pore pressure (e.g. Dymkova and Gerya 2013). The existence of a weak channel between the descending slab and the overriding lithosphere has been shown to facilitate the stability and longevity of subduction (e.g. Crameri et al. 2012). So the conjunction of a high surface temperature and the absence of surface water on Venus could render subduction more difficult to nucleate and less likely to continue for a long time than on Earth. On the other hand, other fluids, such as melt or supercritical CO₂, might help but their effects need to be quantified.

The heat deposited in the mantle by a large impactor (>400 km in diameter) can also have long term effects on mantle convection (Monteux et al. 2007; Gillmann et al. 2016; O'Neill et al. 2017). As the hot material spreads under the surface, it thickens the lithosphere near the edges, promoting the development of downwelling drips and eventually subduction when the local lithospheric strength is reached (Gillmann et al. 2016; O'Neill et al. 2017).

The same sequence of processes is observed when a hot plume impinges under the lithosphere in some numerical models (Gerya et al. 2015; Crameri and Tackley 2016). It can be accelerated when the lithosphere rheology above the incoming plume is artificially softened to mimic melt intrusion: the lithosphere viscosity strongly decreases until it becomes low enough to generate a cold dripping annular sheet around the hot plume (Gerya et al. 2015). In this case, the onset of subduction occurs along a complete ring on the edge of the hot plume. Plumes with an additional chemical buoyancy (i.e. rich in volatiles) could also directly overcome the lithosphere yield stress, as seen in 2D simulations (Ueda et al. 2008). Then, the hot and less dense plume material spreads on the broken lithosphere, which sinks, inducing subduction. 3D laboratory experiments show then that the initiation of subduction in a visco-elasto-brittle lithosphere does not occur along an annulus but along radial rifts (Davaille and Smrekar 2014;

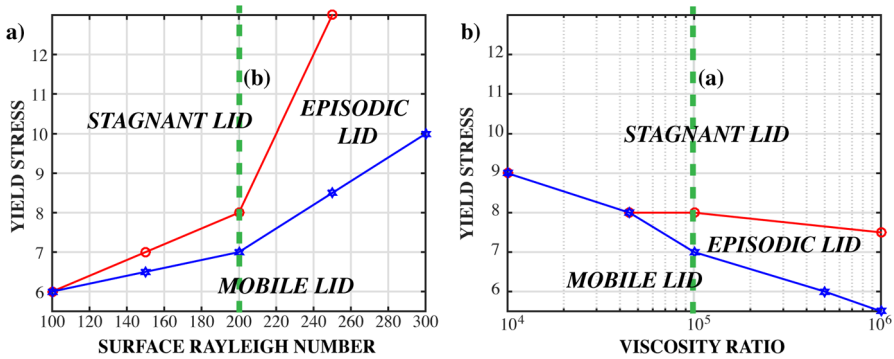


Fig. 5 Different convective regimes in a strongly temperature-dependent fluid which also present a yield stress in the lithosphere (see text). The surface Rayleigh number is calculated with the viscosity at the surface of the system η_0 , and the yield stress has been normalized by $\eta_0\kappa/d^2$. (a) Regimes obtained for a constant viscosity ratio of 10^5 (also indicated on (b) by the grid line); (b) regimes obtained for a constant surface Ra of 200 (also indicated on (a) by the grid line). Adapted from Stein et al. (2004)

Davaille et al. 2017). A rising plume would produce tensile fractures in the experimental lithosphere that forms on the upper fluid surface, and buoyant plume material would upwell through the fractures and spread above the skin, analogous to volcanic flooding. This leads to bending and eventual subduction of the skin along several separate arcuate segments. These segments, as well as the deformations patterns on the surface and the subsurface density variations, would explain well the observations on two large coronae, Quetzalpetlatl and Artemis, two proposed sites of plume-triggered subduction (Sandwell and Schubert 1992a; Schubert and Sandwell 1995). However, in the experiments reported, such events were not precursors of a complete rejuvenation of the experimental surface as was mostly observed in 2D (Armann and Tackley 2012) or 3D (Yoshida and Ogawa 2004; Gerya et al. 2015; Cramer and Tackley 2016) numerical simulations. This discrepancy could result from the 3D geometry and the more complex rheology used in the experiments. However, it shows that this localized-subduction regime is not only different from continuous resurfacing in a mobile lid (e.g. plate tectonics) but also different from the episodic catastrophic and complete resurfacing of a stagnant lid, as had been proposed for Venus (Fowler and O’Brien 1996; Moresi and Solomatov 1998). This new regime is more consistent with models that predict ongoing localized resurfacing of Venus (Bjornnes et al. 2012; O’Rourke et al. 2014).

