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Physics of the Earth and Planetary Interiors 146 (2004) 195–229

PHYSICS OF THE EARTH AND PLANETARY INTERIORS

www.elsevier.com/locate/pepi

Lessons from Venus for understanding mantle plumes on Earth

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Received 27 March 2003; accepted 10 October 2003

Abstract

Mantle plumes are important in the magmatic and tectonic history for both Earth and Venus. The expression of plumes is distinctive on Venus and complementary to that on Earth; therfore, a cross-comparison is useful for better understanding plume magmatism on both planets. In contrast to the Earth, Venus has no observed record of plate tectonics, a low degree of surface erosion, and an apparently short duration for the formation of the present planetary surface. The absence of plate tectonics indicates that all magmatism is ‘intraplate’ and is generated beneath a stagnant lithospheric lid. A low degree of surface erosion preserves the surface structures and short-wavelength topography. The short duration of preserved magmatic activity suggests a global resurfacing event.

Magmatic elements include: (a) individual volcanoes with diameters ranging up to 1000 km, which represent hotspots; (b) annular structures termed coronae with diameters averaging 300 km, but ranging up to 2600 km, and which appear to lack terrestrial (i.e. Earth) analogues; (c) radiating graben-fissure systems extending up to >2000 km in radius, some of which are purely uplift-related while others mark the plumbing system (dyke swarms) of volcanic systems; (d) lava flow fields of scale comparable to terrestrial flood basalts (large igneous provinces (LIPs)); and (e) regions of small shield volcanoes representing shallow-source melting.

There are several hierarchies of magmatic events on Venus, ordered in terms of increasing scale and significance: (1) isolated coronae, volcanoes, flow fields, and radiating graben systems; (2a) individual and small clusters of volcanoes and coronae associated with topographic swells, geoid highs, and triple-junction rifting; these are most clearly indicative of terrestrial-type plumes originating from the deep mantle; (2b) coronae distributed along rifts (chasmata); these are the clearest examples of melt generation associated with rifting; (3) regional concentration of activity in the Beta–Atla–Themis (BAT) region; this is the closest example of a plume cluster event, sometimes termed a ‘superplume event’; and (4) global volcanic resurfacing of the volcanic plains; no terrestrial analogue is confirmed, although the global burst of terrestrial plume activity in the Neoarchean is a possible analogue.

Keywords: Mantle plumes; Venus; Earth; Coronae; Rifts; Large igneous provinces;; Volcanoes

1. Introduction

On Earth, diapiric upwellings from the mantle (mantle plumes) are widely viewed as a cause of major magmatic events, the large igneous provinces (LIPs), as well as associated broad regional uplift and triple-junction rifts (Burke and Dewey, 1973; White and McKenzie, 1989; Campbell et al., 1989; Griffiths and Campbell, 1991; Hill, 1991; Maruyama, 1994; Campbell, 1998, 2001; Condie, 2001; Şengör, 2001; Şengör and Natal’in, 2001; Ernst and Buchan, 2001a,b, 2003). There is less consensus regarding the following: (1) the depth-of-origin of source areas, and

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doi:10.1016/j.pepi.2003.10.012

in particular the location of their distinct geochemical reservoirs (including the radical possibility that all the sources are in the upper mantle, e.g. Anderson, 1995; King and Anderson, 1998); (2) the nature of "superplumes" marking events of atypical significance (e.g. Condie, 2001; Ernst and Buchan, 2003); and (3) the integration of plume tectonics with plate tectonics (e.g. Maruyama, 1994). Some of these questions are being addressed by improving resolution of seismic tomography, increasing application of isotopic systems such as Os, Hf, and He, and the extension of the large igneous province record back beyond the well-studied Mesozoic/Cenozoic record into the Proterozoic and Archean (Ernst and Buchan, 2003).

Some of these frontiers in terrestrial plume research can also be addressed by a comparison with the record of plumes on Venus and Mars. In this review, we focus on the lessons from Venus applicable to improving our understanding of terrestrial plumes.

Venus is similar in size and bulk composition to the Earth and is inferred to have had a vigorous plume history (e.g. Head et al., 1992; Hansen et al., 1997; Nimmo and McKenzie, 1998). Yet there are important differences. One of the most significant is that Venus appears to lack plate tectonics, and therefore all magmatism on Venus would be classified as "intraplate". Thus, Venus allows us to explore the diversity of intraplate magmatic processes free of overprinting by plate tectonics; this should contribute to an increased understanding of terrestrial intraplate magmatism, including that generated by plumes rising from the deep mantle. However, because Venus may have a higher viscosity mantle owing to planetary loss of water (e.g. Arkani-Hamed, 1996), and because of the presence of a 'stagnant lid' (one-plate) (e.g. Solomatov and Moresi, 1996; Reese et al., 1999; Jellinek et al., 2002), intraplate magmatism on Venus may have different characteristics than that on the Earth.

Another dramatic difference on Venus is the lack of significant surface erosion (e.g. Campbell et al., 1997), which affords the opportunity to observe primary surface structures and the short-wavelength topography of intraplate magmatism, again complementing what we can observe on Earth.

In this paper, we first review the types and characteristics of terrestrial (i.e. Earth) intraplate magmatism, particularly those associated with mantle plumes. This is followed by a review of the important classes of volcanic/tectonic features on Venus, with a focus on their characteristics, global distribution, and postulated link with mantle plumes. Finally, we consider the lessons from Venus for understanding terrestrial plumes.

2. Intraplate magmatic settings on Earth and links with mantle plumes

2.1. Plume heads

Upon arrival at the base of the lithosphere, a mantle plume partially melts to form a large igneous province (e.g. Coffin and Eldholm, 1994; Campbell, 2001; Ernst and Buchan, 2003). A second pulse of magmatism can be generated by decompression melting associated with subsequent rifting (Campbell, 1998) (Table 1). Alternative models for the origin of LIPs include 'edge convection' between thick warm cratonic lithosphere and adjacent thin oceanic lithosphere (King and Anderson, 1998; Anderson, 1998), and bolide impact (Jones et al., 2002).

Other tools for the recognition of mantle plume heads on Earth are the presence of triple-junction rifting (Burke and Dewey, 1973; Şengör and Natal'in, 2001; Şengör, 1995), domical uplift recorded in the sedimentary rocks (Cox, 1989; Hill, 1991; Şengör, 2001; Rainbird and Ernst, 2001), positive gravity/geoid anomalies (e.g. Duncan and Richards, 1991), circumferential (concentric) contractional structures (Mège and Ernst, 2001) and radial extensional structures marking dyke swarms (e.g. Halls, 1982; Fahrig, 1987; Ernst et al., 2001). The distribution of possible plume head events through time on Earth can be inferred from the LIP record (Fig. 1) (Ernst and Buchan, 2001a). Plume-head magmatic events occur frequently, perhaps as often as once per 20 Myr for continental plumes (Ernst and Buchan, 2002) and once per 10 Myr for the combined record of continental and oceanic plumes (Coffin and Eldholm, 2001).

2.2. Plume tails

Some linear tracks of small volume magmatism show a systematic age progression along their length, and are termed hotspot tracks. They are most easily recognized in oceanic crust as lines of seamount

Table 1
Categories of intraplate magmatism and links with mantle plumes

Category of mantle feature	Continental crust examples	Oceanic crust examples
Individual plume		
Large igneous province (plume head)	*Mainly flood basalt*: e.g. Columbia River: 17 Ma; 0.16 Mkm^2	*Oceanic plateau* (& *ocean basin floods basalts*): e.g. Ontong Java: 122 Ma, minor pulse at 90 Ma; 44 Mkm^3 (Coffin and Eldholm, 1994, 2001; Arndt and Weis, 2002)
	Mainly flood basalts and sills: e.g. Siberia Traps: 250 Ma; 3 Mkm^3 (Reichow et al., 2002)	*Accreted Oceanic Plateau*: e.g. Wrangellia: 232 Ma, 1 Mkm^3 (Lassiter et al., 1995; Hulbert, 1997)
	Mainly giant radiating dyke swarm: e.g. Central Atlantic Magmatic Province: 200 Ma; up to 3000 km radius (May, 1971; Ernst et al., 1995; Hames et al., 2000, 2003); e.g. Mackenzie swarm: 1270 Ma; up to 2500 km radius (Baragar et al., 1996) Selected *Archean greenstone belts*: e.g. in Slave Province: ∼2700 Ma (Bleeker, 2002); in Abitibi belt: ∼2700 Ma (Sproule et al., 2002)	
Small-volume intraplate magmatism = hotspots (small plumes, diapirs?)	*In Africa*: e.g. Ahaggar, Ankaratra, Jebel Sawda, Tibesti (Burke, 1996; Şengör, 2001) *In Europe*: e.g. Massif Central, Rhenish Massif, Bohemian Massif (Wilson and Patterson, 2001) *In eastern Australia*: N. Queensland, Newer Volcanic Province, may also include hotspot tracks (e.g. Şengör, 2001)	Widespread Pacific seamounts (e.g. Shen et al., 1993)
Plume cluster		
Regional clusters of coeval plumes	e.g. Karoo–Ferrar–Chon Aike event (three or four plume centers) of Africa, Antarctica and South America: 183 Ma (Storey et al., 2001; Elliot and Fleming, 2000; Ernst and Buchan, 2002)	e.g. Ontong Java and Manihiki plateaus (two plume centers): 122 Ma (Coffin and Eldholm, 1994; Larson, 1997; Ernst and Buchan, 2002)
Regional clusters of plumes with age range >10 to <100 Myr	*Associated with current geoid high under Africa*: Gondwana supercontinent breakup: Karoo (180 Ma), Antarctica–India (145 Ma), Parana-Etendeka (135 Ma) (Storey, 1995)	*Associated with current geoid high under Pacific*: Pacific “superplume” event: 90–122 Ma (possibly starting at 145 Ma) (Larson, 1991, 1997; Maruyama, 1994; Ernst and Buchan, 2002)
	Neoproterozoic example: e.g. Iapetus Margin of North America: 565–615 Ma (events 52–54 in Ernst and Buchan, 2001a; Puffer, 2002) *Paleoproterozoic example*: e.g. Superior Province margin, ∼2500 Ma Matachewan and Mistassini plume centers (Ernst et al., 1995; Ernst and Buchan, 2002)	
Linear distribution		
Association with rifting/breakup	Passive margin sequences and seaward dipping reflectors (e.g. Menzies et al., 2002)	e.g. Nova Canton Trough (Larson, 1997)
Not associated with rift	*Age progression = hotspot track* (plume tail): e.g. Snake River Plain associated with Columbia River Province: 5–17 Ma (Parsons et al., 1998); however, note complications discussed in Christiansen et al. (2002)	*Hotspot track (plume tail)*: e.g. Hawaiian-Emperor seamount track (Clague and Dalrymple, 1989; Steinberger and O’Connell, 1998); e.g. Walvis seamount track associated with 135 Ma Parana-Etendeka plume (O’Connor and Duncan, 1990; Peate, 1997)

Table 1 (*Continued*)

Category of mantle feature	Continental crust examples	Oceanic crust examples
	No age progression: e.g. hot-line (Meyers et al., 1998); or edge convection model (King and Anderson, 1998)	
Global event		
Widely distributed plume centers	e.g. 62–65 Ma: Deccan and North Atlantic Igneous Province events (Ernst and Buchan, 2002) e.g. 90 Ma: Ontong Java (second pulse), Madagascar, Alpha Ridge/Queen Elizabeth Islands, Caribbean/Colombia (Ernst and Buchan, 2002) e.g. ∼2700 Ma: Archean greenstone belts with plume association (found in many regions of the world)—possibly mantle overturn events (e.g. Condie, 1998, 2001)	e.g. 118–122. Ontong Java, Kerguelen (e.g. Arndt and Weis, 2002; Coffin and Eldholm, 1994)

Further details, including alternative non-plume origins are discussed in the text.

chains (e.g. Clague and Dalrymple, 1989), which can often be backtracked to plume-head magmatic events marked by LIPs. These are typically interpreted to represent a continued flux of material from a plume tail (that follows a plume head). However, in oceanic areas, there are also widely distributed seamounts that cannot be grouped into linear tracks and may derive from 'mini hotspots' originating in the upper mantle (e.g. Shen et al., 1993; Davis et al., 1995; Courtillot et al., 2003). Hotspot tracks are more difficult to recognize in continental crust. One example is the migrating activity along the Snake River plain in northwestern USA (e.g. Parsons et al., 1998; cf. different interpretation by Christiansen et al., 2002). Other types of linear magmatic belts are discussed in Section 2.5.

2.3. Small-volume mafic magmatic events

Small-volume intraplate basaltic magmatic events (distributed over areas 100–500 km in diameter) occur in Europe, Africa, and elsewhere (e.g. Wilson and Patterson, 2001; Burke, 1996; Şengör, 2001; Ernst and Buchan, 2001a), and are more difficult to assess in a plume context. Models for their origin include (a) mini-plumes originating from the 660 km boundary (Wilson and Patterson, 2001; Burke, 1996); (b) sublithospheric channeling from adjacent more robust plumes (such as the Afar plume feeding the western and north African magmatic centers (e.g. Ebinger and Sleep, 1998); (c) plumes originating from the deep mantle that 'fragment' into smaller plumes on crossing the 660 km boundary (e.g. Arndt, 2000; Courtillot et al., 2003); (d) slab breakoff during continental collision causing diapiric instabilities (e.g. Wilson and Patterson, 2001), and (e) lithospheric delamination above small plumes (e.g. Moore et al., 1999; Şengör, 2001).

2.4. Plume clusters (superplume events)

Plumes head events can be clustered in both time and space (Ernst and Buchan, 2002). For example, the Karoo–Ferrar–Chon Aike event at 183 Ma, which is widespread in southern Africa, Antarctica, and southern South America, probably consists of three or four different plume centers of upwelling and magmatism (Storey et al., 2001; Elliot and Fleming, 2000; Ernst and Buchan, 2002).

It has been suggested that regional clusters of plumes spanning up to 100 Myr are also significant and these have been termed superplume events (Condie, 2001), or plume clusters (Ernst and Buchan, 2002). On the present Earth there are two clusters of hotspots which can be linked with two persistent geoid highs on the Earth, one under the Pacific and the other under Africa (e.g. Crough and Jurdy, 1980; Duncan and Richards, 1991). The geoid highs are caused by regions of lower mantle upwelling (e.g. Maruyama, 1994; Pavoni, 1997; Davies, 1999; Courtillot et al., 2003).

