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Tessera terrain ribbon fabrics on Venus reviewed: Could they be dyke swarms?

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ABSTRACT

Endogenic and exogenic models have been developed over the past 25 years to explain the development of extensive, pervasive, high aspect ratio troughs and graben-like structures ("ribbons") and short- to long-wavelength folds in tessera terrains, commonly associated with crustal plateaus on Venus. Many of the models are complex, and appear to be both internally contradictory and inconsistent with well-established geological principles. They are based on what amount to assumptions regarding (i) the secular evolution of crustal thickness, (ii) crustal mechanical homogeneity and isotropy, (iii) the mechanics of ribbon formation, and (iv) the relative timing of ribbons in the structural geological sequence of tessera terrains. On the basis of both theoretical and observational considerations, this contribution will suggest that coupling the formation of tessera ribbons to the emplacement of dyke swarms relatively late in the structural development of tessera terrains may provide a resolution for the apparent inconsistencies in those published models.

1. Introduction

Among the rocky or terrestrial planets of the inner Solar System, the surface of cloud-shrouded Venus is perhaps the least understood geologically. It has only been visited by four radar-based mapping missions: the Pioneer Venus mission in 1978,[1] the Russian Venera 15 and 16 missions in 1983 (resolution of 1–2 km/pixel[2]), and the Magellan mission in 1990–1994 (resolution of 75–100 m/pixel[3]). Some of the mapped features appear unique to Venus, including the structural fabrics of tessera terrains, commonly spatially associated with crustal plateaus (e.g. Ivanov and Head, 1996; Hansen and Willis, 1996, 1998; Hansen, 2006, 2018; Gilmore and Head, 2018, and references cited by these papers). Much geological research over the past 25 years has focused on elucidating the potential genetic relationship between the tectonic development of crustal plateaus and the structural evolution of associated tessera terrains. Limited by the resolution of the Magellan data, planetary scientists have been constrained to developing tentative interpretations, supported by extrapolations from numerical and analogue modelling. The result has been the development and evolution of tectonic and structural paradigms, many of which are complex, and appear to be both internally contradictory and inconsistent with well-established geological principles (see below). In great part, the complexity, contradictions and inconsistencies appear to stem from attempts to integrate the development of close-spaced, pervasive, purportedly tectonic extensional brittle fabrics (high aspect ratio tessera "ribbons") with at least partially contemporaneous, uniform, ductile fold development, all within the context of tectonic life-cycle models of crustal plateau evolution (see Hansen, 2018; Gilmore and Head, 2018 for current summaries).

This paper proposes that the principal stumbling block in understanding the structural geology of tessera terrains is the general acceptance that ribbons are symmetrical graben-like structures, generated by tectonic, far-field, extensional boundary conditions (tensile regional stress). Instead, this contribution will suggest that they may represent generally blind dyke swarms whose orientation was guided by the regional stress field, but whose emplacement and dilation was primarily driven by internal magmatic pressure. If valid, such a hypothesis would significantly reduce the complexities, internal contradictions and general inconsistencies inherent in much of the current narrative of tessera terrain structural geology.

2. Tessera terrain: evolving paradigms

Scientific perspectives regarding the geological history of the surface of Venus have evolved significantly since the first radar images were received from the space probes. Although no new remote surface

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[1] https://www.nasa.gov/mission_pages/pioneer-venus/

[2] https://nssdc.gsfc.nasa.gov/planetary/venera.html

[3] https://www2.jpl.nasa.gov/magellan/

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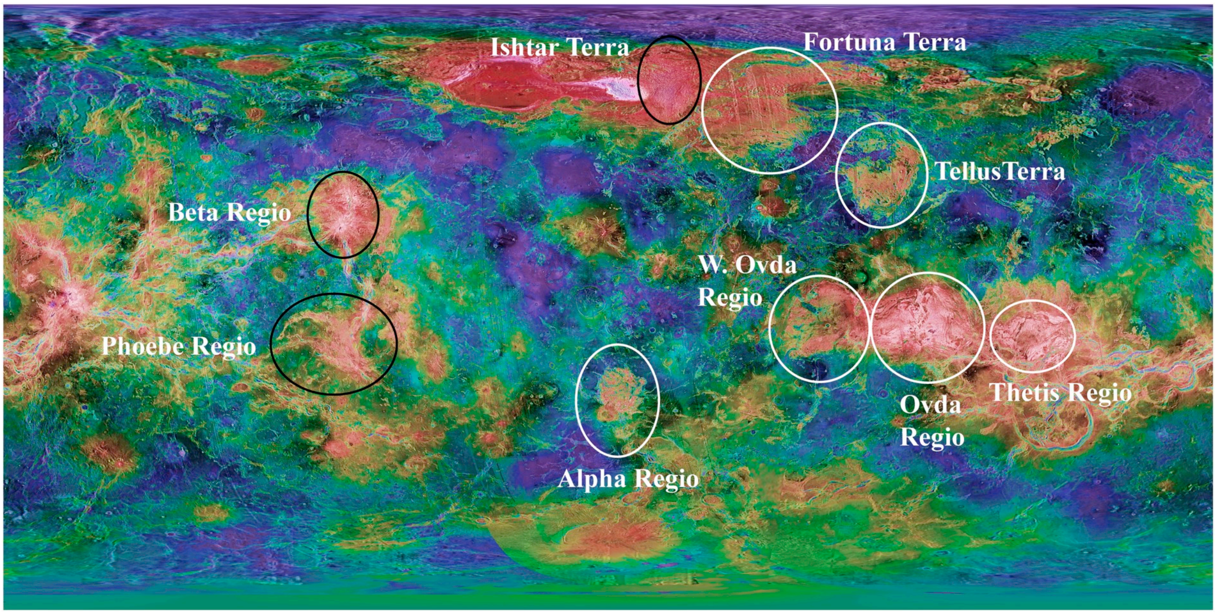

Fig. 1. Generalised planetary distribution of tessera terrains (ellipses) and crustal plateaux. Tessera terrains associated with crustal plateaux are indicated by white ellipses. Adapted from Hansen (2018), and overlain on a base provided by https://astrogeology.usgs.gov/search/map/Venus/Magellan/RadarProperties/Colorized/Venus_Magellan_C3-MDIR_ClrTopo_Global_Mosaic_6600m (Magellan Team: Ford, P., Pettengill, G., Liu, F., Quigley, J. (1993).

data have been obtained from orbit since the Magellan mission, the scientific community has been re-examining observations and derived models for the past 25 years (e.g. see Hansen, 2006, 2018; Gilmore and Head, 2018; and references cited in these papers).

2.1. Venusian global resurfacing

Preliminary global mapping of Venus, with emphasis on impact crater statistics, led to a model of catastrophic (10–100 My), global, volcanic resurfacing of the planet between 300 and 800 My ago (e.g. Bullock et al., 1993; Strom et al., 1994; see also Ivanov and Head, 2013). If valid, such extensive volcanism would provide a planet-wide stratigraphic time-line. According to this model, the surface of Venus is geologically relatively young, with inliers of older features of varying sizes surrounded by younger volcanic plains. Among the older features are crustal plateaus (Figs. 1, 2a and 3): large, steep-sided, flat-topped regions, 1500–2500 km in diameter, standing 1–4 km above the mean planetary radius, associated with thickened crust and relatively shallow isostatic support (e.g. Smrekar and Phillips, 1991; Grimm, 1994; Stofan et al., 1995). They are generally considered to be an integral part of Venus' Ancient Era (e.g. Ivanov and Head, 1996; Hansen et al., 1999, 2000; Ivanov and Head, 2013, 2015a; Hansen and López, 2018; Hansen, 2018; Gilmore and Head, 2018).

More recently, the catastrophic resurfacing hypothesis has been challenged in favour of a non-catastrophic model of progressive equilibrium resurfacing and impact cratering, well encapsulated by the SPITTER formulation (Spatially Isolated Time-Transgressive Equilibrium Resurfacing: Hansen and Young, 2007; see also Hansen and López, 2010, Hansen and López, 2018; Bjonnes et al., 2012). Rather than a specific global event, resurfacing would have occurred at different times in different places. In any event, crustal plateaus are older than adjacent resurfacing, whenever that resurfacing occurred (for a review, see Ivanov and Head, 1996). However, the catastrophic vs equilibrium resurfacing debate is as yet unresolved (e.g. Romeo, 2013; Kreslavski et al., 2015; Ivanov and Head, 2015b). Until it is, the absolute age of venusian surface features remains unconstrained (see Ivanov and Head, 1996).

2.2. Tessera terrain

Tessera terrain on Venus is defined by a distinctive, multi-component structural fabric, commonly spatially correlated with crustal plateaus (Fig. 1) and related lowland inliers (e.g. Bindschadler and Head, 1991; Ivanov and Head, 1996; Hansen and Willis, 1996, 1998). Although Hansen and Willis (1996) defined six different types of tessera terrain, with multiple subtypes of each, subsequent usage of the term refers to sets of extensive, high aspect ratio lineaments that mutually intersect at high angles, associated with localised, effusive volcanic intra-tessera basins (Figs. 2 and 3).

Because they were thought to be potentially critical in discriminating between various tectonic models for crustal plateau formation, the nature and mutual timing relations of these lineaments were the subject of much detailed remote mapping (e.g. Bindschadler and Head, 1991; Ivanov and Head, 1996; Hansen and Willis, 1996, 1998; Hansen et al., 1999, 2000; Ghent and Hansen, 1999; Gilmore et al., 1998; Gilmore and Head, 2000; Ghent and Tibuleac, 2002; Hansen, 2006; Romeo and Turcotte, 2008; Romeo and Capote, 2011; Hansen, 2018; Gilmore and Head, 2018). It is important to note that there is no reason to suppose that the geographically separate tessera terrains are all of the same age (Hansen and Willis, 1996).

