

Venus diapirs: Thermal or compositional?

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ABSTRACT

The Venus surface exhibits quasi-circular structures with a bimodal size distribution; there are at least 17 large (~1300–2600-km diameter) crustal plateaus and volcanic rises, which form major geomorphic features, and ~515 smaller (200-km median diameter) coronae. All of these features—plateaus, rises, and coronae—are interpreted to be the surface signature of mantle diapirs. Diapirs are driven by buoyancy, which is a function of diapir size and diaper-host density contrast, which can be a function of thermal or compositional differences. Plateaus and rises apparently represent the surface signature of deep-mantle thermal diapirs that interacted, respectively, with ancient thin lithosphere and contemporary thick lithosphere. Existing coronae models either do not specify the nature of diapir buoyancy or assume a density difference resulting from thermal differences. I contrast the geological implications of end-member thermal and compositional buoyancy to provide insight into the mechanisms of Venesian diapirism operating at different scales. The analysis indicates that median-size coronae likely represent the surface expression of compositionally driven, rather than thermal, diapirs, whereas plateaus and rises form from large thermal diapirs. The bimodality and surface distribution of Venus' large (plateaus/rises) and small (coronae) diapiric structures might therefore reflect different mechanisms of diapir formation. Thermally driven deep-mantle plumes, initiated along a warm lower boundary layer, rise through the mantle to the lithosphere, forming plateaus and rises; these plumes would transfer heat from the core. Broad mantle upwellings, which presumably formed in response to a cold upper boundary layer, might generate compositional diapirs at relatively shallow levels in

the upper mantle; compositional diapirs, rising to form corona chains, would transfer heat from the mantle. Corona clusters result from compositional diapirs spawned locally by plumes of deep-mantle origin; corona clusters therefore would also represent heat transferred from the mantle, but their formation would be triggered by thermal anomalies from the core-mantle boundary. Isolated coronae probably formed only when the lithosphere was thin, because they occur in regions where the lithosphere is currently too thick to transmit small, diapiric signatures to the surface.

Keywords: Venus, diaper, coronae, plume, thermal evolution.

INTRODUCTION

Venus is presumed to have a heat budget similar to that of Earth, but it cools through a mechanism different from plate tectonics. On Earth, global-scale linear features marking divergent and convergent plate boundaries—regions where lithosphere is formed and recycled, respectively—transect the surface. Earth also displays bimodal hypsometry, reflecting two types of crust with different composition—thick, low-density continental crust and thinner, higher-density oceanic crust. Earth's early crustal differentiation and modern plate tectonics result in a planet surface comprised of large tracts of old (continent) and young (ocean basin) crust. In contrast, Venus displays unimodal hypsometry and a surface characterized by circular structures. The circular structures are both exogenic and endogenic. Approximately 950 impact craters pepper Venus' surface (Schaber et al., 1992; Phillips et al., 1992), recording a mean surface age of ca. 750 Ma (McKinnon et al., 1997). The near-random spatial distribution of craters indicates that Venus lacks large tracts of very old and very young surfaces—that is, a similar average surface age extends across much of the planet. Venus also hosts two distinct

classes of circular endogenic features: at least 17 large (~1300–2600-km diameter), quasi-circular volcanic rises and crustal plateaus (Fig. 1), and ~515 smaller (60–1050-km diameter, 200-km median diameter) quasi-circular coronae (Fig. 2) (Stofan et al., 1992, 1997, 2001; Hansen et al., 1997; Smrekar et al., 1997). Volcanic rises and crustal plateaus are interpreted to be the surface expression of deep-mantle plumes interacting with a contemporary thick lithosphere (McGill et al., 1981; Phillips and Malin, 1984; Smrekar et al., 1997) and an ancient thin lithosphere (Hansen et al., 1997, 1999; Hansen and Willis, 1998; Phillips and Hansen, 1998), respectively. Coronae are also widely accepted as the surface signature of mantle diapirs (Stofan et al., 1991, 1992, 1997; Janes et al., 1992; Squyres et al., 1992). Thus, Venus exhibits a bimodal size distribution of quasi-circular features that are interpreted as genetically related to mantle diapirs.

Diapirs move, in part, due to a density contrast with their surroundings; but density differences can result from temperature, composition, phase type, or a combination of these factors. Volcanic rises and crustal plateaus are interpreted as the surface expression of thermal diapirs (e.g., Phillips and Hansen, 1998). Although workers generally accept that coronae represent the surface signature of mantle diapirs, most do not explicitly define the nature of the diapir buoyancy, or they assume that the buoyancy is thermal (e.g., Janes et al., 1992; Janes and Squyres, 1993; Koch and Manga, 1996; Smrekar and Stofan, 1997). Because the mode and efficiency of heat transfer may vary with the mechanism of diapir buoyancy, evidence concerning the cause of diapir buoyancy should be examined. This paper examines the nature of diapir buoyancy on Venus. The conclusion—that the bimodality of Venus' large (plateaus/rises) and small (coronae) diapiric structures reflects different modes of diapir buoyancy—has implications for internal heat transfer and Venus' evolution.

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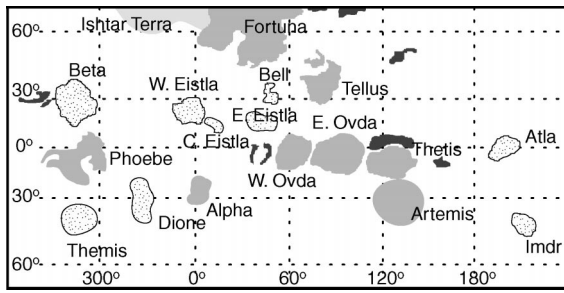


Figure 1. Distribution of crustal plateaus (gray), volcanic rises (stipple), and arcuate areas of ribbon-bearing terrain (black) marking collapsed crustal plateaus (modified from Hansen et al., 1999).

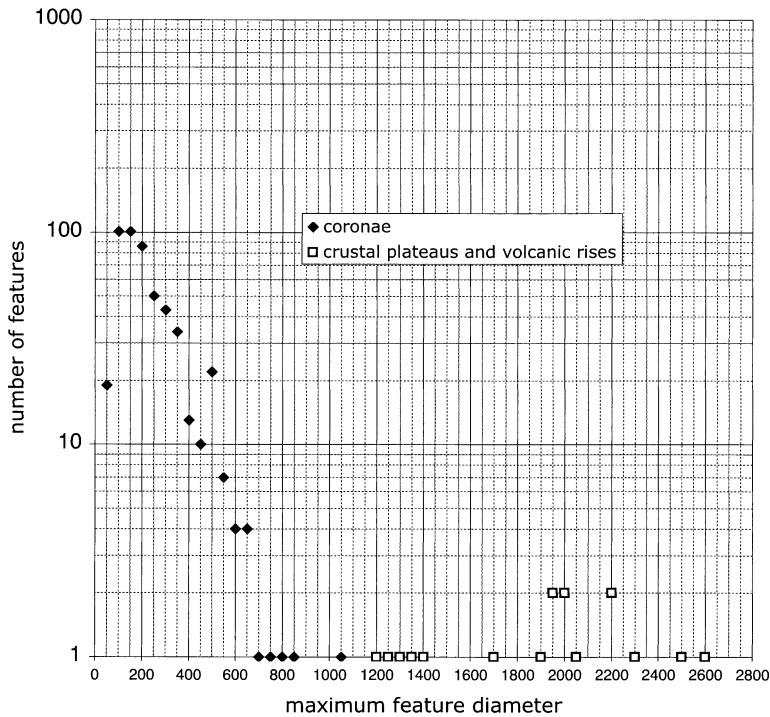


Figure 2. Plot shows number of quasi-circular features interpreted as diapiric structures versus maximum average feature diameter (data from Stofan et al., 2001; <http://planetarynames.wr.usgs.gov/>).

BACKGROUND

Hypsometrically, the Venusian surface is divisible into lowlands, mesolands, and highlands, each of which is represented by various geomorphic features, including volcanic plains, volcanic rises, crustal plateaus, coronae, ridge belts, intermediate to small volcanoes, tessera inliers, impact craters, and Ishtar Terra (Phillips and Hansen, 1994).

The lowlands, or plains—which compose ~70% of the planet’s surface—are defined topographically, volcanically, and structurally. The expansive lowlands feature narrow ridge belts, kipukas of older deformed crust, small

volcanoes, and isolated coronae. The lowlands also preserve regionally extensive suites of delicate tectonic structures, including contractional wrinkle ridges and extensional fractures (Banerdt et al., 1997), which represent small but widespread finite surface strain. The present topographically low plains basins reflect contemporary broad downwellings (Herrick and Phillips, 1992; Rosenblatt and Pinet, 1994). The ages of plains volcanism and mechanisms responsible for it remain mostly unknown. Locally, coronae appear to be volcanic centers providing a major source of plains volcanism (Young et al., 2000; Hansen and DeShon, 2002). Elsewhere, older (?)