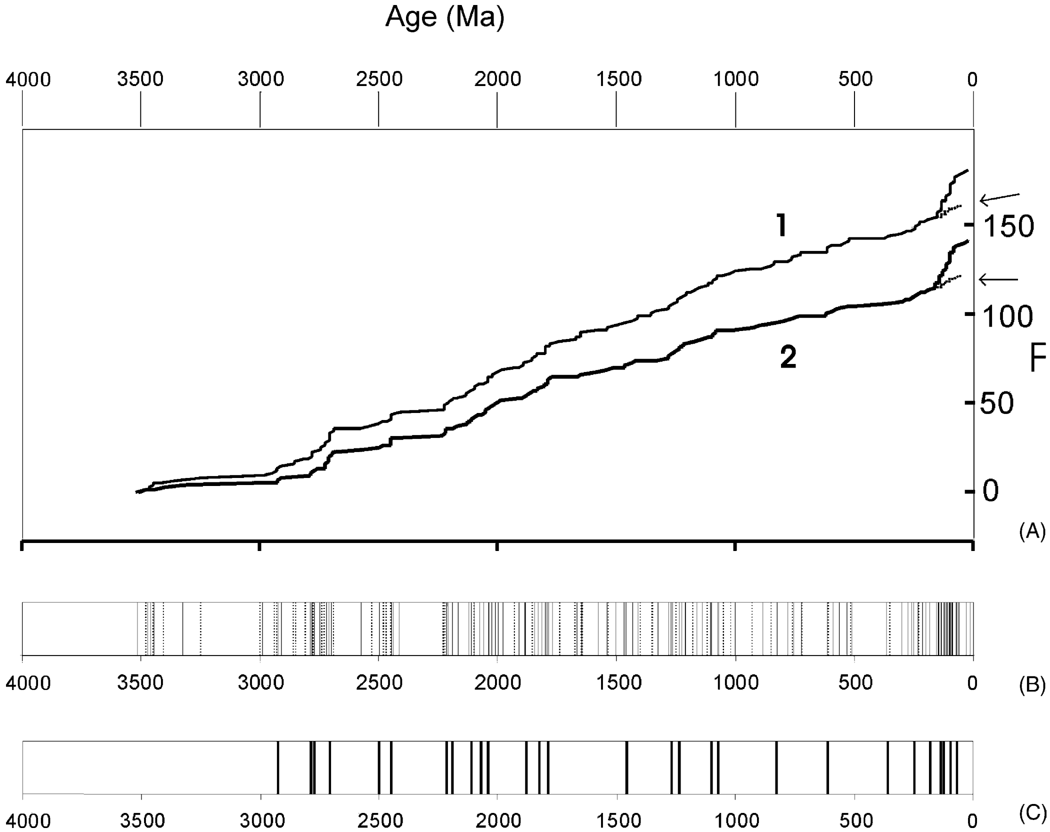


Fig. 1. Distribution of terrestrial plume events in time. (A) Cumulative frequency curves (after Fig. 4 in Ernst and Buchan, 2002) for large mafic magmatic events, and include data rated both "A" and "B" in terms of the reliability of a link with a mantle plume. "A" and "B", indicate a 'strong', and 'probable link, respectively. Rating criteria are defined in Ernst and Buchan (2001a, 2002). Curve 1 includes all data with a 2σ age uncertainty $\leq$50 Myr, while curve 2 includes all data with an age uncertainty $\leq$20 Myr. The steeper the curve the more frequent the events; no events occur where the curve is horizontal. The dotted curves between 150 Ma and the Present (marked by arrows) include only data from continental LIPs. (B) Bar diagram of the data (modified after Ernst and Buchan (2001a, 2002). Events with age uncertainty $\leq$20 Myr are indicated with solid lines whereas those with 20–50 Myr uncertainty are located with a dashed line. (C) Potential plume clusters modified after Ernst and Buchan (2001a, 2002) to include only events rated "A" and "B".

Other examples of plume clusters (superplume events) are recognized in the older geological record. For example, a set of plume centers associated with the progressive breakup of Laurentia from other continents to form the Iapetus Ocean. Plume pulses on the Laurentian margin, can be distinguished at 615, 590, and 563 Ma (events 52–54 in Ernst and Buchan, 2001a; Puffer, 2002). However, uncertain plate reconstructions impede the recognition of such long-duration regionally clustered plumes in the older record. It has also been suggested that the Neoproterozoic supercontinent Rodinia was fragmented by multiple plumes (Li et al., 2003).

2.5. Linear belts of magmatic activity

There are several types of linear belts of mafic magmatism. One class of linear belt is aligned seamounts chains which show a systematic age progression along their length, and are interpreted as hotspot tracks (see earlier discussion in Section 2.2).

Other linear belts of magmatism lack a consistent age progression such as the Cameroon line of west Africa (Marzoli et al., 2000) and may require a different interpretation. Candidate models include the 'hot-line' model in which a line of melting is generated above Rayleigh–Bernard convection rolls in the upper mantle (Meyers et al., 1998). Another non-plume model, termed "edge convection" features convection between hotter and thicker cratonic lithosphere and adjacent thin lithosphere (typically oceanic crust) (e.g. King and Anderson, 1998). Linear belts of "intraplate" magmatism can also been postulated to arise in a back arc setting (e.g. Rivers and Corrigan, 2000) or as the result of overriding of a spreading ridge by the continent (Gower and Krogh, 2002).

2.6. Global burst of activity

Although the terrestrial plume record is fairly continuous (Fig. 1), there are times when global-scale bursts of plume activity may occur. The archetypal 'superplume' event occurred in the mid-Cretaceous, and is marked by the development of huge oceanic plateaus in the Pacific and elsewhere in the period about 80–120 Ma, as well an association with a regional geoid high and increased global rates of extension at spreading ridges (Larson, 1991, 1997; Maruyama, 1994; Condie, 2001). Recent dating demonstrates a nearly identical chronology of LIP activity at Ontong Java in the Pacific and Kerguelen in the Indian Ocean, and provides further support for a global event at about 120 Ma (Arndt and Weis, 2002). The burst of late Archean greenstone magmatism related to plumes has also been cited as evidence of a global magmatic event (e.g. Condie, 2001). However, in the absence of reliable paleocontinental reconstructions for the Archean, it remains uncertain whether these represent a regional or global event. Furthermore, it should be recognized that any of these examples could represent a statistical coincidence; the LIP-plume spectrum on Earth is very dense (Fig. 1) and therefore, coeval but unrelated events are likely to occur frequently.

2.7. Summary

Intraplate magmatism on Earth occurs at a range of scales from individual seamounts and small volcanic centers up to large-scale (Mkm^3) short duration (<10 Myr, and often <1 Myr) events termed large igneous provinces (LIPs). LIPs have been typically linked to mantle plume head events. A second burst of magmatism can be linked with rift-related decompression melting. Smaller magmatic events can be explained by a variety of mechanisms, including shallow-sourced plumes and sublithospheric melting. Linear distributions of events showing an age progression can be linked to plume tails (hotspot tracks) while linear distributions lacking an age progression can have other origins. At the large end of the scale, clusters of plumes are recognized, and there is the possibility that at selected periods in Earth history there were global resurfacing events. The inventory of terrestrial intraplate processes (summarized in Table 1) provides a starting point for understanding magmatism on Venus, where plate tectonics is absent and therefore all magmatism is by definition "intraplate".

3. Volcanic/tectonic features on Venus

The Magellan mission to Venus in the early 1990s provided unprecedented global coverage of Venus at high resolution (~120 m for 98% of the surface) using synthetic aperture radar (SAR). In this paper, we discuss two types of Magellan SAR radar map products: FMAPs (full resolution maps) at 75 m per pixel, and C1-MIDRs (compressed once, mosaicked image data records) at 225 m per pixel. Additional Magellan datasets include radar altimetry to a resolution of about 50 m over a cross-track footprint of 10 km, and radiothermal imaging (for emissivity data). The initial results (Magellan Science Team, 1992) and the ongoing mapping and research efforts are gradually revealing the magmatic and tectonic history of Venus (Table 2). The most significant and enduring conclusions are: resurfacing of the entire planet in a short period, lack of surface erosion, absence of a record of plate tectonics, and importance of mantle plumes and diapirs (e.g. Head et al., 1992; Price and Suppe, 1994, 1995; Hansen et al., 1997; Head and Coffin, 1997; McKinnon et al., 1997; Nimmo and McKenzie, 1998; Basilevsky and Head, 2002).

3.1. Duration of magmatic-driven activity in a stagnant-lid lithospheric regime

The entire planet has been generally inferred to have been resurfaced between about 500 and 300 Myr ago (Phillips et al., 1992; Schaber et al., 1992; Strom et al., 1994; Herrick, 1994; Herrick et al., 1995; Price et al., 1996; Head and Coffin, 1997) in an interval perhaps as

Table 2
Classes of magmatic/tectonic features on Venus

Feature type	Number of features	Size of features	Distribution	Possible tectono-magmatic setting
Volcanoes				
Large volcanoes (Fig. 5) (>100 km in diameter)	168 [1]	Defined as ≥100 km diameter; maximum 1000 km diameter; 60, ≥500 km diameter [1]	Located on volcanic rises, along chasmata and in the plains. Sometimes clustered, sometimes as isolated features	Major plumes/hotspots
Intermediate volcanoes (Fig. 6) (20–100 km in diameter)	289 [1]	Defined as <100, ≥20 km in diameter [1]		Plumes/hotspots?
Shield fields (Fig. 7)	647 [1]	Mean: 50–350 km; mode: 100–150 km; maximum: 1200 × 300 km [1]	Widely distributed within a global stratigraphic unit [6]. Regionally clustered and diachronous on a global scale in part postdating regional plains volcanism [7]; local clusters associated with individual coronae (e.g. [1])	Shallow mantle sources
Circular structures				
Coronae (Figs. 8 and 9)	473 (Including 265 arachnoid subclass) [1]	*Arachnoid subclass*: most: <200 km diameter; maximum: 280 km diameter [1]	Distributed along rifts (chasmata), clustered in volcanic rises, or occurring as isolated features in the plains	Transient plumes (thermals) or diapirs or subsurface intrusions
	513 Type 1 and 2 (‘stealth’ coronae) [1a]	*Other coronae*: most: 200–250 km diameter; maximum: 1000 km diameter; 18, ≥500 km diameter [1]		
	96 [1b]	Note Artemis (with 2600 km diameter) not included in [1]		
Calderae (Fig. 10)	97 [1]	Most: 60–80 km diameter; maximum: 225 × 150 km diameter [1]		Subsurface intrusions
Flow fields (Fig. 11)	208; Including 81 giant flow fields [4]	Defined: >50,000 km^2; maximum: 1,630,000 km^2; 140 to >100,000 km^2; 81 giant flow fields = flow length >500 km [4]	Associated with both volcanoes and coronae	=Large igneous provinces (LIPs)
Radiating graben-fissure systems (Figs. 12–14)			Associated with volcanoes, arachnoids and coronae	Largest are giant radiating dyke swarms derived from mantle plumes
Global reconnaissance survey at C1-MIDR scale (Fig. 12)	163 in global C1-MIDR study [2]	Range: 40 to >2000 km radius; mean: 325 km radius [2]		
Detailed local study at FMAP scale (Fig. 13)	34 in regional FMAP study [3]	15 have ≥300 km radius [3]; 8 have ≥1000 km radius [3]	Area of detailed study bounded by 264–312°E and 24–60°N	

Table 2 (*Continued*)

Feature type	Number of features	Size of features	Distribution	Possible tectono-magmatic setting
Radial fracture centers also termed novae (Fig. 14)	64 [1]	Range: 50–300 km diameter [1]		
Arachnoids: subset having radiating graben (Fig. 9)	Minor portion of arachnoid population [1b]			
Radiating ridge systems				
Arachnoid: subset with radiating ridges (Fig. 9)	Majority of arachnoid population [1b]			??
Crustal plateaus (tesserae)				
Regional uplifts of 'basement' tesserae terrain [5] (Fig. 15)	7	1600–2500 km in diameter		Upwelling or downwelling
Groupings of events				
Clusters of volcanic features (Table 3, and Fig. 3): volcano, corona-, and rift-dominated types)	9		Usually associated with topographic and geoid highs	Plume cluster (superplume event)
Linear distributions of coronae (Table 4 and Fig. 4)			Mainly along Parga Chasma and Hecate Chasma	Rift-related

[1] Crumpler and Aubele (2000); [1a] Stofan et al. (2001), Glaze et al. (2002); [1b] Kostama and Aittola (2001), Aittola and Kostama (2002); [2] Grosfils and Head (1994), Grosfils (1996); [3] Ernst et al. (2003); [4] Magee and Head (2001); [5] Hansen et al. (1999); [6] Basilevsky and Head (1998, 2002); [7] Addington (2001).

short as 10 Myr (Strom et al., 1994). However, a recent reassessment by McKinnon et al. (1997) and review by Basilevsky and Head (2002) suggests that the duration could be much longer. Specifically, the preserved surface history ranges from $(1.47 \pm 0.46)T$ for the mean age of tesserae units, to the youngest rifts $(0.27 \pm 0.39)T$, where T is the mean surface age (Basilevsky and Head, 2002). The new value for T is ca. 750 Ma, but it could be as young as 300 Myr or as old as 1 Ga. (McKinnon et al., 1997). Taking the largest possible value for T (1 Ga), and the ages (including uncertainties) for the oldest and youngest stratigraphic units, the preserved history could span from nearly 2000 Ma ($1.93T$) to present. Using the more realistic mean value for T (750 Ma) but the same extreme values for the age of tesserae and young rifts, the preserved history could span from 1450 Ma to present. Using mean instead of extreme values, the period spanning tesserae formation to young rifting would range from 1100 Ma ($1.47T$) to 200 Ma ($0.27T$) (for T = 750 Ma). These values are significantly larger than previous estimates of 300–500 Myr for mean resurfacing (see discussion in McKinnon et al., 1997), and thus, the plains resurfacing on Venus may not be as catastrophic as traditionally viewed. Nevertheless, the magmatic history is still dramatic by terrestrial standards (see Section 6.1).

