Space Science Reviews (2022) 218:70
https://doi.org/10.1007/s11214-022-00937-9

Dynamics and Evolution of Venus' Mantle Through Time

Tobias Rolf[1,2] · **Matt Weller**[3,4] · **Anna Gülcher**[5] · **Paul Byrne**[6,7] · **Joseph G. O'Rourke**[8] · **Robert Herrick**[9] · **Evan Bjonnes**[3,4] · **Anne Davaille**[10] · **Richard Ghail**[11] · **Cedric Gillmann**[12] · **Ana-Catalina Plesa**[13] · **Suzanne Smrekar**[14]

Received: 25 February 2022 / Accepted: 7 November 2022 / Published online: 28 November 2022

Abstract
The dynamics and evolution of Venus' mantle are of first-order relevance for the origin and modification of the tectonic and volcanic structures we observe on Venus today. Solid-state convection in the mantle induces stresses into the lithosphere and crust that drive deformation leading to tectonic signatures. Thermal coupling of the mantle with the atmosphere and the core leads to a distinct structure with substantial lateral heterogeneity, thermally and compositionally. These processes ultimately shape Venus' tectonic regime and provide the framework to interpret surface observations made on Venus, such as gravity and topography. Tectonic and convective processes are continuously changing through geological time, largely driven by the long-term thermal and compositional evolution of Venus' mantle. To date, no consensus has been reached on the geodynamic regime Venus' mantle is presently in, mostly because observational data remains fragmentary. In contrast to Earth, Venus' mantle does not support the existence of continuous plate tectonics on its surface. However, the planet's surface signature substantially deviates from those of tectonically largely inactive bodies, such as Mars, Mercury, or the Moon. This work reviews the current state of knowledge of Venus' mantle dynamics and evolution through time, focussing on a dynamic system perspective. Available observations to constrain the deep interior are evaluated and their insufficiency to pin down Venus' evolutionary path is emphasised. Future missions will likely revive the discussion of these open issues and boost our current understanding by filling current data gaps; some promising avenues are discussed in this chapter.

Keywords Venus · Mantle dynamics · Interior evolution · Surface tectonics · Thermal history

1 Introduction

Early space missions–such as Pioneer, Venera, and Magellan–indicated that Venus' present tectonic regime differs substantially from the Earth's, but a profound answer for why this is so remains lacking. One major challenge to resolve this issue is the difficulty of observing

Venus: Evolution Through Time
Edited by Colin F. Wilson, Doris Breuer, Cédric Gillmann, Suzanne E. Smrekar, Tilman Spohn and Thomas Widemann

Extended author information available on the last page of the article

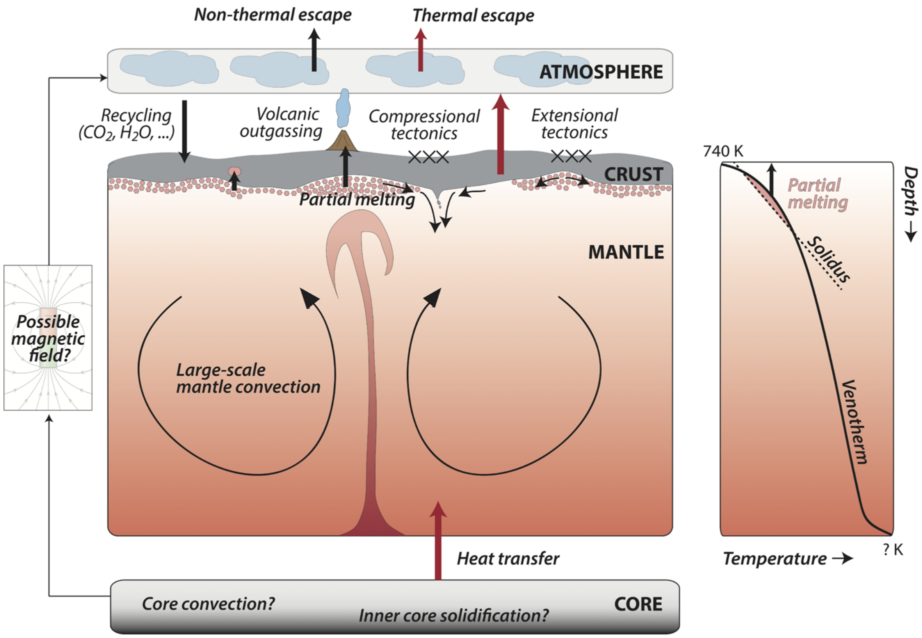

Fig. 1 A schematic of the role of the mantle in planetary evolution. Red arrows denote the exchange of heat; thick black arrows denote other means of transfer such as magma transport from the mantle into the crust and to the surface; thin black arrows indicate flow. The box at right illustrates a hypothetical temperature profile (here named the 'Venotherm'); the dashed line is a hypothetical solidus

robust constraints on Venus, particularly for the deep interior. Nevertheless, those missions provided important insights into Venus' interior structure and dynamics (see e.g., O'Rourke et al. 2023), which are key for understanding the planet's surface tectonics. Bulk density and composition suggest an interior structure that is similar to that of Earth: a massive, iron-rich core overlain by a thick silicate layer (e.g., Margot et al. 2021). The large heat capacity and high viscosity of that silicate mantle cause huge thermal inertia and long dynamic time scales. Over the billions of years of Venus' evolution, however, the mantle is still a highly dynamic system and its evolution determines the state of Venus' interior, the tectonic expressions at the surface, and the interaction with Venus' fluid layers including the atmosphere, the core, and thus potential magnetic field generation and evolution (Fig. 1).

Solid-state convection is the key mechanism of heat transport in the mantle (e.g., Stevenson 2003) and its efficiency determines the temperature in the silicate interior as well as the thickness of Venus' lithosphere and crust. In addition to the thermal impact of convection, upwelling mantle flow moves material from regions of higher pressure to regions of lower pressure (decompression), causing topography at the boundaries of mantle convection cells and controlling the locations and rates of partial melting. Eventually, these partially molten zones form the source regions of intrusive magmatism and extrusive volcanism. Such processes in turn lead to compositional heterogeneity in the interior, determine the thickness and rheological strength of the crust (e.g., Lourenço et al. 2020), and control the outgassing of volatiles into Venus' atmosphere (e.g., O'Rourke and Korenaga 2015). At the bottom of the mantle, thermal exchange determines core cooling and the heat transport efficiency in the planet's central layer. If the mantle allows for sufficient core cooling, an inner core may crystallise at some point during the planet's evolution (e.g., O'Rourke et al. 2018), depending on the core composition (e.g., O'Neill 2021). Core cooling and inner core crystallisation drive convection in the liquid iron-rich core and possibly power a core dynamo. This dynamo can generate a planetary magnetic field, which may interact with the atmosphere in several ways including shielding from stellar winds and volatile loss, but to date this remains strongly debated (see Way et al. 2023; Gillmann et al. 2022).

In summary, the mantle interacts with essentially all other subsystems of a planet over geological time scales. For Venus, these couplings are partly revealed by observational constraints, but many details remain unanswered. This is the case for Venus' present-day state–in principle directly observable–and becomes more challenging for the evolution of Venus' mantle through time. Today, Venus neither features plate tectonics nor generates an Earth-like magnetic field. Evidence for volcanism and tectonics are ubiquitous on the surface, but what caused the divergence between Venus and Earth remains poorly resolved. Were the conditions on Venus less favourable for the development of plate tectonics right from the start, because of differences in the accretion history, or because Venus is closer to the Sun and therefore received more insolation (see Salvador et al. 2023)? Or did the evolution of both planets start off relatively similar before diverging at some later stage because of an endogenic or exogenic trigger?

Previous missions to Venus returned a number of observables useful to constrain interior models, which are summarised in Sect. 3.1 and detailed in Herrick et al. (2023), Ghail et al. (2023), Carter et al. (2023), and Gilmore et al. (2023). Their coverage, resolution, and non-uniqueness leave gaps and uncertainty in Venus' core and mantle dimensions, in how much heat is transferred from the mantle to the atmosphere, and in the thickness, age, and composition of the crust. However, new space missions will launch to Venus within a decade to further fill existing gaps (see Widemann et al. 2023). It is thus timely to review our understanding of Venus' mantle evolution through time and to identify what we know and what we do not yet know.

This work is not the first review on Venus' interior evolution (e.g., Mocquet et al. 2011; Smrekar et al. 2018), but it particularly focuses on how mantle dynamics generate and relate to different regimes of surface tectonics and volcanism, and how this relation may evolve through time. Our goal is to emphasise which possible pathways Venus' mantle may have taken to its present state, and what can be done to reduce the number of feasible scenarios in future. Section 2 provides a background on heat transfer in planetary mantles, the peculiar properties of mantle silicates, and the spectrum of planetary tectonic regimes, driven by mantle convection. Section 3 summarises the observations constraining Venus' contemporary mantle and discusses feasible tectonic regimes relating mantle dynamics, crustal tectonics and volcanism. Section 4 reviews Venus' mantle evolution through time with a focus on the possibility of lateral and particularly temporal variations as well potential triggers. Finally, Sect. 5 gives an overview of possible mantle evolution scenarios for Venus. A perspective is given on how future conceptual understanding and data collection can boost our understanding of Venus' interior and help to resolve unanswered questions specific to the planet's mantle.

2 Planetary Tectonic Regimes

2.1 Basics of Planetary Mantle Convection

The evolution of a planetary body is strongly controlled by its thermal history. Whether the primordial heat accumulated during accretion and differentiation is efficiently kept in the interior or can easily escape determines the planet's cooling rate and the vigour of internal dynamics. Inside a Venus-like planet, heat is transported via thermal conduction or convection (i.e., heat transport via large-scale material flow), where the latter is the much more efficient mechanism for most parts of the mantle. The vigour of convection determines the efficiency of cooling and partial melting of the silicate mantle that leads to crustal production and volcanic outgassing. Thus, interior dynamical processes can be linked to surface expressions such as tectonics and volcanism as well as the evolution of the atmosphere.

A simple analogue to planetary mantle convection is the Rayleigh–Bénard system, where a homogeneous fluid layer of finite thickness D is heated uniformly from below and cooled from above. Heating the fluid leads to unstable density stratification and initiates convective currents once the heating-induced density contrast becomes sufficiently strong. At that point, thermal conduction cannot limit the upward-directed buoyancy anymore and this force overcomes the viscous resistance against the onset of motion inside the fluid layer. Whether convection occurs is determined by the Rayleigh number: $Ra = \alpha g \Delta T \rho D^3/\kappa\eta$, where α is the thermal expansivity, g the gravitational acceleration, ΔT the superadiabatic temperature contrast across the layer, and ρ, κ, η are the layer density, thermal diffusivity, and viscosity, respectively. For a convective instability to grow and generate large-scale convection, a critical value of Ra must be reached, which depends on the layer geometry and boundary conditions. Typically, this value is $\sim 10^3$ (Turcotte and Schubert 2017). Transferred to Venus' mantle, Ra is $O(10^7\text{-}10^8)$, using $D = 3000$ km, $\alpha = 3 \times 10^{-5}$ K^{-1}, $\Delta T = 2500$ K, $g = 8.87$ m s^{-2}, $\rho = 3300$ kg m^{-3}, $\kappa = 10^{-6}$ m^2 s^{-1}, $\eta = 10^{20}\text{-}10^{21}$ Pa s, which implies vigorous, time-dependent convection–despite the huge viscosity of mantle rocks.

A vigorously convecting system develops thin thermal boundary layers (TBLs) near the top and bottom boundary. Across these TBLs, heat is transported via thermal conduction, but this transport cannot accommodate the continuous inflow of heat, so that the TBLs form convective instabilities, expressed as upwellings and downwellings. In comparison to these instabilities, the bulk mantle behaves passively, is essentially stirred around, and adopts a thermal profile that does not actively contribute to driving convection. The larger Ra, the more prone the system is to instabilities and the thinner the TBLs become. On Earth, the TBLs are $\sim$100 km thick and the surface TBL is thought to correspond to the (oceanic) lithosphere.

Convective currents self-organise the relative spacings of up- and downwellings depending on the properties of the convecting layer. In the Rayleigh–Bénard system, these spacings are comparable to the layer thickness, but in planetary mantles several peculiarities promote convection cells with large aspect ratios. These include the strength of the lithosphere (e.g., van Heck and Tackley 2008; Yoshida 2008; Rolf et al. 2014, 2018a), pressure-dependence of mantle viscosity (Bunge et al. 1997; Höink and Lenardic 2008, 2010; Höink et al. 2012; Lenardic et al. 2019) and other material properties (e.g., Hansen et al. 1993), as well as the mantle heating mode (McNamara and Zhong 2005). On Earth, this is manifested in the size of the largest tectonic plates, like the Pacific plate. On Venus, however, there seems to be little indication of such long-wavelength flow structures in the mantle.

2.2 Specific Complexities of Planetary Mantle Convection

2.2.1 Planetary Heating Modes

The Rayleigh–Bénard setup oversimplifies planetary mantle convection. For example, the ratio of the Earth's core and mean surface radii of $\sim$0.55 implies that the plane-layer approximation is inaccurate, impacting the flow patterns in the mantle (e.g., Weller et al. 2016;

Yanagisawa et al. 2016; Guerrero et al. 2018). Venus' radius ratio is not well understood given the uncertainty in core radius (e.g., Margot et al. 2021), but is typically assumed to be similar to Earth's. Due to planetary curvature, the surface boundary comprises a larger area than the core–mantle boundary. This makes the bottom boundary layer more prone to instabilities compared with the plane-layer geometry, if the same heat flow across the core–mantle boundary is considered.

The mantles of Earth and (probably) Venus contain long-lived radiogenic nuclides (in particular ^{40}K, ^{232}Th, ^{235}U, and ^{238}U) that generate internal heat by radiogenic decay, with rates decreasing through time and possibly varying spatially. Internal heating interplays with basal heating from the hot core established during the planet's accretion and differentiation. The rates of basal and internal heating control the stability of the boundary layers, the spacing of its instabilities, and therefore the planform of mantle flow (Moore 2008; Weller et al. 2016; Korenaga 2017). Internally heated convection exhibits different flow patterns than basally heated convection, because the bottom boundary layer is absent or greatly weakened as the interior adopts the temperature of the bottom boundary and inhibits inflow of heat from below (e.g., Mulyukova and Bercovici 2020). This suppresses active hot instabilities, which become diffuse and passive return flows compared with the active and pronounced downwellings (e.g., McKenzie et al. 1974; Parmentier et al. 1994; Sotin and Labrosse 1999).

Both purely basal or purely internal modes of heating are special cases for planets that either are sufficiently ancient to have depleted all their internal heat sources, or sequestered the bulk of them into a nonrecyclable crust, or have conditions where the temperature of the mantle temporarily becomes equal to that of the outer core. For both Earth and Venus, such special cases are extremely unlikely, suggesting that mixed mode heating dominates heat transport and either planet's thermal evolution. Mixed mode convective heating allows for a strong mechanical interaction between both the upper and lower boundary layers (here the upper boundary layer is analogous to the thermal planetary lithosphere). This exerts a first order control on both the thickness and the heat flux through these layers (Moore 2008; Weller et al. 2016). Critically, the mantle then deviates from the classic definition of an adiabatic interior (Weller et al. 2016; Lenardic et al. 2019). Boundary layer interaction through mixed heating is a stark departure from the regime identified in classical theory (Howard 1966; Fowler 1985), where each boundary layer thermally destabilises on its own upon reaching a critical thickness (Sect. 2.1). This emphasises that the heat loss due to convection in planetary mantles deviates from classical convection experiments, and that extrapolation of real behaviour from classic theory is limited.

For any planet, the ratio of basal to internal heating is debatable as it depends on the style of planetary accretion and differentiation (e.g., McKenzie et al. 1974), and is also a function of the planet's evolution and geodynamic regime. For Earth, traditional estimates of ∼90% suggested a mostly internally heated mantle (e.g., Sleep 1990), but more recent estimates indicate a more balanced partitioning of ∼60–70% internal heating (e.g., Lay et al. 2008; Leng and Zhong 2008, 2009). On Earth, the main challenge is to estimate the heat flow across the core–mantle boundary, which depends on the poorly understood thermal conductivity of silicates under deep mantle conditions as well as on the temperature of the core. For Venus, not even the heat flow across the surface has been robustly measured, but indirect estimates exist (Sect. 3.1.3); heat flow across Venus' core–mantle boundary is almost wholly unknown. The absence of an intrinsic magnetic field precludes a thermally driven core dynamo, so that the heat flow conducted along the core adiabat may be an upper bound. However, this scenario assumes an Earth-like core for Venus, which is not guaranteed by the available data (Sect. 3.1.6).

2.2.2 Material Properties and Rheology

The mantle's material properties strongly depend on temperature, pressure, and other factors. As a result, viscosity, thermal expansivity, conductivity, diffusivity, and density vary through space (and time) and the system-characterising Rayleigh number of the bulk mantle does not properly determine the dynamics on local scales. Potential consequences of these variations range from different planforms of mantle flow (e.g., Hansen et al. 1993; Tosi et al. 2010), to different surface heat flows (e.g., Ghias and Jarvis 2008), to differences in obtained surface topography and geoid (e.g., Schmeling et al. 2003).

Viscosity in particular is known to vary over many orders of magnitude throughout the mantle as a function of temperature, pressure, composition, mineral phase, water content, grain size, previous deformation history and various other parameters (see Karato 2010). This material property relates the strain rate with the stress that material experiences. Under different strain-rate versus stress conditions, deformation happens through different mechanisms. At relatively low levels of stress, deformation is viscous. In diffusion creep, viscosity does not explicitly depend on stress, but depends strongly on grain size. At higher levels of stress, but still in the viscous regime, dislocation creep becomes dominant and viscosity has a power-law dependence on stress, yet is insensitive to grain size. In Venus' mantle, both creep mechanisms coexist and which one dominates strongly depends on material properties such as water and melt content, mineralogical phase, deformation texture, and oxygen fugacity (e.g., Kohlstedt and Hansen 2015). Although coexisting creep mechanisms are frequently considered in terrestrial studies of crustal to lithospheric scale, these complexities are often ignored on planetary-scale problems (but cf. Rozel et al. 2014; Dannberg et al. 2017; Schulz et al. 2019 and others). This issue could be especially relevant in connection with grain-size evolution, which under Venus' hot surface conditions may affect crustal rheology very differently than under Earth-like conditions (Bercovici and Ricard 2014).

Water content is another key component of Venus' interior rheology. If Venus' present mantle and shallow crust are drier than Earth's, as suggested by some studies (e.g., Namiki 1995), the strength of Venus' crustal rocks may be enhanced compared to Earth's (e.g., Mackwell et al. 1998) with a commensurate impact on the style of tectonics (e.g., Moresi and Solomatov 1998; Turcotte 1996). Holding all else equal, reduced water content would also have a dampening effect on melting and magmatism as water acts to reduce the mantle solidus (e.g., Green et al. 2014; Ohtani 2020). Although abundant water tends to decrease mantle viscosity, the thermo-tectonic feedback between surface motions and cooling of the mantle may actually result in a more viscous mantle if viscosity is very sensitive to water content (Nakagawa et al. 2015). Abundant water in the mantle transition zone could also help to explain variations in geochemical signatures of basaltic lavas across the planetary surface. If water-rich ambient mantle ascends out of the mantle transition zone into a zone of low-water-solubility, it may undergo dehydration-induced partial melting, thereby filtering incompatible elements out of the depleted rising material (Bercovici and Karato 2003). Finally, water transported into the deepest mantle could lead to chemical reactions with iron-rich materials (e.g., Yuan et al. 2018) that may possibly trigger large-scale geodynamic events (Mao et al. 2021).

Depending on the tectonic regime, water transport in Venus' mantle may differ from that inside Earth. Current data for Venus points to relatively low water content in the atmosphere, but constraints on estimates of the water budget of the mantle remain poor (e.g., Zolotov et al. 1997, and references therein), even though Venus' high D/H ratio may imply that substantial parts of Venus' interior water may have been lost, possibly by volcanic outgassing and atmospheric escape (e.g., Grinspoon 1993; Gillmann et al. 2022). The initial water content of the mantle after magma ocean solidification is a major unknown (Salvador et al. 2023). Parameterized convection models coupled with water transport suggest self-regulation effects that make the present water content only weakly sensitive to the initial content (Sandu et al. 2011)—but these models assume a terrestrial tectonic regime that may be inapplicable for Venus, and more complex fully dynamic models have not confirmed such self-regulation effects (Nakagawa et al. 2015).

In the lithosphere and crust, relatively low temperatures lead to high viscosity and stress that can lead to brittle fracturing at the low confining pressure. The consequence is irreversible plastic deformation associated with strong localization that is thought to enable the formation of narrow weak zones, such as faults and, at larger scales, systems of faults such as rifts. The stress threshold that surface rocks can sustain is often called 'yield stress' and is one crucial property of the planetary lithosphere. If the yield stress is small, convection-induced stresses can fragment the lithosphere and may generate tectonic plates such as on Earth (Sect. 2.3). Such a yield stress approach has been used in many numerical studies to map out the feasibility of a plate-like regime (e.g., Moresi and Solomatov 1998; Trompert and Hansen 1998; Tackley 2000; Stein et al. 2004; van Heck and Tackley 2008; Foley and Becker 2009). A typical, only partly resolved problem, however, is that feasible yield stress ranges in large-scale numerical models are substantially smaller than those inferred from laboratory experiments on terrestrial rocks (e.g., Kohlstedt et al. 1995). Non-uniform yield stress across the surface, for instance due to variations in composition (Lenardic et al. 2003; Rolf and Tackley 2011) or inherited weakening from previous deformation (e.g., Fuchs and Becker 2019, 2021; Miyagoshi et al. 2020) may help to facilitate the formation of plates and to keep the lithosphere mobile even at higher yield stress of the bulk and/or undamaged lithosphere. Such structural inheritance may result from various mechanisms inducing damage. Grain-size evolution is a prominent example as it is manifested in terrestrial shear zones, where grain size is strongly reduced during active deformation (e.g., Okudaira et al. 2017). Subsequent grain growth heals the damage at a strongly temperature-dependent rate, which implies that planetary surface temperature could be a key factor in determining the role of structural damage of the crust and lithosphere. Venus' high surface temperature may prevent preservation of lithospheric weakness that could ultimately have led to the formation of the first tectonic plates on the (relatively cold) surface of the Earth (e.g., Bercovici and Ricard 2014; Foley 2018).

2.2.3 Mineralogy and Compositional Variation

Composition and mineralogy are additional factors that determine state and dynamics of the mantle. A detailed review of Venus' surface composition and mineralogy is given in Gilmore et al. (2023). Depending on the ambient temperature–pressure conditions, mantle rocks undergo a series of phase transitions that modify crystallographic structure. Both density and viscosity change, which possibly promotes separation of distinct layers (Weidner and Wang 2000) and affects mantle flow structure and radial heat transport (Tackley 1996; Bunge et al. 1997). Certain components may become abundant in specific regions of the mantle. For instance, subducted basaltic crust is buoyant at the base of the mantle transition zone, but negatively buoyant just above and below as garnet transitions to (Mg-)perovskite at slightly higher pressure than ringwoodite. As a result, basaltic crust may get trapped in the transition zone of the (Venusian) mantle (e.g., Armann and Tackley 2012; Vesterholt et al. 2021). Episodic breakdown may induce mantle avalanches and possibly trigger dramatic periods of volcanism and large-scale tectonic activity on Venus (e.g., Papuc and Davies 2012; Vesterholt et al. 2021), although some geodynamic models have challenged the relevance of this behaviour for Venus (e.g., Huang et al. 2013). Phase transitions also alter mantle flow by either absorbing or releasing latent heat. Secondary plumes (e.g., Yuen et al. 1998; Yoshida 2004) are likely to form when hot upwellings penetrate an endothermic phase change like that from ringwoodite to perovskite at a pressure of 23 GPa or equivalent depth of 730-740 km inside Venus (Ishii et al. 2018; Trønnes et al. 2019).

Moreover, phase changes depend on mineral assemblage and rock composition, which vary temporally and spatially in the mantle. Although the bulk composition is mostly determined by the accretion history, compositional mantle heterogeneity may arise from melting and freezing processes. For example, compositional layering may be a consequence of fractional crystallisation of an early magma ocean (e.g., Labrosse et al. 2007). Later-stage partial melting induced by hot anomalies in a largely solidified mantle leads to small-scale heterogeneity as different rock components melt under different conditions and have different affinity to partition into the molten phase (e.g., Hofmann 1997). Such processes compete with mantle convective stirring that tends to homogenise lateral variation, but the mixing time scales in the mantle are large and compositional heterogeneity occurs likely across multiple scales in the Earth's present mantle (e.g., Tkalčić et al. 2015), as is also supported by modelling studies (Gülcher et al. 2021). However, terrestrial plate tectonics continuously induces new heterogeneity by deep crustal subduction, which may be less relevant to Venus without plate tectonics.

