

Lithospheric flexure and the evolution of the dichotomy boundary on Mars

Thomas R. Watters

Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington, D. C., USA

Patrick J. McGovern

Lunar and Planetary Institute, Houston, Texas, USA

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[1] The boundary of the Martian crustal dichotomy in the eastern hemisphere is one of the most striking topographic features on the planet. The long wavelength topography of much of the boundary is expressed by a broad rise and an arched ramp that slopes downward from the southern highlands into the northern lowlands and often ends in a steep scarp. Lithospheric flexure of the southern highlands for both a continuous and broken plate boundary with the northern lowlands lithosphere is modeled. We find that the long wavelength topography of the boundary is best fit by a lithospheric deflection profile of a broken lithosphere for reasonable values of the flexural parameter. The lithosphere near the dichotomy boundary may have been weakened by tectonic stress associated with the formation of the crustal dichotomy by subcrustal transport caused by overturn of an early magma ocean. **Citation:** Watters, T. R., and P. J. McGovern (2006), Lithospheric flexure and the evolution of the dichotomy boundary on Mars, *Geophys. Res. Lett.*, *33*, L08S05, doi:10.1029/2005GL024325.

1. Introduction

[2] The Martian hemispheric dichotomy is one of the dominant physiographic features on Mars, only rivaled by the Tharsis volcanic and tectonic province. The dichotomy is expressed in the topography, geology, tectonics, and crustal structure. The transition from the relatively featureless northern lowlands to the heavily cratered southern highlands in the eastern hemisphere is sharply delineated by a boundary where the elevation changes over 2 km (Figure 1). In this hemisphere, the dichotomy boundary is also marked by tectonic features. Fault-controlled fretted valleys and extensional troughs are found in the lowlands and lobate scarp thrust faults in the adjacent highlands [McGill and Dimitriou, 1990; Watters, 2003a, 2003b]. The compressional strain in Amenthes - northern Terra Cimmeria is estimated to be $\sim 0.2\%$ [Watters and Robinson, 1999] and the extensional strain in the northern Arabia Terra - Ismenius region may be on the order of $\sim 0.4\%$ [see Smekar *et al.*, 2004]. These tectonic features appear to have formed during the Late Noachian to Early Hesperian [Maxwell and McGill, 1988; McGill and Dimitriou, 1990; Watters and Robinson, 1999; Watters, 2003b]. The presence of fault scarps and fault-controlled valleys along the dichotomy

boundary suggests that tectonism played a role in its formation and/or modification. A population of subdued circular depressions in the northern lowlands revealed in Mars Orbiter Laser Altimeter (MOLA) data interpreted to be ancient buried impact basins suggests that the northern lowlands crust and the crustal dichotomy formed in the Early Noachian [Frey *et al.*, 2002; Frey, 2006]. This suggests that the present-day boundary has evolved from an ancient dichotomy boundary modified by tectonism, volcanism, deposition and erosion [Watters, 2003a, 2003b; Irwin *et al.*, 2004].

2. Topography of the Dichotomy Boundary

[3] We examined the long wavelength topography along the length of a ~ 2100 km section of the dichotomy boundary in northern Terra Cimmeria (Figure 1). The study area encompasses a ~ 200 km long section of the boundary studied previously [Watters, 2003a]. Along much of its length, the dichotomy boundary consists of a broad rise and an arching ramp (Figure 2). The broad rise is generally asymmetric in profile with gentle slopes on the southern flanks or back rise that slope away from the dichotomy boundary. The northern flanks of the broad rise transition into a relatively steep ramp that slopes down into the lowlands and often terminates in an escarpment (Figure 2). The maximum slopes are reached on the ramp and the scarp that marks the dichotomy boundary.

3. Flexure of Continuous and Broken Lithosphere

[4] Lithospheric flexure results in long-wavelength topography with a distinct deflection profile. Vertical loading results in the downward deflection of the lithosphere accompanied by a flanking upwarp or bulge [see Turcotte and Schubert, 2002; Watts, 2001]. Lithospheric flexure is modeled using analytic solutions for an elastic plate overlying an incompressible fluid subjected to a line load. We modeled flexure of both a broken and a continuous plate. A broken plate is defined as having a vertical plane where both the shear stress and the bending moment are zero. The deflection of a continuous or infinite elastic plate under a line load is given by

$$w = w_0 e^{-x/a} \left(\cos \frac{x}{\alpha} + \sin \frac{x}{\alpha} \right) \quad (1)$$

where w_0 is the maximum amplitude of the deflection at $x = 0$ [Turcotte and Schubert, 2002]. The flexural

