

Hemispheres Apart: The Crustal Dichotomy on Mars

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Key Words

geology, topography, gravity, crustal thickness, tectonics

Abstract

The hemispheric dichotomy is a fundamental feature of Mars, expressed by a physiographic and geologic divide between the heavily cratered southern highlands and the relatively smooth plains of the northern lowlands. The origin of the dichotomy, which may have set the course for most of the subsequent geologic evolution of Mars, remains unclear. Internally driven models for the dichotomy form the lowlands by mantle convection, plate tectonics, or early mantle overturn. Externally driven models invoke one giant impact or multiple impacts. Areal densities of buried basins, expressed by quasi-circular depressions and subsurface echoes in radar sounding data, suggest that the dichotomy formed early in the geologic evolution of Mars. Tectonic features along the dichotomy boundary suggest late-stage modification by flexure or relaxation of the highlands after volcanic resurfacing of the northern lowlands. Subsequent deposition and erosion by fluvial, aeolian, and glacial processes shaped the present-day dichotomy boundary.

INTRODUCTION

At first glance, it is easy to conclude that the landscape of Mars is dominated by the giant volcanoes and vast canyons of the Tharsis volcanic and tectonic province. Many of the landforms there, such as Olympus Mons and Valles Marineris, are on a scale that equals or surpasses their largest counterparts on the other terrestrial planets. A closer look, however, reveals a hemispheric-scale crustal feature—a true divide between the southern and northern hemisphere expressed in many aspects of the geology and geophysics of Mars. The most obvious expression of this hemispheric dichotomy is in the exposed cratering record and in the topography (**Figure 1, 2a**). The north-south asymmetry, as described by Carr (1981), was clear from the first global image mosaics of Mars returned in the 1970s by *Mariner 9* (Mutch et al. 1976) and the Viking Orbiters. The full extent of the contrast between the northern and southern hemispheres has only recently been revealed by the unprecedented robotic exploration effort focused on Mars from orbit and the surface. The hemispheric dichotomy stands out as the most fundamental crustal feature of Mars, manifested in the gravitational and magnetic fields and in the planet's crustal structure. The formation of the dichotomy may have set the stage for the growth of the Tharsis volcanic and tectonic province and most of the subsequent geologic evolution of Mars.

The origin and evolution of the hemispheric dichotomy are still poorly understood. From the initial characterizations of the dichotomy using the first global mosaics, two opposing classes of models for its origin emerged. Externally driven or exogenic models form the northern lowlands by either one giant impact (Wilhelms & Squyres 1984) or multiple impacts (Frey & Schultz 1988). Internally driven or endogenic models form the northern lowlands through removal of the basal lowlands crust by mantle convection (Lingenfelter & Schubert 1973, Wise et al. 1979, McGill & Dimitriou 1990) or the generation of thinner crust by plate tectonics (Sleep 1994). These post-Viking-era views were not only at odds over the mechanisms but also the age of formation of the hemispheric dichotomy. Giant impact models dictate that the hemispheric dichotomy is very old (Early Noachian), forming well before the end of the period of heavy bombardment. In endogenic models, the formation of the dichotomy could have been as late as the Early Hesperian (McGill & Dimitriou 1990, Sleep 1994), the age of fractures and fault-controlled fretted valleys in lowlands (McGill & Dimitriou 1990), and lobate scarp thrust faults in the highlands (Watters & Robinson 1999) along the dichotomy boundary in the eastern hemisphere.

We describe the emerging view of the hemispheric dichotomy from data returned by operational spacecraft in orbit and on the surface of Mars. The topographic, geologic, and geophysical manifestations of the hemispheric dichotomy are discussed. Models for the origin of the hemispheric dichotomy are described in the context of recent data constraining the age of the dichotomy. The possible role of lithosphere flexure and relaxation in shaping the dichotomy boundary is summarized. Finally, late-stage modification of the dichotomy boundary by aeolian and glacial processes is described.

