Thick Shell Tectonics on One-Plate Planets: Applications to Mars

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A theory for stress distributions in thick lithospheric shells on one-plate planets is developed based on the zero frequency equations of a self-gravitating elastic spherical shell overlying a strengthless fluid. Stress distributions in lithospheres are reviewed for both the compensated and flexural modes. In the former case, surface stresses only depend on surface topography, while for the latter case, it is shown for long wavelengths that stress trajectories are mainly dependent on the lithospheric lateral density distribution and not on elastic properties. Computational analyses are carried out for Mars, and it is found that isostatically compensated models correctly predict the graben structure in the immediate Tharsis region and a flexural loading model is satisfactory in explaining the graben in the regions surrounding Tharsis. A three-stage model is hypothesized for the evolution of Tharsis: isostasy with north-south graben formation on Tharsis, followed by flexural loading and radial graben formation on the perimeter of Tharsis, followed by a last stage of loading with little or no regional deformation. This model is consistent with the Martian lithosphere monotonically thickening over geologic time.

INTRODUCTION

In the absence of extensive surface-based geophysical and petrological measurements, the principal observations which relate directly to the interior of a planet are its gravity and topography. Unfortunately, the intrinsic ambiguity in interpreting gravity fields makes this a singularly frustrating data set with which to constrain planetary models. In order to render this ambiguity somewhat tractable, it is generally necessary to make a number of 'reasonable' assumptions about interior models. In addition, any geophysical or geological consequences of the models must also be reasonable, which allows the use of various indirect observations of interior processes (notably photogeologic information) to constrain further the problem (see Phillips and Lambeck [1980] for a thorough review of gravity fields of the terrestrial planets and their interpretation).

One of the more useful ways in which we might utilize gravity data to study the interior of a planet is to find a physical relationship between surface topography and gravity. Applicable models include a topographically loaded elastic lithosphere and an elastic lithosphere responding to convection currents from beneath.

Deformation of an elastic lithosphere offers an additional constraint in that the stress patterns set up by the gravity field and the topography may have observable tectonic consequences in terms of the style and directions of faulting. Much of this paper is concerned with developing the mathematical architecture which will allow us to take advantage of this relationship between gravity, topography, and tectonics in studying the interior of a planet with a thick lithosphere. A

theory for computing stresses in a thick lithospheric shell is developed, and the stress state is reviewed for three types of models: topography that is isostatically compensated, topography that is flexurally loading the lithosphere, and buoyant forces that flexurally uplift the lithosphere. It is shown for long wavelengths that flexural stress trajectories depend mainly on lateral density distribution and less so on lithospheric elastic properties. For the isostatic state, surface stresses are determined entirely by the topography.

The results of this work have been used to investigate the lithospheric stress state on Mars and to constrain interior models of its lithospheric thickness and lateral density distribution. Mars is especially well suited for this type of analysis due to the fact that its long-wavelength topography and gravity fields are dominated by a single feature, the Tharsis plateau. The Tharsis province is a topographically high area roughly 4000 km square centered at 110°W longitude on the equator and is characterized by a pronounced free-air gravity high and extensive volcanism. It is flanked by broad basins with corresponding large free-air gravity lows [Lorell et al., 1972; Phillips and Saunders, 1975]. These features can be represented to a very good approximation by the longest-wavelength harmonics (l = 4), and for these harmonics the correlation between gravity and topography is quite high (see Figure 1).

Earlier versions of lithospheric stress fields computed for flexural loading models were presented by Phillips and Ivers [1979] and Phillips and Lambeck [1980] and suggested a relationship between stress trajectories and observed tectonic features. In this paper it is concluded that the major tectonic features of the Tharsis region are consistent with three stages in its history: one in which the load was fully compensated, a second time interval in which flexural support was operative producing extensive radial grabens, and a late stage with little regional lithospheric deformation.

