Eruption of crystal mush and the formation of steep-sided volcanic domes on Venus

Eruption of crystal mush and the formation of steep-sided volcanic domes on Venus: insight from picritic bodies near Marki, Cyprus.

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Highlights:

- Steep volcanic domes on Venus form by enigmatic eruption of viscous lava
- New model proposes that they represent fault-controlled extrusion of crystal mush
- Implies common magmatic origin for domes and associated, extensive basaltic terrains
- Extrusion mechanism could occur on other stagnant-lid planetary bodies

Declarations of interest: none

Abstract

Steep-sided domes are one of the most striking volcanic landforms on Venus. They may also be key to determining the range of magmatic processes operating on Venus as, in contrast to all other volcanic landforms, they likely represent eruption of viscous lava. Although there have been various explanations for the presence of high-viscosity lavas on a planet dominated by effusive basaltic volcanism, it is often assumed that they are silica-rich. This would necessitate either periodic, large-scale, extensive fractionation of basaltic magma in the Venusian crust, or a mechanism for re-melting an already silica-enriched lower crust. As such, determining the origin of steep-sided domes is important in constraining magmatic processes on Venus, and for understanding geological evolution of stagnant lid regime planets generally. Here, we use observations from the Marki region of the Troodos ophiolite, Cyprus, to propose an alternative model where steep-sided domes form by eruption of crystal mush from the same magmatic systems which fed extensive basaltic terrains with which domes are associated. Steep-sided volcanic landforms near Marki represent extrusion of ‘un-eruptible’, extremely olivine-rich mush onto the palaeo-seafloor, following cessation of widespread basaltic volcanism. Field relations suggest that these bodies formed by localised, repeated extrusion of crystal mush, fed by extensional faults tapping crustal magma chambers. Differential stress enabled eruption of viscous, non-Newtonian magmas with crystal contents >50 vol%, which then built up volcanic edifices on the seafloor. A similar, much larger-scale, mechanism can explain many features of steep-sided volcanic domes on Venus, including their intimate relationship with extensive, basaltic terrains, general morphology, and dome spatial and temporal clustering. This implies that domes share a common magmatic origin with the Venusian basaltic crust, rather than representing a discrete magmatic process, and that they represent periods of magmatic quiescence. It also implies that the contrasting morphology of these domes arises from a fundamental difference in eruptive style, from widespread effusive basaltic magmatism to localised, extensional fault-controlled extrusion of crystal mush. If correct, this mechanism might also explain formation of steep-sided volcanic edifices on other large, stagnant-lid regime planetary bodies.

1. Introduction: steep-sided volcanic domes on Venus

Of the terrestrial planets, Venus is the most similar to Earth in terms of size and density. However, on Venus there is no evidence for the global plate tectonic processes which dominate geological evolution on Earth. Instead, convective regime, crustal structure and volcanism on Venus are generally described by a ‘thick stagnant lid’ model, where a single plate of buoyant lithosphere inhibits mantle upwelling and active volcanism (Solomatov and Moresi, 1996). Volcanic activity within this convective regime can be assumed to relate mainly to plume-type upwellings, i.e. thermal anomalies inducing mantle melting beneath, and periodic puncturing of, the stagnant lid. Due to surface conditions which preclude the presence of liquid water or ice, Venus is a volcanic planet, with 75% of its surface interpreted to be primary volcanics, and 25% categorised as tectonic, i.e. volcanic origin but reworked by tectonics (Kaula, 1990). Geochemical data from Venus is limited, with 3 sites from the Venera-Vega landers returning detailed major element components of Venusian soil (Basilevsky, 1997). Compositions are similar to terrestrial basalts, from which it is inferred that Venus has a similar bulk silicate composition to the Earth, although compositions do have elevated MgO, suggestive of higher mantle melt fractions (Ivanov, 2015). More silica-rich melt compositions on Venus would require either extensive fractionation of primary (basaltic) mantle melts, or in the absence of subduction-related and/or hydrous mantle melting, some process involving re-melting of a Venusian crust that was already silica-enriched.

The surface of Venus is dominated by volcanic landforms (Saunders et al., 1991). Most volcanism is shield-type, with low angle slopes, most likely formed by effusive, basaltic volcanism (Crumpler et al., 1997; Ivanov, 2015; Kaula, 1990). The main Venusian landforms can be divided based on their morphological features (e.g. Tanaka et al. 1997), the principal ones being: shield plains, regional plains, lobate plains, tesserae and steep-sided domes (Ivanov, 2015). Shield plains make up 18.5% of Venus' surface (Hansen, 2005), and are generally similar in morphology, suggesting a common formation process. They are typically associated with small domes interpreted as monogenetic volcanoes formed at the same time as the shield plains (Guest et al., 1992; Ivanov, 2015). Regional plains make up approximately 40% of the surface and may be identified in some instances as lava flows extending dozens to hundreds of kilometres, related to large volcanic centres (Ivanov, 2015). Shield, regional and lobate plains are all easiest to ascribe as the result of effusive volcanism and varying amounts of tectonic deformation. These landforms are consistent with eruption of large volumes of basaltic lava, which, under the high temperature, high pressure conditions of the Venusian surface (average surface temperature of 462°C, and a surface pressure equivalent to a 1 km water column on Earth) can flow very considerable distances.

Tesserae differ significantly from the other landforms in that they are equant or slightly elongate massifs, characterised by ridges and grooves. Tesserae make up approximately 8% of Venus' surface and are the most tectonically deformed regions of Venus, although likely were also initially produced by effusive volcanism (Ivanov and Head, 2015). The only landform present on Venus that does not fit with the observations of low-silica, effusive volcanism are steep-sided domes, sometimes named 'pancake' domes. Steep-sided domes are unusual in that their morphology implies eruption of more viscous lava, or a markedly different eruption style (see Ivanov and Head, 1999, for a summary). These volcanic landforms often occur in lineations or clusters and have characteristically rounded shapes in plan-view, flat tops, clearly pronounced frontal scarps, high elevations (few hundred metres), radial fracture patterns, and sometimes, small volcanic craters and auxiliary necks (Ivanov and Head, 1999). They are on average 20 km in diameter but reach up to 60 km, and have volumes usually exceeding 100 km^3. Two thirds of steep-sided domes are spatially and stratigraphically related to shield plains (Ivanov and Head, 1999; Ivanov, 2015), with domes overlying, or partially embayed by lava flows. These observations have resulted in a variety of explanations for the elevated viscosity of the domes required to produce the steep sides.

Based in part on evidence for a basalt-dominated surface, comparisons were initially drawn between steep-sided domes and seamounts on Earth (Bridges, 1995). Subsequent statistical analysis showed that there is only morphological similarity between smaller seamounts and steep-sided domes (Smith, 1996). Larger seamounts tend towards more traditional 'pointed top' shapes, and diverge markedly from the pancake dome shape, which has characteristically flat top surfaces across all sizes of dome. This suggests a fundamental difference between steep-sided domes on Venus and seafloor volcanoes on Earth, possibly due to differences in magma composition and crystallinity, but also possibly due to magma volume, effusion rate, cooling rate, and/or the pre-existing topography (Smith, 1996). Alternatively, it has been suggested that steep-sided domes more closely resemble terrestrial rhyolitic domes. Fink et al. (1993) suggested that dome emplacement on Venus is consistent with a melt of similar viscosity to terrestrial rhyolite. Modelled cooling times, based on a constant volume theoretical approach to dome relaxation, return emplacement timescales of 650-7400 years (McKenzie et al., 1992), also implying lava viscosities and temperatures consistent with a rhyolitic composition. Geochemical modelling of the Venera-Vega lander compositions shows that rhyolite compositions can be produced by considerable fractionation of Venusian basalts (Shellnutt, 2018, 2013). Ivanov and Head (1999) instead suggested that the juxtaposition of steep-sided domes with basaltic terrains is more consistent with genesis of rhyolitic magma by crustal remelting associated with upwelling plumes. However, the re-melted crust would have to have been non-basaltic, and possibly granitic in composition to begin with, creating an additional quandary. Any model where rhyolitic domes are generated by plume-induced re-melting of the crust or extensive fractionation of basalt would also imply that domes should be more common than they are (Ivanov, 2015); instead, the relative rarity of these domes implies that they are formed via an unusual set of conditions. Eruption of rhyolitic magma might also be inconsistent with primary observations made on the steep-sided domes. The domes are characterised by smooth upper surfaces and only show signs of late-stage fractures (Stofan et al., 2000; Plaut et al., 2004). Stofan et al. (2000) calculated that for present surface conditions, rhyolitic lavas should quickly form thick crusts, ultimately resulting in a much blockier dome morphology than observed. Much higher surface temperature conditions would be required to prevent rhyolitic lavas forming distinct blocky, fractured dome surfaces (Anderson et al., 1998; Stofan et al., 2000). Comparison with radar properties of terrestrial silicic lava domes supports the assertion that Magellan data is instead consistent with a less evolved dome lava composition. In addition, Stofan et al. (2000) noted that depressions in the upper surfaces of some steep-sided domes imply more fluid dome interiors, again suggestive of a basaltic lava composition.