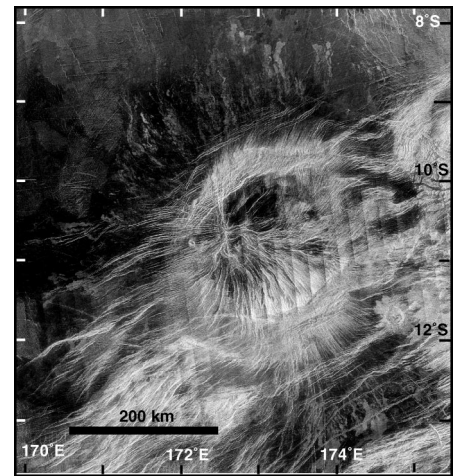


Figure 3. Magellan synthetic aperture radar image of typical median-sized corona. Note radial fractures and lava flows; concentric structures (fractures or folds) that mark annulus are commonly spatially correlated with topographic high. See Stofan et al. (1992, 1997, 2001) for further descriptions and numerous examples.

plains regions, called shield plains, host tens of thousands of small (1–15-km diameter) volcanic shields (Aubele, 1996). The global extent of shield plains is, as yet, poorly documented, but they dominate the region from ~5–60°N and ~90–150°E (Fig. 1) (Aubele, 1996; Hansen et al., 2002).

The mesolands host most of Venus’ coronae and associated chasmata (deep curvilinear to linear troughs) that together form broad chains. Coronae (Barsukov et al., 1984), circular to quasi-circular features that range in size from 60 to 1050 km diameter (Stofan et al., 1992, 2001), are typically marked by a raised rim or annulus that displays concentric annular structures such as fractures, faults, or folds; coronae also display variable tectonic and volcanic features, including radial fractures and extensive lava flow deposits (Fig. 3). (Recent work indicates that Artemis (2600-km diameter), defined by Stofan et al. (1992) as a corona, is more akin to crustal plateaus and volcanic rises [Hansen, 2002; also see Stofan et al., 1997; Hansen et al., 1997]). Coronae occur in three types of settings: (1) broad linear chains, (2) clusters associated with volcanic rises, and (3) isolated features in the plains (Stofan et al., 1997, 2001). Although “corona” originated as a descriptive morphological term, the word has come to imply a volcano-tectonic origin. Coronae are widely interpreted as the surface expression of mantle diapirs; model simulations of diapiric interaction with

lithosphere predict corona surface signatures, namely radial and/or concentric structures, variable volcanic activity, and a topographically raised rim and depressed interior (Squyres et al., 1992; Stofan et al., 1992; Cyr and Melosh, 1993; Janes and Squyres, 1993; Koch, 1994; Koch and Manga, 1996). Current models of corona formation describe a three-stage evolutionary process: (1) a diapir rises through the mantle and raises the overlying brittle crust/lithosphere into a domal uplift with or without radial fractures; (2) the diapir is flattened in a zone of neutral buoyancy, resulting in plateau-like surface topography and concentric structures at the plateau rim; and (3) as the diapir loses buoyancy, elevated topography relaxes gravitationally, resulting in the formation of new concentric structures and locally thickened crust/lithosphere (Stofan et al., 1997).

The highlands, which rise 2–11 km above mean planet radius and comprise the smallest percentage of Venus' surface area, include volcanic rises, crustal plateaus, and Ishtar Terra, a unique feature preserved in the Northern Hemisphere. Volcanic rises and crustal plateaus share size and plan-form shape, but they differ in topographic profile and geologic history. Maximum mean width and maximum median width of rises and plateaus (including Artemis) are 1900 and 2000 km, respectively (Fig. 2).

Nine volcanic rises (Atla, Beta, Bell, Dione, Imdr, Themis, and western, central, and eastern Eistla) are generally distinguished by broad, gentle domal topography and gravity signatures that indicate relatively deep depths of compensation, which are widely interpreted as evidence of thermal support (e.g., Smrekar et al., 1997, and citations therein). Volcanic rises range in diameter from 1300 to 2300 km, and in height from ~1 to 2.5 km (Fig. 1), and they exhibit volcanic processes in the form of large volcanic edifices and extensive flows. Three "corona-dominated" volcanic rises—Themis, central Eistla, and eastern Eistla—exhibit clusters of four to eight coronae (Stofan et al., 1997; Smrekar and Stofan, 1999).

Crustal plateaus display moderately steep-sided, flat-topped plateau forms, with gravity-topography signatures that are indicative of relatively shallow depths of compensation interpreted as evidence of isostatically compensated, thickened crust (e.g., Herrick et al., 1989; Smrekar and Phillips, 1991; Bindschadler et al., 1992; Grimm, 1994a; Simons et al., 1994, 1997). Crustal plateaus also host characteristic ribbon-bearing tessera fabrics (Hansen and Willis, 1996, 1998), a distinctive tectonic fabric defined by periodic ridges and

troughs. Ribbon fabrics provide evidence that a thin (≤ 1 –3 km), strong layer existed above a ductile substrate over much of the region that evolved into a crustal plateau; ribbon formation requires interaction of a large thermal diapir with globally thin lithosphere and likely elevated global temperatures (Hansen and Willis, 1998; Phillips and Hansen, 1998). Although some workers have proposed that crustal plateaus formed above cold mantle downwellings, this model does not address ribbon formation, documented synchronous volcanism and tectonism, or correlation of gravity, topography, and surface structures (Hansen et al., 1999, 2000).

Differences between crustal plateaus and volcanic rises are interpreted to reflect different global lithospheric thickness at the time of their formation. Plateaus represent the interaction of ancient deep-mantle plumes on globally thin lithosphere (Hansen and Willis, 1998; Ghent and Hansen, 1998), whereas rises record the surface signature of contemporary deep-mantle plumes on thick (~100 km) lithosphere (Stofan et al., 1997; Phillips and Hansen, 1998; Hansen et al., 1999, 2000). Phoebe Regio (Grimm, 1994a) and Artemis (Hansen, 2001, 2002), features that are transitional between plateaus and rises, may record a transition from ancient globally thin to contemporary globally thick lithosphere. Large, arcuate tracts of ribbon terrain preserve a record of ancient collapsed crustal plateaus (Phillips and Hansen, 1994, 1998; Hansen et al., 2000).

Ishtar Terra, which is not discussed in this paper, presumably formed during the early period of thin lithosphere and is supported mechanically by a deep root of mantle melt residuum (Hansen and Phillips, 1995; Simons et al., 1997).

ANALYSIS

As discussed above, Venus hosts two types of diapiric structures: (1) >17 large crustal plateaus and volcanic rises (including Phoebe and Artemis); and (2) ~515 smaller coronae (Stofan et al., 1992, 2001) (Fig. 2). Can both of these suites of diapiric structures be driven by similar mechanisms of buoyancy, or are different buoyancy mechanisms required? I contrast the geological implications of end-member thermal and compositional buoyancy to provide insight into the mechanisms of Venusian diapirism operating at different scales. The median size of volcanic rises and crustal plateaus (2000 km) is an order of magnitude larger than the median size of coronae (200 km). I am particularly concerned with median-sized

coronae and smaller; 50% of coronae have diameters less than 201 km, whereas 63% have diameters less than 251 km, and 82% have diameters less than 351 km. So, what is the general size of the diapir with which we should concern ourselves? Although large diapirs could result from the coalescing of small diapirs (Kelly and Bercovici, 1997; Manga, 1997), and therefore large coronae could result from the coalescing of two or more diapirs, it is difficult to envision how small or median-size coronae form from larger diapirs. Therefore, I use the median maximum average corona diameter (200 km) as a maximum proxy size for median corona diapirs. This likely provides an upper limit on diapir size for median-sized coronae and smaller.

Diapir ascent is driven by buoyancy that results from a density contrast between a diapir and its surroundings. Buoyancy is a function of gravity coupled with diapir radius and density. Density differences can result from temperature contrast, compositional or phase contrast, or a combination of these factors. So-called thermal and compositional diapirs interact differently with their respective host material. If a diapir is driven by temperature difference (ΔT) alone (herein called a "thermal," following Griffiths (1986a)), some surrounding material may be entrained into the diapir due to thermal diffusion; in this case, the diapir ascent velocity decreases with progressive entrainment and, hence, time. When driving buoyancy is solely a function of composition (herein broadly defined to include chemical or phase composition, including melt or partial melt, and assuming immiscibility), there is no entrainment; thus, a "compositional" diapir rises at a constant velocity, assuming a uniform and isoviscous and isothermal surrounding composition (Griffiths, 1986a). There are, of course, caveats that should be mentioned. For example, cooling of a diapir could lead to a change in the compositional phase, possibly resulting in a change in buoyancy and thus a change in ascent velocity. At the scale of a terrestrial planet, variable mantle composition could lead to progressively slower ascent velocity over time as a compositional diapir reaches neutral buoyancy. Also, a change in the thermal or viscosity profile of the host material could lead to progressively slower ascent velocity over time as a compositional diapir reaches neutral buoyancy. Given that the Earth's mantle is taken as generally adiabatic away from its upper and lower boundaries (Turcotte and Schubert, 1982; Davies, 1999), the assumption of an isothermal host is probably an acceptable first-order conjecture. The viscosity structure of the Earth's

mantle and Venus' mantle are unknown; in light of this, I assume isoviscosity—which, of course, could be incorrect. In my discussion, thermals lack deeply extended tails and are distinguished from mantle plumes, following Griffiths (1986b); mantle plumes, or plumes, are large thermal diapirs that have tails extending to their source.