State of Stress in a One-Plate Planet—Thick Shell Theory

Our interest here is in developing the appropriate relationships for displacement and stress in thick elastic planetary lithospheres.
where $G$ is the gravitational constant and the $T$ denotes transpose. The term $e_m$ is an eigenvalue of the solution of (1) subject to constraint by the imposed boundary conditions [Kaula, 1963]. It has the dimensions of density times length and is used to represent either an unknown perturbation on the depth of the crust-mantle boundary or an unknown lateral density perturbation in the upper mantle. The term $d$ is a length constant that is specifically accounted for in the integration of (1).

Equation (1) is solved subject to a set of boundary conditions at the surface and in the interior. The surface boundary conditions are the Dirichlet and Neumann conditions for the gravitational potential and conditions on normal and tangential stress. The condition for tangential stress at the free surface,

$$y_4 = 0$$

is actually applied at the deformed surface. In effect, this means that the influence of the elastic properties of the topography on the deformation of the lithosphere is ignored. The error introduced is of the order of the ratio of the total load thickness to the lithospheric thickness. The normal stress condition is applied at the actual deformed surface and includes the observed topography ($Rh_{lm}$) plus that part of the topography taken up by the deflection ($y_1$),

$$y_2 + g_0\rho_i(Rh_{lm} + y_1) = 0$$

where $g_0$ is gravity acceleration at the surface [i.e., $g_0(R)$], $\rho_i$ is the density of the topography, $R$ the radius of the planet, and $h_{lm}$ is a normalized spherical harmonic coefficient for topography.

In the isostatic case, $e_{lm}$ is not treated as an eigenvalue but is specified such that there is isostatic balance in the lithosphere (e.g., the three-element Pratt model discussed by Sleep and Phillips [1979]). In this case, the Dirichlet boundary condition is omitted to keep the system evenly determined, but this condition is implicitly satisfied in the choice of $e_{lm}$. Solutions obtained by this approach do satisfy a simple criterion for isostasy, namely, the flexural deflection $y_1 = 0$.

The elements of the stress tensor are evaluated at the free surface and are derived via a generalized Hooke's Law operator matrix, $R'$ [Arkani-Hamed, 1973]:

$$\begin{bmatrix}
\tau_{rr} \\
\tau_{\theta\theta} \\
\tau_{\phi\phi} \\
\tau_{r\theta} \\
\tau_{r\phi} \\
\tau_{\theta\phi} \\
\end{bmatrix}
T_{lm}(r, \theta, \phi) = R'(r, \theta, \phi)Y'(r)S_{lm}(\theta, \phi)$$

where $S_{lm}(\theta, \phi)$ is a spherical surface harmonic. The total stress tensor $T(r, \theta, \phi)$ is

$$T(r, \theta, \phi) = \sum_{l=1}^{l_{max}} \sum_{m=1}^{l} T_{lm}(r, \theta, \phi)$$

The resultant stresses are rotated into their principal directions and reduced to their deviatoric values.
Stresses in the Sublithosphere

The usual assumption in lithospheric (thin or thick shell) models is that the sublithosphere may be treated as a strengthless fluid of zero rigidity. However, if a long-wavelength load is applied to a planetary surface, it will initially be supported by the finite strength of the deep interior. After a finite amount of geological time, the load-imparted deep stresses will relax, the interior will behave elastically as strengthless, and the load will be borne entirely by the elastic lithosphere, which, by definition, is that portion of the planet which is able to support stresses over long periods of geologic time.

The relaxation time for stresses in the mantle can be bounded by (1) utilizing the most creep resistant of the reliable flow laws and (2) adopting an unusually low interior temperature. Presumably, all other reasonable choices will lead to even faster stress relaxation. The dry olivine flow law of Kohlstedt and Goetze [1974] and a mantle temperature of 1200 K yields, for a stress of 100 bars, a viscosity \( \eta \) of about \( 4 \times 10^{22} \) Pa.

The viscous relaxation time for stresses due to long-wavelength loads on a planet is given approximately by that due to a homogeneous sphere with viscosity equivalent to the sublithospheric mantle viscosity. That is, to first order, the elastic shell or lithosphere is transparent to the stress relaxation, which is also not affected by a liquid core. The relaxation time for a homogeneous sphere of the radius, say, of Mars and with a viscosity \( 4 \times 10^{21} \) Pa is about 3000 years for the harmonics considered in this paper (\( l = 2, 3, 4 \)). This is a geologically short span of time compared to major tectonic changes on a planet and implies essentially that the sublithosphere is not able to support stresses, i.e., may be treated as a medium of zero rigidity.