Alternative explanations for the elevated viscosity of dome-forming magmas include higher levels of crystallinity (e.g. Sakimoto and Zuber, 1995), or a magma rich in gas bubbles (Pavri et al., 1992), although a non-rhyolitic, volatile-rich magma seems unlikely, due to dominance of low viscosity, basaltic lava flows on Venus. Gregg and Fink (1996) and Bridges (1997) also took into account the sluggish character of basaltic eruptions due to the much greater atmospheric pressure and temperature on Venus than on Earth. Analysis of the ambient effects on basalts vs. rhyolites under Venusian and Earth-like conditions supports a basaltic composition for Venusian steep-sided domes (Bridges, 1997) consistent with the cooling models of Stofan et al. (2000). In addition, more recent mathematical modelling on dome emplacement (Quick et al., 2016) using a time-variable volume approach has shown that emplacement times are a lot quicker (2-16 years) than previous estimates (McKenzie et al., 1992), consistent with a basaltic-andesitic composition magma. However, a fundamental problem with invoking a basaltic or andesitic lava source for steep-sided domes is the obvious, distinct contrast in morphology between domes and surrounded volcanic terrains. It is not clear how a change in volatile content or crystal content could result in such a clear transition in eruptive style, without producing a range of intermediate landforms. Similarly, steep-sided domes are intimately related with basaltic terrains, suggesting that the two volcanic landforms are connected, but domes are comparatively rare, implying that the mechanism(s) forming them are somewhat anomalous. As such, despite observed inconsistences between steep-sided domes and features formed by high-viscosity, rhyolitic lavas (Stofan et al., 2000) a full explanation of the morphology of steep-sided domes, especially the stark contrast to the extensive shield plains with which they are associated, remains elusive.

Here we propose an alternative model for the formation of steep-sided volcanic domes which accounts for many of their observed features, but which also circumvents clear issues with advocating eruption of basaltic or rhyolitic lava. Based on field observations near Marki, Cyprus, we propose that these volcanic domes form by extrusion of crystal mush. As such, we propose that they share a common magmatic origin with basaltic terrains on Venus, but form by a fundamentally different extrusive process.

2. Marki: picritic domes within the Troodos ophiolite

The Troodos Massif, which forms one of the main tectonic units comprising the island of Cyprus in the Eastern Mediterranean, is an Upper Cretaceous age ophiolite formed in the Tethyan Ocean around 93-90 Ma (Mukasa and Ludden, 1987). The Troodos Massif is one of the most fully documented ophiolites in the world, and has been pivotal in aiding development of theories on plate tectonics and sea-floor spreading, and magmatic-tectonic-hydrothermal process associated with the formation of oceanic crust (Robertson, 2004). Results of mapping and scientific drilling allow a pseudo-stratigraphy of the main Troodos ophiolitic body to be constructed, consisting of a sequence of serpentinised hazburgite and ultramafic cumulates, overlain by layered and then massive gabbros, a sheeted dyke complex, and extrusive volcanic sequence including overlying volcaniclastic sediment (Dilek and Furnes, 2009; Gass, 1968).The extrusive volcanic sequence can be further divided into a lower basal unit (transitioning downwards into the sheeted dyke complex), and both lower pillow lava (LPL) and upper pillow lava (UPL) units. Differences in appearance allow LPL and UPL to be mapped in the field, although there are also important geochemical differences between the 2 units which have resulted in various theories for the tectonic setting in which the Troodos Ophiolite formed (e.g. Pearce and Robinson, 2010). Most likely, the Troodos complex formed in an extensional oceanic regime, either a back-arc or early fore-arc environment, with variable input from an incipient subduction system, evident from a marked boninitic signature in the UPL unit (Woelki et al., 2018).

UPLs are generally silica-undersaturated, often olivine-bearing basalts. Occasionally, more ultramafic, olivine-rich varieties occur towards the top of the UPL sequence (Malpas and Langdon, 1984). These picritic basalts and ultramafic rocks of the UPL were originally documented by Gass (1958) and Searle and Vokes (1969), and described as varying from shallow intrusive to extrusive. In the area adjacent to the village of Marki (or Margi), approximately 50 km SW from Nicosia, on the northern flank of the Troodos, small ultrabasic lava flows and picritic pillows and flows occur within the UPL. Olivines within these members are typically highly forsteritic (~Fo92) (Gass, 1958; Searle and Vokes, 1969). Malpas and Langdon (1984) noted that the position of ultramafic rocks near the top of the pillow-lava sequence suggests that they were extruded late-stage. They further noted that major element bulk rock chemistry and phenocryst composition demonstrated that removal or addition of olivine phenocrysts can account for the full compositional range of UPL members, following derivation from a basaltic parental magma. As such, olivine-rich members of the UPL were inferred to represent crystal-rich magmas formed from the same magmatic system which fed the entire UPL unit. Gass (1958) viewed olivine-rich picritic and ultramafic bodies in this region as representing both shallow intrusive and extrusive bodies. However, in contrast to the classification of Gass (1958), field relations of larger picritic bodies within this area provide robust evidence that they represent repeated extrusion of viscous, crystal-rich mush of ultramafic to mafic bulk composition, onto the seafloor via an unusual, fault-controlled mechanism. This same mechanism may provide insight into formation of steep-sided volcanic domes on Venus.

3. Field observations and petrology of a Marki picritic body

The region around Marki contains numerous discrete picritic bodies, as shown in Figure 1, based on original mapping and classification by Gass (1958). One of the largest and most prominent of these, ‘Picrite Hill’ is marked in Figure 1. This body sits on top of lavas of the

UPL unit, marked by a shallow angle contact, and adjacent to a N-S trending extensional fault. Highly localised accumulation of 10s of meters of volcaniclastic sediment adjacent to this fault is characteristic of lava/umber relations in the UPL regionally (Constantinou and Govett, 1973; Robertson and Hudson, 1973), and implies that the fault was active during the Cretaceous, forming a half-graben in which sediments accumulated. From Figure 1 it is also apparent that all substantial picritic bodies within the Marki region occur adjacent to extensional faults.

