Tharsis-radial graben systems as the surface manifestation of plume-related dike intrusion complexes: Models and implications

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Several zones of graben (Memnonia, Sirenum, Icaria, Thaumasia, and Claritas Fossae) extend radially away from the Tharsis rise in the southern hemisphere of Mars for distances of up to 3000–4000 km. These graben systems are commonly interpreted to be related to regional tectonic deformation of the Tharsis rise associated with either upwelling or loading. We explore the possibility that these giant Tharsis-radial graben systems could be the surface manifestation of mantle plume-related dike intrusion complexes. Emplacement of dikes causes near-surface stresses that can produce linear graben, and lateral dike emplacement related to plumes on Earth can produce dike swarms with lengths of many hundreds to several thousands of kilometers. We develop a Mars dike emplacement model and explore its implications. We find that the properties (outcrop patterns, widths, and depths) of the extensive Tharsis-radial graben systems are consistent with an origin through near-surface deformation associated with lateral propagation of magma-filled cracks (dikes) from plumes beneath Tharsis, particularly beneath Arsia Mons and Syria Planum. Such dikes are predicted to extend through the crust and into the upper mantle and can have widths of up to several hundred meters. Analyses of summit caldera complexes on Martian volcanoes imply that the magma supply from the mantle into shallower reservoirs is episodic on Mars, and we interpret the graben systems to be large swarms of laterally emplaced giant dikes resulting from the tapping of melt from episodically rising mantle plumes in a buffered magma supply situation. The magmatic interpretation of the Tharsis-radial graben potentially removes one of the conundrums of Tharsis tectonics in which it appeared necessary to require two distinct modes of support for Tharsis in order to explain the presence of radial graben on both the elevated flanks (attributed to isostatic stresses) and outside the rise (more consistent with flexure): dikes capable of forming the observed graben can be emplaced under a wide range of stress fields, including zero stress. The fact that almost no eruptive features are associated with the graben further restricts the ranges of magma density to values between ~3100 and 3200 kg m$^{-3}$ and crustal stress to tensions less than ~30 MPa. Eruptions from giant dikes would be more likely to occur in regions where the crust was thinner, such as the northern lowlands, providing a potential mechanism for emplacement of recently documented Early Hesperian volcanic plains (Hr) there. Dike-related graben systems represent efficient mechanisms of lateral heat transfer in the crust and near-surface environments. Lateral dike intrusions could penetrate the cryosphere and cause melting and release of groundwater, as in the Mangala Valles area, and could also drive hydrothermal circulation systems. The geometries of such dike systems will create barriers which are likely to influence regional to global groundwater flow patterns, which may help to explain the abundance of outflow channel sources in eastern Tharsis. Improved knowledge of the Martian crust and mantle density structure will help to refine this analysis and to provide estimates of the magma densities for dikes underlying specific graben.


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1. Introduction

[1] A number of zones of graben and tension cracks extend generally radially away from the Tharsis rise in the southern hemisphere of Mars [Wise et al., 1979a, 1979b; Plescia and Saunders, 1982], including Memnonia Fossae, Sirenum Fossae, Icaria Fossae, Thaumasia Fossae, and Claritas Fossae (Figures 1 and 2). These graben and tension crack systems consist of from three to more than a dozen linear structures that occur parallel to one another, are discontinuous to continuous, sometimes en echelon, and extend in straight to broadly arcuate patterns for up to 3000-4000 km from central Tharsis (Table 1). In the highlands distal to Tharsis, they cut across a wide range of highland units of Noachian and Hesperian age, and toward Tharsis they are embayed and covered by various units of Hesperian and Amazonian age, but in some places some of them cut these latter units [Scott and Tanaka, 1986; Tanaka et al., 1992]. These data strongly suggest that these graben and tension crack systems formed throughout the period of evolution of Tharsis and the emplacement of its extensive volcanic cover [Tanaka et al., 1992; Banerdt et al., 1992].

[3] Graben and fractures arrayed radially around a central feature have been attributed to (1) faulting due to regional strain associated with isostatic, dynamic, or thermal uplift [e.g., Wise et al., 1979b; Frey, 1979; Banerdt et al., 1982; Sleep and Phillips, 1985; Tanaka and Davis, 1988; Cyr and Melosh, 1993], (2) radial dike emplacement and associated near-surface faulting [e.g., Head and Wilson, 1992; Parfitt and Head, 1993; Grosfils and Head, 1994a, 1994b, 1995], and (3) a combination of these factors [e.g., Grosfils and Head, 1994a; Ernst et al., 1995; Mège and Masson, 1996a, 1996b].

[4] In the Tharsis region of Mars the most common interpretation of the extensive radial graben and fracture systems is that they are related to lithospheric deformation associated with the formation of Tharsis itself. Comparison of the major radial graben systems to stress models [e.g., Banerdt et al., 1992] has shown that more than one lithospheric deformation mechanism is required to explain the enormous extent of the concentric extensional stresses required by the observed radial graben system. For example, concentric extensional stresses on top of the Tharsis rise are produced only by isostatic models, but these produce compressional stress off the rise in the area of the observed graben swarms owing to the large radius of uplift with respect to the planetary radius. In contrast, only flexural loading predicts concentric extensional stresses off the rise. Thus isostatic stresses are required to produce the observed radial graben on top of the rise, and flexural loading stresses are required to produce the graben on the flanks and distal portions of the rise. Several workers have attempted to reconcile these predictions with the temporal sequence of formation of radial graben [e.g., Watters and Maxwell, 1983] in the history of Tharsis [e.g., Golombek and Phillips, 1983; Sleep and Phillips, 1985; Banerdt and Golombek, 1989]. Detailed mapping of the structures and their temporal relationships, and comparison to the history of Tharsis, has revealed, however, the conundrum that many sets of radial fractures that extend the entire distance from the central regions of Tharsis to its periphery appear to have formed simultaneously, or at least closely interspersed in time.

[5] As pointed out by Banerdt et al. [1992, p. 292], “it is not clear how to form such extensive radial fractures in a single event, when the stress models seem to require two distinct events. Resolution of this apparent paradox is required before self-consistent scenarios can be constructed that attempt to link the stress systems with the evolution of the lithosphere and asthenosphere in the Tharsis region.”

More recent analyses using Mars Global Surveyor data suggest that it may be possible to explain most of the faults (both on and off the rise) in a unified fashion by flexure alone, with the faulting within Tharsis being caused by regional deformation not resolved by earlier data sets [Banerdt and Golombek, 2000].

[6] We suggest that another potential solution is the lateral emplacement of magma-filled cracks, that is, dikes, in which the magma pressure, rather than the regional stress field, is the dominant factor causing near-surface stresses that produce linear fractures and graben [e.g., Head and Wilson, 1993; Wilson and Head, 2000]. We assume that the dikes are fed by magma rising from the head of a mantle plume. The stresses caused by plume-related uplift (i.e., least principle stress horizontal and oriented circumferentially to the plume center) define the initial orientations of the intrusions, forming dikes rather than sills. Thereafter the magma pressure in the dikes dominates the regional stress fields and ensures that the intrusions continue to propagate laterally as dikes. Lateral dike emplacement related to extensive mantle plumes can produce dike swarms whose lateral dimensions are measured in hundreds to thousands of kilometers (see reviews by Ernst et al. [1995, 2001]). For example, the Mackenzie dike swarm, the largest known on Earth, consists of a radiating 100° fan of dikes that extend more than 2500 km from the source region [e.g., Fahrig and West, 1986]. These dikes were emplaced in the Canadian Shield about 1.27 Gyr ago in association with a mantle plume; magnetic fabric data show that vertical magma flow in the dikes characterized the central upwelling region and lateral flow dominated at all distances beyond it [Ernst and Barager, 1992].

[7] On the basis of the commonly observed correlation of radial dike emplacement and surface faulting [e.g., Pollard et al., 1983; Pollard, 1987; Rubin, 1992], we explore the possibility that the giant Tharsis-radial graben systems could be the surface manifestation of plume-related dike intrusion complexes [Mège and Masson, 1996a]. We first document the basic characteristics of the graben and fracture systems, then develop a model of giant dike emplacement under Mars conditions, and finally explore the implications of our results.