Diapiric structures form on Venus when the following criteria are met—these criteria might allow us to determine whether a particular feature, or suite of features, formed as a result of thermal or compositional diapiric buoyancy. (1) Diapirs must be able to rise through the mantle at geologically reasonable rates from their depth of origin to the lithosphere. Although we have few temporal constraints for Venus, I assert that >250 m.y. for a diapir to rise from its point of formation to the crust/lithosphere is probably unreasonable. I also assume that diapirs form at a depth at least as great as their diameter. (2) Diapirs, whether thermal or compositional, form through geologically viable processes at the proposed depth of origin. (3) Upon arrival at the base of the brittle surface layer, a diapir must have sufficient buoyancy to deform, or assist in deformation, of the brittle “overburden.” Modeling of corona topography and structures by Koch and Manga (1996) provides constraints relevant to a study of corona diapir buoyancy. Koch and Manga (1996) modeled median-size (200-km diameter) coronae as 100-km diameter diapirs that spread laterally at the depth of neutral buoyancy in a layered system of mantle and crust. Their model predicts the evolution of the surface topography, including a raised rim and depressed central region, and corresponding deformation patterns reminiscent of many coronae. Although their model does not reproduce topographic troughs that surround many coronae, these troughs might be related to topographic loading of coronae (Squyres et al., 1992), crustal relaxation (Janes and Squyres, 1995), or thermal effects related to diapiric cooling (Koch and Manga, 1996), none of which were considered in their model. Koch and Manga's (1996) model uses a density contrast between the diapir and mantle of 100 kg/m³ and a surface-layer thickness of ~5–10 km to produce corona topography and surface structures. They did not specify whether the source of diapir buoyancy was thermal or compositional. Although a model by Smrekar and Stofan (1997) predicts a wide range of corona topographic features, their model assumes thermal buoyancy and also addresses the formation of large end-member coronae generally >600 km in diameter, not the gen-

eral case for coronae of median size, which is 200 km in diameter. Janes et al. (1992) and Janes and Squyres (1993, 1995) modeled the evolution of topography and surface structures of ~350 km-diameter corona. Given that I am concerned with median-sized corona, the results of Koch and Manga's (1996) model are most relevant.

The scaling factor of diapir to surface structure depends on a number of variables. In the case of rises/plateaus and coronae, we have only the surface expression of the presumed diapirs, not the diapirs themselves. Whitehead and Luther (1975) favor a 1:2 diapir:surface feature scaling factor, the scaling factor used by Koch and Manga (1996), but a 1:3 scaling factor may be equally valid (e.g., Hamilton and Stofan, 1996). I consider both scaling factors; for median rises/plateaus (2000-km-diameter surface feature) I consider 700-km and 1000-km diapirs, and for median coronae (200-km-diameter surface feature) I consider 70-km- and 100-km-diameter diapirs.

Because diapir ascent is driven by buoyancy, which is a function of gravity coupled with diapir radius and density, and because density can result from either thermal or compositional factors, we can compare the predicted steady-state ascent velocity of thermal and compositional diapirs of the same size and determine which diapirs can rise through the mantle within a geologically reasonable timeframe.

The ascent velocity of both thermal and compositional diapirs relative to the mantle is given by Stokes' flow formula for a rising, buoyant, viscous sphere (Turcotte and Schubert, 1982; Davies, 1999). Steady-state velocity, v , for a sphere of radius R with a viscosity similar to that of the host is given by:

$$v = -(g * \Delta\rho * R^2)/3 * C * \eta \quad (1)$$

where g is acceleration due to gravity (8.870 m/s² for Venus), $\Delta\rho$ is the density contrast between the diapir and the mantle, C is a constant related to viscosity ratio, and η is viscosity of the mantle. Although viscosity ratio, C , can affect diapir shape, for Earth C is generally assumed to be 1 (Davies, 1999), and a similar assumption is used here for Venus. The viscosity and viscosity structure of Venus' mantle is unknown (Schubert et al., 1997); I follow others (e.g., Parmentier and Hess, 1992; Smrekar and Stofan, 1997) in assuming an Earth-like upper-mantle viscosity of 10²¹ Pa-s. Kaula (1990) argued that Venus' mantle might be somewhat stiffer than Earth's (10²² Pa-s) due to the lack of water; Nimmo

and McKenzie (1998) argued that the viscosity is lower (3×10^{20} Pa-s).

Thermal Diapirs

The density contrast of thermal diapirs is a function of the diapir-mantle temperature differential given by $\Delta T = \Delta\rho/(\alpha*\rho_{\text{mantle}})$ (Turcotte and Schubert, 1982), where α is the coefficient of thermal expansion ($2.4 \times 10^{-5}/\text{K}$). I use 3300 kg/m³ for the density of Venus' mantle and ΔT values ranging from 50 to 400°. (As a point of comparison, other workers have proposed $\Delta T = 100\text{--}300^\circ$ for deep-mantle plumes on Earth (e.g., Campbell and Griffiths, 1990), or 250° (White and McKenzie, 1995). The resulting $\Delta\rho$ values are used to calculate steady-state rise velocity as a function of diapir size (Fig. 4). According to these simple calculations, large thermal diapirs can rise very quickly through the mantle, even with ΔT values of 100°. Diapirs with diameters of 700 and 1000 km can rise through the entire Venus mantle in ~35 and ~17 m.y., respectively, assuming no change in rise rate. If these same size diapirs have ΔT of 300°, they can rise through the entire mantle in ~12 and ~6 m.y., respectively. Thermal diapirs can increase in size due to entrainment during rise; yet our proxy for diapir size (plateaus and rises) reflects the final diapir size rather than initial diapir size. Very large thermal diapirs cool slowly, yet rise quickly; therefore, cooling has little effect on their ability to rise through the mantle. With such fast ascent rates, the large thermal diapirs considered here can traverse the Venus mantle in geologically reasonable time frames. Therefore, it is reasonable, on the basis of ascent velocity, that the diapirs responsible for volcanic rises and crustal plateaus be thermally driven. Formation of thermal diapirs requires a thermal boundary layer. Schubert et al. (1997) proposed that a thermal boundary layer might persist at ~740-km depth in Venus, associated with the spinel-perovskite transition. If such a boundary layer exists, perturbations along the layer could lead to the development of thermal diapirs. If one accepts, however, the view that diapirs form at a depth at least as great as their diameter, then the core-mantle boundary is the more likely nursery for large thermal diapirs. Indeed, it has been argued that large thermal diapirs, or plumes, have formed along the terrestrial core-mantle boundary (e.g., Morgan, 1971, 1972; Griffiths and Campbell, 1990; Sleep, 1990; Davies, 1999), and a similar scenario could be envisioned for Venus (e.g., Phillips and Malin, 1984; Phillips and Hansen, 1994; Smrekar et al., 1997). If these large