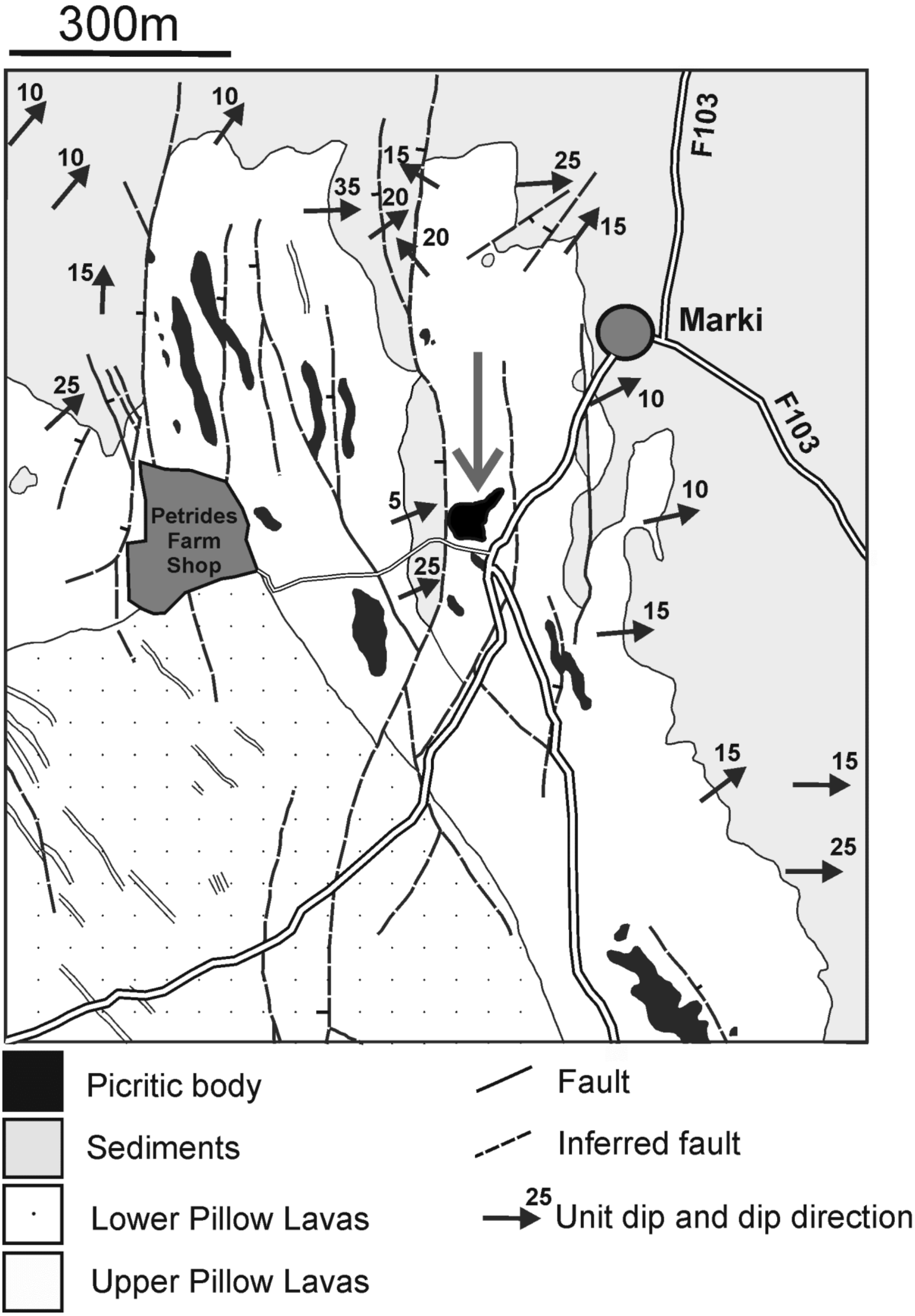

Figure 1. Sketch geological map showing picritic volcanic bodies near Marki, Cyprus, adapted from Gass (1958). Large grey arrow marks Picrite Hill.

Picrite Hill is an approximately 50m diameter body with a tongue extending an additional 40m to the NE. The body rises to a prominent topographic high towards the SW, approximately 25-30m above the underlying UPL pillow lava basement (Figure 2). This central, plug-like portion of the body is massive at the base, but grades upwards into a series of distinct, meter-scale units dipping around 30-35° to the NE (Figures 3), supporting a steep edifice (Figure 2a,b). Towards the top of the body, weak columnar jointing is also evident. The picrite has a uniform, brown appearance in the field which differentiates it from the underlying, grey UPL pillowed terrain, and contains up to 8mm diameter, bright green olivine phenocrysts (Figures 2a, 3a,b). Olivine content varies between units, exceeding 50% by volume in places. Individual units towards the top of the body can be traced laterally, and variably thin towards the NE. This is consistent with elongation of the body in the same direction. The NE elongation of Picrite Hill consists of thinner, interlayered units of lower-crystallinity. Underlying UPLs show a similar, although much lower angle tilt to the NE of 5-10° locally, and up to 30° regionally. As such, field evidence indicates that Picrite Hill consists of interlayered, crystal-rich lavas (>50% in places), with more mobile, runnier lavas (phenocryst contents 30%+), repeatedly erupted onto a shallow palaeoslope on the Upper Cretaceous, pillowed basaltic sea-floor. More elongate picritic bodies in the surrounding area indicate that elsewhere, picritic lavas accumulated in half-grabens on a seafloor which was more faulted and steeply dipping. Poorly-developed slickenslide surfaces on the southern margin of Picrite Hill indicate that the entire body moved downslope, with less viscous lavas flowing further, forming a slight elongation. The body was clearly extruded in a series of events, although poorly developed columnar jointing towards the top of the body indicates relatively quiescent conditions, and minimal flow of crystal-rich lavas. The blockier lower parts of the body, and some units with better developed columnar jointing, may have formed by intrusion of crystal-rich magma into the base of the edifice. Subsequently, the body, especially lowermost units, were cut by various basaltic dykelets (Figure 3a), most likely formed from late-stage fluids squeezed from the body as it cooled. These, in turn, are cross-cut by later, hydrothermal carbonate and zeolite-filled veins.

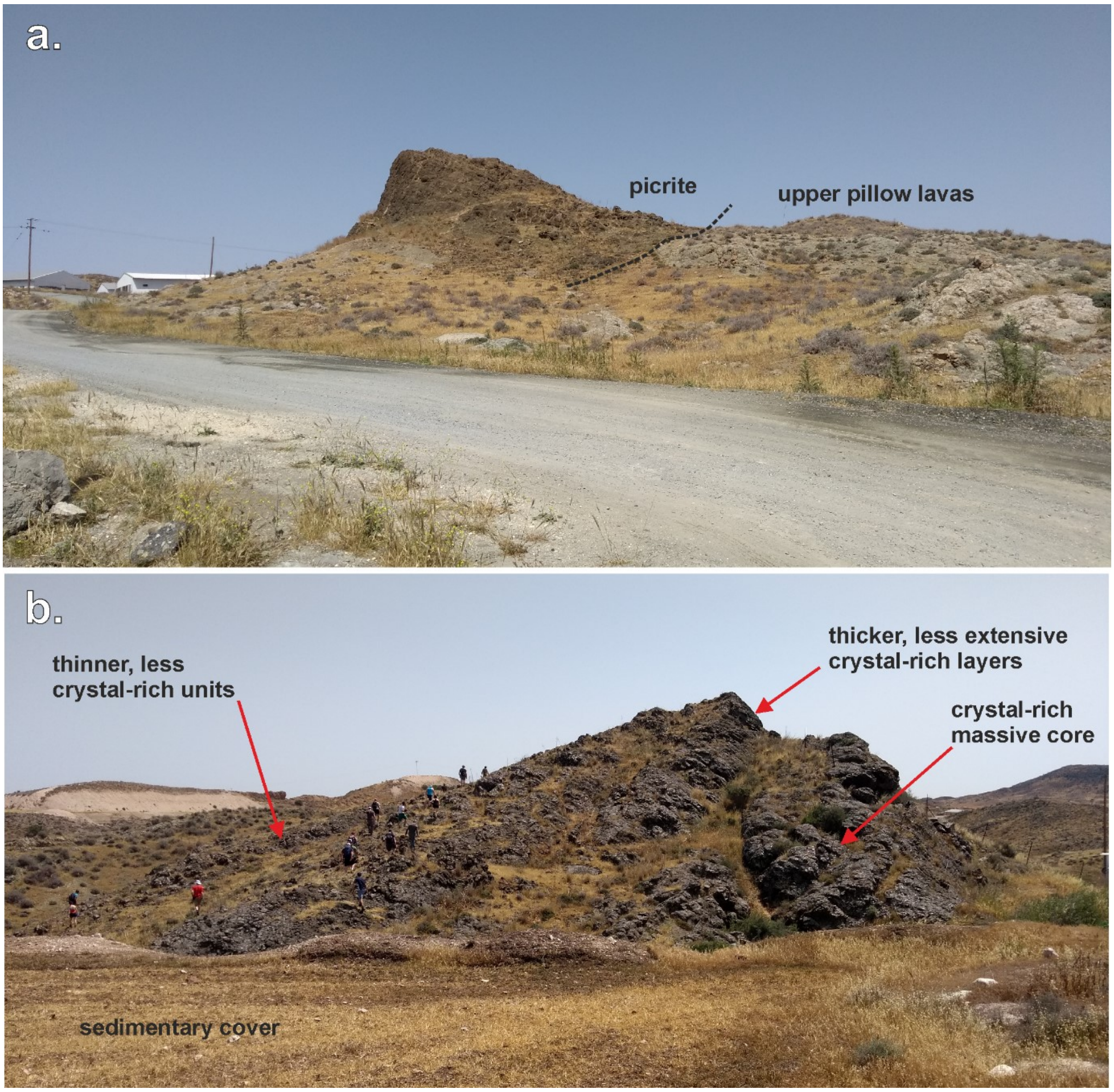

Figure 2 (COLOUR). Photos of Picrite Hill a) Looking WNW showing distinction between picrite and the underlying upper pillow lavas; b) looking E, where individual units are more easily discerned, some extending and thinning to the NE. The entire body can be viewed as interlayered viscous crystal mush with runnier lavas, with a low-angle contact with underlying Upper Pillow Lavas.