The most comprehensive catalogue of volcanic and tectonic features on Venus is that of Crumpler and Aubele (2000). They catalogue 1738 volcanic centers, coronae (raised annular structures) and related features larger than 20 km in diameter, as well as significant lava flow fields and lava channels. Other catalogues include the radiating graben system database of Grosfils (Grosfils, 1996; Grosfils and Head, 1994), the corona database of Stofan et al. (2001, with modifications cited in Glaze et al., 2002), the corona, arachnoid and novae databases of Kostama and Aittola (2001), and the digital tectonic and magmatic features map for Venus (Price, 1995; Price and Suppe,

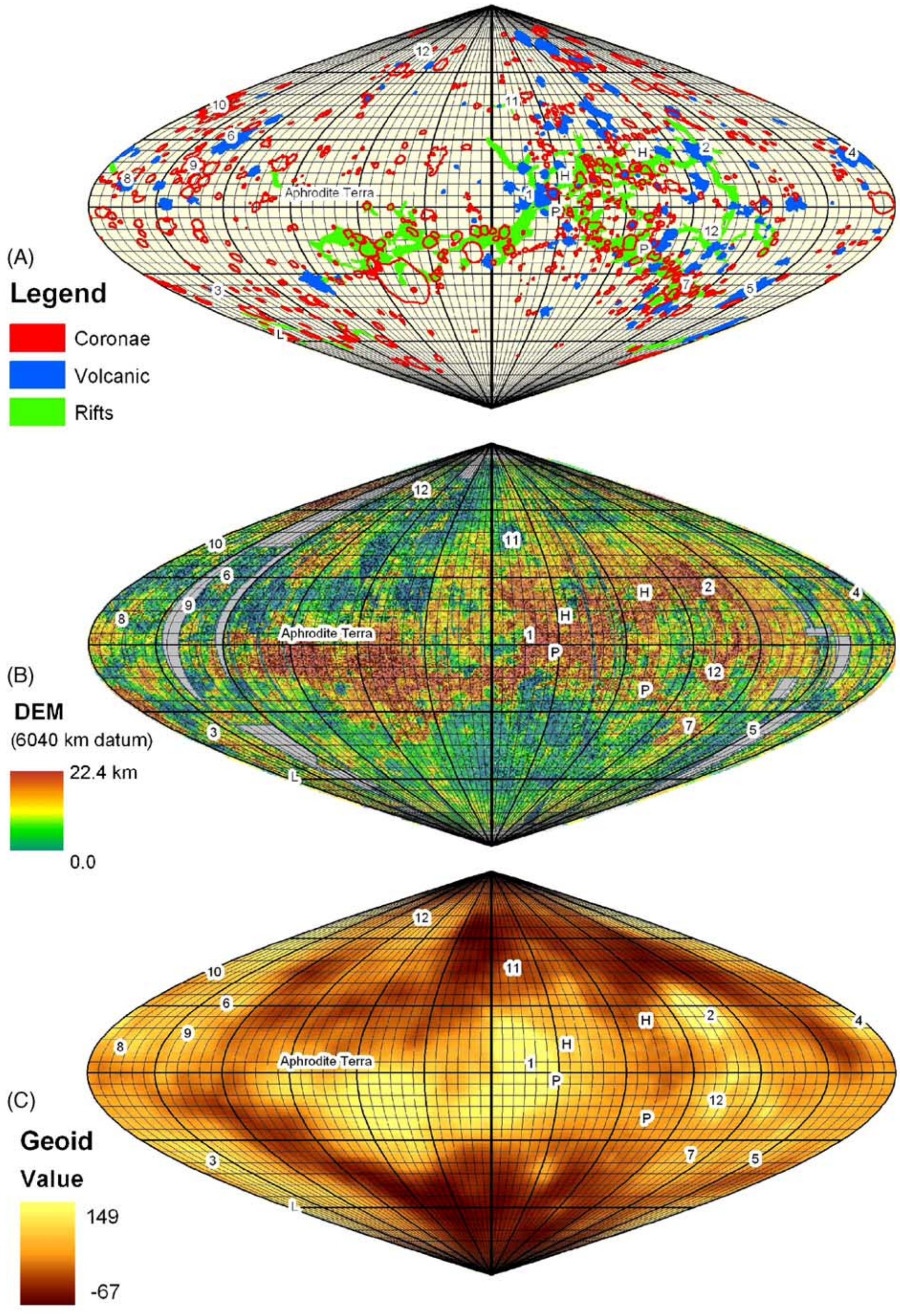


Fig. 2. (A) Global distribution of large volcanoes, coronae and young rifts generated from database of Price (1995). Young rifts (chasmata) were distinguished from old rifts (i.e. fracture belts) based on Basilevsky and Head (2000). Numbers locate volcanism associated with geoid or topographic rises and correlate with numbers in Table 3. Hecate Chasma (H) links Atla Regio (1) and Beta Regio (2), and Parga Chasma (P) links Atla Regio (1) and Themis (7). L is Lada Terra. Displayed in a sinusoidal projection with a central meridian of 180°, and with the edge at 360/0°. Dotted latitude and longitude lines are spaced at 5° intervals. Thick lines are at 30° intervals. (B) Global topography. (C) Global geoid after Herrick (1999).

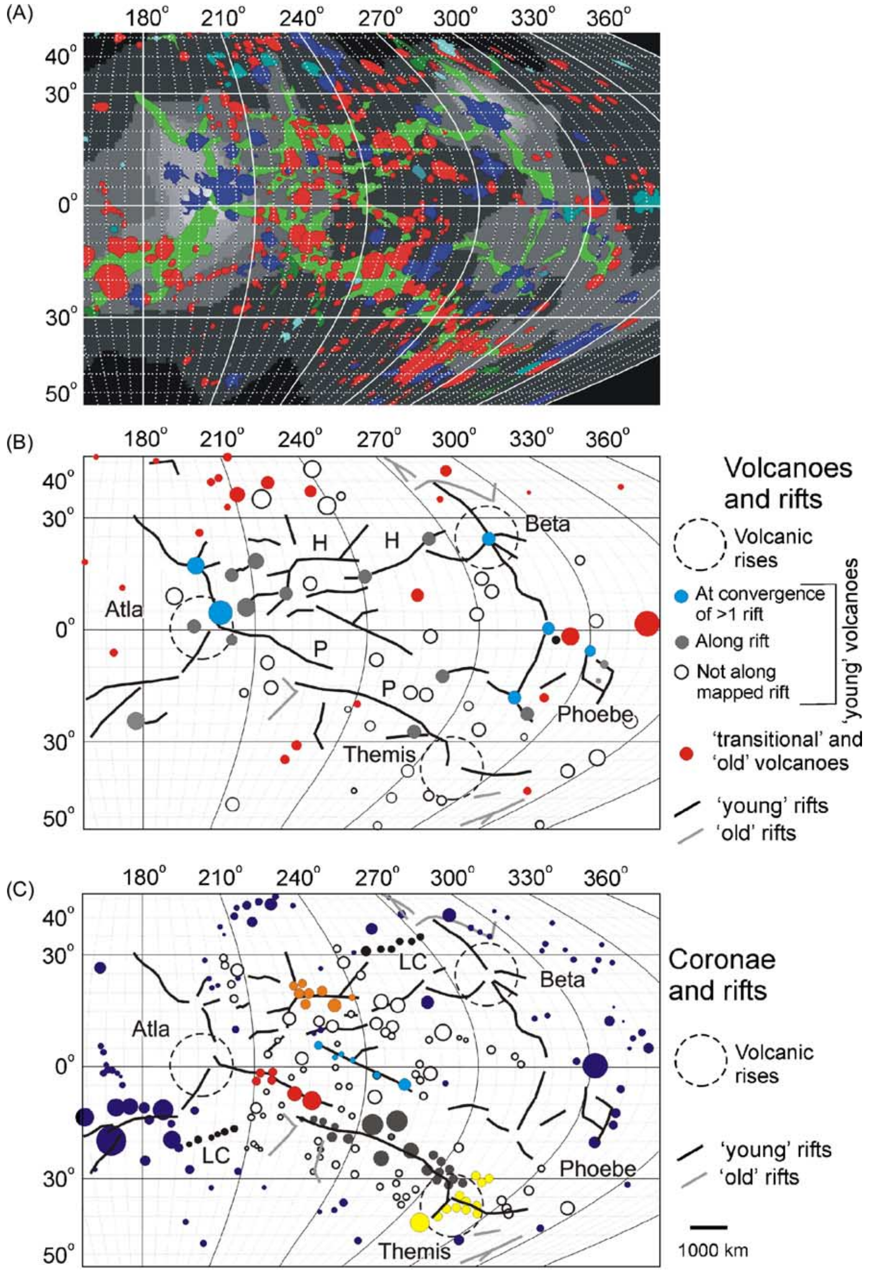


Fig. 3. Distribution of volcanoes, coronae (including arachnoids) and young rifts in the BAT (Beta–Atla–Themis region). (A) Mapped distribution of geological features superimposed on geoid, from Herrick (1999) which is after Price (1995) and Price and Suppe (1995). Coronae are red, rifts are green, and volcanoes are blue (those in dark blue are "young" while those in light blue are "transitional" to "old" in age, based on Basilevsky and Head, 2000). (B) Generalized distribution of volcanoes. (C) Generalized distribution of coronae and rifts. Regional rises located with large circles. Separation of rifts into "young" and "old", and separation of volcanoes into "young" and "transitional/old" is based on Plate 1 in Basilevsky and Head (2000). Symbol size = approximately the minimum dimension of the feature. Sinusoidal projection with a central meridian of 180°. Thin latitude and longitude lines are spaced at 5° intervals. Thick lines are at 30° intervals. LC = lines of coronae not associated with a mapped rift. H is Hecate Chasma and P is Parga Chasma.

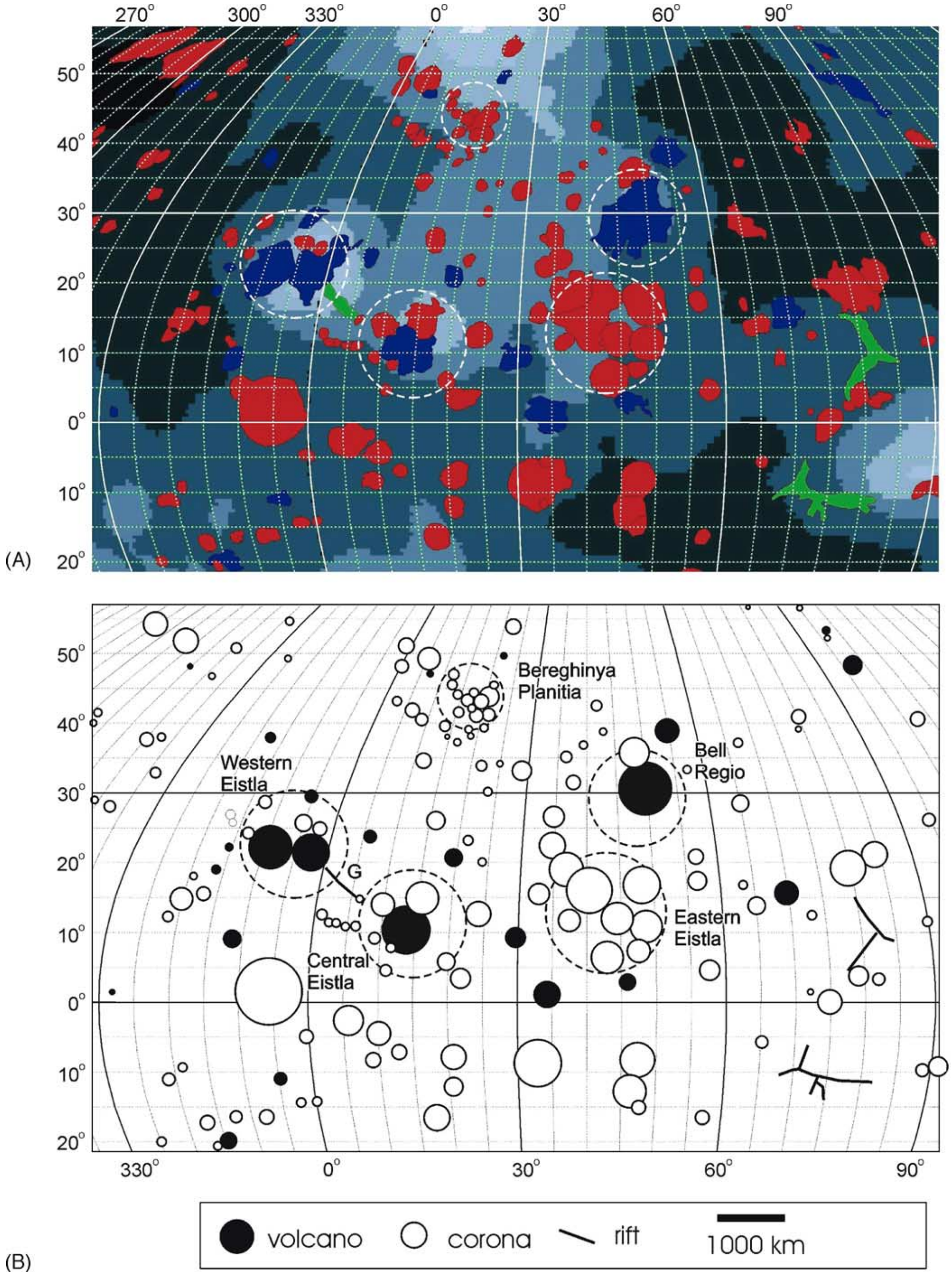


Fig. 4. Distribution of volcanoes, coronae (including arachnoids) and young rifts in the Eistla region. (A) Mapped distribution of features from Herrick (1999) which is after Price (1995) and Price and Suppe (1995). Coronae are red, rifts are green, and volcanoes are blue. (B) Generalized geology. Symbol size = approximately the minimum dimension of the feature. Sinusoidal projection with a central meridian of 40°. Thin latitude and longitude lines are spaced at 5° intervals. Thick lines are at 30° intervals.

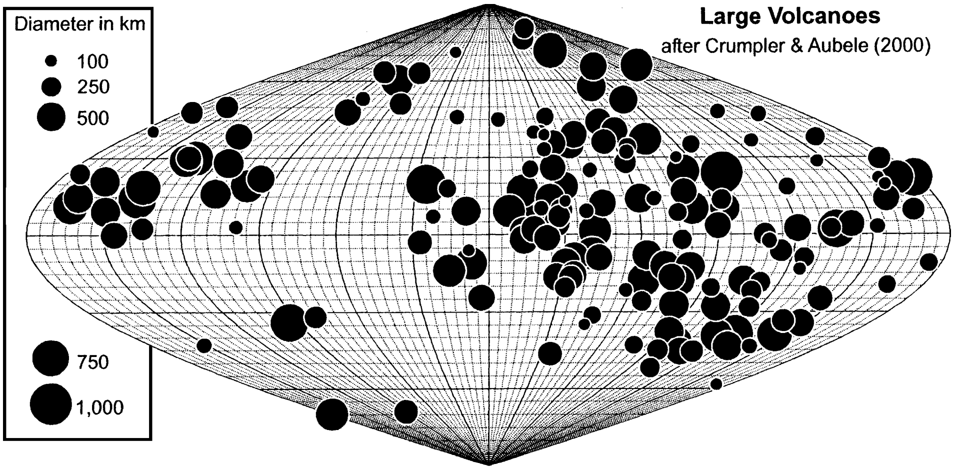


Fig. 5. Distribution of large volcanoes on Venus. Displayed in a sinusoidal projection with a central meridian of 180°. Extracted from catalogue of Crumpler and Aubele (2000). Symbol size is greater than the actual feature size; this was done to allow a greater dynamic range in symbol size, so as to better display the range in feature size.