One set of lineaments forms ridges, interpreted as fold crests, schematically represented as generally concentric to original crustal plateau margins (Figs. 2 and 3; herein "tessera folds" to distinguish them from other fold sets: e.g. Harris and Bédard, 2014a). Short-wavelength (~3 km) tessera folds are interpreted as minor folds on long-wavelength (20–50 km) folds. The degree of shortening represented by the tessera folds is disputed. Ghent and Hansen (1999) define the folds as "gentle" with open inter-limb angles (~155–175°) and correspondingly shallow limb dips ("warps" according to Hansen, 2018), whereas Romeo and Capote (2011) identify strongly compressed tessera folds associated with thrusts (fold-thrust belts; see also Ivanov and Head, 1996).

The other set comprises uniformly close-spaced (2–6 km), pervasive, shallow (<1 km), narrow troughs, several kilometers wide by hundreds of kilometers long with aspect ratios of 50–100+ (Fig. 2b, c and d), commonly oriented at a high angle to large segments of crustal plateau margins (Figs. 2a and 3). These are bounded by symmetrical, opposite-

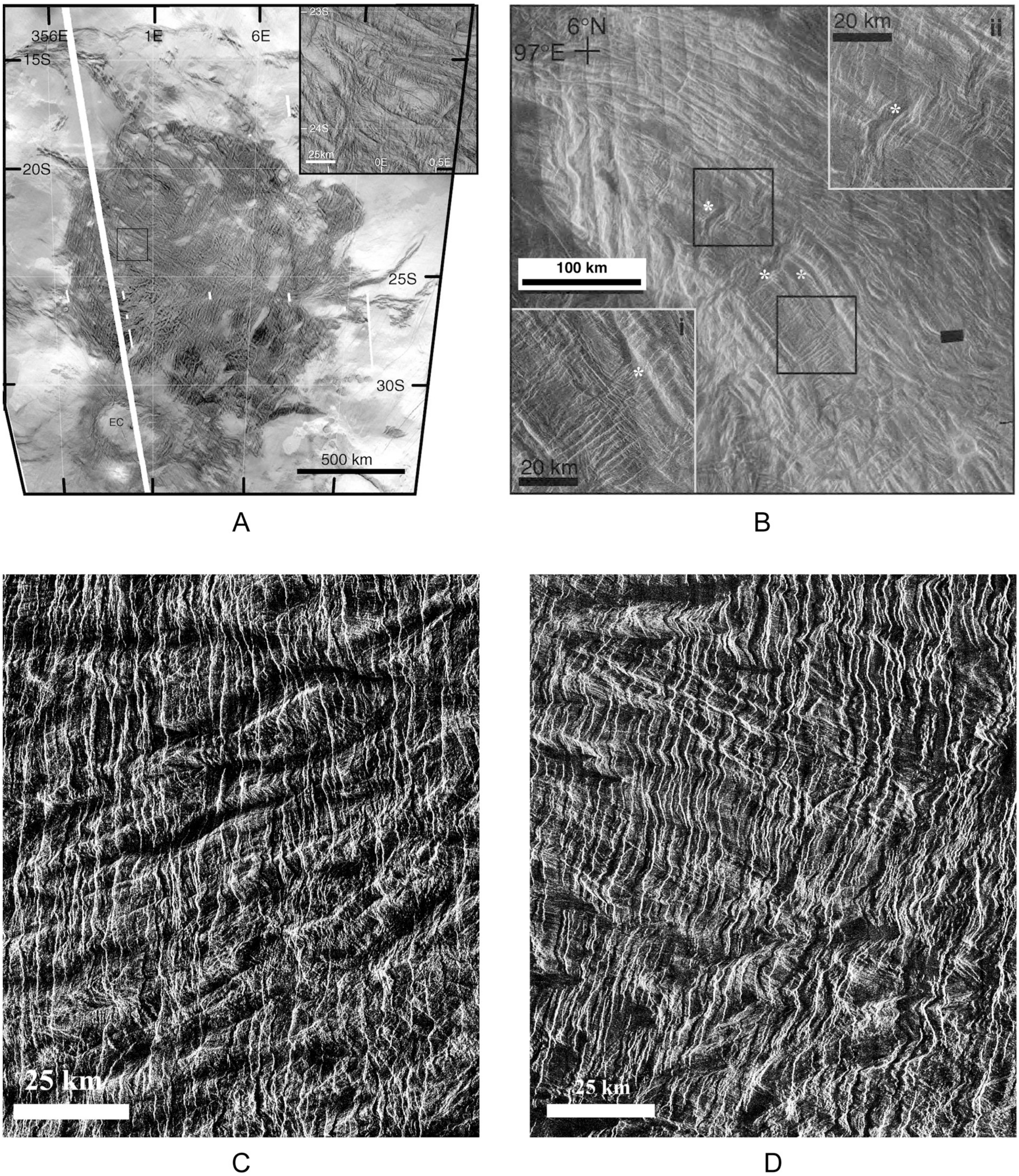

Fig. 2. Radar images of tessera fabrics on Venus. (A) Inverted synthetic aperture radar image of Alpha Regio, a crustal plateau (dark) adjacent to lowlands (bright). Periodically spaced parallel troughs (ribbons) generally trend north-northeast, fold ridges trend northwest to west-northwest. Inset shows detail of periodically spaced parallel ridges and troughs trending north and fold ridges trending west-northwest. The circular feature at the southwest margin of the plateau is younger (Eve Corona, EC). White bold lines are data gaps. Taken from Hansen (2007a, 2007b). (B) Synthetic aperture radar image of tessera terrain in northeastern Ovda Regio (note change of scale compared to A). Periodically spaced parallel troughs (ribbons) trend northeast, fold ridges trend northwest. Details of ribbon fabric (inset i) vs "complex" grabens (inset ii) are discussed in the text. Smooth, dark grey elongate troughs (white asterisks) oriented sub-parallel (inset i) and perpendicular (inset ii), as well as between insets i and ii in the main frame) are interpreted by authors as intra-tessera basins flooded by effusive lava, presumably mafic to ultramafic (see text). Note that ribbon fabric elements appear to both cut across and be truncated by the lavas (inset i). Taken from Phillips and Hansen (1998); see also Hansen et al., 1999; Ghent and Hansen, 1999; Ghent et al., 2005; Hansen (2007c); Hansen and López, 2010). C and D are detailed synthetic aperture radar images of tessera fabric: C is taken from Hansen and Willis (1996, fig. 4a); Hansen and Willis (1998, fig. 1b and 2); Phillips and Hansen (1998, fig. 2) and Hansen et al. (2000, fig 3a: D is taken from their fig. 3b). According to these authors, C illustrates tensile fracture ribbons from Fortuna Tessera, while D shows shear fracture ribbons from Thetis Regio (see Figs. 1 and 5). For further images of tessera fabrics the interested reader is referred to Ivanov and Head (1996, fig. 19); Hansen and Willis (1998, fig. 11); Gilmore et al. (1998, fig. 2a); Ghent and Hansen (1999, fig. 2); Ghent and Tibuleac (2002, fig. 1); Hansen (2006, fig. 3); Hansen (2007c, fig. 4); Hansen (2007a, fig. 8.1-6); Romeo and Capote (2011, figs 5 and 7); Gilmore and Head (2018, figs 6b and 7c).

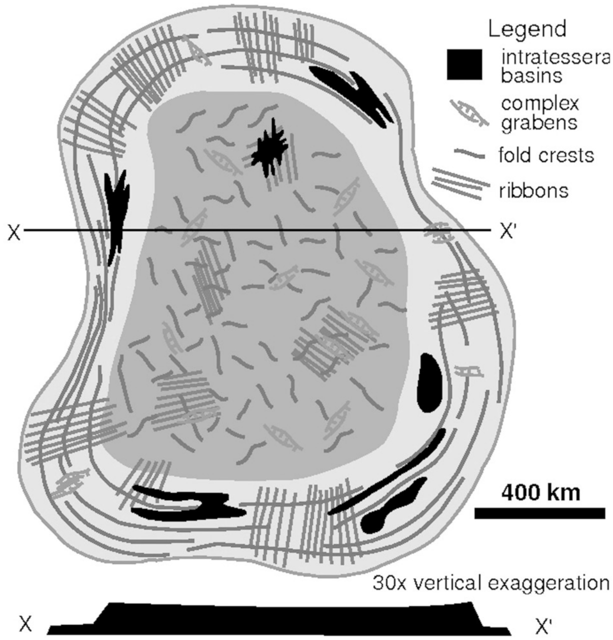

Fig. 3. Schematic, idealised reconstruction of crustal plateau structure to illustrate a marginal fold-belt domain and an interior dome-and-basin domain, with intra-tessera basins with volcanic fill. Not to scale. Taken from Hansen et al. (1999); see also Hansen et al., Hansen et al., 2000). Crustal plateaus observed on Venus appear to be preserved relics of once more extensive features (c.f. Fig. 2a).

facing, steeply dipping, generally sub-parallel to parallel walls (ribbon grabens or simply "ribbons"; see Hansen and Willis, 1998; Hansen et al., 2000). Based on their analysis of the surface fracture topologies in detailed radar images, Hansen and Willis (1996, 1998), Phillips and Hansen (1998) and Hansen et al. (2000) proposed two end-member fracture mechanisms for the formation of tessera ribbons: tensile fracture and shear fracture (Figs. 2c and d; 5; see also Ghent and Hansen, 1999; Gilmore et al., 1998). The dominance of one over the other was observed to vary geographically between crustal plateaus. A second set of short, lenticular, "complex" grabens (referred to herein as simply "grabens"), 10–35 km wide, is also present (Fig. 2b). Note that the Utah Canyonlands have been proposed as a smaller, terrestrial analogue for tessera ribbons (Ivanov and Head, 1996; Hansen et al., 1999, 2000). However, Schultz et al. (2007) presented a detailed dissenting view.