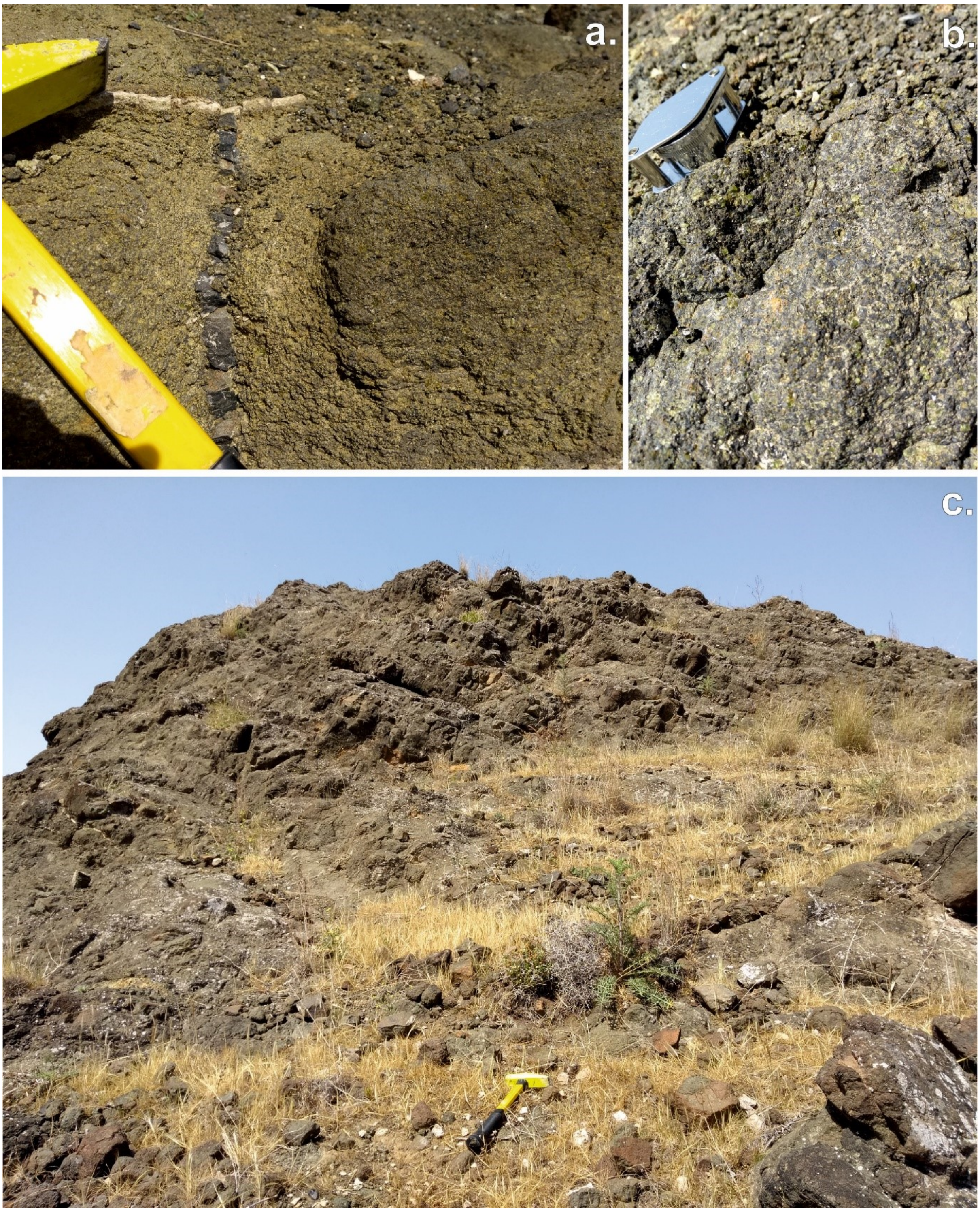

Figure 3 (COLOUR). a) Margins of the body are cut by basaltic dykelets (dark), representing late-stage fluids being squeezed out of the crystal mush, cross cut by later carbonate veins (white); b) crystal-rich unit in the central, more massive centre of the body showing typical appearance of the picrite. c). Photo of the SW portion of the body (looking approx.. N), showing transition from more massive to more layered units vertically upward, development of columnar jointing towards the top of the body, and the general 30°+ dip of the layers to the NE.

Sampling and petrographic analysis reveal variations in phenocryst content and groundmass between units, but also considerable internal variation within some units. All units are picritic, consisting of large, euhedral and sometimes euhedral-to-subhedral mm-sized phenocrysts of olivine, variably altered to serpentine, and some smaller, but again up to mm-sized, phenocrysts of augitic clinopyroxene. Groundmass is variable between units, and in some cases within thin sections from the same unit. In accordance with Malpas and Langdon (1984), picrites can be subdivided as vitrophyric and holocrystalline (Figure 4). Vitrophyric picrites are dominated by the presence of large (up to 8 mm sized) phenocrysts of olivine, ranging up to 50-60% by volume, set in a fine-grained to glassy groundmass, now extensively serpentinised. Olivine phenocrysts are unzoned, generally euhedral, sometimes exhibiting a weak cleavage, but with some slightly corroded boundaries. As noted by Gass (1958), olivines are highly forsteritic (Fo92). Clinopyroxene phenocrysts are much less abundant and generally small, although ranging up to 2 mm sized. Some acicular clinopyroxene is also noted. Holocrystalline picrites have groundmasses with variable grain size. The proportion of olivine phenocrysts is generally lower (40-50%) and olivine is generally slightly smaller and more extensively serpentinised. The groundmass in these units largely comprises plagioclase microlaths, intergrown with clinopyroxene and euhedral magnetite, and some rare orthopyroxene. In places the groundmass plagioclase can be relatively coarse, with laths ranging up to 0.5mm and better described as microphenocrysts. Some samples are, however, transitionary between both types, with groundmass being relatively coarse grained in places, and fine-grained and/or extensively serpentinised in other places, even within the same thin section. As such, petrographic investigation reveals that the body is internally chaotic and highly laterally and vertically variable, as expected from a volcanic edifice formed by multiple events. Phenocryst content of most samples implies that flow under surface conditions would be very limited, consistent with field observations which indicate flow of a few 10s of meters even for the most crystal-poor lavas. Composition of all samples, especially the abundance of unzoned, forsteritic olivine, and the basaltic composition groundmass, is consistent with picrites representing a crystal mush, tapped from magma chambers. Close association of picrites with underlying UPL units is consistent with a common origin, where relatively primitive UPL magmas formed by small degrees of fractionation of olivine and minor clinopyroxene from a basaltic parent, and picrites represent the cognate olivine-rich mush extruded, later on, from the same magmatic system following a prolonged period of fractionation, UPL eruption and rejuvenation of magmatic systems by primitive melt.

