Floor-fractured impact craters on Venus: Implications for igneous crater modification and local magmatism

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Abstract. Regional tectonism and volcanism affect crater modification and crater loss on Venus, but a comparison of Venustian craters to lunar floor-fractured craters suggests that a third style of more localized, crater-controlled magmatism also may occur on Venus. Based on lunar models for such magmatism, Venustian crustal conditions should generally favor crater-filling volcanism over crater-centered floor fracturing. Nevertheless, three craters on Venus strongly resemble extensively modified craters on the Moon where deformation can be attributed to failure over large crater-centered intrusions. Models for crater modification over igneous intrusions indicate typical magmatic pressures beneath these three craters of ~200-300 bars and intrusion depths of the order of 1-6 km. All three craters also share common settings and low elevations, whereas craters embayed by regional volcanism preferentially occur at much higher elevations on Venus. We suggest that the style of igneous crater modification on Venus thus may be elevation dependent, with crater-centered intrusions primarily occurring at low elevations on Venus. This interpretation is consistent with the theoretically predicted variations in magmatic neutral buoyancy depth as a function of atmospheric pressure suggested by other authors.

Introduction

Venera, Arecibo and Magellan radar images reveal a relatively sparse crater population on the surface of Venus [Ivanov et al., 1986; Basilevsky et al., 1987; Campbell et al., 1990; Phillips et al., 1991; Schaber et al., 1992]. Most of these craters exhibit pristine appearances and well-preserved ejecta deposits [Phillips et al., 1991; Schaber et al., 1992]; hence there is little evidence for significant crater modification on Venus by surface erosion or sedimentation [Arvidson et al., 1992; Schaber et al., 1992]. While surface weathering and aeolian activity may modify the radar brightness and roughness of crater surfaces on Venus [Arvidson et al., 1992; Campbell et al., 1992], endogenic processes appear to be the dominant mechanisms for crater loss [Phillips et al., 1992; Schaber et al., 1992].

Few craters show extensive modification by such endogenic processes, but the exceptions are preferentially located near regions of major volcanic and tectonic activity [Basilevsky et al., 1987; Phillips et al., 1992]. In addition, the absence of superposed craters in some of the largest volcanic and tectonic provinces (e.g., Sappho Patera, Alpha Regio) indicate a localized obliteration of impact craters, which is inconsistent with a single planetary surface age [Phillips et al., 1991]. Even if the mean crater density on Venus indicates an average planetary surface age of ~400-600 Ma [Phillips et al., 1992; Schaber et al., 1992], such uncratered regions suggest more recent volcanism and tectonism within a few limited regions [Phillips et al., 1992].

Endogenic crater modification, however, may be more widespread than these observations indicate. Previous studies have focused on only two styles of endogenic crater modification: one in which a crater rim is overrun by external volcanic flows, and one in which the crater floor is deformed by regional tectonism [e.g., Schaber et al., 1992; Schaber and Chadwick, 1994]. Where such craters comprise ~4% and ~12%, respectively, of the identified craters on Venus [Schaber and Chadwick, 1994], however, studies of endogenically modified lunar craters suggest that crater-centered igneous activity can produce a third, more subtle but equally significant, style of endogenic crater modification. On the Moon, deformation and volcanism in craters primarily are confined to the crater floor region [Schultz, 1976] and may reflect interactions between interior crater structures and subsurface mare magmas [Schultz, 1976; Wichman and Schultz, 1991; Wichman, 1993]. Since the pristine appearance of most Venustian craters reflects preservation of the crater wall and rim regions, a comparison of lunar craters to Venustian craters is needed to assess the possible role of such crater-centered magmatism on Venustian crater modification.

This paper uses one of the most distinctive classes of floor-fractured lunar craters to provide an initial assessment of interior crater modification on Venus. After first summarizing the nature of deformation observed in lunar floor-fractured craters, we consider how conditions on Venus might affect similar episodes of crater modification. Next, three craters on Venus are described that show striking similarities to extensively modified lunar craters, and crater-centered laccolith models are developed to assess the nature of the inferred magmatism beneath these craters. Last, since the identified floor-fractured craters all occur at low elevations on Venus, we describe one potential mechanism for producing an elevation-dependent variation in crater-related igneous activity on Venus.