1995). A summary of the geology, topography and geoid of Venus is provided in Figs. 2–4.

We review each of the classes of features on Venus in order to develop a comparison with the terrestrial features in a later section.

3.2. Volcanoes

On Venus, volcanoes are sub-divided based on size. There are 168 large volcanoes (Fig. 5) with diameters >100 km, and 289 intermediate volcanoes (Fig. 6) with diameters ranging from 20 to 100 km. Small (<20 km diameter) volcanoes are abundant on Venus and are grouped into "shield fields" (Fig. 7) (Crumpler and Aubele, 2000).

Large volcanoes are defined as eruptive centers with diameters >100 km (Crumpler et al., 1997, p. 703). Some centers also have concentric features and could be classed as coronae, but are grouped with volcanoes based on the presence of radially dispersed lava flows

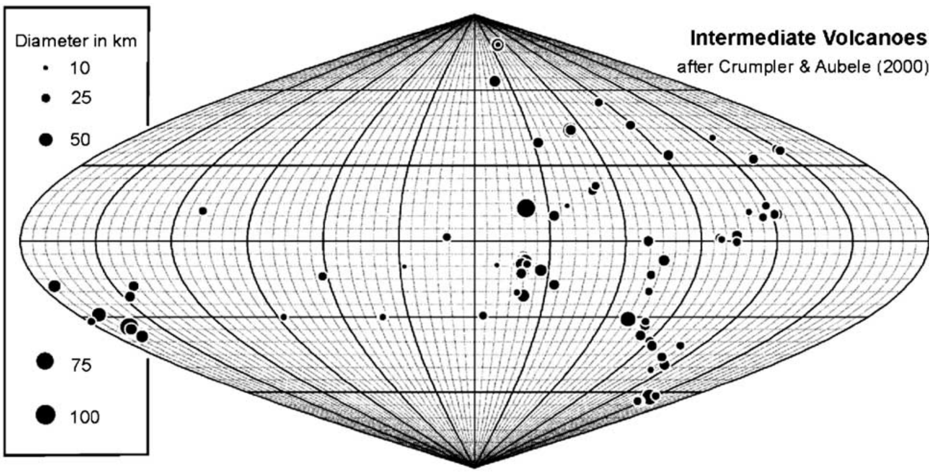


Fig. 6. Distribution of intermediate-size volcanoes on Venus. Displayed in a sinusoidal projection with a central meridian of 180°. Extracted from catalogue of Crumpler and Aubele (2000). As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

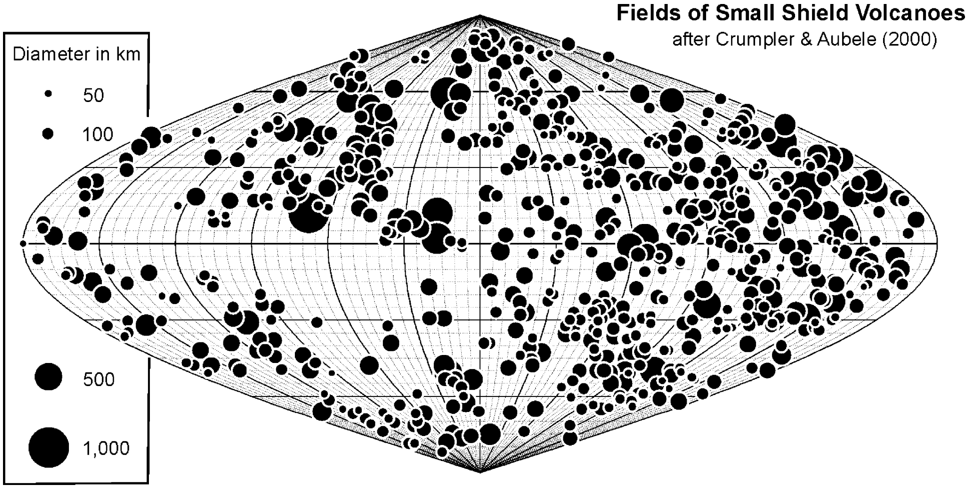


Fig. 7. Distribution of fields of small shield volcanoes on Venus. Displayed in a sinusoidal projection with a central meridian of 180°. Extracted from catalogue of Crumpler and Aubele (2000). As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

(Crumpler and Aubele, 2000). Most large volcanoes on Venus are relatively low in relief, averaging approximately 1.5 km in height and are considered to represent basaltic shield volcanoes. They have wide distribution on Venus (Fig. 5), but do show a preferred association with broad rises and at the junctions of extensive chasmata (deep linear valleys with steep sides generally interpreted as young rifts) and belts of fractures (interpreted as older rifts; Basilevsky and Head, 2000).

Intermediate volcanoes are of several types, including shield volcanoes of presumed basaltic composition (Crumpler and Aubele, 2000). There are also steep-sided morphologies that may represent more felsic compositions. Specifically, "steep-sided domes" and "fluted domes" may be analogues to terrestrial silicic domes. In terms of distribution, many intermediate volcanoes occur in association with coronae (Section 3.4.1).

The largest volcanoes range from more than 500 to over 1000 km in diameter (Crumpler and Aubele, 2000; Stofan et al., 2002). According to Stofan et al. (2002), volcanoes on Venus are larger than on Earth because of the lack of plate motion, and the "volumes of material produced at Venusian hotspots are similar to the overall volcanic volumes produced at terrestrial hotspots" (see further discussion in Section 7.4). However, volcanic edifices on Venus are also broad because of the higher surface temperatures, which allow volcanic flows to spread more widely from central sources. Some large volcanoes are associated with topographic rises, geoid highs, and triple-junction rifting; these are considered derived from a deep source mantle plumes (e.g. Senske et al., 1992). For other large volcanoes lacking these associations, the link with plumes is more indirect (see further discussion in Section 4.2).

3.3. Shield fields

Numerous small volcanoes (<20 km in diameter) have not been individually catalogued, but appear to be widespread and tend to cluster in fields. Six hundred and forty-seven shield fields have been identified (Fig. 7) with diameters ranging from 50 to 350 km, with a mode from 100 to 150 km. In addition to shield-type volcanoes, the shield fields also contain "steep side domes" and "fluted domes".

On the basis of analysis of 179 clusters in 7 quadrangles, Addington (2001) found that 69.8% were spatially associated with one or more volcanic or tectonic features, most commonly coronae (19%), fractures (16.2%), and ridge belts (13.4%). The timing of shield field development has been a matter of some debate. According to one school, the timing of shield field development has been interpreted to have time-stratigraphic significance (e.g. Basilevsky and Head, 1998, 2002), particularly based on the consistency of stratigraphic relations on a detailed geotraverse extending around the planet between latitudes 22.5 and 37.5°N (Ivanov and Head, 2001). This time of widespread shield field development has been variably linked to "among the oldest plains preserved" (Crumpler and Aubele, 2000) or to a relatively late stage in plains development (e.g. Basilevsky and Head, 2002). The other school of thought is that the evolution of Venus is non-directional and that different types of units can form at any time, apart from the tesserae which are widely agreed to have formed early (e.g. Guest and Stofan, 1999). Addington (2001) found that out of the total 179 clusters studied, the relation of 95 among regional plains volcanism could be determined; 41.8% of these (21.8% of the total) postdate the emplacement of regional plains materials. In the Metis Region (255°E, 72°N), the shield fields occur late in the stratigraphic sequence similar to the large shield volcanoes (Dohm, 2003, personal communication). In contrast, detailed mapping by Ivanov and Head (2004) suggests that most shield fields formed prior to regional plains volcanism.

In a terrestrial context, shield fields could be analogous with the widespread seamount distribution (Shen et al., 1993) associated with relatively thin oceanic crust (cf. discussion of seamount connection in Crumpler and Aubele, 2000). In such a case, they would not be linked with plumes, but would result from presumed widespread sublithospheric melting, and may mark regions in which the lithosphere was anomalously thinned. Hansen (2003) suggested that widespread shield fields mark point source partial melting of a hydrated (metamorphosed) deep crust. Those shield fields that are spatially associated with coronae probably have a genetic link with the coronae (see Section 3.4.1).

3.4. Circular features

There are various classes of circular volcanic/tectonic features greater than 20 km in diameter on Venus: These include coronae, arachnoids and calderae (Figs. 8–10).

3.4.1. Coronae

Coronae (type 1) are defined by the presence of a raised annulus of circumferential fractures or ridges, a peripheral moat or trough, and an interior region that can be either topographically positive or negative (Stofan et al., 1992, 1997; Janes et al., 1992; Squyres et al., 1992; Crumpler and Aubele, 2000). Coronae have been divided into nine types on the basis of morphology (Stofan et al., 1997, 2001). Asymmetric and 'multiple' types may originate via emplacement of multiple diapirs, movement of the lithosphere over a stationary diapir, or migration of a diapir under stationary lithosphere (e.g. sub-crustal channeling) (López, 2002).

More recently, an additional class of coronae (type 2, or 'stealth' coronae) has been identified. These have the raised annular topographic signature, but largely lack the fractured annuli that are characteristic of classic type 1 coronae (Stofan et al., 2001, 2003). Statistical analysis indicates that type 1 and 2 coronae have the same size distributions and can be interpreted to have similar formation mechanisms (Glaze et al., 2002). Coronae range in diameter from 60 to over 2000 km, but are generally between 200 and 250 km in diameter (Fig. 8).

Arachnoids (Fig. 9) are an important subclass of coronae, having the same annular structure, but exhibiting strong radial lineaments as well, which can be either extensional (radiating graben-fissure systems) or, more commonly, compressional (radiating ridge systems) (e.g. Kostama and Aittola, 2001; Aittola and Kostama, 2002). Arachnoids are generally smaller than coronae; most have a diameter less than 200 km. The origin of the radial extensional type of arachnoid is discussed in Section 3.6. The formation of the subtype having radial ridge structures is more difficult to understand (Section 3.7).

In the Crumpler and Aubele (2000) catalogue, there are 473 coronae of which 265 belong to the arachnoid subclass (Figs. 8 and 9). Kostama and Aittola (2001) and Aittola and Kostama (2002) recognize 446 coronae of which 96 belong to the arachnoid subclass (Figs. 8 and 9). A recent compilation (Stofan et al., 2001, as slightly modified in Glaze et al., 2002) identifies 513 coronae of which 107 are the newly identified 'stealth' coronae. The different tallies reflect variation in criteria used by various authors and an overall progressive improvement in mapping.

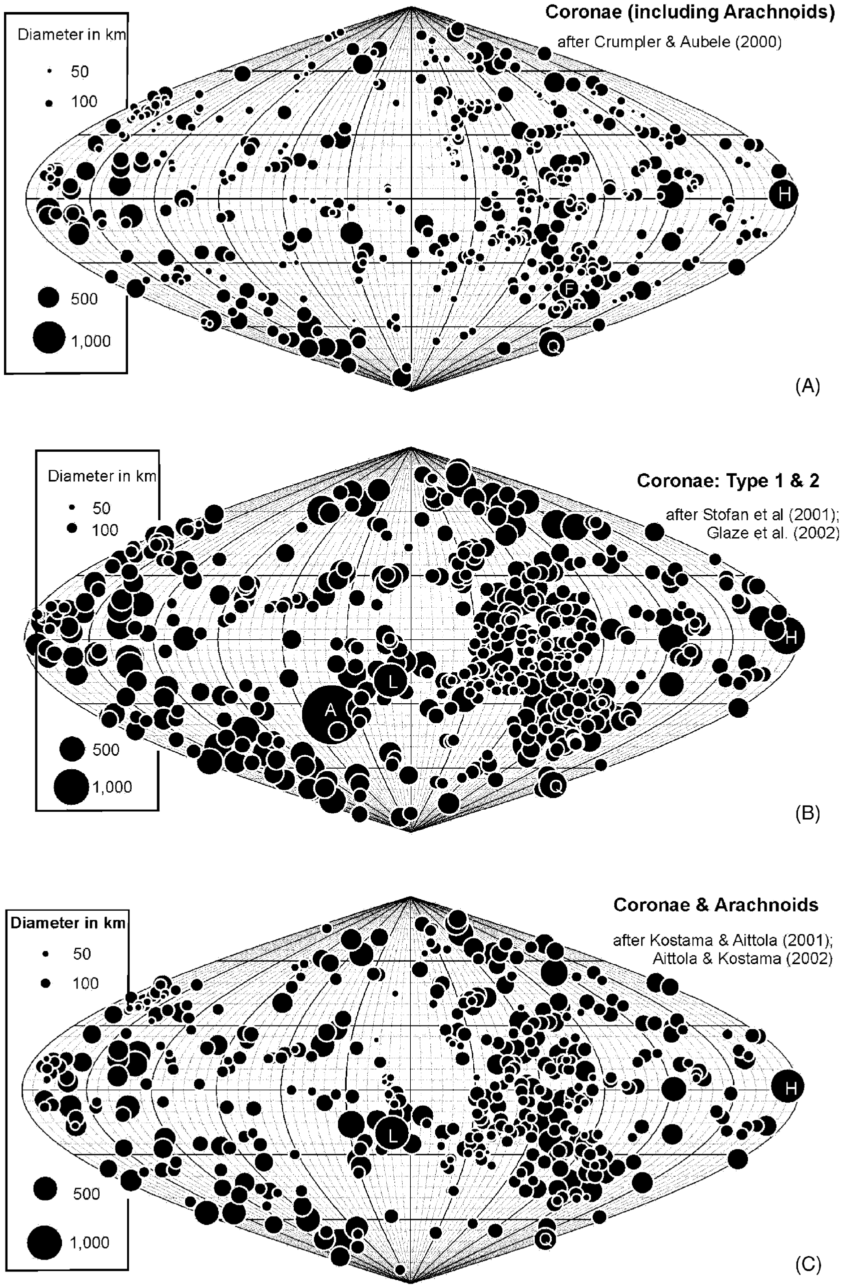


Fig. 8. Distribution of coronae on Venus. Displayed in a sinusoidal projection with a central meridian of 180°. Part A is extracted from catalogue of Crumpler and Aubele (2000), and includes both coronae and the arachnoid subclass of coronae. Part B is after Stofan et al. (2001) as modified in Glaze et al. (2002), and includes both type 1 and 2 coronae. Part C is after Kostama and Aittola (2001) and Aittola and Kostama (2002), and includes both coronae and the arachnoid subclass. A is Artemis, F is Fatua, H is Heng-O, L is Latona, and Q is Quetzalpetlatl. The largest corona, Artemis (2600 km) is absent from the catalogues used for parts A and C, because of uncertainty whether it is the same class of feature as other coronae which are typically much smaller. As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