2. Tharsis Graben and Fracture Zones

[8] Among the several types of extensional structures observed on Mars [e.g., Davis, 1990; Banerdt et al., 1992], two types, simple graben and tension cracks, dominate the southern radial circum-Tharsis systems. Simple graben are linear, straight-walled depressions with a flat floor and are characterized by inward dipping normal faults which have typically undergone displacements of tens to hundreds of
meters. The simple geometry, consistent over hundreds of kilometers, suggests that the two faults are of equal importance, perhaps initiating at a common point and then propagating to the surface [Rubin, 1992]. These simple graben are long (tens to many hundreds of kilometers), relatively narrow (a few kilometers wide), and characterized by spacings of a few tens of kilometers (and locally much less, as in Claritas Fossae and Ceraunius Fossae). Tension cracks, narrow, deep, v-shaped structures, are morphologically distinct from simple graben in that they lack a flat floor [Tanaka and Golombek, 1989].

[9] As pointed out above, several zones of graben and tension cracks extend generally radially away from the Tharsis rise in the southern hemisphere of Mars [Wise et al., 1979a, 1979b; Plescia and Saunders, 1982] (Figures 1 and 2). In order to provide a general characterization of the systems or zones, we have divided them into several groups and subgroups (Figure 2 and Table 1) and have used Viking and MGS image and altimetry data to document their type, system length, segment length, spacing, width, depth, cross-sectional profile and structure (Figures 3–10), and associations.

[10] Memnonia Fossae (Figures 1–4) consist of 4–6 parallel and tangential graben, fractures, and pitted linear structures, typical separated by about 20 km and aligned radially to Arsia Mons, extending from 1100 to 2700 km from the summit of Arsia Mons. Widths are typically 1–2 km, and depths are 100–200 m but can be up to 450 m. Most of the structures in this system cut Noachian and Early Hesperian-aged units distal to Tharsis, but one or two cut Hesperian-Amazonian (AHt3) and Early Amazonian (AHt4) units flanking Arsia Mons [Scott and Tanaka, 1986] (Figure

Figure 1. Mars Global Surveyor (MGS) Mars Orbiter Laser Altimeter (MOLA) [Smith et al., 2001] color-coded topographic map superposed on a MOLA topographic gradient map showing topographic details. Central Tharsis topography above 4000 m is shown in red. Compare with Figure 2, which is a sketch map of the same area and shows the distribution of structures radial to Tharsis and geological units. Note the major transition in surface topography at the margins of the Tharsis rise (below about 2000 m), where radial flows from Tharsis volcanoes give way to ancient deformed and more heavily cratered terrain. Mercator projection.
2) Sirenum Fossae (Figures 1, 2, 5, and 6) consist of 1–6 parallel graben, fractures, and linear pit chains with a typical spacing of about 12–14 km and arrayed radially to Arsia Mons, extending from 1100 km to at least 3700 km from the Arsia Mons summit. They have typical widths of about 1–2 km and depths of 100–300 m, but ranging up to 500 m. Structures of Sirenum Fossae cut Noachian and Early Hesperian-aged units distal to Tharsis, and none intersect the younger flanking Arsia flows of Daedalia Planum. An unnamed set of graben and fissures is located higher on the flanks of Arsia (Figures 2 and 7), extends about 1600 km from the summit of Arsia, and cuts units ranging from Ht1 to At5. These structures range in depth from 50 to 70 m, with one up to several hundred meters deep. A few short segments that parallel this trend are seen distally to the occurrence in Figure 5 about 2000–2800 km away from the summit of Arsia Mons in the Noachian-Hesperian terrain (see Figure 2).

Icaria Fossae form an extensive system of fractures and graben that are oriented radially and subradially to Syria Planum (Figures 1, 2, and 8), extending from about 800 to 4000 km from the summit of Syria Planum. This is a very extensive system that predominantly cuts Noachian and Early Hesperian-aged units. The detailed segment shown in Figure 6 illustrates the relatively high density of these features locally and shows that their depths are ~50–200 m. Claritas Fossae (Figures 1, 2, and 9) form a dense and extensive system of graben and fractures that are generally radial to Syria Planum, extending in a SSE direction from about 1000 to 2800 km from the summit of Syria Planum. Structures there are closely spaced and are generally 25–100

Figure 2. Distribution of graben and fracture systems in southern Tharsis. Dark lines radial to Tharsis show locations of graben and fracture systems mapped at 1:15 million scale [after Scott and Tanaka, 1986; Greeley and Guest, 1987]. Stars indicate location of summits of Arsia Mons (left) and Syria Planum (right) edifices, to which most of the structures are radial. Compare this figure to Figure 1 and note that the radial graben and fracture systems extend well beyond the presently preserved high topography of Tharsis. Boxes show the location of Viking Orbiter enlargements and MOLA altimetry profiles shown in Figures 3–10. Letters represent geologic unit designations; Am is Amazonian Medusae Fossae Formation, and others represent Tharsis-related volcanic units of Hesperian and Amazonian age discussed in the text. A significant part of the area is composed of more heavily cratered units distal to Tharsis and not labeled (compare to Figure 1). Units and boundaries are from Scott and Tanaka [1986] and Greeley and Guest [1987]. Mercator projection.
m deep. Thaumasia Fossae (Figures 1, 2, and 10) consist of a closely spaced set of graben located primarily south of the Thaumasia Rise. Structures there are about 25–150 m deep.

[11] In summary, these graben and tension crack systems consist of from three to more than a dozen linear structures (Figures 2–10) that occur parallel to one another, are discontinuous to continuous, sometimes en echelon, and extend in straight to broadly arcuate patterns for up to 3000–4000 km from central Tharsis (centered primarily at Arsia Mons and Syria Planum) (Figures 1 and 2; Table 1). These data provide a basis on which to assess the plausibility of dike emplacement as a mechanism for formation of the graben and fracture systems in southern Tharsis.

### 3. Giant Dike Model Development

[12] Current understanding of the emplacement of dikes from a rising mantle plume head implies that the following stages occur (Figure 11). The partial melt from a finite vertical extent of the plume head is segregated by compaction into a region of much smaller vertical extent [Maaloe and Scheie, 1982; McKenzie, 1985; Scott and Stevenson, 1986]. This region of concentrated melt is positively buoyant in the host rocks from which it was derived and so can propagate upward as a diapiric body. As it rises it encounters cooler surrounding rocks, and the rheological response of these rocks to its passage becomes progressively less viscoplastic and more viscoelastic. Rubin [1993a] shows how the aspect ratio of the magma body then changes from its original relatively equant shape, becoming progressively more elongate in the plane at right angles to the direction of least principle stress in the host rocks, here horizontal and oriented circumferential to the center of the plume, so that it assumes a more nearly dike-like form. The upper part of the magma body will also encounter increasingly less dense host rocks as it rises: on both Mars and Earth the surface rocks have a significant amount of pore space and become progressively more compacted with increasing depth [Clifford, 1993; Wilson and Head, 1994], and on both planets a significant and probably rapid increase in density occurs at the crust-mantle interface [Zuber et al., 2000]. Thus eventually the magma in the upper part of the rising magma body must reach a level of neutral buoyancy (Figure 11). At all depths shallower than this level the magma is negatively buoyant and cannot rise farther unless it is assisted in some way. The source of such assistance is discussed below.

[13] A critical issue is the interplay between the effects of the host rock rheology and the density contrast between the magma and the host rocks on the behavior of the magma body as it enters the region where the elastic component of the host rock rheology begins to dominate its motion. At some point the combination of buoyancy-driven magma rise speed (and hence strain rate imposed on the surrounding rocks) and host rock rheology becomes such that brittle failure of the host rocks occurs and a true elastic dike begins to propagate upward. Once such a fracture begins to propagate from some location near the top of the diapir (Figure 11), the strain rate imposed on the host rocks will be many orders of magnitude greater than that associated with the slow rise of the diapir. The rheology of the host rocks will be both stress and strain rate dependent, and so once the host rocks have begun to react elastically, rather than plastically, at the dike tip, this tip will continue to propagate elastically. The growing dike will therefore be able to propagate not only upward into cooler rocks, which would in any case be expected to react elastically, but also sideways and downward into hot rocks, which previously reacted in a plastic manner to the slow rise of the diapir.