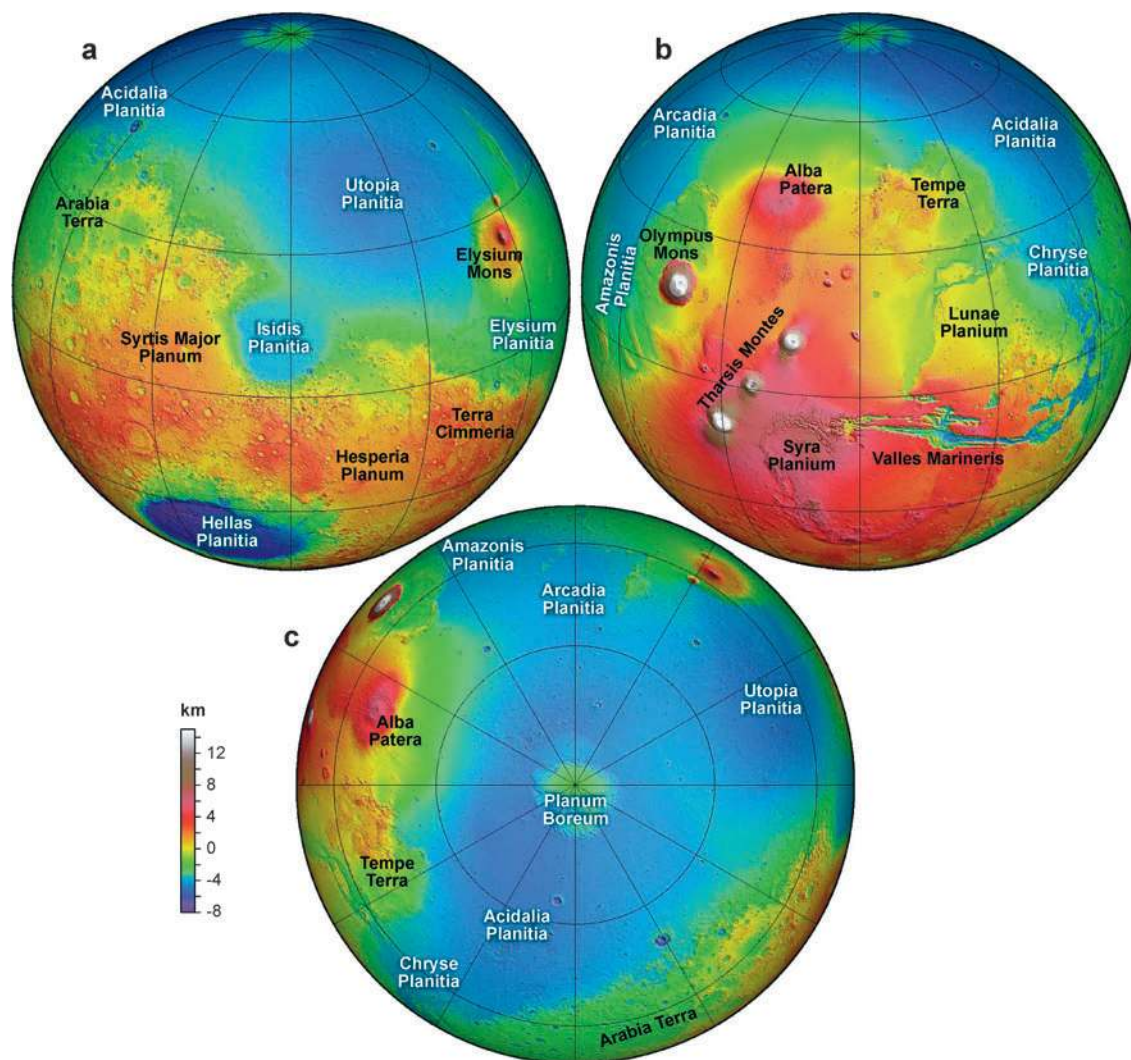


Figure 1

Topographic maps of Mars. (a) The eastern hemisphere centered at 20°N, 90°E; (b) the western hemisphere centered at 20°N, 270°E; and (c) the northern hemisphere centered at 90°N. The topographic maps were derived from MOLA gridded data (Smith et al. 2001). The Martian crustal dichotomy is expressed by the change in elevation between the highlands of the southern hemisphere and the lowlands of the northern hemisphere.

CHARACTERISTICS, AGE, AND ORIGIN OF THE DICHOTOMY

An expression of the hemispheric dichotomy is found in all the principal aspects of the geology and geophysics of Mars. The topography shows an elevation difference

between the highlands and lowlands that is comparable to that between Earth's continents and ocean floors. The dichotomy boundary indicated by the change in crustal thickness, however, is not spatially correlated everywhere with the topographic boundary of the dichotomy. The models for the origin of the hemispheric dichotomy are constrained by the age of the lowlands crust. Evidence of impact basins buried beneath the smooth plains in the lowlands is the key to determining the age of the lowlands crust and when the dichotomy formed.