Because certain instabilities are introduced in the spherical elastic problem [Love, 1911], it is mathematically necessary to take the zero rigidity limit and not treat the sublithosphere as a region of low but finite rigidity. However, the coupled set of first-order differential equations (1) breaks down as \( \mu \to 0 \). This problem has been treated extensively in the literature in terms of the core of the earth [Longman, 1963; Saito, 1974; Crossley and Gubbins, 1975; Wunsch, 1975; Dahlen and Fels, 1978]. For our purposes, the core formulation of Longman [1963], including the core-mantle boundary conditions, is a sufficient description of the sublithospheric zero rigidity region. That is, our description of the sublithosphere is that of a fluid, where the core of the planet is distinguished from the sublithospheric mantle chiefly on the basis of density, and the fluid-to-elastic behavior boundary conditions are applied at the lithospheric boundary.

The equations in the fluid sublithosphere are reduced from the set in (1) and involve only the gravitational potential and its radial derivative. A finite solution at the origin requires only one unknown constant in the solution for the fluid region, and there are two constants introduced at the lithospheric boundary [see Longman, 1963]. The unknown eigenvalue \( e_{in} \) introduces a fourth unknown, and the system is integrated as a three-point (origin, lithospheric boundary, surface) boundary value problem constrained by the four surface boundary conditions for (1), as discussed above.

Stresses in the Lithosphere

The stress field in the lithosphere can be thought of as a superposition of three components: (1) a compensated component, which will depend only on the load, (2) a flexural component, which depends on the displacement field and therefore the elastic properties and thickness of the lithosphere as well as the load, and (3) a hydrostatic component, which will have no effect on the stress differences which produce tectonic features.

Of critical importance to the problem of supporting stresses in the lithosphere is its finite strength, since rocks subjected to stress differences greater than this value will undergo brittle failure. This subject has been discussed at
Fig. 3. Stress results for western longitudes from flexural uplift model (Figure 3a), isostatic model (Figure 3b), and flexural loading model (Figure 3c). Arrows show directions of maximum and minimum principal deviatoric stress. Contours give magnitude of maximum stress difference in kilobars.

length by Phillips and Lambeck [1980]. They conclude that the crustal finite strength of Mars is of the order of 300–500 bars, although this must be considered a lower bound.

Isostasy is usually defined as a weight balancing of surface loads by density anomalies below it in such a way that at some uniform depth the pressure is everywhere constant. In order for this condition to hold in a local sense, it is necessary that shear stresses on vertical planes vanish (in
spherical coordinates $\tau_\theta = \tau_\phi = 0$ so that adjoining columns do not influence one another. On a sphere there are several ways of imposing this constraint in isostatic calculations. A common method is to allow the rigidity of the layer to approach zero and then calculate vertical displacements. Another way is to set $\tau_\theta = \tau_\phi = 0$ and $\tau_\phi = \tau_\phi$ so that there is no shear stress on any vertical plane. However, most stress information is lost, particularly by the second approach since only $\tau_\theta$ and $\tau_\phi$ are obtained. Since we are interested in the orientation of the isostatic stress field and not the displacement of a strengthless crust, we instead require that the vertical load displacements are zero by exactly balancing the surface load by internal buoyancy forces.

The resulting stress field will, to first order, be that which maintains the topographic relief against gravitational slumping and will depend only on the shape, magnitude, and density of the topography itself. In the case of a circularly symmetric load, the stresses will be primarily radial and tensile near the center, grading to circumferential and tensile (‘hoop stresses’) at the edges and radial and compressive beyond the load boundary. We emphasize that the stress field of the surface for an isostatic state does not depend at all on the subsurface structure. In particular, this result does not depend on the type of compensation, be it Pratt or Airy, and the specific internal density distribution.