Finally, melting processes likely cause the growth of secondary crust on the planet's surface. Basalt-rich melts in the upper mantle rise to shallow depths where they cool and solidify to form fresh basaltic crust. Such magmatic transport may happen either via magmatic intrusions in the shallow subsurface, or via extrusive volcanism in cases where melt diapirs reach the surface via so-called heat pipes (Fig. 2b, Turcotte 1989; Moore et al. 2017). Upon such melt formation, latent heat is consumed and migrated to the surface, where it can be released to the atmosphere. Therefore, melting, magmatism, and volcanism strongly influence the cooling history of a planet and with that the dominant tectonic regime at its surface (e.g., Ogawa 2000; Ogawa and Yanagisawa 2014; Lourenço et al. 2016, 2018, 2020; Byrne 2019).

2.3 Diversity of Geodynamic Regimes

Material complexity in the mantle allows for different styles of mantle flow. In turn, this flow induces deformation in the crust and thus determines surface tectonics. Present Earth is in a mobile-lid tectonic regime (with plate tectonics being a peculiar subcategory), but most other known terrestrial planets show different regimes, which are introduced in this section and conceptually visualised in Fig. 2. Venus' current and past regime is then specifically discussed in detail in Sects. 3 and 4.

The standard view of convection (like that from Bénard's experiments) is the mobile lid regime. The name implies that the lid (typically defined as the region inside the top thermal boundary layer) is in continuous motion, with a velocity at least as large as the average velocity of the convecting layer. In other words, the surface mobility (M), defined as the ratio of surface-averaged to volume-averaged velocity, is $M \geq 1$ (e.g., Tackley 2000). M is a useful measure to distinguish mobile from immobile (thus, stagnant) surface regimes. Earth's plate tectonics clearly falls into the mobile category given the observed speeds of tectonic plates. One peculiarity of plate tectonics, however, is the high degree of localisation of deformation into narrow plate boundaries (Fig. 2c). This localisation is not a general characteristic of the

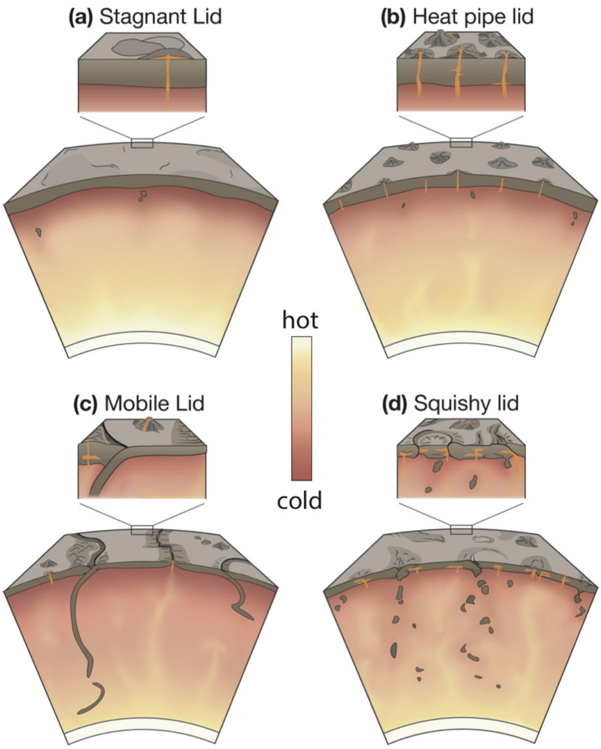

Fig. 2 A conceptual visualisation of geodynamic regimes relevant for the terrestrial planets: (**a**) Stagnant Lid; (**b**) Heat Pipe; (**c**) Mobile Lid; (**d**) (Plutonic) Squishy Lid. The regime could conceivably transition between these modes through time (see Sect. 4). The Episodic lid is an often-proposed specific example of such transitional behaviour between mode (**a**) and (**c**). The colour coding qualitatively indicates temperature, as indicated. The white layer at the bottom of each panel indicates the core of the planet, the brown layer at the top is the lithosphere. The zoom-ins are used to highlight characteristic volcanic and tectonic structures

mobile lid regime as deformation may be more diffuse or feature patterns that are absent in Earth's surface motion. In an Earth-like regime, most deformation is confined to a small fractional area of the surface. This can be expressed via the plateness (P), which quantifies how much more localised tectonic deformation occurs in the investigated regime compared to a reference regime, which is typically the simplest possible, thus an isoviscous regime. A value of $P = 1$ indicates an extremely high degree of localisation, while $P = 0$ means that localisation is as poor as in the isoviscous case (Tackley 2000). Therefore, on Earth $P \to 1$, while a wider range of values ($0 \ll P < 1$) is representative of the mobile lid.

The viscosity of mantle rocks is so sensitive to temperature that the surface boundary layer becomes a quasi-rigid lid and decouples from the convecting mantle below (e.g., Solomatov 1995). This regime is often called 'stagnant lid' (Fig. 2a). Small viscosity contrasts between cold lithosphere and hot mantle degrade the decoupling and lead to a mixed regime ('sluggish lid') in which the surface lid is mobile, but at reduced rates ($M < 1$). The strong temperature dependence of mantle rocks makes a stagnant lid seemingly inevitable, unless additional processes lead to localised failure of the stagnant layer. The stagnant lid is thus sometimes seen as the default mode of planetary mantle convection, with present Mars being an archetype of this regime. The stagnant lid is also likely the terminal mode of mantle evolution when the planet has cooled so much that convection is not maintained any longer (e.g., O'Neill et al. 2016; Stern et al. 2018). In the stagnant-lid regime, surface velocity is much smaller than interior velocity ($M \ll 1$). Surface deformation is not particularly focussed into narrow weak zones, so that the characteristic plateness is also small ($P \ll 1$).

The transition between the mobile lid and stagnant lid regimes is determined by the competition between stress induced into the lithosphere and the integrated strength of that

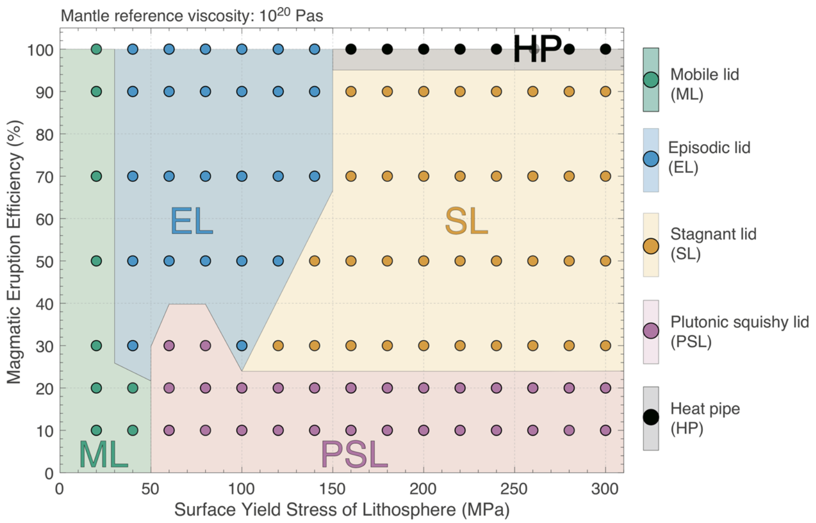

Fig. 3 A geodynamic regime diagram as a function of the surface yield stress of the lithosphere and the magmatic eruption efficiency. Here, the yield stress is defined as the maximum stress material can sustain without deforming plastically; its values cannot be directly compared to laboratory inferred values. Each dot indicates a geodynamic model; its colour denotes the geodynamic regime and is determined on the basis of average surface mobility and plateness–a measure for the localization of surface deformation (Tackley 2000). Background colours illustrate regime fields qualitatively. The regime diagram is also a function of other parameters, such as mantle viscosity (here: 10^{20} Pa s). The plotted data is taken from Lourenço et al. (2020)

lithosphere (Fig. 3). Lid mobility may occur when stresses inside the lithosphere overcome its internal strength locally to induce weakness by failure. The transition to mobile lid is probably not sharp, but includes a transitional range. One flavour of this transition is episodicity. In this context (but not generally), this means that episodes of pronounced surface mobility intersect the evolution in the stagnant-lid mode. Stress may build up gradually during evolution, for instance due to mantle cooling or crustal and lithospheric thickening (e.g., Fowler and O'Brien 1996). At some point the lid is mobilised and tectonic recycling cools the mantle and reduces lid thickness. In turn, the stress in the lithosphere is lowered and active recycling stops again until stress has built up again to initiate another episode of recycling. The scales of such resurfacing events may range from regional (e.g.,Noack et al. 2012; Karlsson et al. 2020; Weller and Kiefer 2020) to global (Turcotte 1993; Armann and Tackley 2012). In its global form, this 'episodic lid' regime has been used as an explanation for Venus' quasi-random distribution of impact craters that imply uniform surface age, but the necessity of such a catastrophic event has been challenged (Sects. 3.2, 4.2). During overturn phases, surface mobility and plateness are similar to the mobile lid characteristics but, between such phases, these diagnostics are representative of stagnant lid behaviour. Temporal averages would thus strongly depend on the time scales of overturn episodes.

The regimes outlined so far do not consider melting and magmatism. As heat transfer across a stagnant lid is much less efficient than across a mobile lid, heat is trapped inside the planet and leads to a hotter mantle. This trapped heat enhances melting and magmatic activity. Latent heat consumption and transport of the hot, buoyant magma via volcanic eruption together facilitate extraction of interior heat, buffer mantle temperatures, and act as a 'mantle thermostat' (e.g., Ogawa and Yanagisawa 2011). This mode has been described as a 'heat pipe' (Fig. 2b, e.g., O'Reilly and Davies 1981; Moore and Webb 2013), as the ascending magma is thought to rise through narrow vertical channels. At present, such a regime may apply to Jupiter's moon Io, which is heated by tidal friction (e.g., Tyler et al. 2015). The 'heat pipe' mode may be particularly relevant for the early phases of terrestrial planet evolution (Stern et al. 2018), such as during the Hadean and Archean epochs on Earth (e.g., Kankanamge and Moore 2016). If Venus featured a stagnant-lid throughout all its history, it could have been in a heat pipe mode during some phases, too (e.g., Turcotte 1989). On modern Venus, there is also evidence for recent and possibly active volcanism (e.g., Smrekar et al. 2010; Bondarenko et al. 2010, D'Incecco et al. 2017, 2021; Campbell et al. 2017; Byrne 2019), but its rate is not well-known.

Heat-piping leads to large volumes of extrusive volcanism, but not to large-scale horizontal motion as in tectonically mobile regimes. The ability of rising magma to reach the surface strongly depends on its buoyancy and overpressure, but also on the strength of the lithosphere and crust. A thick and strong crust prevents magma from reaching the surface and could instead promote the emplacement of magmatic intrusions within the crust. On the other hand, Io features the largest volcanic heat flow in the solar system and may have a thick strong crust, but its lower part may be substantially weakened by magmatic emplacements which could consume up to 80% of the total magma delivered to the crustal base (Spencer et al. 2020). Intrusion could be the dominant mode of terrestrial magmatism as several proposals have been made that only 10–20% of magmatism is extrusive (Crisp 1984; Cawood et al. 2012), but such ratios are spatially highly variable and sensitive to the detailed tectonic setting and structure of the crust (e.g., White et al. 2006). For other planets including Venus, such details are typically unknown, leaving the ratio of extrusive and intrusive magmatism as a major unknown.

If hot crustal intrusions are sufficiently abundant, however, they weaken the surface lid from inside and can lead to yet another tectonic regime, the 'plutonic-squishy lid' (Lourenço et al. 2018, 2020). This regime is characterised by a strong lithosphere that is fragmented into a set of small tectonic units by warm and weak regions caused by plutonism (Fig. 2d). The lithosphere is much thinner and more mobile than in the 'heat pipe' and 'stagnant lid' regime (though not as mobile as in the mobile regime, thus $M < 1$ and $M_{PSL} < M_{ML}$) and deformation is expected to be more localised (thus $P_{PSL} > P_{SL}$). The Archean Earth may have displayed such a regime, but it could also be relevant for Venus given this planet's hot and therefore soft lithosphere (e.g., Gerya 2014; Byrne et al. 2021) and the hints of lateral motion documented in present tectonic features without evidence for Earth-like subduction (Sect. 3.1).

The regimes discussed here may not capture all possible behaviours. Sub- and mixed regimes may exist (Loddoch et al. 2006; Rozel et al. 2015). Moreover, it is challenging to interpret a regime with regards to the state of the mantle: is the displayed regime a snapshot representing the current state of the mantle or an accumulated consequence of planetary evolution of hundreds of millions to billions of years? Describing a planet using the single geodynamic regime observed today seems infeasible in most cases, including Venus (Sect. 4). Determining how and when regime transitions occur is important for interpreting the preserved geological record. One challenge is that evidence from earlier regimes can be (and, on Venus, likely have been) overprinted by signatures from the modern regime. Also, transitions may not be clearly distinct events, but stretch over long time scales (e.g., Weller and Kiefer 2020) and possibly lead to different, coexisting tectonic styles for different parts of a planet's surface (e.g., Robin et al. 2007; Capitanio et al. 2019).

3 Present-Day Mantle Dynamics on Venus

3.1 Observational Constraints

This section briefly reviews the available constraints with the greatest relevance for Venus' deep interior. A more detailed account on these constraints is given in this collection by Ghail et al. (2023), Gilmore et al. (2023), and Herrick et al. (2023).

3.1.1 Gravity and Topography

In the absence of seismic data, gravity and topography are the most powerful constraints on Venus' interior – and at shorter wavelengths – on the strength of the lithosphere and crustal thickness variations. The long–wavelength (i.e., below spherical harmonic degree 40) gravitational and topographic response of a mantle density anomaly depend on its scale, density contrast, depth in the mantle, and on the stress propagation from there through the crust and lithosphere to the surface. Thus, gravity and topography provide key inferences on the viscosity structure of Venus' interior at long wavelengths (e.g., Hager et al. 1985; Kiefer et al. 1986; Zhang and Christensen 1993; Rudolph et al. 2015) and ultimately on the planet's tectonic regime (e.g., Steinberger et al. 2010; Huang et al. 2013).

From the Magellan mission, Venus' topography is available at 10–25 km horizontal resolution and a nominal vertical resolution of 80 m that strongly depends on local topographic gradients (Pettengill et al. 1992). The Magellan gravity field is on average resolved at $\sim$270 km (corresponding to spherical harmonic degree 70, Konopliv et al. 1999), but resolution varies from 170–540 km (spherical harmonic degrees $\sim$35–110) across the surface. Future missions aim to deliver higher resolution (Widemann et al. 2023). As discussed below, gravity and topography analyses can be used to estimate elastic thickness in many areas, but the low resolution of the gravity data results in larger errors (e.g., Anderson and Smrekar 2006). Within these errors, there is evidence for significant variations in crustal and elastic thickness over short spatial scales over much of Venus. Short–wavelength variations may have crustal sources (Fig. 4, Steinberger et al. 2010; Benešová and Čížková 2012).

Lacking seismological data, internal density anomalies are poorly known for Venus, which complicates interpretation of surface gravity with respect to mantle viscosity structure. To overcome this, Steinberger et al. (2010) assumed that Venus' mantle density anomalies are statistically similar to Earth's. Other authors used numerical models to predict synthetic density distributions consistent with Venus' evolution (e.g., Pauer et al. 2006; Armann and Tackley 2012; Orth and Solomatov 2011, 2012; Benešová and Čížková 2012; Huang et al. 2013; King 2018; Rolf et al. 2018b). For these approaches, the spectral representation via a gravity power spectrum is particularly useful (Fig. 4), as models cannot be expected to fit the observed signals from Venus' interior directly, but make predictions with statistically similar amplitudes and scales.

3.1.2 Crater Statistics

Topography and gravity constrain the current state of Venus' interior, but reveal little temporal information. Such insights require reconstruction of the planet's surface chronology, which—lacking dating of Venus' crustal rocks—is mostly based on cratering statistics and the relative age of geologic features. Compared with the stagnant-lid bodies—Mars, Mercury, and the Moon—Venus displays much fewer craters whose distribution taken alone is

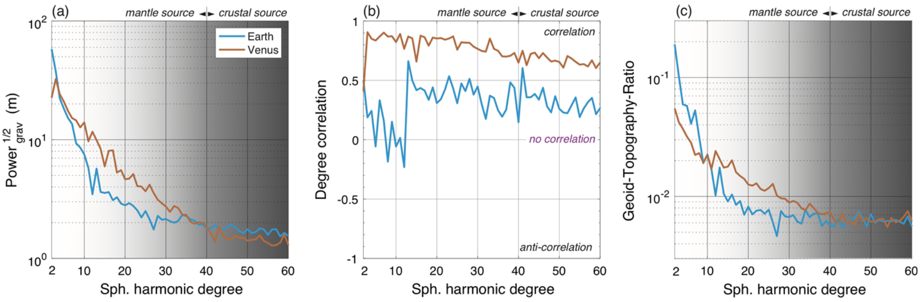

Fig. 4 Power spectra of Venus' present (**a**) surface gravity field, (**b**) degree correlation between gravity and topography, and (**c**) geoid–topography ratio. Power spectra are as defined by Steinberger et al. (2010). The respective spectra for Earth are shown in blue for comparison. In (**a**) and (**c**) the background shading indicates which parts of the spectra may dominantly have deep/mantle (bright) and shallow/crustal sources (dark) according to Steinberger et al. (2010). The gravity model SHGJ180UA01 is used (https://pds-geosciences.wustl.edu/mgn/mgn-v-rss-5-gravity-l2-v1/mg_5201/gravity/). Topography data is obtained from Venus' shape given by Wieczorek (2007) (https://github.com/MarkWieczorek/web/tree/master/spherical-harmonic-models-topography).

not distinguishable from random (Phillips et al. 1992; Strom et al. 1994). The planet's dense atmosphere tends to obliterate small meteoroids, leading to a lack of craters with diameters below ∼2 km (Herrick and Phillips 1994). Considering this atmospheric screening, Venus' craters suggest a young age of ∼150–1000 Myr (Phillips et al. 1992; Strom et al. 1994; McKinnon et al. 1997; Herrick and Rumpf 2011; Le Feuvre and Wieczorek 2011). Overall, there is no consensus on the average surface age, the age of various geological units (e.g., Kreslavsky et al. 2015), the rate of surface weathering or the degree of volcanic embayment of impact craters (Herrick et al. 2023).

In broad terms, accommodating the relatively uniform surface age can be done by the end members of completely wiping the surface clean of craters every few hundred Myr or by having resurfacing occurring everywhere at similar rates but with different mechanisms (e.g., Romeo and Turcotte 2010; Bjonnes et al. 2012). These end members have different implications regarding the long-term evolution of the mantle. The former 'catastrophic' resurfacing—singularly or recurring episodically—implies that mantle temperature and other conditions are not smoothly evolving with time. Global-scale excursions occur, with rapid periods of cooling during a resurfacing episode. The episodes are separated by longer quiescent periods during which the mantle temperature decreases only slowly or even increases due to radiogenic heating (e.g., Turcotte 1993; Nimmo 2002; Armann and Tackley 2012; Rolf et al. 2018b). The latter 'equilibrium' case has a global resurfacing rate (which can be accommodated with different mechanisms) that implies enough surface disruption caused by the interaction between mantle, lithosphere, and crust to eliminate at least several hundred million years of Venus' cratering record, regardless of surface spatial location relative to mantle convection pattern (Herrick et al. 2023). Apart from these end members, the scenario best supported by both the impact crater record (Phillips et al. 1992) and by the removal of extended impact ejecta blankets (Phillips and Izenberg 1995) is regional equilibrium resurfacing. This scenario allows local resurfacing patches on the scale of 100s to ∼1000 km to occur in different locations at different times (O'Rourke et al. 2014). This scenario is consistent with the scale of a variety of volcanic features, including plume-induced subduction (Davaille et al. 2017).

3.1.3 Crustal and Lithospheric Thickness

The thermal state of Venus' interior is tied to the thickness of the lithosphere via conductive heat transfer. It also sets the conditions for the occurrence of magmatism and volcanism (partly) determining crustal thickness. To date, no direct measurements of lithospheric and crustal thickness have been made, but indirect estimates exist (Ghail et al. 2023). Absent seismic data, crustal thickness is typically inferred from inversion of surface gravity and topography (e.g., Maia and Wieczorek 2022). Estimated mean thicknesses vary greatly, from ∼10 km (James et al. 2013), to ∼20–25 km (Jiménez-Díaz et al. 2015; Maia and Wieczorek 2022) to as much as ∼60 km (Steinberger et al. 2010). Lateral variations are substantial, as Anderson and Smrekar (2006) suggested thickness ranges from 0 to ∼90 km, whereas James et al. (2013) reported a range of ∼5–60 km (Fig. 5a-b). However, such values depend on the non-unique choice of a mean thickness and the assumption that no crust should be thicker than ∼70 km–the depth at which Venus' basaltic crust transitions into eclogite. Dense eclogitic crust may be recycled through delamination, unless the crustal root grows too quickly, which could be the case for Maxwell Montes (Namiki and Solomon 1993).

For models of mantle convection, crustal thickness analysis is most easily incorporated in spectral form (Fig. 5c). Wei et al. (2014) combined observed admittance spectra of gravity and topography with convection models to infer which parts of the spectra arise from mantle and from crustal sources and estimated a crustal thickness range of 28–70 km. Thicknesses exceeding 50 km are limited to Ishtar Terra, Ovda Regio, and Thetis Regio, where the gravity and topography signatures suggest isostatic compensation via a thick crustal root (e.g., Smrekar and Phillips 1991; James et al. 2013). In addition, Yang et al. (2016) separated Venus' gravity and topography into dynamic and isostatic components to derive a crustal thickness range of 12–65 km. However, as crustal growth and destruction are strongly tied to volcanism, magmatism and convective processes acting on the crustal base, the long-term evolution of the interior can impose substantial variations of Venus' crustal thickness through time (Sect. 4).

The thickness of the lithosphere, in which the crust is embedded, is defined based on either its elastic strength or temperature profile (Ghail et al. 2023). The thermal thickness is given by the thickness of the conductive thermal boundary layer and depends on the (unknown) temperature and viscosity of the interior. The geoid and topography can be interpreted as indicating average thickness of 200–400 km, possibly less below the volcanic highlands (Solomatov and Moresi 1996; Moore and Schubert 1997). If geoid and topography are explained purely by thermal isostatic adjustment of Venus' stagnant lid, the thermal thickness may be as great as 600 km (Orth and Solomatov 2011). However, a thinner lithosphere (100–150 km) may be more compatible with melt generation rates estimated at Venus' hotspots (Smrekar and Parmentier 1996; Nimmo and McKenzie 1998). The elastic thickness is always less than the thermal thickness and can be estimated from flexural models applied to geological features on Venus (Solomon and Head 1990; Johnson and Sandwell 1994; O'Rourke and Smrekar 2018; Borrelli et al. 2021) and from global admittance maps (Anderson and Smrekar 2006). For most of the planet, the elastic thickness could be as little as 20 km (Anderson and Smrekar 2006), which may indicate a warm lithosphere possibly promoted by intrusive magmatism (Lourenço et al. 2020; Plesa and Breuer 2021) and plume–lithosphere interactions (Gülcher et al. 2020). Admittance for tessera plateaus can be interpreted as evidence of thin elastic lithosphere, reflecting the presumed more ancient time of tessera formation (Maia and Wieczorek 2022), or as simply Airy compensation due to a thick crust (Anderson and Smrekar 2006). Topographic fitting to flexural models can be

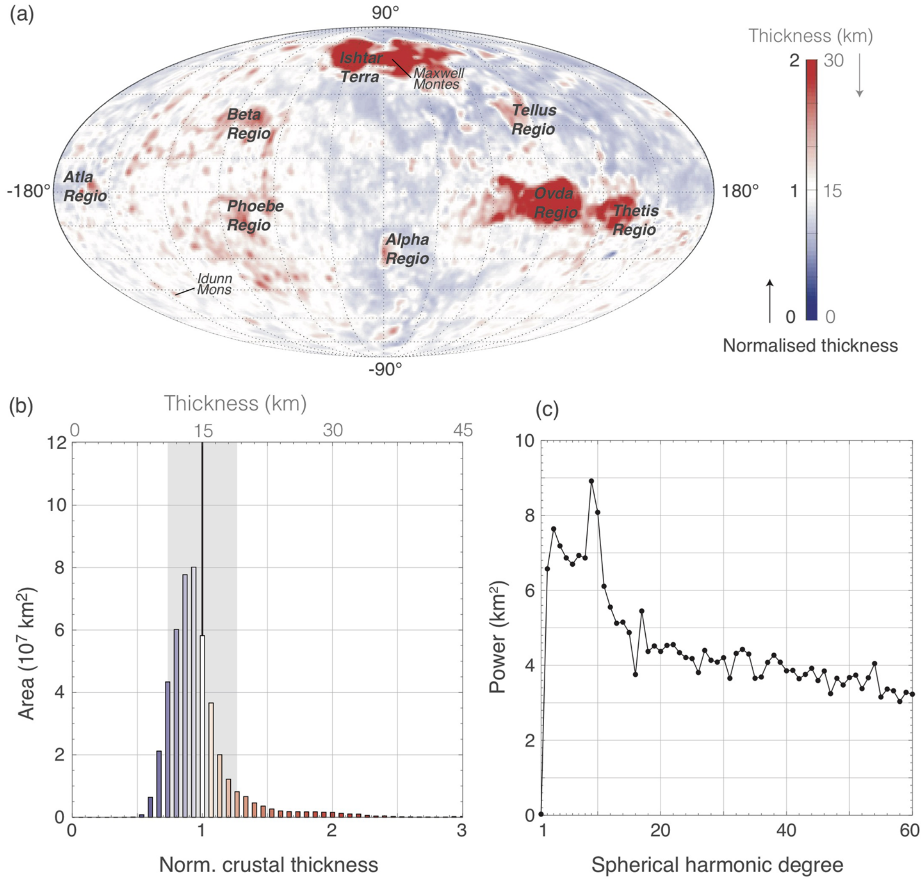

Fig. 5 (**a**) A Mollweide projection of inferred crustal thickness on Venus, centred at 0°E (data from James et al. (2013), assuming a mean thickness of 15 km and a mantle load depth of 250 km). Red and blue indicate higher- and lower-than-average thickness, respectively. (**b**) A histogram representation of panel (**a**). The black vertical line denotes the mean value; the grey-shaded region is bounded by one standard deviation. (**c**) A power spectrum of crustal thickness variations for spherical harmonic degrees 1–60

used to estimate elastic thickness from which heat flux can be inferred (e.g., O'Rourke and Smrekar 2018), based on assuming a particular rheology and strain rate.