Both ribbons and grabens are generally perpendicular to the tessera fold crests (Fig. 2), and to large segments of the crustal plateau margins (Figs. 2a and 3). Ribbons were generally taken to reflect tectonic extension by symmetrical normal faulting in response to far-field, tectonic tensile stress of unspecified origin (e.g. Hansen and Willis, 1996, 1998; Gilmore et al., 1998; Ghent and Hansen, 1999; Hansen et al., 2000; Ruiz, 2007).

2.3. *Crustal plateau paradigms*

Hansen (2006, 2018) and Gilmore and Head (2018) provide recent comprehensive overviews of early models of crustal plateau formation, which sought to explain associated crustal thickening by either crustal flow above mantle downwellings (e.g. Bindschadler et al., 1992; Gilmore and Head, 2000), or magmatic underplating and crustal inflation associated with mantle upwellings, i.e. plumes (e.g. Phillips and Hansen, 1998; Ghent and Hansen, 1999; Hansen et al., 1999, 2000). In either model, plateau roots represent thickened crust (see also James

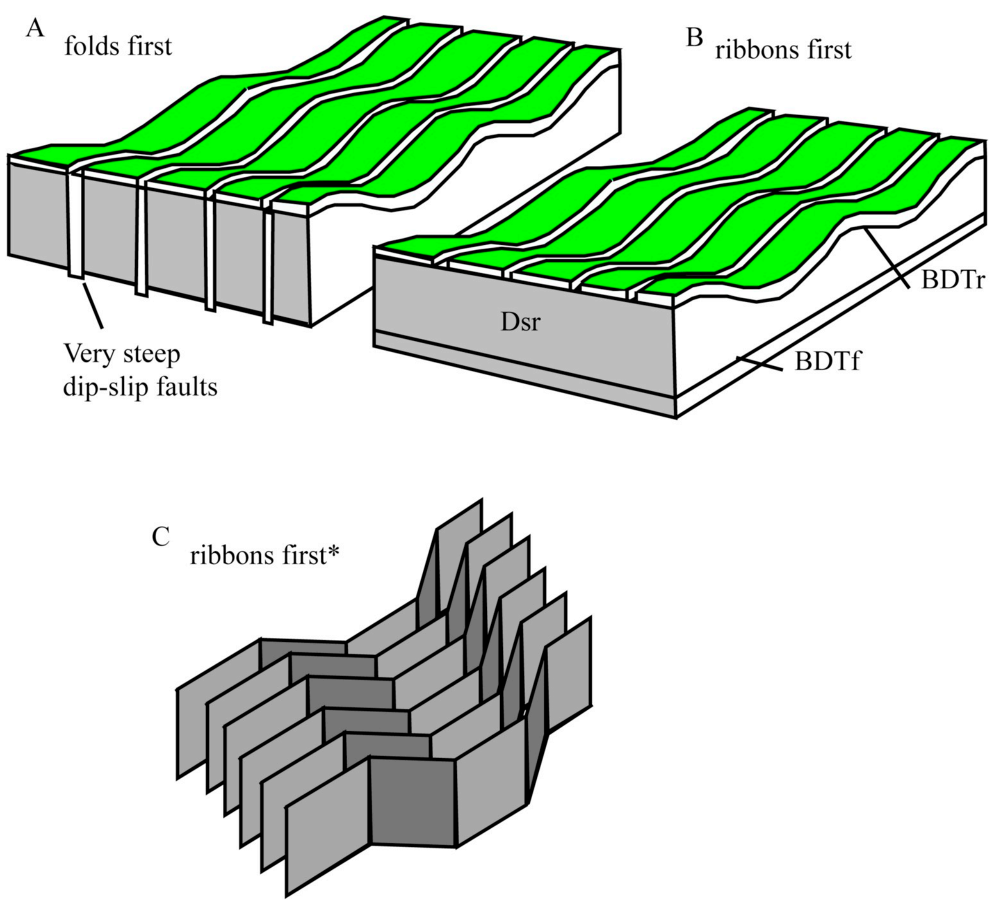

Fig. 4. Schematic diagrams of temporal relations between tessera ribbons and folds. A and B are adapted from Hansen and Willis (1998); see also Ghent and Hansen, 1999), according to whom (i) if folds predate ribbons the crust must be dissected by closely spaced, near vertical faults; (ii) alternatively, ribbons that predate folds are the result of tensile or shear fractures within a brittle membrane above a ductile substrate (Ds); and (iii) depth to the brittle-ductile transition (BDT) increases with time and fold formation. BDTr: BDT at time of ribbon formation. BDTf: BDT at time of fold formation. Dsr: Ds during early ribbon formation. C: However, ribbons and their associated faults would have formed a significant anisotropy; horizontal and two-dimensional (linear) if surficial, or vertical and three-dimensional (planar) if more deeply rooted. Either would be oriented at a low-angle to the principal shortening direction associated with observed folds. Accordingly, folding of pre-existing or contemporaneous ribbons (B) should have resulted in widespread kink, chevron or box-style deformation of tessera ribbons. Not to scale.

et al., 2013; Harris and Bédard, 2014a), while lowland inliers with associated tessera terrain features represent viscously collapsed plateau relics within the younger volcanic plains. Most importantly, both models were used to predict diagnostic time sequences for the generation of tessera terrain structures: downwelling would produce initial folding, while upwelling would result in initial extensional features (ribbons).

However, reported remote observations were contradictory (Fig. 4), variously due to data resolution, different locations, and a model-driven approach (see 2.4 and 2.5 below). Some workers reported that ribbons formed prior to spatially associated tessera folds, which were in turn cut by grabens (e.g. Hansen and Willis, 1996, 1998; Ghent and Hansen, 1999; Hansen et al., 1999, 2000). Romeo and Capote (2011), however, reported extension post-dating shortening, although they considered that ribbons and grabens can be contemporaneous. Other workers reported that the folds in question either preceded ribbon formation (e.g. Gilmore et al., 1998 [vigorously challenged by Hansen et al., 2000]; see also Ivanov and Head, 1996; Gilmore et al., 1997); Romeo et al., 2005; Gilmore and Head, 2018), or occurred in different time sequences in different locations (e.g. Bindschadler and Head, 1991; Romeo and Capote, 2011), or were broadly synchronous (e.g. Ivanov and Head, 1996; Hansen, 2006, 2018). However, all appear to agree that grabens post-date tessera folds of all wavelengths.

The widespread punctuation of structural sequences by relatively limited effusive volcanism in intra-tessera basins (Figs. 2 and 3) is similarly temporally complex (e.g. Hansen et al., 2000; Hansen, 2006, 2018; Gilmore and Head, 2018). Intra-tessera basins tend to occur in troughs between fold crests, and both truncate and are cut by ribbons (Fig. 2b) with which they are, at least in part, contemporaneous.

Subsequently, both crustal plateau formation models were challenged, as summarized by Romeo and Capote (2011) and Hansen (2006, 2018). On the one hand, the ability of crustal flow above mantle downwelling to provide the observed crustal plateau topology by crustal thickening, plateau growth, and eventual plateau collapse by viscous relaxation, was called into question on rheological, thermal and temporal criteria (e.g. Kidder and Phillips, 1996; Mackwell et al., 1998; Nunes et al., 2004; Nunes and Phillips, 2007). On the other hand, the upwelling (plume) hypothesis provided no mechanism to produce early shortening by folding, nor the local intensity of that shortening in tessera terrains (e.g. Ghent et al., 2005; Hansen, 2006; Romeo and Capote, 2011). In response to this apparent "dead-end", Romeo and Turcotte (2008) proposed a thematically derived, "pulsating continents" model, based on theoretical consideration of variations in force balance across the vertical crust/mantle boundaries at the edges of crustal plateau: a model that, in turn, was challenged by Hansen (2018).

In addition, given the absence of unique age relations between ribbons and folds in tessera terrains, as acknowledged by Ghent and Hansen (1999) and Ghent et al. (2005), it was difficult to muster systematic, unequivocal, structural geological observations in support of either the upwelling or downwelling models (Hansen, 2006). Nonetheless, based on detailed observations of cross-cutting and superposition relationships in Tellus Regio (see Fig. 1 for location), Gilmore and Head (2018) presented a comprehensive model of crustal plateau development that begins with large, crustal fragments comprising multiple, internally folded, ~50–100 km diameter, ovoid "blocks" delimited by curvilinear "troughs" (their "pre-tessera terrain"; equivalent to basin-and-dome terrain of Hansen and Willis, 1996: see Fig. 3). They proposed that the plateau margins were further deformed in response to drifting, and possible collisions, of rigid plateau interiors (see also Harris and Bédard, 2014a, 2014b). In this paradigm, tessera ribbon fabrics are specifically identified by Gilmore and Head (2018) as occurring late in the structural sequence (see also Ivanov and Head, 1996; Romeo and Capote, 2011), some even post-dating volcanic plains associated with the purported resurfacing of the planet (see 2.1 above).

2.4. Crustal thickness and isotropy

According to Ghent et al. (2005), the lack of unique age relations led Ghent and Hansen (1999); see also Hansen et al., 1999, Hansen et al., 2000) to seek mechanical arguments to ascertain the relative timing of ribbons and folds. Tessera terrain folding is observed on both short- and long-wavelengths. The early version of their updated model started with an initially hot, thin surface layer (hereafter "crust"[4]), whose base was defined by the regional brittle-ductile transition, that developed above a mantle upwelling and progressively thickened with secular cooling. This was based on the visual distinction of end-member tensile-fracture vs shear-fracture ribbons (Hansen and Willis, 1998; see Fig. 2c and d; Fig. 5 and 2.2 above). Hansen and Willis (1998) and Ghent and Hansen (1999) then argued from simple considerations of wavelength to layer thickness ratio and application of classical brittle boudinage theory and buckling theory (e.g. Biot, 1961; Ramberg, 1955, 1962), that it was more geologically plausible (*"reasonable"*, according to Phillips and Hansen, 1998) that the close-spaced ribbon fabrics developed first on hot, thin crust ($<$1–3 km: Ghent and Tibuleac, 2002). Subsequently, tessera folds would have developed with progressively increasing wavelengths as the crust cooled and thickened (see also Hansen and Willis, 1998; Phillips and Hansen, 1998; Hansen et al., 2000; Hansen, 2006). Hansen and Willis (1998); see also Ghent and Hansen, 1999) explicitly applied the boudinage model to both tensile and shear fracture ribbons. However, Gilmore et al. (1998) applied a different, kinematic concept wherein paired, conjugate, surficial normal faults are modelled as intersecting at a fundamental, horizontal discontinuity, the depth of which is a function of fault dip and spacing. This model, in principal applicable to a shear fracture mechanism, was also applied by Hansen et al. (2000) to obtain a thickness estimate for initial, hot, thin crust. Note that both Gilmore et al. (1998) and Hansen et al. (2000) assumed fault dips of 60°. However, Schultz et al. (2007) questioned the validity of simple derivation of crustal thickness from structural wavelength related to paired normal faults. In a detailed critique, they showed that the method is necessarily based upon unwarranted assumptions regarding fault symmetry, depth of fault initiation and intersection, fault dip, and distribution of slip on fault planes, among others.