Floor-Fractured Craters on the Moon

Floor-fractured craters have been identified near volcanic plains on both the Moon and Mars [Schultz, 1972; 1976; 1978;
Pike, 1972; Young, 1972; Whitford-Stark, 1974], and also may occur on Mercury [Schultz, 1977]. The excellent image resolutions and topographic data available for the Moon, however, provide the best template for modeling such crater modification and for identifying similar structures on Venus.

Lunar Crater Modification

By definition, floor-fractured craters contain systems of prominent radial, concentric, or polygonal fractures (~0.5-2.0 km in width) within the crater floor region [Schultz, 1976]. These fractures rarely extend beyond the crater rim and typically appear to be unrelated to more regional fracturing outside a crater. In addition to these fractures, however, a number of other crater modification features also develop (Figure 1), including graben-like moat structures, coherent floor uplifts, and various degrees of crater-centered volcanism. Like the fracture patterns, these features are typically confined to the crater interior, and their development appears to reflect a progression of increasing crater modification [Schultz, 1976].

In the least modified craters (e.g., Schltiter, Figure 1a), fracturing is limited and frequently is concentrated near the center of the crater floor. Crater relief (crater rim height, floor depth and central peak heights) is relatively well preserved in such craters. More extensively modified craters also preserve fracturing in the central floor, but both the crater profile and the distribution of fracturing change with continued crater modification. Specifically, crater depths decrease with greater degrees of floor deformation, and the majority of failure becomes concentrated near the floor edge (Figure 1b). In the most extensively modified craters (e.g., Posidonius, Haldane, Figure 1c), floor uplift then produces a moat-like structure around a coherently uplifted central floor plate [Schultz, 1976; Wichman, 1993]. Such craters are much shallower than pristine craters of the same size, but central peak reliefs typically are unaffected by the floor uplift [Schultz, 1976]. Consequently, the central massifs of a few lunar craters rise above their respective crater rims [Schultz, 1976; Wolfe and Et-El, 1976]. Extended rim/wall failure also has produced peripheral concentric graben outside the crater rims in a few cases [Schultz, 1976].

Volcanic units in floor-fractured craters range from small (Alphonsus-type) dark mantling deposits to floor-flooding mare patches. Such units typically occur near fractures along the crater floor edge or within the peripheral floor moats [Schultz, 1976; Wolfe and Et-El, 1976]. In some cases, these units have completely buried both the moat and floor plate, leaving only the uplifted central peaks exposed within a flooded crater rim [Schultz, 1976].

Modification Models

Two mechanisms have been proposed for the evolution of lunar floor-fractured craters: topographic relaxation and crater-centered magmatism. The viscous relaxation of crater topography is easily modeled, and can explain the observed broad crater floor uplifts fairly well [e.g., Hall et al., 1981]. However, this mechanism has difficulties accommodating some of the more detailed aspects of extensive lunar crater modification. For instance, the relaxed crater models of Hall et al. [1981] typically have difficulties reproducing the well-defined concentric moat structures in heavily modified lunar craters. Also, while viscous relaxation should preferentially occur at long topographic wavelengths, the majority of lunar floor-fractured craters are less than 40 km in diameter [Schultz, 1976]. Indeed, in several regions, it is the smallest craters which show the greatest degrees of apparent relaxation [Wichman, 1993]. Finally, while viscous relaxation requires a localized or regional heating of the crust (possibly linked to volcanism), the repeated development of isolated mare
ponds and other volcanic features in lunar floor-fractured craters strongly suggests a direct correlation of crater modification to subsurface magmatism [Schultz, 1976].

The alternative interpretation holds that floor-fractured craters record deformation over shallow crater-centered igneous intrusions [e.g., Schultz, 1976]. Although this mechanism is less susceptible to numerical modeling and testing than is viscous relaxation, Figure 2 illustrates an idealized three-stage model that can potentially account for most of the elements observed in lunar floor-fractured craters. The model begins with the lateral growth of a sill-like magma body into a crater-centered breccia lens, but surface deformation during this stage is minimal, as only minor, localized deformation occurs at depth along the intrusion edge [Johnson and Pollard, 1973]. When the intrusion begins to expand beyond the crater floor, however, decreasing brecciation beneath the crater wall impedes further horizontal intrusion growth. Consequently, continued magma injection eventually initiates successive stages of vertical intrusion growth through flexure and/or uplift of the crater floor [Wichman, 1993]. (Although flexure requires an initial elastic response in the uplifted breccias, this elasticity can reflect either a near-surface, coherent impact melt unit, or breccia welding by disseminated impact melts and residual impact heating.)