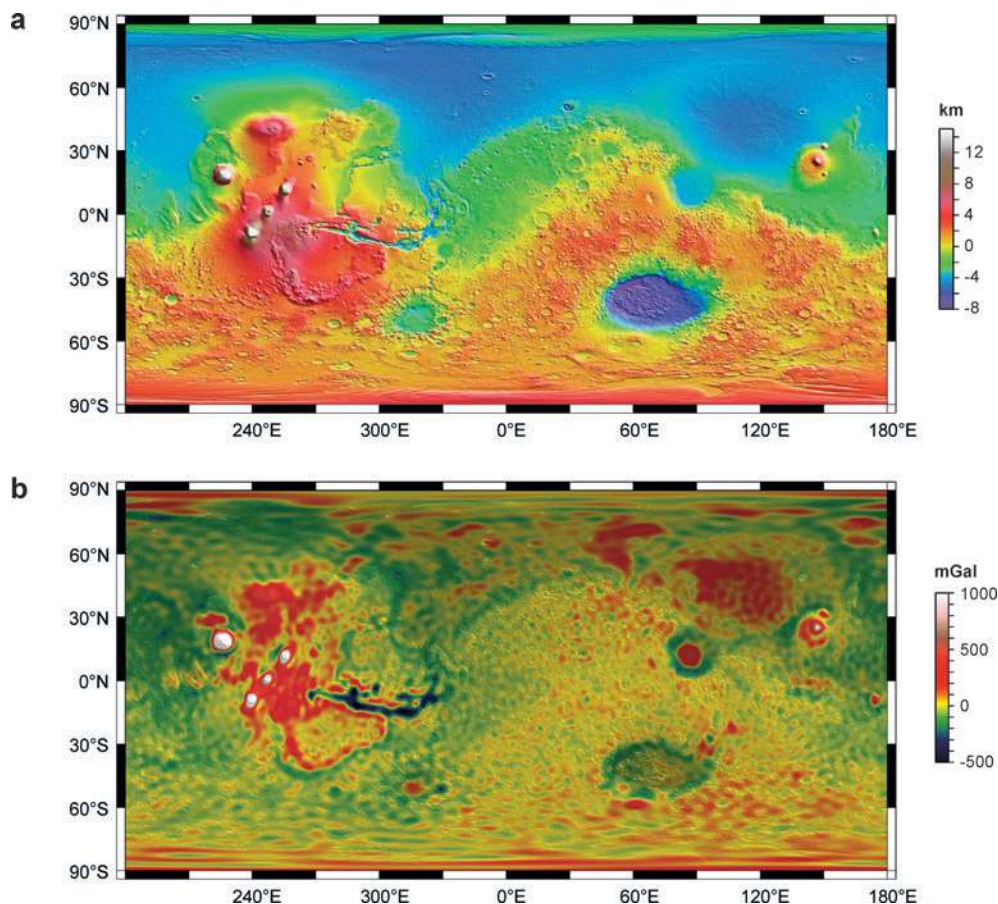


Figure 2

Topography, free-air gravity, Bouguer gravity, crustal thickness, and magnetic field maps of Mars. (a) The topographic map is from MOLA gridded data (Smith et al. 2001). (b) Free-air gravity model JGM95J evaluated to degree and order 95. (c) Bouguer anomaly map with degree 1 and degree 2 terms omitted from Neumann et al. (2004). (d) Degree 1 to 85 crustal thickness model from Neumann et al. (2004). The black line is the 40 km contour. (e) Radial magnetic field altitude-normalized at 200 km from Purucker et al. (2000). Note that the color scale is nonlinear.