[14] We assume that the magma within the diapir and within the growing dike suffers negligible cooling and exhibits Newtonian rheology (we show later that the widths of giant dikes penetrating close to the surface are always large enough that this is true; a similar conclusion was reached in a detailed study of the lateral propagation of giant dikes on Earth by Fialko and Rubin [1999]). The rise of the diapir will have been so slow that there will be minimal stress differences supported across its boundaries, and so we can equate the pressure inside the diapir to the horizontal component of the lithostatic load (approximated as a hydrostatic pressure [see McGarr, 1988]) at some level. These two pressures cannot be equal at all depths because, except exactly at a neutral buoyancy depth, the densities of the magma and the host rocks are different and therefore the vertical gradients of the pressures within them are different. This subtlety would be important in any detailed study of the motion of diapirs in planetary mantles but for our

### Table 1. Measured Characteristics of Radial Graben and Fracture Systems in Southern Tharsis

<table>
<thead>
<tr>
<th>Area (Name and Figure)</th>
<th>Map (MC Designation and I-Number)</th>
<th>Type (G, Graben; F, Fracture; P, Pits)</th>
<th>Range of Widths, km</th>
<th>Range of Depths, m</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Memnonia Fossae, Figure 3</td>
<td>MC-16 SE, I-1187 MC-16 SW, I-1188</td>
<td>G, F, P</td>
<td>1–2</td>
<td>100–200, up to 400</td>
<td>mean spacing ∼20 km (18 in ~350 km); Mangala source</td>
</tr>
<tr>
<td>Memnonia Fossae, Figure 4</td>
<td>MC-16 SW, I-1188</td>
<td>G, F</td>
<td>1–2</td>
<td>100–200, up to 300</td>
<td>mean spacing ∼23 km (14 in ~320 km)</td>
</tr>
<tr>
<td>Sirenum Fossae, Figure 5</td>
<td>MC-16 SE, I-1187 MC-24 NC, I-1355</td>
<td>G, F, P</td>
<td>2–4</td>
<td>100–200</td>
<td>mean spacing ∼13 km (12 in ~150 km)</td>
</tr>
<tr>
<td>Sirenum Fossae, Figure 6</td>
<td>MC-29 NE, I-1339</td>
<td>G</td>
<td>1</td>
<td>100–150</td>
<td>Two structures are ∼150 km apart. modified by flows; profile 0890 intersects crater</td>
</tr>
<tr>
<td>Daedalia Planum, Figure 7</td>
<td>MC -17 SW, I-1189</td>
<td>G, F</td>
<td>1–2</td>
<td>&lt;100</td>
<td>mean spacing ∼19 km (9 in ~190 km)</td>
</tr>
<tr>
<td>Icaria Fossae Figure 8</td>
<td>MC-24 NE, I-1601</td>
<td>G, F</td>
<td>1–2</td>
<td>100–200</td>
<td>mean spacing ∼20 km (13 in ~260 km)</td>
</tr>
<tr>
<td>Claritas Fossae, Figure 9</td>
<td>MC-17 SE, I-1190</td>
<td>G, F</td>
<td>2–3, largest 6–8</td>
<td>50–100</td>
<td>mean spacing ∼20 km (13 in ~350 km)</td>
</tr>
<tr>
<td>Thaumasia Fossae, Figure 10</td>
<td>MC-25 NW, I-1263</td>
<td>G</td>
<td>2–4</td>
<td>50–100</td>
<td>mean spacing ∼18 km (20 in ~350 km)</td>
</tr>
</tbody>
</table>
Figure 3. Eastern Memnonia Fossae. (top) Viking image mosaic and (bottom) MOLA altimetry profiles showing the details of several of the graben and fractures in the Memnonia Fossae region of southern Tharsis (see Figure 2 for location). Arrows in profiles show location of graben and fractures.
The purpose can be dealt with by making the depth of pressure equivalence, $S$, hereafter called the compensation depth, a variable in the analysis and bearing in mind that this depth will correspond to some unspecified point on the boundary of the diapir.

The dike initially propagates upward buoyantly from its depth of origin toward the neutral buoyancy level of its magma. It increases in width and horizontal extent in the manner modeled by Lister [1990]. On reaching the neutral buoyancy level it grows upward from, downward from, and laterally along that level (Figure 11). The excess pressure at the dike center on the neutral buoyancy level is equal to $(\Delta \rho \ g S)$, where $\Delta \rho$ is the density difference between the magma and the host rocks averaged over the vertical distance $S$ and $g$ is the acceleration due to gravity. We show in section 4 that $S$ can be as large as 10–20 km on both Mars and Earth, so that if $\Delta \rho \approx 300$ kg m$^{-3}$, the excess pressure acting on the dike can be of order 10–20 MPa. Over most of the lateral extent of a giant dike, lateral growth will continue long after vertical growth has finished [Fialko and Rubin, 1999], and so the stability and location of the final position of the upper tip of the dike will be determined by the static pressure conditions in the vertical direction above the diapir in a way analogous to that proposed by Rubin and Pollard [1987] for rift-zone dikes leaving much shallower magma reservoirs in shield volcanoes. The ability of the upper dike tip to approach the surface is therefore controlled by two requirements. First, the pressure in the magma at the compensation depth $S$ must be at least great enough to support the gravitational weight of all of the magma above that depth. Second, the stress intensity at the upper tip of the dike must be equal to or greater than the effective fracture toughness of the host rocks. Pollard [1987] and Rubin [1993b] showed that fracture toughness is not an inherent material property but is a function of the host rock tensile strength and the dike tip geometry, including the likely presence of a gas-filled region into which magma is too viscous to propagate. At negligible confining stresses and in the absence of a gas cavity, $K$ is of order $3 \text{ MPa m}^{1/2}$ [Rubin, 1993b], whereas Parfitt [1991] found that shallow basaltic dikes, which probably do have gas accumulations in their tips, generally behave as though $K$ has a value close to 100 MPa m$^{1/2}$. While the dike is still growing upward, the pressure distribution within it will be partly controlled by the need to overcome wall friction, and there will be a relatively low magma pressure just below the dike tip [Rubin, 1993a]. Magmatic volatiles will be exsolved in this region and will collect into a gas pocket, and so we expect the effective fracture toughness here to be of order 100 MPa m$^{1/2}$. If the pressure at the compensation depth is great enough, the magma column can extend above the surface. In practice, of course, this means that an eruption will occur, with the magma discharge rate being a function of the amount by which the compensation pressure exceeds the weight of the magma column. The depth to which the lower tip of the dike will extend also depends on the static stress conditions as the downward growth comes to an end. Any bubbles of gas (actually supercritical fluid) formed by volatiles exsolved nearer the lower tip during the growth phase will tend to drift upward, rather than collect at the lower dike tip, and so we expect the effective fracture toughness at the lower tip to be small; if no exsolved volatile is present, it should approach the theoretical value of zero [Lister, 1990]. We have investigated the extent to which our model is sensitive to the value assumed and find that the results are indistinguishable for any value less than a few tens of MPa m$^{-1/2}$.