With the onset of floor uplift, doming of the crater floor by early flexure accounts for the central, radial to polygonal fracture patterns observed inside the least modified lunar craters. Then, as later flexure initiates deep-seated failure near the intrusion edge (Figure 2b), deformation shifts toward the edge of the crater floor and eventually decouples the uplifted floor materials from the rest of the crater. At this stage, flexural deformation becomes relatively insignificant, and both the floor plate and the observed moat structures result from simple uplift inside peripheral ring faults (Figure 2c). This pistonlike uplift of materials over an intrusion is analogous to the uplifts observed over some terrestrial laccoliths [Johnson and Pollard, 1973; Corry, 1988]. Finally, the escape of magmas to the surface through the peripheral ring faults can accommodate the preferential development of volcanism along crater floor edges on the Moon [Schultz, 1976].

Application to Venus Surface (Theory)

Although the high surface temperatures on Venus should favor topographic relaxation, the observed crater depths on Venus suggest that this mechanism has had only a limited to negligible effect on crater topography [Grimm and Solomon, 1988]. Therefore, given both the widespread evidence for past volcanism on Venus [Head et al., 1992] and the apparent frequency of magmatism in lunar floor-fractured craters [Schultz, 1976], this paper focuses on the implications of the igneous intrusion model for Venusian crater modification. Since surface failure in this model is directly related to intrusion growth, however, differences in both crustal and magmatic conditions may alter the typical expression of such magmatism on Venus from that observed on the Moon. In particular, crustal composition, crustal coherency, surface gravity, and surface temperature all may affect the interaction of magmas with crater-centered breccias, and thus may influence both the likelihood and the surface expression of crater-centered magmatism on Venus.

First, intrusion depths should reflect differences in crustal composition and density. On the Moon, crustal anorthosites are nearly equivalent in density or slightly less dense than most mare basalt magmas [Solomon, 1975]. Therefore, since lunar magmas typically developed only a limited hydrostatic head, extensive
The differences in crustal composition and surface gravity suggest that: (1) crater-centered intrusions may be less likely on Venus than on the Moon, and that (2) crater-filling volcanism is
more likely on Venus than lunarlike crater floor fracturing. While the effects of Venustian surface conditions may not favor deformation over crater-centered intrusions, however, such crater modification is still possible, especially for intrusions developing shortly after crater formation. Since a survey of Venustian craters has revealed several cases of lunarlike crater floor fracturing, some consideration must be given to the role of crater-centered igneous activity in crater modification on Venus.

**Floor-Fractured Craters on Venus**

Floor-fractured lunar craters contain a variety of radial, concentric, and polygonal fracture patterns, but the scarpsbounded floor uplifts with their associated wide moat structures present the most distinctive signatures of crater modification. Such moats tend to be both wider (up to ~8~10 km) and of greater relief (up to hundreds of meters) than the central fracture patterns produced by earlier stages of deformation, and should be clearly visible in Magellan imagery. In addition, such moat structures accompany the most advanced stages of crater modification on the Moon, and are difficult to explain via viscous relaxation. Consequently, although patterns of radial and polygonal features may reflect minor viscous relaxation in some large Venustian craters [Wichman and Schultz, 1993], moat-delimited floor plate structures should indicate the most likely sites for lunarlike, igneous crater modification on Venus.

**Examples**

Three craters on Venus show unequivocal examples of moat-delimited floor plate uplifts (Figure 3), while six other craters show less clear evidence for shallow moats around portions of their crater floors (Table 1). The first, and smallest, clear example is located between Anio Planitia and Ovda Regio at ~18S, 70E (Figure 4). Although only 19 km in diameter, this crater exhibits an annular rim deposit, a sharp crater rim (especially to the west), and a clear scarp structure surrounding the central floor region. The crater appears to be superposed upon a set of slightly arcuate graben fractures, which are part of a larger regional trend perpendicular to Ovda Regio [Solomon et al., 1992].

