Strain at radially fractured centers on Venus

Peter M. Grindrod,¹ Francis Nimmo,² Ellen R. Stofan,¹,³ and John E. Guest¹

Received 9 February 2005; revised 24 August 2005; accepted 22 September 2005; published 2 December 2005.

[1] Radially fractured centers (RFCs) on Venus are distinctive, ~100 km long radiating systems of graben, fissures, and fractures arranged around a central topographic high. We use cycle 1 and cycle 3 Magellan images to determine the depths and wall dip angles of several large radial graben, using two different methods, at four RFCs: Becuma Mons, Didilia Corona, H’uraru Corona, and Pavlova Corona. We record graben depths of 0.3–0.9 km and mean wall dip angles of about 36°, which we interpret as a primary talus slope rather than a fault plane. Using the calculated depths and a more realistic fault dip of 60°, we find extension of between 0.5 and 1 km on graben between about 5 and 10 km wide. We use this extension to estimate the hoop strain at each RFC. Maximum measured strains on the flanks are typically between ~1 and 4% and may be up to 7% toward the center. We conclude that this strain is too large to be explained by current plume-related uplift models and also by magma chamber inflation. We conclude that dikes are probably responsible for the majority of the strain at the large radial graben.


1. Introduction

[2] The Magellan mission to Venus produced synthetic aperture radar (SAR) images of 98% of the entire surface. These images revealed 163 distinctive radiating systems of graben, fissures and fractures [Grosfils and Head, 1994]. These radial systems have been interpreted as both (1) the surface expression of dikes at a shallow depth below the surface [e.g., McKenzie et al., 1992; Grosfils and Head, 1994; Ernst et al., 1995] and (2) faults forming as the result of domical uplift [e.g., Stofan et al., 1991; Janes et al., 1992; Cyr and Melosh, 1993]. The radiating systems have previously been found at, and labeled as, novae [Krassilnikov and Head, 2003], arachnoids [Crumpler and Aubele, 2000; Krassilnikov, 2002], giant radiating dike swarms [Ernst et al., 1995], coronae [Stofan et al., 1992, 2001a], and large volcanoes [Head et al., 1992; Crumpler et al., 1997]. We use the term radially fractured centers (RFCs) [Schubert et al., 1991] as they appear to be the result of common processes found at a variety of feature types.

[3] RFCs are characterized by graben and fractures arranged radially around a central region, often with concentric structures surrounding the central region [Ernst et al., 1995]. The radial fractures extend from 40 to 2000 km from the central region, with an average maximum radius of ~325 km [Grosfils and Head, 1994]. The topography of the central region is highly variable, but over half have positive topography [Grosfils and Head, 1994; Krassilnikov and Head, 2003]. The radius of the radial fractures is on average 2.5 times longer than the radius of the associated topography [Grosfils and Head, 1994]. Fractures close to the central region are often graben with resolvable walls and floors, which narrow to a single radial bright lineament with increasing distance from the center [Ernst et al., 1995].

[4] Two interpretations have been proposed for the formation of RFCs: (1) diking and (2) uplift. The first hypothesis describes radial graben and fractures as the result of dike propagation at shallow crustal depths [McKenzie et al., 1992; Grosfils and Head, 1994; Ernst et al., 1995; Koenig and Pollard, 1998]. Recent studies of surface stresses and deformation associated with dikes rely on numerical calculations and field studies of terrestrial dikes [e.g., Pollard et al., 1983; Mastin and Pollard, 1988; Rubin and Pollard, 1988; Rubin, 1992; Gudmundsson, 2003]. The second hypothesis interprets RFCs to be the result of fracturing due to uplift, caused by the impingement of rising mantle diapirs on the upper part of the lithosphere, as part of the corona-forming process [e.g., Stofan et al., 1991, 1992, 1997; Janes et al., 1992; Squyres et al., 1992; Cyr and Melosh, 1993; Koch, 1994; Koch and Manga, 1996]. However, it is likely that RFCs are the result of a combination of both processes, and, on the basis of comparisons with terrestrial dikes, fractures due solely to uplift are less abundant and more centrally located than those associated with dike emplacement [Grosfils and Head, 1994, 1995, 1996].