To determine the stress intensities at the dike tips and the shape of the dike after it reaches a stable vertical profile, we need the detailed variation with depth, $z$, of the density in the host rocks, $\rho_c$, and in the magma, $\rho_m$. A suitable model for the density structure of the Earth’s continental crust is that developed by W. Mooney et al. (http://ncweb-east.
Figure 5. Sirenum Fossae. (top) Viking image mosaic and (bottom) MOLA altimetry profiles showing the details of several of the graben and fractures in the Sirenum Fossae region of southern Tharsis (see Figure 2 for location). Arrows in profiles show location of graben and fractures.

wr.usgs.gov/study/CrustalStructure/crust/.2000) (Figure 12a). We have approximated this discrete layer model by a more continuous function defined by contiguous straight line segments as shown in Figure 12b. Information on the structure of the interior of Mars derived from tracking the motion of the Mars Global Surveyor spacecraft [Smith et al., 1999; Zuber et al., 2000] suggests a crust with a mean density of 2900 kg m$^{-3}$ overlying a mantle with a mean
density of 3500 kg m\(^{-3}\). In the region to the south and southwest of Tharsis containing the Martian graben studied here the crust—mantle boundary is at a depth of about 60 km [Smith et al., 1999; Zuber et al., 2000]. We have taken this basic information (Figure 12c) and derived from it the smoother density distribution shown in Figure 12d, based on the trends of the terrestrial profile. Given the uncertainties in the host rock densities, we feel justified in neglecting the

Figure 6. Distal Sirenum Fossae. (top) Viking image mosaic and (bottom) MOLA altimetry profiles showing the details of several of the graben and fractures in the extreme western portion of Sirenum Fossae region of southern Tharsis (to the west of the area of Figure 2 and thus not shown there). Arrows in profiles show location of graben and fractures.
variation with depth of the density, $\rho_m$, of the magma within the dike. In practice, some exsolution of carbon dioxide and water will have taken place during dike emplacement, but unless the upper dike tip approaches extremely close to the surface, the typical pressures everywhere in the system will be high enough that the change in magma density due to the presence of gas bubbles will not be large. With these assumptions we can evaluate the lithostatic compressive stress (pressure) in the host rocks as a function of depth $z$ as

$$S(z) = P_{\text{surf}} + g \int \rho_m(z) dz. \quad (1)$$

---

**Figure 7.** Daedalia Planum. (top) Viking image mosaic and (bottom) MOLA altimetry profiles showing the details of several of the graben and fractures in the Daedalia Planum region of southern Tharsis (see Figure 2 for location). Arrows in profiles show location of graben and fractures.

**Figure 8.** Icaria Fossae. (top) Viking image mosaic and (bottom) MOLA altimetry profiles showing the details of several of the graben and fractures in the Icaria Fossae region of southern Tharsis (see Figure 2 for location). Arrows in profiles show location of graben and fractures.
At any chosen compensation depth, \( C \), let the host stress be \( S_{\text{comp}} \), then the pressure in the magma at any depth \( z \) is

\[
P(z) = S_{\text{comp}} - \rho_m g (C - z).
\]

(2)

The local driving pressure, \( P_d \), wedging the dike open at any depth \( z \) is then just

\[
P_d(z) = P(z) - S(z) + T,
\]

(3)

where \( T \) is the absolute value of any horizontal regional tensile stress which may be present in the crust. Banerdt et al. [1992] argue that the range of such stresses for the Tharsis region of Mars is likely to span a few tens of Mpa, and so we have used the values 0, 10, 30, and 60 MPa for both terrestrial and Martian models in what follows.

(17) If the dike reaches a stable vertical configuration with its upper tip at depth \( U \) and its lower tip at depth \( L \), then the half-height of the dike is

\[
A = \frac{L}{C_0} U / 2.
\]

The stress intensities at the upper and lower dike tips, \( K_u \) and \( K_l \), respectively, are then obtained [Broek, 1974, equation (3.35)] from

\[
K_u = \left( \pi A^{1/2} \right)^{-1} \int_{y=-A}^{y=A} (A + y) (A - y) P_d(y) dy
\]

(4)

and

\[
K_l = \left( \pi A^{1/2} \right)^{-1} \int_{y=-A}^{y=A} (A - y) (A + y) P_d(y) dy.
\]

(5)

Finally, using the model of Rubin and Pollard [1987], if \( \rho_{cu} \) and \( \rho_{cl} \) are the mean density of the host rocks containing the upper and lower halves of the dike, respectively, then the mean dike width, \( W \), is given approximately by

\[
W = \left\{ \frac{1}{2} \left( \frac{A}{2} \right)^2 P_0 (A - \rho_m / \rho_c) \right\} A^2.
\]

(6)

(18) We implement the model as follows. First, values are chosen for the magma density, \( \rho_m \), and the regional tensile stress, \( T \). Then, a series of values is assumed for the pressure compensation depth, \( C \). The static height of the magma column which would just be supported by the pressure \( S(C) \) is evaluated and compared to \( C \). If the column height is greater than \( C \), the outcome is noted as a potential eruption; otherwise, the value is stored as the depth of the upper dike tip, \( U \). Since the bottom tip of the dike must be located at a depth greater than \( C \), a series of values, \( L \), progressively greater than \( C \) is taken for the depth of the lower dike tip, and for each of these, equations (4), (5), and (6) are used to evaluate \( K_u \), \( K_l \), and \( W \). When \( L \) is just greater than \( C \), \( K_l \) is always positive, but its value decreases as \( L \) increases. When \( K_l \) reaches some chosen small value, the lower dike tip is regarded as having become trapped and the growth process is stopped (as noted above, \( K_l \) values from zero up to a few tens of MPA m\(^{1/2}\) all lead to similar solutions). The values of \( K_u \) and \( W \) are then examined. If \( K_u \) is greater than or equal to the critical value representing the effective fracture toughness (we use 100 MPa m\(^{1/2}\)), and if \( W \) is

Figure 9. Claritas Fossae. (top) Viking image mosaic and (bottom) MOLA gridded altimetry profiles showing the details of several of the graben and fractures in the Claritas Fossae region of southern Tharsis (see Figure 2 for location). Arrows in profiles show location of graben and fractures.
positive at all depths, the solution is regarded as physically possible and is stored. In other cases, either or both of $K_u$ and $W$ are negative. These solutions correspond to events in which small batches of magma would have pinched off from the diapir and propagated upward as discrete, detached dikes as discussed above; such solutions are discarded since they do not represent stable giant dikes.

4. Model Results

[19] Using the above analysis, we arrive at a family of physically possible giant dikes characterized by three parameters: the assumed magma density, $\rho_m$, the depth of pressure compensation, $C$, and the regional crustal tensile stress, $T$. For each such dike our modeling predicts its mean width, $W$, and the depth to its upper tip, $U$ (or the fact that the tip reaches the surface and an eruption occurs). Figure 13 shows the details of the relationships for dikes on the Earth in the case where the regional tension is zero. The magma densities used span the entire range between the deep mantle density and the shallow crustal density. In Figure 13a, mean dike width is seen to increase with increasing depth of compensation for a given magma density and to decrease with increasing magma density for a given compensation depth. The boundaries of the parameter space occupied by viable giant dikes are indicated by dashed lines. As expected, too deep a compensation depth leads to the generation of a series of small buoyant dikes which rise and stall in the crust, and no giant dike can be formed (right-hand side of diagram). Also, the lower the density of the magma, the more readily a surface eruption occurs (upper left part of diagram). Finally, large compensation depths coupled with dense magmas lead to giant dikes which fail to fracture a pathway to as shallow a depth as could be reached on buoyancy grounds alone and are emplaced with a significant excess pressure in their upper tips (bottom middle of diagram). Figure 13b shows the depth to the upper dike tip for the same set of magma densities and compensation depths as Figure 13a, with the same boundaries indicated. The depths to the lower dike tips are not indicated directly in these figures but are generally 1.4–1.6 times the corresponding compensation depths.

[20] In order to show the influence of the presence of a regional tectonic tensile stress in a systematic way, Figure 14 repeats the information for $T = 0$ on Earth shown in Figure 13 and also gives the results for the three other tensile stress values used: 10, 30, and 60 MPa. As before, part a of the figure shows the mean dike widths, and part b shows the depths to the upper dike tip. The trends seen in Figure 13 are repeated but with some systematic changes. As $T$ increases, the range of compensation depths from which giant dikes can be generated decreases, and the sources are required to be ever closer to the base of the crust. Also as $T$ increases, the mean widths of giant dikes increase, and there is a tendency for upper dike tips to be located at greater depths below the surface.

[21] Figure 15 shows the complete set of equivalent results for Martian giant dikes. The variations of mean dike width values with $\rho_m$ and $T$ shown in Figure 15a are very similar to those of terrestrial giant dikes. However, there is a difference in the trends of depths to upper dike tips between Figures 14b and 15b: upper dike tip depths increase with $T$ throughout the range of values used for the Earth, but at the largest tension values on Mars the trend reverses: increasing the tension further decreases the depth to the upper dike tip, and also (60 MPa panel of Figure 15a) disproportionally increases the mean dike width. These differences are the
consequence of the nonlinearities introduced by the magma and rock properties being the same on the two planets, whereas the gravity, and hence all stresses depending on depth beneath the surface, is systematically smaller on Mars.