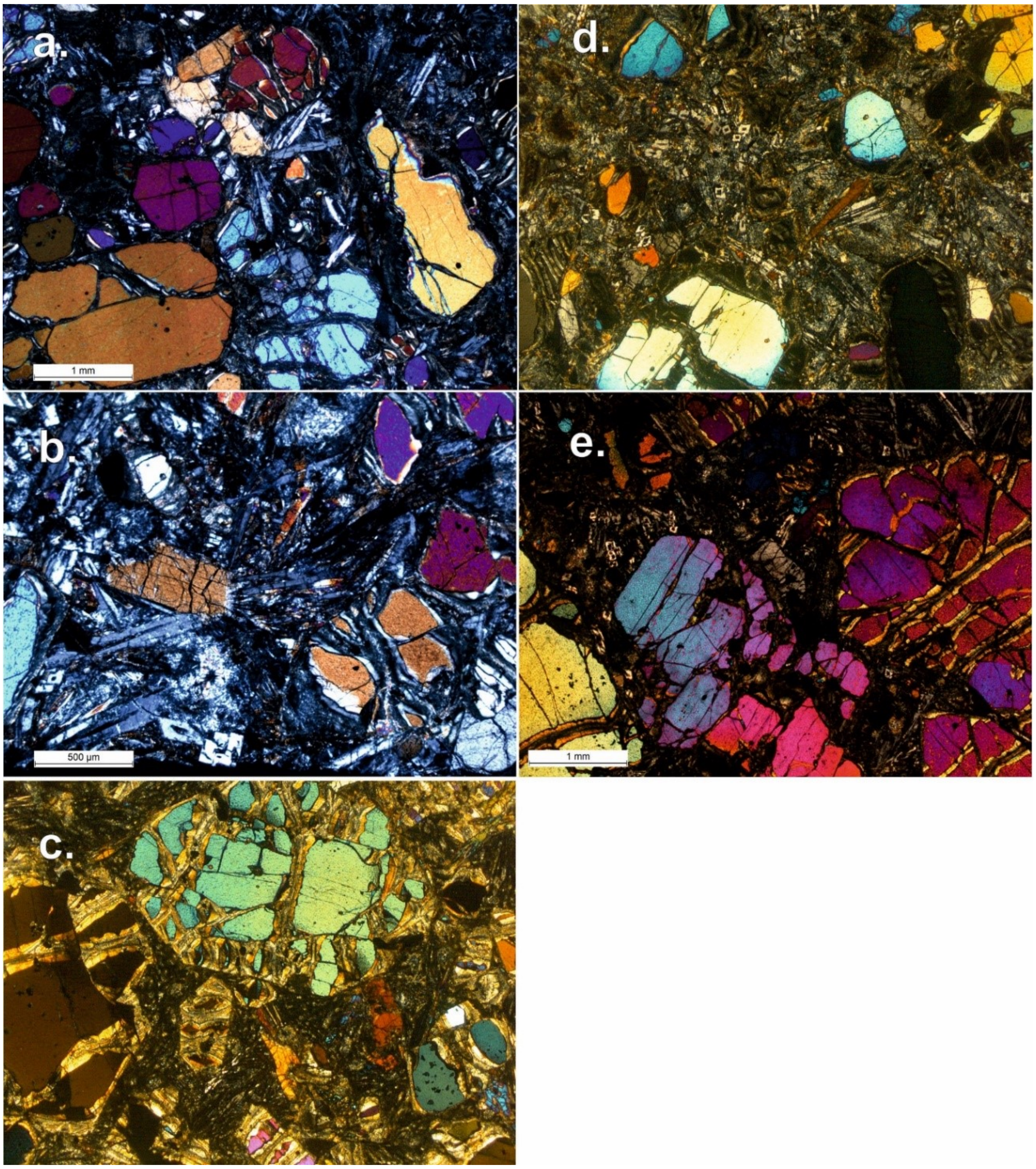

Figure 4 (COLOUR). Photomicrographs of Marki picrite. (a) and (b) are from the same ‘holocrystalline picrite’ unit, after the classification of (Malpas and Langdon, 1984), showing large olivine and minor clinopyroxene phenocrysts, with varying alteration, in a finer-grained basaltic groundmass, variably altered. In places, large plagioclase microlaths are noted in the groundmass; in pockets within individual sections (b) the grain size groundmass increases substantially, and large, radiating plagioclases are apparent. (c) ‘Vitrophyric’ picrite. Extensive serpentinisation of large, euhedral olivines is variable, largely due here to surface weathering. Minor, smaller clinopyroxene and plagioclase is also present. Groundmass here is extensively serpentinised, although still glassy in places. (d) Same section as (c) showing variability in groundmass and phenocryst content within samples. Serpentinised groundmass here contains larger microlaths of plagioclase as well. (e) Other ‘vitrophyric’ picrite sections have phenocryst contents exceeding 50%, with large, euhedral, interlocking olivine.

Phenocryst content of a number of units is close to, or exceeds 45-55%, the range of values over which eruption of a basaltic magma becomes unfeasible (Marsh, 1981), and implies an apparent magmatic viscosity between one and several orders of magnitude higher than a crystal-free magma (Champallier et al., 2008; Okumura et al., 2016). Development of weak columnar jointing and the limited lateral extent of more crystal-rich units suggest that they also did not flow significantly once erupted. Many units contain clusters of mm-sized, interlocking crystals of olivine, which would have inhibited flow, consistent with surface flow of a few meters at most. Flow of less viscous, lower crystal content lavas was a few 10s of meters. Therefore, picritic lavas were able to build a steep-sided volcanic edifice, in marked contrast to the extensive, relatively flat terrain of UPL pillows on which this, and other similar bodies, rest. The observation that all picrite bodies rest on the uppermost extrusive sequence is consistent with the inference of an extrusive origin, and implies that picrite was erupted as the final stage of magmatism. Marked differences in unit dip between picrite and UPL confirm that Picrite Hill formed as a steep-sided edifice, rather than having a current morphology imposed by secondary processes such as erosion. The association of pictritic bodies with extensional faults provides an obvious mechanism for explaining how such crystal-rich magmas were erupted. Vertical extent of these faults, and degree of fault movement, are difficult to ascertain. However, throughout of the main body of the Troodos there is extensive evidence for propagation of extensional faults deep into the crustal sequence, with larger, detachment faults extending through extrusives and sheeted dyke units to the boundary of massive gabbros, i.e. extending to depths of deep crustal magma chambers (Dilek and Furnes, 2009). In the Troodos ophiolite there is also evidence for periods of crustal thinning through extension, during magmatically quiescent periods, resulting in detachment faulting and block movement throughout the ophiolitic sequence (Robertson and Xenophontos, 1997). The fault adjacent to Picrite Hill extends at least several hundreds of meters laterally, and is part of a series of approximately N-S trending faults on the northern flank of the Troodos which extend many km. In the Marki area, extensional faults are associated with locally thick deposits of volcaniclastic sediment, accumulated in half-grabens. A type example of the relationship of umbers with UPL pillows, which implies on-axis, fault-controlled hydrothermal discharge (Constantinou and Govett, 1973), is found 100m W of Picrite Hill. As such, the fault with which Picrite Hill is associated likely propagated deep enough into the crust to tap parts of the magmatic system with which UPL units were derived. Extension on this fault would then have allowed viscous crystal mush within this magmatic system, accumulated during prolonged basaltic magmatism, to extrude onto the sea floor. Strong N-S alignment of faults within the Marki region means that many of the picritic bodies are lens shaped and elongated N-S. However, Picrite Hill, associated with one of the most extensive fault systems is, aside from elongation in one direction representing preferred flow, an approximately circular body with a clear, central complex. Absence of any basaltic units within the body suggests that this extrusion occurred after normal magmatism had ceased, although adjacent volcaniclastic sediment and an absence of any sedimentation implies that the process occurred shortly after formation of the main body of UPL within this area. Picrite Hill consists of multiple units, implying that formation of the body was episodic, possibly related to pulses of movement on the main feeder fault. Repeated opening of this fault during the final stages of crustal extension of the Troodos, and a large difference in pressure from the surface to the tapped magmatic system, would have allowed crystal mush to flow and extrude onto the seafloor. The highly non-Newtonian nature of this viscous crystal mush explains why it could be extruded onto the seafloor from deep within the crust due to a large pressure differential, but would then only flow limited distances once on the surface (e.g. Champallier et al., 2008; Okumura et al., 2016). Coarser groundmass within some units is consistent with eruptions of crystal mush into, or associated with, a hot lava pile, and relatively rapid formation of the body.

Lower parts of the body might have formed, later on, by injection of mush into the bottom of a lava pile.

4. Extrusion of crystal mush on Venus

An interesting observation of the Marki region is the juxtaposition of two contrasting volcanic landforms: pillowed and sheet-like basaltic lava flows of the UPL overlain by steeper picritic volcanic domes and lenses. All observations imply that picritic bodies and basaltic terrains share a common origin, with picrite formed by extrusion of crystal mush from the same magmatic system in which UPL basalts evolved. This observation is consistent with the variable modal, forsteritic olivine content of UPL basalts (Gass, 1958). The high olivine content of picrites means that they are classified as ultramafic bodies, in contrast to basaltic (mafic) pillows, although it is solely the extreme difference in crystal content, and possibly a small difference in volatile content, which explains the marked contrast in eruption style. Considerable extensional faulting within the Troodos must have been required to extrude this picritic crystal mush, which would otherwise have remained trapped within the crust. The presence of unzoned, highly forsteritic olivine with only minor clinopyroxene implies that this mush likely formed in a relatively high temperature, and therefore relatively deep, magmatic system.

A similar, although much larger scale process of crystal mush extrusion could be invoked to explain the formation of steep-sided volcanic domes on Venus. In the proposed model, small degrees of fractionation of mantle-derived basaltic melt takes place within extensive magmatic systems in the Venusian crust. Over prolonged periods of time, periodic eruption of basaltic lava and reinjection of primitive magma into the magmatic system results in substantial accumulation of crystal mush, perhaps analogous to extensive ultramafic cumulate sequences typical of terrestrial ophiolitic sequences. During magmatically quiescent periods there is no new input of magma into the crust, and effusive volcanism ceases. Extensional faulting provides a mechanism for allowing accumulated crystal mush within this magmatic system to extrude onto the surface. The high crystal content of this mush results in a markedly different volcanic landform, steep-sided domes, in contrast to underlying basaltic terrains. At a later stage, renewed magmatic input would result in a return to effusive type basaltic eruptions.