The absence of a second, spatially periodic, extensional, brittle fabric with a characteristic wavelength, different to that of the observed ribbons, was taken to indicate the absence of layering (anisotropy), specifically the absence of layers of different mechanical thickness, within the initial tessera crust (Phillips and Hansen, 1998; Hansen and Willis, 1998; Gilmore et al., 1998; Ghent and Hansen, 1999; Hansen, 2006; Ruiz, 2007). This assumption led these authors to propose that all pre-existing macroscopic structures and fabrics, surficial and internal, had been mechanically, thermally or magmatically eliminated ("healed" or "annealed") in all tessera terrains (see also Ivanov and Head, 1996; Hansen et al., 2000ß). This was considered a requirement in order to provide an initial, thin, mechanically homogeneous and isotropic crust within which pervasive ribbon fabrics might form by simple, tectonic, brittle extension over a weak substrate (e.g. Phillips and Hansen, 1998; Hansen et al., 2000). The thinness of the crust, homogenization of at least its surface by viscous relaxation, and magmatic sealing of internal fractures and anisotropies (e.g. layering) were attributed to very high geothermal gradients, coupled with a potential surface temperature of ~1000 K (e.g. Phillips and Hansen, 1998). However, this isotropic and mechanically homogeneous model crust was challenged by Romeo and Capote (2011) who presented fundamental arguments in favour of a more geologically realistic, layered (anisotropic), and likely fractured, crust (e,g, volcanic layering and

[4] "Crust" is used here as a shorthand for the hot, thin surface layer whose base was rheologically defined by authors, as opposed to the classical compositional definition.

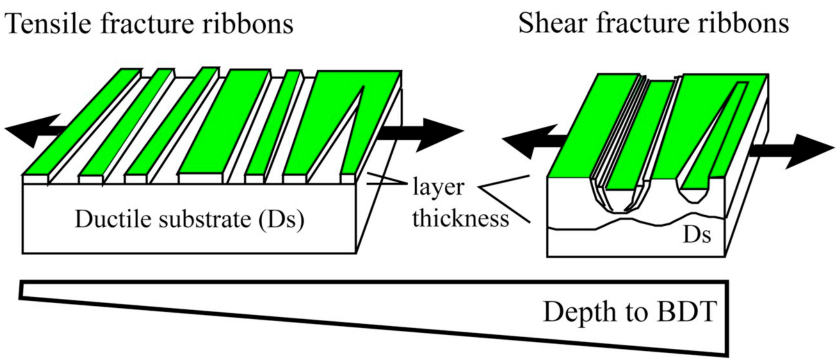

Fig. 5. Tensile and shear fracture end-member block models of ribbons, after Hansen et al. (2000); see also Hansen and Willis, 1998; Ghent and Hansen, 1999), according to whom (i) ribbon trough formation, by either mechanism, requires a ductile substrate at relatively shallow depths; and (ii) opening of tensile fractures would be favoured by a sharp brittle-ductile transition (BDT) at shallow depth, whereas a deeper BDT would favour shear fracture ribbon structures. Not to scale.

internal décollements).

2.5. Exogenic bolide paradigm

According to Hansen (2006), formation of tessera ribbon fabrics required a sharp viscosity decrease at the base of the thin crust, because the purportedly isotropic crust must extend in brittle mode, yet shorten in ductile mode (potentially contemporaneously). Hence, she proposed that the crust was underlain by a liquid; most probably a silicate magma (see also Gilmore et al., 1998; Ruiz, 2007). Combining this rationale with the numerous geological, geophysical and temporal contradictions inherent in both the endogenic upwelling and downwelling models for the origin of crustal plateaus and their tessera terrains (see 2.3. above), Hansen (2006, 2007a, 2007b, 2007c, 2015) proposed a new exogenic model: crustal plateaus as huge lava ponds generated by massive mantle melting, a consequence of large bolide impacts into a thin planetary lithosphere (see also Hamilton, 2005, 2019). In this context, a ßtessera terrain represents congealed *"scum"* on the surface of a lava pond, deformed by thermally-driven convection in the underlying magma. Hansen (2006, 2007a, 2007b, 2007c, 2015) emphasized that formation of *"congealed"* (crystallized) pond scum via secular cooling accounted for both the proposed progressive thickening of an initial hot, thin surface layer, and its purported mechanically homogeneous and isotropic nature. In addition, according to the model, isostatic adjustment of buoyant, depleted residuum in the upper mantle would result in plateau uplift: subsequent delamination, if it occurred, could be responsible for plateau collapse.

This exogenic model was challenged by Romeo and Turcotte (2008) and Romeo and Capote (2011) on several counts: (i) it is unclear whether large bolides can catalyze significant mantle melting (e.g. Ivanov and Melosh, 2003), and (ii) it is difficult to transmit stress capable of deforming a brittle crust of kilometric thickness and driving locally strong shortening strains within it from an underlying liquid magma. In addition, (iii) the presence of short-wavelength folds of constant trend over hundreds of kilometers is inconsistent with the irregular and unstable geometries of convecting cells in the substrate (as opposed to the dome-and-basin structure in crustal plateau cores: see Fig. 3). Furthermore, this model-driven exogenic interpretation stands in stark contrast to the observation-based accretionary crustal plateau model of Gilmore and Head (2018; see 2.3 above). In any event, the exogenic model does not address the possibility that plateaus associated with tessera terrain may not be mafic in composition (e.g. Gilmore, 2015; Gilmore et al., 2015; however, see Wroblewski et al., 2019).

3. Internal inconsistencies and assumptions

Inconsistencies tend to develop from generalised assumptions that are uncritically accepted as part of a given narrative. Numerous inconsistencies are apparent within much of the published work on tessera terrain development in the context of crustal plateaus. In what follows, I will suggest that most inconsistencies can be traced to the interpretation of tessera ribbon fabrics as long, narrow, extensional grabens-like structures generated by far-field, tectonic tensile stress. Such an interpretation imposes significant difficulties for the formulation of an internally consistent, global model for the development of tessera terrains. Accordingly, the rest of this paper will examine how current models for the mechanics and relative timing of the formation of ribbons represent the principal impediments to understanding the structural geology of tessera terrains on Venus.

3.1. Ribbon mechanics

As reviewed above, many published papers proposed that the development of homogeneously distributed, closely spaced, pervasive, short wavelength, extreme aspect ratio tessera ribbons by brittle fracturing occurred at low confining pressure over vast areas of thin crust by extension under tensile stress generated by unspecified far-field tectonic boundary conditions. However, the assumption that one can treat the ribbon fabrics as classical (symmetrical) boudinage features is inconsistent with geological observations, as well as with analogue and numerical experiments, all of which show that tectonic pull-aparts adjacent to a free surface form clusters of asymmetrical, similar-facing dominoes (e.g. compare Ramberg, 1955; Rast, 1956; with Fossen and Gabrielsen, 1996; Brun, 2002; Montesi and Zuber, 2003; Corti, 2005), incompatible with tessera ribbons that purportedly comprise pairs of symmetrical, opposite-facing, steep-dipping, tectonic normal faults.

In the classical (dynamic) boudinage model (e.g. Ramberg, 1955; Rast, 1956) a more viscous layer is sandwiched between bounding, less viscous material. The spatially periodic pulling apart of the more viscous layer occurs via the transmission of shear stress to the bounding interfaces as the bounding less viscous material yields. However, some proponents of the extensional ribbon model couched their hypothesis as the tectonic extension of a thin *surface layer*. In order to apply the classical boudinage model, Hansen and Willis (1998) proposed that the venusian atmosphere represented the upper bounding less viscous material. However, the timing of the formation of the present, dense, CO2-rich (96.5%) venusian atmosphere remains unknown (e.g. Grinspoon, 2018; Kane et al., 2019), as does the age of the formation of the tessera ribbon fabrics (see 2.1 above). Regardless, the viscosity of the current venusian atmosphere is only ~50% greater than that of present-day Earth (~3 pa.s vs ~2 pa.s, respectively; e.g. Hassanalian et al., 2018); it is not comparable with the viscosity of a venusian subcrustal substrate, even if the latter was fully molten. Nonetheless, Hansen and Willis (1998) and Ghent and Hansen (1999) applied the classical boudinage model to derive layer thickness from ribbon geometries for the initial thin crust during purported tectonic extension (Fig. 4b).