Compositional variations caused by partial melting and crustal formation could also play a major role in lithosphere stability. In a pioneering study, Dupeyrat and Sotin (1995) observed that buoyant depleted mantle tends to accumulate below the lithosphere and suppress downwellings, while dense eclogite promotes recycling of the basaltic crust and limits crustal thickness to the depth of the basalt-eclogite transition. This study was done with a constant viscosity, but the results hold with a plastic rheology (e.g. Johnson et al. 2014), although a purely temperature-dependent viscosity would stabilize the base of the crust.

(iii) Coupled models

Ogawa (2000) showed that episodicity can also be produced in a more complex model coupling magmatism, strongly temperature-dependent viscosity and the density cross-over between basalt and harzburgite below the spinel to perovskite transition. There, basalt is less dense for several tens of km at the top of the lower mantle because the transition to perovskite occurs at greater pressure in the basalt mineralogy (Ringwood 1991), while it

is relatively dense throughout most of the mantle. So substantial crustal recycling can occur despite a high-viscosity lid, with the recycled crust building up above the core-mantle boundary (CMB) and at the base of the transition zone. The episodic breakdown of local chemical layering around 660 km depth caused by the composition-dependent phase transition resulted in massive “flushing events” causing vigorous magmatic pulses that might account for volcanic plain formation on Venus, and even global resurfacing events. This process remains dominant even with a moderately mobile lithosphere (Papuc and Davies 2012) or a lithosphere plastic rheology (Armann and Tackley 2012). The episodicity time-scale depends on the lithosphere rheology and the chemical barrier amplitude, with a stronger lithosphere and a weaker barrier promoting transition to a stagnant lid regime (Ogawa and Yanagisawa 2014).

(iv) Interaction between mantle convection and a deformable crust

Even though Venus appears to be presently in a kind of stagnant-lid regime, its surface presents much larger and more complex deformation patterns than for example Mercury or Mars, which are considered as archetypes of a stagnant-lid planet. For example, Harris and Bedard (2015) propose the existence of a northward drift and indentation of Lakshmi Planum in western Ishtar Terra. They interpret the large displacement of this area of continent-like crust on Venus as resulting from mantle plume traction and pressure acting against its deep lithospheric mantle « keel » in analogy with the northward displacement of India under the action of La Reunion mantle plume (Cande and Stegman 2011).

On the other hand, all the above assumes that the crust stays strongly coupled to the lithospheric mantle. Venus' crust is estimated to be on average about 30 km. Given the high surface temperature, the lower crust is expected to be much softer than the lithospheric mantle underneath, producing a strong decoupling (Buck 1992; Azuma et al. 2014; Ghail 2015). The model proposed by Buck (1992) assumed a heat flow on Venus scaled from that of Earth (70 mW/m^2) and mid ocean ridge-like processes of formation, spreading, and subduction of plates, all occurring beneath a weak crust decoupled from the mantle. Assuming a 20 km-thick, anorthosite crust, Buck (1992) showed that a lower crust viscosity of 10^{18} Pa s would be sufficient to decouple it from the mantle underneath. This viscosity value is comparable to the “sticky air” viscosity value employed in numerical models to generate one-sided subduction (e.g. Cramer et al. 2012). Although the crust is too buoyant to be entrained in the mantle, it could still deform and flow due to the pressure gradients related to surface topography and the mantle-crust relative motions (Buck 1992). However surface observations do not provide a good match for this model, which predicts that the lowlands are sites of mantle upwelling and numerous subducting slabs that should be observable in the gravity data.