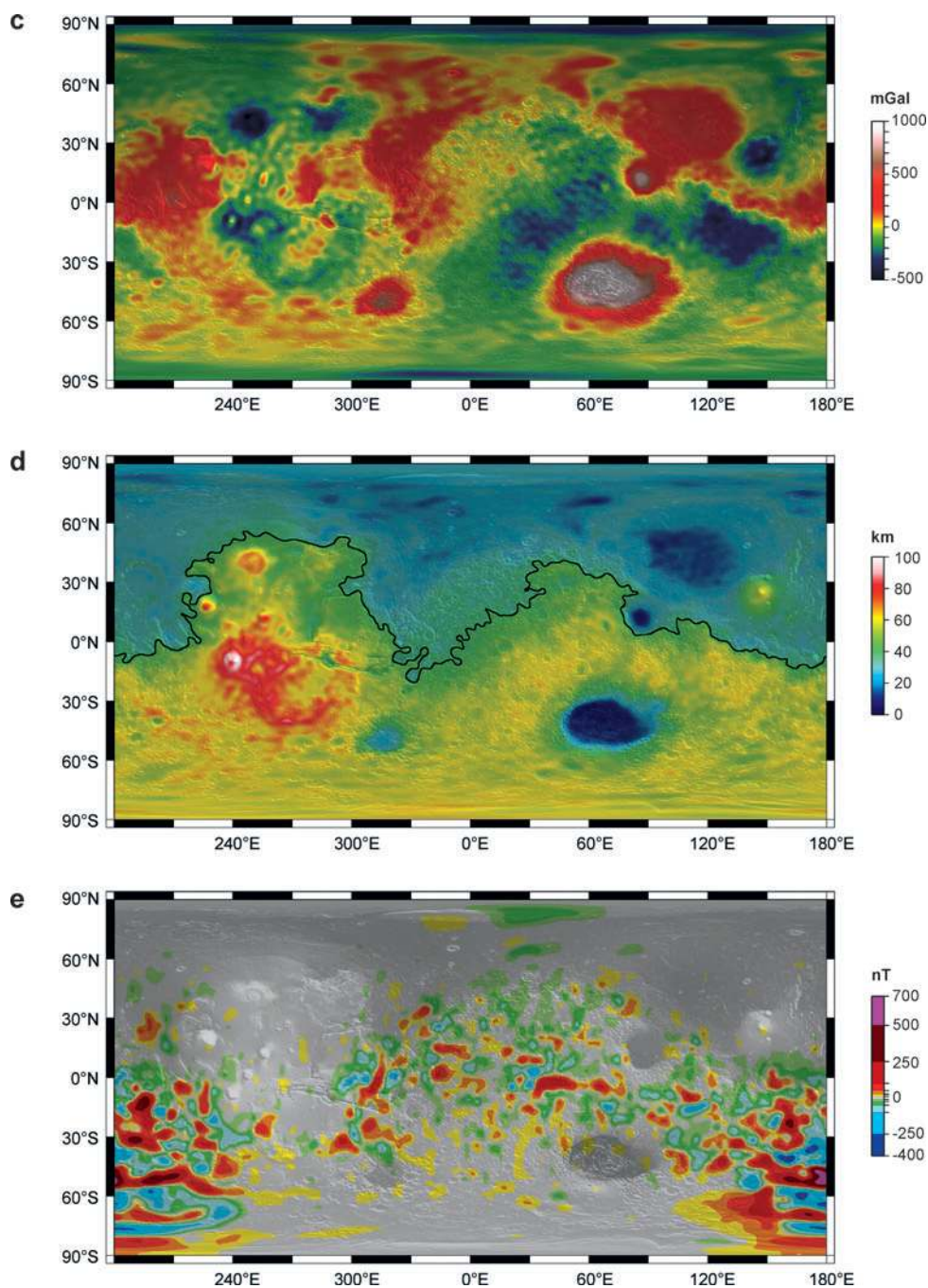


Figure 2

(Continued)

MOLA: Mars Orbiter Laser Altimeter

MGS: Mars Global Surveyor

DEM: digital elevation model

Topographic Expression of the Dichotomy

The topography of Mars has been characterized with great accuracy by the Mars Orbiter Laser Altimeter (MOLA) on the Mars Global Surveyor (MGS) orbiter (**Figure 1, 2a**) (Smith et al. 1998, 1999, 2001; Zuber et al. 2000). MOLA data have a maximum vertical range resolution of approximately 38 cm, a footprint of 168 m, and along-track shot spacing of 300 m (Smith et al. 1998, 2001). Although the cross-track resolution is variable, gridded and interpolated global digital elevation models (DEMs) with spatial resolutions of ~ 460 m/pixel and better can be generated (see Neumann et al. 2001).