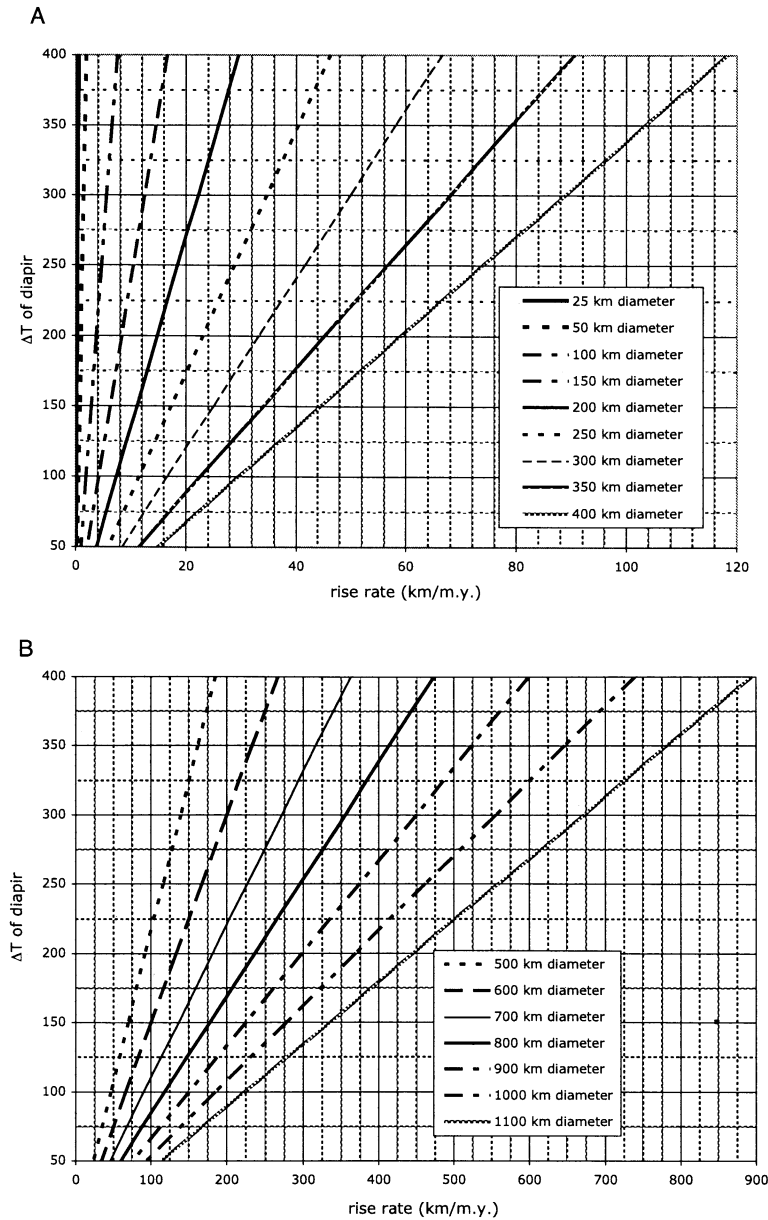


Figure 4. Plots of rise rate of thermal diapirs as function of ΔT and initial diapir diameters of (A) 25–400 km and (B) 500–1100 km.

thermal diapirs formed at the postulated ~ 740 km thermal boundary, they would represent heat transferred from the mantle; however, if the diapirs formed along the core-mantle thermal boundary, they would represent heat from the core transferred through the mantle (Stacey and Loper, 1984).

What about small thermal diapirs? Small thermals, of the size that form median-sized coronae, have rise rates roughly an order of magnitude slower than large thermal diapirs. Diapirs with ΔT of 100° and diameters of 70 and 100 km rise <2 km/m.y.; even with ΔT values of 400° , rise rates for these small dia-

pirs are <10 km/m.y. (Fig. 4B). As slow as these rise rates are, they are exaggerated because of thermal diffusion. Given that small diapirs have large surface areas relative to volume, small thermals lose heat, and therefore buoyancy, relatively quickly. The rate of cooling and the subsequent effect on ascent velocity can be estimated. We can calculate the temperature of a diapir at various depths as a function of size and the initial temperature difference between the diapir and its surroundings, $\Delta T_{\text{initial}}$ (Rititake, 1959). Figure 5 illustrates the thermal evolution of thermal diapirs with radii of 35, 40, 50, and 88 km, and

$\Delta T_{\text{initial}} = 300^\circ$. Substituting calculated ΔT into the buoyancy equation (1) at each time step yields information on the ascent rate of these various-sized thermals (Fig. 6). To place bounds on ascent velocity, we apply uniform temperature to the entire diapir equal to the temperature at distances 0.5R and 0.8R from the center, respectively (Fig. 5). Using the temperature at 0.5R yields an upper bound on ascent velocity, because a uniform high temperature is applied to the entire diapir, whereas in reality, only 12.5% of the diapir volume is at or above this high temperature. Conversely, using a temperature at 0.8R means that 50% of the diapir is at or above the assumed uniform temperature; this yields a correspondingly lower, and more likely, estimate of ascent velocity (Fig. 6). Despite use of high $\Delta T_{\text{initial}}$ and generous cooling estimates, small thermals cannot rise very far.

Diapirs with diameters of 70 and 100 km ($\Delta T_{\text{initial}} = 300^\circ$) ascend the farthest in the first 20 m.y., then slow significantly. Thermals of 70-km diameter are effectively stopped after 20 m.y. and 25 km; 100-km-diameter thermals rise slowly for ~ 20 m.y., and more slowly still for the next 20 m.y. Even after 100 m.y., a 100-km diameter thermal only rises ~ 145 km. In addition, after 32–65 m.y., these thermals have very low $\Delta\rho$ values (≤ 2 kg/m 3). Therefore, these small thermals cannot rise from a depth of 150–200 km, much less have sufficient buoyancy force to impart on a brittle surface to form a corona. Therefore, two criteria for corona formation cannot be met by small thermal diapirs: (1) small thermals cannot rise through the mantle before they cool, and (2) small thermals do not have buoyancy forces large enough to form an active diapir on a brittle surface layer. If we accept Koch and Manga's (1996) modeling of median-sized coronae, a 100-km-diameter diapir requires $\Delta\rho \sim 100$ kg/m 3 to modify a 5–20-km-thick surface layer. If such a diapir is thermally driven, $\Delta T_{\text{initial}}$ must be 1260° , and the diapir cannot cool during ascent. Such a high ΔT seems difficult to justify geologically. Thus, formation of median-size coronae by small thermals does not appear to be a geologically plausible proposition.

Compositional Diapirs

Consider a corona-forming diapir driven by compositional, rather than thermal, density contrast. If driving buoyancy is due to chemical composition or a phase change, or both, there is no material entrainment, and the diapir rises at a constant velocity, assuming a uniform (and isoviscous and isothermal) sur-

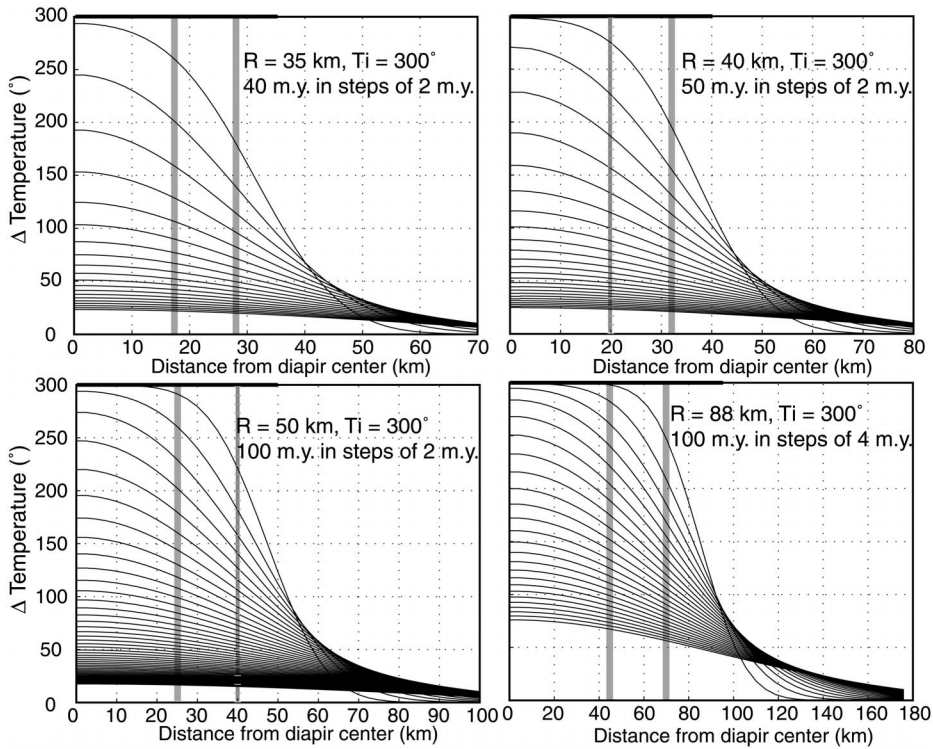


Figure 5. Cooling history of thermal diapirs following method of Rikitake (1959). R = diapir radius; initial $\Delta T = 300^\circ$. Vertical gray lines indicate T used for buoyancy calculations (Fig. 6) at $0.5R$ and $0.8R$, representing 12.5% and 50% diapir volume above temperature at that location, respectively.