Decoupling of the crust and mantle have different results depending on whether the lithospheric mantle is assumed to be strong (brittle) or weak (ductile). Decoupling due to ductile, lower crustal flow between the brittle upper crust and brittle lithospheric mantle effects surface topography, as described for Venus by Buck (1992). Ghail (2015) advocates for decoupling of the brittle crust above both a weak lower crust and a weak lithospheric mantle to allow for sufficient delamination to produce significant heat loss. These models have also been explored for the Earth as well, taking the form of the “crème brûlée” model versus the “jelly sandwich” model (Burov and Watts 2006). Oceanic crust is generally too thin to be ductile. The thicker, felsic continental crust is predicted to be ductile at its base in most locations. If the mantle lithosphere underlying the continental crust is brittle, a “jelly sandwich” forms, with two brittle layers separated by a weak ductile layer. In the “crème brûlée” model there is no strong mantle lithosphere layer. Determining the applicability of this rheological structure in a given region requires estimation of the thicknesses of the

crust and elastic layers and the rheology of the crust and mantle, as well strain rate. On Earth, these data are most frequently consistent with a weak lower crust and strong mantle lithosphere (“jelly sandwich”) and occasionally consistent with no strong lithospheric layer (“crème brûlée”) (Burov and Watts 2006). There is evidence of lower crustal flow locally on Venus (Banerdt and Golombek 1988; Zuber and Parmentier 1995; Smrekar and Phillips 1991; Karimi and Dombard 2017), consistent with a strong mantle lithosphere. Both large values of elastic thickness and the presence of high plateaus and mountain belts (up to 12 km above the mean) argues for a strong mantle lithosphere layer, as on Earth (Burov and Watts 2006). However the range of crustal and elastic thicknesses on Venus suggest that the rheological structure is also variable. Thus decoupling between the mantle and crust may occur regionally.

3.4 Venus Mantle Evolution

The evolution of a planetary mantle is controlled by the energy equation

$$MC_p dT_m/dt = Q_{\text{top}} + Q_{\text{melt}} + H_R + Q_{\text{core}}$$

where M is the mantle mass, C_p the heat capacity, T_m its potential temperature, Q_{top} is the heat flux out of the planet, Q_{core} the heat flux out of the core, H_R the internal heating by radioactive elements and Q_{melt} the heat flux produced by melting and crust production.

Q_{top} is the most important term and it is controlled by the mantle convective regime, and the mobility of the surface plate. As we already saw, those in turn depend strongly on the surface temperature and the presence of water, which are determined by the interactions between the mantle and its atmosphere.

Depending of the models, the atmosphere in turn is mostly or totally constituted by exsolution of the volatiles stored in the liquid mantle at the end of accretion (so called “Magma Ocean stage”) and progressively released as the magma ocean solidifies. And the surface temperature of the planet depends on the distance to the Sun, but also on the atmospheric content and the presence or not of a greenhouse.

The D/H ratio implies that Venus has lost substantial quantities of water over its history, but when and how is still unclear. Two end-member scenarios have been proposed. Models considering a dense H₂O atmosphere in presence of a hydrodynamic escape of H twice as large on Venus as on the Earth due to its proximity to the Sun suggest that the H escape could have been sufficient to dry the magma ocean before its complete solidification and that therefore Venus solid-mantle stage would have been born dry (e.g. Gillmann et al. 2009; Hamano et al. 2013). On the other hand, models considering a CO₂–H₂O atmosphere suggest the possibility of water ocean condensation on Venus at the end of a rapid (less than a few Myr) magma ocean stage, especially if clouds or Venus rotation are considered (e.g. Kasting 1988; Way et al. 2016; Salvador et al. 2017).

Then, the convective regime in the solid mantle stage depends very strongly whether its surface is mobile or not, as already discussed. A stress-free surface allows mantle material to approach the surface and cool efficiently, whereas a rigid surface hampers heat loss from the planet, resulting in a hot planet even when the internal heat sources are reduced. An endothermic phase change inside the mantle does not help (Armann and Tackley 2012; Yang et al. 2015). In the stagnant lid regime, an Earth-sized planet struggles to lose its heat and the maximum heat flux is about 20 mW/m², which is not enough to extract the radioactive heat produced by the long-lives radiogenic elements likely to be in similar abundance as in the Earth’s mantle (e.g. Reese et al. 1998). Scaling laws linking the heat transfer to