The MOLA topography indicates that the relief between the southern highlands and the northern lowlands is everywhere >2.5 km and reaches a maximum >6 km in some areas (**Figure 1, 2a**) (Frey et al. 1998; Smith et al. 1998, 1999, 2001; Zuber et al. 2000). The hypsometry (the frequency distribution of elevations) of Mars is truly bimodal, with the lowlands peak ~ 5.5 km lower than the highlands peak and a greater range of elevation than Earth (**Figure 3a**) (Aharonson et al. 2001). The topographic expression of the dichotomy boundary (the transition between the highlands and lowlands) is highly variable. Throughout half of the northern hemisphere, between $\sim 40^\circ\text{E}$ to 220°E , the dichotomy boundary is marked by a dramatic elevation change and punctuated by a relatively steep scarp (**Figure 1, 2a**). In northeastern Arabia Terra, the maximum relief at the dichotomy boundary is ~ 2.5 km, and in the northern Terra Cimmeria region, the boundary has a maximum relief of ~ 3.5 km (Watters 2003a,b). In much of the eastern hemisphere, the highlands at the dichotomy boundary generally slope toward the northern lowlands; however, there are areas where this trend is reversed (see Frey et al. 1998). Some portion of the relief of the dichotomy boundary in the eastern hemisphere may be due to the impact basins in the highlands, lowlands, and on the dichotomy boundary. The Hellas basin has a prominent annulus of ejecta material that contributes up to 2 km of relief in the highlands (Zuber et al. 2000, Zuber 2001). The Utopia basin in the northern lowlands is comparable in size to Hellas, and although it is significantly shallower in depth than Hellas, Utopia contributes to the maximum relief between the highlands and lowlands. The rims of both basins, however, do not coincide with the dichotomy boundary. Only the Isidis basin is superimposed directly on the dichotomy boundary, its rim walls and floor locally defining the boundary. Modeling suggests that lithospheric loading by the Tharsis rise will result in a topographic high in Arabia Terra antipodal to Tharsis (Phillips et al. 2001). This Arabia bulge may also contribute to the relief of the dichotomy boundary in the eastern hemisphere. Tharsis loading may have also resulted in some uplift in the northern lowlands, possibly reflected by the broad western rim of the Utopia basin (Phillips et al. 2001). From $\sim 330^\circ\text{E}$ to 50°E , the dichotomy boundary is less pronounced in the topography (**Figure 2a**). The maximum relief between the highlands of western Arabia Terra and the lowlands of Acidalia Planitia is much less than that of northeastern Arabia. The maximum relief is generally <1 km at the dichotomy boundary. Contributing to the lower relief at the dichotomy boundary is that much of the highlands of western Arabia are lower than the adjacent highlands of Terra Sabaea and Noachis Terra. Western Arabia could be described as a broad

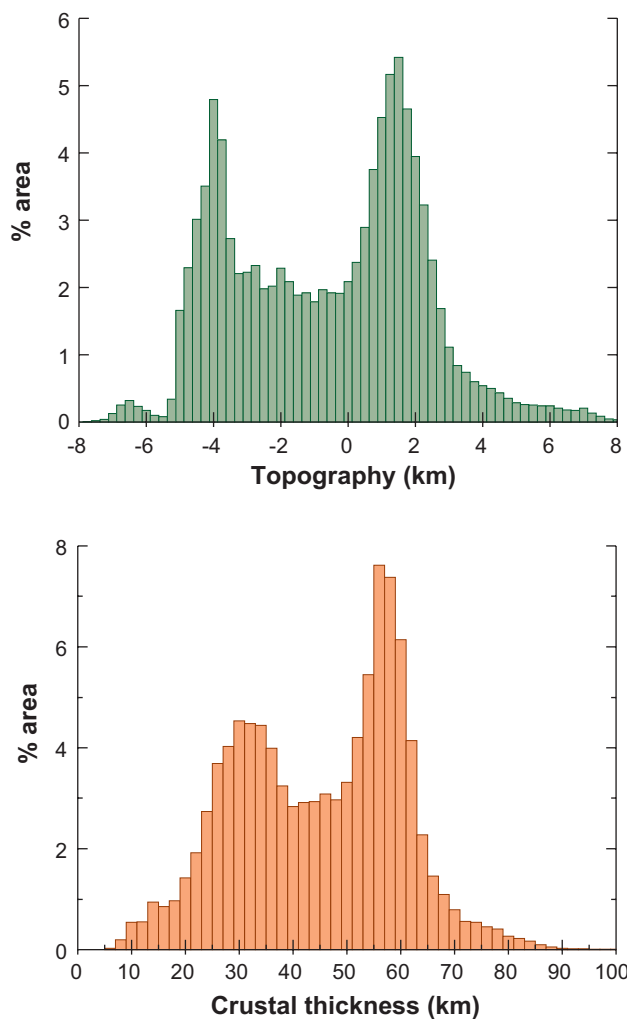


Figure 3

Histograms with respect to percentage of surface area for the topography (hypsoetry) and crustal thickness for Mars, derived from equal area projections of MOLA topography (Smith et al. 2001) and crustal thickness (Neumann et al. 2004). (a) Hypsoetry of Mars, in 0.25 km bins. (b) Histogram of crustal thickness, in 2 km bins.

bench at intermediate elevation between the highlands of Terra Sabaea and Noachis Terra and the lowlands of Acidalia Planitia (**Figure 1, 2a**). Between $\sim 220^\circ\text{E}$ to 330°E , the topography of the dichotomy boundary is dominated by Tharsis. The boundary in eastern Tharsis is formed by the highlands and highland plains of Xanthe Terra, Lunae Planum, and Tempe Terra. The topographic boundary of the dichotomy is defined in northern and western Tharsis by the volcanic edifices and plains of Alba Patera and Olympus Mons and the Tharsis Montes (**Figure 1, 2a**). The location of the dichotomy boundary beneath Tharsis as defined by the change in crustal thickness is unknown (Zuber et al. 2000). As discussed below, the physiographic expression of the dichotomy boundary seen in imaging does not correspond everywhere to the most abrupt transition in crustal thickness between the highlands and lowlands (Zuber et al. 2000, Neumann et al. 2004).