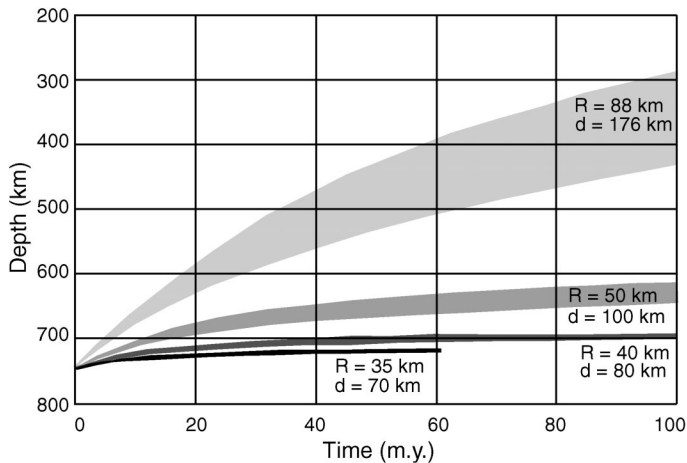


Figure 6. Rate of thermal diapir ascent with assumed starting depth of 750 km, $\Delta T_{\text{initial}} = 300^\circ$, and diapir radii (R) and diapir diameter (d) shown. Buoyancy was calculated using a uniform ΔT for each thermal with ΔT taken at 12.5% and 50% of the volume of each thermal at or above the uniform temperature (see Fig. 5). Here, the upper edge of each shaded region represents upper limit of ascent velocity for a given size thermal.

rounding composition and the caveats mentioned above (Griffiths, 1986a). Due to the lack of entrainment, the final size of the diapir, as constrained by corona diameter, provides a good estimate of the original diapir size. We

can determine steady-state velocity for 100-km diameter diapirs with a range of density contrasts (Fig. 7). A density contrast of 100 kg/m^3 (used by Koch and Manga (1996)) has a steady-state velocity of 23.3 km/m.y. Such

a compositional diapir can rise through 200 km of the upper mantle in less than 9 m.y. Compositional diapirs of 100-km diameter with density contrasts of 200 kg/m^3 and 300 kg/m^3 can rise 200 km in 4.3 and 2.9 m.y., respectively. The difference in rise rates between small compositional diapirs and small thermals is striking (Fig. 7). A 150-km-diameter compositional diapir with $\Delta\rho = 25 \text{ kg/m}^3$ can rise at the same rate ($\sim 13 \text{ km/m.y.}$) as a 150-km-diameter thermal with ΔT of 350° . A 100-km-diameter compositional diapir with $\Delta\rho$ of 100 kg/m^3 will rise at roughly the same rate as a 200-km-diameter thermal with ΔT of 325° . Furthermore, in contrast to thermal diapirs, compositional diapirs will not lose buoyancy as they rise unless mantle (host) density decreases (and therefore $\Delta\rho$ decreases) at shallower depth. In fact, if a compositional diapir consists of partial melt, it could increase in buoyancy, and therefore ascent velocity, as it rises, given a small contribution from thermal buoyancy (e.g., Tackley and Stevenson, 1993). A thermal component of coronae-forming diapirs might also be important in the surface evolution of the diapir (Janes and Squyres, 1995).

Given that small, compositional diapirs can theoretically ascend relatively rapidly through an Earth-like mantle viscosity, geologically plausible mechanisms to form compositional diapirs on Venus can be explored. Consider a case in which compositional diapirs might be generated by partial melt processes, using the Earth as a proxy. Fractional melting of mantle peridotite on Earth yields alkali basaltic magma with a density of $\sim 3100\text{--}3150 \text{ kg/m}^3$ at a depth of $>150 \text{ km}$ (Campbell and Hill, 1988), resulting in a $\Delta\rho$ of $\sim 150 \text{ kg/m}^3$. On Venus, fractional melting of the mantle might occur over broad thermal upwellings or over mantle plumes. If similar melting ensued in a mantle source region at such deep depths in Venus, then a 100-km-diameter basaltic diapir ($\Delta\rho$ of $\sim 150 \text{ kg/m}^3$) can rise at a rate of $\sim 35 \text{ km/m.y.}$, allowing it to pass through 200 km in less than 6 m.y. with $\eta = 10^{21} \text{ Pa}\cdot\text{s}$. Even 70-km- and 50-km-diameter basaltic diapirs with $\Delta\rho = 150 \text{ kg/m}^3$ can rise at rates of 17 and 8.75 km/m.y. , taking ~ 12 and 23 m.y., respectively, to rise 200 km. A 50-km-diameter compositional diapir could form a 100-km-diameter corona—the approximate lower size limit of coronae observed on Venus. Thus, median-sized and smaller coronae might result from the interaction of compositional diapirs that rise through the mantle and interact with a brittle surface layer.

But how do such compositional diapirs form? Broad thermal upwellings or plumes

may be able to spawn small compositional (melt or partial melt) diapirs, which could, in turn, rise rapidly and form coronae as they interacted with Venus' lithosphere (e.g., Tackley and Stevenson, 1991, 1993; Phillips and Hansen, 1994). However, formation and rise of a diapir requires that all material (melt in this case) rises as a coherent body, and therefore, melt must either be generated essentially instantaneously or it must be contained after formation but prior to rise. Given that melt production rates depend on a host of variables including T, P, mantle and melt composition, and volatile content, instantaneous melt formation for a 100-km-, 70-km-, or even a 50-km-diameter diapiric body (volumes of 523,600, 179,600, and 65,500 km³, respectively) is unlikely. If melt rises soon after formation, then a large volume cannot collect and rise coherently as a diapir. Therefore, formation of melt diapirs also requires trapping or containment of diapiric melt at depth until the entire diapiric volume has formed. The volume of material that can be formed and contained prior to ascent would presumably limit the size of compositional diapirs. Because compositional diapirs do not increase in size as they rise, the volume of required low-density material is equal to the final diapir volume, and thus directly related to our surface feature proxy. Koch and Manga (1996) used the widely accepted 1:2 ratio of diapir:surface feature diameter. Although Hamilton and Stofan (1996) argue for a 1:3 scaling factor, a decrease in the diapir size for a given size of resulting corona leads to a decrease in the rise rate of the diapir, or requires an increase in $\Delta\rho$, to maintain rise rate (Fig. 8). For example, a 100-km-diameter diapir with $\Delta\rho\sim 100$ kg/m³ can rise at a rate of ~ 23 km/m.y., whereas a 70-km-diameter diapir would require $\Delta\rho\sim 200$ kg/m³ to rise at the same rate. In addition, the value of $\Delta\rho$ is critical to the formation of a corona once the diapir rises through the mantle (Koch and Manga, 1996). Although it might be possible that environmental factors on Venus balance these variables and that 50–100-km-diameter volumes of melt simply remain trapped at depth until they achieve enough buoyancy, it is also possible that compositional diapirs are not comprised solely of melt, but instead represent a mixture of mantle crystalline material and matrix melt, or entirely of mantle melt residuum.

A melt percolation model previously proposed for terrestrial ocean crust formation might be appropriate to Venus. Following an earlier suggestion (Whitehead et al., 1984), Schouten et al. (1985) proposed that melt generated beneath the mid-Atlantic Ridge might

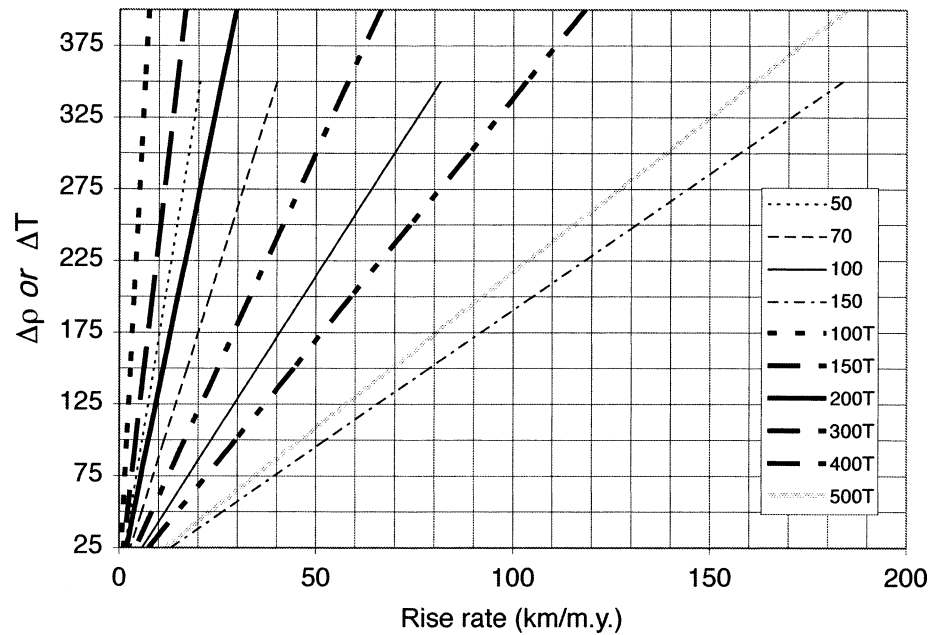


Figure 7. Plot of rise rate for compositional diapirs ($\Delta\rho$ (kg/m³); thin line) and thermals (ΔT ; thick line) as a function of diapir diameter. Note that compositional diapir diameters range from 50 to 150 km, whereas diameters of thermals range from 100 to 500 km.