The second example, Barrymore, is nearly 50 km in diameter and occurs in ridged lowland plains near Imdr Regio at ~52S, 196E (Figure 5). Again the crater interior contains a clear moat structure and a scarpbounded central floor region typical of extensively modified lunar craters. Both the moat and floor appear to have been resurfaced after impact, possibly during emplacement of the surrounding dark plains units. Wrinkle ridges transect both the moat and floor units. In Magellan topography, the moat structure is up to 200 m in depth and total (rim-floor) crater depth is ~450 m (Figure 5c).

The third, and largest (~85 km diameter), example is on the edge of Eistla Regio at ~25N, 25E. This crater, Mona Lisa, exhibits a complex crater floor morphology (Figure 6), in which a central dark floor unit has breached two rings of radar-bright ridges and isolated massifs to form moatlike extensions against the northern half of the crater wall. The innermost ring structure nearly surrounds the central dark floor region and it has a ring/rim crest diameter ratio similar to other peak ring craters on Venus (Figure 7, Mona Lisa). Consequently, we interpret this feature to be the peak ring resulting from crater formation. The outer ring feature, however, has a ring/rim crest diameter ratio more comparable to floor/rim crest diameter ratios in other Venustian craters (Figure 7, Mona Lisa). Since floor plate uplift on the Moon [Wichman, 1993] and in the other two examples on Venus (Figure 7) closely correlates with pristine crater floor diameters, we interpret this outer floor ring in Mona Lisa as the edge of an uplifted floor plate. The mottled, hummocky terrain located south and west of the central peak ring may represent an uplifted section of the initial crater floor edge which appears to be buried by dark materials to the north (Figure 6b). A set of possible moat fractures also is visible in the southern crater floor (Figure 6a, arrows).

Mona Lisa also exhibits a clear set of central floor fractures suggestive of domal floor uplift (Figure 6a). These fractures postdate emplacement of the dark floor materials and follow two primary patterns: a ring of concentric fractures near the edge of the central floor region, and a set of radial/polygonal fractures within this ring. Since the ejecta pattern and the scalloped southern crater rim at Mona Lisa indicate crater formation by an oblique impact from the north [Schultz and Gault, 1991; Schultz, 1992], the northward offsets of both fracture patterns from the crater rim center may reflect an initial asymmetry of the transient cavity. The offset locations of both central peaks and basin uplifts (relative to the final crater rim) support similar asymmetries in a

**Table 1. Venustian Craters Showing Floor Moat Structures**

<table>
<thead>
<tr>
<th>Name</th>
<th>Type*</th>
<th>Diameter, km</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation, m†</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>unnamed</td>
<td>clear</td>
<td>19</td>
<td>-18.5</td>
<td>70.5</td>
<td>-400</td>
<td>Figure 4</td>
</tr>
<tr>
<td>Barrymore</td>
<td>clear</td>
<td>50</td>
<td>-52.3</td>
<td>195.55</td>
<td>-400</td>
<td>Figure 5</td>
</tr>
<tr>
<td>Mona Lisa</td>
<td>clear</td>
<td>86</td>
<td>25.55</td>
<td>25.1</td>
<td>-900</td>
<td>Figure 6</td>
</tr>
<tr>
<td>Antonina</td>
<td>equivocal</td>
<td>15.5</td>
<td>28.2</td>
<td>106.8</td>
<td>-950</td>
<td>Buried most structure?</td>
</tr>
<tr>
<td>Caroline</td>
<td>equivocal</td>
<td>17</td>
<td>6.85</td>
<td>306.3</td>
<td>-50</td>
<td>Uplifted floor section</td>
</tr>
<tr>
<td>Jeanne</td>
<td>equivocal</td>
<td>19.5</td>
<td>40.05</td>
<td>331.45</td>
<td>-600</td>
<td>Partial moat structure</td>
</tr>
<tr>
<td>Lagerlof</td>
<td>equivocal</td>
<td>59</td>
<td>81.15</td>
<td>284.5</td>
<td>-750</td>
<td>Partial moat structure</td>
</tr>
<tr>
<td>Teasdale</td>
<td>equivocal</td>
<td>35</td>
<td>0.75</td>
<td>76.25</td>
<td>750</td>
<td>Overprinted by nearby rift</td>
</tr>
<tr>
<td>Tanya</td>
<td>equivocal</td>
<td>14</td>
<td>-19.3</td>
<td>282.7</td>
<td>1100</td>
<td>Partial moat structure, rim may be volcanically embayed</td>
</tr>
</tbody>
</table>