VBF: Vastitas Borealis Formation

OMEGA: Observatoire pour la Minéralogie, l'Eau, les Glaces, et l'Activité

MOC: Mars Orbiter Camera

THEMIS: Thermal Emission Imaging System

Geology of the Northern Lowlands

The geologic contrast between the southern highlands and the northern lowlands is strongly correlated with the elevation difference between the two provinces. The Noachian cratered highlands are the oldest exposed terrain on Mars, recording the period of heavy bombardment (Scott & Carr 1978, Scott & Tanaka 1986, Greeley & Guest 1987, Tanaka & Scott 1987). The smooth plains of the northern lowlands mostly consist of Amazonian age sedimentary surficial deposits of the Vastitas Borealis Formation (VBF) (Tanaka et al. 2003, 2005). The VBF covers Vastitas Borealis as well as Utopia, Acidalia, and northern Arcadia Planitiae, and it is interpreted to be sediments deposited by Late Hesperian outflow channels (Tanaka et al. 2005). Notable exceptions are areas of Amazonis and Utopia Planitiae that have been resurfaced by relatively young lava flows from the Elysium and Tharsis volcanic centers (Fuller & Head 2002). The circum-Chryse outflow channel system deeply incised the dichotomy boundary on the southwestern edge of the Arabia bench (Margaritifer Terra) and to the east of Tharsis (Xanthe Terra and Lunae Planum) (Tanaka et al. 2003, 2005; Rodriguez et al. 2006). The VBF plains are remarkably flat and smooth at scales down to ~100 m (Kreslavsky & Head 1999, Smith et al. 2001), and MOLA pulse width data indicate the lowland plains have a nearly uniform RMS vertical roughness of ~1 m (Garvin et al. 1999). The large number of subtly expressed, sub-parallel wrinkle ridges in the northern lowlands suggests that the VBF is underlain by ridged plains volcanic materials that appear to be Early Hesperian in age (Withers & Neumann 2001, Head et al. 2002). It is estimated that a minimum of ~100 m of VBF sediments cover an accumulation of ridged plains volcanic material in the northern lowlands up to several kilometers thick (Head et al. 2002). A volcanic origin for the ridged plains material is supported by recent data from the Observatoire pour la Minéralogie, l'Eau, les Glaces, et l'Activité (OMEGA) instrument on the European Space Agency's *Mars Express* orbiter that shows evidence of olivine-rich deposits over the floors and ejecta of some fresh impact craters in the northern lowlands (Mustard et al. 2005). This observation suggests that the same Late Noachian to Early Hesperian volcanic event that resurfaced large areas within the highlands also resurfaced the northern lowlands crust (Head et al. 2002).

The transition zone between the highlands and lowlands is geologically complex in some areas, particularly in the eastern hemisphere where the dichotomy boundary is often flanked by knobby and fretted terrain (Sharp 1973). Consisting of knobs, mesas, and fault-controlled fretted valleys, fretted terrain is generally interpreted to be remnants of highland materials resulting from widespread erosion, mass wasting, and retreat of the dichotomy boundary scarp (Scott et al. 1978; Greeley & Guest 1987; Tanaka et al. 2003, 2005). Mars Orbiter Camera (MOC) and Thermal Emission Imaging System (THEMIS) images (on the MGS and Mars Odyssey orbiters, respectively), however, show that fretted terrain in Aeolis Mensae is comprised of friable massive and layered deposits with boulder-free talus (Irwin et al. 2004). These data suggest that fretted terrain may have formed in aeolian sediments deposited along the dichotomy boundary in the Late Noachian–Early Hesperian when fluvial activity was declining (Irwin et al. 2004). In other areas, the transition zone is marked

by relatively young volcanic deposits. The Early to Late Amazonian Medusae Fossae Formation includes large, rolling, discontinuous mounds of loess or volcanic ash in the lowlands along the dichotomy boundary in Amazonis and Elysium Planitia (Greeley & Guest 1987, Tanaka et al. 2005). This material has been deeply eroded by the wind, leaving extensive fields of yardangs (Ward 1979).