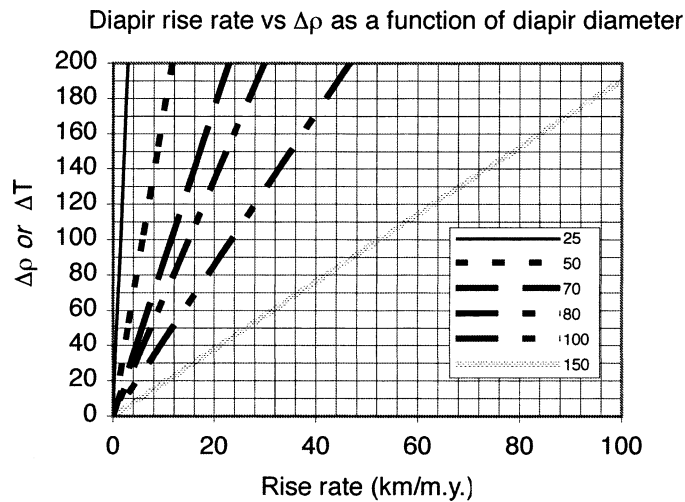


Figure 8. Plot of compositional diapir rise rate as function of $\Delta\rho$ (0–200 kg/m³) and diapir diameter (25–150 km).

rise, percolating interstitially through and into the overlying mantle to form a layer of melt-saturated(?) crystalline mantle. They further proposed that this layer of melt-matrix mantle can develop internal gravitational instabilities, causing low viscosity melt to migrate through porous flow to the tops of the perturbations, form melt diapirs, and rise to form regions of thickened oceanic crust. This model has been dismissed on Earth, however, because it is argued that melting beneath a mid-ocean ridge environment results from adiabatic decom-

pression of upwelling mantle, and therefore partial melt-saturated mantle would not be expected to form a layer (e.g., Choblet and Parmentier, 2001). Given that Venus lacks evidence of terrestrial plate boundaries (e.g., Solomon et al., 1992), perhaps the Schouten et al. (1998) model can be modified for Venus. On Venus, partial melting of the mantle might be triggered by heating from below, as noted earlier, rather than by decompression. Melt might rise and percolate interstitially through the overlying crystalline mantle; the intersti-

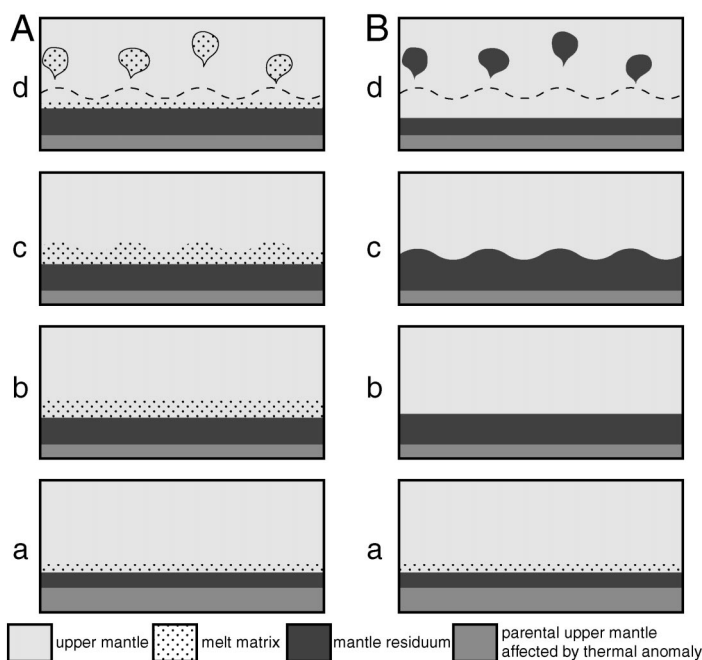


Figure 9. Cartoon illustrating model for formation of low-density layers above broad mantle thermal upwelling or mantle plume head (modified from Schouten et al. (1985)). In model A, partial melt remains in system, whereas in model B, partial melt rises interstitially upward out of system. Both models begin with partial melting of a parental upper mantle (a); partial melt collects in layer of melt matrix (mixture of melt and crystalline mantle) above mantle residuum; with continued partial melting, melt matrix rises and either collects in melt matrix layer above residuum (A.b) or percolates upward through mantle and is lost to system (B). With continued partial melting, melt matrix layer and residuum thicken (b and c); at some point, based on layer thickness and local environmental variables, thickened melt matrix layer (A.c) or residuum layer (B.c) may become buoyantly unstable and spawn melt, or melt-matrix diapirs (A.d) or residuum diapirs (B.d). In model A, residuum diapirs could also spawn from residuum layer. In either case, melt matrix layer or residuum layers could remain buoyantly stable at depth for a geologically significant period of time with instability triggered due to an event quite unrelated to their formation. See text for discussion.

tial melt might arrest at some level within the mantle, forming a layer of melt-matrix—that is, a layer in which the interstices contain melt (Fig. 9A). This melt-matrix layer would have a lower density than the surrounding mantle due to low-density interstitial liquid. As this layer continued to grow and thicken (Fig. 9Ab) it might, at some point, become buoyantly unstable and begin to rise along instabilities or perturbations (Fig. 9Ac); melt would collect at these perturbations and ultimately rise as melt diapirs when buoyancy forces became high enough, as proposed in the original terrestrial model. Other scenarios can also be envisioned. Instabilities in the melt-saturated layer could spawn diapirs of a composition similar to the layer itself—diapirs comprised of matrix melt and crystalline mantle (Fig. 9Ad) rather than pure melt diapirs. These compositional diapirs would rise to in-

teract with the overlying lithosphere, forming coronae. It is also possible that the residuum layer could become buoyantly unstable (Fig. 9B), and it, too, might spawn compositional diapirs of residuum that would rise through the overlying mantle and interact with the brittle crust to form coronae. The $\Delta\rho$ values that might be reasonably generated are difficult to estimate. Parmentier and Hess (1992) argue that partial melt of Venus-like mantle would result in a residuum with $\Delta\rho$ of 40–50 kg/m³, equivalent to a ΔT of 500° ($\Delta\rho$ depends on the value used for α ; Parmentier and Hess (1992) use $3 \times 10^{-5}/\text{K}$, whereas I use $2.5 \times 10^{-5}/\text{K}$, following Turcotte and Schubert (1982)). Compositional diapirs of 100-km diameter with $\Delta\rho$ of 40–50 kg/m³ have raise rates of 10–12 km/m.y., rising through 200 km in less than 20 m.y. The $\Delta\rho$ of a melt-matrix layer would depend both on the $\Delta\rho$ of

the melt and the percent liquid in the layer. Thus, diapirs might be comprised of melt, melt-crystalline mixture (melt-saturated mantle), or residuum. Presumably, the more tensile the tectonic environment, the more easily melt would rise interstitially and move out of the system entirely. In this case, only residuum would form a low-density layer. On Venus, such melt that rises out of the mantle system might be responsible for the formation of a relatively widespread, but as yet poorly documented, shield plains unit defined by Aubele (1996). The generation, migration, and trapping of Venus' mantle melt is an extremely complex problem, as evidenced by many terrestrial studies (e.g., Choblet and Parmentier, 2001; Braun et al., 2000; Jha et al., 1994), and it is beyond the scope of the current contribution and an exciting arena for further work. Theoretical and experimental petrologic modeling could specifically address whether large enough volumes of compositional phases could be derived from the mantle to accommodate the formation of median-sized coronae; what specific conditions, or set of conditions, would be required; and if the necessary conditions might have been time-specific to Venus' evolution, which could in turn be tested for consistency with geological mapping. Geological mapping can also test the proposed model by considering the temporal evolution of adjacent coronae along individual corona chains. Broadly synchronous formation of adjacent coronae within individual corona chains would be consistent with the model proposed here, whereas broadly diachronous formation of adjacent coronae would be inconsistent with the proposed model.

Coronae Formation

Any model of corona formation should also consider the global spatial distribution of the coronae. The majority of coronae (68%) occur along chasmata, or fracture belts; other coronae occur in association with volcanic rises (21%) and as isolated features (11%) (Stofan et al., 1997, 2001). Coronae concentrated in chains along fracture belts or chasmata would result from compositional diapirs formed by the model proposed here for broad passive-mantle upwellings. Similar processes would also lead to coronae associated with volcanic rises formed by deep-mantle plumes, as discussed below. Isolated coronae, which are perhaps more difficult to explain, are discussed later.

Coronae associated with the corona-dominated volcanic rises Themis, central Eistla, and eastern Eistla (Smrekar and Stofan,

1999), as well as those associated with Artemis (Hansen, 2001, 2002), could represent compositional diapirs spawned by deep-mantle plumes. That is, a plume could cause fractional melting of the mantle at the top of the plume, resulting in the formation of melt-matrix or residuum layers that spawn small compositional diapirs that, in turn, rise and form coronae associated with the volcanic rise structure, which resulted from the plume itself. Smrekar and Stofan (1999) argued that coronae in corona-dominated volcanic rises could not form through breakup of a deep-mantle upwelling, because such a model would require simultaneous formation of clustered coronae. However, estimating the relative timing among coronae is extremely difficult, given the nature of the remote Magellan data sets and the rich tectonic and volcanic histories recorded by coronae (Baer et al., 1994; Copp et al., 1998; Guest and Stofan, 1999; Rosenberg and McGill, 2001; Hansen and DeShon, 2002). Therefore, timing between adjacent coronae is poorly constrained by present geological mapping. In addition, I do not propose that thermal plumes themselves break up. Such an argument would be flawed, given that small thermals cannot rise great enough distances (Fig. 4B), and given experimental studies that indicate that plumes do not generally break up as they travel through homogeneous material (Griffiths and Campbell, 1991). Large (thermal) plumes can, however, generate separate small compositional diapirs in the same way that a broad (thermal) mantle upwelling can cause partial melting in the upper mantle and thus produce compositional diapirs. These small compositional diapirs would rise and form coronae associated with the volcanic rise structure, which resulted from the plume itself.