It is generally assumed implicitly that the stress differences in the isostatic state are small enough that the finite strength of the lithosphere is easily able to support them. This is not necessarily the case. Jeffreys [1932, 1943] solved the problem of a floating crust as a mechanical equilibrium problem (without reference to an elastic stress-strain relation) in order to find the density distribution which would lead to the minimum stress. For both plates and spherical shells he found that the minimum stress configuration was nearly isostatic and the minimum value for the maximum stress difference is

$$\tau_{\text{max}} = \rho g h$$  \hspace{1cm} (7)

where $h$ is the harmonic coefficient of the topography, or

$$\tau_{\text{max}} = \frac{1}{2} \rho g H$$  \hspace{1cm} (8)

where $H$ is the height of the feature above its surroundings. Equations (7) and (8) can be generalized to include stress differences caused by density anomalies within the lithosphere by computing an equivalent topographic height corresponding to the anomaly.

It is possible to have a maximum stress less than that predicted by (7) if there is active construction of topography by lithospheric compressive stress, as possibly in the Tibetan plateau on earth [Bird, 1978; Molnar and Tapponier, 1978]. In general, however, large horizontal lithospheric stresses that are unrelated to density heterogeneities in the lithosphere are unlikely on planets that do not have plate tectonics. Global thermal contraction, which produces an isotropic and homogeneous membrane stress, is a possible exception [see Solomon, 1978].

Pure isostasy is an end-member of a suite of flexural states involving varying degrees of compensation. In general, the isostatic stresses generated by the load will be added to the flexural stresses due to strains within the lithosphere. Although a system will always tend toward isostasy in order to decrease its stresses [Phillips and Lambeck, 1980], the finite strength of the lithosphere ensures that there will always be some component of stress.

Some general properties of flexural models can be investigated using thin shell theory [see, e.g., Turcotte et al., 1981]. In particular, since membrane stresses dominate bending stresses for long-wavelength loading, then the stress traject-
For \( l = 2, 3, \) and \( 4, \) \( g_l = 1, \) so that \( \alpha_l \) is approximately independent of \( l. \) The shape of the displacement field \( W \) depends on the relative contributions of the harmonic coefficients \( W_{lm}. \) If \( \alpha_l \) is independent of \( l, \) then the shape of \( W \) only depends on the shape of the load, as spectrally characterized by the coefficients \( h_{lm}, \) and is not dependent on the properties of the lithosphere (e.g., \( E, t_L, \Delta \rho \)) represented in \( \alpha_l. \) The magnitude of the displacement relative to the load is still dependent, however, on the approximately \( l \)-independent value of \( \alpha_l \) given in (11). The stresses are obtained directly from the displacement field, so it follows that the stress trajectories will be relatively insensitive to the properties of the lithosphere.

The subsurface density distribution (i.e., lateral density anomalies), on the other hand, may have a significant effect on stress trajectories. This is because the density anomalies themselves modify the load on the lithosphere. The magnitude of this load will depend on the depth assigned to these inhomogeneities, since deeper anomalies must necessarily be larger in order to satisfy the gravity boundary condition. Since the gravity does not correlate perfectly with the topography, the observed stress trajectories will be some combination of a pure topographic load and a pure gravitational load. This is in contrast to either the compensated case or the case of a homogeneous lithosphere flexurally loaded of the surface, which, as discussed above, depend only on the topography.

**APPLICATION TO MARS**

**Introduction**

Phillips and Saunders [1975] found that the gravity and topography of Mars could be divided into two distinct populations. One consists of the Tharsis plateau and the adjoining Amazonis and Chryse basins, and the other is composed primarily of older terrane and covers most of the rest of the planet. This latter population appears to be largely compensated at a depth no greater than 100–150 km, assuming an Airy type model of isostasy. Based on spectral properties do not depend on the elastic properties of the lithosphere. This is seen in the spherical harmonic coefficient \( \alpha_l, \) which for thin shell theory relates elastic displacement \( W_{lm} \) to the applied load \( h_{lm}, \)