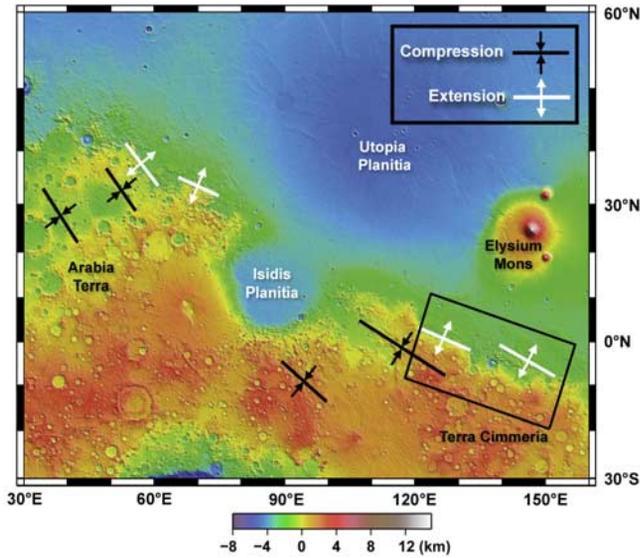


Figure 1. Location of extensional (troughs) and compressional (lobate scarps) tectonic features along dichotomy boundary in the eastern hemisphere overlaid on a color-coded shaded-relief map derived from MOLA $1/128$ degree per pixel resolution gridded data. Black box shows the approximate location of the study area.

parameter α is related to the flexural rigidity and the elastic thickness T_e by

$$\alpha = \left(\frac{ET_e^3}{3(1-\nu^2)(\rho_m - \rho_c)g} \right)^{\frac{1}{4}} \quad (2)$$

where ρ_m and ρ_c are the density of the mantle and overlying material, respectively, g is the acceleration due to gravity, ν is Poisson's ratio, and E is Young's modulus. The deflection of a broken or semi-infinite elastic plate is given by

$$w = w_0 e^{-x/a} \left(\cos \frac{x}{\alpha} \right) \quad (3)$$

The most pronounced difference between the deflection of a continuous plate and a broken plate supporting a line load is in the amplitude of the deflection. For a given load the amplitude of the deflection or maximum height of the rise for a broken plate is twice as large as for a continuous plate [see *Turcotte and Schubert, 2002; Watts, 2001*]. In addition to a smaller maximum height of the rise, the slope of the downward deflection or ramp is less for a continuous lithosphere. We compared model results with the long wavelength topography of two well-preserved areas of the dichotomy boundary in northern Terra Cimmeria (eastern and western sections of the study area). These areas exemplify the rise and ramp morphology shared by other sections of dichotomy boundary in the eastern hemisphere (e.g., the northern Arabia Terra–Ismenius region [see *Watters, 2003b, Figure 9*]). A series of deflection profiles were generated over a range w_0 and α for a continuous lithosphere. Although the maximum downward deflection and ramp slope in eastern (Figure 3a) and western (Figure 3b) Terra Cimmeria can be obtained for reasonable

parameter values (east: $\alpha = 115$ km, $w_0 = 130$ m; west: $\alpha = 95$ km, $w_0 = 118$ m), in both cases the height of the rise is underestimated. Better fits to the topography are obtained with deflection profiles for a broken plate (Figures 3a and 3b) (east: $\alpha = 155$ km, $w_0 = 220$ m; west: $\alpha = 154$ km, $w_0 = 387$ m). The results for a broken lithosphere are consistent with those obtained modeling flexure of an elastic plate subjected to end load [*Watters, 2003a*]. The ramp slope in both models is sensitive to relatively small (tens of kilometers) variations in the flexural parameter, decreasing with increasing α . From equation (2), the best fit flexural parameters correspond to an elastic thickness T_e of ~ 30 km, assuming the mean density of the highland of $2900 \text{ kg} \cdot \text{m}^{-3}$, a density of the Martian mantle of $3400 \text{ kg} \cdot \text{m}^{-3}$, and a Young's modulus of the lithosphere of $E = 100$ GPa.

[5] We suggest that Late Noachian - Early Hesperian flexure of the highlands was induced by vertical loading along the dichotomy boundary from the emplacement of volcanic material in the northern lowlands [*Watters, 2003a*]. There is evidence of buried ridged plains in the northern lowlands that appear to be similar to Late Noachian - Early Hesperian volcanic units in the southern highlands [*Zuber, 2001; Withers and Neumann, 2001; Head et al., 2002*]. This northern lowlands ridged plains volcanic material is estimated to be up to several kilometers thick [*Head et al., 2002*]. Gravity data from Mars Global Surveyor indicates large positive free-air anomalies along the dichotomy boundary in the eastern hemisphere [see *Zuber et al., 2000; Yuan et al., 2001*]. Positive bouguer gravity anomalies paralleling the dichotomy boundary in the northern lowlands in the eastern hemisphere [*Neumann et al., 2004, Figure 3*] may indicate relatively thick accumulations of basalt-like volcanic material. Thus, this volcanic sequence is the source of the modeled line load.