As an example, Artemis, a 2600-km-diameter circular feature composed of a central high region, circular chasma or trough, and outer rise, may represent the surface expression of just such a plume. Exterior radial fractures, a large circular trough, and concentric contractional and extensional trough structures can be accommodated within a plume model; five interior coronae and one trough corona that formed broadly synchronously with the chasma and trough structures could have formed from relatively small compositional diapirs initiated by the interaction of the plume head with the mantle (Hansen, 2002). In this case, it seems that the plume provided the thermal anomaly that led to formation of a compositional layer(s) that spawned corona diapirs and contributed to thinning of the lithosphere, pre-

paring the way for the corona diapirs to form coronae.

Depending on specific conditions, it might be possible for the melt-matrix and/or residuum layer(s) to remain at depth for quite some time, in the same way that some low-density salt layers have locally remained at depth for more than 400 m.y. on Earth. Following a salt analogy, displacement or disruption of the low-density melt-matrix layer might require a tectonic trigger to initiate layer instability (e.g., Jackson et al., 1994). The tectonic triggering mechanism could be genetically related to formation of the melt-matrix layer, such as a rising plume head in the case of Artemis, or a growing, broad mantle upwelling, or it might be genetically unrelated, as is the case of some salt tectonics.

In any case, once the compositional diapir rises, it interacts with the lithospheric crust, and either forms a corona or not, depending on specific environmental factors, such as strength of the surface layer. Although a detailed discussion of the interactions of diapirs with brittle surface layers is beyond the scope of this paper, it is worth considering some factors learned from recent studies of terrestrial salt dynamics. Recent studies of salt dynamics illustrate that salt diapirs form as reactive, passive, or active diapiric structures (e.g., Jackson et al. 1994). Coronae are probably most like active diapiric structures, and, as such, they require high diapir pressure (buoyancy) relative to overburden strength (Schultz-Ela et al., 1993; Jackson et al., 1994). Given that Venus' dry crust is quite strong (e.g., Mackwell et al., 1998) coronae diapirs likely require high diapir pressure (a function of $\Delta\rho$), and coronae probably form more easily within thin, or thinning, rather than thick lithosphere/crust. As noted above, small thermals would not have sufficient pressure to behave as active diapirs. In addition, the presence of coronae can provide clues about the thickness of the brittle crust/lithosphere at the time of corona formation; even with high $\Delta\rho$, resulting from compositional buoyancy, the brittle surface layer must yield to a diapir to form an active diapiric structure. This notwithstanding, there are several ways that Venus' diapiric structures are quite different from even active salt diapirs. The viscosity of salt diapirs is not highly temperature dependent, as is likely the case for Venus diapirs; surface erosion and depositional loading are dominant processes in salt dynamics, yet these processes are likely minor to nonexistent on Venus. Large thermal diapirs (and perhaps small compositional diapirs as well) have a thermal component that can affect both the crust/lithosphere ("over-

burden" in salt dynamics) rheology and diapir rheology, as well as contribute thermal buoyancy forces that decay with time. Venus diapirs probably do not generally maintain connection with a source layer, and ratios of diapir diameter versus crust/lithosphere thickness are probably almost an order of magnitude larger for 200-km-diameter coronae as compared to salt diapirs. In active salt systems, diapir width approximates overburden height (Schultz-Ela et al., 1993); application of this ratio to median corona diapirs would imply 100-km-thick crust/lithosphere, in strong contrast to corona modeling, which indicates layer thickness of $\sim 5\text{--}20$ km (Janes and Squyres, 1995; Koch and Manga, 1996).

Several workers have attempted to reproduce corona surface topography and structures (type and orientation) with diapiric models controlled by diapir diameter, $\Delta\rho$, and crust/lithosphere thickness. Janes and Squyres (1995), using a dry diabase flow law for Venus' surface crust (after Mackwell et al., 1998), determined that Venus' crust/lithosphere must be <10 -km thick to form coronae of ~ 350 -km diameter. Smaller coronae would require thinner crust/lithosphere with the same flow law. Median-sized coronae likely formed with crust/lithosphere thickness $\leq 5\text{--}10$ -km thick and $\Delta\rho$ values of ~ 100 kg/m³ (Koch and Manga, 1996), which can be attributed to compositional differences. In the case of both broad mantle upwellings and deep-mantle plumes, the upwellings and/or plumes provide the thermal energy to cause partial melting and result in the formation of compositional layer(s) that ultimately spawn small compositional diapirs. Upwellings and plumes also contribute to thinning of the brittle crust/lithosphere, preparing the way for compositional diapirs to form coronae in chains or clusters.

The formation of small or median-sized isolated coronae is more difficult to explain, and as such, these isolated coronae may provide a test of the hypothesis developed herein. Isolated coronae generally occur in the lowlands, where current thermal lithosphere thickness is $\sim 100\text{--}300$ km (Phillips et al., 1997). Clearly, if a small diapir—thermal or compositional—flattened along the base of such lithosphere, no corona would form at the surface. In addition, lowland gravity-topography relations are consistent with contemporary broad downwellings rather than upwellings or plumes (Herrick and Phillips, 1992; Rosenblatt and Pinet, 1994). However, if the global Venusian lithosphere changed over time from an ancient, thin lithosphere to a contemporary, thick lithosphere (e.g., Grimm, 1994b; Hansen and Willis, 1998; Phillips and Hansen, 1998),

then isolated small- to median-sized coronae preserved in the lowlands may have formed when lithosphere was thin, and thus retain a part of Venus' ancient history. Type 2 coronae of Stofan et al. (2001, originally called "stealth coronae" (Tapper, 1997)), marked by clear topographic rings but lacking a significant fracture annulus, occur most commonly as isolated coronae in the lowlands. The lack of fracture annuli could result from veiling of these coronae structures by younger volcanic processes and thus would be consistent with formation of these coronae early in Venus' history. Such a proposal can be tested with the detailed mapping currently under way across much of Venus. Detailed geological mapping may be able to delineate when these features formed, in a relative sense, with respect to the surrounding terrain. An old relative age of formation would be consistent with the proposed model, whereas a young relative age could indicate that the proposed hypothesis is seriously flawed, particularly regarding the features can be shown to have formed recently in a global sense. A young age of these features would be inconsistent with the proposed hypothesis, because small diapiric structures, even driven by compositional buoyancy, could not form within the thick lithosphere of recent times.

The analysis herein suggests that the bimodal size distribution of Venus' diapiric structures could be related to differences in diapir buoyancy. Large volcanic rises and crustal plateaus (~1300–2600-km diameter) result from large thermal diapirs or plumes in which buoyancy results from a combination of large volumes with small $\Delta\rho$ due to small ΔT . Large thermals can form along either the postulated 740 km spinel-perovskite thermal boundary or along the core-mantle boundary. The relatively large size and limited number of rises and plateaus is consistent with formation at one or two specific locations on Venus. In contrast, small diapiric structures, represented by 200-km-diameter, median-sized coronae, apparently result from compositional diapirs, with buoyancy derived from high $\Delta\rho$ values distributed over small volumes. High $\Delta\rho$ values are most easily justified as a function of composition, as are small diapiric volumes. Small thermals simply cannot rise very far before they cool. Large thermals cool slowly and therefore can rise; but large compositional diapirs are difficult to justify geologically because they require generation and ponding of large volumes of compositionally distinct material. The arguments presented herein suggest that on Venus, we should expect large thermal diapiric structures and

small compositional diapiric structures. Precisely what set of conditions nucleates large, thermally driven diapirs versus small compositional diapirs likely depends on a number of variables that may have changed through Venus' history. Given that 82% of coronae are less than ~350 km, whereas rises and plateaus range in size from 1300 to 2600 km, the distinction between small compositional diapirs and large thermal diapirs appears to be relatively great.

Given the apparent trade-off between small compositional diapirs and large thermal diapirs, perhaps the most difficult diapiric structures to rationalize in terms of the hypothesis presented herein are those of intermediate diameter (large coronae of Stofan et al., 1992). Therefore, the geological history and evolution of the large corona structures can provide tests of the proposed hypothesis. The size of thermal diapirs is presumably limited by rise rate versus cooling rate over a required rise distance. If thermals can only be generated at one of two boundary layers (core-mantle boundary or the postulated spinel-perovskite transition), then we can place plausible constraints on the minimum thermal diapirs and their resulting surface structures. For example, a 352-km surface structure might require a 176-km diameter thermal. With $\Delta T_{\text{initial}}$ of 300 K, this thermal could not rise from either thermal boundary layer in a geologically reasonable time frame (Fig. 6). Even if it could rise to interact with the lithosphere, it would have an extremely low $\Delta\rho$ value, and thus it would not result in formation of a surface structure recognized as a corona. A compositional diapir of this size also seems difficult to justify geologically unless two or more individual diapirs coalesced. Further analysis is necessary, but in this intermediate range, the choice of variables will likely be critical.