3.1.4 Heat Flux and Thermal Emissivity

Venus' average surface heat flux is typically assumed to be smaller than Earth's (∼80 mW/m^2, Davies 2013). Mantle convection simulations in the stagnant lid regime promote low estimates of 10–40 mW/m^2 (Solomatov and Moresi 1996; Gillmann and Tackley 2014; Rolf et al. 2018b; Uppalapati et al. 2020), but episodic overturns may cause a pronounced temporal increase (Armann and Tackley 2012; Gillmann and Tackley 2014; Rolf et al. 2018b; Uppalapati et al. 2020). These average fluxes do not reflect spatial variations. Based on viscoelastic relaxation models, Karimi and Dombard (2017) suggest a higher-than-average flux of 55–90 mW/m^2 at Mead Crater, although the applicability of their model assumptions has been challenged (Ruiz et al. 2019). Impact-crater formation models indicate that Mead's multiring-structure implies much lower flux (<28 mW/m^2, Bjonnes et al. 2021). In contrast, many of Venus' coronae may feature high fluxes (>95 mW/m^2, O'Rourke and Smrekar 2018), whereas domes in the proximity of coronae may display intermediate fluxes (Borrelli et al. 2021). The spatially varying estimates imply a thermally heterogeneous upper mantle and/or lateral thickness variations of the crust and lithosphere, which numerical models of Venus' mantle dynamics can shed further light on (Sect. 4.3).

In addition, thermal emissivity provides constraints on recent volcanic activity and therefore on the thermal conditions in the interior that may make magmatism feasible. Emissivity measurements for Venus' southern hemisphere by Venus Express revealed a number of regions of anomalously high emissivity. These high values do not imply thermal anomalies, but rather are consistent with fresh, unweathered basaltic composition (Smrekar et al. 2010). Each of these features had previously been identified as having the broad topographic rises and major volcanoes analogous to terrestrial hotspot features (McGill 1994). The gravity anomalies for nine such features are interpreted as indicative of active mantles plumes (Smrekar and Phillips 1991; Kiefer and Hager 1991; Smrekar 1994), which offers insight into the planform of Venus' mantle flow (e.g., Huang et al. 2013; Rolf et al. 2018b). The presence of high emissivity and gravity anomalies at both large-scale rift-dominated and corona-dominated features is consistent with current activity at small and large plumes (Smrekar et al. 2010), rather than requiring different convective regimes to allow their formation (Jellinek et al. 2002). The inferred number of hotspots places bounds on mantle viscosity ($\leq 10^{20}$ Pa s, if the mantle is mostly internally heated) and core temperature ($\geq$1700 K, Smrekar and Sotin 2012). Future data may reveal evidence of additional locations of recent volcanism and find evidence of additional small-scale plumes at depth, thus providing further insights into mantle temperature and planform.

3.1.5 Surface Tectonic Features Linked to the Deep Interior

Observations of the distributions, types, and spatial and temporal relations of tectonic and volcanic landforms provide important constraints for models of Venus' interior. These constraints include estimates of surface strains, areas of crustal shortening and extension, regions of stratigraphically young volcanism, and the extent and distribution of likely active surface features (see Ghail et al. 2023). A key question for this chapter is to what extent are interior processes manifested at Venus' surface? Gravity–topography admittance ratios obtained during the Magellan mission indicated that some portions of the planet are dynamically supported. For example, the large volcanic rises – such as Beta, Atla and Themis Regiones – and several smaller rises are consistent with their being supported by large mantle upwellings (e.g., Smrekar and Phillips 1991). It is notable that the volcanic flows hosting geochemical evidence of incomplete weathering (e.g., Smrekar et al. 2010; Brossier et al. 2020) are those interpreted to be comparable to hotspots on Earth (Smrekar 1994).

Another issue is whether mantle forces drive large-scale horizontal motion on Venus? There is no morphological evidence for Earth-like oceanic plate movement and convergent (rather than roll-back or retrograde) subduction on Venus today. Nonetheless, because of the high surface temperature, there may be a weak layer within the lower crust or upper mantle, akin to the rheological layering in continental lithosphere on Earth (Buck 1992; Ghail 2015). If so, tractions from mantle flow on this low-strength layer may drive surface deformation (e.g., Leftwich et al. 1999). Rheological data for Venus' lithosphere is sparse, but there is a geological basis for interpreting lateral motions on the planet. For example, Harris and Bédard (2013, 2014) documented evidence for Lakshmi Planum having collided with Ishtar Terra. Their work proposed that, akin to how continents move on

Earth, mantle flow on Venus may have pushed against a deep lithospheric keel of Lakshmi Planum, driving it northwards over perhaps 1000s of kilometres. Similar deep-seated mantle flow may have led to major shear displacements between Ovda and Thetis Regiones. Recent work has suggested more modest horizontal motions, too, again potentially driven by mantle flow. Numerous portions of Venus' lowlands feature intersecting bands of extensional and shortening structures (termed 'groove belts' and 'ridge belts', respectively) that delineate smooth plains-filled lows (Byrne et al. 2021). Many such intersecting belts show evidence of lateral displacements and paint a picture of these plains-filled lows being mechanically coherent crustal blocks that have moved with respect to one another. Calculations from gravity-induced mantle flow show that the tractions arising from mantle motion today may transfer sufficient force to the surface at every location where these blocks have been observed—consistent with this motion having taken place geologically recently, and perhaps even ongoing (Byrne et al. 2021).

Various modelling studies have focused on the formation of Venus' prominent coronae, which are thought to have been formed by mantle upwelling impinging on the lithosphere (Stofan et al. 1992). The observed variety of their morphological forms can be explained by the spectrum of development of individual coronae (e.g., Stofan et al. 1991; Smrekar and Stofan 1997; Koch and Manga 1996; Hoogenboom and Houseman 2006; Gülcher et al. 2020). Models of plume upwelling generally evolve from domes to depressions; the opposite is true for models of corona formation above dripping or delaminating lithosphere (Hoogenboom and Houseman 2006; Piskorz et al. 2014). For coronae that form over small upwellings, a major unresolved question is the origin of the upwelling and their relationship to circulation patterns in the mantle. Are they shallow upwellings or do they rise up from the core–mantle boundary, like classical plumes on Earth (e.g., DePaolo and Manga 2003, French and Romanowicz 2015) and those inferred for large scale (1000-2000 km) features such as Atla Regio (e.g., Smrekar 1994)?

3.1.6 Magnetic Field

Pioneer Venus Orbiter provided the upper limit on the intrinsic magnetic field: any magnetization is $\leq 10^{-5}$ times weaker than Earth's magnetic field today (Phillips and Russell 1987). Other missions have failed to detect any intrinsic magnetism and this apparent dearth of signal has been used to exclude the existence of a magneto-hydrodynamic dynamo in a convecting, electrically conductive core of Venus (see Gillmann et al. 2022) Theory predicts that a dynamo inside Venus would be apparent in available data if it existed (e.g., Stevenson 2003, 2010).

The lack of an intrinsically generated magnetosphere today may inform models of mantle convection. Core convection is primarily affected by core cooling, which is determined by mantle heat transfer. The described (absence of) observation is most useful assuming an Earth-like core for Venus that is partially liquid and its liquid part is chemically homogeneous. Then, the heat flow across the core–mantle boundary today is below the critical value required to drive core convection (e.g., Labrosse 2015; Nimmo 2015). If Venus has no solid inner core, the dynamo has to be sustained by thermal convection alone and the critical value equals the heat flow along the core adiabat. Depending on the core's thermal conductivity, this could range from $\sim$5–15 TW (see Lay et al. 2008 and references therein). With a solid inner core, the latent heat of freezing and the gravitational energy release associated with the partitioning of light elements into the liquid following inner core growth provide additional sources of power and lower the threshold for a dynamo to roughly half the total adiabatic heat flow (Blaske and O'Rourke 2021).

However, using these arguments to constrain Venus' modern core heat flow is problematic, because Venus' core may not be like Earth's. The estimates for the moment of inertia (Spada et al. 1996; Margot et al. 2021) and tidal Love number (Konopliv and Yoder 1996) are too inaccurate even to rule out that Venus' core is fully solid or has a liquid layer too thin to sustain a dynamo (Dumoulin et al. 2017). A fully solidified core still appears unlikely with at least some fraction of sulphur (S) in the Venus core as this would imply core temperatures close to the Fe-FeS eutectic. These are likely too low (e.g., Boehler 1998) to allow for partial melting of the mantle that is prerequisite for volcanic activity. However, core heat flow also depends on heat transport efficiency on the mantle side and if Venus' mantle plumes are merely hot thermals (Jellinek et al. 2002), they may extract as little heat as 3 mW/m^2 from the core. That is an integrated heat flow of less than 0.6 TW for the range of possible core radii (3500 ± 500 km, Margot et al. 2021).

3.2 Which Geodynamic Regime for Present Venus?

A key question is which tectonic regime matches the observational constraints summarised best? The mobile lid regime implies a comparably thin lithosphere as well as efficient heat transfer through the surface. The typically suggested reduced heat flux on Venus with respect to Earth then points to a thicker lithosphere favoured by less vigorous mantle flow and a more viscous mantle. However, predictions of Venus' modern geoid from mantle flow models (e.g., Steinberger et al. 2010; Rolf et al. 2018b) do not point to systematically higher mantle viscosities. Subduction—the ultimate characteristic of mobile lid convection—cools the mantle efficiently, and a colder mantle triggers less volcanism, confined to regions underlain by hotter-than-average mantle (see e.g., Bondarenko et al. 2010; Smrekar et al. 2010; Smrekar and Sotin 2012). Although active subduction does not occur on Venus today, there are indications for subduction-like processes (e.g., Schubert and Sandwell 1995), either in the form of localised retrograde (Sandwell and Schubert 1992) and/or as plume-induced subduction (Davaille et al. 2017; Zampa et al. 2018). A continuous network of divergent and convergent plate boundaries is still lacking, so that the indication of subduction processes does not imply planet-wide mobile lid tectonics (but see Sect. 4.3). Byrne et al. (2021) argue that in many locations Venus' present lithosphere appears fragmented into dozens of coherently moving crustal blocks – like tectonic plates, albeit the largest identified block is only $\sim 1.9 \times 10^6$ km^2 and thus much smaller than any major tectonic plate on Earth. The small size could be an expression of merely crustal deformation processes rather than fully-developed mobile lid tectonics. This in turn suggests smaller horizontal forces to drive lateral motion of plate-size structures, and has consequences for the rheology of Venus' crust and upper mantle.

On Earth, a low-viscosity asthenosphere is evident in the post-seismic deformation of large earthquakes (e.g., Hu et al. 2016) and known to promote long-wavelength flow in the upper mantle and large tectonic plates (e.g., Höink and Lenardic 2008). On Venus, such evidence is lacking and the asthenosphere may be less pronounced—in particular if its strength is related to water content (Green et al. 2014; Masuti et al. 2016) and Venus' interior is relatively dehydrated. The lacking or weakly pronounced asthenosphere is manifested in the high correlation of Venus' gravity and topography (Fig. 4b), which can also be expressed via the spectral admittance ratio (e.g., Kiefer et al. 1986). At long wavelengths (spherical harmonic degrees < 10), the Earth features positive geoid and negative topography, leading to negative admittance. On Venus, in contrast, the admittance is positive, which requires stronger coupling of the mantle and the lithosphere, hence the absence of a pronounced asthenosphere (e.g., Kiefer et al. 1986; Steinberger et al. 2010; Rolf et al. 2018b). This absence is also supported by the large geoid anomalies associated with Atla and Beta Regiones that require stronger coupling of upwelling plumes to Venus' surface layer (Smrekar and Phillips 1991). In contrast, deformation experiments on crustal plagioclase support a strong rheological contrast between crust and mantle at Venus' Moho conditions. This promotes a decoupling of both layers, which reduces the magnitude of mantle tractions transmitted to the crust and impedes large-scale lateral motion (Azuma et al. 2014; Katayama 2021). Smaller-scale surface mobility is still feasible, especially if facilitated by a weak lower crust (e.g., Arkani-Hamed 1993; Ghail 2015; Byrne et al. 2021). A weaker lower crust may however have difficulty to support some of Venus' highest topography, so that this feature may not be global.

Venus' crater population supports the infeasibility of a large-scale mobile plate-like regime today as it would efficiently renew the surface, deform, and erase craters. Although Venus' cratering record supports a geologically young surface, the mobile-lid framework has difficulty accounting for the apparent random distribution of Venus' craters and the implied age uniformity (Sect. 3.1.2). For example, terrestrial tectonic reconstructions since 200 Ma indicate substantial variations in surface age, even after ignoring the presence of anomalously old continents (e.g., Coltice et al. 2013). Finally, continuous subduction of surface material would foster heat transport from Venus' core into the mantle making dynamo action and inner core nucleation more likely. However, other aspects—such as stable stratification of the core—could explain the lack of dynamo action even if the mantle were in a mobile-lid regime today (Smrekar et al. 2018).

An episodic lid regime with an ongoing resurfacing episode is difficult to distinguish from the mobile lid regime given the (intermittent) surface motions, but peak rates of surface velocities and eruption rate are likely higher during short episodes than in a long-term stable mobile lid. If resurfacing happens via one major zone of convergence as promoted by some numerical models (e.g., Karlsson et al. 2020), the deep mantle density distribution becomes heterogeneous at hemispheric scale and causes a too large offset between Venus' centre of mass and centre of figure (Bindschadler et al. 1994; King 2018). If the resurfacing process operates on localised, regional scales, this issue is less problematic, but still a sudden increase in volcanism would occur due to the lithospheric thinning following the onset of resurfacing. This is challenging to accommodate with the apparent rates of recent volcanism on Venus. An ongoing resurfacing episode on Venus is thus similarly unlikely as a continuous mobile regime. However, resurfacing episodes happen with undetermined frequency (e.g., Uppalapati et al. 2020), and are separated by long stagnant-lid periods especially during mature stages of the planet's evolution (Armann and Tackley 2012). Therefore, a current quasi-stagnant state with long tectonic quiescence between episodes is difficult to rule out based on available data (Rolf et al. 2018b). During a long-lasting stagnant lid state, the mantle becomes mechanically decoupled from the shallow lid because of the high viscosity contrast between mantle and crust. The mantle does not exert enough forces on the lid to coherently move it horizontally in a plate-like fashion. Under this regime, surface heat loss is greatly reduced, but it remains unclear how such a reduction accounts for locally elevated fluxes, such as suggested for coronae structures (Sect. 3.1.4). Convection-induced deformation in a subcrustal lid with a weak lower crust (Ghail 2015; Byrne et al. 2021) and plumes impinging and eroding the lithospheric base (Smrekar and Stofan 1997; O'Rourke and Smrekar 2018; Gülcher et al. 2020) together could provide an explanation.

With inefficient heat loss during a stagnant-lid period, Venus' mantle would have difficulty losing its heat. If such a period is established after previous extended periods of mobile lid tectonics, the mantle even heats up – despite the background trend of decaying radiogenic heat sources – until the mantle temperature has adjusted to the rate of radiogenic heat production. During the heating process, the mantle and core slowly equilibrate thermally, decreasing heat transfer and making a core dynamo less feasible. Moreover, the upper mantle heats up towards its solidus, resulting in substantial magmatism and volcanism. Moderate rates of localised volcanic resurfacing are not at odds with the relatively uniform surface age inferred from Venus' crater distribution (e.g., Kreslavsky et al. 2015). With purely volcanic resurfacing, the uniform surface age requires lava flows to be distributed broadly across the surface, as indeed indicated on geological maps of Venus' surface (e.g., Ivanov and Head 2011). In the absence of cold sinking slabs, lateral variations in mantle temperature may be small in the stagnant-lid regime, so that a global layer beneath Venus' lithosphere could be partially molten and feed spatially random volcanism. Whether this form of equilibrium resurfacing is a feasible scenario to generate a uniform surface age consistent with the crater distribution remains debated (e.g., Romeo and Turcotte 2010; cf. Bjonnes et al. 2012). However, O'Rourke et al. (2014) show that scales of volcanism of several 100 to $\sim$1000 km are consistent with the crater population.

The relevance of the stagnant lid for Venus also depends on this regime's definition. Often defined by small surface mobility, it is undefined what 'small' means. Although less than in the mobile lid, some mobility is permitted, in particular if the main manifestation of the stagnant lid is the dominance of heat transport via conduction through the lithosphere (Byrne et al. 2021). Also, magmatic processes potentially induce mobility (e.g., Noack et al. 2012; Lourenço et al. 2016, 2020), depending on the efficiency of magma eruption. Heat pipes can operate at any non-zero eruption efficiency, but at high efficiency less magma production and therefore a lower temperature is required in the mantle to make heat piping the main mode of planetary resurfacing. The estimated rates of volcanic activity on modern Venus do not reflect a dominance of heat piping. The prospect of a heat-pipe regime is further challenged by the estimates of Venus' crustal thickness, which typically suggest an average crustal thickness of only a few 10s of km (Sect. 3.1.3), whereas evolution models of Venus' mantle with maximum eruption efficiency—resembling the heat-pipe regime—typically predict much larger global crustal thicknesses of $\sim$100 km (Armann and Tackley 2012; Rolf et al. 2018b). The high volcanic fluxes required to form such thick crust are difficult to reconcile with the age of Venus' present crust (e.g., Turcotte 1989). Reduced magma eruption efficiency ($<$20%) can substantially decrease the thicknesses of the present crust in the stagnant lid regime (Fig. 6). At this point, crustal intrusions control the thickness and strength of the crust so that the plutonic-squishy lid can be entered (Sect. 2.3; Lourenço et al. 2020).

In this regime, Venus' groove and ridge belts serve as the regions where a lot of magma is intruded into the crust, coexisting with relatively coherent crustal blocks (Byrne et al. 2021). Recent volcanism then represents the relatively small extrusive portion of magma that makes it to the surface, but the total volume of magmatism is much higher, so that the rate of volcanism is not directly linked to the thermal state of the upper mantle. Extrusions are more likely where the rising magma has anomalously high buoyancy or where the integrated compressive strength of the crust is low. On Venus, extension in a subcrustal lid around regions of major mantle melting is in line with regional topography variations (Ghail 2015) and may keep the compressive strength low to facilitate the rise of magma to the surface. Idunn Mons may be an example for such a region (D'Incecco et al. 2017). The ubiquitous presence of novae, coronae, and other volcanic features (e.g., Stofan and Smrekar 2005) is another hint to such a spatially heterogeneous regime in which tectonic and volcanic resurfacing interact and widespread volcanism is consistent with a relatively uniform surface age on scales $>$1000 km (O'Rourke et al. 2014). Most of the buoyant magma would intrude the crust, keeping it warm and sufficiently weak for crustal flow and surface deformation required to eliminate crater signatures.

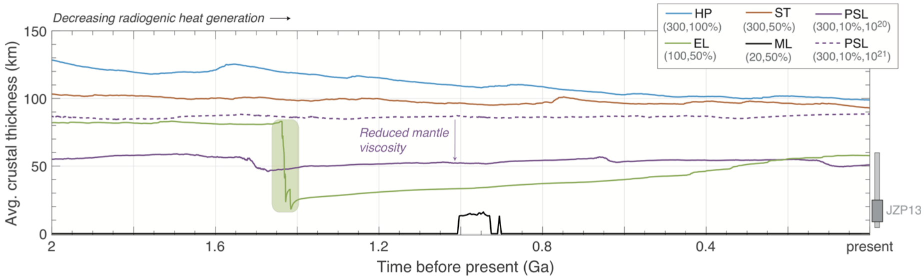

Fig. 6 A time series of mean crustal thickness during 2 Gyr of model evolution in different geodynamic regimes (data from Lourenço et al. 2020). HP: Heat Pipe; ST: Stagnant Lid; PSL: Plutonic Squishy Lid; EL: Episodic Lid; ML: Mobile Lid. Values in parentheses indicate the surface yield stress in MPa, and the magmatic eruption efficiency in %. In the EL model, the shaded area indicates a large-scale overturn episode. The grey bars at the right margin of the plot indicate estimates of Venus' present crustal thickness (James et al. 2013, dark: mean, bright: range). All cases assume a mantle viscosity of 10^{20} Pa s, except one (dashed, 10^{21} Pa s)

In summary, Venus' present dynamic regime remains hard to be defined with the available data. Observations and predictions by numerical models can be combined, but a complete answer to how Venus loses its interior heat and causes deformation of the lithosphere and crust does not yet exist; future missions will provide new data that will help to resolve this issue (Sect. 5.2). For the time being, a regime with a globally mobile lid can be ruled out – this includes, but in a more general way, the Earth's specific regime of plate tectonics – and so can be an ongoing (or recently ceased) global overturn event in the episodic lid regime. However, Venus is also not in the classical stagnant-lid state without any lid mobility such as Mars, Mercury, or the Moon (see e.g., Tosi and Padovan 2021). Improved estimates of strain on Venus' surface could help to distinguish between the different regime evolutions, as characteristic values tend to differ by orders of magnitude (e.g., Grimm 1994). Another promising avenue invokes the interplay of magmatism and tectonics to generate a hybrid regime, such as the plutonic-squishy lid. The interaction between tectonics and magmatism as well as between mantle and crust vary across Venus' surface due to mantle flow patterns and crustal variations resulting from the cooling and deformation history of the planet. Therefore, understanding Venus' current regime also requires insights on past tectonic states.

4 Venus' Mantle Dynamics Through Time

For Earth, the mantle during the Archean (>2.5 Ga) was several 100 K hotter than at present and operated under a different regime, leading to a different style of tectonics (e.g., Moyen and van Hunen 2012). Some aspects of Venus' current geology may be analogous to those present on early Earth (e.g., Ghail et al. 2023). As discussed in this section, Venus may have experienced similar regime transitions, in particular with a more mobile period in its past. However, details on the style of mobility during that epoch and when it occurred still need to be revealed.

4.1 Shaping Venus' Interior Through Time

The thermal evolution of Venus' mantle is determined by the balance of interior heating and the net flux of heat leaving the mantle (e.g., Smrekar et al. 2018). The net flux is mostly the sum of heat loss from the interior to the atmosphere, heat transfer from the core to the mantle, and heat transfer related to melting and magmatism via latent heat consumption (release) upon melting (freezing), as well as extraction of heat through extrusive volcanism. Internal heat generation comprises mostly radiogenic heating, but can also involve shock heating induced by asteroidal impacts and heating by tidal dissipation (negligible for Venus). Even if the present thermal state, fluxes and heating rates were well known, backward integration of the present heat budget is challenging because the different contributions to the heat budget are nonlinearly coupled and depend on the geodynamic regime and its history (see e.g., Korenaga 2008).

4.1.1 Evolution of the Upper Mantle and Crust

The well-known half-lives of the principle radiogenic isotopes expected for Venus suggest a decay of internal heating by a factor of about three over Venus' lifetime (Turcotte and Schubert 2017). Jellinek and Jackson (2015) proposed that ablation from early impactors on Earth may have caused the loss of an isolated geochemical reservoir with a lower ^{142}Nd/^{144}Nd ratio lower than ordinary chondrites. Such a loss could imply a 20–45% lower radiogenic heat production on Earth compared to Venus, potentially enough to explain their divergent evolution pathways (Weller and Lenardic 2015). Stronger radiogenic heating increases mantle temperature and decreases viscosity, effectively resulting in weaker coupling to the mantle and lower stress in the lithosphere. Therefore, planets operating with high mantle temperatures are more likely in the stagnant lid regime (e.g., Stein et al. 2013; Weller et al. 2015). However, whether ejected enriched material would not re-accrete to the planet and why impact ablation should happen on Earth, but not on Venus has not been discussed by Jellinek and Jackson (2015). Moreover, measurements on enstatite chondrites (Boyet et al. 2018) suggest a nucleosynthetic origin of the Earth's apparently anomalous ^{142}Nd composition (e.g., Burkhardt et al. 2016); removing this effect leads to almost indistinguishable Nd ratios of chondrites and the accessible Earth. This leaves the proposal of Jellinek and Jackson (2015) as a possibility that cannot be disproven, but it is neither sufficiently supported by available data.