This model can account for a number of observed characteristics of steep-sided volcanic domes. (1) Firstly, it readily explains the intimate association of steep-sided domes with basaltic terrains, especially extensive lava flows and coronae, on Venus. Crystal mush would be formed at depth during the prolonged, high temperature, low-degree fractionation of basaltic magma inferred for large-scale Venusian volcanism. This mush is then parental to the volcanic domes. The same magmatic process explains formation of various volcanic terrains and volcanic domes, explaining their close association without needing to invoke a contrasting magmatic processes. (2) Extrusion of crystal mush can only occur during magmatically quiet periods. This is consistent with observed clustering of steep-sided domes at certain stratigraphic heights within volcanic terrains (Ivanov and Head, 1999). However, unlike other proposed mechanisms, dome formation by mush extrusion could occur at any point during more widespread basaltic volcanism, explaining why domes are sometimes partly embayed or covered by flows following renewed basaltic magmatism. (3) At the same time, high crystal contents of mush explain why steeper domes are supported on the high temperature, high-pressure Venusian surface, and why there is a marked contrast in morphology between domes and lava flows. (4) The process of mush extrusion is unusual, and dependent on both cessation in normal magmatism and extensional faulting, which would explain why domes have unique characteristics and why there is not, instead, a gradual transition from lava flows or shield terrains to steep-sided domes. (5) However, extrusion of mafic to ultramafic mush would explain the flatter, more rounded-top morphology of volcanic domes on Venus, in contrast to the blockier texture expected with eruption of more rhyolitic lava. Prolonged eruptions of mush through the same fault system would, presumably, eventually mainly occur through injection into the base of volcanic piles, consistent with many observations. However, piles could also accumulate through repeated eruptions and build-up of lava flows. (6) As eruptions of mush are fault-controlled this model also explains why steep-sided domes on Venus are not ubiquitous, but instead, rather unusual features. (7) Extrusion of mush would explain why Radar properties of the domes are similar to those of surrounding lava fields (Ford, 1994; Stofan et al., 2000), and finally, (8) extrusion of crystal mush would also be consistent with a single phase of eruption, or pulsed/episodic emplacement of domes that has been inferred by some authors based on observed aspect ratios (Fink et al., 1993; Pavri et al., 1992).

However, it is important to note that picritic extrusions in Cyprus only indicate a possible mechanism for formation of volcanic domes on Venus, rather than providing a terrestrial analogue. There are clear differences between picritic bodies at Margi and Venusian volcanic domes in terms of (1) scale of eruptions and magma volumes, (2) the external shape of bodies (circular vs elongate), (3) the tectonic regime in which volcanic bodies occur, and (4) the contrasting extent of weathering and consequent modification in dome morphology on Earth. Landforms around Marki are small bodies less than 100 meters in diameter. Most bodies are also elongate and not circular. In terms of dimensions, therefore, Venusian steep-sided domes are 3 to 4 orders of magnitude larger than any feature noted in the Marki region. However, the scale of volcanic activity on Venus associated with shield and lava flow terrains is similarly orders of magnitude larger than that associated with oceanic crust formation in the Troodos massif. Importantly, the relative extent of dome-related volcanism on Venus is not inconsistent with the vast scale of magmatism recorded by extensive lava flows and shield terrains. This is demonstrated by using a simple mass balance calculation to determine the volume of basaltic magma required to produce Picrite Hill. The volume of basaltic magma required to fractionate the volume of Fo92 crystals present in a 25m high hemispherical dome with a crystal fraction of 0.5, corresponds to an erupted basaltic lava flow 1m thick and roughly 1 km in diameter. Scaling this to a 20km steep-sided dome implies an erupted basaltic lava flow of 10m thick and 700 km diameter. Given the large uncertainties with this type of calculation, the scale of dome size to extent of basaltic volcanism for both Marki and Venus are, to a first approximation, reasonable.

A similar scale issue occurs when attempting to find other terrestrial analogues for Venusian volcanic domes. Pavri et al. (1992) noted that Venusian domes are typically 2-3 orders of magnitude larger than terrestrial rhyolitic domes, although also noted that association of domes with coronae, and the probability that magmatic reservoirs on Venus would, due to the absence of plate motion, grow to unusually large sizes, could be used to invoke a model for dome formation from rhyolitic magma following fractionation of basaltic parental melt. Similarly, association of domes with extensive magmatic terrains and large-scale fractionation of basalt can also be used, here, to invoke volumetric extrusion of crystal mush. The contrasting nature of plate tectonic vs stagnant lid regimes on Earth and Venus, in tandem with strongly contrasting surface conditions, suggest that the search for true terrestrial analogues for all volcanic landforms is of limited value. However, it is also possible that smaller mush-type eruptions occur on the Venusian surface. Hansen (2005) used Magellan data to conducted detailed geological mapping on Venus. This work identified numerous small shield and shield-like features distinct from extensive lowland lava flows in which they are found. Hansen (2005) considered the possibility that these features represent some type of point-source partial re-melting of the highland crust. They could, alternatively, represent smaller scale extrusion of basaltic crystal mush.

Most picritic bodies near Marki are elongate. However, most elongate bodies are small, some likely representing single eruptions (Gass, 1958). The much more substantial Picrite Hill is an approximately circular body, with an elongate tongue formed by limited preferential flow of less viscous lavas in one direction. Extrusion of material onto a gently-sloping seafloor, consistent with the low-angle contact between picrite and UPL, explains the observed shape of the body. Elsewhere in this area, a highly-faulted and tilted palaeo-seafloor consisting of a series of half-graben accounts for the shape of picritic bodies. Large extensional faults in the Marki region have a strong N-S trend, consistent with the regional spreading direction of the main Troodos body (Robertson and Xenophontos, 1997). It is generally agreed that the Troodos Ophiolite formed in a Supra-Subduction Zone environment, with extension due to subduction initiation and/or back-arc spreading (Pearce and Robinson, 2010). The nature of faulting within the Troodos body, and at Marki specifically, has controlled the morphology of picritic extrusions to a significant degree. Differences in local and planet-wide magmatic-tectonic regime on Venus (Ivanov, 2015; Ivanov and Head, 2011), surface conditions, scale of eruption, and mechanical response of the crust and dyke propagation (Foster and Nimmo, 1996; Karato and Barbot, 2018; Mikhail and Heap, 2017) could easily result in fundamental differences in morphology of volcanic landforms.

Extrusion of crystal mush at Marki is facilitated by extensional faulting related to seafloor spreading after normal magmatic processes ended. The absence of a comparable mechanism on Venus would, therefore, necessitate some alternative process to induce fracturing and faulting of the Venusian crust. Extensional faulting has been inferred for other parts of Venus. Rift-like valleys within the Aphrodite Terra can be traced for 1000s km in some instances, and elsewhere, extensional tectonism has produced marked belts of deformation, persisting over 100s of km (Ivanov and Head, 2011). However, although a correlation between rifting and coronae, and large-scale volcanism generally, has been noted, volcanic domes on Venus have no discernible relationship with such this type of large-scale rifting (Airey et al., 2017). Large-scale rifting on Venus is often assumed to result from crustal doming associated with upwelling plumes (Ivanov and Head, 2015), which explains a correlation with extensive, i.e. basaltic, volcanism. As such, volcanic domes formed by extrusion of crystal mush might be expected to post-date such rifting, or be associated with a smaller scale extensional process. Instead, correlation of domes with coronae and other similar magmatic landforms, and associated with extensive basaltic terrains, suggests that domes could be associated with smaller scale faulting to shallow magmatic systems. Significant differences in crustal strength between Venus and Earth (e.g. Karato and Barbot, 2018; Mikhail and Heap, 2017) should result in a significant deflection of the brittle-ductile transition to much shallower levels in the hotter, Venusian crust. As such, there will be fundamental differences in emplacement of bodies of magma, supporting the presence of magma chambers at comparatively higher levels in the Venusian crust. For example, Mikhail and Heap (2017) invoked a model where large bodies of magma associated with coronae are emplaced at high levels in the Venusian crust via diapirism, due to the inhibition of fracture propagation and dyke emplacement. Magma chambers feeding the volcanic plains with which steep-sided domes are associated could, therefore, be at comparatively shallow depths in the hotter, more ductile Venusian crust, meaning that extensional faults required to tap them would be smaller scale features.

Whether steep-sided domes are associated with smaller-scale extensional faulting is not clear. Domes often have surfaces cut by radial and/or concentric fractures, although these can readily be explained by fracturing of cool, solid rinds (Ivanov, 2015; Pavri et al., 1992). However, other observations of steep-sided volcanic domes are consistent with a fault controlled extrusion process. The notable dome cluster near Alpha Regio, for example, consists of 7 partially overlapping domes with a strong E-W trend, associated with a series of very obvious graben which strike NE across the associated volcanic plain (Pavri et al., 1992). Elsewhere, domes appear to converge on radial fracture patterns in surrounding terrains. Pavri et al. (1992) suggested that such relationships, and lineations in domes, dome clustering, evidence for interconnecting grabens and other surface lineations, are all evidence for shared feeder dykes. These features can also be explained by a fault-related extrusion model. Unfortunately, resolution in data means that it is challenging to accurately determine and infer relationships between domes and surrounding terrains. A fault controlled process of dome formation is, however, feasible.