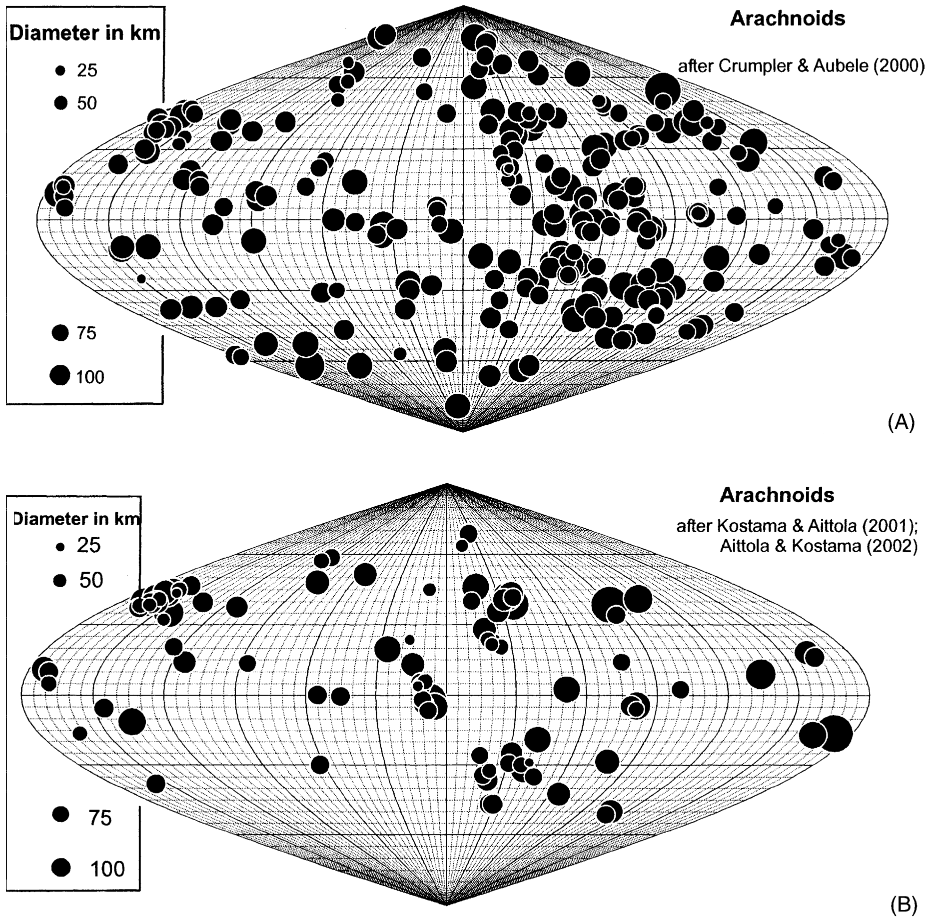


Fig. 9. Distribution of arachnoids on Venus. Displayed in a sinusoidal projection with a central meridian of 180°. Part A is extracted from catalogue of Crumpler and Aubele (2000). Part B is after Kostama and Aittola (2001) and Aittola and Kostama (2002). As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

According to the current classical model, coronae form as a consequence of impingement of a diapir (i.e. plume lacking a tail) at the base of the lithosphere. This causes an initial upwelling with associated magmatism followed by flattening of the diapir and gravitational relaxation leading to the development of the annular structure (Stofan et al., 1992, 1997; Janes et al., 1992). However, there have been difficulties in modeling the annular structure, resulting in several competing models: deformation of the surface by upwelling (Musser Jr. and Squyres, 1997), the delamination model of Smrekar and Stofan (1997), and the spreading drop model of Koch and Manga (1996). Some of the complexities may reflect differing geological histories and therefore various origins for different coronae (Copp et al., 1998; Ivanov and Head, 2003). In all the above models, the size of the corona reflects the size of the underlying diapir. In contrast, a more recent model links the size of the corona to the size of the melt zone rather than the size of the underlying plume, which can be substantially larger (Dombard et al., 2002). There is also a model in which coronae represent the surface response above large layered intrusions (Campbell and Campbell, 2003).

There is less agreement regarding the origin of the mantle diapirs; possibilities include: (1) sec-

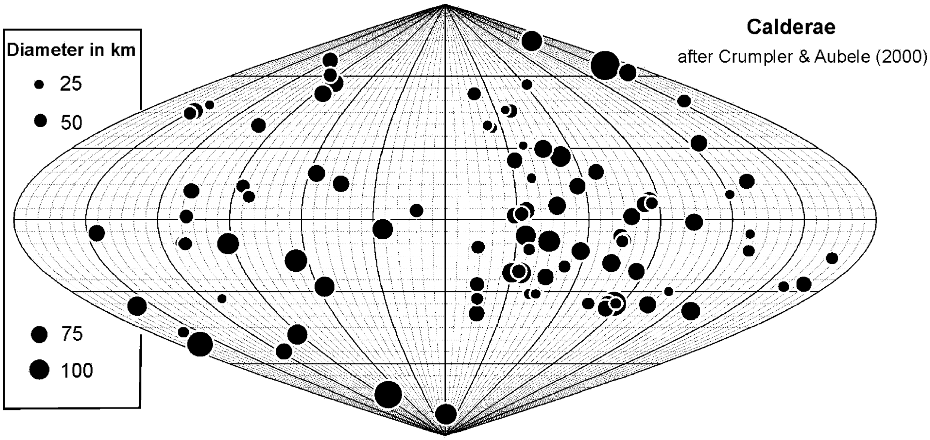


Fig. 10. Distribution of calderae on Venus. Displayed in a sinusoidal projection with a central meridian of 180°. Extracted from catalogue of Crumpler and Aubele (2000). As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

ondary convection from an upper mantle boundary layer (Phillips and Hansen, 1994); (2) compositional plumes in the upper mantle generated as melt instabilities within regions of mantle upwelling beneath zones of lithospheric extension (Hansen et al., 1997); (3) plumes that lose their tails during ascent through a mantle which is convecting vigorously (Herrick, 1999); (4) diapirs (tail-less plumes) that originate at mantle boundaries across which there is little temperature contrast (Jellinek et al., 2002); and (5) large plumes originating from the deep mantle, but whose topographic expression is limited by melt zone size and effectiveness of lithospheric penetration by the plume (Dombard et al., 2002). Recent laboratory experiments have demonstrated that small transient plumes (coronae) can coexist with and be focused toward large persistent plume conduits (e.g. Atla Regio, Beta Regio; Figs. 2 and 3) (Johnson and Richards, 2003). There are also models suggesting coexistence of mid-mantle (from below the 660 km boundary) and deep mantle plumes (Cserepes and Yuen, 2000).

Coronae occur at volcanic rises, as isolated features in the plains, and most commonly, along chasmata (rift) systems (e.g. Stefanick and Jurdy, 1996; Hansen et al., 1997; Crumpler and Aubele, 2000). The association with volcanic rises may reflect thermal diapirs resulting from the breakup of a deep-sourced thermal plume (Stofan et al., 1995), whereas coronae along chasmata may represent compositional diapirs (Hansen et al., 1997). Type 1 coronae are preferentially associated with chasmata and fracture belts, while type 2 ('stealth') coronae are found mainly as isolated features in the plains (Stofan et al., 2001; Glaze et al., 2002). The coronae associated with chasmata (and fracture belts) have the same mean diameters as those in the plains, but the coronae associated with hotspots (volcanic rises) are significantly larger (Glaze et al., 2002). The greater size for coronae in clusters may reflect a greater size for the underlying plumes.

Artemis corona (Fig. 8B, but not in 8A and C) is so much greater in size (annulus diameter of 2600 km) than the next largest coronae (1000 km radius) that its origin tends to be considered independently of the rest of the corona population. Several interpretations of Artemis have been offered. Hansen (2002) interprets Artemis as the product of a deep-sourced plume. Spencer (2001) also recognizes diapir involvement, and in addition, identifies the central region as a metamorphic core complex generated during uplift (associated with plume tail buoyancy) that exhumes deep crust or uppermost mantle. However, some of the largest coronae have also been linked to proto subduction zones (Schubert and Sandwell, 1995; Schubert et al., 1994, 2001, p. 678); e.g. Artemis, Latona, Heng-O, and Fatua (Fig. 8) with annulus diameters of 2600, 800, 900, and 500 km, respectively. Arguments against the subduction model have been discussed by Hansen et al. (1997). A further insight comes from detailed mapping of radiating graben-fissure systems associated with large coronae (e.g. Harris et al., 2002). Specifically, Fatua and Heng-O coronae have radiating graben systems (interpreted as radiating dyke swarms; Section 3.6) that may extend for >1000 km away from the center. Each of these coronae is associated with a minor topographic and gravity high (Schubert et al., 1994), but the giant scale of the associated radiating swarms suggests a significant underlying mantle feature (plumes or diapirs). The search for terrestrial analogues for corona is discussed in Section 7.6.

3.4.2. Calderae

There are 97 calderae on Venus, and their size range is typically 60–80 km (Fig. 10) (Crumpler and Aubele, 2000). They represent volcanoes that have experienced a central calderae collapse of magnitude 1–3 km (measured from calderae rim to floor). In some cases, they are also associated with radiating graben systems (Section 3.6) and thus evacuation of the central collapse may reflect subsurface dyke injection as well as volcanism. Calderae are generally smaller than arachnoids, which in turn are generally smaller than coronae (cf. Figs. 8–10). Crumpler and Aubele (2000) suggest that calderae, arachnoids and coronae represent a continuum of genetically related structures.

3.5. Large lava flow fields

On Venus large flow fields represent direct analogues of terrestrial flood basalts (Head and Coffin, 1997). They may develop through emplacement of numerous thin flow units (Byrnes and Crown, 2002). A global survey of Venus (Magee and Head, 2001) recognizes 208 volcanic flow fields that exceed 50,000 km^2 in size (Fig. 11). These “large” flow fields average 220,000 km^2, but range up to 1.6 Mkm^2, collectively constituting 11% of the plains areas on Venus. There is a sub-population of 81 classified as exceptionally large or ‘great’ flow fields on the basis of having a maximum flow length greater than 500 km (Magee and Head, 2001). As summarized by Magee and Head (2001), the flow fields most typically derive from coronae (37%), large volcanoes (25%), and fractures within rifts and fracture belts (20%). Approximately 36% are associated with (and presumably fed by) radiating fissure-graben systems.

The majority (up to 74%) of all large flow fields is located within zones of extension (major rifts and fracture belts), and largely postdates the onset of extensional deformation. Many have also been deformed by subsequent fracturing along their associated rifts. This association suggests that the rift-generated decompression melting model of White and McKenzie

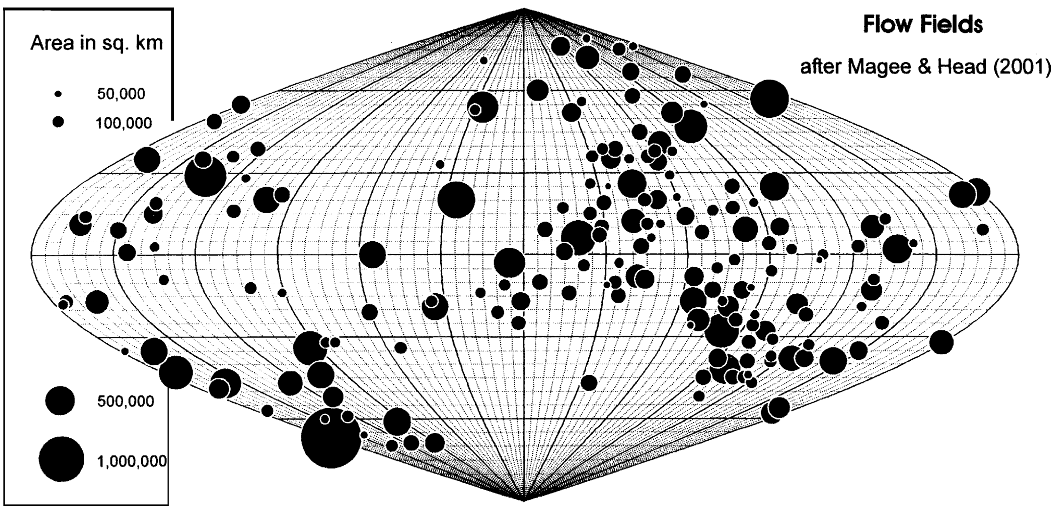


Fig. 11. Distribution of large flow fields on Venus. Size of symbol scaled to flow field area. Displayed in a sinusoidal projection with a central meridian of 180°. Based on database from Magee and Head (2001). As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

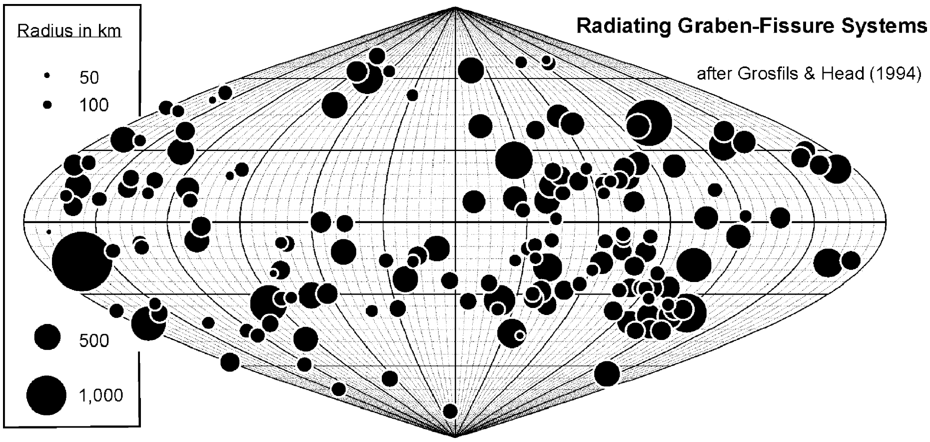


Fig. 12. Distribution of radiating graben-fissure systems on Venus based on reconnaissance study. 163 systems identified by Grosfils and Head (1994) are shown as solid circles. Displayed in a sinusoidal projection with a central meridian of 180° As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

(1989) is applicable; the intersection of mantle upwelling with zones of extension controls the location and volume of melt produced (Head and Coffin, 1997, p. 428).