internal heating and mantle temperature suggest that this would favored high mantle temperatures and therefore wide-spread partial melting under Venus lithosphere (Limare et al. 2015). If widespread melting occurs, a magmatic “heat pipe” mechanism (e.g. O’Reilly and Davies 1981; Stevenson and McNamara 1988; Moore et al. 2017; Moore and Webb 2013) could happen and has been considered for Venus (Spohn 1991; Van Thienen et al. 2005; Armann and Tackley 2012). Actually, melting has been found quite efficient at removing heat from a planet because of two effects (e.g. Nakagawa and Tackley 2012): 1) latent heat of melting and 2) rapid cooling of the molten material at the surface. But a stagnant lid regime throughout Venus’ history would lead to unrealistically high rates of crustal production and Earth-like core/mantle heat flow (Armann and Tackley 2012), and therefore probably dynamo action, which is not observed (see next section).

The most efficient way to remove heat while meeting the present-day observations is for Venus to have had some type of mobile lid regime; either episodic overturn of the lithosphere (Parmentier and Hess 1992; Turcotte 1993, 1995; Armann and Tackley 2012), that may have frozen (Ogawa and Yanagisawa 2014), or plate tectonics that subsequently froze (Solomatov and Moresi 1996). An episodic lid mode with periodic global resurfacing would suppress core cooling during quiescent periods (Fig. 4) and could better match the present-day amplitude of the geoid and topography (Armann and Tackley 2012). But those models were done considering a constant surface temperature. However, the planet surface temperature depends on its atmosphere composition, which in turn depends on the outgassing of the mantle (especially water) and therefore on its tectonic regime. Coupled models show that it therefore varies with time (Fig. 6, after Gillmann and Tackley 2014). The feedbacks between the different ingredients of such models are complex, resulting in particular in hysteresis behavior between the main tectonics regimes (e.g. Weller et al. 2015; Weller and Lenardic 2017). The current lack of data to constrain Venus history (and even present-state) therefore leads to a multiplicity of possible evolutions.

As shown in this short review, climate influences tectonics through the role of surface temperature in a planet’s tectonic regime (i.e., stagnant lid versus plate tectonics). On the other hand, the tectonic regime affects climate through volatile cycling between the surface and interior. The tectonic regime also influences whether a magnetic field can be generated by dictating the core cooling rate. And finally, the strength of the magnetic field influences atmospheric escape, and therefore long-term climate evolution (e.g. Foley and Driscoll 2016; Gillmann and Tackley 2014). To decipher planetary evolution therefore requires to consider the whole coupled system.

4 Core Dynamics

The absence of a magnetic field on Venus today points toward the absence of dynamo action in its electrically conducting metallic core. Dynamo action requires that (i) convective motions are vigorous in the liquid core and can overcome magnetic diffusion, i.e. $V R_{cv} / \kappa_m > 1$, and (ii) the Coriolis force has a strong effect on the flow, so that $V / \Omega_v R_{cv} < 1$ (where V is the typical velocity of core motions, Ω_v is Venus rotation rate, R_{cv} is Venus core radius and κ_m is the magnetic diffusivity). Given the size of Venus core, the latter condition is fulfilled even though Venus rotation is slow (Hide 1956; Christensen and Aubert 2006). So dynamo action should occur as soon as (i) is valid. However, due to the incremental accretion of Venus-size planets, their cores should be stratified where the abundance of light elements increases with radius (Jacobson et al. 2017). Only