Another significant difference between terrestrial and Venusian domes is surface conditions and the effects of weathering. Picrites near Marki were erupted onto a shallow Cretaceous seafloor, and at a pressure not dissimilar from that of the Venusian surface. However, temperature of the surface would have been significantly lower, and the palaeoseafloor preserved at Marki was first affected by seafloor processes, and has, since emergence, been shaped by active weathering processes. Bodies like Picrite Hill would originally have been concealed beneath sedimentary cover consisting of radiolarian mudstone, which is found adjacent to Picrite Hill (Fig.1), and deep-sea pelagic carbonates ranging from Maastrictian to Oligocene in age (Lefkara formation) (Robertson 1997). Extensional faults in the Marki region, as well as facilitating eruption of crystal mush, also produced seafloor depressions which resulted in rapid accumulation of sediments, including, locally, thick umber deposits. However, background sediments are not found within Picrite Hill, suggesting that the body was erupted relatively soon after cessation of seafloor spreading and normal volcanism, and that extrusion, although pulsed, was rapid. Extensional faults are covered by sediments immediately to the north of Marki suggesting that they did not remain active for long after the end of magmatic activity. It is likely, therefore, that the palaeoseafloor at Marki was rapidly covered in sediment.

Field relationships demonstrate that although eruption of mush at Picrite Hill was facilitated by extensional faulting, this faulting had little influence on dome morphology. The extent to which the morphology of Picrite Hill has been modified by more recent weathering and erosion remains unclear. Although there is a clear difference in colour between UPL units and picrite, hardness and resistance to weathering appear to be similar. The Marki region is relatively arid and there are, currently, low rates of removal of material from the area adjacent to Picrite Hill, and low degrees of vegetation. Abrupt changes in dip between UPL and picrite units are consistent with formation of Picrite Hill as a steep-sided lava pile. The volume of material removed from this body is not clear. However, patterns in veining of UPL units immediately below and adjacent to Picrite Hill indicate that the lateral extent of the body was probably not significantly greater than presently outcropped. As such, the body appears to be well preserved, and clearly formed as a steep sided volcanic dome. Whilst the core of the body is more massive, thinner units on the sides and towards the top of Picrite Hill also suggest that crystal-rich lavas formed flows with relatively smooth, flat-tops. Spheroidal weathering gives parts of Picrite Hill a blocky appearance. However, examination of individual units suggests that erupted mush was closer in appearance to basaltic lava flows, rather than to rhyolitic flows.

The model for dome formation presented here should be readily testable in future, as it (1) implies that the domes are mafic to ultramafic rather than silica-rich, and (2) invokes an extensional-fault controlled mechanism for extrusion. The model implies that lavas forming volcanic domes are crystal-rich, and that non-hydrostatic stress provides a mechanism for mobilising crystal-rich mushes and facilitating eruption. The effects of crystal content, crystal size and crystal shape on magma viscosity remain poorly constrained. However, crystal-rich magmas are expected to be highly non-Newtonian, and the effect of increasing crystal content will be to increase viscosity by up to orders of magnitude (Champallier et al., 2008; Okumura et al., 2016). Pavri et al. (1992) used various modelling methods to constrain the effect of composition of dome morphology, and concluded that observed dome shapes and fracture patterns are consistent with silicic, high viscosity lava. However, as noted by Pavri et al. (1992), no single numerical approach produces a consistent model for dome formation, and the starting assumption for many models, i.e. Newtonian magma rheology, may be incorrect. Stofan et al. (2000) used an alternative approach to constrain dome composition, based on thermal models of crust development on erupted lava. They concluded that observed fracture patterns are consistent with fluid, lower viscosity interiors, at least for some of the domes observed in Magellan data. Petrographic examination and late-stage veining both demonstrate that erupted mush at Marki contained basaltic liquid. As such, extruded mush was a high temperature fluid, more comparable to basaltic lava. Superficially at least, such a mush would not be expected behave like a high viscosity rhyolitic liquid, and would not form a thick crust and blocky surface texture inconsistent with observed Venusian domes (Stofan et al., 2000). This is supported by observations at Marki. None of the units observed in the lava dome shows evidence for formation of thick crusts. Furthermore, in the adjacent area there are examples of pillowed picritic basalts. Although these have a lower olivine crystal content, their appearance does suggest that olivine-rich mushes behave more like high viscosity basalts than like rhyolites.

Stofan et al. (2000) also suggested that surface depressions on some Venusian domes indicate a fluid dome interior. However, these features are not inconsistent with the proposed model of mush extrusion. As noted previously, parts of the core of Picrite Hill may have formed by injection of mush directly into the base of the lava pile. There is also a clear variation in lava viscosity between units, with limited flow of less crystal-rich units. It is, therefore, conceivable that ongoing eruption might produce lava domes on Venus with lower viscosity interiors. Importantly, there are considerable variations in dome morphology on Venus (e.g. Pavri et al. 1992), from flat-topped or shield-like domes, to domes with heavily fractures surfaces, domes with tiered shapes, complex shapes and domes with significant surface depressions. This complexity in morphology is not inconsistent with variations, both spatial and temporal, in magma viscosity as a function of crystal content, as observed at Marki.

Finally, the model presented here does not require an additional magmatic process such as re-melting of parts of the Venusian crust. This is consistent with a simpler model for magmatism on Venus and similar stagnant-lid regime planets, where volcanism arises due to plume-related mantle melting, and where basaltic volcanism dominates. Uniquely, this model is able to account for marked differences in dome morphology compared to surrounding terrains, and also explains the intimate association of domes with basaltic lava plains. If correct, this crystal mush model may be applicable to other stagnant-lid bodies. The appearance of steep-sided domes on Venus might imply a unique, large-scale extensional mechanism which promotes extrusion of crystal mush. Alternatively, surface conditions on Venus might simply favour formation, preservation and ready identification of volcanic features formed by mush extrusion. This raises the possibility that mush extrusion occurred on other bodies such as Mars, Mercury and the Moon during magmatically quiet periods, resulting in formation of distinct volcanic landforms.

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References

Anderson, S.W., Stofan, E.R., Plaut, J.J., and Crown, D.A., 1998, Block size distributions on silicic lava flow surfaces: Implications for emplacement conditions, Geol. Soc. Am. Bull., 110, 1258-1267.

Airey, M.W., Mather, T.A., Pyle, D.M., Ghail, R.C., 2017. The distribution of volcanism in the Beta-Atla-Themis region of Venus: Its relationship to rifting and implications for global tectonic regimes. J. Geophys. Res. 122, 1626–1649. https://doi.org/10.1002/2016JE005205

Basilevsky, A.T., 1997. Venera 8 landing site geology revisited. J. Geophys. Res. 102, 9257–9262. https://doi.org/10.1029/97JE00413

Bridges, N.T., 1997. Ambient effects on basalt and rhyolite lavas under Venusian, subaerial, and subaqueous conditions. J. Geophys. Res. 102, 9243–9255. https://doi.org/10.1029/97JE00390

Bridges, N.T., 1995. Submarine Analogs to Venusian Pancake Domes. Geophys. Res. Lett. 22, 2781–2784. https://doi.org/10.1029/95GL02662

Champallier, R., Bystricky, M., Arbaret, L., 2008. Experimental investigation of magma rheology at 300 MPa: From pure hydrous melt to 76 vol.% of crystals. Earth Planet. Sci. Lett. 267, 571–583. https://doi.org/10.1016/j.epsl.2007.11.065

Constantinou, G., Govett, G.J.S., 1973. Geology, Geochemistry, and Genesis of Cyprus Sulfide Deposits. Econ. Geol. 68, 843–858. https://doi.org/10.2113/gsecongeo.68.6.843

Crumpler, L.S., Aubele, J.C., Senske, D.A., Keddie, S.T., Magee, K.P. and Head, J.W., 1993. Volcanoes and centers of volcanism on Venus. In Venus II: Geology, Geophysics, Atmosphere and Solar Wind Environment, eds Bougher S.W., Hunten D.M., Phillips, R.J. University of Arizonia Press. ISNB 0-8165-1830-0.

Dilek, Y., Furnes, H., 2009. Structure and geochemistry of Tethyan ophiolites and their petrogenesis in subduction rollback systems. Lithos 113, 1–20. https://doi.org/10.1016/j.lithos.2009.04.022

Fink, J.H., Bridges, N.T., Grimm, R.E., 1993. Shapes of Venusian Pancake Domes Imply Episodic Emplacement and Silicic Composition. Geophys. Res. Lett. 20, 261–264. https://doi.org/10.1029/92GL03010

Ford, P.G., 1994. Radar Scattering Properties of Steep-sided Domes on Venus. Icarus 112, 204–218. https://doi.org/10.1006/icar.1994.1178

Foster, A., Nimmo, F., 1996. Comparisons between the rift systems of East Africa, Earth and Beta Regio, Venus. Earth Planet. Sci. Lett. 143, 183–195. https://doi.org/10.1016/0012-821X(96)00146-X

Gass, I.G., 1958. Ultrabasic Pillow Lavas from Cyprus. Geol. Mag. 95, 241–251.