Difference in Crustal Thickness

The correlation between the topography and geology of the hemispheric dichotomy is reflected in the impact crater record. The population of impact craters declines abruptly along a northward transect of the crustal dichotomy boundary owing to Hesperian resurfacing of the lowlands. Is this surficial expression of the dichotomy reflected in the internal structure of Mars? This question can be addressed by examining the variation in crustal thickness. Subsurface mass variations have been estimated by subtracting the free-air gravity [obtained from Doppler tracking of the MGS spacecraft (Albee et al. 2001)] from the MOLA surface topography (Bouguer gravity), assuming uniform crustal density for broad regions of the crust (**Figure 2b,c**) (Zuber et al. 2000, Neumann et al. 2004). The resulting Bouguer gravity is inverted by downward-continuation to a crust-mantle interface, and the crustal thickness is the difference between the surface and mantle relief (Zuber et al. 2000, Neumann et al. 2004). It is important to note that this method requires an independent estimate of the mean crustal thickness. Like the hypsometry of the topography, the distribution of the crustal thickness is distinctly bimodal, with peaks at ~ 32 km and ~ 58 km (**Figure 3b**) (Neumann et al. 2004). The hemispheric dichotomy is thus expressed by a degree-1 harmonic structure in the crust (Neumann et al. 2004). The transition in crustal thickness is marked by the local minimum of 40 km in the distribution (**Figure 2d, 3b**).

From $\sim 50^\circ\text{E}$ to 220°E , the 40-km contour in crustal thickness coincides with the topographic and geologic manifestation of the dichotomy boundary. This is not the case, however, in western Arabia Terra ($\sim 330^\circ\text{E}$ to 50°E). Here, the 40-km contour in crustal thickness is several thousand kilometers south of the dichotomy boundary defined by the geology and topography. Much of the heavily cratered Arabia Terra bench is in the thin crustal zone of the northern lowlands (**Figure 2d**); thus, Arabia Terra has a crustal thickness that is more consistent with the northern lowlands than the highlands (Neumann et al. 2004). This suggests that the Arabia bench may be exposed northern lowlands basement (Zuber et al. 2000), lowlands crust that survived resurfacing by volcanic plains and sediments.

The crustal thickness of much of the Tharsis region is not representative of the northern lowlands (Neumann et al. 2004). However, there are zones of relatively thin crust in Tharsis (**Figure 2d**). The most notable zone is associated with the Echus Chasma-Kasei Valles outflow channel system.

Age of the Dichotomy

In the Viking era of Mars exploration, very little was known about the nature of the crust below the smooth plains of the northern lowlands. Viking Orbiter imaging of

QCD: quasi-circular depression

the lowland plains provided few clues as to whether the underlying crust was similar in age to the southern highlands crust (i.e., Early Noachian) or if the lowlands crust was the same age as the Hesperian to Amazonian surficial units (Scott & Carr 1978, Scott & Tanaka 1986, Greeley & Guest 1987, Tanaka & Scott 1987). One clue that the northern lowlands crust beneath the plains material might be ancient was from the absence of partially buried craters and knobs visible in Viking Orbiter images in the circular area of Utopia Planitia. This was interpreted to be evidence of a 3300-km-diameter buried impact basin (McGill 1989), requiring considerable time between its formation and its Hesperian surface age to achieve its present, highly subdued profile.

The first window into the age of the northern lowlands crust was provided by the MOLA topography. The topographic data unveiled a large, circular depression in Utopia Planitia. Its coincidence with a large positive free-air gravity anomaly confirmed the existence of the Utopia basin (Smith et al. 1999, Zuber et al. 2000). A buried basin of the scale of Utopia requires the northern lowlands basement crust to be ancient because impactors of the necessary size would not have escaped early accretion (Zuber 2001). Also, subtly expressed in the MOLA topography is a large number of quasi-circular depressions (QCDs), interpreted to be evidence of a population of buried impact craters and basins in the northern lowlands (Frey et al. 2002). The cumulative frequency curves (areal density as a function of diameter) of the QCD population have the same distribution as the exposed crater population. The areal density of the QCDs indicates that the number of impact craters in the northern lowlands is comparable to the number of exposed impact craters in the southern highlands (**Figure 4**) (Frey 2006a,b). The QCD-based N(200) crater density (a measure of the cumulative number of basins with diameters >200 km per million square