4. Implications of a Broken Lithosphere

[6] The analytic model fits (Figures 3a and 3b) suggest a weakened or broken lithosphere at the dichotomy boundary. The question then becomes how the lithosphere may have been broken and how this is related to the origin of the crustal dichotomy. One way a broken lithosphere could have been formed is an early stage of plate tectonics [*Lenardic et al., 2004*] where the dichotomy boundary in

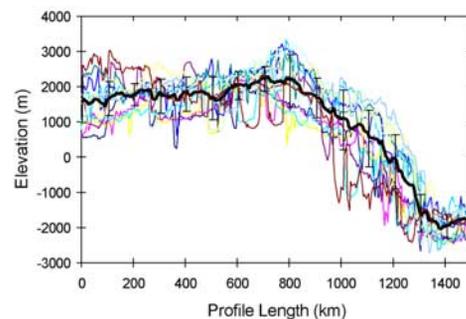


Figure 2. Long wavelength topography of the dichotomy boundary in northern Terra Cimmeria. The black line is the mean of 13 profiles across this 2,100 km segment with ± 1 standard deviation error bars. Vertical exaggeration is $\sim 136:1$.

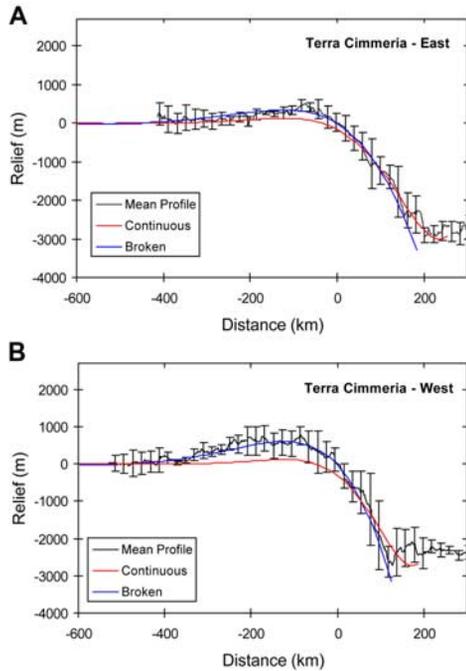


Figure 3. Topographic profile across dichotomy boundary compared to deflection profiles for a continuous and broken plate. (a) Topographic profile in eastern Terra Cimmeria (black curve) is the mean of the 4 profiles with ± 1 standard deviation error bars. (b) Topographic profile in western Terra Cimmeria (black curve) is the mean of the 7 profiles with ± 1 standard deviation error bars. Best fit models for a continuous and broken plate are shown (see text for parameters). The orientation of the topographic profiles is parallel to the sides of the box shown Figure 1. Vertical exaggeration of plots is $\sim 85:1$.

the eastern hemisphere is analogous to a terrestrial passive margin [see *Sleep, 1994; Watters, 2003a*]. Another possibility is that the crustal dichotomy formed by early transport of subcrustal material by degree-one mantle convection which thinned and thickened the crust above zones of upwelling and downwelling [*Zhong and Zuber, 2001; Zhong et al., 2004*]. New evidence, however, indicates the crust of Mars formed very soon after planetary accretion, within about 50 My of solar system formation [*Solomon et al., 2005*]. This suggests there may not have been sufficient time for early plate tectonics or convection driven subcrustal transport to have formed the northern lowlands crust [*Solomon, 2004*]. The rapid formation time suggests the crust formed by differentiation of a global magma ocean [*Solomon et al., 2005*]. Crystallization of an early magma ocean could result in a gravitationally unstable mantle that rapidly overturned [*Elkins-Tanton et al., 2003, 2005*]. If the scale of mantle overturn was hemispheric, the crustal dichotomy could have formed by very early subcrustal transport.

[7] Subcrustal transport results in lithospheric stresses analogous to those generated by surface erosion, where extension occurs where crust is removed and compression occurs in the adjacent lithosphere [*Stein et al., 1979*]. Because subcrustal transport results in the replacement of crust with mantle material, compression is induced in the lowlands and extension induced in the highlands. The fiber

stresses resulting from subcrustal transport assuming a semi-infinite load are given by

$$\sigma_{xx} = \frac{Eh\rho_r z}{\alpha^2(\rho_m - \rho_c)} \sin\left(\frac{x}{\alpha}\right) \exp\left(-\frac{|x|}{\alpha}\right) \quad (4)$$

where z is the vertical distance in the lithosphere relative to the neutral surface defined as positive up, ρ_c is the density of the crust, ρ_m is the density of the mantle, α is the flexural parameter, and h is the depth of the material removed [*Stein et al., 1979*].