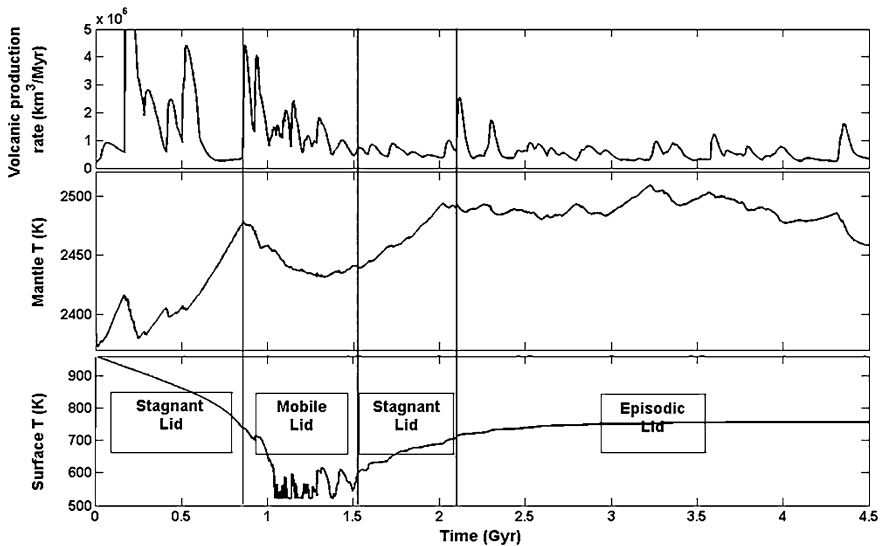


Fig. 6 Comparative evolution of volcanic production rate, surface temperature, and volume averaged mantle temperature with time for the preferred coupled model of Gillmann and Tackley (2014). Also indicated are the different convective regimes. The transition from mobile lid to stagnant lid is progressive. Early evolution (before 700 m.y.) follows an episodic, but mostly stagnant, lid pattern

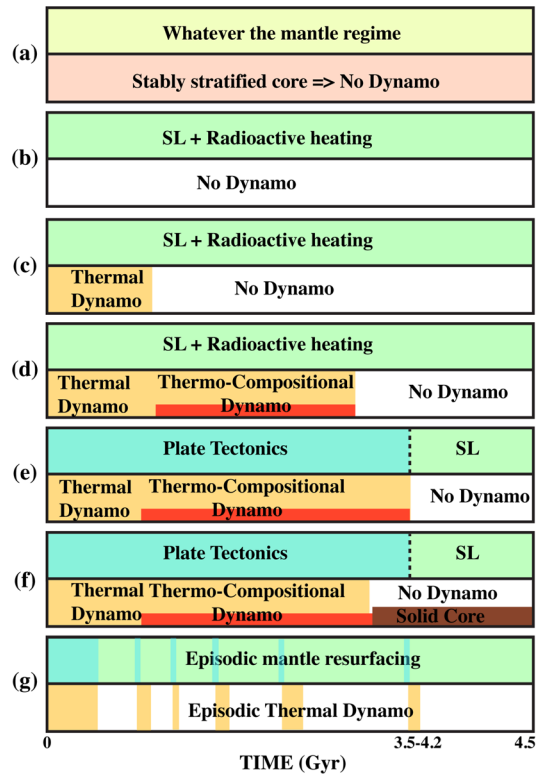
a late and very energetic impact, such as the Earth–Moon impact, is capable of eliminating this stratification through mechanical stirring. In the absence of such an impact, Venus core may have retained its original stratification, which would preclude any convection, and therefore dynamo action, to develop (Jacobson et al. 2017; O’Rourke et al., 2018, submitted for publication). Venus’ slow and retrograde rotation has been attributed to diverse tidal effects (Dobrovolskis and Ingersoll 1980; Correia and Laskar 2003; Leconte et al. 2015). On the other hand, it could be the signature of a large collision (Baines et al. 2013; Raymond et al. 2013), which then could have homogenized its core. In that case, convection still needs buoyancy sources, which can be provided either if a sufficient super-isentropic temperature gradient exists across the core (thermal convection) or if an inner core crystallizes and releases light elements (compositional convection). Both mechanisms require that the core is efficiently cooled, and therefore that convection in Venus mantle is extracting heat from the core fast enough. The value of the thermal conductivity of iron at core conditions is very controversial (e.g. Konopkova et al. 2016; Ohta et al. 2016). High values ($80\text{--}100\text{ W m}^{-1}\text{ K}^{-1}$) would render dynamo action impossible to sustain (even on Earth!), while lower values ($<50\text{ W m}^{-1}\text{ K}^{-1}$) would allow it (O’Rourke et al., 2018, submitted for publication).

Hence at least seven different scenarios can be envisioned, six of which are closely related to mantle convective regimes (Fig. 7, modified from Breuer et al. 2010):

(a) Venus core has remained stably stratified and therefore convection and dynamo cannot develop.

(b) Venus never had a dynamo because its mantle cools too slowly in the stagnant lid regime, especially if radioactive heating is high and the cooling effect of melting is not considered: the thermal gradient across its core–mantle boundary was never big enough for thermal convection to become powerful enough, nor for the core to cool down sufficiently to

Fig. 7 Different scenarios for mantle convection (green indicates a Stagnant Lid regime, -SL-, and blue a mobile upper surface) and core dynamo action (yellow indicates dynamo action, and white indicates no dynamo; pink indicates a stably stratified core). The red bars show core crystallization



start solidification. However, a dynamo should eventually develop when the inner core starts to grow (Stevenson et al. 1983; Nimmo and Stevenson 2000).