* Clear implies most scarp traceable around most of crater floor plate, equivocal indicates presence of a less well-defined moat structure or the identification of moat scarps on only one side of the crater floor.
† Elevations derived from global Magellan topography, referenced to mean radius of 6051.8 km.
number of other oblique impact structures on the Moon and Venus [Schultz and Gault, 1991; Schultz, 1992; Wichman and Schultz, 1992].

Geologic and Topographic Setting

These examples of floor-fractured craters on Venus all occur near the edge of lowland volcanic plains (Figure 3), and in all three cases, the neighboring highlands show a strong component of volcanic activity [Crumpler et al., 1993]. Many of the tessera structures in eastern Ovda Regio are embayed by younger dark lavas [Solomon et al., 1992], whereas both Imdr Regio and Eistla Regio contain large shield volcanoes. Similarly, most of the equivocal floor-moat structures also occur at low elevations (Table 1). In contrast, the identified examples of externally embayed, volcanically modified craters on Venus [Schaber et al., 1992] preferentially occur at much higher elevations (Figure 3). Consequently, we suggest that the character of igneous crater modification on Venus may be a function of regional surface elevation. After first discussing the implications of the laccolith modification model for intrusion properties beneath these craters, we will describe one mechanism by which such elevation-dependent crater modification can be explained.

Intrusion Models for Venus

Extending the lunar model for igneous crater modification to Venus, the identified most structures and floor plate uplifts suggest that shallow, crater-centered intrusions lie beneath at least three Veniran craters. In addition, however, since the modeled growth of these intrusions resembles that of terrestrial laccoliths, this model can provide estimates for the magmatic pressure driving deformation and for intrusion depth [Wichman, 1993]. The basis and implications of such model inversions are summarized below for the three identified floor-fractured craters on Venus.

Model Relations

Terrestrial studies of laccolithic intrusions [Johnson and Pollard, 1973; Pollard and Johnson, 1973] have shown that intrusion thicknesses during different stages of laccolith growth can be described by different theoretical relations. For instance, at the onset of vertical intrusion growth (when deformation is dominated by elastic bending), the maximum, central thickness of a circular laccolith \( w \) primarily depends on the magma pressure driving deformation (i.e., the superlithostatic pressure in the intrusion, \( P_d \)); intrusion radius, \( a \); and the effective (flexural) elastic thickness of the overburden, \( T_e \), according to the relation

\[
P_d = \frac{5.33 w_m B T_e^3}{a^4}.
\]

In this equation, \( B \) is the elastic modulus of the uplifted floor materials.

Later, after peripheral failure decouples the uplifted overburden from the surrounding strata, flexure becomes negligible, and further uplift primarily reflects a flotation of the overburden. Hence the final intrusion thickness can be simply related to the driving magma pressure and to magma density [Pollard and Johnson, 1973] by the relation

\[
P_d = w_m \left( \frac{2k}{a^4} + \gamma_m \right),
\]

where \( k \) is the magma yield strength (an essentially negligible quantity) and \( \gamma_m \) represents the unit magma weight in the intrusion.
Figure 5a. Barrymore crater, 50 km diameter, located near Imdr Regio (section of C1-MIDR 45s202;1).

Since elastic deformation over a laccolith should preclude significant thinning of the uplifted crater floor sections, the change in crater depth ($\Delta d_a$) during crater modification should reflect $w_{el}$, and therefore can be used in equation (2) to estimate magma pressures at depth. Then, if magma pressure is essentially constant during intrusion growth, we can use this value to solve equation (1) for the effective elastic thickness ($T_e$) of the uplifted floor plate. Assuming that the uplifted floor plate diameter delineates the size of a crater-centered intrusion ($a$), this value provides a minimum estimate for the depth of intrusion beneath the crater floor [Wichman, 1993].