[22] If the whole range of magma densities illustrated here is actually present (3000–3300 kg m$^{-3}$ in the case of Mars and 2950–3200 kg m$^{-3}$ in the case of the Earth), then on both planets dike widths should commonly be as large as 400 m. However, the distributions of dike widths shown in Figures 14 and 15 imply that the average widths of Martian giant dikes are about 30% greater than those of their terrestrial counterparts, a trend in the same sense as, but smaller than, that inferred on the basis of simpler models by Wilson and Head [1988], Wilson and Parfitt [1989], and Wilson and Head [2000].

[23] We noted earlier that the lower tips of the giant dikes modeled here were located at depths of ~1.5 times the compensation depth. The mean compensation depth on Mars is ~75 km (Figure 15b) and on Earth is ~50 km (Figure 14b). This difference, coupled with the somewhat greater mean dike width, means that giant dikes on Mars will typically hold almost twice as much magma per unit distance along strike as dikes on Earth. Also, on both planets the entire range of dike widths can be exhibited by dikes penetrating to very shallow depths in the crust (certainly less than 1500 m on Earth and 2500 m on Mars), thus implying that a very wide range of sizes of surface graben could in principle be produced.

[24] Figures 14b and 15b show that if magma of a fixed density is being delivered into a spreading giant dike from a diapir, then if the compensation depth (the effective source depth of the magma) decreases with time the depth to the top of the dike should increase with time. A trend of very slightly decreasing compensation depth might be expected during a dike emplacement event. This is not because the diapir feeding the dike is ascending as a whole: if the diapir penetrates the host rocks in a plastic fashion, its mean rise rate, based on diapir rise models by Marsh [1982, 1984], will be less than a meter per year [Wilson et al., 2001]. However, the diapir must deform in order to deliver magma into the dike, and if, as we postulate, there are minimal stresses acting across the boundary between the diapir and its surroundings, this implies that the deep mantle must accommodate a deformation regime which allows mantle rocks immediately beneath the diapir to move upward. If the location of the main connection between the diapir and the dike is anywhere other than the very top of the diapir, therefore, there will be some small effective upward motion of the source. We show in the next section that the extent of this upward motion does not need to be more than a few kilometers, however, and so we would not expect the depth of the upper dike tip to change by more than a very few hundred meters. Nevertheless, this might be enough to lead to conditions in which the growing dike initially intersected the surface immediately above the diapir, thus leading to a proximal eruption, but ceased to be able to reach the surface as it migrated sideways and depleted the diapir, thus forming lateral dikes which had no eruptions from their distal portions but nevertheless had associated graben.

[25] Ernst et al. [1995, Table 1] found evidence that members of giant dike swarms on Earth commonly have widths of ~20–60 m, with some values ranging up to ~200 m. There are no direct methods of measuring the depths of the bottoms of these dikes, but indirect methods imply lower limits on the depths of 12–20 km. The lower left parts of the panels in Figure 14a show that dike widths in the 20–60 m range correspond to those predicted for magmas with densities less than about 3000 kg m$^{-3}$, whereas widths up to 200 m imply densities in the 3000–3150 kg m$^{-3}$ range. Even the smallest of these widths and densities would require the bottoms of the dikes to penetrate to depths of at least 50 km, implying that the lower limits estimated by existing techniques are significant underestimates.

[26] We stress that the absolute density values given here must be viewed with caution: the real control is the difference in density between the magma and the crust. Changes in the assumed crust-mantle density values, and in the shape of
the density profile, would change the implied magma densities systematically. However, the degree of convergence between our model results and the properties of giant dikes on Earth encourages us to apply the modeling to Mars.

5. Application to Martian Giant Dikes

5.1. Magma Extraction from Mantle Plumes

[27] We expect that, as on Earth, the main sources of giant dikes on Mars will be diapirs stalled just below the base of the crust. We therefore begin by examining the lower left parts of the panels of Figures 15a and 15b, where compensation depths are 60–80 km. Giant dikes can propagate to shallow levels from these depths for magma densities up to 3200 kg m$^{-3}$. These dikes will have mean widths up to 400 m (Figure 15a). Their tops will be at depths commonly <3000 m (Figure 15b), and their lower edges will be at depths between 100 and 120 km (about 1.5 times the compensation depths). The volumes of such dikes will depend on the lateral distance that they are able to propagate. We shall shortly infer that graben systems on Mars between ~1000 and ~3000 km in lateral extent are

Figure 12. (a) Model for the density structure of the Earth’s continental crust (W. Mooney et al., http://neweb-east.wr.usgs.gov/study/CrustalStructure/crust/, 2000). (b) This discrete layer model was approximated by a more continuous function defined by contiguous straight line segments. (c) For the area of interest on Mars, the crust-mantle boundary is at a depth of about 60 km [Smith et al., 1999; Zuber et al., 2000]. (d) A smoother density distribution for Mars based on trends in the terrestrial profile.
underlain by giant dikes, and using this distance, we find total magma volumes in the dikes of 20,000–60,000 km$^3$. 

If this magma were derived from a diapiric plume head with a horizontal diameter of 500 km, then the equivalent vertical extent of the magma in the diapir would need to be $\sim$300 m. If this melt were segregated from a residuum that had undergone 5% partial melting, the original vertical extent of the zone of partial melting would have had to be $\sim$6 km. Clearly, then, during the lifetime of a single plume there is no difficulty in accommodating multiple episodes of incremental plume rise, partial melting in a zone a few kilometers in vertical extent, and melt segregation, followed finally by melt extraction and giant dike formation. However, an important issue is the fractional amount of partial melting which must occur in order to initiate dike propagation, since this, together with the volume of the plume head, limits the volume of magma available to be injected into the dike. We do not address this issue in detail but show later in this section that magma cooling limitations imply that a single giant dike may be formed from a series of magma pulses provided that the total duration of the episodes is no more than $\sim$100 years.

The velocity of propagation of a laterally propagating dike is limited only by the ability of the magma to flow fast enough to keep pace with the elastically advancing tip and is therefore determined by the mean dike width, $W$, the magma viscosity, $\eta$, and the pressure gradient, $dP/dz$, acting on the magma in the horizontal direction. For a magma density close to 3100 kg m$^{-3}$ (which yields viable dike models in Figure 15) the absolute pressure in the magma at the midpoint of the dike is found in our models to lie in the range 450–550 MPa. If the dike is able to propagate freely, the pressure at the propagating lateral tip will tend to be as small as possible, which means that it will be buffered by the exsolution of the least soluble magmatic gas [Lister, 1990; Rubin, 1993a]. Data on the solubility of CO$_2$ in magmas [Harris, 1981; Wilson and Head, 1981] show that if the magma contains, say, 0.1 wt% CO$_2$, this pressure will be about 45 MPa. At the beginning of propagation, with the dike tip $\sim$250 km (a plausible plume radius) from the center of the diapir, the pressure difference acting horizontally between the magma source and dike tip will be $\sim$450 MPa. Rubin [1993a] showed that most of the pressure change along a propagating dike occurs near the dike tip, and so the typical pressure gradient acting on the flowing magma will be $dP/dz \sim 100$ Pa m$^{-1}$ near the end of propagation, with the tip 3000 km from the source, the gradient will be an order of magnitude smaller at about 10 Pa m$^{-1}$. The flow speed under these conditions of a basaltic magma with viscosity, say, 100 Pa s into a typical giant dike with a mean width of $\sim$200 m would range from $\sim$35 m s$^{-1}$ initially to $\sim$10 m s$^{-1}$ near the end of dike growth (the magma motion is fully turbulent, and so the exact value of the viscosity is irrelevant). The average of these values corresponds to magma flowing into the dike at a rate of $\sim$1 km$^3$ s$^{-1}$. On the assumption that this rate could be accomplished, the emplacement time of a 3000 km long dike would be of order 20 hours. We note that the above pressure gradients and magma flow speeds are very similar to those found by Fialko and Rubin [1999] in a more detailed model of lateral magma flow in giant dikes on Earth. To some extent this is fortuitous, since we are postulating a significantly greater level of neutral magma buoyancy than in their model, such that we can neglect contributions to the driving pressure gradient from topographic changes (i.e., lateral changes in crustal thickness) or variations in the vertical gradient of the tectonic stresses along the dike path. However, the similarity means that we can draw on some of Fialko and Rubin’s [1999] thermal results. These show not only that magma cooling in dikes of the thicknesses implied by our models is negligible but also that the amount of heat transferred to the dike walls on the $\sim$1 day timescale of emplacement that we find is not large, not enough to cause significant melting of dike wall rock, for example. Initial interactions between magma and water or ice in the shallow crust would therefore be negligible, unless perhaps overpressurized groundwater were released by fracturing of a cryospheric seal as a result of surface deformation [e.g., Head and Wilson, 2002].
These speeds and timescales would, of course, be maintained only if the diapiric body supplying the magma were able to deform at the required rate. If the deformation were accomplished by rise of the mantle rocks beneath the base of the diapir, the required speed over a circular area with a radius of 250 km would need to be \( \frac{4}{C_2} \frac{m}{ms} \). Whether the Martian mantle can deform at this rate is problematic, given that typical terrestrial mantle deformation speeds within mantle plumes are \( \frac{3}{C_2} \frac{10}{C_2} \frac{mm}{s} \); however, presumably the material in the core of a mantle plume from which melt has just been extracted has a smaller bulk viscosity, because of the presence of some residual melt, than typical mantle rock. We can establish an absolute minimum required deformation speed by using the magma cooling rate. A thermal cooling wave will penetrate the distance \( \lambda = 100 \) to the center of a 200 m thick dike in a time of \( (\lambda^2/k) = 10^{10} s = 330 \) years. To avoid significant cooling during emplacement, therefore, the minimum emplacement time of such a dike needs to be less than \( 100 \) years = \( 3 \times 10^9 \) s. The implied magma supply rate into the dike is then \( 2 \times 10^5 \) m\(^3\) s\(^{-1}\), and the corresponding deep mantle deformation speed using the same geometry as before is \( 10^{-4} \) mm s\(^{-1}\). This value is closer to, but still \( 100 \) times faster than, normal mantle deformation rates.