[8] The removal of 3.5 km of lowlands crust results in fiber stresses with a near-surface maximum >400 MPa assuming values for the flexural parameter and elastic thickness consistent with those obtained from the flexural models ($\alpha \cong 154$ km, $T_e = 30$ km, $\rho_c = 2900$ kg m $^{-3}$, $\rho_m = 3400$ kg m $^{-3}$) (Figure 4). These stresses are sufficient to cause a zone of significant normal and thrust faulting along the early Noachian dichotomy boundary and may have resulted in a broken lithosphere.

[9] The Late Noachian – Early Hesperian age of the tectonic features along the present-day dichotomy boundary indicates that they are not the result of extension and compression related to the early formation of the dichotomy (Figure 1). The surface expression of any tectonic features formed in the very Early Noachian would not be preserved. The highland lobate scarps and the fractures and fault controlled valleys in fretted terrain in the lowlands reflect tectonic stresses related to the modification of the dichotomy boundary [*Watters, 2003a, 2003b; Guest and Smrekar, 2004; McGovern and Watters, 2004; Nimmo, 2005*]. Flexure induced bending stresses reach a maximum in the near-surface of >400 MPa several hundred kilometers east of the rise (near the base of the scarp) for $T_e \sim 30$ km [*Watters, 2003a*]. Compressional stresses reach a maximum of only about 20 MPa in the highlands, ~ 500 km from the base of the scarp. Estimates of the stress necessary to initiate thrust faulting at even relatively shallow depths (2 to 3 km) are on the order of 100 MPa [see *Watters et al., 2000*]. This suggests that other compressional stresses must have contributed to the formation of the lobate scarp thrust faults [*Watters, 2003a, 2003b; McGovern and Watters, 2004*].

[10] Erosion of highland material at the dichotomy boundary scarp and from degradation of the highlands are

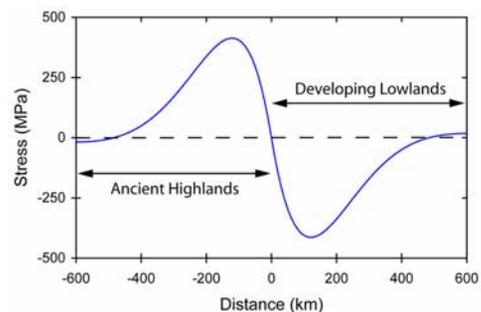


Figure 4. Very Early Noachian near-surface extensional stresses (positive) due to subcrustal transport of 3.5 km of lowlands crust reach a maximum in the highlands ($x < 0$) while compressional stresses (negative) reach a maximum in the developing lowlands ($x > 0$).

possible sources of compressional stress. Although compressional stresses of >100 MPa may be generated by the erosion of several hundred kilometers of highland material along the dichotomy boundary [Watters, 2003a], such a large amount of erosion would have substantially erased the topographic expression of the flexure. Erosion of several hundred meters of material from the highlands could result in an additional few MPa of compression, assuming uniaxial strain [Turcotte and Schubert, 2002]. However, studies of degradation of the highlands suggest that eroded material has been locally redistributed, not removed [Craddock et al., 1997].

[11] A likely source of compressional stress on Mars during the Late Noachian - Early Hesperian is global contraction. Volcanic resurfacing and tectonism peaked during the Late Noachian - Early Hesperian [Greeley and Schneid, 1991; Watters, 1993; Head et al., 2002]. This is reflected by the global-scale emplacement and subsequent deformation of ridged plains volcanics, including those in the northern lowlands [Watters, 1993; Head et al., 2002]. Thus, a compressional stress bias in the highlands from global contraction may have played an important role in the tectonic modification of the ancient dichotomy boundary.

[12] The scenario proposed here for the formation of the crustal dichotomy suggests that significant tectonic stresses in the very Early Noachian resulted in a broken lithosphere along the developing dichotomy boundary. The ancient dichotomy boundary was modified by loading from the emplacement of Late Noachian - Early Hesperian volcanic material in the northern lowlands. Loading of the northern lowlands induced flexure in the southern highlands. The combination of bending stresses and a Late Noachian - Early Hesperian compressional stress bias in the lithosphere formed the tectonic features associated with the dichotomy boundary in the eastern hemisphere.

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P. J. McGovern, Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston, TX 77058, USA.

T. R. Watters, Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington, D. C. 20560, USA. (watters@si.edu)