Model Results

In applying these relations to Venus, we have used the preliminary morphometric relations of Sharpton [1992] for fresh Venusian craters to calculate the initial apparent crater depth ($d_a$) and Magellan altimetry to estimate the present crater floor depth (Table 2). Although our estimated present crater depths therefore may be biased toward shallower values by the low spatial resolution of the Magellan altimeter [Leberl et al., 1991], much of this...
Alternatively, if magma enters the crater from a very shallow magma source, the derived magma pressures may reflect lateral variations in lithostatic pressure resulting from crater topography. The close correlation between derived magma pressures and calculated topographic effects in both Table 4 and in lunar craters [Wichman, 1993] tends to support the last alternative.

Intrusion depths are less well constrained for the Venusian craters than the driving magma pressures. Equation (1) indicates that effective plate thicknesses on Venus range from <1 to ~2 km (Table 4), but these thicknesses assume that the crater floor deforms as a single coherent unit. A layered section of impact melts and breccia units, however, can have an effective flexural thickness ~2-3 times thinner than its actual stratigraphic thickness [Pollard and Johnson, 1973]. Consequently, effective floor plate thickness provides only a minimum estimate for the intrusion depth beneath a crater, and the derived \( T_e \) values on Venus suggest likely intrusion depths range from ~1 km beneath the smallest crater to ~4-6 km at Mona Lisa. Since significant brecciation on the Earth apparently extends to depths of the order of \( 1/5-1/10 \) the crater diameter [Pohl et al., 1978, 1988], these depth estimates (despite their uncertainty) are consistent with crater modification by relatively shallow, breccia-hosted intrusions within all three Venusian craters.

Figure 6a. The crater Mona Lisa, 85 km diameter, located on the edge of Eistla Regio (section of C1-MIDR 30n027;1). Arrows indicate a possible set of exposed, most-bounding fractures near the southern crater wall.

Bias apparently reflects poor recovery of crater rim topography in the Magellan data. Comparison of the Magellan topography data to radar clinometric data in the crater Danilova indicates an error in the crater floor elevation of only ~100 m [Lebert et al., 1991]. Consequently, in sufficiently large craters (crater floor diameters >4-5 altimeter footprints) like Barrymore and Mona Lisa, the Magellan altimeter should provide representative values for both the central crater floor elevation and the apparent crater depth. Although the apparent crater depths in smaller craters may still be biased, the derived depth for the crater in northern Anio Planitia provides at least an end-member case for the model calculations. If the crater depth is underestimated, it places an upper limit on both the floor uplift and the possible magma pressures during crater modification.

For the material constants in Table 3, the modeled uplift values (Table 2) translate into magmatic driving pressures of ~220-370 bars at Mona Lisa, ~240-290 bars at Barrymore, and ~170 bars at the unnamed crater in Aino Planitia (Table 4). Such magma pressures indicate maximum column lengths of ~22 km, ~17 km, and ~10 km, respectively (Table 4), assuming that the magma columns are hydrostatic and that crustal density is 3000 kg/m³. Since the entire magma column is situated in crustal units under these assumptions, these modeled column lengths, in combination with the initial apparent crater depths, indicate crustal thicknesses on Venus of at least 11 to 25 km. Much thinner crustal models also can produce such pressures, however, if (as on the Moon) the magma column extends into subcrustal mantle compositions.

Mona Lisa

Figure 6b. Sketch map of the crater interior at Mona Lisa, showing correlation of our inferred floor plate uplift to the outermost ring of floor peaks and massifs, and also showing the distribution of floor materials within the southern, least flooded, sections of the peripheral moat structure.
Discussion

The common, low elevations of the identified floor-fractured craters on Venus also may indicate a systematic variation in Venusian magmatism as a function of altitude. Specifically, the volcanically embayed craters of Schaber et al. [1992] preferentially occur at much higher elevations than the preserved floor-fractured craters. This distribution could result if floor-fractured craters are rapidly obliterated by regional volcanism at high altitudes, but it also could reflect a transition between different regimes of crater modification. Since the depth and stability of magmatic neutral buoyancy zones (NBZ's) on Venus should vary with atmospheric pressure [Head and Wilson, 1992], elevation-dependent changes in magmatic conditions can produce such a transition by favoring different mechanisms of igneous crater modification.