Other authors compared a kinematic model of ribbon formation by symmetrical paired conjugate normal faults (see 2.4 above) with terrestrial crustal-scale analogues, such as the Basin and Range (e.g. Gilmore et al., 1998; Ghent and Tibuleac, 2002) or the East African Rift (e.g. Gilmore et al., 1998). However, the East African Rift is a type example of localised, as opposed to distributed, faulting, whereas the absence of strain localisation is a defining characteristic of tessera ribbon fabrics. Furthermore, Ruiz (2007) drew direct comparisons between numerical and analogue studies of extended terrestrial lithosphere and tessera ribbon development. However, the results of such laboratory studies are inconsistent with their application to tessera ribbon fabrics. First, inspection of the top surfaces of such analogue models clearly show that distributed brittle faults are short and discontinuous, and that the faults are not systematically developed in opposite-facing pairs (e.g. Corti, 2005, figs. 2 and 4). Second, Montesi and Zuber (2003) state that extensional brittle faulting, even if initially distributed, tends to localise in *"more efficient"* localised faults (see also Koptev et al., 2018). In short, it is inconsistent to compare any of the published models of tessera ribbons with either classical boudinage, or with lithospheric extension.

3.2. Ribbon timing

The foregoing mechanical, dynamic and kinematic inconsistencies notwithstanding, a lack of unique age relations between tessera ribbon and fold development led to the formulation and general acceptance of an initially hot, thin crust that progressively thickened and deformed at longer wavelengths with global secular cooling, on the explicit basis of purported mechanical arguments and assumed geological plausibility (e.g. Ghent and Hansen, 1999; Hansen et al., 2000; Ghent et al., 2005). However, ribbon fabrics and short- to medium-wavelength tessera folds are now thought to have formed contemporaneously (e.g. Hansen, 2006, 2018; see also Romeo and Turcotte, 2008), or even last in the structural sequence (Gilmore et al., 1998; Gilmore and Head, 2018). In addition, numerical modelling indicated that shorter wavelength folding could form *after* the longer wavelength folds, even in thickened model crust (Ghent et al., 2005). The latter authors also reported that none of their numerical models developed a brittle surface layer, or a brittle-ductile transition. Instead, they developed a semi-brittle surface zone with associated crystal-plastic creep, thereby weakening the model "crust". Consequently, shortening strains by folding in excess of 40% were required to contribute any significant model plateau relief. Despite these caveats, the current tessera terrain narrative still invokes progressive crustal plateau thickening, now with contributions from deformation (folding) and magmatic inflation, in addition to secular cooling (e.g. Hansen, 2006, 2018).

Two inconsistencies are apparent here. First, the crustal-scale open folds, described by most authors, may warp the crust (Hansen, 2018), but they cannot significantly thicken it. Tectonic thickening by folding would require extensive development of recumbent or thrust nappes (e.g. Romeo et al., 2005; Romeo and Turcotte, 2008; Romeo and Capote, 2011). Second, given that (i) medium-wavelength folds can form contemporaneously with ribbons, (ii) short-wavelength folds can form at the end of the folding sequence, and (iii) ribbon fabrics can even form last, the structural sequences of tessera terrains no longer point to progressive crustal thickening, in any event.

The notion that tessera ribbon fabrics imply extensive, uniform, distributed, pervasive, brittle extension of thin crust whose base represents a steep rheological gradient led to yet another assumption, presented as a requirement of the model (e.g. Phillips and Hansen, 1998; Hansen and Willis, 1998; Gilmore et al., 1998; Ghent and Hansen, 1999; Hansen, 2006; Ruiz, 2007): that this initial crust was mechanically homogeneous and isotropic. Hansen and Willis (1996) had already recognised that extensive, penetratively distributed, homogeneous, brittle, extensional strain would be problematic in a mechanically inhomogeneous medium. Indeed, it would be especially problematic at the low confining pressures within thin crust (see Hanmer, 1989 for discussion of natural terrestrial examples of distributed brittle deformation and confining pressure). Hence the proposed model requirement for a mechanically homogeneous, isotropic crust decoupled from the underlying substrate by either a brittle-ductile transition, or by molten magma (e.g. Hansen, 2006). However, as argued in the next section, tessera terrain crust was most likely internally anisotropic.

3.3. Crustal anisotropy

In order to obtain quantitative estimates of crustal shortening, Ghent and Hansen (1999); see also Hansen, 2006) made several explicit assumptions: (i) that the shortening brittle crust maintained a constant thickness during (parallel-style) folding; (ii) that no homogeneous layer-parallel shortening occurred prior to fold amplification; and (iii) that the base of the thin, brittle surface layer was the regional brittle-ductile transition (see also Hansen and Willis, 1998; Phillips and Hansen, 1998; Hansen et al., 2000). These authors were concerned with bending of the entire thickness of a thin crust, from which to derive estimates of progressively changing crustal thickness with increasing fold wavelength through time. Building on these assumptions, Hansen (2006) stated that *"progressive thickening of a fold layer with earlier formed folds carried piggy-back on younger folds requires an exponential decrease in viscosity with depth"*. However, as pointed out by Romeo and Capote (2011), this is inconsistent with both well-established geological field observations and classical analogue modelling. Specifically, a multilayer of uniform average viscosity with embedded, thinner, more competent sublayers, can reproduce the venusian observations in response to bulk, layer-parallel shortening (e.g. Ramberg, 1962, 1963, 1964; Ramsay, 1967 pp. 372–386; Hobbs et al., 1976). These classical publications demonstrate that less competent layers can thicken by homogeneous layer-parallel shortening until they reach a critical thickness where buckling instabilities will begin to rapidly amplify (see also Biot, 1961). Furthermore, Hansen (2006) describes *"straight limbs, angular hinges, and no homogeneous layer shortening prior to fold formation"*. This is inconsistent with a model that requires a mechanically homogeneous, isotropic crust, which would deform by bending associated with internal tangential longitudinal strain (e.g. Ramsay, 1967, fig. 7–63). Rather, the observations appear to describe chevron folds, which would point to the presence of a *strong* internal mechanical anisotropy within the deforming crust (e.g. Weiss, 1969; Cobbold et al., 1971), inconsistent with the explicit rejection of internal crustal anisotropies (see 2.4 above). In any event, parallel-style and kink-, chevron- or box-style folding would all tend to generate décollement surfaces at the base of a fold train, potentially within the crust (e.g. Suppe, 1983; Mitra, 2003).

The development of pervasive, tessera ribbon fabrics prior to, or contemporaneously with, perpendicular folding (Figs. 2 and 3) of any wavelength resolvable in the Magellan radar images, is problematic. Ribbons and their associated faults would have formed a significant anisotropy; horizontal and two-dimensional (linear) if surficial (see geometry of Fig. 4b), or vertical and three-dimensional (planar) if more deeply rooted (see geometry of Fig. 4a). Either would be oriented at a low-angle to the principal shortening direction associated with observed folds. Accordingly, early-formed ribbons should have been widely deformed (shortened) in kink, chevron, or box fold styles (e.g. Weiss, 1969; Cobbold et al., 1971; Fig. 4c). Had this occurred, it may have resembled so called "lava flow" terrain, wherein *"ribbons and folds appear to be deformed into chevron-like folds, with the resulting fabric appearing similar to that of a piece of rumpled corduroy"* (Hansen and Willis, 1996, fig. 3; see also Hansen et al., 2000, plate 1d; Hansen, 2006, fig. 13). However, such deformation of ribbon fabrics is apparently rare (Hansen, 2006). The purported early nature of ribbon fabrics in the structural sequence of tessera terrains is thus inconsistent with the apparent absence of generalised ribbon shortening.

4. Dyke swarms

Citing Hansen and Willis (1998), Hansen et al. (2000) state that *"the critical characteristic of ribbon fabric is not the dip of trough-bounding*

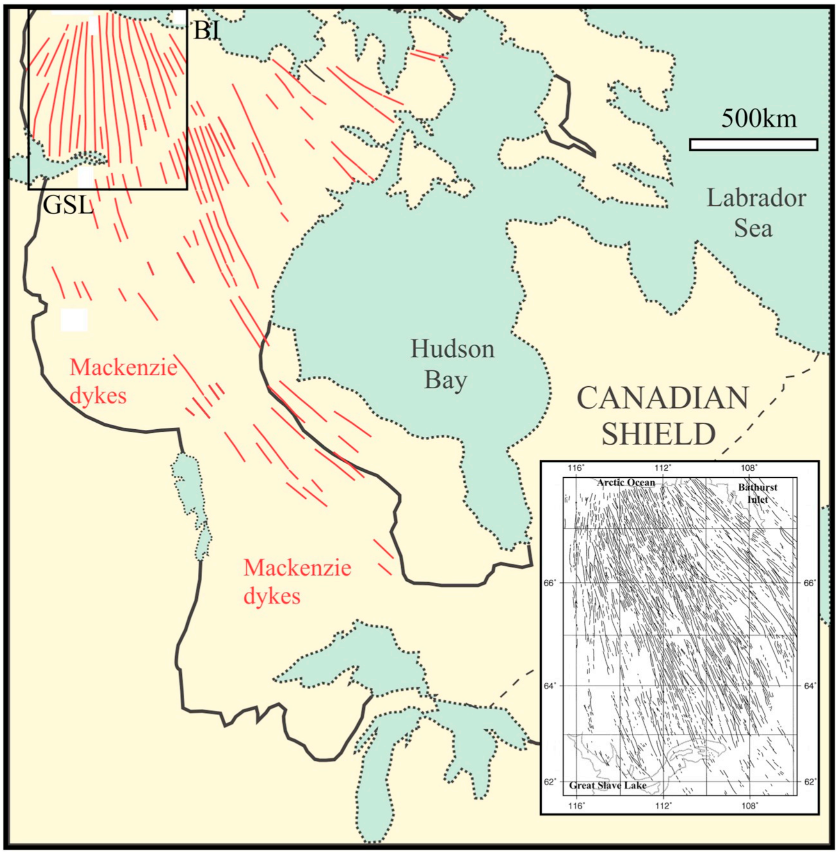

Fig. 6. Schematic representation of the Mackenzie Dyke Swarm, Canada, derived and generalised from aeromagnetic data. Black lines mark the limit of Paleozoic cover rocks. Adapted from Ernst and Buchan (2001). Inset, located by rectangular outline at upper left, is a line drawing of part of the Mackenzie dyke swarm (linear traces derived from aeromagnetic data), Slave Craton, Northwest Territories, Canada, taken from Pilkington and Roest (1998). This figure highlights both how the dyke swarm in the Slave Craton resembles tessera ribbon fabrics on Venus at the crustal plateau scale (see Fig. 2a), and how the linear vs radial aspect of the dyke swarm is a function of scale within the overall dyke swarm (compare this with the apparent contrast between Figs. 2a and 3).

structures but rather the regular spacing of ribbon ridges and troughs, which reflect a structural wavelength and imply a very shallow depth to the brittle-ductile transition". This raises the question: where else in the Solar System have regularly spaced, long, high aspect ratio, linear features, at least locally associated with surficial troughs and/or grabens, been observed over wide areas of the order of thousands to millions of square kilometers? Dyke swarms on the terrestrial planets (e.g. Ernst et al., 1995, 2001; Ernst and Buchan, 2001; Aspler and Ernst, 2003, and references therein).