5.2. Graben Formation Above Giant Dikes

We now predict the ranges of geometries of the graben that would be expected to have formed above giant dikes penetrating close to the surface on Mars and compare these to our observations. Mége and Masson [1996b] discuss at length the various issues which make it difficult
to establish an unambiguous relationship between the geometry of a graben and the geometry of, and stress conditions related to, an underlying dike. They give a range of possible relationships for graben in the Tharsis region determined from the treatments given by Pollard et al. [1983], Mastin and Pollard [1988], and Rubin and Pollard [1988]. We prefer to use the relationships found by Rubin [1992], who analyzed two dike-induced graben in Iceland using a model taking account of the inelastic response of the crust to the fractures forming the graben walls. He found that the ratio of the graben width to the depth to the dike top was a factor of \( \frac{2}{3} \) in one case and 2.9 in the other. We adopt the value 3.5 and note that this ratio is not expected to be a function of gravity [Rubin, 1992]. Figure 15b shows that for compensation depths in the upper mantle, dike top depths commonly range from zero to 3000 m, which would imply graben widths up to 10 km. Table 1 shows that all of the measured graben widths lie in the range 1–4 km, entirely consistent with expectations and implying that the Martian graben were produced as a result of a smaller range of conditions than the wide range we have explored in the modeling. Rubin [1992] also found that for his Icelandic examples the ratio of dike width to amount of vertical subsidence of the graben floor was 1.0 in one case and 1.5 in the other. These values are broadly consistent with the idea that the extension represented by the width of the graben is approximately equal to the width of the underlying dike: for graben boundary faults dipping at \( \sim 60^\circ \) the ratio would be \( 2 \tan 30^\circ = 1.15 \). Overall, we adopt the value 1.25 as a plausible estimate. The dike widths in Figure 15a, commonly in the range 50–500 m, then imply graben subsidence depths of 40–400 m; Table 1 shows that the vast majority of unmodified graben depths lie in the range 100–200 m, again consistent with expectations.

Figure 15. (a) Mean widths and (b) depths to upper dike tips for giant dikes on Mars as a function of depth of diapir pressure compensation and magma density for each of four values of the regional crustal tectonic tensile stress.
Table 2. Parameters of Giant Dikes and Associated Graben Systems on Mars*  

<table>
<thead>
<tr>
<th>Horizontal Regional Tensile Stress, Mpa</th>
<th>Range of Depths From Which Dikes Can Propagate, km</th>
<th>Depths of Upper Dike Tips, km</th>
<th>Mean Dike Widths, m</th>
<th>Implied Range of Graben Widths, km</th>
<th>Implied Range of Graben Depths, m</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>60–110</td>
<td>0–3.0</td>
<td>0–420</td>
<td>0–9.5</td>
<td>0–335</td>
</tr>
<tr>
<td>10</td>
<td>60–95</td>
<td>0–3.8</td>
<td>0–500</td>
<td>0–13.3</td>
<td>0–400</td>
</tr>
<tr>
<td>30</td>
<td>60–85</td>
<td>0–5.7</td>
<td>80–620</td>
<td>0–20</td>
<td>65–500</td>
</tr>
<tr>
<td>60</td>
<td>60–70</td>
<td>0–3.6</td>
<td>460–850</td>
<td>0–12.6</td>
<td>370–680</td>
</tr>
</tbody>
</table>

*The maximum ranges of values are given; see text for discussion of most likely values.

Table 2 summarizes the predicted dike and graben geometries derived from the theoretical model. As an illustration of the more detailed inferences that may be made from these results, we note, for example, that the upper limit to graben width of 4 km shown in Table 1 corresponds to a depth to dike top of no more than ~1500 m, and Figure 15b implies that the simplest explanation is that the dike magmas involved had densities less than ~3200 kg m$^{-3}$. Similarly, the common maximum graben width of 200 m corresponds to a dike width of ~250 m, and Figure 15a shows that if the regional tensional stress is greater than about 30 MPa, the dike width can be less than ~250 m only if the magma density is greater than ~3200 kg m$^{-3}$. This is in direct conflict with the inference that the magma density be less than 3200 kg m$^{-3}$ just obtained from the graben widths, and so we conclude that regional tensional stress levels in the Tharsis province were generally less than ~30 MPa when the giant dikes were emplaced.

Next we note that it is rare for eruptions to be associated with the graben studied here. Figure 15b shows that to ensure that the upper dike tip does not reach the surface, we require a magma density greater than ~3100 kg m$^{-3}$ if the regional tension is 30 MPa, greater than ~3150 kg m$^{-3}$ if the regional tension is 10 MPa, and greater than ~3200 kg m$^{-3}$ if the regional tension is 30 MPa. But we have already concluded that the regional tension must be less than ~30 MPa and that the magma density must be less than ~3200 kg m$^{-3}$. Thus we infer that various combinations of magma density in the range 3100–3200 kg m$^{-3}$ and regional tension between zero and 30 MPa are consistent with the general absence of eruptions. We stress here, as for the terrestrial examples given earlier, that these values would change somewhat if the density structure of the lithosphere were different from the model adopted. However, they are certainly robust enough to indicate the broad conditions required for the formation of the features observed.

Finally, we explore the consequences of intruding giant dikes into a Martian crustal structure different from that underlying Tharsis. The equivalent of the calculations described above was carried out with the depth assumed for the crust-mantle boundary changed from 60 to 40 km, a value more appropriate for the northern lowlands on Mars [Zuber et al., 2000]. The shape of the density profile shown in Figure 12d was retained; that is, we simply scaled all of the depths to 2/3 of their original values. The results show the following patterns: (1) The range of depths below the crust-mantle boundary from which viable giant dikes can be produced is reduced to ~65% of its previous value; (2) the depths to dike tops for a given magma density are also reduced to ~65% of their previous values; (3) the mean dike widths for a given magma density are reduced to ~50% of their previous values; (4) to generate a given dike width and depth to dike top for the same distance of the compensation depth below the crust-mantle boundary requires a denser magma; and (5) if we wished to ensure that generally eruption-free graben with the widths and depths found in the Tharsis systems were also created in these thinner crustal conditions, then the restrictions on regional tensional stress would be similar to before, with values less than ~30 MPa being relevant, but the restrictions on magma density would be much tighter, with only values close to 3100 kg m$^{-3}$ being allowed. The implication is that for a similar spread of magma densities, eruptions from giant dikes would be more likely to occur in regions where the crust was thinner.