\[
W_{lm} = \alpha_l h_{lm}
\]  

where

\[
\alpha_l = \frac{\rho_o \rho R^4 f_l + 1 - \nu) D}{f_l^3 + 4f_l^2 + \psi(1 - \nu)(f_l + 2) + \Delta \rho_o R^4 (f_l + 1 - \nu) D}
\]  

where \( f_l = -l(l + 1), \) the flexural rigidity \( D = E t_L^3/12(1 - \nu), \) \( \psi = 12R^2 t_L, \) \( t_L \) is the thickness of the elastic shell, \( \Delta \rho \) is the density contrast with the sublithosphere, and \( E \) and \( \nu \) are Young's modulus and Poisson's ratio, respectively. For long wavelengths (small \( l), \) the first two terms in the denominator of (10), which represent the contribution of bending stresses, can be neglected relative to the last two terms, which relate to membrane stresses. In this case, \( \alpha_l \) reduces to

\[
\alpha_l = \frac{-\rho_o \rho R^2}{E t_L g_l + \Delta \rho_o R^2}
\]  

where

\[
g_l = 1 + \frac{(1 + \nu)}{(f_l + 1 - \nu)}
\]  

**Fig. 5.** Maximum stress difference as a function of planetary radius for \( C_{22} \) harmonic using isostatic model. The term \( h_{op} \) is the observed topographic height and \( h_{op} \) is the equivalent height of the topography plus the density anomaly at the crust-mantle interface.
considerations, Lambeck [1979] showed that the stresses implied by the non-Tharsis component of the gravitational potential are no larger than those found in the earth's compensated topography, even if all density anomalies are assumed to be near the surface. We will, therefore, concentrate our attention on the Tharsis region.

Phillips and Saunders [1975] evaluated isostasy, using spherical harmonics through degree and order 8 based on gravity from Sjogren et al. [1975] and topography of Christensen [1975]. They found that complete Airy compensation at a single depth would not satisfy the data. Depths of compensation of 1100 and 600 km were required for the second and third degree tesseral harmonics, respectively, while a depth of less than 100-150 km was needed for the $l = 4$ through $l = 8$ harmonics, corresponding mostly to ancient (non-Tharsis) terrane. Because the wavelengths are much greater than the depths of compensation, their analysis is valid for Pratt models as well. The great disparity between these depths argues against a simple Pratt or Airy model of compensation. Either (1) a more complex type of compensation (passive or dynamic) must be invoked [e.g., Sleep and Phillips, 1979] or (2) Tharsis is only partially compensated, implying some degree of flexural support.

Numerical Experiments

Three classes of models have been investigated in an attempt to gain some insight into the structure and history of the Tharsis region: flexural uplift model, an isostatic model, and a flexural loading model. These models are shown schematically in Figure 2. In each case $\rho_s = \rho_0 = 3.0$ g/cm$^3$, $\rho_m = 3.5$ g/cm$^3$, and $\lambda = \mu = 5 \times 10^{11}$ dyne/cm$^2$ in the lithosphere. The term $\Delta \rho$ in Figure 2 represents a lateral density anomaly in the lithosphere, determined for a flexural uplift solution or set by the requirements for compensation in an isostatic solution. The topography of Bills and Ferrari [1978] and the gravity of Sjogren et al. [1973], complete through degree and order 4, were used as boundary conditions (Figure 1). More recent gravity models are available [Gapcynski et al., 1977; Christensen and Baimino, 1979], but they are not significantly different at these long wavelengths. Likewise, new earth-based radar determinations of Tharsis elevations [Downs et al., this issue] do not have an important effect on the spherical harmonic coefficients for long-wave-length topography. We have examined the effect on the solutions of including the contribution of the nonhydrostatic contribution to the second zonal harmonic $J_2$ [Kaula, 1979]. We found only a small change (<5%) in stress magnitude and a negligible change in stress trajectories.