The development of regularly spaced, uniform width, long, narrow mafic dykes with extreme aspect ratios in vast swarms is a characteristic feature of parts of Mars, of Venus outside of tessera terrains, and of Earth (e.g. McKenzie et al., 1992; Mège and Masson, 1996; Ernst et al., 1995, 2001, 2003; Wilson and Head, 2002, fig. 10; Scott et al., 2002,5.2; Krassilnikov and Head, 2003; Mège et al., 2003; Studd et al., 2011; Buchan and Ernst, 2019; Magee et al., 2019. Dyke swarms trend across these planetary surfaces without deviation in trend or width, over extreme distances (e.g. Fig. 6).

4.1. Theoretical considerations

The relationship between long, narrow grabens and dyke emplacement on terrestrial planets and the Moon has long been debated (e.g. Golombek, 1979; Magee et al., 2019 and references therein). Briefly, there are three principal scenarios under discussion: (i) passive emplacement of dykes in fractures beneath grabens generated by a regional tensile stress of far-field tectonic origin (e.g. Tanaka et al., 1989; Mège and Masson, 1996; Wyrick et al., 2014); (ii) forceful emplacement of dykes as pressurised cracks driven by magma and evolved gas pressures of sufficient magnitude to overcome brittle failure strength and ambient lithostatic pressure that results in graben formation above blind dykes (e.g. Delaney and Pollard, 1981; Pollard et al., 1983; Mastin and Pollard, 1988; McKenzie et al., 1992; Rubin and Pollard, 1988; Ernst et al., 1995; Wilson and Head, 2002; Scott et al., 2002; Townsend et al., 2017); and (iii) a hybrid scenario with both tectonic and magmatic contributions to extension, dyking and graben formation (e.g. Rubin, 1992, 1995; Mège and Masson, 1996; Koenig and Pollard, 1998; Mège et al., 2003; Biggs et al., 2009; Koehn et al., 2019), or the exploitation of pre-existing cracks by rising magma (e.g. Wyrick et al., 2014). The existence of dykes or dyke swarms at depth beneath long, narrow surface grabens is generally accepted (e.g. Murray and Pullen, 1984; Mège and Masson, 1996; Ernst et al., 2001; Scott and Wilson, 2002; Scott et al., 2002; Wilson and Head, 2002; Aspler and Ernst, 2003; Mège et al., 2003; Goudy and Schultz, 2005; Schultz et al., 2007, 2010; Patterson et al., 2016; Magee et al., 2019; see, however, Wyrick et al., 2014). Ongoing debate is principally focused on the theoretical mechanisms and processes that might genetically link dyke emplacement and graben formation (e.g. Mège et al., 2003; Biggs et al., 2009; Schultz et al., 2007, 2010; Wyrick et al., 2014; Townsend et al., 2017; Koehn et al., 2019).

Forceful dyke intrusion is modelled in terms of pressurised, vertical, discoid cracks that dilate and propagate, driven by internal magmatic and evolved gas pressures (e.g. Pollard et al., 1983; Mastin and Pollard, 1988; McKenzie et al., 1992; Rubin and Pollard, 1988; Wilson and Head, 2002; Scott et al., 2002). The model dyke mid-point is located at the neutral buoyancy level for the magma in the crust, and propagates vertically and laterally. Blind "non-feeder" dykes that do not breach the surface are common (e.g. Mège and Masson, 1996; Wilson and Head, 2002). The analysis is predicated on the mechanically brittle and elastic behaviour of the host rock associated with such blind dykes (e.g. McKenzie et al., 1992; Pollard et al., 1983; Townsend et al., 2017). According to the simple version of the model (Fig. 7), dyke injection into a homogeneous, isotropic crust involves the dilation of a vertical pressurised crack, which perturbs the stress field related to the lithostatic pressure gradient, especially at the top of the dyke. Two tensile stress maxima form symmetrically to either side of, and above the dyke

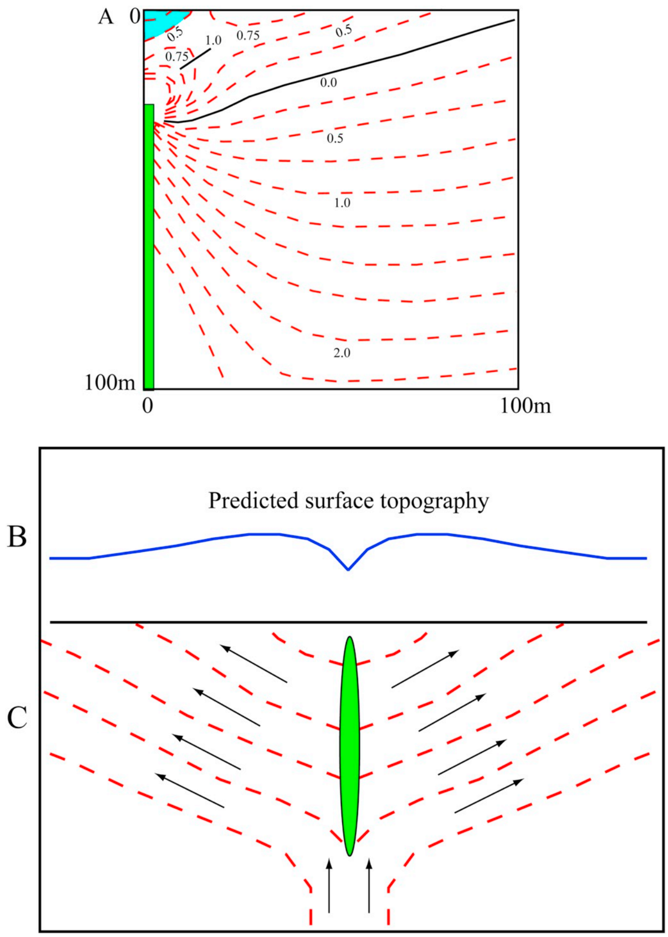

Fig. 7. Schematic basic model of a surface graben developed above local stress anomalies generated by the emplacement of a blind magmatic dyke modelled as a pressurised elliptical crack. A: Distribution of maximum compressive stress trajectories (dashed lines with Mpa values) adjacent to and above a pressurised blind dyke (shaded bar at left). Maximum compressive stress drops below 0.5 Mpa in the shaded zone at top left. Horizontal scale is lateral distance from the dyke; vertical scale is depth. See main text for detailed description. Adapted from Pollard et al. (1983). B and C: Lower image is a schematic of displacement trajectories (dashed lines and arrows) associated with inflation of a pressurised blind dyke (shaded ellipse at centre). Flat horizontal line is the schematised surface. Upper image is the model prediction for the surface topography above the dyke. Not to scale. Adapted from Schultz et al. (2010).

tip, and a zone of compressive stress forms over the top of the dyke (Fig. 7a). The local tensile stress maxima lead to the formation of steeply dipping normal faults that bound a graben floor above the dyke, and the local compressive stress acts to resist further propagation of the pressurised crack toward the surface (e.g. Gudmundsson, 1990; Fig. 7a). In addition, in the case of horizontal dyke propagation, magma flow does not significantly contribute to the vertical pressure gradient within the dyke, i.e. it does not contribute to further vertical dyke propagation toward the surface. The surface expression of the upward and outward displacements, associated with the locally generated stress field adjacent to the dyke tip (Fig. 7c), is manifested as a ridge-trough-ridge profile perpendicular to the dyke plane (Fig. 7b). The bending associated with this topographic profile gives rise to fracturing and eventual normal faulting that may flank the trough (Schultz et al., 2004, 2010).

However, the foregoing is not universally accepted. Rubin (1993) challenged the assumption of linear elastic crustal behaviour; however, Mastin and Pollard (1988) explicitly considered inelastic behaviour in their modelling. According to Mège et al. (2003), the model of a symmetrical graben forming above a single, blind dyke is unrealistic because most grabens are asymmetrical (see also Melosh and Williams, 1989; Schultz et al., 2007). Moreover, they suggested that individual grabens are underlain by magma reservoirs, topped by swarms of multiple dykes (see also Scott et al., 2002). In addition, Mège et al. (2003) considered that large dykes form by the successive intrusion of thinner dykes, thereby rendering graben formation by dyking difficult. They proposed that regional tectonic stress fields are of sufficient strength to produce narrow grabens without input from magmatic pressure, while allowing that, in some circumstances, dyke emplacement could contribute instantaneously to catalysing the brittle fracture of rock that was already about to fail. Analogue modelling of dyke emplacement led Wyrick et al. (2014) to suggest that the formation of grabens above blind dykes requires pre-dyke tectonic fractures or faults. They concluded that magma is a relatively passive actor in dyke emplacement, and that grabens in themselves are not evidence of the presence of a dyke at depth. Other workers (e.g. Rubin, 1995; Mège and Masson, 1996; Koenig and Pollard, 1998; Wilson and Head, 2002; Scott and Wilson, 2002; Scott et al., 2002; Biggs et al., 2009; Koehn et al., 2019) found that successful dyke emplacement involves components of both regional tectonic extension and forceful intrusion. For example, Wilson and Head (2002) suggested that the role of regional stress is to influence the orientation of sheet-like intrusions, and that local magmatic pressure does the work of fracture formation and dilation. They also recognised the role of magma speed in favouring high strain rates, and thereby brittle failure. In addition, while Schultz et al. (2004, 2010) provided a theoretical criterion for distinguishing the regional tectonic and local magmatic components of a dyke-related stress field manifested in surface topography, normal to the dyke in question (Fig. 7b), this does not appear to have been widely applied to natural examples.