[5] This paper focuses on the radial graben located close to the central region, which have been interpreted to be the result of uplift alone. We used full-resolution Magellan SAR images to study 4 RFCs (Figure 1). Becuma Mons, Didilia Corona, H’uraru Corona and Pavlova Corona all have positive central topography (Figure 2) and lava flows,
features which are indicative of, but not exclusive to, large volcanoes on Venus [e.g., Head et al., 1992; Stofan et al., 1992; Crumpler et al., 1997; Stofan et al., 2001b]. We selected these features on the basis of their well-defined radar signatures in cycle 1 and/or cycle 3 images. At each RFC we identify graben which are typically several kilometers wide near the summit region, and decrease in width significantly with increasing distance from the central region (Figure 3). We determine the depths and wall dip angles at these graben for comparison with previous estimates at other normal faults on Venus. We use the derived depth values to calculate the extension that has occurred at each graben and thus the total hoop strain observed at each RFC studied. We then discuss the possible causes of the

Figure 1. Cycle 1 left-looking Magellan SAR image of each RFC studied: (a) Becuma Mons, 34.0°N, 21.9°E; image extents are 32–35.5°N, 20–24°E; (b) Didilia Corona, 19.0°N, 38.0°E; image extents are 17–21°N, 36–40°E; (c) H'uraru Corona, 9.0°N, 68.0°E; image extents are 7–11°N, 66–70°E; and (d) Pavlova Corona, 14.3°N, 38.9°E; image extents are 12–16°N, 36–40°E. All images are Mercator projection with a scale bar of ~100 km. Boxes indicate the location of higher-resolution images in Figure 3.
Figure 2. Topography of each RFC studied: (a) Becuma Mons, (b) Didilia Corona, (c) H’uraru Corona, and (d) Pavlova Corona. Images correspond to those in Figure 1. Scale bars show planetary elevation in kilometers. Black lines show the dominant trend of all the radial fractures at each RFC. Topography data are from Ford and Pettengill [1992].
observed strain, including plume-related uplift, magma chamber inflation and subsurface dikes.

2. Measurements

2.1. Graben Depth and Slope

Depth and slope measurements were complicated by the fact that the graben were too small to be resolved individually in the altimetric data. A good estimate of the depth and slope of the graben can however be calculated by exploiting the incidence properties of the active radar system on Magellan. We calculated the depth and slope of the graben walls by using a stereo [Connors, 1995] and/or symmetry [Weitz, 1993] method depending on coverage.

The stereo method of Connors [1995] was applied to those graben which had both cycle 1 (left look) and cycle 3 (stereo look) SAR coverage. Width measurements of the west facing graben walls were taken parallel to the radar range direction at regular intervals. The range direction for cycle 1 and cycle 3 images was assumed to be 82° east of north for all RFCs studied [Connors, 1995]. Those graben with a near north-south orientation were the most well defined and thus our measurements tend to focus on these graben. From these measurements we were able to calculate

Figure 3. Cycle 1 left-looking Magellan SAR image of large radial graben measured at (a) Becuma Mons, (b) Didilia Corona, (c) H’uraru Corona, and (d) Pavlova Corona. The graben measured are indicated by single white lines and identification numbers which correspond to those given in the results. All images are sinusoidal projection with a scale bar of ~10 km.
graben depth and slope angle as a function of distance from the center of the RFC. As the slopes determined were apparent slopes in the range direction, correction to the true topographic slope was carried out by dividing the tangent of the apparent slope by the cosine of the acute angle between the dip direction of the face and the range direction of the radar [Connors, 1995]. Uncertainties in the depths and slopes are the result of each width measurement having a total uncertainty of ±1 range resolution [Connors, 1995]. The stereo method returns two possible depth and slope values for each measurement point; to resolve this ambiguity we followed the method of Connors [1995]. As no cycle 2 (right look) SAR images were available, and Magellan altimetric echoes were too large ([10], both azimuth and range directions [Connors, 1995]), the viewing geometry ambiguity in each case was resolved using stereoscopic impression. We found that the most common viewing geometry was that the graben walls were foreshortened in both cycle 1 and cycle 3 images, with layover also occurring in cycle 3 images at H’uraru Corona.