### 6. Discussion and Implications

This analysis illustrates that the vast Tharsis-radial graben and fracture systems are plausibly interpreted as giant radiating dike systems emanating from the central part of Tharsis and forming dike complexes whose vertical extents span the majority of the crust and extend into the mantle. On the basis of these findings we address several implications and outline areas of further study.

#### 6.1. Origin and Orientation of Radial Structures

On the basis of geological relationships described in section 2, the extensive graben in the southern Tharsis region (Figures 1 and 2) appear to radiate from two different centers, Arsia Mons and Syria Planum. It is highly likely that these two edifices represent the surface manifestation of ascending plumes. As we have shown above, the size and geometry of the radial graben are consistent with an origin by radial dike emplacement from plumes rising below Arsia Mons and Syria Planum, development of near-surface stresses during dike emplacement, and faulting to produce the fractures and graben.

Graben systems are commonly not purely radial to their source edifices, but they can clearly be traced back to the interpreted source region. This is typical of radial dike systems as they propagate outward from regions characterized by the source-controlled, overpressure-dominated proximal stress field to distal regions where the regional stress field dominates [e.g., Pollard, 1987; Grosfils and Head, 1994a, 1994b; Ernst et al., 1995]. As a corollary is that the longer dikes emplaced during buffered magma supply conditions [Parfitt and Head, 1993] are more likely to be arcuate than shorter dikes that are emplaced in unbuffered magma supply conditions. Because dikes can become...
increasingly nonradial with distance from the source, we caution that multiple lines of evidence must be employed in order to locate source regions of dike-induced graben where only parts of the system are preserved [Anderson et al., 1998; Anderson and Dohm, 2000; Dohm et al., 2000; Anderson et al., 2001].

[33] In the case of the graben mapped in this analysis (Figures 1 and 2), stereographic reprojection (Figure 16) shows that they are generally radial to the centers in Arsia and Syria and do not show the major distal deviation often seen in the other examples cited above. This is very likely due to the fact that stresses are concentrated at the advancing dike tip which lies at the neutral buoyancy zone at the base of the crust, and thus the surface orientation is not very susceptible to shallower stresses concentrated in the upper crust (though there is some influence of shallow stresses, evidenced by the presence of en echelon offsets between some graben segments: Figures 3–10). This, together with the minimal regional stresses apparently necessary for graben formation over dikes, appears to account for the dominantly radial trends out to great distances from the source region.

6.2. Timing of Radial Structures

[39] The range of tectonic and volcanic activity associated with Tharsis spans from the Noachian to Amazonian periods [Scott and Tanaka, 1986; Tanaka et al., 1992], and it is anticipated that radial dike systems were emplaced throughout this period, as suggested by the stratigraphic relationships between the structures [e.g., Watters and Maxwell, 1983]. In principle, giant radial dikes are very likely to be linked to periods of enhanced magma production and supply (buffered conditions [e.g., Parfitt and Head, 1993]), which are most likely to occur in the initial periods of plume penetration and flattening or at times of subsequent peak supply (e.g., during episodic pulses in plume tail flow). For the Arsia Mons examples, it is clear from the stratigraphic relationships that radial graben occurred during the early, middle, and late stages in the formation of Arsia but that the longest graben were concentrated in the earlier stages of evolution. For the Syria-radial systems, graben formation was concentrated in the early to intermediate periods, and there is little evidence for Amazonian-aged activity.

6.3. Volume of Magma Intruded in Different Events and Implications for Magma Reservoirs and Magma Supply

[40] Wilson et al. [2001] have shown that the volumes and estimated total lifetimes of the Tharsis volcanoes imply that they were each built with a mean magma supply rate of \( \sim 0.05 \text{ m}^3 \text{ s}^{-1} \). Summit caldera morphologies and positions indicate that multiple magma reservoirs existed at slightly different places at different times [Scott and Wilson, 2000] and thus that temporal gaps existed between periods of magma supply. Magma supply rates to establish new reservoirs are estimated by Wilson et al. [2001] to be \( \sim 1-10 \text{ m}^3 \text{ s}^{-1} \) for periods of 100,000 years, with an initial pulse of more than 150 \text{ m}^3 \text{ s}^{-1} over at least a few weeks required for reservoir initiation. Wilson et al. [2001] examined volumes of long flows on Tharsis and concluded that they must represent eruption rates of \( \sim 100-300 \text{ m}^3 \text{ s}^{-1} \) that occurred for periods of a few to tens of years, implying buffered eruption conditions [Parfitt and Head, 1993] in which a high magma supply rate from the mantle is continuously maintained, in contrast to unbuffered conditions characterizing smaller-volume inelastic caldera collapse events. These relationships suggested to Wilson et al. [2001] that the Tharsis volcanoes were built episodically; active phases lasting between 1 and 10 Myr alternated with quiescent periods of \( \sim 10-100 \) Myr. Ten to perhaps as many as a hundred cycles of enhanced magma supply may have been involved.

[41] The individual dike emplacement events modeled here and interpreted to underlie the radial graben are of the magnitude and frequency that are also typical of buffered magma supply situations, in which a high magma supply rate from the mantle is continuously maintained [Parfitt and Head, 1993]. For example, a circumferential traverse around the Arsia-centered radial graben reveals about 30 parallel graben, while a similar traverse around the Syria Planum-centered graben shows at least 50 structures (Figure 2). Thus these features may be a record of the buffered events that initiate new phases of magmatic activity in the Tharsis volcanoes. Closer examination of their crosscutting stratigraphic relationships with underlying units, as well as their grouping, could provide more detailed evidence about the history of magma supply in the Tharsis volcanoes and also about the state and orientation of lithospheric stress at those times. The number and implied frequency of these major events can be understood in terms of instabilities in the extraction of melt from mantle plume heads in general [Scott and Stevenson, 1986; White et al., 1995] and on Mars in particular [e.g., Harder, 1998, 2000].

[42] Although volcanic edifices, particularly those on Tharsis, are visually impressive in terms of their magnitude, detailed analyses have shown that a significantly larger portion of a magmatic event is intrusive, rather than extrusive [Crisp, 1984], and this is particularly true in terms of the presence of extensive radial dike swarms. For example, the volume of the Arsia Mons edifice is estimated to be \( \sim 1.5 \times 10^6 \text{ km}^3 \) [Smith et al., 2001]. Estimates of the volume of magma in one major diking event calculated here range from 20 to 60 \times 10^3 \text{ km}^3, which is 1–4% of the total volume of the edifice.

6.4. Number of Events in Individual Systems and Implications for Regional Strain

[43] Golombek et al. [1996] measured the amount of extension across Tempe Terra by examining fault scarp widths and deformed craters. Measurements of fault scarp width and slopes for simple graben and rifts were used to estimate fault throw, and these data, together with estimates of fault dips, were used to estimate extension. Golombek et al. [1996] found about 2–3% strain across the Tempe Terra region, in a direction normal to the radially oriented structures. The two methods indicated an average extension for single normal fault scarp of \( \sim 100 \text{ m} \). Golombek et al. [1996] point out that estimates elsewhere around Tharsis show significant differences (Thaumasia \( \sim 0.5\% \); Sirenum, 0.1%) although strain across individual structures with smaller widths can be much higher (5–20%). Our calculations show that a significant amount of the observed strain could be the result of dike emplacement, in which vertical
meters in width create extensional stresses resulting in near-surface fissures and graben.