4.2. Surface and field observations

"Assessment of subsurface structure in areas of volcano-tectonic activity is critical to evaluating relationships, for example, between regional extension, faulting and dyke intrusion" (Schultz et al., 2010). Other than Earth (e.g. Aspler and Ernst, 2003; Van Kranendonk et al., 2006; Koehn et al., 2019), this is not yet possible on terrestrial planets. Hence, we must look to remotely observable surface features for clues regarding the presence of magmatic features at depth. Low rates of erosion on Venus and Mars favour the preservation of surface features (e.g. Ferrill et al., 2004). While long, narrow troughs or grabens are taken by some to indicate the presence of subsurface dykes, this is not universally accepted (see 4.1 above). However, some long, narrow troughs or grabens are internally decorated by rimless pit chains, generally interpreted as collapse features (e.g. Schultz, 1988, Schultz, 1989; Ferrill et al., 2004, fig. 1; Wyrick et al., 2004, fig. 6; Mège and Masson, 1996; Scott et al., 2002, fig. 2; Scott and Wilson, figs. 1 and 6; Mège et al., 2003, figs. 3 and 6; Davey et al., 2013; Patterson et al., 2016). Most observations pertain to Mars (e.g. Fig. 8), though they are also reported from Venus (e.g. Patterson et al., 2016) and Earth (e.g. Ferrill et al., 2004; Magee et al., 2019). The debate continues to focus on the causal agency that creates the subsurface volume into which collapse can occur (e.g. Wyrick et al., 2004, fig. 6; Mège and Masson, 1996).

Pit chains have been ascribed to collapse of loose surface material into subsurface void space created by faulting (e.g. Tanaka et al., 1989; Ferrill et al., 2004, figs. 2 and 4; Davey et al., 2013). However, others have proposed that rimless pit chains in long narrow graben troughs are indicative of the presence of blind magmatic dykes in the subsurface beneath and parallel to the troughs containing the pits (e.g. Fig. 8; Mège et al., 2000; Scott and Wilson, 2002; Scott et al., 2002). Accordingly, the subsurface void space into which surface materials collapse has been variously attributed to (i) dyke magma draw-down consequent upon a potentially cyclical drop in magmatic pressure (e.g. Schultz, 1988, 1989; Patterson et al., 2016); (ii) the thermally/magmatically driven escape of volatiles from either groundwater or a melted

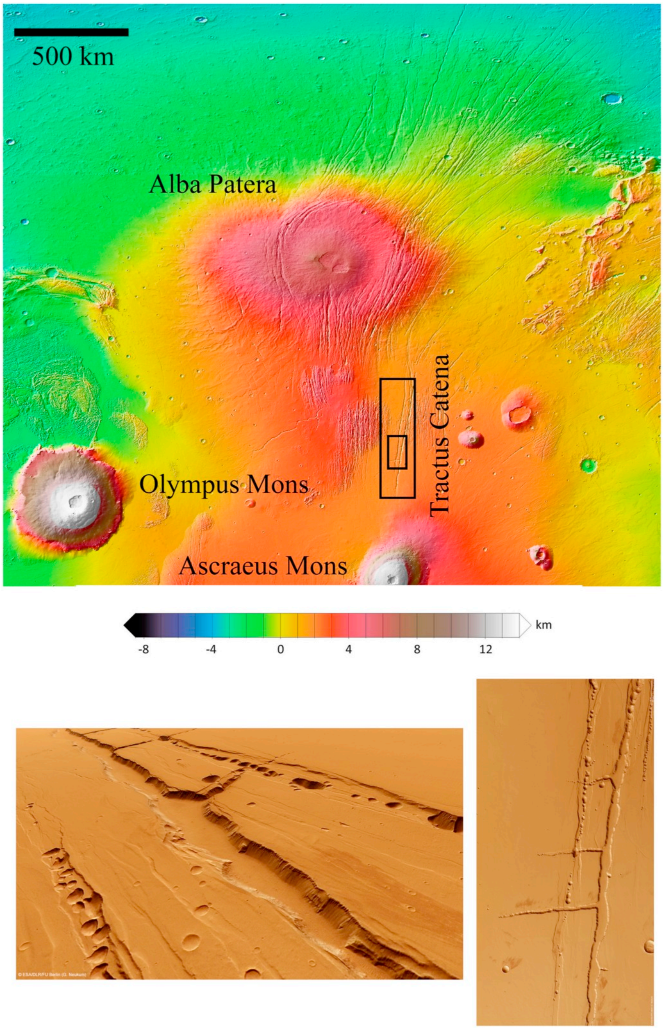

Fig. 8. Pit chains associated with long narrow troughs/grabens of very high aspect ratio on Mars, comparable with the hypothesised origin of tessera ribbons on Venus, as presented in the main text. Top: Setting of Alba Patera (aka Alba Mons), located north of the Tharsis Uplift, cut by a swarm of narrow, high aspect ratio troughs. The larger box emcompasses the location of Tractus Catena. (commons.wikimedia.org/wiki/File:Alba_Mons_MOLA_zoom_64.jpg). Lower right: Detail of long narrow troughs containing pit-chains in Tractus Catena, corresponding to the smaller box in the upper image. Lower left: a perspective view of similar features elsewhere in Tractus Catena (Esa.int/Science_Exploration/Space_Science/Mars_Express/The_pit-chains_of_Mars_a_possible_place_for_life). Note that the lower images represent visible spectrum data at 1.5–20 m/pixel, compared to ~75 m/pixel for Venus radar data.

cryosphere (e.g. Mège and Masson, 1996); or (iii) leakage of evolved volatiles from a foam layer or gas cavity within the top of a stalled dyke (e.g. Scott and Wilson, 2002). However, Magee et al. (2019) find that they cannot validate these various proposed mechanisms to explain pit chain craters.

Terrestrial observations, both ancient (e.g. Aspler and Ernst, 2003; Van Kranendonk et al., 2006) and modern (e.g. Koehn et al., 2019), indicate that surficial rifts may be genetically associated with dyke emplacement. Other terrestrial observations demonstrate that dyke swarms commonly cut through thick continental crust, and propagate horizontally over extreme distances (1000s km) regardless of pre-existing geological features of any scale (e.g. Baragar, 1977; Fahrig, 1987; Baragar et al., 1996; Ernst et al., 1995, 2001; Buchan and Ernst, 2019; Magee et al., 2019; Fig. 6). In addition, terrestrial dyke swarms are known to be commonly polyphase, comprising multiple, geologically short-lived events, widely separated in time (~10–50 myr; e.g. Cadman et al., 1993; Davies and Heaman, 2014). Hence a variety of timing relationships could be developed with respect to what otherwise appears superficially to be a single, short-lived dyke swarm. Terrestrial observation of natural dykes also demonstrates that their emplacement does not necessarily require regional tensile stress. For example, synkinematic dykes can be emplaced under conditions of high confining pressure in the lower continental crust, even as that crust is deforming crystal-plastically (e.g. Hanrner and Lucas, 1985, fig. 2.9; Hanmer, 1997, p. 64–65; Hanmer et al., 1997; Williams and Hanmer, 2005; Mills et al., 2007). The orientation of such dykes is kinematically controlled, as opposed to stress controlled (Hanmer, 1997; Hanmer et al., 1997). However, the orientation of regionally extensive dyke swarms may be stress controlled, even if their dilation may be primarily driven by magmatic pressures (e.g. Wilson and Head, 2002).

5. Discussion

Is consideration of dyke swarms observed on the terrestrial planets relevant to understanding tessera ribbons fabrics on Venus? I suggest that many of the apparent inconsistencies noted above (see 3) could be resolved if tessera ribbons were not the product of purely regional tectonic extension, and if they were not temporally confined to the early part of the tessera deformation history. A number of the nested, apparently inconsistent hypotheses and contradictory observations regarding tessera ribbon fabrics, reviewed and examined in this contribution, would be better explained if tessera ribbons were in fact associated with dyke swarms.

Principal among these inconsistencies is the generally accepted requirement for an initially *thin* tessera crust, mechanically homogeneous and isotropic, both internally and at the surface, in order to account for the penetratively distributed, uniform development of tessera ribbon fabrics. This model is inconsistent with observed tessera fold geometries that resemble chevron-style deformation, which would require the presence of a well-developed, intra-crustal, planar anisotropy. Notwithstanding, the notion of a homogeneous, isotropic crust then led to the further requirement for geologically speculative mechanical or thermally-driven processes that would "heal", "anneal" and seal a hot, thin, early crust. These mechanical and thermal postulates in turn provided the underpinning for the model-driven impact bolide "lava pond" paradigm, wherein the pond "scum" would form a thin, homogeneous, isotropic surface layer within which the ribbons would develop uniformly. The initial justification of hot, thin, early crust that thickened with secular cooling and the formation of progressively longer wavelength folds is itself inconsistent with observation. First, it was based on inappropriate application of both the classical dynamic boudinage model and a kinematic model of symmetrical extension by paired conjugate normal faults to determine crustal thickness from structural wavelength considerations. Second, if the formation of tessera ribbon fabrics was indeed coeval with, and cross-cutting with respect to short- and medium-wavelength folds, then there is no basis for proposing an early thin tessera crust in the first place. It is, therefore, inconsistent to present the structural geology of tessera terrains in terms of secular crustal thickening. Moreover, if tessera ribbon fabrics had indeed preceded long-wavelength folds, they should have been extensively crenulated by kink- or chevron-style deformation.