More established is that small variations in surface temperature affect the relationship between radiogenic heating and the geodynamic regime, and may have acted as important triggers for transitions in Venus' tectonic regime (Sect. 4.2). Depending on the thermo-tectonic history, the result can be multiple tectonic states, feasible for the same conditions. To date, the link between radiogenic heat production, mantle thermal state, and tectonic regime has mostly been inferred from theoretical scalings and numerical models employing a pseudo-plastic rheology (Sect. 2.2). As discussed, stress imparted to the lithosphere through convection decreases with increasing mantle temperature. Reaching the yield stress to induce plastic failure is thus more complicated and the feasibility of developed mobile-lid tectonics on a younger, hotter planet is reduced. On the other hand, lithospheric stress increases with increasing slope of the lithospheric base (e.g., Fowler 1985; Wong and Solomatov 2015) making it sensitive to the aspect ratio of convection cells. If Venus mantle established sufficiently wide convection cells at some point, initiation of subduction-like processes would have been facilitated (Wong and Solomatov 2015), in particular when the planet was not in the heat-pipe regime anymore (Kankanamge and Moore 2016). Another possibility to enable lithospheric weakening in an earlier, hotter mantle—and thus a challenge to the notion of the prevalence of early hot stagnant lid mantle convection—is via grain-size-dependent damage evolution as the deformational work driving grain size reduction does not decrease in a hotter mantle (Foley 2018). In contrast, an additional argument for the promotion of stagnant lid behaviour independent of rheological arguments may come from the physics of mantle systems heated from below and within. In such systems, high levels of radiogenic heating tend to inhibit convection, and thus stress imparted via internal convective velocities, due to transition in convective planforms from predominantly sheet-like to plume-like (Weller et al. 2016; Weller and Lenardic 2016; Lenardic et al. 2021).

This discussion typically assumes a uniform radiogenic heat production across the mantle, which is not the case in a planetary mantle with melting-induced differentiation. Incompatible radiogenic elements preferentially partition into the liquid phase upon melting, which leads to an enriched crustal layer, impacts mantle cooling history (e.g., Ogawa 2018), and can alter surface heat flux (e.g., Lourenço et al. 2018; Vilella and Deschamps 2021). However, even with extreme partitioning that forces almost all radiogenic elements into the melt, numerical models of Venus' evolution in the stagnant lid—or rather heat pipe—regime cannot produce a thin crustal layer consistent with other estimates for Venus (Armann and Tackley 2012). As discussed in Sect. 3.2, a regime supporting high volumes of intrusive magmatism—such as the plutonic squishy lid—may be more feasible in this regard as it facilitates remixing of radiogenically enriched material into the mantle. The effects on radiogenic partitioning could then be less pronounced than they would be otherwise (Lourenço et al. 2018).

The melting-induced crustal layer features thickness variations reflecting the lateral variations of temperature and flow in the mantle in a time-integrated sense. Crustal heterogeneity locally increases the stress within the lithosphere and facilitates surface mobilisation (Lourenço et al. 2016). This mobilisation is reinforced when the crustal root transforms into eclogite (at ∼60–70 km depth on Venus) as this denser phase induces additional buoyancy and stress to trigger episodic mobilisation of an otherwise stagnant lid (e.g., Rolf et al. 2018b). Thinning the crustal layer by a mobile episode may let the stress in the lithosphere drop below the yield strength, promoting tectonic quiescence. After shutting down recycling, surface heat loss decreases and the mantle gradually heats up, favouring larger volumes of melting, magmatism and volcanism until the crust and lithosphere eventually experience sufficient melting-induced heterogeneity and stress to reinitiate surface mobility. As radiogenic heat production decreases with time, heating up the mantle and powering magmatism takes progressively more time, possibly increasing the interval between surface recycling events until they may eventually fade (e.g., Armann and Tackley 2012; Vesterholt et al. 2021).

The importance of melting-induced heterogeneity and crustal tectonics for Venus depends on the ability of magma to propagate through the crust and reach the surface. Venus is widely covered in basaltic volcanic rocks, so that the magma eruption efficiency is clearly non-zero. Without knowing the total volume of generated magma in the mantle, refined estimates are difficult to make, but important as different eruption efficiencies can lead to different tectonic regimes (Sect. 2.3, Fig. 3). Reducing the eruption efficiency in the stagnant lid regime tends to increase the average age of Venus' crust, but also creates larger spatial age variations (Uppalapati et al. 2020), which may be difficult to reconcile with Venus' crater distribution. Assuming a continuous stagnant lid, dominantly intrusive volcanism could strongly reduce the mechanical lithospheric thickness and better match present elastic thickness estimates for Venus, especially if intrusions are placed at shallow depths (∼50 km, Plesa and Breuer 2021). Magmatic inclusions cool less efficiently than surface lava so that pockets of high melt fractions may be preserved whose evolution is difficult to address in global models (see Abe 1995; Rozel et al. 2017; Lourenço et al. 2018). Moreover, magma eruption varies with the integrated crustal strength and the buoyancy of rising magma, whose ascent through the crust may also be affected by permeability barriers

(Schools and Montési 2018). For all these complexities, the interaction of magmatism and tectonism remains incompletely understood–even in terrestrial settings–and demands future research.

4.1.2 Evolution of the Deep Mantle and Core–Mantle Coupling

The lack of a magnetosphere is a characteristic that any evolution scenario of Venus' mantle has to match (Sect. 3.1.6). However, without sufficient knowledge of the state and structure of the core, including the existence of an inner core, different evolutionary scenarios remain possible (see Smrekar et al. 2018). Speculatively, a basal magma ocean could exist in Venus' lower mantle, thick enough to suppress core cooling, but too thin to support a dynamo by itself today (O'Rourke 2020). Alternatively, Venus could have experienced a 'gentle' accretion relative to Earth. In the absence of mechanical stirring provided by late energetic impacts, Venus' core could have retained a stable primordial chemical stratification (Jacobson et al. 2017). With a stably stratified or solid core, even rapid core cooling would not produce a dynamo and planetary magnetic field.

Realistic predictive models of planetary dynamo generation are still to be developed (see Wicht and Sanchez 2019), however, simplified thermal evolution models of Venus' interior suggest that the prospects for an internal dynamo were more favourable in the past when Venus' interior was hotter (Nimmo 2002; Driscoll and Bercovici 2013, 2014; O'Rourke et al. 2018; Gillmann et al. 2022). Detecting crustal remanent magnetism could indicate that an internal dynamo once operated on Venus. Although the surface of Venus is hot, modern temperatures are still below both the Curie point and the expected blocking temperatures of magnetite, a common magnetic carrier. The Pioneer Venus Orbiter and Venus Express would have detected large magnetised surface regions northwards of 50°S, at least if magnetization is coherent over horizontal spatial scales comparable to the orbital altitude ($\sim$150 km, Russell et al. 2007). However, crustal magnetization may still be undetected on Venus if located near the south pole or if the spatial scales of preserved crustal magnetization at present day are small.

Recent models of Venus' core–mantle–atmosphere coupling (O'Rourke et al. 2018) point out that an initially hot and chemically homogeneous core should remain at least partially liquid today. If Venus has an Earth-like core, the absence of a dynamo is easiest to explain if the thermal conductivity of core material is at the high end of recent estimates (i.e., >100 $\mathrm{W m^{-1} K^{-1}}$). Internal heating in the core and/or dense insulating layers at the base of the mantle could keep the core fully molten. In particular, a basal magma ocean slows down core cooling, but similarly requires slow mantle cooling to avoid solidification. For this, a continuous stagnant lid regime would be more favourable as it keeps the mantle hotter than one (episodically) cooled by active resurfacing. However, the stagnant lid does not rule out a core dynamo in general, with Mercury being a solar system example for such a planet (e.g., Christensen 2006). If melting and magmatism provide a sufficient heat sink for the mantle, a core dynamo may be active on Venus until $\sim$0.3 Ga in the stagnant lid regime with sufficient extraction ($>50\%$) of magma to cool the mantle, but only until $\sim$3 Ga when magma is inefficiently extruded and/or the magma volume is small (Driscoll and Bercovici 2014).

Thermal insulation of the core may also result from compositional layering of the mantle. When entering the lower mantle, basalt becomes denser than olivine, making recycled crustal material denser than the ambient lower mantle and possibly leading to piles of basaltic material atop the core-mantle boundary. Such ponding can happen to some degree in all regimes discussed in Sect. 2.3, but most pronounced in the episodic lid or mobile lid where sufficiently strong, deeply subducting slabs provide an efficient way of downward transport (Lourenço et al. 2020). The presence of such relatively dense piles may prevent Venus' core from solidifying completely (O'Rourke et al. 2018). If originating from the surface and being transported efficiently across the mantle, the settled material may initially be cold. The temperature contrast would temporarily increase heat transfer from the core (King 2018; Rolf et al. 2018b) and possibly enable intermittent thermally-driven dynamo activity. In the longer term, the accumulated dense layer would heat up, in particular when enriched in radiogenic nuclides, insulating the core and suppressing core cooling. Depending on the competition between the thermal and compositional buoyancy, the dense layer may be permanent or remix back into the mantle if not continuously fed by new settling material.

On Earth, most of the dense material atop the core–mantle boundary is organised in two antipodal provinces centred at the equator (Garnero et al. 2016 and references therein). For Venus, seismological constraints are lacking, but the small offset of ∼280 m (Bindschadler et al. 1994) between Venus' centre of mass (CoM) and centre of figure (CoF) rules out strong hemispheric asymmetry in the thickness of the dense layer. This finding seemingly argues against a recent global resurfacing episode, which produces much too large an offset (King 2018). However, the hemispherical-scale variations may distort the offset only for a relatively short time span (∼100 Myr, Fig. 7); afterwards, propagation of active resurfacing zones and the reorganisation of mantle flow may override the hemispherical anomaly in the deep mantle and promote a smaller-scale structure that could have much less impact on the CoM-CoF offset. Clearly, the feasibility of such a scenario depends on Venus' lower mantle properties. Given their uncertainty, the observed small offset may not definitively rule out an episodic resurfacing event on Venus, but at least places bounds on the timing of the latest resurfacing episode: recent cessation less than 150–200 Myr ago seems infeasible. This prediction, in turn, is consistent with the decay of long-wavelength gravity anomalies observed over a time scale of ∼100–150 Myr after overturn cessation (Rolf et al. 2018b). Minimum estimates of Venus' mean surface age are similar (e.g., Herrick and Rumpf 2011; Le Feuvre and Wieczorek 2011), but such a young surface can also be generated without lithospheric overturn when magma eruption efficiency is high (Uppalapati et al. 2020). However, mantle overturn and deep recycling wipe out the pattern of mantle plumes established prior to an overturn episode and re-establishing that pattern tends to take much longer after the cessation of active resurfacing (Rolf et al. 2018b).

4.2 Regime Transitions: Triggers and Time Scales

As emphasised previously, the planetary tectonic regime changes through time in response to the thermal and compositional evolution of the mantle. A relevant question for this paper is what could trigger a regime transition and on what time scale?

4.2.1 Surface Temperature Variations

Being closer to the Sun than Earth, Venus would be expected to receive greater solar insolation, but Venus' much higher albedo ultimately leads to smaller absorption of solar energy, at least at present. Nevertheless, Venus' surface is almost 500 K hotter than the Earth's surface, because surface temperature is controlled by Venus' greenhouse atmosphere (Gillmann et al. 2022). Higher surface temperature weakens crustal rocks due to the strong temperature dependence of viscosity. An increase in surface temperature persisting over geological time propagates through the lithosphere into the mantle, acts to reduce the stress imparted

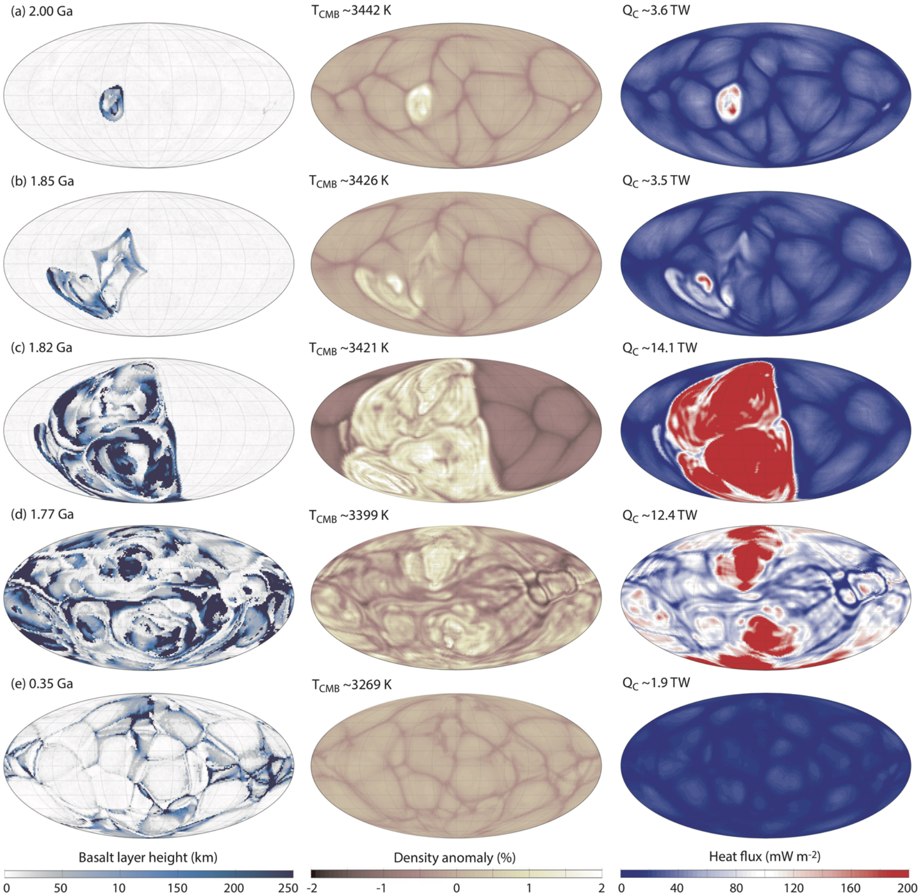

Fig. 7 The evolution of basalt heterogeneity (left column), the deviation from mean density (middle), and heat flux across the core–mantle boundary (Q_C, right) from a 3D thermochemical evolution model of Venus (case 'E50' from Rolf et al. 2018b). Basalt heterogeneity is plotted as the height of the column in which basalt is the dominant composition (i.e., basalt fraction $>50\%$). Each row indicates a different time. Substantial lid mobility is observed between $\sim$1.85 and $\sim$1.65 Ga here. The respective temperature at the core–mantle boundary (T_{CMB}) is indicated

to the lithosphere from the convecting mantle, and may induce transition from a mobile to a stagnant lid regime (e.g., Lenardic et al. 2008; Weller et al. 2015). Moreover, higher surface temperature advances healing of previously accumulated damage, since the growth and recovery of mineral grains is faster. As a consequence, reactivation of previously weakened tectonic structures and the formation of plates may be more complicated on Venus than on Earth (e.g., Landuyt and Bercovici 2009; Foley et al. 2012; Bercovici and Ricard 2014).

Both processes inherently require geological time scales to operate. Relaxing this temporal requirement and assuming that the time scale of surface temperature change is less than the mantle mixing time allows mantle temperatures to remain relatively unchanged. Under this condition, an increase in surface temperature reduces the thermal contrast across the lithosphere. If the associated viscosity contrast falls below a critical value ($\sim 10^5$ as suggested by models and scaling analysis), a previously stagnant lithosphere may enter a transitional regime (Moresi and Solomatov 1998). Although the viscosity contrast across Venus' lithosphere is not well determined, comparison to viscosity estimates of the Earth's crust and asthenosphere could make this situation relevant for Venus. If so, the transition could in turn lead to resurfacing and to a comparably young surface, like that of Venus, without the need for global overturn episodes induced by yielding of the lithosphere (Noack et al. 2012). In contrast, Gillmann and Tackley (2014) concluded that lower surface temperatures can trigger transition to mobile lid behaviour. Such relatively lower temperatures foster volcanic activity thereby potentially increasing water content in the atmosphere, which causes surface temperatures to increase again until the interior transitions back into a stagnant and subsequently into an episodic lid regime. Lateral variations in surface temperature can also induce spatial variations in rheology and–in extreme cases of tidally-locked planets–induce hemispherically different tectonics (Meier et al. 2021), but on Venus balancing by the thick atmosphere keeps such lateral variations small.

4.2.2 Stochastic Triggers

A fundamental question is whether a planet's state can always be categorised by a regime that is distinct from others by a number of diagnostics (Sect. 2.3). This depends on the frequency of regime transitions, but also on whether such transitions are reversible. If melting-induced crustal growth triggers overturn events (e.g., Armann and Tackley 2012; Rolf et al. 2018b; Vesterholt et al. 2021), the rate of magmatism may partly determine overturn frequency. Then, overturns may be more frequent in the early phases of evolution when stronger volcanism facilitates crustal growth, and then feature longer intervals until they eventually cease (Armann and Tackley 2012; Vesterholt et al. 2021). With sufficiently long intervals, the mantle may reach a thermal state representative of the stagnant-lid regime in between overturns, but it may still spontaneously enter another isolated resurfacing episode. Although overturn timing may be unpredictable because of the chaotic nature of mantle convection (Wong and Solomatov 2016), the triggers for such spontaneous transitions can include localised lithospheric thinning (Wong and Solomatov 2015), the merging of several upwellings into a stronger one (Loddoch et al. 2006), plume-induced subduction (Crameri and Tackley 2016), or sub-lithospheric, small-scale convection (Solomatov 2003). These mechanisms together add a degree of stochasticity to the evolution of a convecting planet, on top of that arising from different initial states.

Stochasticity in mantle convection is associated with complex feedbacks. Critical system parameters such as surface temperature, internal heating rate, and yield strength couple nonlinearly to the convective system (Crowley and O'Connell 2012; Lenardic and Crowley 2012; Weller and Lenardic 2012, 2018; Weller et al. 2015; Lenardic et al. 2016). This coupling leads to a hysteresis of states in which otherwise identical parameters lead to non-unique tectonic states (Fig. 8). Within the hysteresis window, stagnant, episodic, or mobile states are equally allowable and stable, with none energetically preferred over another. For vigorous mantle convection, as expected for Venus, the region of multistable states may extend over a wide range of lithospheric yield strengths and surface temperatures (Weller and Lenardic 2018). In this framework, the observed tectonic state is inherently controlled by the specific history of the planet. A planet such as early Venus, which may have been in a mobile lid state before transitioning to a stagnant lid through a surface temperature increase for example, may not transition back to a mobile lid state by a reduction in surface temperature alone (e.g., Weller et al. 2015). With all things held equal, either planetary tectonic state represented by Earth and Venus (State A, B in Fig. 8a) is inherently allowable from the same initial condition. Under the surface temperature regime (Fig. 8b), both Venus and Earth are

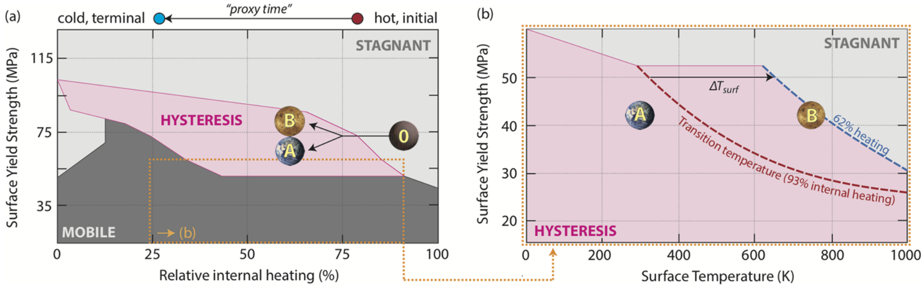

Fig. 8 Tectonic regime diagrams indicating regions of hysteresis (multiple stable tectonic regimes), showing (**a**) the surface yield strength and relative internal heating (the latter is a time proxy: X% indicates that 100-X% of radiogenic heat source are exhausted), and (**b**) yield strength and surface temperature parameter space. The pink threshold lines in (**a**) indicate the yield strength and internal heating combination required to leave an (early) mono-tectonic stagnant lid state 0. With the yield strength held equal, either state (mobile A, stagnant B) is allowable from the same initial state 0. In (**b**), the dashed lines emphasise a widening of the hysteresis window; larger surface temperature changes (ΔT_{surf}) are required to leave the region of multiple states and enter a mono-tectonic state as radiogenics become depleted. Earth and Venus are plotted merely illustratively. These plots are based on Weller and Lenardic (2018)

plotted in the positions of the currently observed surface temperatures, indicating that both planets could currently be within the bistable temperature space. However, this strongly depends on the poorly constrained effective yield strength of the planets' lithospheres. Thus, planets such as Earth and Venus could presently feature different tectonic regimes even under identical present-day conditions, if their surface temperature evolved differently in the past. This finding holds true for other key system parameters such as the yield strength, the global heat budget, radiogenic heating rate, and also for largely stochastic effects occurring in chaotic vigorous mantle convection.

4.2.3 Impact Events

A peculiar class of stochastic events are impacts, which primarily determine the accretion of the terrestrial planets including initial structure, composition, and heat budget. An extremely large, early impact collision could have melted large parts of Venus' early mantle (Davies 2008) and depleted it from most of its water (see Salvador et al. 2023). Water could have been delivered afterwards, but coupled orbital–interior–atmosphere models suggest that such late accretion consisted mostly of relatively dry enstatite chondrites, as otherwise Venus' present atmosphere would be too rich in volatiles (Gillmann et al. 2020).

Apart from water delivery, impacts may trigger changes in mantle dynamics and in interior–atmosphere coupling as reviewed in detail in Gillmann et al. (2022). On Earth, such impacts particularly affected the Hadean and Archean mantle (O'Neill et al. 2017, 2020) which was still hotter and perhaps more comparable to present Venus. Impact energy causes shock heating of the interior and potentially drives magmatic pulses that could trigger resurfacing. The delivered impact energy cannot easily be extracted from the system again, which adds to the discussion of hysteresis above. Under Venus' conditions, a single large impactor could enforce substantial volatile release into the atmosphere, maintaining high surface temperature and promoting a stagnant lid (Gillmann et al. 2016). An otherwise identical evolution lacking such an impact may instead evolve through a period of relatively cold surface conditions that are more prone to lithospheric mobility. On Venus' young surface any detectable relics of such large impacts have been obliterated (e.g., Ruedas and Breuer 2018), but even if preserved, interpreting such anomalies with regards to the triggering of large-scale tectonic events is not unique. As a general problem, the differences between Venus' and Earth's bombardment history are not known well enough to reliably use them as arguments for the diverging pathways of both planets.

4.3 Can We Establish Venus' Geodynamic Regime Evolution?

Tesserae may be among the oldest surfaces preserved on Venus today (e.g., Ivanov and Head 2011; Kreslavsky et al. 2015) and thus possibly important windows into Venus' past. Magellan radar imaging and altimetry reveal the collision of three distinct tessera regions in Tellus Regio (Gilmore and Head 2018), pointing to a phase of surface mobility at some period after the formation of those tesserae. The collision of Lakshmi Planum and Ishtar Terra (Harris and Bédard 2013, 2014; Sect. 3.1.5), and implied 2000–3000 km of crustal convergence may further support such a mobile period (Kiefer 2013). Although ridge belts that possibly formed by crustal convergence over downwelling mantle are ubiquitous on Venus' volcanic plains, they probably have accommodated far less horizontal deformation than tesserae and Ishtar Terra (Moruzzi and Kiefer 2020; Kiefer and Weller 2021), so they may originate from a period with less lithospheric mobility. However, expected regional variations in convergence and deformation rates at the same time would complicate this argument and probably allow for alternative scenarios. In contrast to the evidence of an (undated) earlier mobile epoch, most geophysical evidence points to stagnant lid-like behaviour on Venus at present (Sect. 3.2), which would require a transition in tectonic regime to explain.

Tessera terrains feature characteristic 'ribbons', large-aspect-ratio trough-and-graben structures of debated origin (see Hanmer 2020, and references therein). At least two types of ribbons have been suggested, tensile-fracture and shear-fracture ribbons both of which require a shallow brittle-ductile transition within Venus' crust ($<$1-2 km, Hansen and Willis 1998). Moreover, the apparent regular spacing and similarity of such ribbons may reflect a thermal control on the thickness of the deformed layer (Ruiz 2007), with which the local heat flux at the time of ribbon formation can be estimated. For a brittle–ductile transition at 1–3 km depth (Ghent and Tibuleac 2002) – in support of prevailing locally hot lithospheric conditions – Ruiz (2007) proposed a heat flux range of 130–780 mW/m^2 assuming present surface temperature, well above all estimates for the present (Sect. 3.1.4). If the surface were 100–150 K hotter when the ribbons formed, heat flux could be much reduced to 20–130 mW/m^2 with a brittle–ductile transition at $\sim$3 km depth. High heat fluxes likely place the solidus inside Venus' crust and support arguments for a weak lower crust (e.g., Ghail 2015) and for intrusive crustal magmatic bodies relevant for the plutonic-squishy lid regime. Additional constraints for Venus' past heat flux come from impact crater morphology, which suggest low heat flux ($\leq$28 mW/m^2) during the formation of Mead (Bjonnes et al. 2021), in line with predictions from mantle convection modelling in the stagnant-lid regime. In contrast, Karimi and Dombard (2017) inferred a substantially higher background heat flow in the vicinity of Mead (55–90 mW/m^2), which simply reflects the uncertainty of heat flux estimates for Venus.