Atmospheric pressure variations can affect igneous crater modification on Venus in two ways. First, the predicted increase in regional NBZ depths with elevation on Venus [Head and Wilson, 1992] can modulate or preclude the interaction of regional magmas with crater-centered breccias, depending on the relative depth extent of impact brecciation beneath a crater. Second, the effect of atmospheric pressures on vesiculation can also reduce magma densities at higher elevations, and thus affect the stability of magmas within a breccia-defined buoyancy trap. In combination, therefore, these two factors suggest that crater structures are most likely to interact with magmas at low elevations on Venus, whereas regional volcanism may be relatively unaffected by impact structures at higher elevations.

Figure 6c. Magellan topography contours for Mona Lisa showing deepest resolved sections of the crater within the peripheral moat (100-m contour interval).

Figure 7. Comparison of floor plate morphometry from Figures 4, 5, and 6 to the morphometry of other crater structures on Venus. Dots show floor diameters as a function of rim crest diameter for 62 simple and central peak craters on Venus; open circles show floor diameters for eight peak ring basins; and open triangles denote dependence of peak ring diameters on rim crest diameters from Alexopoulos and McKinnon [1992]. The values for Mona Lisa₀ and Mona Lisaₐ indicate diameters for the innermost, central, dark floor region and the outermost massif ring, respectively. Note the close correlation of modeled floor plate diameters to the expected floor diameters at each crater and the similarity in diameter of the central dark floor at Mona Lisaₒ to the trend of peak ring diameters in other two-ring basins.
The possible effects of atmospheric pressure and NBZ depth on igneous crater modification can be illustrated more fully by considering a hypothetical crater 40 km in diameter on the Venusian surface (Figure 8). For reference, such a crater would have an initial apparent depth (after rim/wall collapse) of the order of 1.0-1.25 km, based on the morphometric relations of Sharpton [1992]. Also, the inferred breccia densities at the terrestrial Ries crater [Erlensohn and Polk, 1977; Polk et al., 1978] indicate that significant brecciation (Δρ >100 kg/m³) probably extends an additional 1 to 2 km beneath the floor of such a crater. Because Venusian surface temperatures may anneal deep-seated crustal fractures, brecciation at greater depths may not be preserved and is not considered in this discussion.

Several different styles of crater-centered magmatic activity can occur in this reference crater (Figure 8) due to the range of possible NBZ depths on Venus [Head and Wilson, 1992]. If NBZ depths are less than the apparent crater depth (Figure 8, dike a), then magma extrusion should simply fill the crater. Alternatively, if NBZ depths are large compared to the combination of crater depth and low-density breccia thickness (Figure 8, dike c), then breccia interactions with the magma should be minimized and crater-centered magmatic activity may be negligible. Only if NBZ depths are comparable to the combined thickness of significant brecciation and apparent crater depth are magmas likely to encounter a breccia-defined intrusion site favoring laccolith formation (Figure 8, dike b).

As summarized by Table 5, the NBZ depths calculated for dry basaltic melts with 0.4 wt % CO₂ [Head and Wilson, 1992] are consistent with the inferred variation in crater-centered magmatism on Venus. For the lowest plains surfaces on Venus (~2 km below mean planetary radius (MPR)), the indicated depth of neutral buoyancy is ~1200 m below the surface. Since the apparent depth of a 40-km crater is ~1000-1250 m, magmas in such an NBZ would be no more than ~200 m below the crater floor (Table 5). Hence crater-centered laccolith formation is unlikely. Instead, a combination of volatile exsolution at these depths with even minor magmatic pressures (<20-40 bars) should result in the effusion of such magmas onto the crater floor.