(c) An early thermal dynamo existed but ceased when the core-mantle heat flux became too low (typically < 10 TW), and the inner core growth has not started yet (Stevenson et al. 1983).

(d) Even if a stagnant lid regime is considered throughout the life of Venus, the timing of dynamo action, core crystallization and dynamo arrest depends strongly on the amount of volcanism out of Venus surface and the amount of radioactive elements present in Venus mantle and core (Driscoll and Bercovici 2014). Thermal and thermochemical dynamo action could have existed in the past for low radioactive heating and/or high melting rates (Driscoll and Bercovici 2014).

(e) The lack of dynamo today is directly due to the catastrophic resurfacing of Venus 0.3–1.0 Gyr ago, when it is associated to the transition from the cooling-efficient plate tectonics to stagnant lid. Plate Tectonics could have allowed core crystallization and thermo-compositional convection. But a period of warming of both mantle and core could have followed the transition to stagnant lid, leading to inner core melting and the arrest of dynamo action (Stevenson 2003).

(f) Plate Tectonics could have been so efficient at cooling Venus core, which is smaller than the Earth's, that the whole core could have solidified before the 1.0–0.3 Gyr transition, arresting dynamo action (e.g. Dumoulin et al. 2017).

(g) If Venus resurfacing is the consequence of episodic global overturn of the mantle or global lithosphere foundering (e.g., Turcotte 1993; Armann and Tackley 2012), Venus may

also have episodic pulses of dynamo action and magnetic field. During each overturn, the interior is strongly cooled and it is possible that the heat flow out of the core then exceeds the critical isentropic core heat flow, allowing for convection and dynamo action to proceed until core cooling becomes insufficient again.

Any trace of an ancient magnetic field would be helpful to discriminate among these scenarios. The Venusian surface temperature is close to the Curie temperature of the main magnetic carrier minerals, except for magnetite and hematite (e.g., Dunlop and Ozdemir 1997). Most of the crust, except for a thin layer at the surface, should be above the Curie point. This makes it difficult, but not impossible, to detect any remanent crustal magnetic field that would be the signature of an early episode of dynamo activity. It would be interesting in particular to look for any small spatial scale field (e.g. Schubert and Soderlund 2011).

5 Future Measurements

Venus is the only rocky planet other than Earth with a young surface age. The Venus Exploration and Analysis Group has laid out both a comprehensive suite of measurements and a proposed series of missions to explore this complex and active planet (www.lpi.usra.edu/vexag). In this section we outline a suite of measurements that are needed to further elucidate the evolution of Venus' interior and how the interior is coupled with the surface geology, atmosphere, and climate.

Gravity and Geodesy Data Gravity data provides the best available constraints on interior structure. However, the constraints are limited by current data. The average crustal thickness is known to only a factor of ~ 3 . The core diameter has an uncertainty of several hundred km (Konopliv and Yoder 1996), and even its state is uncertain (Dumoulin et al. 2017). Modeling of the gravity and topography data enables estimation of elastic thickness for features where the lithosphere experiences flexure in response to a load, such as a volcano. However, flexural bending occurs at relatively short spatial scales, and is currently only possible over a small fraction of Venus, typically with very large error bars (Anderson and Smrekar 1999). Future orbiters can provide higher precision and a uniformly higher resolution gravity field (Andrews-Hanna et al. 2016; Dumoulin et al. 2017; Mazarico et al. 2017), sufficient to enable elastic thickness estimation globally, and much improved crustal thickness. Future topography measurements could also greatly improve estimation of spin rate (Hensley et al. 2017), which is believed to be variable based on comparison of Magellan and Venus Express data (Mueller et al. 2012). Together with improved gravity resolution, *k2* and thus core size values can be derived with much better precision.