[9] Where there was a well-defined radar signal from both graben walls we applied the symmetry method of Weitz [1993], for determining the depth of impact craters, to estimate graben depths and slopes of the graben walls. For the stereo method to be applicable the graben walls must have slopes less than the incidence angle of the radar (i.e., there is no radar layover). As the stereo analysis found that no slopes had been laid over in any of the cycle 1 images studied, we extended this assumption to all slopes for the symmetry method. Width measurements of both graben walls were taken parallel to the range direction, and correction to the true topographic slope was carried out as for the stereo method [Connors, 1995]. Uncertainties in scarp width are a function of range resolution [Connors, 1995], which is about 100 m at the latitudes studied here [Saunders et al., 1992].

2.2. Extension

[9] The minimum extension recorded at each graben is the amount needed to account for the current width of the walls, assuming that the extension is accommodated by normal faults. The minimum amount of extension is therefore given by

\[
\Delta W = \frac{2d}{\tan \theta}
\]

(1)

where \(d\) is the depth of the graben, and \(\theta\) is the fault dip angle. To calculate extension we used the depth measurements calculated from the stereo method and a presumed fault dip angle of 60°, the rationale and implications of which we discuss later.

2.3. Hoop Strain

[10] The total hoop strain accommodated at an RFC as a function of distance from the center was calculated by summing the extension measured at all graben, after interpolating to a common set of distances, and dividing by the circumference around the RFC at a given distance

\[
\varepsilon(r) = \frac{\sum \Delta W(r)}{2\pi r}
\]

(2)

where \(\Delta W(r)\) is the extension as a function of distance from the center, \(r\), at each graben. To estimate the hoop strain at the entire RFC we extended our strain measurements to those graben which have a similar width/length appearance to those that were measured, but which were not defined well enough in the radar images to be measured accurately. This was done by assuming similar values of strain on all the unmeasured graben of similar width, length and general appearance. Inspection of Figure 3 shows that a conservative estimate is that we have only measured half of the graben responsible for the total strain. The possible burial of other large graben by lava flows, or obscuring by subsequent tectonism, means that our total strain estimates are most likely minimums. Measurements of hoop strain were restricted to those distances from the center of the RFC which overlapped for each graben studied, providing a limited but acceptable data window.

3. Results

3.1. Graben Depth and Slope

[11] Figure 4 shows individual graben depths as a function of distance from the center of the RFC for each graben measured. At all features there is a general trend of decreasing depth with increasing distance from the center regardless of which method was used. Depths nearest the summit ranged from about 0.3 to 1.0 km and depths furthest from the summit ranged from about <0.1 km to 0.7 km. Different graben from the same RFC show similar depth ranges and errors. Becuma Mons has the shallowest graben which are also nearest the summit region. Those graben for which both stereo and symmetry methods were taken show good agreement between methods. Graben at both H’uraru and Pavlova Coronae show similar depth values for both methods, with those recorded by the symmetry method showing a greater decrease with distance. The RMS error in depth between the two methods for graben 1 at Didilia Corona is 0.05 km, and 0.03 and 0.06 km for graben 1 and 2 at Pavlova Corona.

[12] Figure 5 shows individual graben wall dip angles as a function of distance from the center of the RFC for each graben measured. Graben wall dip angles recorded by each method have typical values of between 30° and 50°, with the mean slope from both methods being about 36°. The slope values recorded by the symmetry method show a gradual and significant decrease with increasing distance, probably as a result of the increased uncertainties in the measurements. The stereo method results are almost constant with distance and show no overall trend, and probably reflect the true measured slope. Those graben for which both stereo and symmetry methods were taken show good agreement between methods, particularly nearest to the summit, although not as good as the depth measurements. The RMS error in slope between the two methods for graben 1 at Didilia Corona is 1.11°, and 0.99 and 1.02° for graben 1 and 2 at Pavlova Corona.

[13] All the wall slopes were found to have a dip of typically 30–50°, comparable to low-angle normal or detachment faults on Earth, which are thought to be the result of rotation of the stress axes with depth [Wernicke, 1995]. However, studies of terrestrial normal faults suggest that this is probably an unrealistically low fault dip [e.g.,
Jackson and White, 1989], and it is most likely that postextension mass wasting has also contributed to the current shallow dip of the walls [Malin, 1992]. This means that the observed extension and strain could be overestimated. Therefore we assume a primary fault dip angle of $60^\circ$ [e.g., Jackson and White, 1989; Schultz, 1995] when calculating the extension and strain at each graben. Assuming a typical normal fault geometry changes the magnitude of the results by a factor of $\frac{1}{2}$, but does not change the qualitative conclusions.