The global relevance of heat flux estimates from a single morphological structure is undetermined, but by analogy to Earth, likely not all that representative. Apart from Mead, several other multi-ring basins are preserved on Venus and are not spatially clustered. If the conditions required to form a multi-ring morphology at these sites are similar as for Mead, the estimated low heat flow may have global relevance (Bjonnes et al. 2021). Recent maps for Earth (e.g., Davies 2013) indicate low heat flux comparable to those derived for Mead crater are common in and near cratonic shields, whereas heat flux is much higher at mid-ocean ridges. Cratons on Earth are the locations of thickest lithosphere; if the tesserae on Venus are similar in nature to cratons, then estimates from these locations are not globally representative.

Independent of the presence of continent-like terrains on Venus, the analogy to heat flux variations on Earth could be challenged as our world is under a regime of plate tectonics whereas Venus is not. Still, heat flux variations are also expected under a stagnant lid regime reflecting variations in crustal thickness and upper mantle temperature linked to the planform of mantle down- and upwellings. Global models of Venus' mantle convection (Rolf et al. 2018b; Uppalapati et al. 2020), however, indicate relatively small variations in surface heat flux, at least if the mantle evolved under stagnant-lid conditions for a sufficiently long time (Fig. 9). In addition, the change over the last billion years is small due to inefficient mantle cooling. Stronger partitioning of radiogenic elements in the crust facilitates surface heat loss, but not greatly, and spatial variations are not more pronounced. Under the assumed model conditions, heat flux lows correlate with regions of thick basaltic crust across which conductive heat transfer is poorer. At the same time, these regions correspond to the zones of strongest magmatic/volcanic activity and thus strong heat transfer by magmatic processes. If magmatic activity is capable of mobilising the surface lid locally, spatial variations in heat flux can be much larger (Noack et al. 2012). On a global scale, this effect is seen during episodic overturn events. During such an event (Fig. 9c–g), heat flux contrasts strongly between recently recycled regions and regions where relatively thick crust is preserved. After cessation of the resurfacing event, the strong variations typically decay on a time scale of 100–200 Myr. At this stage, the variations of predicted heat fluxes across the surface could be within a factor of about two (Fig. 9e–f). If such modelled variations are representative of Venus, this could imply that local estimates of Venus' surface heat flux–such as those made for Mead–may be within a factor of two or less compared to Venus' average, unless the estimate is made for a period during (or shortly after the cessation of) large-scale tectonic lid mobilisation. During those times, the average heat flux would not be related to the present value. High local heat flux estimates such as those derived from ribbon formation (Ruiz 2007) would thus–if confirmed–manifest a period of previous lid mobility on Venus.

Making the assumption that a past mobile epoch did occur on Venus, the question is when and how abruptly the transition to the present state occurred. Even for the Earth – which transitioned from some early regime into the mobile lid regime – neither the timing nor the spatiotemporal evolution of such a transition (gradual or abrupt, regional or global onset) are well established. For Venus conditions, global mantle flow models support long time scales for regime transitions (Weller and Kiefer 2020; Fig. 10). Whereas this time scale is likely sensitive to both mantle structure and convective vigour, tectonic stability is controlled by the system's sensitivity to perturbations (Weller and Lenardic 2018) and by the growth of thermal boundary layers. For a conditionally stable system (i.e., those in the hysteresis windows in Fig. 8), a relatively small perturbation can initiate a transition, such as changes in the global yield strength, water content of the crust, or surface temperature changes of 5–10% (Weller and Kiefer 2020). Once initiated, the perturbation disrupts the established mobile lid pattern on a time scale of $\sim$500 Myr (dashed line in Fig. 10). As instabilities grow, the system enters the transitory (or episodic-like) state, oscillating between extreme activity and quiescence. Each overturn is marked by plume generation leading to destabilisation of the lithosphere, followed by cessation of yielding and thermal boundary

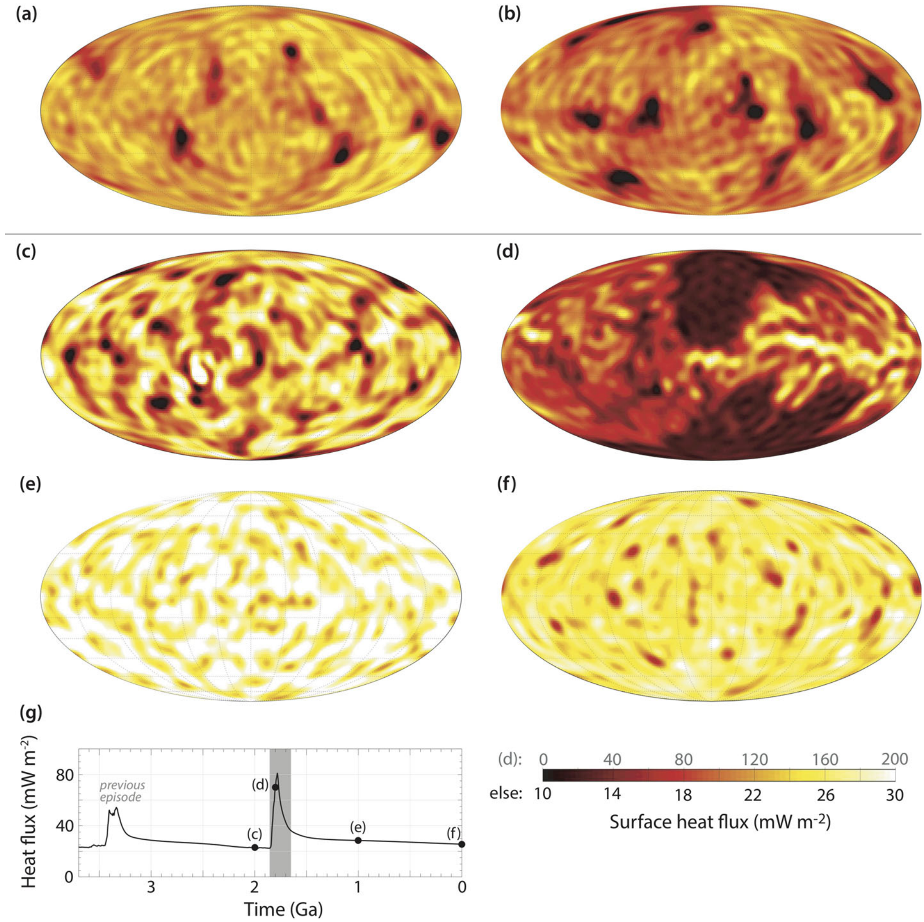

Fig. 9 Maps of surface heat flux (in the spherical harmonic degree range 0-32) for two models from Rolf et al. (2018b). (**a**) Case 'S2' in the stagnant-lid regime at 0 Ga (mean: 22 ± 3 mW/m^2, range: 6–29 mW/m^2) and (**b**) at 1 Ga (20 ± 4 mW/m^2, 4–26 mW/m^2). (**c–f**) Case 'E50' in the episodic regime with a global overturn event ending at ∼1.65 Ga at (**c**) 2 Ga (23 ± 6 mW/m^2, 7–58 mW/m^2), (**d**) 1.8 Ga (71 ± 43 mW/m^2, ∼0–226 mW/m^2), (**e**) 1 Ga (29 ± 4 mW/m^2, 20–39 mW/m^2), (**f**) 0 Ga (26 ± 3 mW/m^2, 15–33 mW/m^2). Note the anomalous color bar for panel (**d**). (**g**) Evolution of mean surface heat flow for case 'E50', the main resurfacing episode is indicated by the grey box

layer thickening. Although the number and recurrence interval of overturns may be stochastic, each overturn has a minimum operating time scale of 100–300 Myr. Consequently, each different tectonic state likely can operate over 500-1000 Myr time frames, with multiple states requiring several billion years to transition fully.

There are several implications for this behaviour and these time frames of operation. At any given time, a planet that has undergone a tectonic transition would be in a form of dynamic thermal disequilibrium. Properties such as mantle temperature and heat flow would be out of synchronicity with the planet's observed tectonic expressions by perhaps as much as a billion years (Weller and Kiefer 2020). For example, portions of Venus' mantle would be warming at differing rates after a transition (as opposed to assumptions of secular cooling trends), which presents substantial challenges for mission data interpretation. However, the

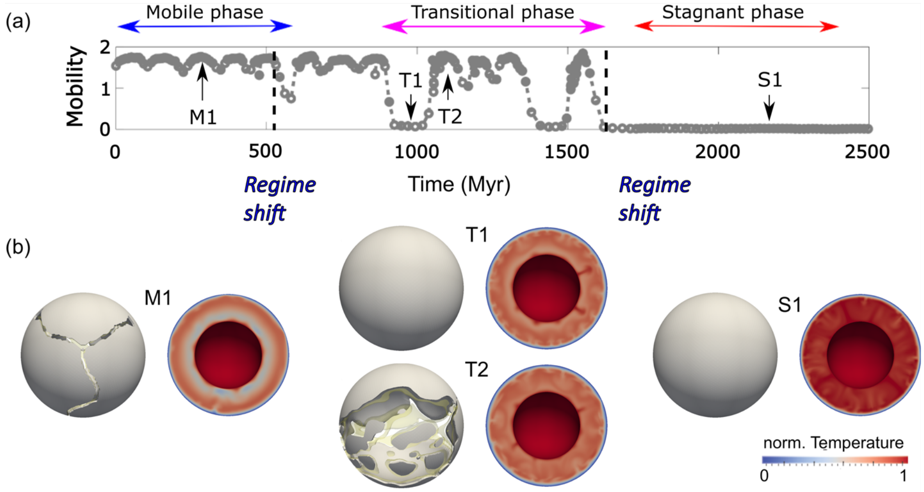

Fig. 10 Oscillation in tectonic states in 3D thermo-tectonic numerical models. The transition follows from a destabilised mobile lid (yield stress increased by 8% at time 0), through an episodic (transitional) state, into a stagnant lid state. (**a**) Time series of surface mobility; time is dimensionalised assuming a mantle overturn time of 100 Myr. Dashed lines indicate regime shifts. (**b**) Snapshots of surface viscosity (left) and temperature (right) that are representative for each regime: mobile (M1), transitional (T1, T2), and stagnant (S1). For viscosity, grey shells indicate high viscosity "plates", yellow bands are regions of active yielding. Temperature is normalised to the temperature drop across the mantle. This figure is modified from Weller and Kiefer (2020)

overall behaviour described may elucidate several outstanding questions regarding Venus' evolution. During a transition, rapid shifts in surface behaviour are predicted that would be discontinuous in nature, often restricted to regional or hemispheric extents. These local scale events may then not be reflective of the global state, in terms of activity, temperature, and heat flux. As a result, the surface of Venus may record differing convection modes and styles of tectonics, with some portions showing extreme activity, yet others reflecting tectonic quiescence. This evolutionary pattern appears consistent with many Venusian geological enigmas, such as inferences of the apparent crustal mobility required to form Ishtar Terra in addition to some tesserae (Gilmore and Head 2018), the hemisphere-scale variation in both volcanic rates and intensity between the BAT region and surrounding areas (Crumpler et al. 1993), the presence of multiple differing and apparently simultaneous styles of mantle upwelling, such as mantle plumes and coronae (Johnson and Richards 2003; Robin et al. 2007; Smrekar et al. 2010), as well as the general and more widespread partial infilling on impact crater floors (Herrick and Rumpf 2011).

Although suggestive, this concept leaves several major questions unanswered. As mentioned above, during a regime transition, a stagnant lid would be effectively indistinguishable from a sufficiently long quiescent period following an overturn. Given the range of proposed surface ages, this disequilibrium potential and the ambiguity regarding stagnant lid convection versus quiescent overturn period could imply that Venus is in an ongoing transition phase today. How long this transition has been ongoing needs to be further resolved. Existing insights are mostly based on numerical simulations, but observable diagnostics need to be defined for future missions (like those in Sect. 5.2) in order to distinguish between the different possibilities predicted by such models.

5 Synthesis and Future Perspectives

5.1 Mantle Dynamics and Evolution on Venus

As discussed throughout this chapter, our understanding of Venus' mantle evolution remains fragmentary to date due to insufficient data for developing robust models of the subsurface. Given the clearly different surface expressions, Venus' mantle likely differs from that of Earth in various aspects. A key difference could be the absence of a weak asthenosphere on Venus as suggested by gravity and topography data (Sects. 3.1.1, 3.2). The asthenosphere is often considered as key for allowing large-scale horizontal surface motions, and thus plate tectonics, on Earth. A reduced water content in Venus' interior is a possible explanation for this discrepancy, but drier conditions need to be confirmed by future observations. Alternatively, different concentrations of carbon dioxide and different degrees of partial melting may also suppress an asthenosphere (e.g., Sifré et al. 2014).

Divergent surface tectonics could also arise from a different crustal structure—like a distinct weak lower crust (e.g., Ghail 2015), possibly analogous to Earth's crust in the Archean (e.g., Ghail et al. 2023). Crustal structure is shaped by the interplay of tectonics and magmatism. Further understanding is needed of this coupling and how mantle and crustal interactions shape the surface. Such insights include establishing the coupling between the interior and the thermo-compositional state of the atmosphere, which is tied to the interior via mantle outgassing and feeds back with surface tectonics by modulating the surface temperature (Gillmann et al. 2022). The coupling between Venus' subsystems (Fig. 1) implies that—as for the Earth—the planet can only be understood as one integrative system: addressing the feedback between the subsystems is and will be essential.

Models are powerful tools with which to shed light on the coupling processes, as we have reviewed here. A major challenge, however, is to evaluate the predictive power of those models for Venus, in particular regarding time scales. Existing studies propose a wide range of possibilities of Venus' present and past interior dynamics (Fig. 11). These options are only converging in the sense that Venus' lithosphere does not feature large-scale coherent horizontal motion in contrast to Earth. Some works suggest that Venus is in the stagnant lid regime, but upon closer inspection this regime is likely different from that of the classical stagnant-lid bodies, such as the Moon or Mercury. The current tectonic state may be transient, with Venus in transition between a past mobile state and a future stagnant state (Weller and Kiefer 2020), or in a quiescent state between episodic mantle overturns (e.g., Rolf et al. 2018b). Moreover, the tectonic state may be linked to the evolution of volcanism, such as in the plutonic-squishy lid regime, where intrusive magmatism may keep the lithosphere hot and more prone to tectonic deformation (Lourenço et al. 2020). The widespread presence of coronae and other volcanic features on Venus (e.g., Stofan and Smrekar 2005), and the possibility of discrete, mobile crustal blocks (Byrne et al. 2021) could be indicative for such a volcano-tectonic regime. From our general understanding of terrestrial planet evolution, the ratio of intrusive to extrusive magmatism increases through time (e.g., Stern et al. 2018), but it is difficult to establish where Venus is currently situated on that trend. However, several inferences of recent volcanic activity (e.g., Smrekar et al. 2010; Brossier et al. 2020) suggest that Venusian magmatism may not (yet) be entirely intrusive (see Byrne 2019).

Observed surface tectonics suggest that Venus' surface has been more mobile in previous epochs, but little is known about the timing, extent, and duration of such mobility. Present geophysical constraints are insufficient to properly distinguish the different models and Venus' sparse and spatially random crater population seems insufficient to reveal these details either, especially on absolute time scales (Herrick et al. 2023; Sect. 3.1.2). On Earth,

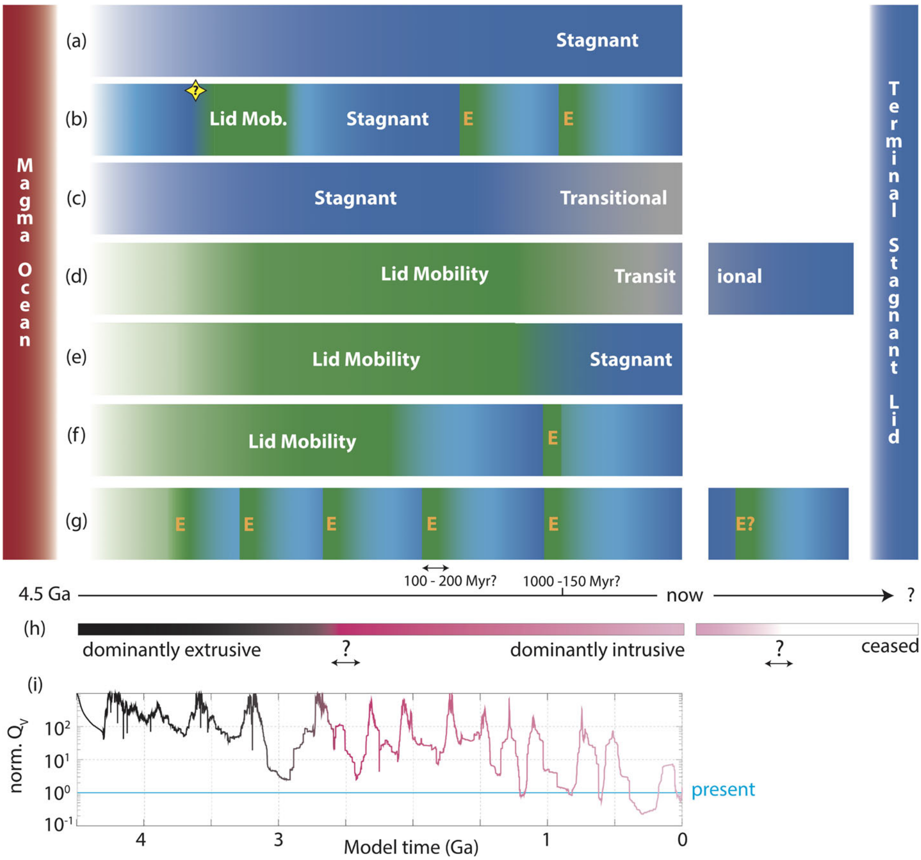

Fig. 11 Proposed evolutionary scenarios for Venus. (**a**) A continuous stagnant lid (O'Rourke and Korenaga 2015); (**b**) regime transitions controlled by surface temperature (Gillmann and Tackley 2014), possibly triggered by an impact (yellow star, Gillmann et al. 2016); (**c**) the transition from early stagnant lid to a transitional regime with localised resurfacing, triggered by surface temperature (Noack et al. 2012); (**d**) the transition from an earlier epoch of lid mobility through a transitional state towards a (future) stagnant lid (Weller and Kiefer 2020); (**e**)+(**f**) an early epoch of mobility motivated by low surface temperature to allow for liquid surface water (Way and Del Genio 2020), transitioning into a stagnant or an episodic lid state; and (**g**) episodic resurfacing with a progressively increasing interval between the episodes (E, Armann and Tackley 2012); future overturns may occur. The time axis is largely unconstrained, except for a few estimates on overturn duration or time since last overturn from modelling studies. All scenarios may start from a magma ocean and terminate in a stagnant lid (Stern et al. 2018). (**h**) Qualitative evolution of magmatism, from dominantly extrusive to more intrusive to intermittent and ultimately faded activity (see Byrne 2019). Panel (**i**) provides a quantitative example of such a scenario; shown is a (smoothed) time series of volcanic heat flux Q_V in a plutonic-squishy lid evolution (normalised to the present-day value) based on Lourenço et al. (2020). This case assumes a mantle viscosity of 10^{21} Pa s, a lithospheric yield stress of 80 MPa and a magmatic eruption efficiency of 10%

dating of tectonic events beyond the limit of seafloor reconstructions relies on geochronology and radiometric dating of preserved rocks, but such an option is infeasible for Venus for the time being. However, potential future boosts for improving our understanding are discussed in Sect. 5.2. Apart from the timing and duration of previous lid mobility, the question arises as to whether this period was similar to mobility on Earth (that is, plate tectonics). On present Earth, mantle plume and plate boundary locations are largely uncorrelated, but laboratory experiments and structural observations on Venus point towards plume-induced subduction, favoured by Venus' hotter and thus weaker lithosphere (Davaille et al. 2017). A similar mechanism may have triggered (proto-)subduction on early Earth (Gerya et al. 2015; Fischer and Gerya 2016; Baes et al. 2020).

If the analogy with early Earth holds, present Venus would be a key observatory for the Earth's past, but why would Earth have evolved from its early tectonic mode into modern plate tectonics, when Venus is still in the early-state mode? Could Venus' evolution have been 'slower', meaning that it has not yet entered the plate tectonics mode? Cooling rate is largely controlled by mantle viscosity, thus a higher viscosity in Venus' mantle could have trapped heat inside more efficiently. Reduced water content could explain such increased viscosity, but would imply that Venus either accreted differently than Earth or was dried out soon after—perhaps by an extreme impact collision (Davies 2008). Also, such a 'slower' evolution would suggest a mobile epoch yet to come, but hints of such a mobile regime are already manifested in the planet's preserved geological record (Sect. 4.3). This could controversially point to a 'faster' evolutionary pace on Venus instead. The role of a hotter lithosphere in this regard is discussed in Ghail et al. (2023). Other possibilities are recurring episodes of mobility or that plate tectonics may develop only under specific circumstances (for water content and surface temperature for example) that the Earth happened to have at some point, but Venus did not (see Stern et al. 2018).

A planet is a highly dynamic system with strongly nonlinear behaviour: small changes can induce large and unpredictable consequences. Coupled atmosphere–interior models (e.g., Gillmann and Tackley 2014) indicate substantial variation of surface temperature through time, more than enough to induce transition from one mode of tectonics to another; bolide impacts could—stochastically—further trigger such transitions (Sect. 4.2). Even if the immediate trigger subsequently vanishes, the transition in tectonic mode may not automatically reverse because of hysteresis. Once pushed off the evolutionary path into a different mode, various processes may establish some of Venus' peculiarities such as lacking an asthenosphere and having high surface temperature, which are both maintained by and help maintain the planet's tectonic regime. In this light, self-organisation could explain the divergent evolutions of two very similar planets. Whether such an explanation applies for the tectonic divergence of Venus and Earth remains difficult to answer. Some avenues to further resolve these issues are given below.

5.2 Future Boosts for Understanding the Venus Mantle?

5.2.1 Expected Insights from Confirmed Future Missions

The upcoming decade(s) will see various space missions targeting Venus. NASA's VERITAS and DAVINCI missions, together with ESA's EnVision mission, will look for active volcanism and tectonics, return high-resolution gravity and topography data, provide compositional maps of the surface, and place bounds on estimates of mantle outgassing rates by measuring noble gas concentrations (see Widemann et al. 2023). Some key questions with particular relevance for the mantle that could potentially be resolved with the upcoming mission data are discussed here.

5.2.1.1 How Did Mantle Cooling History Contol the State of Venus' Core? The mantle provides the boundary conditions of the core and controls its cooling history. A better understanding of Venus' core would thus inform our understanding of the mantle evolution. VERITAS will provide much improved measurements of the planet's k_2 tidal Love number, a characteristic for the planet's tidal deformation from which the size of Venus' core and its state can be estimated, in combination with the moment of inertia factor (MoI, Cascioli et al. 2021). The radius of the core pins the thickness of the mantle and thus places bounds on the temperature and pressure range at and above the core–mantle boundary. Margot et al. (2021) found the first estimate of core size (∼3500 km) using Earth-based radar observations of Venus' spin. However, this method has large error bars (±300 km at best), and cannot determine core state. Scaling core size from that of Earth places Venus' lowermost mantle probably just outside the stability field of post-perovskite, which is known to influence the dynamics of the Earth's lowermost mantle (e.g., Čížková et al. 2010) and outer core (e.g., Amit and Choblet 2009). Assuming an Earth-like mantle density profile and a transition pressure of ∼125 GPa (Murakami et al. 2004; Trønnes et al. 2019), the occurrence of post-perovskite is inconsistent with the range of core sizes estimated by Margot et al. (2021).

The improved data will not only better characterise the size of the core, but also its state, whether it is liquid, solid, or partially both. This will be an important constraint for mantle interior models (Dumoulin et al. 2017), and for telling us how much core cooling should be accounted for in these models. In turn, we will be able to place bounds on core–mantle boundary temperature and the formation of mantle plumes as well as their excess temperature, at least in the deep mantle. For example, if future k_2 and MoI measurements point to a fully solidified core, a mantle evolution in the permanent stagnant lid regime becomes less feasible as not enough heat could have escaped the interior. In contrast, a fully solid core would imply strong core cooling and point to a relatively cold mantle, which was cooled more efficiently like in a mobile lid regime. Such information will also impact the feasibility of a basal magma ocean inside Venus today, and thus our understanding of the planet's magnetic field history (O'Rourke et al. 2019). Mantle cooling rate further strongly depends on the mantle viscosity. To constrain the latter, VERITAS will deliver estimates of Venus' tidal phase lag with a potential accuracy of 0.05° (Cascioli et al. 2021).