In order to interpret the stress trajectory plots, the following conventions should be borne in mind. All stress results are shown in terms of horizontal principal deviatoric stresses at the surface. Thus the maximum stress $\sigma_1$ will always be positive (tensile), and the minimum stress $\sigma_3$ will always be negative (compressive). We have plotted both the $\sigma_1$ and $\sigma_3$ directions (the intermediate stress $\sigma_2$ does not generally enter into failure criteria). If only one stress direction is shown, the conjugate (maximum or minimum) principal stress direction is vertical. When displayed in this way, the relationship between the stresses and the expected faulting (using the criteria of Anderson [1951]) is straightforward. Diverging arrows indicate normal faulting, or graben formation. Converging arrows indicate reverse faulting, or compressional ridge formation. In both cases the features are expected to be orthogonal to the stress directions. If both stress directions are shown, the strict interpretation is strike slip faulting along planes roughly bisecting the angle between $\sigma_1$ and $\sigma_3$. However, this state is very sensitive to the vertical load, and local perturbations in the topographic height could cause either normal or reverse faulting in the directions shown. Note that while principal stresses are always orthogonal, this relationship between the stress trajectories is distorted at high latitudes by the mercator projection. Contours show the magnitude of the maximum stress difference $\sigma_1 - \sigma_3$. This is the relevant parameter in

Fig. 7. Maximum surface stress difference in kilobars and maximum load thickness in kilometers as a function of lithospheric thickness from flexural loading model.
most failure criteria and is equal to twice the maximum shear stress.

In the flexural uplift model, the present topography is caused by flexural doming of the lithosphere in response to an upward force. We have modeled this with a buoyant force due to a Pratt-type overcompensation, but other modes of uplift give virtually identical stress trajectories at the surface. These stress trajectories (shown for a lithosphere thickness of 200 km in Figure 3a) predict east-west grabens in the highest part of Tharsis, showing virtually no correlation with the observed tectonic features; the observed grabens are primarily radial to Tharsis. Since the predicted stress differences are great enough to induce fracture (i.e., >500 bars), we conclude that either flexural uplift played no major role in the history of Tharsis, or all obvious tectonic evidence for such an episode has been obliterated.

Stress trajectories for the isostatic case are shown in Figure 3b. The density contrast Ω0 was determined using the three-element Pratt isostatic model proposed by Sleep and Phillips [1979], but as discussed earlier, the exact choice of models has no effect on the surface stresses. In this model, the mean crustal thickness of 150 km thins to about 100 km below Tharsis, the lithospheric thickness is 400 km, and the maximum value of Ω0 is about 0.1 g/cm². The stresses correlate very well with tectonics out to a radius of about 40° from the center of Tharsis, but the correlation breaks down at greater distances. This can be seen clearly in Figure 4, which shows the stresses from Figure 3b along with major tectonic features mapped by Scott and Carr [1978] in the Valles Marineris region and those shown on the USGS shaded relief map of the Phaethontis quadrangle, southwest of the main Tharsis elevation.

Figure 5 shows a plot of the maximum stress difference as a function of planetary radius. The Jeffreys prediction (equation (7)) is borne out when the sum of the surface and equivalent subsurface topography is considered (hₐ). The maximum stress difference near the surface is predicted by the topography alone (hₐ).

A series of cases were run using the flexural loading model in order to check the prediction by thin shell theory that stress trajectories should be insensitive to variations in the physical parameters. This was found to be the case for elastic constants in the range of 10¹⁰–10¹² dynes/cm², crustal thicknesses from 50 to 150 km, and lithospheric thicknesses from 100 to 400 km. The stress trajectories shown in Figure 3c were computed using a crustal thickness of 150 km and a lithosphere thickness of 200 km. The anomalous crustal thickness (from the density eigenvalue) was small, about +2 km relief on the crust-mantle boundary under Tharsis. Depending on the choices of crustal (tₐ) and, particularly, lithospheric (tₗ) thicknesses, it is possible to achieve a range of eigenvalues in the flexural solution from negative through positive. Those specific values of tₐ and tₗ which drive the stress to approximately zero provide a 'pure' flexural solution. That is, both the topography and gravity can be explained by the loading of a laterally homogeneous elastic lithosphere. This is analogous to the thin shell results of Willemann and Turcotte [this issue].