3.2. Extension

Extension as a function of distance from the center is shown for each graben in Figure 6, both by using the measured slope values and the assumed $60^\circ$ wall slopes. The extension is heavily dependent on the dip of the graben walls, and therefore the extension calculated from the presumed talus slope angles are probably overestimating the extension by a factor of typically 2. The more realistic extension calculated from a constant $60^\circ$ wall dip shows a general trend of decreasing with increasing distance from the center. Typical values of extension of about 1 km are recorded close to the summit at most individual graben, decreasing to about 0.5 km by the termini. Becuma Mons is the exception with extension at the graben ranging typically from about 0.4 to 0.1 km with increasing distance from the center. Again, as a result of depending on the depth, those graben for which both stereo and symmetry methods were used show similar values of extension, with both methods showing the systematic decrease with increasing distance.

3.3. Hoop Strain

The hoop strain as a function of distance from the center of the RFC is shown for each RFC in Figure 7. At each feature both the hoop strain from the measured graben and all likely graben responsible for the observed strain are given for comparison, assuming a wall dip angle of $60^\circ$. Despite the limited data points in each case, there is a definite general trend of decreasing hoop strain with increasing distance from the center of each RFC. Hoop strain values vary for each RFC, but are generally large even at significant distances (e.g., 4% at 30 km for Pavlova Corona), suggesting that in these cases the maximum hoop strain experienced at the center could be of the order of 7%. Becuma Mons has a steeply decreasing hoop strain profile, suggesting that most of the strain is concentrated at the center of the RFC. On the other hand, Didilia Corona has a shallow strain profile at distances of about 90 km, suggesting a broader strain distribution at this feature.

4. Discussion

4.1. Measurements

The depths measured at the graben at the RFCs are comparable to previous high-resolution studies of graben of similar widths [Connors, 1995; Connors and Suppe, 2001]. Both the stereo and symmetry methods for depth determination returned values which were generally in the 0.3–0.9 km range, values which correspond well with the general value of 0.7 km calculated by the stereo method for 4 individual fault transects [Connors, 1995]. Our calcu-
lated depth ranges also correspond well with the 170 non-imbricated surfaces related to normal faults calculated by the stereo method, the majority of which have depths of less than 1 km [Connors and Suppe, 2001]. Because of the lack of erosion on Venus, the depth of the graben is equal to the throw on the underlying fault, and therefore our depth measurements are not affected by any postextension collapse [Connors and Suppe, 2001]. This inference assumes that any postextension collapse only increases the width of the slope, and does not affect the depth of the graben. The faults are therefore assumed to lie in the middle of their talus slopes [Connors and Suppe, 2001]. However, our depth results may represent a lower limit in some cases because of lava flows filling the graben. This is especially important at Didilia and Pavlova Coronae, and to a lesser extent H’uraru Corona, where radar dark volcanic units have infilled the graben to an unknown extent. This effect appears to be greatest at Didilia Corona where a large flow unit has buried most fractures apart from those we measured in the northern quarter. This possible underestimation in depth has the effect of our values of extension and hoop strain being minimum values.

[17] Both the stereo and symmetry methods reveal slopes which are significantly lower than the ~65° expected dip for normal faults according to Anderson’s theory of faulting [Connors and Suppe, 2001]. The slopes derived by the stereo method have a mean angle of 36.0°, which corresponds closely with the mean value of 36.4° calculated by Connors and Suppe [2001], and are within the 31–39° range expected for a talus slope of well-drained, unconsolidated material [Chandler, 1973; Selby, 1993]. Therefore it most likely that the topographic slopes that we have measured are not indicators of primary normal fault geometry, but primary talus slopes from collapsed fault scarps [Connors and Suppe, 2001]. Therefore to determine the extension and hoop strain at the RFCs, the assumption of a constant graben wall slope of 60° does not seem unreasonable [e.g., Schultz, 1995; Golombek et al., 1996; Rathbun et al., 1999; Connors and Suppe, 2001].