6.5. Influence of Giant Dike Emplacement on the Cryosphere and Hydrosphere

Emplacement of dikes of this magnitude must have an influence on various aspects of the geological evolution of the substrate. Emplacement may be rapid enough (hours to days) that the initial heat pulse is likely to have a minimal effect on the environment. However, as soon as emplacement is complete, the dike represents a vertical thermal anomaly that spans virtually the entire hydrosphere and cryosphere. Its presence should set up heating of the surrounding groundwater system and produce extensive hydrothermal circulation over the course of its cooling history [McKenzie and Nimmo, 1999]. Dikes with the typical widths found here will take several tens of years to cool conductively [Wilson and Head, 1994], and cooling rate will be enhanced by recirculating groundwater. One might expect significant deposition of hydrothermal minerals in the vicinity of the margins of the dike and perhaps as surface deposits in the vicinity above the dike associated with the graben and fractures [e.g., Gulick, 1998]. Furthermore, if the near-surface stresses associated with dike emplacement crack the cryosphere, such buoyantly rising hot water might form hot springs and surface hydrothermal deposits in the vicinity. If the cryosphere has sealed the groundwater system under hydrostatic pressure prior to dike emplacement, then fracturing associated with dike emplacement may crack the cryospheric seal, leading to the catastrophic release of groundwater. Such a situation is envisioned to have occurred in the case of Mangala Valles, in the northern part of Memnonia Fossae, where Mangala emerges from one of the graben in the system (Figure 17) [see Head and Wilson, 2002, and references therein].

6.6. Influence of Giant Dikes on Groundwater Flow

Examination of Figure 2 shows that the formation of such significant dikes in the crust could have a major effect on regional and global groundwater circulation. Clifford [1993] describes a north-south global circulation of ground-
resulting vertical recycling of magma would enhance gas convection within the dike while it was cooling. The abundance of groundwater in this area and potentially influence or contribute to an ing the interpreted relation of graben, dikes, and outflow channels. (a) The Memnonia Fossae graben extend from the Tharsis rise, and Mangala Valles appears to have its source at one of the Memnonia graben. (b) One explanation for this relationship is shown in the block diagram, where dikes are seen to propagate radially from overpressurized magma reservoirs in Tharsis. As the dikes approach the surface, they create near-surface stress fields that form graben, cracking the cryosphere and releasing groundwater confined by the cryosphere. Outflow continues until the cryosphere freezes and reseals or hydrostatic equilibrium is reached.

6.7. Gas Exsolution from Giant Dikes

The great width of a giant dike could also lead to convection within the dike while it was cooling. The resulting vertical recycling of magma would enhance gas exsolution and gas loss. This could result in gas pressure buildup, leakage, and collapse to form craters over the dike and within the graben. It could also lead to vulcanian eruptions, such as those envisioned in some dark halo deposits on the Moon [e.g., Head et al., 2002a], where gas buildup creates overpressures that cause catastrophic disruption of the lid, or even more prolonged plinian eruptions, as suggested for some large elongate craters associated with graben near Alba Patera on Mars [Scott et al., 2002; Scott and Wilson, 2002]. In any case, if all of the gas estimated to occur in Martian basalts degasses from the dike, it would contribute about 0.1% of the total volatiles thought to be contributed by the volume of early Hesperian extrusive lavas represented by the ridged plains [e.g., Head et al., 2002b].

6.8. Eruptions Along Giant Dikes and the Origin of Plains

[47] The laterally (along-strike) irregular and discontinuous occurrence of graben and fractures indicates that the tops of the dikes are not necessarily at the same depth along their total length. Differences in surface topography, and local and regional geology and stresses, could also cause such irregular distribution. As a result, surface eruptions might be expected to occur at some locations along the strike of the dike but not necessarily everywhere. These eruptions would tend to form flood deposits and plains, not small edifices characterized by multiple, low-volume eruptions. Other than pits and depressions locally enlarged by collapse, positive edifices and volcanic landforms are not clearly associated with the graben and fractures. Scott [1980, 1982] has mapped a number of features that might be candidates, but a clear association has not been firmly established. One possible manifestation of surface eruptions could be the distribution of volcanic plains in regions associated with the graben. At least part of the along-strike irregularities could be due to surface effusion and burial of the graben. Analysis of the geological maps [Scott and Tanaka, 1986] shows that there are often patches of Noachian and Hesperian plains that could overlie and obscure some of the graben and fractures and represent localized surface eruptions.

6.9. Variations in Crustal Thickness and Implications for Resurfacing of the Northern Lowlands

[48] If giant dikes are intruded into a Martian crustal structure different from that underlying Tharsis (~60 km) and more like that of the northern lowlands (~40 km) [e.g., Zuber et al., 2000], we find that for a similar spread of magma densities, eruptions from giant dikes would be more likely to occur in regions where the crust was thinner. Recent analyses of the stratigraphy of the northern lowlands has documented the likely presence of widespread deposits of Early Hesperian ridged plains (HR) of volcanic origin [Head et al., 2002b]. No specific source edifices have been documented within the northern lowlands, but our findings suggest that radial dikes from nearby sources in the adjacent uplands (e.g., Tharsis, Tempe Fossae, Alba Patera, Elysium, etc.) could easily reach into all parts of the northern lowlands and produce eruptions to provide the resurfacing that apparently occurred in the Early Hesperian, prior to the emplacement of the Vastitas Borealis Formation [e.g., Head...
et al., 2002b]. Reduction in magma density favors eruption even further. If the basaltic andesite compositions in the northern lowlands interpreted from Thermal Emission Spectrometer (TES) data [Bandfield et al., 2000; Christensen et al., 2000] represent the underlying rock composition (see also discussion by Noble and Pieters [2001]), then density may have been an additional factor and may imply different compositions in the northern plume sources from those in the south of Tharsis.

6.10. Implications for Linear Magnetic Anomalies

[49] We note that some of the magnetic lineations recently observed on Mars [Acuna et al., 1999; Connerney et al., 1999; Purucker et al., 2000] have dimensions similar to those of the dike systems described here and are subparallel to them. Dike intrusion has been proposed as a possible cause of such lineations [e.g., Wilson and Head, 2000; Nimmo, 2000], but further study is needed to determine if this specific mechanism is applicable. An outstanding question is related to timing. If the magnetic dynamo on Mars is limited to its earliest history, primarily during crustal formation (see summary by Stevenson [2001]), then this largely predates the observed surface geological record and the features described in this paper almost certainly postdate this period. However, if the initiation of Tharsis accompanied or overlapped with initial crustal formation [e.g., Phillips et al., 2001], these linear magnetic anomalies could be related to earlier dike intrusion phases not currently preserved in surface deformation.

6.11. Implications for Models for the Origin and Evolution of Tharsis

[50] The origin of Tharsis and the relationships between its volcanic and tectonic features have been a matter of controversy for years (see summaries by Banerdt et al. [1992] and Tanaka et al. [1991]). Numerous models have been proposed to explain the observed topography, gravity, and stress state, including that Tharsis is a huge uplifted dome [e.g., Phillips et al., 1973; Mège and Masson, 1996b], that it formed isostatically by magmatic intrusion [e.g., Sleep and Phillips, 1979, 1985], and that it formed primarily by volcanic loading [e.g., Solomon and Head, 1982]. Recent gravity and topography data from MGS have shown that loading of the lithosphere by the Tharsis rise in the Noachian can explain much of the global shape and long-wavelength gravity field of Mars [e.g., Phillips et al., 2001].

[51] The radial graben discussed in this paper (Figures 1 and 2) form an important part of the observational constraints for models of the origin of Tharsis [e.g., Banerdt et al., 1992; Tanaka et al., 1991]. The interpretation of these features as the surface manifestation of large dikes emplaced radially from magmatic centers in Tharsis removes the need to call on Tharsis-related regional extensional stresses for their formation and orientation. Similarly, regional strain measurements made from analysis of the offset on these radial features should be considered in the context of strain associated with dike emplacement and not solely regional tectonic strain associated with the evolution of Tharsis. Finally, the timing of formation of these features should be considered in the context of both tectonic and magmatic models. Tectonic models often call on time-dependent stresses associated with, for example, uplift or loading to produce the radial graben. Magmatic models such as the one outlined here produce the graben from stresses associated with plume and reservoir overpressurization and lateral propagation of magma-filled cracks. Thus the graben are associated primarily with magmatic history and can form at any time throughout the history of the plumes. The range of ages of the radial graben described in section 2 lends support to the magmatic models.

[52] The magmatic interpretation of the Tharsis-radial graben potentially removes one of the conundrums of Tharsis tectonics [e.g., Banerdt and Golombek, 2000], in which it appeared to be necessary to require two distinct modes of support for Tharsis in order to explain the presence of radial structures (graben) on both the elevated flanks (where they were attributed to isostatic stresses) and outside the Tharsis topographic rise (where they were seen to be more consistent with flexure).

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References


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