The flexural loading stress trajectories are complementary to those of the isostatic case. For distances greater than about 40° the tectonic features are well predicted, but closer to Tharsis the model fails (Figure 6). This is most striking in the Valles Marineris region, where the 'hand-off' from one stress regime to the other occurs quite sharply at 70°W longitude. However, the sudden change in horizontal stress from compressive to tensile, or vice versa, occurring at numerous locations (Figure 3), is somewhat misleading without considering the intermediate stress Ω2. For example, in the western part of Valles Marineris (Figure 6a), the intermediate stress is horizontal and tensile; it increases in magnitude from west to east. In the vicinity of 75°W longitude, Ω2 is approximately equal to Ωₙ in magnitude. Therefore, continuing eastward, it exceeds Ωₙ and becomes Ωₙ, while Ωₚ becomes Ω₂, i.e., the intermediate stress is horizontal and compressive. Thus, when considering all three principal stresses, the geographical variation in stress pattern is functionally smoothly varying. Nevertheless, the isostatic and flexural stress patterns are significantly different and match the geology well in distinct geographic localities.

For the flexure model the stresses are undoubtedly large enough to cause extensive failure near the surface. This is true even for thick lithospheres, as is shown in Figure 7. Also shown in Figure 7 is the thickness of the constructional load (observed topography plus computed deflection) as a function of lithospheric thickness. For relatively thin lithospheres, an enormous thickness of volcanics is necessary, which would require the magmatic depletion of a not insignificant fraction of the mantle volume.

**Discussion**

We have investigated a suite of models intended to represent a broad spectrum of possible modes of static support for the Tharsis region of Mars. We consider that dynamic models are less plausible because of the requirement that specific topographic features be supported for significant fractions of Martian history. Of the three basic passive configurations: (1) flexural uplift, (2) isostasy, and (3) flexural loading, only the first type appears to be ruled out by the surface geology. Both isostasy and flexural loading appear to be represented by the orthogonality of tensile stress trajectories with observed grabens.

The grabens are, in fact, consistent with some very general statements concerning lithospheric disposition and normal faulting. In an isostatic configuration, there will be a tendency for grabens to form on and be confined to positive topographic features. This is due to edifice stresses, or simply stated, due to the tendency of stress to remove topographic variations [Artyushkov, 1973]. Grabens can potentially form in all directions given an isotropic distribution of topographic gradient vectors. If there is a preferred map direction of high topographic gradients, then grabens will tend to form with strikes perpendicular to these gradient directions. Thus, examination of the topography in Figure 2b leads to the immediate conclusion, under the assumption of isostasy, that north-south graben will form on Tharsis,
tending toward north-northeast strikes in the northern latitudes. This is confirmed by both the stress trajectories in Figure 3b and the observed geology.

Long-wavelength flexural loading on Mars would tend to create radial grabens, due to membrane stresses, well outside the Tharsis load under a variety of assumptions regarding lithospheric parameters. This result is seen here (Figure 3c) as well as in the thin shell results of Willemann and Turcotte [this issue]. That this flexural regime has operated sometime in the history of Mars is suggested by the orthogonality of flexural stress trajectories and grabens in the regions well outside the Tharsis province.

Since the isostatic and flexural regimes both appear to be represented in the observed tectonics, we suggest that both modes existed but during different times in Martian history. A fundamental question concerns why it is appropriate to use the present-day topography and gravity fields to construct stress models to constrain earlier regimes of lithospheric tectonics. That is, do we have any way, in general, of supposing that gravity and topography were the same when the various grabens formed? This question should be addressed separately to the isostatic and to the flexural states.