[18] Previous estimates of extension at normal faults on Venus have concentrated on individual faults [e.g., Connors, 1995; Connors and Suppe, 2001] or chasmata belts [e.g., Rathbun et al., 1999; Connors and Suppe, 2001; Bleamaster and Hansen, 2004]. Using two different methods, Rathbun et al. [1999] and Connors and Suppe [2001] calculated comparable values of extension (~20 km) at the same large (~200 km wide) rift zones associated with Beta Regio. Bleamaster and Hansen [2004] estimated the cumulative extension across Jana Chasma to be ~2.5 km over approximately 100 km wide rift area. Our results of ~0.5 to 1 km of extension at individual graben of typically 3 to 10 km in width reveal a similar extension to fault width relationship, although obviously much variation exists between individual graben and RFCs.

[19] The hoop strain recorded by the radial graben at the RFCs is generally larger and much more locally confined than other estimates of strain as a result of uplift on Venus. Rathbun et al. [1999] measured the hoop strain and asso-
ciated uplift at three major rifts at Beta Regio, a rift-dominated topographic rise [Stofan et al., 1995]. They recorded strain values of up to 2% and uplift of 3 km, neglecting the central volcanic construction area. The strain at Beta Regio has a profile that is likely to have been about 5% at the centre, if not obscured by volcanic infilling of rifts, and follows a similar trend to that at the RFCs but over a much larger distance, effectively reaching zero by about 1500 km [Rathbun et al., 1999]. The strain we have observed at the RFCs is a minimum and yet is up to about 5% at a distance of ~25 km. If this trend is extrapolated to the center of the features, by using a linear or Gaussian fit, then the strain is likely to be of the order of 5–10% for the 4 RFCs we have studied, regardless of extrapolation method. Therefore it is necessary to consider what mechanism is likely to cause such concentrated high levels of strain.

4.2. Models of Hoop Strain

[20] Here we consider two classes of mechanism which might be responsible for generating the inferred strain profiles. The first class is extension as a result of topographic uplift, caused by either convective plume buoyancy or magma chamber inflation. As we demonstrate below, the likely strains associated with these mechanisms are at least an order of magnitude smaller than the observations. The second mechanism is subsurface diking, leading to surface extension which is consistent with the inferred strains.

[21] The first mechanism that we consider is that of plume-related uplift. We compare our results with those of Janes et al. [1992] and Janes and Squyres [1993], who model the elastic deformation in terms of displacement and stresses at several RFCs due to a buoyant mantle diapir at depth. They interpret the entire topography present at the features to be the result of uplift and then determine the best fit parameters, and resultant stresses, needed to produce the corresponding amount of uplift. The topographic shapes of the features studied are similar to, but slightly larger than, those studied here, with central heights of between about 0.9 and 2 km and diameters of about 200–400 km, compared with our values of between about 0.5 and 1 km in height and diameters of 150–200 km. To produce the required topographies, Janes et al. [1992] and Janes and Squyres [1993] predict maximum radial and hoop stresses at the centre of between 100 and 250 MPa. We can compare these results to our inferred strain levels values by using the relationship between hoop strain, $\varepsilon_h$, and radial and hoop stress, as given by [Timoshenko and Goodier, 1970],

$$\varepsilon_h = \left( \frac{1}{E} \right) (S_r - \nu S_h)$$  \hspace{1cm} (3)

where $E$ is Young’s modulus, $\nu$ is Poisson’s ratio, $S_r$ is the radial stress, and $S_h$ is the hoop stress. Using a value of 250 MPa for both radial and hoop stress [Janes et al., 1992;
Janes and Squyres, 1993], and assuming typical values of 5 × 10^{10} \text{ N m}^{-2} [Jousset et al., 2000] and 0.25 [McTigue, 1987] for Young’s modulus and Poisson’s ratio respectively, gives us a maximum value of hoop strain of 0.3% at the center of the feature. This value of hoop strain is an order of magnitude smaller than that observed at the RFCs, and would produce a central high of about 2 km [Janes et al., 1992], more than twice that observed at the RFCs.