Topography, Radar Imaging, and Interferometry Venus' dense clouds prevent visual imaging from orbit. A single orbital radar instrument can produce high resolution synthetic aperture radar (SAR) imaging and topography, as well as interferometric surface deformation. Collectively these datasets are essential for understanding the geologic evolution of Venus. As discussed above, despite the leap in understanding enabled by topography and SAR imaging from Magellan, in the 3 decades since this radar system was built, new instrumentation has been designed that can provide orders of magnitude improvement in resolution (Ghail et al. 2017; Smrekar et al. 2018). These data sets are essential for understanding such questions as how has Venus resurfaced and what tectonic and volcanic processes are

active. Every time Mars and the Moon have been revisited with an order of magnitude improvement in resolution, entirely new processes have been discovered (Zimbelman 2001). This bodes well for not only addressing key questions about Venus' geology, but also for unanticipated discoveries on such a young planet. The potential for active processes makes Venus an ideal location for making the first interferometric measurements beyond Earth.

Surface Composition and Mineralogy Gilmore et al. (2017) discuss in detail the geochemical data provided by several landers in the 1970s and the information gleaned from 1 micron data from the VIRTIS spectrometer (Helbert et al. 2008) on the recently concluded Venus Express Mission. From the standpoint of interior evolution, key questions are whether or not all the tessera plateaus are in fact felsic, which implies formation in the presence of water, and what processes are producing present day volcanism. As described above, emissivity in the 1.02 micron band derived from data VIRTIS data is consistent with a felsic composition at Alpha Regio (Gilmore et al. 2015), the only tessera plateau covered by VIRTIS data. Similarly all of the hotspots covered by VIRTIS data, Dione, Imdr, and Themis Regiones plus Quetzelpetlatl Corona, have evidence of recent volcanism. Are all hotspots active? Are hotspots the only source of present day volcanism? If so, this might suggest that only the lower mantle, the source of large mantle plumes, contains enough volatiles to permit pressure release melting (Smrekar and Sotin 2012).

Although VIRTIS made tremendously exciting discoveries, it was originally designed for observing a comet on the Rosetta mission, not for observing Venus. A near infrared spectrometer designed for observing the surface of Venus could provide greatly enhanced signal to noise ratios in 6 surface bands. Such observations could demonstrate conclusively if other tessera plateaus are not only felsic, but also decipher the iron content to $\pm x\%$ (Dyar et al. 2017). Conducting this investigation requires high resolution topography in addition. Tessera terrain is highly deformed at the scale of 1–2 km. Current altimetry resolution of 12–25 km does not resolve these variations, and can introduce an average topographic bias on the order of 0.5 km. Surface temperature variations that are a function of altimetry much be removed as a data reduction step for surface emissivity. A 0.5 km elevation bias can introduce an error into surface emissivity that could be mistaken for a felsic composition. Thus topography capable of resolving the deformation structures on tessera is needed as an input for determining their global composition.

Noble Gases and Volatile Evolution Noble gases carry rich information on planetary origins (especially of volatiles), and on the history of planetary degassing and atmospheric processes using radiogenic isotopes produced on timescales ranging from the first several 10 s of m.y. to the present. Two noble gases, Ar and Ne, have been measured with reasonable precision for Venus (Donahue and Pollack 1983). However key gases such as Kr, Xe, and He are effectively unknown. High precision measurement of these gases and their isotopes constrain exogenic and endogenic volatile sources and loss mechanisms to the noble gas mantle-crust-atmosphere system. Noble gas provide a fingerprint of the building blocks of planetary formation. Volcanic degassing releases primordial and radiogenic/fissiogenic isotopes produced at different times after accretion, which may be lost via varying escape mechanisms. Comet and asteroid impacts may also enrich noble gases and change their relative abundances in the atmosphere, as discerned through isotopic abundances and ratios. Acquiring high precision data for noble gases on Venus will not only elucidate the early history of Venus but also that of other terrestrial planets by providing an additional data point to discriminate between processes dependent on mass, distance to the sun, and geologic activity.