5.2.1.2 How Much Heat Does Venus Lose Today? The heat loss from the mantle through the surface of the planet is a cornerstone for every thermal evolution model. VERITAS will measure global topography and gravity at much improved resolution over presently available data. EnVision will also measure topography and gravity (e.g., Rosenblatt et al. 2021). Such refined data is expected to inform models on upper-mantle density anomalies, crustal thickness, and eventually elastic thickness from which thermal gradients and thus surface heat flux can be estimated. The improved resolution could also provide a better picture of lateral heat flux variations, which can be linked to hotter or colder regions of the mantle and possibly to convective structures such as deep mantle plumes. VERITAS will additionally deliver improved measurements of surface thermal emissivity, positive anomalies of which may be linked to plume locations (e.g., Smrekar et al. 2010). Moreover, thermal emissivity is an important observable factor for the detection of recent volcanic activity (e.g., D'Incecco et al. 2017, 2021). Improved estimates of the volcanic eruption rate at Venus could pin the amount of heat leaving the interior via volcanic processes, thus providing constraints on the temperature of the uppermost mantle or the important partitioning of intrusive magmatism and eruptive volcanism.

5.2.1.3 Are Venus' Tesserae Felsic and Older than the Average Crust? High-resolution gravity and topography data will not only allow for improved estimates of elastic thickness and thermal gradients, but also help to further map out the crater population, in particular in terms of searching for buried craters. Subsurface radar measurements from EnVision may support this search. Buried craters may also be apparent in high resolution topography, as on other rocky bodies, or could still be retained in the gravity field if measured at sufficient resolution. Given the small population of preserved craters on Venus, the detection of buried craters, particularly a few large ones (e.g., Karimi et al. 2018), would place new constraints on the absolute model age of Venus' crust and therefore on the rates of resurfacing the internal evolution models must accommodate. This, again, is directly linked to Venus' tectonic regime. In particular, refining the distribution of Venus' modified craters could help to identify whether parts of Venus' surface, specifically tesserae, are older than the average surface age. Deformed craters on tesserae could be seen in high resolution topography, as evidenced in the stereo topography (see Herrick et al. 2023). Higher-resolution mapping of the tesserae's bounding structures can additionally help to determine the structural relationship between the tesserae and their surroundings and thus to constrain the formation mechanism of the tesserae. The advanced mapping of tesserae regions, the search for buried craters and in particular the determination of the processes modifying craters will reveal Venus' resurfacing history in detail.

VERITAS and EnVision will also measure iron content from orbit, providing proxy data for the SiO_2 content of the surface rocks (Dyar et al. 2020; Helbert et al. 2021). Such data for the whole planet will be acquired, so that the composition of tessera regions will be further refined. Alpha Regio—the only tessera region captured by previous data—has a composition that is consistent with, but does not uniquely mean, felsic rock (Gilmore et al. 2015, 2017). Determining whether this finding holds for all Venusian tesserae is important for our understanding of the interior evolution, as a generally felsic composition would require the presence of near-surface water during tessera formation. The DAVINCI mission (Garvin et al. 2022) shall deliver new measurements of Venus' D/H ratio with a precision sufficient to distinguish between the different scenarios proposed for the origin of Venus' water. This will help to constrain how wet the near-surface of Venus may be. If such a wet environment propagates into the deeper interior, it has implications for the viscosity and the degree of partial melting in the upper mantle and therefore volcanic outgassing rates, for which DAVINCI's improved measurement of atmospheric noble gas concentrations will be new constraints. Water in Venus' shallow interior will also affect the strength of the crust. Targeted surface deformation maps—another expected outcome of the VERITAS mission—will further support this line of insight by revealing currently active features for various parts of the planet. Finally, DAVINCI's descending probe (VDI) will take high-resolution images of Alpha Regio when approaching the surface. Such images may offer additional visual information for the formation and evolution of what are possibly Venus' most ancient tectonic features.

5.2.2 Additional Desired Observations Beyond Planned Mission Plans

5.2.2.1 In-Situ Heat Flow The future data to be delivered by DAVINCI, VERITAS, and EnVision will greatly advance our understanding of Venus' interior and mantle, but gaps will naturally remain. For example, surface heat flux will remain an estimate and not be measured in-situ, which would be the ultimate data to pin down Venus' heat loss and the thermal state of the upper mantle. As discussed in Sect. 4.3, however, extrapolating data from single locations to a global characteristic value is difficult when lateral variations are important, so that future heat flow measurements would ideally be taken at various sites.

5.2.2.2 Seismology Seismology would be the most powerful tool to map out the planet's density distribution. With that, the thermo-compositional structure of the interior could be further revealed, including the (spatially varying) thickness of Venus' crust. Such constraints would ultimately inform thermo-magmatic evolution models about the rates of crustal production and destruction and therefore about the volume and timing of volcanic resurfacing. Mapping the relatively shallow crustal boundary could be feasible with relatively low-magnitude seismic events; deeper structures would be more challenging to map, but useful for understanding Venus' deeper mantle structure. Major unknown aspects are for instance the thickness of the mantle transition zone and its spatial variations, which provide information about the thermal state of that zone (e.g., Lawrence and Shearer 2008), the identification of possible heterogeneity in a deep layer comparable to the D" region on Earth (e.g., Cobden et al. 2015), and confirmation of the size and state of the core that are indirectly determined by measurements of k_2 and the MoI. The potential of future seismic investigation is detailed in Widemann et al. (2023); further background and concepts for (future) seismology on Venus are for instance given by Stevenson et al. (2015) and Kremic et al. (2021).

5.2.2.3 Electromagnetics As discussed above, the water content in the crust and upper mantle is a crucial parameter to understand rheology and deformation on Venus.

Electromagnetic methods, for instance magnetotellurics, provide a tool to constrain water, but also melt and carbon dioxide content (e.g., Sifré et al. 2014). Such measurements are best made from the surface, but aerial sounding is a possibility and can still achieve exploration depths exceeding the lithospheric thickness with sufficiently low electromagnetic frequencies (Grimm et al. 2012). If Venus' crust is dry, this method could provide an independent measurement of crustal and lithospheric thickness. If the crust is wet, containing 100s ppm of water, exploration depth is limited to much shallower depth and lithospheric properties can no longer be estimated; however, such a finding would still reveal the wet nature of Venus' upper mantle. Knowing the water content of the upper mantle would further inform our models of its viscosity, jointly with improved gravity and topography measurements by EnVision and VERITAS, and shed light upon whether Venus' interior features an asthenosphere, and how the mantle couples to the lithosphere and crust. An issue for such a measurement is the availability of electromagnetic sources. Lightning in Venus' atmosphere (e.g., Russell et al. 2007)—if found to be present—could provide one such source at frequencies that allow for wave penetration as deep as 100 km into Venus' interior (Grimm et al. 2012).

5.2.3 Future Conceptual Approaches and Modelling

The next generation of missions will deliver new observations with which to improve our understanding of Venus' interior, but a remaining challenge will be to incorporate these observations into a common dynamic framework and ideally to place at least relative time scales on the processes to which these observations can be attributed. To do so, future advances in modelling the evolution of the planet are needed. Of particular importance will be work dedicated to revealing the coupling between subsystems—such as the core and the mantle, the mantle and the atmosphere (including surface processes), and tectonics and volcanism. Some important questions to address with such models are how much water does Venus' interior contain, how is it distributed through time and space, and how does it affect mantle viscosity, melt generation, and the properties of the crust? Improved understanding of how magma migrates through the crust and how much of it makes it to the surface as a function of crustal properties is another target of future modelling efforts. However, such models must also account for the properties and complexities of Venus' crustal rocks, their composition, and mineral assemblage—which, as we know from Earth—can be far from homogeneous even over relatively modest spatial scales. Identifying and analysing Earth materials analogue to Venusian rocks (e.g., Filiberto et al. 2020), as well as detailed understanding of mineral physics of mantle material under high temperature and pressure conditions, are thus also necessary to decipher Venus' still enigmatic mantle structure and evolution.

Acknowledgements The authors thank the International Space Science Institute in Bern/Switzerland for hosting and the editorial board (D. Breuer, C. Gillmann, S. Smrekar, T. Spohn, T. Widemann, C. Wilson) for organising the workshop "Venus Evolution Through Time" from which this work originates. The authors thank W. Kiefer and W. Moore for thoughtful reviews that helped to improve the original manuscript. TR acknowledges funding from the Research Council of Norway (grants 223272, Centre for Earth Evolution and Dynamics, and 276032, PLATONICS).

Funding Note Open access funding provided by University of Oslo (incl Oslo University Hospital).

Declarations

Competing Interests The authors declare no competing interests.

References

Abe Y (1995) Basic equations for evolution of partially molten mantle and core. In: TERRA-PUB, pp 215–230

Amit H, Choblet G (2009) Mantle-driven geodynamo features—effects of post-perovskite phase transition. Earth Planets Space 61:1255–1268. https://doi.org/10.1186/bf03352978

Anderson FS, Smrekar SE (2006) Global mapping of crustal and lithospheric thickness on Venus. J Geophys Res 111:E08006. https://doi.org/10.1029/2004je002395

Arkani-Hamed J (1993) On the tectonics of Venus. Phys Earth Planet Inter 76:75–96. https://doi.org/10.1016/0031-9201(93)90056-f

Armann M, Tackley PJ (2012) Simulating the thermochemical magmatic and tectonic evolution of Venus' mantle and lithosphere: two-dimensional models. J Geophys Res 117:E12003. https://doi.org/10.1029/2012je004231

Azuma S, Katayama I, Nakakuki T (2014) Rheological decoupling at the Moho and implication to Venusian tectonics. Sci Rep 4:4403. https://doi.org/10.1038/srep04403

Baes M, Sobolev S, Gerya T, Brune S (2020) Plume-induced subduction initiation: single- or multi-slab subduction? Geochem Geophys Geosyst 21(2):e2019GC008663. https://doi.org/10.1029/2019GC008663

Benešová N, Čížková H (2012) Geoid and topography of Venus in various thermal convection models. Stud Geophys Geod 56:621–639. https://doi.org/10.1007/s11200-011-0251-7

Bercovici D, Karato S-I (2003) Whole-mantle convection and the transition-zone water filter. Nature 425:39–44. https://doi.org/10.1038/nature01918

Bercovici D, Ricard J. (2014) Plate tectonics, damage and inheritance. Nature 508:513–516. https://doi.org/10.1038/nature13072

Bindschadler DL, Schubert G, Ford P (1994) Venus' center of figure-center of mass offset. Icarus 111:417–432. https://doi.org/10.1006/icar.1994.1153

Bjonnes EE, Hansen VL, James B, Swemson JB (2012) Equilibrium resurfacing of Venus: results from new Monte Carlo modeling and implications for Venus surface histories. Icarus 217:451–461. https://doi.org/10.1016/j.icarus.2011.03.033

Bjonnes EE, Johnson BC, Evans AJ (2021) Estimating Venusian thermal conditions using multiring basin morphology. Nat Astron 5:498–502. https://doi.org/10.1038/s41550-020-01289-6

Blaske CH, O'Rourke JG (2021) Energetic requirements for dynamos in the metallic cores of super-Earth and super-Venus exoplanets. J Geophys Res 126:e2020JE006739. https://doi.org/10.1029/2020je006739
Boehler R (1998) Fe–FeS eutectic temperatures to 620 kbar. Phys Earth Planet Inter 96:181–186. https://doi.org/10.1016/0031-9201(96)03150-0
Bondarenko NV, Head JW, Ivanov MA (2010) Present-day volcanism on Venus: evidence from microwave radiometry. Geophys Res Lett 37:L23202. https://doi.org/10.1029/2010gl045233
Borrelli ME, O'Rourke JG, Smrekar SE, Ostberg CM (2021) A global survey of lithospheric flexure at steep-sided domical volcanoes on Venus reveals intermediate elastic thicknesses. J Geophys Res 126:e2020JE006756. https://doi.org/10.1029/2020je006756
Boyet M, Bouvier A, Frossard P, Hammouda T, Garcon M, Gannoun A (2018) Enstatite chondrites EL3 as building blocks for the Earth: the debate over the 146Sm–142Nd systematics. Earth Planet Sci Lett 488:68–78. https://doi.org/10.1016/j.epsl.2018.02.004
Brossier JF, Gilmore MS, Toner K (2020) Low radar emissivity signatures on Venus volcanoes and coronae: new insights on relative composition and age. Icarus 343:113693. https://doi.org/10.1016/j.icarus.2020.113693
Buck WR (1992) Global decoupling of crust and mantle: implications for topography, geoid and mantle viscosity on Venus. Geophys Res Lett 19:2111–2114. https://doi.org/10.1029/92gl02462
Bunge H-P, Richards MA, Baumgardner JR (1997) A sensitivity study of three-dimensional spherical mantle convection at 10^8 Rayleigh number: effects of depth-dependent viscosity, heating mode, and an endothermic phase change. J Geophys Res 102:11991–12007. https://doi.org/10.1029/96jb03806
Burkhardt C, Borg LE, Brennecka GA, Shollenberger QR, Dauphas N, Kleine T (2016) A nucleosynthetic origin for the Earth's anomalous 142Nd composition. Nature 537:394–398. https://doi.org/10.1038/nature18956
Byrne PK (2019) A comparison of inner solar system volcanism. Nat Astron 4:321–327. https://doi.org/10.1038/s41550-019-0944-3
Byrne PK, Ghail RC, Celâl Sengör AM, James PB, Klimczak C, Solomon SC (2021) A globally fragmented and mobile lithosphere on Venus. Proc Natl Acad Sci 118:e2025919118. https://doi.org/10.1073/pnas.2025919118
Campbell BA, Morgan GA, Whitten JL, Carter LM, Glaze LS, Campbell DB (2017) Pyroclastic flow deposits on Venus as indicators of renewed magmatic activity. J Geophys Res 122:1580–1596. https://doi.org/10.1002/2017je005299
Capitanio FA, Nebel O, Cawood PA, Weinberg RF, Chowdhury P (2019) Reconciling thermal regimes and tectonics of the early Earth. Geology 47:923–927. https://doi.org/10.1130/G46239.1
Carter L, Gilmore M, Ghail R, Byrne P, Izenberg N, Smrekar S (2023) Sedimentary processes on Venus. Space Sci Rev
Cascioli G, Hensley S, De Marchi F, Breuer D, Durante D, Racioppa P, Iess L, Mazarico E, Smrekar SE (2021) The determination of the rotational state and interior structure of Venus with VERITAS. Planet Sci J 2:220. https://doi.org/10.3847/PSJ/ac26c0
Cawood PA, Hawkesworth CJ, Dhuime B (2012) The continental record and the generation of continental crust. Geol Soc Am Bull 125:14–32. https://doi.org/10.1130/b30722.1
Christensen U (2006) A deep dynamo generating Mercury's magnetic field. Nature 444:1056–1058. https://doi.org/10.1038/nature05342
Čížková H, Cadek O, Matiska C, Yuen DA (2010) Implications of post-perovskite transport properties for core–mantle dynamics. Phys Earth Planet Inter 180:235–243. https://doi.org/10.1016/j.pepi.2009.08.008
Cobden L, Thomas C, Trampert J (2015) Seismic detection of post-perovskite inside the Earth. In: Khan A, Deschamps F (eds) The Earth's heterogeneous mantle. Springer, Cham, pp 391–440. https://doi.org/10.1007/978-3-319-15627-9_13
Coltice N, Seton M, Rolf T, Müller RD, Tackley PJ (2013) Convergence of tectonic reconstructions and mantle convection models for significant fluctuations in seafloor spreading. Earth Planet Sci Lett 383:92–100. https://doi.org/10.1016/j.epsl.2013.09.032
Crameri F, Tackley PJ (2016) Subduction initiation from a stagnant lid and global overturn: new insights from numerical models with a free surface. Prog Earth Planet Sci 3:30. https://doi.org/10.1186/s40645-016-0103-8
Crisp JA (1984) Rates of magma emplacement and volcanic output. J Volcanol Geotherm Res 20:177–211. https://doi.org/10.1016/0377-0273(84)90039-8
Crowley JW, O'Connell RJ (2012) An analytic model of convection in a system with layered viscosity and plates. Geophys J Int 188:61–78. https://doi.org/10.1111/j.1365-246x.2011.05254.x
Crumpler LS, Head JW, Aubele JC (1993) Relation of major volcanic center concentration on Venus to global tectonic patterns. Science 261:591–595. https://doi.org/10.1126/science.261.5121.591

Dannberg J, Eilon Z, Faul U, Gassmöller R, Moulik P, Myhill R (2017) The importance of grain size to mantle dynamics and seismological observations. Geochem Geophys Geosyst 18:3034–3061. https://doi.org/10.1002/2017gc006944

Davaille A, Smrekar SE, Tomlinson S (2017) Experimental and observational evidence for plume-induced subduction on Venus. Nat Geosci 10:349–355. https://doi.org/10.1038/ngeo2928

Davies JH (2008) Did a mega-collision dry Venus' interior? Earth Planet Sci Lett 268:376–383. https://doi.org/10.1016/j.epsl.2008.01.031

Davies JH (2013) Global map of solid Earth surface heat flow. Geochem Geophys Geosyst 14:4608–4622. https://doi.org/10.1002/ggge.20271

DePaolo DJ, Manga M (2003) GEOLOGY: deep origin of hotspots–the mantle plume model. Science 300:920–992. https://doi.org/10.1126/science.1083623

D'Incecco P, Müller N, Helbert J, D'Amore M (2017) Idunn Mons on Venus: location and extent of recently active lava flows. Planet Space Sci 136:25–33. https://doi.org/10.1016/j.pss.2016.12.002

D'Incecco P, Filiberto J, López I, Gorinov DA, Komatsu G (2021) Idunn Mons: evidence for ongoing volcano-tectonic activity and atmospheric implications on Venus. Planet Sci J.2(5):215. https://doi.org/10.3847/PSJ/ac2258

Driscoll P, Bercovici D (2013) Divergent evolution of Earth and Venus: influence of degassing, tectonics, and magnetic fields. Icarus 226:1447–1464. https://doi.org/10.1016/j.icarus.2013.07.025

Driscoll P, Bercovici D (2014) On the thermal and magnetic histories of Earth and Venus: influences of melting, radioactivity, and conductivity. Phys Earth Planet Inter 236:36–51. https://doi.org/10.1016/j.pepi.2014.08.004

Dumoulin C, Tobie G, Verhoeven O, Rosenblatt P, Rambaux N (2017) Tidal constraints on the interior of Venus. J Geophys Res, Planets 122:1338–1352. https://doi.org/10.1002/2016je005249

Dyar MD, Helbert J, Maturilli A, Müller NT, Kappel D (2020) Probing Venus surface iron contents with six-band visible near-infrared spectroscopy from orbit. Geophys Res Lett 47:e2020GL090497. https://doi.org/10.1029/2020GL090497

Filiberto J, Trang D, Treiman AH, Gilmore MS (2020) Present-day volcanism on Venus as evidenced from weathering rates of olivine. Sci Adv 6:eaax7445. https://doi.org/10.1126/sciadv.aax7445

Fischer R, Gerya T (2016) Early Earth plume-lid tectonics: a high-resolution 3D numerical modelling approach. J Geodyn 100:198–214. https://doi.org/10.1016/j.jog.2016.03.004

Foley BJ (2018) The dependence of planetary tectonics on mantle thermal state: applications to early Earth evolution. Philos Trans R Soc A 376:20170409. https://doi.org/10.1098/rsta.2017.0409

Foley BJ, Becker TW (2009) Generation of plate-like behavior and mantle heterogeneity from a spherical, viscoplastic convection model. Geochem Geophys Geosyst 10:Q08001. https://doi.org/10.1029/2009gc002378

Foley BJ, Bercovici D, Landuyt W (2012) The conditions for plate tectonics on super-earths: inferences from convection models with damage. Earth Planet Sci Lett 331–332:281–290. https://doi.org/10.1016/j.epsl.2012.03.028

Fowler AC (1985) Fast thermoviscous convection. Stud Appl Math 72:189–219. https://doi.org/10.1002/sapm1985723189

Fowler AC, O'Brien SGB (1996) A mechanism for episodic subduction on Venus. J Geophys Res, Planets 101:4755–4763. https://doi.org/10.1029/95je03261

French SW, Romanowicz B (2015) Broad plumes rooted at the base of the Earth's mantle beneath major hotspots. Nature 525:95–99. https://doi.org/10.1038/nature14876

Fuchs L, Becker T (2019) Role of strain-dependent weakening memory on the style of mantle convection and plate boundary stability. Geophys J Int 218:601–618. https://doi.org/10.1093/gji/ggz167

Fuchs L, Becker TW (2021) Deformation memory in the lithosphere: a comparison of damage-dependent weakening and grain-size sensitive rheologies. J Geophys Res, Solid Earth 126:e2020JB020335. https://doi.org/10.1029/2020jb020335

Garnero EJ, McNamara AK, Shim S-H (2016) Continent-sized anomalous zones with low seismic velocity at the base of Earth's mantle. Nat Geosci 9:481–489. https://doi.org/10.1038/ngeo2733

Garvin JB, Getty SA, Arney GN, Johnson NM, Kohler E, Schwer KO, Sekerak M, Bartels A, Saylor RS, Elliott VE, Goodloe CS, Garrison MB, Cottini V, Izenberg N, Lorenz R, Malespin CA, Ravine M, Webster CR, Atkinson DH, Aslam S, Atreya S, Bos BJ, Brinckerhoff WB, Campbell B, Crisp D, Filiberto JR, Forget F, Gilmore M, Gorius N, Grinspoon D, Hofmann AE, Kane SR, Kiefer W, Lebonnois S, Mahaffy PR, Pavlov A, Trainer M, Zahnle KJ, Zolotov M (2022) Revealing the mysteries of Venus: the DAVINCI mission. Planet Sci J 3:117. https://doi.org/10.3847/PSJ/ac63c2

Gerya TV (2014) Plume-induced crustal convection: 3D thermomechanical model and implications for the origin of novae and coronae on Venus. Earth Planet Sci Lett 391:183–192. https://doi.org/10.1016/j.epsl.2014.02.005