[22] Obviously it is possible that parameters different to those used by Janes et al. [1992] and Janes and Squyres [1993] may produce the required hoop strain at the surface. Petrological differences are not expected to have a major effect on the results of the uplift model [Janes et al., 1992], and thus it is more likely that any change would be the result of differing lithospheric properties. To increase the surface strain requires a reduction in the elastic thickness of the lithosphere. Hoogenboom et al. [2004] determined the elastic thickness at Didilia and Pavlova Coronae as 0 and 44 km, with top-loading and bottom-loading signatures, respectively. The low elastic thickness at Didilia might permit a larger strain to be generated by plume-related uplift, but the elastic thickness value at Pavlova greatly exceeds the value of 8 km used by Janes and Squyres [1993] and suggests that regional uplift is not an important contributor to strain in this latter case. Although elastic thickness measurements are not available for Becuma Mons and H’uraru Corona, the lack of broad-scale positive topography suggests that plume-related strain is relatively unimportant for these features.

[23] Different models of plume-related uplift are difficult to compare as stress results are not usually published. However, some models [e.g., Koch, 1994; Koch and Manga, 1996; Musser and Squyres, 1997] predict similar radial and hoop stress distributions to Janes et al. [1992] and Janes and Squyres [1993] which, when dimensional, return even smaller values of hoop strain. Other models of uplift applicable to doming on Venus concentrate on large-scale hot spots and thus predict diameters of uplift much larger than the width of the RFCs [e.g., Nimmo and McKenzie, 1996; Smrekar and Parmentier, 1996; Smrekar and Stofan, 1997; Vezolainen et al., 2004]. We therefore need to consider a more localized source of uplift than plumes.

[24] An alternative source of uplift which we consider here as a possible source of the inferred strain is that caused by the inflation of a magma chamber. We modeled the strain at the surface as the result of uplift caused by a pressurized spherical cavity (representing a magma chamber) in an elastic half space [McTigue, 1987]. Despite the simplification of a spherical magma body, which in reality is more likely to be a complex array of tubes and magma lenses in a much larger volume of rock which deforms plastically.
[Davis et al., 1974], a spherical point source often represents deformation at volcanoes quite well [e.g., Murray and Guest, 1982; Berrino et al., 1992; Williams-Jones and Rymer, 2002; Sturkell et al., 2003]. The radial, \( S_r \), and hoop, \( S_h \), stresses are given by [McTigue, 1987]

\[
S_r = -2Pe^3 \left[ \frac{(2-v)}{(p^2 + 1)^{3/2}} - \frac{3}{(p^2 + 1)^{5/2}} \right]
\]

(4)

\[
S_h = 2Pe^3 \left[ \frac{(1-2v)}{(p^2 + 1)^{3/2}} + \frac{3v}{(p^2 + 1)^{5/2}} \right]
\]

(5)

where \( e \) is the ratio of sphere radius, \( a \), to depth to the center of the sphere, \( f \), \( p \) is the radial distance, \( r \), divided by \( f \), and \( P \) is the pressure increase in the sphere.

[25] Substituting for radial and hoop strain in equation (3) gives the hoop strain as

\[
\varepsilon_h = \frac{2Pe^3}{E} \left[ \frac{(1-v^2)}{(p^2 + 1)^{3/2}} \right]
\]

(6)

[26] The modeled hoop strain can thereby be calculated as a function of \( r \) and compared to the hoop strain measured at the RFCs. We set \( E \) at a constant value of \( 5 \times 10^{10} \) N m\(^{-2}\) [Joussel et al., 2000], \( v \) at 0.25 [McTigue, 1987], and \( P \) at 200 MPa [McGovern and Solomon, 1998], and then solved for the best fit sphere radius \( a \) and sphere depth \( f \).