At the other extreme, predicted NBZ depths for the highest highland elevations (MPR+3.6 km) are of sufficient magnitude that magmas may be below the lowest density impact breccias (Table 5). For slightly lower highland elevations (near MPR+1 km), however, the modeled depths of neutral magmas can still allow regional magmas access to a crater-centered NBZ within the breccia lens. The apparent lack of floor-fractured craters at these elevations thus may indicate a depth limit (~1.0-1.5 km) for the extent of significant unannealed brecciation beneath the crater floor. This depth limit is consistent with the inferred thickness of low-density breccias at the Ries crater (~1-2 km), but it also may reflect the consolidating effects of thermal annealing on deep-seated impact breccias over time. Alternatively, the lack of crater-centered intrusions at these elevations also may reflect increased levels of magma vesiculation. Since magmatic devolatilization significantly reduces magma densities at elevations over MPR+1 relative to magmas below MPR [Head and Wilson, 1992], the near-surface region of degassing, positively buoyant magmas should extend to greater depths at higher elevations on Venus. Consequently, depending on the source magma pressures and dike heights outside a crater, breccia-hosted buoyancy traps within the highlands may simply be too narrow to initiate crater-centered intrusion growth.

In contrast, for surface elevations close to the observed floor-fractured craters (MPR-800 m), the modeled depth of neutral buoyancy allows magmas to intersect near-surface regions of significant brecciation beneath a crater, but the atmospheric pressures...
Concluding Remarks

Crater-centered igneous activity on Venus can theoretically produce endogenic crater modification resembling both mare-filled and floor-fractured craters on the Moon, but crater-filling volcanism is more likely due to the effects of buoyancy forces on Venusian magma conduits. Thus, despite the pristine appearance of crater rim and ejecta units, it is possible that many of the dark floored craters on Venus may record crater-centered igneous activity. Nevertheless, although crater-centered intrusions may be rare on Venus, three Venusian impact craters show unequivocal similarities to extensively modified floor-fractured craters on the Moon, and thus are identified as possible sites for crater-centered intrusions.

If these intrusions are then modeled as shallow, laccolith-like magma bodies, theoretical relations can be used to constrain both local magmatic conditions and regional variations in Venusian magmatism. For the identified floor-fractured craters on Venus, such models indicate crater-centered intrusions with magma pressures of ~170-370 bars at depths of ~1-6 km. Further, since the identified floor-fractured craters occur at much lower elevations than the vast majority of externally modified, volcanically embayed craters on Venus, we also suggest that igneous crater modification on Venus is at least partially dependent on local surface elevation.

Regional NBZ depth and Crater Modification

**Table 5. Modeled Neutral Buoyancy Depths**

<table>
<thead>
<tr>
<th>Surface Elevation, km</th>
<th>$d_{NBZ}$, m</th>
<th>$d_{NBZ} - d_{CR}$, m†</th>
<th>Intrusion Level</th>
</tr>
</thead>
<tbody>
<tr>
<td>-1.8</td>
<td>1224</td>
<td>0-225</td>
<td>crater floor</td>
</tr>
<tr>
<td>-0.8</td>
<td>1815</td>
<td>600-800</td>
<td>significant brecciation</td>
</tr>
<tr>
<td>1.2</td>
<td>2516</td>
<td>1300-1500</td>
<td>base of significant brecciation</td>
</tr>
<tr>
<td>3.6</td>
<td>3213</td>
<td>2000-2200</td>
<td>below significant brecciation</td>
</tr>
</tbody>
</table>

Data are after Head and Wilson (1992), assuming dry magma with 0.4 wt% CO₂. * Referenced to mean planetary radius of 6051.8 km. † Difference between NBZ depth and apparent depth of reference crater (40-km diameter).

Figure 8. Sketch illustrating the potential variation of igneous crater modification on Venus as a function of regional NBZ depths. In this illustration, the crater diameter is ~40 km with an apparent crater depth of 1.25 km; and the dikes a, b, and c are centered on NBZ's at depths of ~1225 m, ~2050 m, and ~3250 m, respectively, relative to the surface outside the crater. The hachured region beneath the crater floor denotes a breccia-centered NBZ and is the most likely site for a crater-centered intrusion. Note that crater-centered igneous activity ranges from crater-flooding volcanism to shallow regional magma depths, to a breccia-centered intrusion at slightly deeper NBZ depths, to potentially negligible magma/breccia interactions at large regional NBZ depths. The indicated limits of significant and light brecciation reflect the decay of impact brecciation with depth inferred from terrestrial gravity studies at the Ries impact crater, Germany [Pohl et al., 1978].
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