3.6. Radiating graben-fissure systems

A global reconnaissance study using C1-MIDR scale (225 m per pixel) maps revealed 163 radiating graben-fissure systems (Fig. 12), which range in radius from 40 to >2000 km with an average radius of 325 km (Grosfils and Head, 1994, 1995, 1996; Grosfils, 1996). More detailed mapping is resulting in an increase in the number of radiating systems and their size. For example, detailed mapping using FMAP images (75 m per pixel) in the region of Guinevere Planitia and northern Beta Regio (Fig. 13), an area located between 264–312°E and 24–60°N (14 Mkm^2), identified 34 radiating systems of which 16 have radii greater than 300 km and 8 have radii greater than 1000 km (Ernst et al., 2003) in contrast to only 5 radiating systems identified for the same area in the reconnaissance-scale study (Grosfils and Head, 1994). In addition, 26 linear (straight) systems with lengths greater than 300 km have been mapped, of which 6 have lengths greater than 1000 km. Nineteen circumferential (circular or sub-circular systems often circumscribing a volcanic center) systems are recognized, some associated with catalogued coronae, while others may mark additional uncatalogued coronae.

There are other terms that have been applied to subtypes of radiating graben systems: novae (radially fractured centers) and arachnoids. Novae are characterized by radiating graben systems converging to a

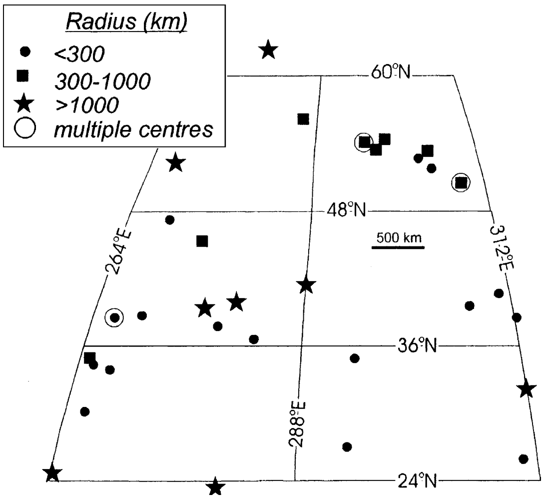
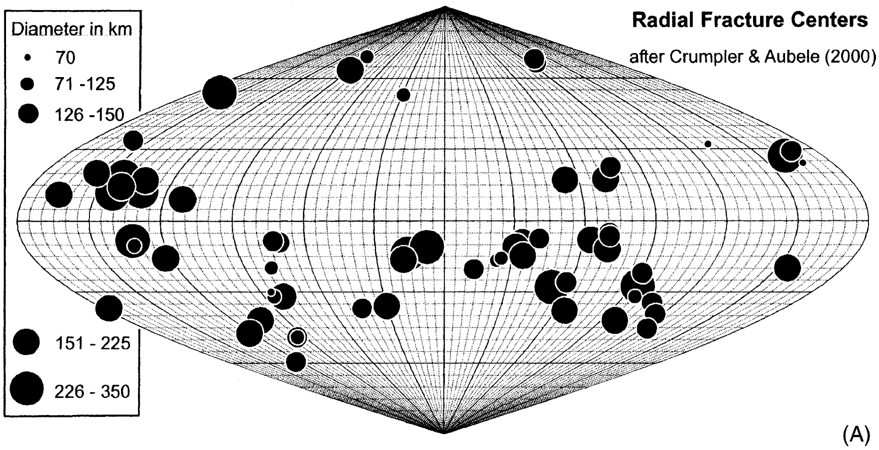


Fig. 13. Distribution of magmatic centers in Guinevere Planitia/Beta Regio region of Venus based on mapping of radiating graben-fissure systems (Ernst et al., 2003). Thirty-four radiating systems have been identified of which eight radiating systems have a radius >1000 km, and 16 have a radius greater than 300 km. In most cases, the graben-fissures systems converge on volcanoes, or coronae.



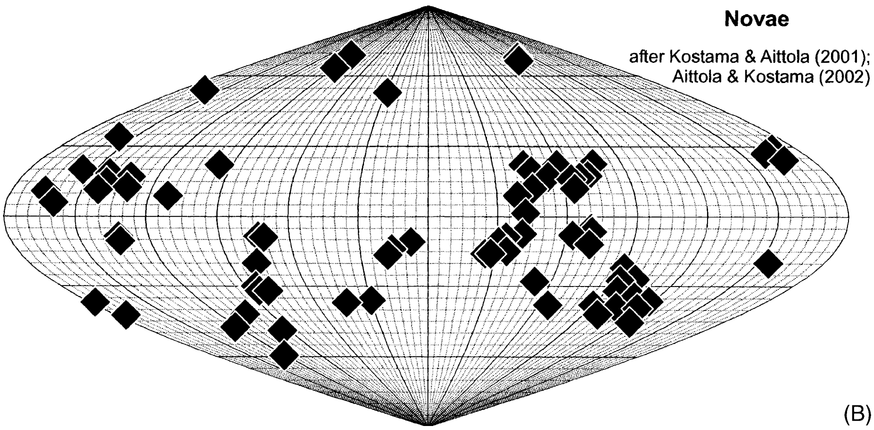


Fig. 14. Distribution of radial fracture centers (novae) on Venus. Features binned into size classes for display purposes. Displayed in a sinusoidal projection with a central meridian of 180°. Part A is extracted from catalogue of Crumpler and Aubele (2000). Part B is from Kostama and Aittola (2001) and Aittola and Kostama (2002). As in Fig. 5, symbol sizes were exaggerated to better display the range in feature sizes.

point. The "Catalog of Volcanic structures of Venus" (Crumpler and Aubele, 2000) lists 64 novae ('radially fractured centers'), which are 100–300 km in diameter (Fig. 14). All but 9 of the 64 radial fractured centers of Crumpler and Aubele (2000) are associated with radiating systems identified by Grosfils and Head (1994). Kostama and Aittola (2001) and Aittola and Kostama (2002) identify 74 novae. Novae may simply represent the better-preserved, central regions of giant radiating graben-fissure systems (e.g. Ernst et al., 2003).

Arachnoids were also discussed earlier (Section 3.4.1) in the context of their circular corona-like structures. Herein, we focus on their radiating elements. The radiating elements can be either extensional in origin or compressional. Although existing catalogues do not differentiate between these types, the compressional type (radiating wrinkle ridge-type) apparently predominates over the extensional (graben) type (e.g. Kostama and Aittola, 2001). This is consistent with the observation that only 22 of the 265 arachnoids in the classification of Crumpler and Aubele (2000) correspond to the listing of radiating fracture systems (Grosfils and Head, 1994; Grosfils, 1996).

A detailed analysis of novae and their relationship with coronae has been presented by Krassilnikov (2001) and Krassilnikov and Head (2003). They suggest an evolutionary sequence from an initial central uplift ("upraised") to various post-relaxation topographies: flat, depressed (“negative”), or “annular”. The final topography depends on the thickness of the brittle lithosphere and the ascent height of the plume. Novae with annular structures are transitional to coronae, and the annular structures are of two types: concentric ridges or concentric graben (just as for coronae). Novae with plateau-like structures are interpreted to result from the interaction of plumes (diapirs) with regional zones of extension (rifting).

Comparison of the distribution of radiating graben-fissure systems in Grosfils and Head (1994) with volcanoes, coronae (+arachnoids) in the database of Crumpler and Aubele (2000) reveals that radiating graben-fissure systems are associated with 45 of the 169 large volcanoes, 20 of the 209 coronae, 22 of the 265 arachnoids, and 1 of the 97 calderae. However, this is undoubtedly an underestimate, since more detailed mapping (at FMAP scale, 75 m per pixel) is revealing that radiating graben-fissure systems are associated with almost every known volcano, corona, or caldera in the Guinevere Planitia/Beta Regio region (Table 1 in Ernst et al., 2003).

Another way to look at the link between radiating systems and volcanoes and coronae is from the characteristics of the radiating systems sampled by Grosfils and Head (1994). They observe that 53% of the radiating systems are associated with concentric (corona-type) structures, and that in half of these cases, the radiating systems extend significantly beyond the concentric structures. Thus, about half are associated with coronae. Also, 53% (including some of the coronae) have domical topography interpreted as a central volcanic edifice.

Some radiating graben systems are purely tectonic features generated from radial extension associated with domal uplift. However, there is a growing consensus (e.g. McKenzie et al., 1992; Grosfils and Head, 1994; Ernst et al., 1995, 2001, 2003) that many radiating graben systems on Venus, particularly the largest, are the surface expression of underlying radiating dyke swarms. Specifically, Grosfils and Head (1994) found that 118 of their 163 radiating systems were interpreted to be underlain by dykes on the basis of criteria such as: (1) lava emanating from graben; and (2) systems extending far beyond any central uplift (ruling out a purely tectonic origin by uplift) (Grosfils and Head, 1994, 1995, 1996). Long-narrow extensional structures (graben and fissures) form above dykes when the top edge of a dyke does not reach the surface. Additional linear graben-fissure systems (that may or may not be underlain by dykes) are associated with rift zones and may be influenced by pre-existing structural fabrics.

3.7. Arachnoids (with radiating ridges)

As discussed earlier, arachnoids are a subset of coronae (Section 3.4.1) having radiating structures (Fig. 9). Those with radiating extensional structures were discussed in Section 3.6. Those having radiating wrinkle ridge structures are the subject of this section and represent the majority of the arachnoid population. Although the origin of these arachnoids has not been specifically modeled, the presence of a radiating pattern of ridges implies compression possibly resulting from collapse of an original domical structure.

3.8. Crustal plateaus (tesserae)

The oldest crust on Venus is represented by complexly deformed regions called tesserae (e.g. Ivanov and Head, 1996; Gilmore et al., 1998; Hansen et al., 1999, 2000). Tesserae occur as large clusters (e.g. Aphrodite Terra) and arc-like segments that may extend for thousands of kilometers (Ivanov and Head, 1996). About 8% of Venus is currently covered by tesserae, but Ivanov and Head (1996) conclude that tesserae represent a basement unit distributed over at least 55% of the surface of Venus, but largely covered by younger volcanic units. Hansen et al. (1999) disagrees and considers that tesserae represent local thickening of a globally thin and ancient Venusian lithosphere. In any case, the largest areas of tesserae are associated with crustal plateaus (Fig. 15), which represent steep-sided, flat-topped quasi-circular regions 1600–2500 km in diameter (e.g. Hansen et al., 1999). Seven crustal plateaus are identified (Fig. 15): Fortuna Tesserae, Tellus Tesserae, Alpha Regio, Eastern Ovda Regio, Western Ovda Regio, Phoebe Regio, and Thetis Regio. In addition, some of the arc-like segments of tesserae observed elsewhere may represent the outer boundary of crustal plateaus that experienced central collapse and flooding (Tewksbury, 2003). Crustal plateaus are distinct from volcanic rises (Section 4.1) by having small gravity anomalies, low gravity to topography

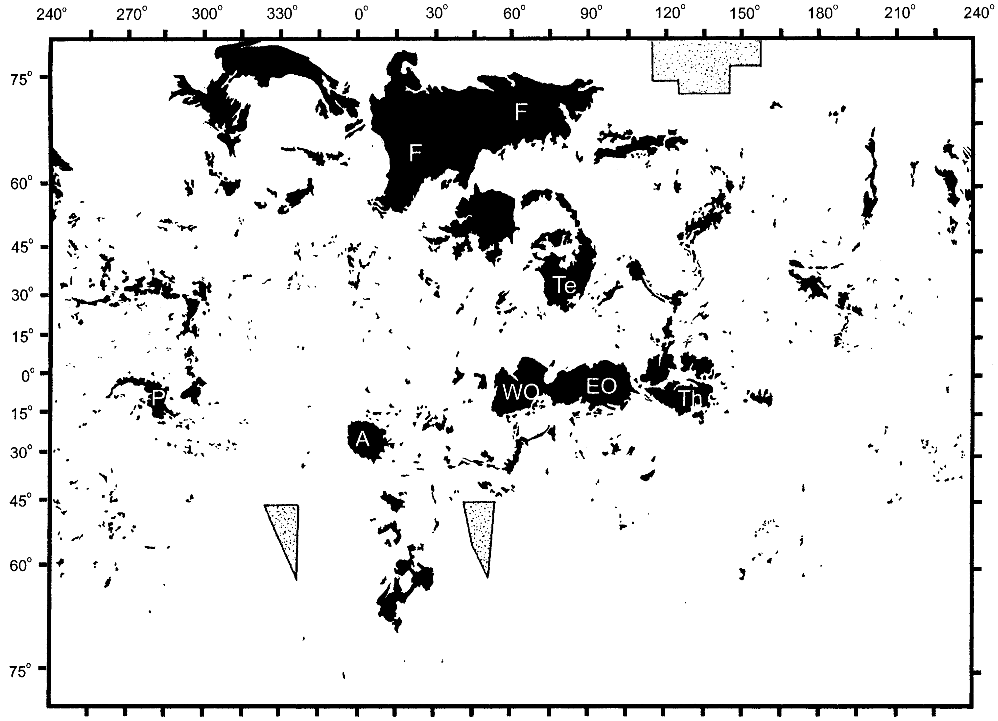


Fig. 15. Crustal plateaus (labelled with letters) that may be linked to mantle plumes. F is Fortuna, Te is Tellus, Th is Thetis, EO is Eastern Ovada, WO is Western Ovada, A is Alpha, and P is Phoebe (locations after Hansen et al., 1997). Displayed in a sinusoidal projection with a central meridian of 180°. Tesserae distribution after Ivanov and Head (1996).

ratios, and shallow apparent depths of compensation. Two general end-member models have been developed for plateau formation: "hotspot" or mantle upwelling (e.g. Phillips and Hansen, 1994; Phillips and Hansen, 1998) and "coldspot" or mantle downwelling (Bindschadler et al., 1992; Bindschadler, 1995; Ivanov and Head, 1996). Fig. 15 shows the distribution of inferred plumes according to the upwelling model for tesserae.