[27] The magma chamber inflation model fails to recreate the observed hoop strain levels with valid parameters. In each case the returned best fit parameters include sphere radii which are larger than the depth to their center, meaning that the model is invalid. Only by increasing the magma chamber pressure or reducing the strength of the material can the inflation model account for the strain. McGovern and Solomon [1998] used a finite element method to model the evolution of stress and deformation in growing volcanic edifices on Venus. They found that a pressure increase of 200 MPa in a magma chamber with a depth to center of about 2 km was sufficient to cause rotation of the principal stress axes so that magma ascended to the surface was favored. This value however is much higher than terrestrial chamber pressures [e.g., Giudmundsson et al., 1997], but is likely to be lower if smaller increments of volcanic building were considered [McGovern and Solomon, 1998]. There is no justification in increasing the magma chamber pressure in the inflation model above already exceptionally high levels. It is possible, however, that the strength of the material has been reduced to some extent because of preexisting fracturing, although it is not likely that it would reduce the Young’s modulus to sufficiently low levels. Again using Pavlova Corona as an example, the Young’s modulus would need to decrease by one or two orders of magnitude to account for the inferred hoop strain with pressure increases of 200 and 20 MPa respectively.

[28] Therefore both the plume and magma chamber-related uplift models fail to account for the high levels of concentrated hoop strain we have observed at the 4 RFCs. It is therefore highly likely that subsurface dikes have made a major contribution to extension and strain at the large radial graben, as well as to the other long, narrow presumably associated fractures. Analytical studies of the location of tensile stress maxima associated with dikes show that deep dikes are more likely to produce a single large graben, whereas shallow dikes produce a pair of narrow fissures either side of the dike [e.g., Pollard et al., 1983; Mastin and Pollard, 1988; Koenig and Pollard, 1998]. This observation is similar to the fracture patterns at the 4 RFCs studied here, and at other similar features, where there is often a transition from a single, large, radial graben to one or two bright lineaments [e.g., Grosfils and Head, 1994; Koenig and Pollard, 1998]. This suggests that the dike surfaces propagate at a near constant horizontal level, not far from the surface, and possibly from a common source such as a central magma chamber.

[29] Although no definite dike dimensions can be determined from the graben properties alone, Mège and Masson [1996] give a range of estimates of dike width and depth which have been made from previous analytical and analogical models. The minimum dike width is roughly 1.5 times the extension [Mastin and Pollard, 1988] while the maximum could be up to about 4.5 times the extension [Pollard et al., 1983]. Taking a general value of extension at a single graben of 1 km gives the width of the dike responsible of between 1.5 and 4.5 km. This is two orders of magnitude larger than typical dike widths on Earth [Ernst et al., 1995], and one order of magnitude than those postulated for the Tharsis region on Mars [Wilson and Head, 2002], although kilometer-sized dike widths might be present at the Great Dike of Zimbabwe on Earth and Valles Marineris on Mars [McKenzie and Nimmo, 1999]. Similarly, the depth to dike top is between about a sixth of the graben width to a full graben width beneath the surface [Mastin and Pollard, 1988]. Taking a general graben width of 5 km gives the depth to the top of the dike of between about 0.8 and 5 km. As none of the graben in our study are sources of lava flows it is likely that the dikes responsible for the graben have insufficient magma pressures to penetrate the surface, and probably lie toward the deeper end of the range.

[30] No definite values of dike heights can be obtained from the graben dimensions. However, it is likely that the dike height is much larger than the depth to the top of the dike [Mège and Masson, 1996]. Assuming conservative dike dimensions of 1 km width, 10 km height and 50 km length, and a rectangular dike cross section, results in a dike volume of 500 km$^3$. If we assume there are six dikes responsible for the strain we have observed, then the total volume of magma in these dikes alone is 3000 km$^3$, just over half the volume of the positive topography feature at Pavlova Corona, assuming a conic profile [e.g., Stefan et al., 2001b] of 0.8 km height and 80 km radius. Therefore the magma contained within the inferred dikes could be responsible for a significant fraction of the volcanic edifice construction. The volume of magma is similar to previous estimates of magma chamber volumes at Hatsheshupt Patera and Kali Mons [Cook et al., 1998], and corresponds to a sphere of about 5 km radius. Unfortunately, since none of the RFCs we have studied have caldera structures, it is not possible to independently verify the magma chamber dimensions.
For a dike to propagate, the magma overpressure in the dike should be comparable to the lithostatic pressure. Although terrestrial magma overpressures for dikes are typically \(\sim 1–20\) MPa [e.g., Pollard et al., 1983; Pollard, 1987; Cervelli et al., 2002], petrological studies of Mount St. Helens indicate magma chamber pressures of 160–300 MPa [Gardner et al., 1995]. The latter values are comparable to the estimates obtained by McGovern and Solomon [1998]. An overpressure of 200 MPa would permit a dike to ascend to ascend from \(\sim 7\) km depth and suggests that dike propagation is mechanically plausible.