As stated above, under general isostatic conditions, grabens would tend to form on the elevated regions of Tharsis, and their presence argues that isostasy prevailed at some point in the evolution of this region of Mars. The only question is whether the orthogonal relationships between isostatic stress trajectories and grabens are a coincidence; i.e., was the topography substantially different when the grabens formed? There are at least (at least) three possibilities: (1) the general north-south elongation of Tharsis with steeper east-west slopes has existed for a substantial portion of the history of Tharsis, (2) the present relationship between topography and stress is coincidental, and the north-south grabens represent a fundamental anisotropy in lithospheric strength which dominated over whatever topographic shape existed at the time of graben formation, or (3) the relationship is coincidental. The strike of the grabens was in a number of directions, but all except the north-south grabens have been obliterated by subsequent volcanic events. We have no way of deciding between these three possibilities, but it is our view that they are listed above in terms of decreasing plausibility.

As also stated above, under general conditions of flexural loading, radial graben will form on the perimeter of Tharsis. Again, the question is whether the observed orthogonal relationships between stress trajectories and grabens are coincidental, and here we are somewhat more concerned with the detailed shape of the load. For example, the youngest volcanic plains unit on Tharsis is younger than the radial grabens. If there is a causal relationship between the calculated stresses and the grabens, then one would conclude that the addition of the youngest volcanic unit had only an insignificant effect on the direction of previously existing stress trajectories. Additionally, the long-wavelength gravity field also contributes to the stress field in the flexural case, and the same arguments apply as to the suitability of the present gravity field in stress modeling. However, the issue is less critical because the generally small magnitude of the eigenvalues in the solution of equation (1) suggests that the flexure is dominated by the topographic load; i.e., the gravity field due to the internal density distribution plays only a secondary role.

In summary, there is no way to demonstrate unequivocally that the present topography and gravity field are appropriate to predict ancient tectonic events. However, it is certainly a plausible assumption. More importantly, the argument can be turned around. That is, we dismiss the argument that the orthogonality between tensile stress trajectories and grabens is a coincidence and use this observation as the strongest argument that our procedure has been appropriate, and thus we can constrain the style of compensation of the Tharsis province.

We close this paper with a hypothesis for the evolution of Tharsis. It is based on the suppositions that (1) the north-south grabens are older than the radial graben [e.g., Wise et al., 1979], and (2) there has been only minor graben formation since the emplacement of the youngest volcanic plains unit. If this is true, then the evolution of Tharsis proceeded from a state of isostatic compensation to one of flexural loading with lithospheric failure to one of flexural loading without attendant graben formation. Such a scenario is consistent with a lithospheric monotonically increasing in thickness over the evolution of Tharsis. This is, of course, the expected course of events as heat generation declines over the history of the planet.

Figure 8 is a cartoon of the three major stages in the evolution of Tharsis according to this hypothesis. The isostatic state may represent the time when most of the elevation of Tharsis was attained. According to stage I, Tharsis is built by a combination of constructional volcanics and isostatic uplift. The first of these processes has been discussed by Finnerty and Phillips [1981], who review volcanic construction in a system wherein all the differentiated igneous products remain (vertically) above the source region. This situation tends to be isostatic because mass is neither added nor removed from the Tharsis region. Whether complete isostasy can be achieved depends upon the thermal state of the lithosphere. Construction takes place due to a volume increase in the conversion of mantle minerals to a less dense crustal mineralogy. If some of the magmatic products move laterally outside the Tharsis region, then there will be isostatic uplift with only a small fraction of magma removed required to generate a significant uplift contribution.

Since isostatic stresses are proportional to topographic height (equation (7)), at some point in the construction/uplift process the stresses would exceed the finite strength of the near-surface rocks, and grabens would form in the elevated regions.

In stage II, the lithosphere has thickened to the point that complete isostatic compensation of the growing volcanic pile did not take place. This may also be a time when magmatic material is added from regions in the interior that are not immediately beneath Tharsis. That is, mass may not be conserved in the Tharsis region, and thus the isostatic conditions described by Finnerty and Phillips [1981] may not be operative. The volcanics at this stage behave as a flexural load on the lithosphere. The load increases to the level that lithospheric failure takes place in the form of grabens radial to Tharsis.

By stage III time, the lithosphere has thickened to the extent that the addition of new volcanics does not significantly perturb the stress levels, and no additional regional failure takes place. These new units are represented by the youngest volcanic plains and the shield volcanoes, the latter
References


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