Seismology Of all possible measurements, a robust seismic investigation of Venus is both the most valuable for illuminating interior structure and dynamics, and the most challenging to obtain. Seismic data could determine crustal thickness and layering, core radius and state, mantle temperature, and types and frequency of tectonic or volcanic events with far greater precision than other data sets. The most information would be obtained using a network of stations, which would capture the most seismic events and provide a better inventory of global activity. Valuable information also can be derived from a single seismic station (Panning et al. 2015). Some of the technologies required for a long duration lander, a requirement for seismic monitoring, are in development, but the high data rate needed for seismology is not yet available. An alternative approach is to search for seismically generated waves in the atmosphere. The dense atmosphere of Venus creates a strong coupling between fault-induced motion of the surface and atmospheric waves (Garcia et al. 2005) that can potentially be detected from either an aerial platform or from orbit (Cutts et al. 2015). Although these methods do not yield the three components of ground motion recorded with surface seismic stations and thus do not provide as much information as a surface network, such measurements could give information on the types and frequency of events on Venus and provide a pathfinder for future surface measurements.

Heat Flow Given the similarity in radiogenic content measured by the Venera landers (e.g. Gilmore et al. 2017), along with the size of the planet and its young surface age, one of the major questions for Venus is how does it lose its heat in the absence of plate tectonics. Heat flow measurements would discriminate between classes of tectonic evolution models, which make predictions about present day heat flow depending on whether or not Venus had episodic or steady rates of heat loss.

Magnetic Field All measurements of the magnetic field to date have been conducted from orbit, and do not preclude an extremely weak dynamo (Luhmann et al. 2015). As discussed above, the surface temperature of 460C precludes preservation of crustal remanent magnetization for most minerals, although the Curie temperatures of magnetite and hematite are well above this value. Measurements of the magnetic field from the surface, an entry probe, or aerial platform could reveal either remanent magnetization or a very weak dynamo, either of which would provide strong constraints on models of interior evolution.

6 Conclusions

Magellan revealed Venus as a highly complex planet. Subsequent observations and analyses suggest it is also a dynamic planet, with likely active volcanism and outgassing. Hot plumes from the core-mantle boundary create surface uplift and recent volcanism (Smrekar et al. 2010; Davaille et al. 2017), yet do not draw enough heat from the core to drive a significant dynamo. One of these locations appears to have induced subduction (Davaille et al. 2017). One question is whether or not the only sites of active volcanism are at hotspots (Smrekar et al. 2010). If so, this could point to a dry upper mantle and more volatile-rich lower mantle (Smrekar and Sotin 2012). Evidence for geologically recent pyroclastic flows point to the role of volatile outgassing (Campbell et al. 2017). A significant increase in sulfur content over the lifetime of Venus Express further suggests present day outgassing (Marcq et al. 2013). If volatiles are coming out of the interior, subduction and delamination may recycle them back into the interior (Elkins-Tanton et al. 2007).

As Earth's virtual twin, Venus represents a control case for understanding rocky planet evolution, including Earth. Venus has long been considered an analog for early Earth, and may hold clues for understanding the conditions under which Earth, and possibly exoplanets, transition from a single plate to plate tectonics. As more Earth-sized exoplanets are discovered, the objective is to understand just how Earth-like, and habitable, they may be. The conditions that drove Venus and Earth down such divergent evolutionary paths is one of the primary questions for understanding habitability. Models of geologic evolution are required to ingest the limited information on exoplanets—size, density, insulation, and eventually upper atmospheric composition—and make informed predictions. The role of plate tectonics, though long-term formation of the atmosphere and recycling of volatiles into the mantle, has emerged an essential element of this prediction. Venus holds the key to predicting the conditions that will or will not produce plate tectonics.

Although there are a host of compelling questions about our sister planet, several fundamental questions stand out for interior evolution. Did Venus experience catastrophic resurfacing? The catastrophic resurfacing hypothesis is the basis for episodic models of interior evolution. Are these models valid, or is Venus losing its heat through smaller scale resurfacing, perhaps in a mode distinct from either mobile or stagnant lid convection? Does Venus have remanent magnetization? Evidence for a dynamo would provide another key constraint on thermal and compositional evolution. Does Venus have subduction? Could it produce transform faults that are precursors to plate tectonics? What's the average surface age? Could it be closer to 100 my than 1 by? This question is essential to understating heat loss mechanisms and overall evolution. Are Venus' tesserae plateaus analogs to Earth's continents and a fingerprint of interior water? Is Venus still outgassing today? Volatiles exert a strong control on crust and mantle rheology, melting and volcanism, as well as on climate. Numerous missions have been proposed to answer these questions (Ghail et al. 2017; Smrekar et al. 2018).

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