Gerya TV, Stern RJ, Baes M, Sobolev SV, Whattam SA (2015) Plate tectonics on the Earth triggered by plume-induced subduction initiation. Nature 527:221–225. https://doi.org/10.1038/nature15752
Ghail R (2015) Rheological and petrological implications for a stagnant lid regime on Venus. Planet Space Sci 113–114:2–9. https://doi.org/10.1016/j.pss.2015.02.005
Ghail R, Smrekar SE, Byrne PK, Gilmore MS, Herrick RR, Ivanov MA, Plesa AC, Rolf T, Sabbeth L, Schools JW, Shellnut JG (2023) Volcano and tectonic constraints on the evolution of Venus. Space Sci Rev
Ghent RR, Tibuleac IM (2002) Ribbon spacing in Venusian tessera: implications for layer thickness and thermal state. Geophys Res Lett 29:61. https://doi.org/10.1029/2002GL015994
Ghias SR, Jarvis GT (2008) Mantle convection models with temperature- and depth-dependent thermal expansivity. J Geophys Res, Solid Earth 113:B08408. https://doi.org/10.1029/2007jb005355
Gillmann C, Tackley P (2014) Atmosphere/mantle coupling and feedbacks on Venus. J Geophys Res, Planets 119:1189–1217. https://doi.org/10.1002/2013je004505
Gillmann C, Golabek GJ, Tackley PJ (2016) Effect of a single large impact on the coupled atmosphere-interior evolution of Venus. Icarus 268:295–312. https://doi.org/10.1016/j.icarus.2015.12.024
Gillmann C, Golabek GJ, Raymond SN, Schönbächler M, Tackley PJ, Dehant V, Debaille V (2020) Dry late accretion inferred from Venus's coupled atmosphere and internal evolution. Nat Geosci 13:265–269. https://doi.org/10.1038/s41561-020-0561-x
Gillmann C, Way MJ, Avice G, Breuer D, Golabek GJ, Höning D, Krissansen-Totton J, Lammer H, O'Rourke JG, Persson M, Plesa AC, Salvador A, Scherf M, Zolotov MY (2022) The long-term evolution of the atmosphere of Venus: processes and feedback mechanisms. Space Sci Rev 218:56. https://doi.org/10.1007/s11214-022-00924-0
Gilmore MS, Head JW (2018) Morphology and deformational history of Tellus Regio, Venus: evidence for assembly and collision. Planet Space Sci 154:5–20. https://doi.org/10.1016/j.pss.2018.02.001
Gilmore MS, Mueller N, Helbert J (2015) VIRTIS emissivity of Alpha Regio, Venus, with implications for tessera composition. Icarus 254:350–361. https://doi.org/10.1016/j.icarus.2015.04.008
Gilmore MS, Treiman A, Helbert J, Smrekar S (2017) Venus surface composition constrained by observation and experiment. Space Sci Rev 212:1511–1540. https://doi.org/10.1007/s11214-017-0370-8
Gilmore M, Helbert J, Brossier J, Carter L, Darby D, Filiberto F, Gerya T, Ghail R, Ivanov M, Izenberg N, Müller N, Santos A, Smrekar S (2023) Surface composition and mineralogy of the Venus surface. Space Sci Rev
Green DH, Hibberson WO, Rosenthal A, Kovács I, Yaxley GM, Falloon TJ, Brink F (2014) Experimental study of the influence of water on melting and phase assemblages in the upper mantle. J Petrol 55:2067–2096. https://doi.org/10.1093/petrology/egu050
Grimm RE (1994) Recent deformation rates on Venus. J Geophys Res 99:23163–23171. https://doi.org/10.1029/94JE02196
Grimm RE, Barr Milnar A, Harrison K, Stillman D, Neal K, Vincent MA, Delory G (2012) Aerial electromagnetic sounding of the lithosphere of Venus. Icarus 217:462–473. https://doi.org/10.1016/j.icarus.2011.07.021
Grinspoon DH (1993) Implications of the high D/H ratio for the sources of water in Venus' atmosphere. Nature 363:428–431. https://doi.org/10.1038/363428a0
Guerrero JM, Lowman JP, Deschamps F, Tackley PJ (2018) The influence of curvature on convection in a temperature-dependent viscosity fluid: implications for the 2-D and 3-D modeling of moons. J Geophys Res, Planets 123:1863–1880. https://doi.org/10.1029/2017je005497
Gülcher AJP, Gerya TV, Montési L, Munch J (2020) Corona structures driven by plume–lithosphere interactions and evidence for ongoing plume activity on Venus. Nat Geosci 13:547–554. https://doi.org/10.1038/s41561-020-0606-1
Gülcher AJP, Ballmer MD, Tackley PJ (2021) Coupled dynamics and evolution of primordial and recycled heterogeneity in Earth's lower mantle. Solid Earth 12:2087–2107. https://doi.org/10.5194/se-12-2087-2021
Hager BH, Clayton RW, Richards MA, Comer RP, Dziewonski (1985) Lower mantle heterogeneity, dynamic topography and the geoid. Nature 313:541–545. https://doi.org/10.1038/313541a0
Hanmer S (2020) Tessera terrain ribbon fabrics on Venus reviewed: could they be dyke swarms? Earth-Sci Rev 201:103077. https://doi.org/10.1016/j.earscirev.2019.103077. 2020
Hansen VL, Willis JJ (1998) Ribbon terrain formation, southwestern fortuna tessera, Venus: implications for lithosphere evolution. Icarus 132:321–343. https://doi.org/10.1006/icar.1998.5897
Hansen U, Yuen DA, Kroening SE, Larsen TB (1993) Dynamical consequences of depth-dependent thermal expansivity and viscosity on mantle circulations and thermal structure. Phys Earth Planet Inter 77:205–223. https://doi.org/10.1016/0031-9201(93)90099-u
Harris LB, Bédard JH (2013) Crustal evolution and deformation in a non-plate-tectonic Achaean Earth: comparisons with Venus. In: Modern approaches in solid Earth sciences evolution of Archean crust and early life, pp 215–291. https://doi.org/10.1007/978-94-007-7615-9_9

Harris LB, Bédard JH (2014) Interactions between continent-like 'drift', rifting and mantle flow on Venus: gravity interpretations and Earth analogues. Geol Soc (Lond) Spec Publ 401:327–356. https://doi.org/10.1144/sp401.9
Helbert J, Maturilli A, Dyar MD, Alemanno G (2021) Deriving iron contents from past and future Venus surface spectra with new high-temperature laboratory emissivity data. Sci Adv 7:eaba9428. https://doi.org/10.1126/sciadv.aba9428
Herrick RR, Phillips RJ (1994) Effects of the Venusian atmosphere on incoming meteoroids and the impact crater population. Icarus 112:543–546. https://doi.org/10.1006/icar.1994.1180
Herrick RR, Rumpf ME (2011) Postimpact modification by volcanic or tectonic processes as the rule, not the exception, for Venusian craters. J Geophys Res 116:E02004. https://doi.org/10.1029/2010je003722
Herrick RR, Bjonnes E, Carter L, Gerya T, Ghail R, Gillmann C, Gilmore M, Hensley S, Ivanov M, Izenberg N, Müller N, O'Rourke JG, Rolf T, Smrekar SE, Weller M (2023) Resurfacing history and volcanic activity of Venus. Space Sci Rev
Hofmann AW (1997) Mantle geochemistry: the message from oceanic volcanism. Nature 385:219–229. https://doi.org/10.1038/385219a0
Höink T, Lenardic A (2008) Three-dimensional mantle convection simulations with a low-viscosity asthenosphere and the relationship between heat flow and the horizontal length scale of convection. Geophys Res Lett 35:L10304. https://doi.org/10.1029/2008gl033854
Höink T, Lenardic A (2010) Long wavelength convection, Poiseuille & Couette flow in the low-viscosity asthenosphere and the strength of plate margins. Geophys J Int 180:23–33. https://doi.org/10.1111/j.1365-246x.2009.04404.x
Höink T, Lenardic A, Richards M (2012) Depth-dependent viscosity and mantle stress amplification: implications for the role of the asthenosphere in maintaining plate tectonics. Geophys J Int 191:30–41. https://doi.org/10.1111/j.1365-246x.2012.05621.x
Hoogenboom T, Houseman GA (2006) Rayleigh-Taylor instability as a mechanism for corona formation on Venus. Icarus 180:292–307. https://doi.org/10.1016/j.icarus.2005.11.001
Howard LN (1966) Convection at high Rayleigh number. In: Görtler H (ed) Applied mechanics. Springer, Berlin, pp 1109–1115. https://doi.org/10.1007/978-3-662-29364-5_147
Hu Y, Bürgmann R, Banerjee P, Feng L, Hill EM, Ito T, Tabei T, Wang K (2016) Asthenosphere rheology inferred from observations of the 2012 Indian Ocean earthquake. Nature 538:368–372. https://doi.org/10.1038/nature19787
Huang J, Yang A, Zhong S (2013) Constraints of the topography, gravity and volcanism on Venusian mantle dynamics and generation of plate tectonics. Earth Planet Sci Lett 362:207–214. https://doi.org/10.1016/j.epsl.2012.11.051
Ishii T et al (2018) Complete agreement of the post-spinel transition with the 660-km seismic discontinuity. Sci Rep 8:6358. https://doi.org/10.1038/s41598-018-24832-y
Ivanov MA, Head JW (2011) Global geological map of Venus. Planet Space Sci 59:1559–1600. https://doi.org/10.1016/j.pss.2011.07.008
Jacobson SA, Rubie DC, Hernlund J, Morbidelli A, Nakajima M (2017) Formation, stratification, and mixing of the cores of Earth and Venus. Earth Planet Sci Lett 474:375–386. https://doi.org/10.1016/j.epsl.2017.06.023
James PB, Zuber MT, Phillips RJ (2013) Crustal thickness and support of topography on Venus. J Geophys Res, Planets 118:859–875. https://doi.org/10.1029/2012je004237
Jellinek AM, Jackson MG (2015) Connections between the bulk composition, geodynamics and habitability of Earth. Nat Geosci 8:587–593. https://doi.org/10.1038/NGEO2488
Jellinek AM, Lenardic A, Manga M (2002) The influence of interior mantle temperature on the structure of plumes: heads for Venus, tails for the Earth. Geophys Res Lett 29:1532. https://doi.org/10.1029/2001gl014624
Jiménez-Díaz A, Ruiz J, Kirby JF, Romeo I, Tejero R, Capote R (2015) Lithospheric structure of Venus from gravity and topography. Icarus 260:215–231. https://doi.org/10.1016/j.icarus.2015.07.020
Johnson CL, Richards MA (2003) A conceptual model for the relationship between coronae and large-scale mantle dynamics on Venus. J Geophys Res 108:5058. https://doi.org/10.1029/2002je001962
Johnson CL, Sandwell DT (1994) Lithospheric flexure on Venus. Geophys J Int 119:627–647. https://doi.org/10.1111/j.1365-246x.1994.tb00146.x
Kankanamge DGJ, Moore WB (2016) Heat transport in the Hadean mantle: from heat pipes to plates. Geophys Res Lett 43:3208–3214. https://doi.org/10.1002/2015gl067411
Karato S-I (2010) Rheology of the deep upper mantle and its implications for the preservation of the continental roots: a review. Tectonophysics 481:82–98. https://doi.org/10.1016/j.tecto.2009.04.011
Karimi S, Dombard AJ (2017) Studying lower crustal flow beneath Mead basin: implications for the thermal history and rheology of Venus. Icarus 282:34–39. https://doi.org/10.1016/j.icarus.2016.09.015

Karimi S, Ojha L, Lewis K (2018) Searching for larger buried craters on Venus. In: 48th lunar planet sci conf, #2831

Karlsson RVMK, Cheng KW, Crameri F, Rolf T, Uppalapati S, Werner SC (2020) Implications of anomalous crustal provinces for Venus' resurfacing history. J Geophys Res, Planets 125:e2019JE006340. https://doi.org/10.1029/2019je006340

Katayama I (2021) Strength models of the terrestrial planets and implications for their lithospheric structure and evolution. Prog Earth Planet Sci 8:1. https://doi.org/10.1186/s40645-020-00388-2

Kiefer WS (2013) Making Ishtar Terra, Venus: mobile lid tectonic, continental crust, and implications for liquid water and planetary evolution. In: 44th lunar planet sci conf, #2541

Kiefer WS, Hager BH (1991) A mantle plume model for the equatorial highlands of Venus. J Geophys Res 96:20947–20966. https://doi.org/10.1029/91JE02221

Kiefer WS, Weller MB (2021) Venus, Earth's divergent twin: observations constraining the transition from a mobile lid planet to a stagnant lid planet. In: 52nd lunar planet sci conf, #1792

Kiefer WS, Richards MA, Hager BH (1986) A dynamic model of Venus's gravity field. Geophys Res Lett 13:14–17. https://doi.org/10.1029/GL013i001p00014

King SD (2018) Venus resurfacing constrained by geoid and topography. J Geophys Res, Planets 123:1041–1060. https://doi.org/10.1002/2017je005475

Koch DM, Manga M (1996) Neutrally buoyant diapirs: a model for Venus coronae. Geophys Res Lett 23:225–228. https://doi.org/10.1029/95GL03776

Kohlstedt Dl, Hansen LN (2015) Constitutive equations, rheological behavior, and viscosity of rocks. Treatise Geophys 2:441–472. https://doi.org/10.1016/b978-0-444-53802-4.00042-7

Kohlstedt DL, Evans B, Mackwell SJ (1995) Strength of the lithosphere: constraints imposed by laboratory experiments. J Geophys Res, Solid Earth 100:17587–17602. https://doi.org/10.1029/95jb01460

Konopliv AS, Yoder CF (1996) Venusian k_2 tidal Love number from Magellan and PVO Tracking Data. Geophys Res Lett 23:1857–1860. https://doi.org/10.1029/96gl01589

Konopliv AS, Banerdt WB, Sjogren WL (1999) Venus gravity: 180th degree and order model. Icarus 139:3–18. https://doi.org/10.1006/icar.1999.6086

Korenaga J (2008) Urey ratio and the structure and evolution of Earth's mantle. Rev Geophys 46:2007RG000241. https://doi.org/10.1029/2007rg000241

Korenaga J (2017) Pitfalls in modeling mantle convection with internal heat production. J Geophys Res, Solid Earth 122:4064–4085. https://doi.org/10.1002/2016jb013850

Kremic T, Amato M, Gilmore M, Kiefer W, Johnson N, Sauder J, Hunter G, Thompson T (2021) Venus surface platform study final report. NASA Glen Research Center NP-2021-11-102-GRC, 50 pages. LPI contribution 2660. https://www.lpi.usra.edu/vexag/documents/reports/Venus-Surface-Platform-Study-Final_11-4-21.pdf

Kreslavsky MA, Ivanov MA, Head JW (2015) The resurfacing history of Venus: constraints from buffered crater densities. Icarus 250:438–450. https://doi.org/10.1016/j.icarus.2014.12.024

Labrosse S (2015) Thermal evolution of the core with a high thermal conductivity. Phys Earth Planet Inter 247:36–55. https://doi.org/10.1016/j.pepi.2015.02.002

Labrosse S, Hernlund JW, Coltice N (2007) A crystallizing dense magma ocean at the base of the Earth's mantle. Nature 450:866–869. https://doi.org/10.1038/nature06355

Landuyt W, Bercovici D (2009) Variations in planetary convection via the effect of climate on damage. Earth Planet Sci Lett 277:29–37. https://doi.org/10.1016/j.epsl.2008.09.034

Lawrence JF, Shearer PM (2008) Imaging mantle transition zone thickness with SdS-SS finite-frequency sensitivity kernels. Geophys J Int 174:143–158. https://doi.org/10.1111/j.1365-246x.2007.03673.x

Lay T, Hernlund J, Buffett BA (2008) Core–mantle boundary heat flow. Nat Geosci 1:25–32. https://doi.org/10.1038/ngeo.2007.44

Le Feuvre M, Wieczorek MA (2011) Nonuniform cratering of the Moon and a revised crater chronology of the inner solar system. Icarus 214:1–20. https://doi.org/10.1016/j.icarus.2011.03.010

Leftwich TE, von Frese RRB, Kim HR, Noltimier HC, Potts LV, Roman DR, Tan L (1999) Crustal analysis of Venus from Magellan Satellite Observations at Atalanta Planitia, Beta Regio, and Thetis Regio. J Geophys Res, Planets 104:8441–8462. https://doi.org/10.1029/1999je900007

Lenardic A, Crowley JW (2012) On the notion of well-defined tectonic regimes for terrestrial planets in this solar system and others. Astrophys J 755:132. https://doi.org/10.1088/0004-637x/755/2/132

Lenardic A, Moresi L-N, Mühlhaus H (2003) Longevity and stability of cratonic lithosphere: insights from numerical simulations of coupled mantle convection and continental tectonics. J Geophys Res, Solid Earth 108:2303. https://doi.org/10.1029/2002jb001859

Lenardic A, Jellinek AM, Moresi L-N (2008) A climate induced transition in the tectonic style of a terrestrial planet. Earth Planet Sci Lett 271:34–42. https://doi.org/10.1016/j.epsl.2008.03.031

Lenardic A, Jellinek AM, Foley B, O'Neill C, Moore WB (2016) Climate-tectonic coupling: variations in the mean, variations about the mean, and variations in mode. J Geophys Res 121:1831–1864. https://doi.org/10.1002/2016JE005089

Lenardic A, Weller M, Höink T, Seales J (2019) Toward a boot strap hypothesis of plate tectonics: feedbacks between plates, the asthenosphere, and the wavelength of mantle convection. Phys Earth Planet Inter 296:106299. https://doi.org/10.1016/j.pepi.2019.106299

Lenardic A, Seales J, Moore W, Weller M (2021) Convective and tectonic plate velocities in a mixed heating mantle. Geochem Geophys Geosyst 22:e2020GC009278. https://doi.org/10.1029/2020GC009278

Leng W, Zhong S (2008) Controls on plume heat flux and plume excess temperature. J Geophys Res 113:B04408. https://doi.org/10.1029/2007jb005155

Leng W, Zhong S (2009) More constraints on internal heating rate of the Earth's mantle from plume observations. Geophys Res Lett 36:L02306. https://doi.org/10.1029/2008gl036449

Loddoch A, Stein C, Hansen U (2006) Temporal variations in the convective style of planetary mantles. Earth Planet Sci Lett 251:79–89. https://doi.org/10.1016/j.epsl.2006.08.026

Lourenço DL, Rozel AB, Tackley PJ (2016) Melting-induced crustal production helps plate tectonics on Earth-like planets. Earth Planet Sci Lett 439:18–28. https://doi.org/10.1016/j.epsl.2016.01.024. 2016

Lourenço DL, Rozel AB, Gerya T, Tackley PJ (2018) Efficient cooling of rocky planets by intrusive magmatism. Nat Geosci 11:322–327. https://doi.org/10.1038/s41561-018-0094-8

Lourenço DL, Rozel AB, Ballmer MD, Tackley PJ (2020) Plutonic-squishy lid: a new global tectonic regime generated by intrusive magmatism on Earth-like planets. Geochem Geophys Geosyst 21:e2019GC008756. https://doi.org/10.1029/2019gc008756

Mackwell SJ, Zimmermann ME, Kohlstedt DL (1998) High-temperature deformation of dry diabase with application to tectonics on Venus. J Geophys Res, Solid Earth 103:975–984. https://doi.org/10.1029/97jb02671

Maia JS, Wieczorek MA (2022) Lithospheric structure of Venusian crustal plateaus. J Geophys Res 127:e2021JE007004. https://doi.org/10.1029/2021JE007004

Mao H-K, Hu Q, Yang L, Liu J, Kim DY, Meng Y, Zhang L, Prakapenka VB, Yang W, Mao WL (2021) When water meets iron at Earth's core-mantle boundary. Natl Sci Rev 4:870–878. https://doi.org/10.1093/nsr/nwx109

Margot J-L, Campbell DB, Giorgini JD, Jao JS, Snedeker LG, Ghigo FD, Bonsall A (2021) Spin state and moment of inertia of Venus. Nat Astron 5:676–683. https://doi.org/10.1038/s41550-021-01339-7

Masuti S, Barbot SD, Karato S-I, Feng L, Banerjee P (2016) Upper-mantle water stratification inferred from observations of the 2012 Indian Ocean earthquake. Nature 538:373–377. https://doi.org/10.1038/nature19783

McGill GE (1994) Hotspot evolution and Venusian tectonic style. J Geophys Res 99:23149–23161. https://doi.org/10.1029/94JE02319

McKenzie DP, Roberts JM, Weiss NO (1974) Convection in the Earth's mantle: towards a numerical simulation. J Fluid Mech 62:465. https://doi.org/10.1017/s0022112074000784

McKinnon WB, Zahnle KJ, Ivanov BA, Melosh HJ (1997) Cratering on Venus: models and observations. In: Bougher SW, Hunten DM, Phillips RJ (eds) Venus II: geology, geophysics, atmosphere, and solar wind environment. University of Arizona Press, Tucson, pp 969–1014

McNamara AK, Zhong S (2005) Degree-one mantle convection: dependence on internal heating and temperature-dependent rheology. Geophys Res Lett 32:L01301. https://doi.org/10.1029/2004gl021082

Meier TG, Bower DJ, Lichtenberg T, Tackley PJ, Demory B-O (2021) Interior dynamics of tidally locked super-earths: the case of LHS 3844b. Astrophys J Lett 908:L48. https://doi.org/10.3847/2041-8213/abe400

Miyagoshi T, Kameyama M, Ogawa M (2020) Tectonic plates in 3D mantle convection model with stress-history-dependent rheology. Earth Planets Space 72:70. https://doi.org/10.1186/s40623-020-01195-1

Mocquet A, Rosenblatt P, Dehant V, Verhoeven O (2011) The deep interior of Venus, Mars, and the Earth: a brief review and the need for planetary surface-based measurements. Planet Space Sci 59:1048–1061. https://doi.org/10.1016/j.pss.2010.02.002

Moore WB (2008) Heat transport in a convecting layer heated from within and below. J Geophys Res 113:B11407. https://doi.org/10.1029/2006JB004778

Moore WB, Schubert G (1997) Venusian crustal and lithospheric properties from nonlinear regressions of highland geoid and topography. Icarus 128:415–428. https://doi.org/10.1006/icar.1997.5750

Moore WB, Webb AAG (2013) Heat-pipe Earth. Nature 501:501–505. https://doi.org/10.1038/nature12473

Moore WB, Simon JI, Webb AAG (2017) Heat-pipe planets. Earth Planet Sci Lett 474:13–19. https://doi.org/10.1016/j.epsl.2017.06.015

Moresi L, Solomatov V (1998) Mantle convection with a brittle lithosphere: thoughts on the global tectonic styles of the Earth and Venus. Geophys J Int 133:669–682. https://doi.org/10.1046/j.1365-246x.1998.00521.x

Moruzzi SA, Kiefer WS (2020) Thrust faulting on Venus: tectonic modeling of the Vedma Dorsa Ridge Belt. In: 51st lunar planet sci conf, #1430

Moyen J-F, van Hunen J (2012) Short-term episodicity of Archaean plate tectonics. Geology 40:451–454. https://doi.org/10.1130/g322894.1

Mulyukova E, Bercovici D (2020) Mantle convection in terrestrial planets. In: In Oxford research encyclopedia of planetary science. https://doi.org/10.1093/acrefore/9780190647926.013.109

Murakami M, Hirose K, Kawamura K, Sata N, Ohishi Y (2004) Post-Perovskite phase transition in $MgSiO_3$. Science 304:855–858. https://doi.org/10.1126/science.1095932

Nakagawa T, Nakakuki T, Iwamori H (2015) Water circulation and global mantle dynamics: insight from numerical modeling. Geochem Geophys Geosyst 16:1449–1464. https://doi.org/10.1002/2014GC005701

Namiki N (1995) Tectonics and volcanism on Venus: constraints from topographic relief, impact cratering, and degassing. PhD Thesis, Massachusetts Institute of Technology. 240 p

Namiki N, Solomon SC (1993) The gabbro-eclogite phase transition and the elevation of mountain belts on Venus. J Geophys Res 98:15025. https://doi.org/10.1029/93je01626

Nimmo F (2002) Why does Venus lack a magnetic field? Geology 30:987. https://doi.org/10.1130/0091-7613(2002)030<0987:wdvlam>2.0.co;2

Nimmo F (2015) Thermal and compositional evolution of the core. In: Schubert G (ed) Treatise on geophysics, 2015, vol 9. Elsevier, Amsterdam, pp 201–219. https://doi.org/10.1016/b978-0-444-53802-4.00160-3. Chap 9.08

Nimmo F, McKenzie D (1998) Volcanism and tectonics on Venus. Annu Rev Earth Planet Sci 26:23–51. https://doi.org/10.1146/annurev.earth.26.1.23

Noack L, Breuer D, Spohn T (2012) Coupling the atmosphere with interior dynamics: implications for the resurfacing of Venus. Icarus 217:484–498. https://doi.org/10.1016/j.icarus.2011.08.026

Ogawa M (2000) Coupled magmatism–mantle convection system with variable viscosity. Tectonophysics 322:1–18. https://doi.org/10.1016/s0040-1951(00)00054-8

Ogawa M (2018) The effects of magmatic redistribution of heat producing elements on the lunar mantle evolution inferred from numerical models that start from various initial states. Planet Space Sci 151:43–55. https://doi.org/10.1016/j.pss.2017.10.015

Ogawa M, Yanagisawa T (2011) Numerical models of Martian mantle evolution induced by magmatism and solid-state convection beneath stagnant lithosphere. J Geophys Res 116:E08008. https://doi.org/10.1029/2010je003777

Ogawa M, Yanagisawa T (2014) Mantle evolution in Venus due to magmatism and phase transitions: from punctuated layered convection to whole-mantle convection. J Geophys Res, Planets 119:867–883. https://doi.org/10.1002/2013je004593

Ohtani E (2020) The role of water in Earth's mantle. Nat Sci Rev 7:224–232. https://doi.org/10.1016/j.pss.2017.10.015

Okudaira T, Shigematsu N, Harigane Y, Yoshida K (2017) Grain size reduction due to fracturing and subsequent grain-size-sensitive creep in a lower crustal shear zone in the presence of a CO_2-bearing fluid. J Struct Geol 95:171–187. https://doi.org/10.1016/j.jsg.2016.11.001

O'Neill C (2021) End-member Venusian core scenarios: does Venus have an inner core? Geophys Res Lett 48:e2021GL095499. https://doi.org/10.1029/2021GL095499

O'Neill C, Lenardic A, Weller M, Moresi L, Quenette S, Zhang S (2016) A window for plate tectonics in terrestrial planet evolution? Phys Earth Planet Inter 255:80–92. https://doi.org/10.1016/j.pepi.2016.04.002

O'Neill C, Marchi S, Zhang S, Bottke W (2017) Impact-driven subduction on the Hadean Earth. Nat Geosci 10:793–797. https://doi.org/10.1038/ngeo3029

O'Neill C, Marchi S, Bottke W, Fu R (2020) The role of impacts on Archaean tectonics. Geology 48:174–178. https://doi.org/10.1130/g46533.1

O'Reilly TC, Davies GF (1981) Magma transport of heat on Io: a mechanism allowing a thick lithosphere. Geophys Res Lett 8:313–316. https://doi.org/10.1029/gl008i004p00313

O'Rourke JG (2020) Venus: a thick basal magma ocean may exist today. Geophys Res Lett 47:e2019GL086126. https://doi.org/10.1029/2019gl086126

O'Rourke JG, Korenaga J (2015) Thermal evolution of Venus with argon degassing. Icarus 260:128–140. https://doi.org/10.1016/j.icarus.2015.07.009

O'Rourke JG, Smrekar SE (2018) Signatures of lithospheric flexure and elevated heat flow in stereo topography at coronae on Venus. J Geophys Res, Planets 123:369–389. https://doi.org/10.1002/2017je005358

O'Rourke JG, Wolf AS, Ehlmann BL (2014) Venus: interpreting the spatial distribution of volcanically modified craters. Geophys Res Lett 41:8252–8260. https://doi.org/10.1002/2014GL062121