Geological mapping focused on large coronae should yield clues about their formation. The hypothesis states that these large coronae should have formed as a result of thermal buoyancy rather than compositional buoyancy; that is, these features should be more plume-like in their evolution than the median-sized coronae. Consider, for example, the six coronae larger than 650 km. The largest corona, Artemis, displays a history that is extremely different from that of median-sized coronae (e.g., Brown and Grimm, 1995, 1996; Spencer, 2001; Hansen, 2002). Artemis (35°S/135°E; 2600-km diameter) is more akin to crustal plateaus and volcanic rises and should not be considered a corona; Artemis likely represents a thermal rather than compositional diapir (Hansen, 2002), and therefore it can be

explained by the analysis presented here. Similar studies of the other large coronae should be undertaken. Hengo (2°N/355°E; 1050-km diameter), marked by a circular paired ridge and trough similar in some regards to Artemis Chasma, might also be more akin to plateaus and rises and perhaps formed from a 500-km-diameter thermal; this hypothesis can be tested with detailed geological mapping. Is Hengo's evolution more similar to that of Artemis, as would be predicted by the hypothesis proposed here, or to median-sized coronae, which would contradict the current hypothesis? Zisa Corona (12°N/221°E; 850 km diameter) appears to be an ~850 by 225 km elongate structure that may represent a composite of two coronae. Therefore, this structure does not actually have an 850-km diameter, and it might represent the composite signature of two median-sized coronae that formed from compositional diapirs. If this is the case, Zisa could be explained by the analysis presented here; if, however, geological mapping demonstrates that Zisa should be considered a single corona with geological evolution similar to that of median-sized coronae, then the proposed hypothesis should be reevaluated in light of those results. Atahesik (19°S/170°E; 810-km diameter) displays many features typical of median-sized coronae, including pervasively developed radial fractures, concentric folds and fractures, and extensive volcanic flows (Hansen and DeShon, 2002); except for its 810-km diameter, this feature seems more akin to coronae than plateaus/rises, and as such it remains a puzzle; yet, Atahesik also displays extremely deep and steep chasmata, similar to Artemis, and thus it might have a more plume-like geological history signature. Therefore, studies aimed at understanding the detailed tectonovolcanic evolution of this large and distinctive feature will be important in evaluating the hypothesis outlined herein. Quetzalpetlatl (68°S/357°E; 780 km diameter) will also require detailed geological mapping to determine its geological history and whether it is morphologically like plateaus/rises or coronae, or something else entirely. Shiwanokia Corona (42°S/280°E), which is associated with coronae-dominated volcanic rise Themis Regio, deserves detailed geological mapping aimed at understanding its tectonovolcanic evolution. Shiwanokia Corona might be a hybrid structure with both compositional and thermal diapiric components, and thus it could provide critical data to either discount or corroborate the proposed model. Additional detailed geologic study is required of these large hybrid coronae structures, which may re-

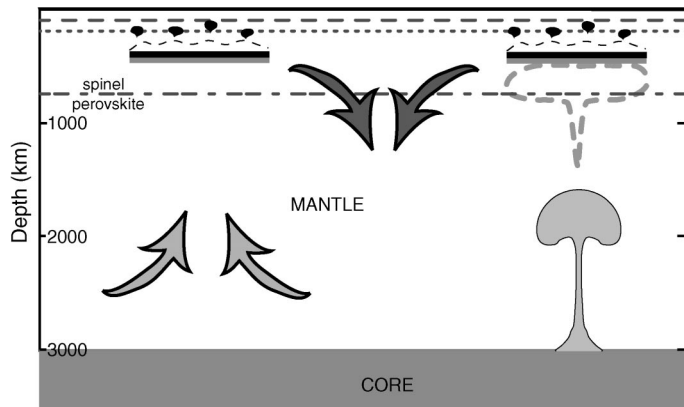


Figure 10. Scaled illustration of large thermals, or plumes, and small compositional diapirs within Venus' mantle. Top black line represents crust/lithosphere thickness during time of early, thin lithosphere; horizontal dashed line represents 100-km-thick lithosphere; horizontal dotted line marks 200-km depth for reference; possible 740-km spinel-perovskite thermal boundary layer (Schubert et al., 1997) is also shown for reference. Plumes represent heat extracted from core and transferred through the mantle; these thermals result in formation of crustal plateaus and volcanic rises on thin and thick lithosphere, respectively. Coronae form due to small compositional diapirs spawned in chains above broad mantle upwellings (left), or as clusters above some deep-mantle plumes (right). Median-sized isolated coronae can only form in thin crust, and therefore likely formed during time of ancient, thin lithosphere.

flect relatively unique conditions during Venus' evolution.

DISCUSSION

Simple buoyancy calculations taken together with geological reasoning indicates that the bimodal size of Venus' diapiric structures can be attributed to the nature of diapir buoyancy. Small diapiric structures (coronae) apparently result from diapirs driven by compositional buoyancy, whereas large diapiric structures—plateaus and rises—result from diapirs driven by thermal buoyancy. Therefore, the nature of diapir buoyancy is important to consider in postulating where and how diapirs form, and the implications for heat transfer within Venus.

Figure 10 illustrates implications of the proposed hypothesis. I assume that Venus, like the Earth, has two thermal boundary layers and is internally heated. A lower warm boundary layer at the core-mantle interface forms the nursery for deep-mantle plumes, which produce crustal plateaus and volcanic rises under conditions of thin and thick lithosphere, respectively. An upper cool boundary layer defined by the lithosphere thickens over time by secular cooling. Early in Venus' history, when the lithosphere was thin, small, spatially isolated compositional diapirs formed within the mantle, then rose and interacted with the thin lithosphere, forming isolated coronae.

Perhaps the scale and distribution of these coronae reflect overall mantle heterogeneity at the time of their formation. Also, during the early period of thin lithosphere, the upper boundary layer would presumably cool, and it might locally become buoyantly unstable and sink into the mantle, resulting in downwelling. The mechanism of downwelling is unknown, but it might involve crumpling and folding of the crust/lithosphere and detachment at depth, or involve recycling of the crust/lithosphere like Earth's plate tectonics (Phillips and Hansen, 1998).

As the lithosphere thickens with time, the lower warm boundary layer will continue to generate deep-mantle plumes, but the signature of plumes on the surface changes from that of crustal plateaus to one of volcanic rises. The upper cool boundary layer thickens and becomes stronger as a result, thereby influencing the mode and wavelength of deformation. The internally heated mantle will respond to downwellings by the formation of broad mantle (thermal) upwellings elsewhere (Davies, 1999); above such upwellings, the lithosphere will become thin relative to the lithosphere overlying the downwellings. In the current model, it is these large-scale mantle upwellings that could spawn a lower density compositional layer that forms the source for compositional diapirs that rise to form chains of coronae and associated chasmata. The role of the upwelling is twofold: (1) it triggers the

formation of a low-density compositional layer by fractional melting and upward migration, concentration, and "trapping" or stalling of the melt into the interstices of mantle above, and (2) it causes extension and thinning of the overlying lithosphere, thereby triggering the upward migration of compositional diapirs from the source layer. These compositional diapirs would rise through the mantle and interact with the relatively brittle lithosphere and form coronae. During the transition from thin to thick lithosphere, compositional diapirs formed above deep-mantle plumes would rise, and, with their parent thermal plume, form Artemis, or coronae-dominated volcanic rises (Hansen, 2002). Under conditions of thick lithosphere, deep-mantle plumes might or might not spawn a compositional layer that could, in turn, provide the source for compositional diapirs, depending on P-T conditions, among other factors. Plume-mantle interactions could also result in the formation of a compositional layer, but the presence of the layer might not yield coronae at the surface because of thicker and stronger lithosphere.

SUMMARY

The results of simple buoyancy calculations, taken with existing structural and geologic constraints, indicate that Venusian median-sized coronae most likely represent the surface interaction of compositional diapirs with the lithosphere and cannot result from thermal diapirs. Rather, corona diapirs are likely to be compositionally driven, although broad mantle upwellings or deep-mantle plumes might initially trigger the melting. Median-sized coronae thus reflect heat derived from the mantle. In contrast, Venus' large volcanic rises and crustal plateaus represent the surface signatures of deep-mantle plumes on contemporary thick lithosphere and ancient thin lithosphere, respectively. These also are diapiric structures, but they differ from coronae in that they range in size from 1600 to 2600 km in diameter—more than an order of magnitude larger than median-size coronae—and they reflect deep-mantle plume origins resulting from thermal rather than compositional buoyancy forces. Requisite plume size indicates that they likely originate at the core-mantle boundary and reflects cooling of the Venusian core rather than cooling of the mantle.

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