It is also possible to determine if dikes of the inferred widths can form by studying the critical width for dike ascent to occur. Petford et al. [1994] give an expression for this critical width, \(w_c\), as

\[
w_c = 1.5 \left[ \frac{c_p (T_m - T_s)}{L (T_m - T_r)} \right]^{\frac{1}{2}} \left( \frac{\mu s H}{\kappa \Delta p} \right)^{\frac{1}{4}}
\]

where \(c_p\) is the heat capacity of the magma, \(T_m\) is the melting temperature, \(L\) is the latent heat, \(T_i\) is the initial temperature of the magma, \(\mu\) is the viscosity, \(\kappa\) is the diffusivity, \(H\) is the dike height, \(g\) is gravity, \(\Delta p\) is the density contrast between the magma and the surrounding rock and \(Ts\) is the surface temperature. Parameters are defined by using typical basalt values of \(c_p = 1200\) J kg\(^{-1}\) K\(^{-1}\), \(T_i = 1500\) K, \(L = 560\) kJ kg\(^{-1}\), \(T_m = 1600\) K, \(\mu = 104\) Pa s, \(\kappa = 8 \times 10^{-2}\) m\(^2\) s\(^{-1}\), and \(\Delta p = 100\) k Pa m\(^{-3}\) [Nimmo, 2000], and setting \(g = 8.87\) m s\(^{-2}\) and \(T_s = 740\) K. For a dike of 10 km height, the critical width for dike ascent to occur is about 7 m. This calculated critical dike width is much smaller than the observed graben widths and again suggests that dike propagation is mechanically feasible.

Nimmo and McKenzie [1996] argue that time-averaged melt generation rates of individual Hawaii-scale plumes on Venus probably does not exceed \(\sim 0.4\) km\(^3\)/yr, and are more likely to be close to 0.1 km\(^3\)/yr. Using a melt generation rate of 0.1 km\(^3\)/yr would result in the volume of magma estimated above forming in 30 kyr. Lipman [1995] gives the magma supply rate for Mauna Loa as 0.028 km\(^3\)/yr. Using this more conservative estimate for the volume of magma estimated above forming in 30 kyr. We used the extension to estimate the hoop strain recorded at 4 RFCs, and found varying levels of strain at each feature, but strain levels are generally high and concentrated within a narrow region. The observed strain is too large to be explained by previous plume models of uplift and also by magma chamber inflation. We therefore conclude that subsurface dikes must have made a significant contribution to the formation of the large radial graben at the RFCs, as well as being responsible for the numerous smaller fractures present.

5. Conclusions

We have measured the depths and wall dip angles of several large radial graben using two different methods. We find depths of \(<0.1\) to \(>1\) km, with most within the range of about 0.3 to 0.9 km. We found the dip angles of the graben walls to be about 36\(^\circ\), consistent with primary talus slopes from collapsed fault scarps. By assuming an original fault dip angle of 60\(^\circ\) we determined the extension at radial graben to be of the order of 0.5 to 1 km for graben typically between 5 and 10 km wide. We used the extension to estimate the hoop strain recorded at 4 RFCs, and found varying levels of strain at each feature, but strain levels are generally high and concentrated within a narrow region. The observed strain is too large to be explained by previous plume models of uplift and also by magma chamber inflation. We therefore conclude that subsurface dikes must have made a significant contribution to the formation of the large radial graben at the RFCs, as well as being responsible for the numerous smaller fractures present.

Acknowledgments. This work was supported by a Natural Environment Research Council studentship (PMG) and the Royal Society (FN). We thank Sue Smrekar and an anonymous reviewer for comments which greatly improved the paper.

References


P. M. Grindrod, J. E. Guest, and E. R. Stofan, Department of Earth Sciences, University College London, Gower Street, London WC1E 6BT, UK. [p.grindrod@ucl.ac.uk]

F. Nimmo, Department of Earth and Space Sciences, University of California, 1156 High St., Santa Cruz, CA 95064, USA.