O'Rourke JG, Gillmann C, Tackley P (2018) Prospects for an ancient dynamo and modern crustal remanent magnetism on Venus. Earth Planet Sci Lett 502:46–56. https://doi.org/10.1016/j.epsl.2018.08.055

O'Rourke JG, Buz J, Fu RR, Lillis RJ (2019) Detectability of remanent magnetism in the crust of Venus. Geophys Res Lett 46:5678–5777. https://doi.org/10.1029/2019GL082725

O'Rourke JG, Wilson C, Borrelli M, Byrne PK, Dumoulin C, Ghail R, Gülcher A, Jacobson S, Korablev O, Spohn T, Way M, Weller M, Westall F (2023) Venus, the planet: introduction to the evolution of Earth's sister planet. Space Sci Rev

Orth CP, Solomatov VS (2011) The isostatic stagnant lid approximation and global variations in the Venusian lithospheric thickness. Geochem Geophys Geosyst 12:Q07018. https://doi.org/10.1029/2011gc003582

Orth CP, Solomatov VS (2012) Constraints on the Venusian crustal thickness variations in the isostatic stagnant lid approximation. Geochem Geophys Geosyst 13:Q11012. https://doi.org/10.1029/2012gc004377

Papuc AM, Davies GF (2012) Transient mantle layering and the episodic behaviour of Venus due to the 'Basalt Barrier' mechanism. Icarus 217:499–509. https://doi.org/10.1016/j.icarus.2011.09.024

Parmentier EM, Sotin C, Travis BJ (1994) Turbulent 3-D thermal convection in an infinite Prandtl number, volumetrically heated fluid: implications for mantle dynamics. Geophys J Int 116:241–251. https://doi.org/10.1111/j.1365-246x.1994.tb01795.x

Pauer M, Fleming K, Cadek O (2006) Modeling the dynamic component of the geoid and topography of Venus. J Geophys Res 111:E11012. https://doi.org/10.1029/2005je002511

Pettengill GH, Ford PG, Wilt RJ (1992) Venus surface radiothermal emission as observed by Magellan. J Geophys Res 97:13091–13102. https://doi.org/10.1029/92je01356

Phillips RJ, Izenberg NR (1995) Ejecta correlations with spatial crater density and Venus resurfacing history. Geophys Res Lett 22:1517–1520. https://doi.org/10.1029/95GL01412

Phillips JL, Russell CT (1987) Upper limit on the intrinsic magnetic field of Venus. J Geophys Res 92:2253. https://doi.org/10.1029/ja092ia03p02253

Phillips RJ, Raubertas RF, Arvidson RE, Sarkar IC, Herrick RR, Izenberg N, Grimm RE (1992) Impact craters and Venus resurfacing history. J Geophys Res 97:15923–15948. https://doi.org/10.1029/92je01696

Piskorz D, Elkins-Tanton LT, Smrekar SE (2014) Coronae formation on Venus via extension and lithospheric instability. J Geophys Res 119:2568–2582. https://doi.org/10.1002/2014JE004636

Plesa A-C, Breuer D (2021) The effects of intrusive magmatism on the mechanical lithosphere thickness of Venus. In: 52nd lunar planetary science conference, p #2130

Robin CMI, Jellinek AM, Thayalan V, Lenardic A (2007) Transient mantle convection on Venus: the paradoxical coexistence of highlands and coronae in the BAT region. Earth Planet Sci Lett 256:100–119. https://doi.org/10.1016/j.epsl.2007.01.016

Rolf T, Tackley PJ (2011) Focussing of stress by continents in 3D spherical mantle convection with self-consistent plate tectonics. Geophys Res Lett 38:L18301. https://doi.org/10.1029/2011gl048677

Rolf T, Coltice N, Tackley PJ (2014) Statistical cyclicity of the supercontinent cycle. Geophys Res Lett 41:2351–2358. https://doi.org/10.1002/2014gl059595

Rolf T, Capitanio FA, Tackley PJ (2018a) Constraints on mantle viscosity structure from continental drift histories in spherical mantle convection models. Tectonophysics 746:339–351. https://doi.org/10.1016/j.tecto.2017.04.031

Rolf T, Steinberger B, Werner SC, Sruthi U (2018b) Inferences on the mantle viscosity structure and the post-overturn evolutionary state of Venus. Icarus 313:107–123. https://doi.org/10.1016/j.icarus.2018.05.014

Romeo I, Turcotte DI (2010) Resurfacing on Venus. Planet Space Sci 58:1374–1380. https://doi.org/10.1016/j.pss.2010.05.022

Rosenblatt P, Dumoulin C, Marty J-C, Genova A (2021) Determination of Venus' interior structure with EnVision. Remote Sens 13:1624. https://doi.org/10.3390/rs13091624

Rozel A, Besserer J, Golabek GJ, Kaplan M, Tackley PJ (2014) Self-consistent generation of single-plume state for Enceladus using non-Newtonian rheology. J Geophys Res, Planets 119:416–439. https://doi.org/10.1002/2013je004473

Rozel AB, Golabek GJ, Näf R, Tackley PJ (2015) Formation of ridges in a stable lithosphere in mantle convection models with a viscoplastic rheology. Geophys Res Lett 42:4770–4777. https://doi.org/10.1002/2015gl063483

Rozel AB, Golabek GJ, Jain C, Tackley PJ, Gerya T (2017) Continental crust formation on early Earth controlled by intrusive magmatism. Nature 545:332–335. https://doi.org/10.1038/nature22042

Rudolph ML, Lekic V, Lithgow-Bertelloni C (2015) Viscosity jump in Earths mid-mantle. Science 350:1349–1352. https://doi.org/10.1126/science.aad1929

Ruedas T, Breuer D (2018) 'Isocrater' impacts: conditions and mantle dynamical responses for different impactor types. Icarus 306:94–115. https://doi.org/10.1016/j.icarus.2018.02.005

Ruiz J (2007) The heat flow during the formation of ribbon terrains on Venus. Planet Spa Sci 55:2063–2070. https://doi.org/10.1016/j.pss.2007.05.003

Ruiz J, Jimenez-Diáz A, Egea-Gonzalez I, Parro LM (2019) Comments on 'Using the viscoelastic relaxation of large impact craters to study the thermal history of Mars' (Karimi et al. (2016) Icarus 272:102–113) and 'Studying lower crustal flow beneath Mead basin: implications for the thermal history and rheology of Venus' (Karimi & Dombard (2017) Icarus 282:34–39). Icarus 322:221–226. https://doi.org/10.1016/j.icarus.2018.10.009

Russell CT, Zhang TL, Delva M, Magnes W, Strangeway WHY (2007) Lightning on Venus inferred from whistler-mode waves in the ionosphere. Nature 450:661–662. https://doi.org/10.1038/nature05930
Salvador A, Avice G, Breuer A, Gillmann C, Jacobson S, Marcq E, Raymond S, Sakuraba H, Scherf M, Way M (2023) Magma ocean, water, and the early atmosphere of Venus. Space Sci Rev
Sandu C, Lenardic A, McGovern P (2011) The effects of deep water cycling on planetary thermal evolution. J Geophys Res 116:B12404. https://doi.org/10.1029/2011jb008405
Sandwell DT, Schubert G (1992) Evidence for retrograde lithospheric subduction on Venus. Science 257:766–770. https://doi.org/10.1126/science.257.5071.766
Schmeling H, Marquart G, Ruedas T (2003) Pressure- and temperature-dependent thermal expansivity and the effect on mantle convection and surface observables. Geophys J Int 154:224–229. https://doi.org/10.1046/j.1365-246x.2003.01949.x
Schools JW, Montési LGJ (2018) The generation of barriers to melt ascent in the Martian lithosphere. J Geophys Res, Planets 123:47–66. https://doi.org/10.1002/2017je005396
Schubert G, Sandwell DT (1995) A global survey of possible subduction sites on Venus. Icarus 117:173–196. https://doi.org/10.1006/icar.1995.1150
Schulz F, Tosi N, Plesa A-C, Breuer D (2019) Stagnant-lid convection with diffusion and dislocation creep rheology: influence of a non-evolving grain size. Geophys J Int 220:18–36. https://doi.org/10.1093/gji/ggz417
Sifré D, Gardés E, Massuyeau M, Hashim L, Hier-Majumder S, Gaillard F (2014) Electrical conductivity during incipient melting in the oceanic low-velocity zone. Nature 509:81–85. https://doi.org/10.1038/nature13245
Sleep NH (1990) Hotspots and mantle plumes: some phenomenology. J Geophys Res 95:6715–6736. https://doi.org/10.1029/jb095ib05p06715
Smrekar SE (1994) Evidence for active hotspots on Venus from analysis of Magellan Gravity Data. Icarus 112:2–26. https://doi.org/10.1006/icar.1994.1166
Smrekar SE, Parmentier EM (1996) The interaction of mantle plumes with surface thermal and chemical boundary layers: applications to hotspots on Venus. J Geophys Res, Solid Earth 101:5397–5410. https://doi.org/10.1029/95jb02877
Smrekar SE, Phillips RJ (1991) Venusian highlands: geoid to topography ratios and their implications. Earth Planet Sci Lett 107:582–597. https://doi.org/10.1016/0012-821X(91)90103-O
Smrekar SE, Sotin C (2012) Constraints on mantle plumes on Venus: implications for volatile history. Icarus 217:510–523. https://doi.org/10.1016/j.icarus.2011.09.011
Smrekar SE, Stofan ER (1997) Corona formation and heat loss on Venus by coupled upwelling and delamination. Science 277:1289–1294. https://doi.org/10.1126/science.277.5330.1289
Smrekar SE, Hoogenboom T, Stofan ER, Martin P (2010) Recent hotspot volcanism on Venus from VIRTIS emissivity data. Science 328:605–608. https://doi.org/10.1126/science.1186785
Smrekar SE, Davaille A, Sotin C (2018) Venus interior structure and dynamics. Space Sci Rev 214:88. https://doi.org/10.1007/s11214-018-0518-1
Solomatov VS (1995) Scaling of temperature- and stress-dependent viscosity convection. Phys Fluids 7:266–274. https://doi.org/10.1063/1.868624
Solomatov VS (2003) Initiation of subduction by small-scale convection. J Geophys Res, Solid Earth 109:B01412. https://doi.org/10.1029/2003jb002628
Solomatov VS, Moresi L-N (1996) Stagnant lid convection on Venus. J Geophys Res, Planets 101:4737–4753. https://doi.org/10.1029/95je03361
Solomon SC, Head JW (1990) Lithospheric flexure beneath the Freya Montes Foredeep, Venus: constraints on lithospheric thermal gradient and heat flow. Geophys Res Lett 17:1393–1396. https://doi.org/10.1029/gl017i009p01393
Sotin C, Labrosse S (1999) Three-dimensional thermal convection in an iso-viscous, infinite Prandtl number fluid heated from within and from below: applications to the transfer of heat through planetary mantles. Phys Earth Planet Inter 112:171–190. https://doi.org/10.1016/s0031-9201(99)00004-7
Spada G, Sabadini R, Boschi E (1996) The spin and inertia of Venus. Geophys Res Lett 23:1997–2000. https://doi.org/10.1029/96gl01765
Spencer DC, Katz RF, Hewitt IJ (2020) Magmatic intrusions control Io's crustal thickness. J Geophys Res 125:e2020JE006443. https://doi.org/10.1029/2020JE006443
Stein C, Schmalzl J, Hansen U (2004) The effect of rheological parameters on plate behaviour in a self-consistent model of mantle convection. Phys Earth Planet Inter 142:225–255. https://doi.org/10.1016/j.pepi.2004.01.006
Stein C, Lowman JP, Hansen U (2013) The influence of mantle internal heating on lithospheric mobility: implications for super-earths. Earth Planet Sci Lett 361:448–459. https://doi.org/10.1016/j.epsl.2012.11.011

Steinberger B, Werner SC, Torsvik TH (2010) Deep versus shallow origin of gravity anomalies, topography and volcanism on Earth, Venus and Mars. Icarus 207:564–577. https://doi.org/10.1016/j.icarus.2009.12.025

Stern RJ, Gerya T, Tackley PJ (2018) Stagnant lid tectonics: perspectives from silicate planets, dwarf planets, large moons, and large asteroids. Geosci Front 9:103–119. https://doi.org/10.1016/j.gsf.2017.06.004

Stevenson DJ (2003) Styles of mantle convection and their influence on planetary evolution. C R Géosci 335:99–111. https://doi.org/10.1016/s1631-0713(03)00009-9

Stevenson DJ (2010) Planetary magnetic fields: achievements and prospects. Space Sci Rev 152:651–664. https://doi.org/10.1007/s11214-009-9572-z

Stevenson DJ, Cutts J, Mimoun D, Arrowsmith S, Banerdt B, Blom P, Brageot E, Brissaud Q, Chin G, Gao P, Garcia R, Hall J, Hunter G, Jackson J, Kerzhanovic V, Kiefer W, Komjathy A, Lee C, Lognonné P, Lorenz R, Majid W, Majorradi M, Nolet G, O'Rourke J, Rolland L, Schubert G, Simons M, Sotin C, Spilker T, Tsai V (2015) Probing the interior structure of Venus. Keck Institute of Space Studies, California Institute of Technology, Pasadena. https://doi.org/10.26206/C1CX-EV12. 85 pages

Stofan ER, Smrekar SE (2005) Large topographic rises, coronae, large flow fields and large volcanoes on Venus: evidence for mantle plumes? In: Foulger GR, Natland JH, Presnall DC, Anderson DL (eds) Plates, plumes, and paradigms. Geological Society of America special papers, vol 388, p 861. https://doi.org/10.1130/SPE388

Stofan ER, Bindschadler D, Parmentier EM, Head J (1991) Corona structures on Venus: models of origin. J Geophys Res 96:20933–20946. https://doi.org/10.1029/91JE02218

Stofan ER, Sharpton VL, Schubert G, Baer G, Bindschadler DL, Janes DM, Squyres SW (1992) Global distribution and characteristics of coronae and related features on Venus: implications for origin and relation to mantle processes. J Geophys Res 97:13347–13378. https://doi.org/10.1029/92je01314

Strom RG, Schaber GG, Dawson DD (1994) The global resurfacing of Venus. J Geophys Res 99:10899–10926. https://doi.org/10.1029/94je00388

Tackley PJ (1996) On the ability of phase transitions and viscosity layering to induce long wavelength heterogeneity in the mantle. Geophys Res Lett 23:1985–1988. https://doi.org/10.1029/96gl01980

Tackley PJ (2000) Self-consistent generation of tectonic plates in time-dependent, three-dimensional mantle convection simulations. Geochem Geophys Geosyst 1:2000GC000036. https://doi.org/10.1029/2000gc000036

Tkalčić H, Young M, Muir JB, Davies DR, Mattesini M (2015) Strong, multi-scale heterogeneity in Earth's lowermost mantle. Sci Rep 5:18416. https://doi.org/10.1038/srep18416

Tosi N, Padovan S (2021) Mercury, Moon, Mars: surface expressions of mantle convection and interior evolution of stagnant-lid bodies. In: Marquardt H, Ballmer MD, Cottaar S, Konter J (eds) Mantle convection and surface expressions. AGU monograph series. Wiley, New York, pp 455–489

Tosi N, Yuen DA, Cadek O (2010) Dynamical consequences in the lower mantle with the post-perovskite phase change and strongly depth-dependent thermodynamic and transport properties. Earth Planet Sci Lett 298:229–243. https://doi.org/10.1016/j.epsl.2010.08.001

Trompert R, Hansen U (1998) Mantle convection simulations with rheologies that generate plate-like behaviour. Nature 395:686–689. https://doi.org/10.1038/27185

Trønnes RG, Baron MA, Eigenmann KR, Guren MG, Heyn BH, Løken A, Mohn CF (2019) Core formation, mantle differentiation and core-mantle interaction within Earth and the terrestrial planets. Tectonophysics 760:165–198. https://doi.org/10.1016/j.tecto.2018.10.021

Turcotte DL (1989) A heat pipe mechanism for volcanism and tectonics on Venus. J Geophys Res, Solid Earth 94:2779–2785. https://doi.org/10.1029/jb094ib03p02779

Turcotte DL (1993) An episodic hypothesis for Venusian tectonics. J Geophys Res 98:17061–17068. https://doi.org/10.1029/93je01775

Turcotte DL (1996) Magellan and comparative planetology. J Geophys Res, Planets 101:4765–4773. https://doi.org/10.1029/95je02295

Turcotte DL, Schubert G (2017) Geodynamics, 3rd edn. Cambridge University Press, Cambridge, 2017

Tyler RH, Henning WG, Hamilton CW (2015) Tidal heating in a magma ocean within Jupiter's moon Io. Astrophys J Suppl Ser 218:22. https://doi.org/10.1088/0067-0049/218/2/22

Uppalapati S, Rolf T, Crameri C, Werner SC (2020) Dynamics of lithospheric overturns and implications for Venus's surface. J Geophys Res, Planets 125:e2019JE006258. https://doi.org/10.1029/2019je006258

van Heck HJ, Tackley PJ (2008) Planforms of self-consistently generated plates in 3D spherical geometry. Geophys Res Lett 35:L19312. https://doi.org/10.1029/2008gl035190

Vesterholt Al, Petersen KD, Nagel TJ (2021) Mantle overturn and thermochemical evolution of a non-plate tectonic mantle. Earth Planet Sci Lett 569:117047. https://doi.org/10.1016/j.epsl.2021.117047

Vilella K, Deschamps F (2021) Heat-blanketed convection and its implications for the continental lithosphere. J Geophys Res, Solid Earth 126:e2020JB020695. https://doi.org/10.1029/2020jb020695

Way M, Del Genio AD (2020) Venusian habitable climate scenarios: modeling Venus through time and applications to slowly rotating Venus-like exoplanets. J Geophys Res, Planets 125:e2019JE006276. https://doi.org/10.1029/2019JE006276

Way M, Ostberg C, Foley BJ, Gillmann C, Höning D, Lammer H, O'Rourke JG, Persson M, Plesa AC, Salvador A, Scherf M, Weller M (2023) Synergies between Venus and exoplanetary observations. Space Sci Rev

Wei D, Yang A, Huang JS (2014) The gravity field and crustal thickness of Venus. Sci China Earth Sci 57:2025–2035. https://doi.org/10.1007/s11430-014-4824-5

Weidner DJ, Wang Y (2000) Phase transformations: implications for mantle structure. In: Karato SI, Forte A, Liebermann R, Masters G, Stixrude L (eds) Earth's deep interior: mineral physics and tomography from the atomic to the global scale. Geophysical monograph series, vol 117, pp 215–235. https://doi.org/10.1029/gm117p0215

Weller MB, Kiefer WS (2020) The physics of changing tectonic regimes: implications for the temporal evolution of mantle convection and the thermal history of Venus. J Geophys Res 125:e2019JE005960. https://doi.org/10.1029/2019je005960

Weller MB, Lenardic A (2012) Hysteresis in mantle convection: plate tectonics systems. Geophys Res Lett 39:L10202. https://doi.org/10.1029/2012gl051232

Weller MB, Lenardic A (2015) Diverging worlds: bi-stability, the evolution of terrestrial planets and its application to Venus and Earth. In: 46th lunar and planetary science conference, p #2670

Weller MB, Lenardic A (2016) The energetics and convective vigor of mixed-mode heating: velocity scalings and implications for the tectonics of exoplanets. Geophys Res Lett 43:9469–9474. https://doi.org/10.1002/2016gl069927

Weller MB, Lenardic A (2018) On the evolution of terrestrial planets: bi-stability, stochastic effects, and the non-uniqueness of tectonic states. Geosci Front 9:91–102. https://doi.org/10.1016/j.gsf.2017.03.001

Weller MB, Lenardic A, O'Neill C (2015) The effects of internal heating and large scale climate variations on tectonic bi-stability in terrestrial planets. Earth Planet Sci Lett 420:85–94. https://doi.org/10.1016/j.epsl.2015.03.021

Weller MB, Lenardic A, Moore WB (2016) Scaling relationships and physics for mixed heating convection in planetary interiors: isoviscous spherical shells. J Geophys Res, Solid Earth 121:7598–7617. https://doi.org/10.1002/2016jb013247

White SM, Crisp JA, Spera FJ (2006) Long-term volumetric eruption rates and magma budgets. Geochem Geophys Geosyst 7:Q03010. https://doi.org/10.1029/2005GC001002

Wicht J, Sanchez S (2019) Advances in geodynamo modelling. Geophys Astrophys Fluid Dyn 113:2–50. https://doi.org/10.1080/03091929.2019.1597074

Widemann T et al (2023) Venus evolution through time: key science questions, selected mission concepts and future investigations. Space Sci Rev

Wieczorek M (2007) Gravity and topography of the terrestrial planets. In: Treatise on geophysics, 2nd edn. Planets and moons, vol 10, pp 165–206. https://doi.org/10.1016/b978-044452748-6/00156-5

Wong T, Solomatov VS (2015) Towards scaling laws for subduction initiation on terrestrial planets: constraints from two-dimensional steady-state convection simulations. Prog Earth Planet Sci 2:18. https://doi.org/10.1186/s40645-015-0041-x

Wong T, Solomatov VS (2016) Variations in timing of lithospheric failure on terrestrial planets due to chaotic nature of mantle convection. Geochem Geophys Geosyst 17:1569–1585. https://doi.org/10.1002/2015gc006158

Yanagisawa T, Kameyama M, Ogawa M (2016) Numerical studies on convective stability and flow pattern in three-dimensional spherical mantle of terrestrial planets. Geophys J Int 206:1526–1538. https://doi.org/10.1093/gji/ggw226

Yang A, Huang JS, Wei D (2016) Separation of dynamic and isostatic components of the Venusian gravity and topography and determination of the crustal thickness of Venus. Planet Space Sci 129:24–31. https://doi.org/10.1016/j.pss.2016.06.001

Yoshida M (2004) Influence of two major phase transitions on mantle convection with moving and subducting plates. Earth Planets Space 56:1019–1033. https://doi.org/10.1186/bf03352544

Yoshida M (2008) Mantle convection with longest-wavelength thermal heterogeneity in a 3-D spherical model: degree one or two? Geophys Res Lett 35:L23302. https://doi.org/10.1029/2008gl036059

Yuan L, Ohtani E, Ikuta D, Kamada S, Tsuchiya J, Naohisa H, Ohishi Y, Suzuki A (2018) Chemical reactions between Fe and H_2O up to megabar pressures and implications for water storage in the Earth's mantle and core. Geophys Res Lett 45:1330–1338. https://doi.org/10.1002/2017GL075720

Yuen DA, Cserepes L, Schroeder BA (1998) Mesoscale structures in the transition zone: dynamical consequences of boundary layer activities. Earth Planets Space 50:1035–1045. https://doi.org/10.1186/bf03352198

Zampa LS, Tenzer R, Eshagh M, Pitoňák M (2018) Evidence of mantle upwelling/downwelling and localized subduction on Venus from the body-force vector analysis. Planet Space Sci 157:48–62. https://doi.org/10.1016/j.pss.2018.03.013

Zhang S, Christensen U (1993) Some effects of lateral viscosity variations on geoid and surface velocities induced by density anomalies in the mantle. Geophys J Int 114:531–547. https://doi.org/10.1111/j.1365-246x.1993.tb06985.x

Zolotov MY, Fegley Jr B, Lodders K (1997) Hydrous silicates and water on Venus. Icarus 130:475–494. https://doi.org/10.1006/icar.1997.5838

Authors and Affiliations

Tobias Rolf[1,2] · Matt Weller[3,4] · Anna Gülcher[5] · Paul Byrne[6,7] · Joseph G. O'Rourke[8] · Robert Herrick[9] · Evan Bjonnes[3,4] · Anne Davaille[10] · Richard Ghail[11] · Cedric Gillmann[12] · Ana-Catalina Plesa[13] · Suzanne Smrekar[14]

✉ T. Rolf
Tobias.Rolf@geo.uio.no

1 Centre for Earth Evolution and Dynamics (CEED), University of Oslo, Norway

2 Institute of Geophysics, University of Münster, Münster, Germany

3 Department of Earth, Environmental & Planetary Sciences, Brown University, Providence, RI, USA

4 Lunar and Planetary Institute/USRA, Houston, TX, USA

5 Institute of Geophysics, Department of Earth Sciences, ETH Zürich, Switzerland

6 Department of Marine, Earth, and Atmospheric Sciences, North Carolina State University, Raleigh, NC, USA

7 Department of Earth, Environmental, and Planetary Sciences, Washington University in St. Louis, St. Louis, MO, USA

8 School of Earth and Space Exploration, Arizona State University, Tempe, AZ, USA

9 Geophysical Institute, University of Alaska Fairbanks, AK, USA

10 Laboratoire FAST, CNRS and Université Paris-Saclay, Orsay, France

11 Earth Sciences, Royal Holloway, University of London, Egham, UK

12 Department of Earth, Environmental & Planetary Sciences, Rice University, Houston, TX, USA

13 German Aerospace Centre, Institute of Planetary Research, Planetary Physics, Berlin, Germany

14 Jet Propulsion Laboratory, NASA, Pasadena, CA, US