The second major inconsistency is the assumption that the surface expression of tessera ribbons as long, narrow troughs, can *only* be explained by brittle tectonic extension driven by far-field stresses, structurally manifested as opposite facing pairs of either tensile fractures, or symmetrical, conjugate shear fractures associated with down-dip slip (e.g. Hansen and Willis, 1998; Hansen et al., 2000). This model is presented without accounting for the lack of strain localisation within the geographically extensive tessera ribbon fabrics. Parallels proposed between tessera ribbons and observations of terrestrial examples of lithospheric-scale extension are invalid because the latter either involve families of asymmetrical, uniformly facing dip-slip faults (e.g. Basin and Range), or important strain localisation (e.g. East African Rift), neither of which correspond to tessera ribbon fabrics. Equally, comparisons drawn with experimental simulations of surficial tectonic extension are invalid because the cited modelling produces short, discontinuous brittle fractures that are not systematically developed in opposite-facing pairs. Moreover, they clearly predict the localisation of deformation.

5.1. Resolving observational and model inconsistencies by a dyking process?

Gilmore et al. (1998) discussed the potential for once voluminous molten rock at shallow (< 1 km) crustal depths, contemporaneous with tessera terrain deformation on Venus (see also Banks and Hansen, 2000; Hansen et al., 2000; Hansen, 2006, 2018; Ruiz, 2007), and explicitly asked whether it would have given rise to dyke injection. Phillips and Hansen (1998) and Hansen et al. (2000) invoked magmatic injection into thin crust in order to seal pre-existing fractures and other anisotropies as a pre-requisite for the formation of extensive, geometrically uniform, brittle tessera ribbon fabrics. Gilmore et al. (1998) also discussed the theoretical role that "suprahydrostatic" or magma pressure could play in promoting the propagation of regional fracturing under the pressure conditions of the deeper crust. Their paper specifically refers to Tractus Fossae on Mars (Gilmore et al., 1998, fig. 11), a swarm of close-spaced, long, narrow troughs developed north of the Tharsis uplift (cf. Fig. 8). However, they postulated that significant dyking should have resulted in more volumetrically important volcanism than is observed, which would have hidden the purportedly tectonically extensional ribbon fabrics. While this may be true for dykes that breach the planetary surface, it would not be the case for blind dykes. Nonetheless, the absence of such volumetric lava in tessera terrains led Gilmore et al. (1998) to reject the notion of dykes as a significant feature in tessera terrains. However, as already noted (e.g. Hansen et al., 1999, 2000; Hansen, 2006, 2018; Gilmore and Head, 2018), widespread, albeit volumetrically limited, effusive volcanism in intra-tessera basins does indeed occur throughout the tessera structural sequence (see Figs. 2b and 3). This is permissive of the lavas of the intra-tessera basins having been fed by those ribbon-associated dykes that managed to breach the surface. However, this begs the question as to why major volcanic surface deposits are not observed within crustal plateaus. I can only speculate, but given that preserved crustal plateaus are relics of once more extensive crustal-scale features (compare Figs. 2a and 3; compare also Fig. 6 and its inset), and that extensive mantle melting may have occurred multiple times in multiple places (see 2.1 above), the disposition of ribbons observed today may not require local sources of magma, or eruptive lava, to have been located directly beneath the current plateau surfaces. I further speculate that such a scenario of crustal relics coupled with planet-wide volcanic resurfacing would allow for potentially limitless magmatic availability. Furthermore, crustal shortening or foundering in areas now hidden from view would relax limits on crustal extension consequent upon the injection of giant dyke swarms.

Remote surface observations of tessera terrain on Venus are at least suggestive of the presence of blind dykes. Pit chains have been identified, broadly parallel to spatially associated ribbon fabrics (Hansen, 2006). They cross-cut all tessera fabric elements (Hansen, 2018), but are not commonly reported; possibly a partial function of the resolution of the Magellan imagery (cf. Fig. 8). Tessera ribbon fabrics have been reported to cut across folds of all wavelengths, as well as geological boundaries between intra-plateau domains, without change in width, spacing or trend (e.g. Ghent and Hansen, 1999; Gilmore et al., 1998; Gilmore and Head, 2018). Locally, tessera ribbon fabrics occur as two mutually near-orthogonal sets (Ivanov and Head, 1996, fig. 22; Ghent and Hansen, 1999). Collectively, these observations are readily explicable if ribbon fabrics are surface manifestations of dyke swarms. In contrast, they are difficult to account for as the extensive, uniform, pervasively distributed, tectonically-driven, brittle extension of thin crust in response to far-field driven regional tensile stress, especially in the context of a model that requires an initial, mechanically homogeneous and isotropic crust for such ribbons to develop.

The relationship between the emplacement of generally blind dykes and surface deformation and collapse is not universally accepted, nor are there universally accepted, readily observable/applicable surface criteria for identifying the presence of subsurface dykes. In addition, as noted above, there is ongoing debate regarding the relative roles of regional tension and internal magmatic pressure in dyke emplacement per se. Accordingly, proposing subsurface dyke swarms as a potential solution for inconsistencies in the tessera ribbon paradigm is necessarily speculative, especially in light of the resolution of currently available data. Furthermore, while I will refer to "giant dyke swarms" in the following discussion, theoretical considerations of the potential general relationship between blind dykes and surface deformation (troughs or grabens) do not require that potential tessera dykes be of the same width as those of established "giant" dyke swarms. With these caveats in mind, I will now discuss the possible role of dyking in the development of tessera terrain fabrics.

As noted above (4.2), giant mafic dyke swarms are capable of cutting across thick lithosphere. Their presence in tessera terrains would impose no constraints on either crustal thickness or mechanical rheology pertaining at the time of their emplacement. There would therefore be no requirement to invoke either initially thin, mechanically homogeneous and isotropic crust, or loosely defined mechanical and thermal mechanisms for crustal and surface "healing", "annealing" or sealing. In addition, there would be no reason to (incorrectly) apply the either classical boudinage model or a model of symmetrical extension by paired conjugate normal faults to gauge the thickness of the tessera terrain crust during ribbon fabric formation. Indeed, invoking the tectonic extension of a surface layer predicts domino-style, pull-aparts that will produce asymmetric grabens grouped into similar-facing families (see 3.1 above), a feature not characteristic of tessera ribbons on Venus. One clear advantage of the blind dyke hypothesis proposed here in this regard is its prediction of the symmetrical tessera ribbons reported from venusian crustal plateaus.

Absent the need to explain initially hot, thin, mechanically uniform and isotropic crust, there would be no reason to invoke an exogenic "lava pond" model to explain tessera terrains and their associated crustal plateaus (however, see Hamilton, 2005, 2019). Moreover, there would be no basis for postulating extension of such a model to terrestrial Archean greenstone belts (Hansen, 2006, 2015, 2018), and ignoring the latter's well-known complex internal stratigraphic, structural and tectonic histories (e.g. de Wit and Ashwal, 1997; Bleeker, 2002; see also Hanmer et al., 2004, 2006).

While terrestrial giant mafic dyke swarms are certainly known to be associated with the localisation of tectonic rifting and cratonic/continental separation (e.g. Ernst and Bleeker, 2010), this can be a consequence, not necessarily a cause. Large expanses of giant dyke swarms are uniformly developed and contain no internal indications of strain localisation (Ernst et al., 1995, 2001; Ernst and Buchan, 2001; Fig. 6). Indeed, Mège et al. (2003) speculated that distributed brittle surface deformation without localisation may be facilitated by the presence of underlying dyke swarms. Their presence in tessera terrains on Venus could potentially account for regional extension without localisation of brittle deformation.

What controls the general orientation of tessera ribbons, especially if they are indeed radial to the original crustal plateaus with which they are spatially associated (Fig. 3)? As noted above (4.1), regional stress fields can influence the orientation of sheet-like intrusions. In addition, both the radial and concentric patterns observed in giant dyke swarms (e.g. Ernst et al., 1995, 2001, 2003; Buchan and Ernst, 2019) can be interpreted in terms membrane stresses related to the effects of topography (e.g. Turcotte et al., 1981). However, the specifics of such a relationship with respect to tessera ribbons on Venus are beyond the scope of this contribution.

Finally, if tessera ribbon fabrics are indeed related to underlying dyke swarms, mafic or otherwise, then the reported surface observations from Venus would allow that they could have been intruded at any time in the tessera terrain deformation sequence, including at or near the end. If valid, this could account for the absence of generalised, widespread folding of tessera ribbon fabrics, as well as reports of ribbons cross-cutting both folds and each other.

6. Conclusions

Models developed over the past quarter century to explain the structural geology of tessera terrains commonly spatially associated with crustal plateaus on Venus are unnecessarily complex and appear to be both internally contradictory and inconsistent with well-established geological principles. In great part, this appears to stem from attempts to integrate the development of high aspect ratio tessera ribbons with, at least partially, coeval fold development, within two further nested models of progressively thickening crustal plateaus and a mechanically homogeneous and isotropic crust. The principal stumbling block in understanding the structural geology of tessera terrains appears to be the assumption that ribbons are graben-like structures, generated by tectonic, far-field, extensional boundary conditions, without strain localisation. While acknowledging that the relationship of dykes and surface deformation, as well as the relative roles of regional and magmatic stresses in dyke emplacement and propagation are subject to debate, this contribution suggests that tessera ribbons may represent the surface expression of generally blind mafic dyke swarms, associated with widespread, volumetrically limited volcanic surface leakage. Dyke orientation would have been potentially guided by regional stress patterns, while emplacement and dilation, relatively late in the tessera structural sequence, would have been primarily driven by internal magmatic pressure. If valid, this necessarily speculative hypothesis would significantly reduce the complexities and inconsistencies inherent in the current narrative of tessera terrain structural geology. Until further observations can be derived, either from existing or new, higher resolution data, attempts to define a robust tectonic model for tessera terrains on Venus, including this contribution, will be model-driven and remain in the realm of speculation based on circumstantial evidence.

Declaration of Competing Interest

There are no competing interests related to this paper.

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