4. Clusters of plumes and their association with topographic and geoid highs

On Venus, some volcanoes and coronae occur in clusters, which can be associated with topographic rises and geoid highs (Figs. 3 and 4 and Table 3). (In Section 5, we discuss the remarkable association between coronae and rift systems.)

4.1. Volcanic rises on Venus

Nine volcanic rises have been identified on Venus: Atla, Beta, Bell, Dione, Imdr, Themis, and Western, Central, and Eastern Eistla Regio (Figs. 2–4 and Table 3). They range in diameter from 1000 to 2500 km, and in height from about 1 to 2.5 km. They are distinguished by their overall broad, gentle, dome-like topography, and by their gravity signature, which implies relatively deep apparent depths of compensation (Senske et al., 1992; Stofan et al., 1995; Hansen et al., 1997; Smrekar et al., 1997) and associated crustal thinning, at least for Atla and Beta Regio (McKenzie, 1994). They are sub-divided into three groups based on their dominant morphology: Atla and Beta Regio are classified as rift-dominated, Dione, Western Eistla, Bell and Imdr Regio as volcano-dominated, and Themis, Eastern Eistla, and Central Eistla Regio as corona-dominated (Stofan

Table 3
Clusters of volcanic activity on Venus (Figs. 2–4)

Label	Clusters	Characteristics
Volcanic rises: rift-dominated type		
1	Atla Regio diameter: 1200–1600 km	Apparent depth of compensation: 175 km [1] Topographic rise: swell height 2.5 km [1] Magmatism: 7 volcanic centers, most importantly Maat and Ozza Mons. Sapas Mons is to the west, but still within the geoid high Triple-junction rifting: Hecate, Parga, Dali/Diane Chasmata
2	Beta Regio diameter: 1900–2500 km	Apparent depth of compensation: 225 km [1] Topographic rise: swell height 2.1 km [1] Magmatism: Theia Mons Triple-junction rifting: Devana Chasma (two arms) and Žverine Chasma (eastern end of Hecate Chasmata)
Volcanic rises: volcano-dominated type		
3	Imdr Regio diameter: 1200–1400 km	Apparent depth of compensation: 260 km [1] Topographic rise: swell height 1.6 km [1] Magmatism: unnamed volcano; no associated coronae
4	Western Eistla Regio diameter: 2000–2400 km	Apparent depth of compensation: 200 km [1] Topographic rise: swell height 1.8 km [1] Magmatism: two major shield volcanoes (Sif Mons, Gula Mons), several smaller edifices and two coronae on its northern flank
5	Dione Regio	Apparent depth of compensation: 130 km [1] Topographic rise: swell height 0.5 km [1] Magmatism: Ushas, Innini, and Hathor Mons; no associated coronae
6	Bell Regio diameter: 1100–1400 km	Apparent depth of compensation: 125 km [1] Topographic rise: swell height 1.2 km [1] Magmatism: Tepev Mons, Nefertiti Corona
7	Themis Regio diameter: 1650–2300 km	Apparent depth of compensation: 100 km [1] Topographic rise: swell height 1.5 km [1] Magmatism: five major coronae and lies at the southeast end of the Parga Chasma corona chain”
Volcanic rises: corona-dominated type		
8	Central Eistla Regio diameter: 1000–1400 km	Apparent depth of compensation: 120 km [1] Topographic rise: swell height 1.0 km [1] Magmatism: dominated by three coronae including Sappho corona
9	Eastern Eistla Regio diameter: 1600–1800 km	Apparent depth of compensation: 65 km [1], 75 km [2] Topographic rise: swell height 1.0 km [1] Magmatism: Cluster of about 12 coronae. Includes four main coronae, the largest of which is Pavlova Corona
Arachnoid groups		
10	Bereghinya Planitia	Magmatism: At least 10 arachnoids [3]
11	Ganiki Planitia	Magmatism: 5 Arachnoids [3]
Groups lacking a distinct topographic swell		
12	Mnemosyne Regio	Magmatism: A corona cluster that lacks a distinct swell [1,4]

[1] Stofan et al. (1995); [2] Schubert et al. (1994); [3] Kostama (2002); [4] Stofan and Head (1990).

et al., 1995, p. 23317). Additional plume upwellings that are recognized in Nimmo and McKenzie (1996) are Ulfrun Regio, Farida (0°N, 34°E), and Eastern Phoebe (17°S, 51°W).

Volcanic rises are interpreted as the surface manifestation of mantle upwellings, and represent the clearest expression of terrestrial-style, deep-sourced mantle plumes on Venus because of the association with a topographic rise, a geoid high, triple-junction rifting, and LIP-scale magmatism (see plume/hotspot criteria in Burke and Dewey, 1973; Şengör, 2001; Ernst and Buchan, 2003). Particularly significant is the Beta–Atla–Themis (BAT) region (Fig. 3) where at least 4 large volcanoes are each associated with triple-junction rifting; in this area an additional 10 large volcanoes occur along or at the termination of single rifts, and an additional 13 large volcanoes are not associated with rifts.

In the broad Eistla region, there are four regional upwellings (topographic rises): Western Eistla, Central Eistla, Eastern Eistla and Bell Regio that are also marked by geoid highs (Fig. 4 and Table 3). Each appears to be represented by a clustering of volcanoes and coronae. However, unlike the Beta–Atla–Themis region, these rises are not associated with rifts, apart from one major rift (Guor Linea) associated with Western Eistla.

4.2. Other (largely older) large volcanoes on Venus

There are numerous isolated volcanoes on Venus that lack associated rifts and geoid highs, and yet the volcanoes are of similar scale to those volcanoes having these associations (Figs. 3 and 4). From stratigraphic considerations these isolated volcanoes are generally considered to be older than those near the rifts (Plate 1 in Basilevsky and Head, 2000). One interpretation is that they are not plume-related, but arise from shallow melting producing large bodies of magma (e.g. Fig. 18 in Stofan et al., 1992). However, there is no significant morphological difference in size between these isolated volcanoes and those associated with the triple-junction rifts and geoid highs.

Further insight into the nature of these isolated volcanoes is available from detailed mapping of graben-fissure systems in the Guinevere Planitia/Beta Regio region (Fig. 13) which reveals eight radiating graben-fissure systems that exceed 1000 km in radius (in a 14 Mkm^2 area) (Ernst et al., 2003). All are interpreted to represent giant radiating dyke swarms and are linked with arrival of a mantle plume head. However, only two are associated with a geoid high and triple-junction rifting. It is interpreted that the remaining five represent older plumes whose thermal anomalies have decayed. All but one of these giant radiating systems has an associated central volcanic edifice and so this association of giant radiating dyke swarms with isolated volcanoes (i.e. those not on a geoid high and lacking rifts) may indicate that in general, the isolated volcanoes mark older plumes whose thermal anomalies (and associated topographic uplift) have since decayed.

5. Linear distributions of coronae

5.1. Corona–rift (chasmata) association

Coronae occur in isolation in plains areas, as well as in clusters often associated with topographic rises (see Section 4), but also in chains associated with rifts (chasmata) (Figs. 3 and 4 and Table 4). In this section, we focus on the link with chasmata, which has been extensively discussed in the literature (see reviews in Stofan et al., 1997; Hansen et al., 1997). The main associations are with chasmata in the BAT region, specifically along the Parga and Hecate

Table 4
Linear distributions of coronae on Venus

Label (Figs. 2 or 3)	Corona–chasmata association
P	Parga Chasma: 27 coronae distributed along a 10,000 km long rift [2] Apparent depth of compensation: 125 km [1]
H	Hecate Chasma: 46 coronae distributed along a 8000 km long rift [2] Apparent depth of compensation: 150 km [1]
A	Along rifts of Aphrodite Terra system [3]
L	Alpha-Lada and Derceto-Quetzalpetlatl extensional belts in northern Lada Terra [4]
LC	Lines of coronae not associated with a mapped rift

[1] Schubert et al. (1994); [2] Hamilton and Stofan (1996), Stofan et al. (1997), Hansen et al. (1997); [3] Hansen and Phillips (1993), Stefanick and Jurdy (1996); [4] Baer et al. (1994).

Chasmata. There are 46 coronae along the 8000 km Hecate Chasma (Hamilton and Stofan, 1996) and at least 27 coronae along the 10,000 km long Parga Chasma. There is also a significant distribution of coronae along the rifts in the Aphrodite Terra region (Stefanick and Jurdy, 1996), and along the Alpha-Lada and Derceto-Quetzalpetlatl extensional belts in northern Lada Terra (Baer et al., 1994) (Fig. 2).

The coronae distributed along Hecate Chasma are about equally divided among pre-, syn- and post-tectonic classes (Hamilton and Stofan, 1996), a relationship that is also observed between coronae and extensional belts in northern Lada Terra (Baer et al., 1994). This heterogeneous association suggests that rifting and coronae formation is broadly contemporaneous (Stofan et al., 1997), although a significant age difference between early and late coronae cannot be ruled out. The absence of an age progression along chains of coronae on Venus would appear to rule out an origin by hotspots tracks (Stofan et al., 1997).

The association of coronae with chasmata suggests that lithospheric extension along chasmata causes enhanced decompression melting producing diapirs resulting in coronae. This is supported by Rayleigh–Taylor instability analysis, which predicts a shallow (150–200 km) depth-of-origin for coronae along Hecate Chasma based on an average coronae spacing of 457 ± 28 km (Hamilton and Stofan, 1996). A specific mechanism that requires only a minor degree of pressure release to begin self-perpetuating process to generate a compositional plume is outlined in Tackley and Stevenson (1991, 1993) and is discussed in Hansen et al. (1997). There are also terrestrial analogues that may be appropriate. Hamilton and Stofan (1996) suggest that the generation of chains of coronae by Rayleigh–Taylor instability is similar to the generation of spaced upwellings and volcanism at terrestrial mid-ocean ridges due to diapiric processes in the underlying asthenosphere (Crane, 1985). However, it is also possible that a deep-sourced plume rising beneath a linear zone of weakness (the rift) will exploit that weakness and cause the generation of lines of coronae.

5.2. Other linear distributions of coronae

On Venus it appears that there are also other linear distributions of coronae that are not localized along a mapped chasmata (LC in Fig. 3 and Table 4). Do these linear distribution mark older flooded rifts, or some other sort of line of weakness in the lithosphere that localized the emplacement of lines of coronae.

As noted by Hansen et al. (1997), older chasmata will loose their thermal buoyancy and sink, and may be progressively flooded by volcanism. Therefore, linear distributions of coronae may be a clue to the identification of older rift systems in which the rift has been obscured by younger flooding events.

6. Mantle overturn events

On Earth there are even larger LIP events that may have a link with a pulse of numerous mantle plumes, or may reflect a global mantle overturn event (Stein and Hofmann, 1994). We focus on the widely discussed possibility that a global magmatic event on Venus resulted in flooding of the plains in a relatively short period.

6.1. Plains volcanism

Based on a uniform distribution of impact craters, it has been inferred that global resurfacing occurred in very short period of time, perhaps as short as 10 Myr (Strom et al., 1994). Episodic resurfacing can also be obtained from mantle modeling (e.g. Ogawa, 2000). The statistical analysis of Price et al. (1996) concluded that large volcanoes, flood-type lava flow fields, rifts, and coronae have statistically fewer craters, and are therefore, significantly younger than the mean plains age. So, if catastrophic flooding occurred, it would have to be focused on the plains.

Plains units cover 80% of the surface area of Venus or ∼368 Mkm^2, and their inferred volume (assuming an average plains thickness of 2.5 km) is about 920 Mkm^3 (Head and Coffin, 1997). The magmatism responsible for flooding the plains would be equivalent to about 21 plumes of the scale of the largest LIP on Earth (Ontong Java 44 Mkm^3; Coffin and Eldholm, 2001), or would be equivalent to 184 of the largest continental LIPs (estimated to be about 5 Mkm^3 in volume; see Table 1 in Ernst and Buchan, 2001a). Using an average terrestrial continental LIP rate of once every 20 million years (Ernst and Buchan, 2002), or a combined continental and oceanic LIP rate of once every 10 Myr (Coffin and Eldholm, 2001; Ernst and Buchan, 2002), the time period required for resurfacing would range from 210 to 1840 Myr (for a one plume per 10 Myr rate), or from 420 to 3680 Myr (for a one plume per 20 Myr rate).

How do these duration estimates based on assumed terrestrial plume size and frequency compare with the actual estimates for the duration of plains volcanism on Venus. According to Price et al. (1996), the duration of plains volcanism may be between 400 and 200 Myr (values based on a mean plains age of 300 Ma). However, using the current mean-age estimate of 750 Myr with uncertainties (McKinnon et al., 1997), an age of duration from 1000 to 500 Myr or even 2000 to present would be possible (Section 3.1). This longer duration of plains volcanism would be marginally consistent with a resurfacing rate of one plume per 10 Myr. So it is just possible that the resurfacing could be accomplished by a normal terrestrial rate of plume head generated magmatism.

An alternative is that the plains were resurfaced by a thin volcanic 'veil' derived from widespread volcanism from small (1–15 km diameter) volcanoes (Hansen and Bleamaster, 2002) in which case the volume of resurfacing would be greatly reduced, and a catastrophic event would not be required.

6.2. *Beta–Atla–Themis region*

Magmatic activity (Figs. 5–14) is concentrated in the Beta–Atla–Themis region (Fig. 3) corresponding to about 9.2×10^7 km^2 or about 20% of the surface of Venus (Head and Coffin, 1997). This area of regionally abundant volcanoes largely postdates the earliest plains emplacement and corresponds to the region of three major rifted volcanic rises: Beta Regio, Atla Regio, and Themis Regio, all of which are presumed to mark young mantle plumes. The Atla plume center has been inferred to be the youngest (Nagasawa et al., 1998; Stoddard and Jurdy, 2003). The rifts (chasmata), which radiate from and connect the plume centers, are also young (Basilevsky and Head, 2000). Additional evidence for the youth of the BAT region and the significance of its magmatism are provided by a recent study of Venusian coronae by Johnson and Solomon (2002). All coronae in the database of Stofan et al. (1992) were assessed for the presence of a gravity anomaly in order to test which remains uncompensated; 35 compensated and 102 uncompensated coronae were identified. The uncompensated coronae are interpreted to be young and show a remarkable clustering in the BAT region, confirming that this region is geologically young (Johnson and Solomon, 2002).