Gass, I.G., 1968. Is Troodos Massif of Cyprus a Fragment of Mesozoic Ocean Floor. Nature 220, 39-. https://doi.org/10.1038/220039a0

Gregg, T.K.P., Fink, J.H., 1996. Quantification of extraterrestrial lava flow effusion rates through laboratory simulations. J. Geophys. Res. 101, 16891–16900. https://doi.org/10.1029/96JE01254

Guest, J.E., Bulmer, M.H., Aubele, J., Beratan, K., Greeley, R., Head, J.W., Michaels, G., Weitz, C., Wiles, C., 1992. Small volcanic edifices and volcanism in the plains of venus. J. Geophys. Res. 97, 15949–15966. https://doi.org/10.1029/92JE01438

Hansen, V. L. (2005). Venus's shield terrain. Bulletin of the Geological Society of America, 117(5-6), 808-822. https://doi.org/10.1130/B256060.1

Ivanov, M.A., 2015. Volcanic complexes on Venus: Distribution, age, mechanisms of origin, and evolution. Petrology 23, 127–149. https://doi.org/10.1134/S0869591115020046

Ivanov, M.A., Head, J.W., 2015. The history of tectonism on Venus: A stratigraphic analysis. Planet. Space Sci. 113, 10–32. https://doi.org/10.1016/j.pss.2015.03.016

Ivanov, M.A., Head, J.W., 2011. Global geological map of Venus. Planet. Space Sci. 59, 1559–1600. https://doi.org/10.1016/j.pss.2011.07.008

Ivanov, M.A., Head, J.W., 1999. Stratigraphic and geographic distribution of steep-sided domes on Venus: Preliminary results from regional geological mapping and implications for their origin. J. Geophys. Res. 104, 18907–18924. https://doi.org/10.1029/1999JE001039

Karato, S., Barbot, S., 2018. Dynamics of fault motion and the origin of contrasting tectonic style between Earth and Venus. Sci. Rep. 8. https://doi.org/10.1038/s41598-018-30174-6

Kaula, W.M., 1990. Venus - A Contrast in Evolution to Earth. Science (80). 247, 1191–1196. https://doi.org/10.1126/science.247.4947.1191

Malpas, J., Langdon, G., 1984. Petrology of the Upper Pillow Lava suite, Troodos ophiolite, Cyprus, in: Gass, I.G., Lippard, S.J., Shelton, A.W. (Eds.), Ophiolites and Oceanic Lithosphere. Geological Society, London, Special Publication No. 13, pp. 155–167.

Marsh, B.D., 1981. On the CRystallinity, Probability of Occurrence, and Rheology of Lava and Magma. Contrib. Mineral. Petrol. 78, 85–98. https://doi.org/10.1007/BF00371146

McKenzie, D., Ford, P.G., Liu, F., Pettengill, G.H., 1992. Pancake-like Domes on Venus. J. Geophys. Res. 97, 15967–15976. https://doi.org/10.1029/92JE01349

Mikhail, S., Heap, M.J., 2017. Hot climate inhibits volcanism on Venus: Constraints from rock deformation experiments and argon isotope geochemistry. Phys. Earth Planet. Inter. 268, 18–34. https://doi.org/10.1016/j.pepi.2017.05.007

Mukasa, S.B., Ludden, J.N., 1987. Uranium-lead isotopic ages of plagiogranites from the troodos ophiolite, cyprus, and their tectonic significance. Geology 15, 825–828. https://doi.org/10.1130/0091-7613(1987)15<825:UIAOPF>2.0.CO;2

Okumura, S., Kushnir, A.R.L., Martel, C., Champallier, R., Thibault, Q., Takeuchi, S., 2016. Rheology of crystal-bearing natural magmas: Torsional deformation experiments at 800 degrees C and 100 MPa. J. Volcanol. Geotherm. Res. 328, 237–246. https://doi.org/10.1016/j.jvolgeores.2016.11.009

Pavri, B., Head, J.W., Klose, K.B., Wilson, L., 1992. Steep-sided domes on venus - characteristics, geologic setting, and eruption conditions from magellan data. J. Geophys. Res. 97, 13445–13478. https://doi.org/10.1029/92JE01162

Pearce, J.A., Robinson, P.T., 2010. The Troodos ophiolitic complex probably formed in a subduction initiation, slab edge setting. Gondwana Res. 18, 60–81. https://doi.org/10.1016/j.gr.2009.12.003

Plaut, J.J., Anderson, S.W., Crown, D.A., Stofan, E.R., van Zyl, J.J., 2004. The unique radar properties of silicic lava domes. The unique radar properties of silicic lava domes. J. Geophys. Res. 109. https://doi.org/10.1029/2002JE002017

Quick, L.C., Glaze, L.S., Baloga, S.M., Stofan, E.R., 2016. New approaches to inferences for steep-sided domes on Venus. J. Volcanol. Geotherm. Res. 319, 93–105. https://doi.org/10.1016/j.jvolgeores.2016.02.028

Robertson, A.H., Hudson, J.D., 1973. Cyprus umbers - chemical precipitates on a tethyan ocean ridge. earth Planet. Sci. Lett. 18, 93–101. https://doi.org/10.1016/0012-821X(73)90039-3

Robertson, A., 2004. Development of concepts concerning the genesis and emplacement of Tethyan ophiolites in the Eastern Mediterranean and Oman regions. Earth-Science Rev. 66, 331–387. https://doi.org/10.1016/j.earscirev.2004.01.005

Robertson, A.H.F., Xenophontos, C., 1997. Cyprus, in: Encyclopedia of European and Asian Regional Geology. Springer Netherlands, Dordrecht, pp. 160–171. https://doi.org/10.1007/1-4020-4495-X_23

Sakimoto, S.E.H., Zuber, M.T., 1995. The spreading of variable-viscosity axisymmetrical radial gravity currents - applications to the emplacement of venusian pancake domes. J. Fluid Mech. 301, 65–77. https://doi.org/10.1017/S0022112095003806

Saunders, R.S., Arvidson, R.E., Head, J.W., Schaber, G.G., Stofan, E.R., Solomon, S.C., 1991. An overview of Venus geology. Science 252, 249–252. https://doi.org/10.1126/science.252.5003.249

Searle, D.L., Vokes, F.M., 1969. Layered ultrabasic lavas from cyprus. Geol. Mag. 106, 515-. https://doi.org/10.1017/S001675680005929X

Shellnutt, J.G., 2018. Derivation of intermediate to silicic magma from the basalt analyzed at the Vega 2 landing site, Venus. PLoS One 13.

https://doi.org/10.1371/journal.pone.0194155

Shellnutt, J.G., 2013. Petrological modeling of basaltic rocks from Venus: A case for the presence of silicic rocks. J. Geophys. Res. 118, 1350–1364. https://doi.org/10.1002/jgre.20094

Smith, D.K., 1996. Comparison of the shapes and sizes of seafloor volcanoes on Earth and ``pancake'' domes on Venus. J. Volcanol. Geotherm. Res. 73, 47–64. https://doi.org/10.1016/0377-0273(96)00007-8

Solomatov, V.S., Moresi, L.N., 1996. Stagnant lid convection on Venus. J. Geophys. Res. 101, 4737–4753. https://doi.org/10.1029/95JE03361

Stofan, E.R., Anderson, S.W., Crown, D.A., Plaut, J.J., 2000. Emplacement and composition of steep-sided domes on Venus. J. Geophys. Res. 105, 26757–26771. https://doi.org/10.1029/1999JE001206

Tanaka, K.L., Senske, D.A., Price, M., Kirk, R.L., 1997. Physiography, geomorphic/geological mapping, and stratigraphy of Venus. In Venus II: Geology, Geophysics, Atmosphere and Solar Wind Environment, eds Bougher S.W., Hunten D.M., Phillips, R.J. University of Arizona Press. ISNB 0-8165-1830-0.

Woelki, D., Regelous, M., Haase, K.M., Romer, R.H.W., Beier, C., 2018. Petrogenesis of boninitic lavas from the Troodos Ophiolite, and comparison with Izu-Bonin-Mariana fore-arc crust. Earth Planet. Sci. Lett. 498, 203–214. https://doi.org/10.1016/j.epsl.2018.06.041