In summary, the BAT region is marked by clustering of young coronae (representing transient plumes heads; "thermals" in the terminology of Jellinek et al., 2002) and also persistent plume conduits marked by large volcanoes (Johnson and Richards, 2003). On Earth, such a grouping of related plumes would be termed a plume cluster or "superplume event" (Condie, 2001; Ernst and Buchan, 2002; cf. also Maruyama, 1994).

7. Discussion: lessons from Venus for understanding plumes on the Earth

There are abundant magmatic/tectonic centers on Venus (Figs. 5–14), all of which occur in an global intraplate setting, since no plate tectonics has been recognized except for possible limited incipient subduction occurring along the rims of the largest coronae (Schubert et al., 1994; Schubert and Sandwell, 1995). So, all of these centers are interpreted to be related to mantle hotpots. However, the origin and nature of these hotspots may vary. The large volcanoes and clusters of coronae associated with volcanic rises (Section 4.1) represent large plumes originating from the deep mantle. Lines of coronae associated with the rift zones are smaller in diameter and probably originate at shallow levels in response to lithospheric thinning associated with the rifting. The numerous isolated large volcanoes and coronae found elsewhere on the planet have a less clear origin, but may represent thermally relaxed older plumes.

There are lessons from Venus for understanding mantle plumes on the Earth. However, we must keep in mind that while Venus seems to offer a complementary view of Earth from the point of plumes, it is possible that some distinctive differences may complicate the between-planet comparison. For instance, the viscosity contrast across the Venusian mantle may have been less than that of the Earth (e.g. Jellinek et al., 2002) and the overall mantle Venusian viscosity may have been higher owing to planetary loss of water (e.g. Moore and Schubert, 1997; Nimmo and McKenzie, 1996; Arkani-Hamed, 1996). This caveat may particularly apply in the search for terrestrial analogues of coronae (Section 7.6).

7.1. Interpretation as large igneous provinces

The volume of magmatism (combined intrusive and extrusive components) associated with individual large (>100 km diameter) volcanoes is comparable to terrestrial large igneous provinces (LIPs). Specifically, for a conical-shaped volcano, 100 km in diameter, and one kilometer high, the edifice volume is 2600 km^3. However, a volcano 500 km in diameter and 1.5 km in height comprises a volume of nearly 100,000 km^3. The terrestrial definition of large igneous provinces (Coffin and Eldholm, 1994, 2001) requires an areal extent of magmatism of >100,000 km^2, which corresponds to a volume criterion of 100,000 km^3 (Ernst and Buchan, 2001a). Thus, on Venus, volcanoes >500 km in diameter are of terrestrial LIP scale. Crumpler and Aubele (2000) list 60 volcanoes that are >500 km in diameter.

To these estimates must be added the volume of the intrusive component, which is more difficult to estimate. Many volcanoes on Venus are associated with radiating graben-fissure systems interpreted to mark underlying dyke swarms (e.g. Grosfils and Head, 1996; Grosfils, 1996; Ernst et al., 2003), which are observed to be of similar scale to terrestrial radiating dyke swarms such as the Mackenzie event of the Canadian Shield (e.g. McKenzie et al., 1992; Ernst et al., 1995). The Mackenzie radiating swarm has been estimated to consist of 60,000 line-kilometers of dyke, with an average thickness of 30 m and a depth extent of 50 km for a total volume of about 80,000 km^3 (Fahrig, 1987). An additional volume related to crustal feeder chambers and associated with crustal underplating would be expected. So a significant proportion of the large volcanoes (>100 km in diameter) on Venus have a magma volume (combined extrusive and intrusive) that must match or exceed the volumes inferred for terrestrial LIPs.

7.2. Testing models for LIP generation

The magmatic record for Venus has implications for competing models for flood basalt (LIP) generation. LIPs on Earth are generally linked to starting plumes (plume head stage) of deep-source plumes (Ernst and Buchan, 2003 and references therein). Second pulses of magmatism have been linked with rift-generated decompression melting (Campbell, 1998). Recently there has been strong interest in the edge-driven convection model (Anderson, 1998; King and Anderson, 1998). In this model, the convection across the boundary between thick hot cratonic (usually Archean) lithosphere and thinner oceanic lithosphere drives flood basalt generation. The essential feature for this model is the presence of a step-change in lithospheric thickness.

Recent study has shown that Venus has a remarkably consistent elastic lithospheric thickness of 29 ± 6 km (Barnett et al., 2002). It should be noted that Barnett et al. (2002) were not able to determine lithospheric thickness in the vicinity of the tesserae (nor within rifts). This is very different than for Earth where significant regional variations are observed: low values at spreading ridges and active mountain belts and high values (35–40 km) beneath continental shields and old oceanic lithosphere. This suggests that the edge convection model, which has been applied to some terrestrial LIPs, may not apply Venus because of its more constant (elastic) lithospheric thickness. As a corollary, the application of the edge convection model to terrestrial LIPs becomes less certain. However, this interpretation needs to be more rigorously evaluated. The only places on Venus with significant changes in lithospheric thicknesses are the chasmata (rifts), and in these settings either the edge convection model (Anderson, 1995, 1998) or the decompression melting model (White and McKenzie, 1989) could apply.

7.3. Deep-sourced plumes and plume clusters

The maximum size of plumes on Venus is similar to that inferred for Earth. The topographic rises such as Atla and Beta Regio with their triple-junction rifting and central volcanoes are about 2000 km across. This is of similar size to the largest individual plume identified on Earth (e.g. 122 Ma Ontong Java, 200 Ma Central Atlantic Magmatic Province, and 1267 Ma Mackenzie (Ernst and Buchan, 2002). In addition, this is similar to the inferred size of a plume that originates from the base of the mantle and flattens against the lithosphere (e.g. Griffiths and Campbell, 1991). The largest coronae on Venus (Ivanov and Head, 2003), which include the 2600 km diameter Artemis, 900 km diameter Heng-O, and 800 km diameter Quetzalpetlatl, suggest that plumes (producing coronae) are of similar maximum size. Thus, the plume record for Venus provides no evidence for individual plumes larger than those consistent with models for 'normal' plume ascent from the deep mantle.

On Earth, larger events such as the Pacific superplume event have been interpreted as marking a cluster of plumes (e.g. of Ernst and Buchan, 2002). A similar interpretation applies to the BAT region of Venus. Furthermore, the scale of the BAT region, almost 10,000 km across, is similar to the scale inferred for plume cluster events (also called superplume events) on Earth (Condie, 2001) such as the 122 Ma Ontong Java, Manihiki Plateau plume cluster (Ernst and Buchan, 2002). Therefore, further studies of the BAT region should provide insights into plume cluster events on Earth.

As noted earlier, the isolated large volcanoes on Venus are not associated with topographic swells or geoid highs or rifts, but are otherwise similar in scale to those associated with the young volcanic rises, and their radiating graben systems can be of similar scale. It is therefore possible that the isolated volcanoes represent older plume centers whose topographic swell and geoid anomaly have decayed, and the associated rifts have been flooded by lava. However, it is also possible that they represent a class of diapirs that were never associated with geoid highs.

7.4. Integrated plume heads and tail magmatism

As discussed earlier, the lithosphere on Venus appears broadly stationary (stagnant), and therefore continued flow along a plume tail would continue to produce magmatism that would augment the LIP generated by the plume head. The accumulated plume tail magmatism can be very significant. Recent estimates (Campbell and Davies, 2003) indicate that the strongest plume tail has a melt production 1/50 of that of the plume head. For a more typical plume tail, the rate would be about 1/250 of the plume head rate. So it would roughly take 50–250 Myr of plume tail activity to produce an equivalent volume of magmatism as produced by a plume head. Interestingly, geological constraints indicate that the uplift of Beta Regio (Figs. 2 and 3) may have occurred over the past 500 Myr (Vezolainen et al., 2003).

An important research frontier on Venus will be whether two stages of activity can be distinguished at a volcanic edifice, a underlying brief flood basalt (plume head) event, and a more protracted edifice building (plume tail) stage. Potential for such recognition is best in areas where measurable (although limited) movement of magmatic centers can be recognized (Chapman and Kirk, 1996; Grosfils and Ernst, 2002, 2003).

7.5. Plumbing system of plumes

Venus also has an important story to tell regarding the nature and distribution of magma chambers in plume center regions. On Earth, magma is observed to be distributed outward from the plume center regions both within the crust via radiating dyke swarms (Ernst et al., 1995, 2001; Ernst and Buchan, 2001b), and probably at the base of the crust as an underplate. It is also probable that mantle material is distributed by sublithospheric channeling (e.g. Ebinger and Sleep, 1998).

The important plume center regions on Earth are obscured both by erosion and by deformation associated with continental closure at the end of a Wilson cycle. On Venus, the extremely low erosion rates imply preservation of the surface expression of plumes. Some plume center regions on Venus are marked by multiple magmatic centers (magma chambers?), which are located by several generations of radiating systems (e.g. Ernst et al., 2001; Harris and Ernst, 2001; Grosfils and Ernst, 2003). Ongoing studies of magma chamber distributions above plume center regions are useful for understanding the crustal plumbing system above a plume on both Earth and Venus, and for comparisons with models (e.g. Baragar et al., 1996; Ernst et al., 2001).

7.6. Do coronae have terrestrial analogues?

Coronae (and arachnoids) are widespread on Venus, and may also occur on Mars (Watters and Janes, 1995), but are not definitively recognized on Earth.

On Earth, there are regions with volcanoes of similar size to average Venusian coronae (see Table 1), such as in northern Africa (Burke, 1996; Şengör, 2001), Europe (Wilson and Patterson, 2001), and eastern Australia (e.g. Şengör, 2001). These features have been considered terrestrial analogues for Venusian coronae (Herrick, 1999, for the African features; Wilson and Patterson, 2001 for the European examples). Importantly, none of these terrestrial cases exhibit the characteristic annular structure. In order to address this critical problem, a model by Herrick (1999) suggests that coronae represent hotspot development above a slowly moving or stationary plate (the latter is characteristic of Venus). Such conditions are rare on Earth apart from Africa. Africa has been stationary for the past 30 Myr (Burke, 1996), and Herrick (1999) infers that many of the small Neogene volcanic features associated with domical uplift in Africa may represent proto-coronae. The absence of the annular structures is explained by the relatively short duration of 30 Myr since the African features began to form; the annular structures occur late stage in the life cycle of a corona and so the African features are still too young to have developed annular structures.

There are other types of annular structures on Earth that may represent terrestrial analogues of coronae. These include: (1) arcuate dyke swarms ("circumferential" swarms) hundreds of km in diameter (Ernst and Buchan, 1998); and (2) Paleozoic anorogenic ring complexes such as those belonging to the Aïr massif in Niger, west Africa. The most notable of the latter is the 65 km in diameter, 200–300 m wide, steeply inward dipping, 400 Ma Meugueur–Meugueur troctolite ring, interpreted to be a cone sheet (Moreau et al., 1995). A third idea from Campbell and Campbell (2003) is that large layered intrusions, such as the 375 km × 300 km Bushveld complex in South Africa, represent terrestrial analogues for Venusian coronae. An applicable mechanism may be suggested by work of Diakov et al. (2002) regarding subsidence above magmatic intrusions associated with the Siberian Traps. However, erosion will inevitably remove the topographic record of annular fault structures, and make their recognition ore difficult on Earth.

In contrast, Jellinek et al. (2002) argue that coronae only form in a planetary body lacking plate tectonics (e.g. on Venus and Mars), and thus would not be expected on Earth. The absence of plate tectonics means that cold slabs are not carried into the deep mantle and therefore, the temperature transitions in the deep mantle and across the core–mantle boundary are small, which results in the formation of "thermals" (ascending diapirs that lack tails). Resolution of whether coronae actually occur on Earth is extremely important in understanding mantle dynamics for both Earth and Venus.

7.7. Small scale magmatism

Venus exhibits a remarkable range in scale of magmatism. The largest features are large volcanoes and coronae. However, magmatism ranges down to small shields a few km across. As discussed earlier (Section 2.5), the large volcanoes and clusters of coronae can be linked to terrestrial style plumes, while lines of coronae are associated with rifts, or may be linked with 'hot-lines'. However, the distribution of fields of small shield volcanoes does not fit a plume model. Some regions of shield volcanism may represent a distributed style of magmatism above an individual plume (e.g. Head et al., 1992). However, the distribution of small shields seems to be time stratigraphic (Basilevsky and Head, 1998, 2002) and thus may reflect a time of widespread thin lithosphere during development of the plains. Venusian shield fields on thin lithosphere may represent a style of magmatism analogous to scattered seamount distributions on terrestrial oceanic crust. The record for Venus demonstrates that widespread small-scale basaltic volcanism can be generated from sublithospheric or shallow mantle processes. This provides added support for ideas that some intraplate magmatism on Earth must also have a shallow source (cf. Courtillot et al., 2003).

7.8. Nature of rifting

Another useful application of comparative studies of Venus and Earth, relates to the nature of rifting associated with plume volcanism. The triple-junction rifting associated with topographic swells and volcanic centers in the BAT region (Fig. 3) is a clear example of plume-generated rifting. Furthermore, all the rifting in the BAT region and elsewhere on Venus should have intraplate origin, owing to the absence of plate tectonics on Venus. This rift record may be compared with the terrestrial rift record. Terrestrial rifts have a variety of origins (Şengör, 1995; Şengör and Natal'in, 2001), which may be grouped into active (plume-related) or passive (plate boundary-related). Distinguishing between these can be difficult on the Earth, and there are ambiguities (see rift catalogue in Şengör and Natal'in, 2001). It should prove useful to characterize the rifts on Venus in terms of degree of asymmetry, width to depth, spacing and scale of associated magmatism along rifts, in order to better characterize and recognize “active” (i.e. plume-generated) rifting on Earth.

Acknowledgements

Robert Herrick kindly provided a digital copy of his version of the map (Fig. 2 in Herrick, 1999) originally produced by Price (1995). Larry Crumpler and Jayne Aubele are thanked for a digital version of their comprehensive database of volcanic centers on Venus. Keri Magee shared her flow field database, and Ellen Stofan, her corona database. John Morgan helped with drafting diagrams and with editing. Detailed comments on an early version of the paper by Jim Head are appreciated. Discussions with Ken Buchan, as always, resulted in valued insights into the nature of giant radiating dyke swarms, large igneous provinces and mantle plumes. Detailed reviews by Simon Hanmer, James Dohm, and an anonymous reviewer are gratefully acknowledged. This is Geological Survey of Canada contribution 2003162.

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