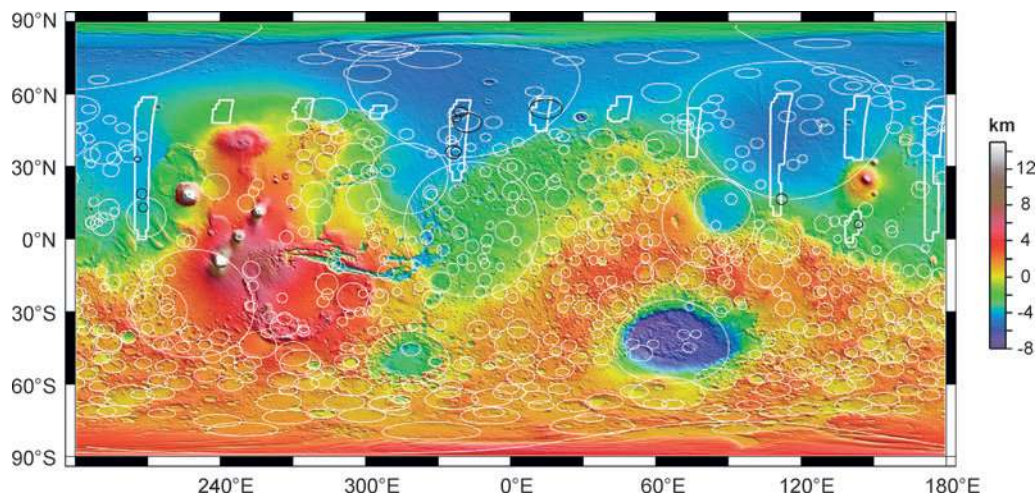


Figure 4

Locations of MARSIS buried basins and QCDs. The locations and inferred diameters from MARSIS echoes are shown in black on MOLA color-coded shaded relief. The locations of QCDs with diameters greater than 200 km mapped by Frey (2006a) are plotted in white.

kilometers) for the northern lowlands crust is 2.5 (Frey 2006a,b). QCDs, however, have also been found in the highlands. The ratio of buried basins to exposed basins with diameters >200 km is 6 to 1; in the lowlands, the ratio is 20 to 1 (Frey 2006a). The $N(200)$ crater density of the ancient highlands (exposed basins plus QCDs) is estimated to be ~ 4.5 , with an upper limit of ~ 8.5 based on extrapolation of the largest highlands basins (exposed basins and QCDs) (Frey 2006a). Thus, although the areal density of QCDs in the northern lowlands is comparable to that of the exposed southern highlands, it is lower than the total highlands that include highland QCDs. This suggests that the age of the northern lowlands crust is younger than the highlands crust (Frey 2006a,b). However, because of the likely possibility that some buried basins in the highlands and the lowlands are not expressed by QCDs, the QCD-based $N(200)$ crater densities are lower limits.

The second window into the age of the northern lowlands crust is from radar sounder data. The Mars Advanced Radar for Subsurface and Ionospheric Sounding (MARSIS) instrument onboard the European Space Agency's *Mars Express* spacecraft has returned the first data on the unexposed subsurface (Picardi et al. 2005). MARSIS is a multifrequency synthetic aperture orbital sounding radar consisting of a 40-m dipole antenna that operates in four frequency bands between 1.3 and 5.5 MHz in its subsurface modes. It has a spatial resolution of 10 to 30 km cross-track and 5 to 10 km along-track, with a free-space range resolution of approximately 150 m. The smooth, flat northern lowland plains provide ideal operating conditions for the MARSIS sounder because the low vertical roughness results in minimum surface clutter that can be confused with echoes from subsurface features (Picardi et al. 2005, Watters et al. 2006). Radargrams, time-delay renderings of the sounding data, for a number of MARSIS orbits over the northern lowlands have parabolic-shaped echoes. Simulations and MARSIS data over exposed impact basins show that parabolic-shaped echoes result from surface echoes from basin walls with slope-normals in the direction of the spacecraft. The parabolic echoes with no source in the surface as determined from clutter models generated using MOLA topography are interpreted to be from off-nadir and near-surface features (Picardi et al. 2005, Watters et al. 2006). Radargrams projected to ground-range geometry show that parabolic echoes project into circular arcs with constant curvature on the surface, and these arcs are used to infer the probable location and size of the basin. The left/right ambiguity in the MARSIS data (the instrument receives surface and subsurface radar backscatter from the right and left of the ground track simultaneously), however, does not allow a unique determination of the location of the center of the basin unless the same rim wall is detected in multiple MARSIS orbits.

To date, the MARSIS data have provided evidence of 11 buried basins in the northern lowlands, ranging in diameter from approximately 130 to 470 km in Chryse, Acidalia, Amazonis, Elysium, and Utopia Planitia (Figure 4) (Watters et al. 2006). An independent crater age of the northern lowlands can be determined from the MARSIS basins. The $N(200)$ crater density for the area of the northern lowlands surveyed is ~ 1.9 (range ~ 1.6 to 2.6) (Watters et al. 2006). This suggests an Early Noachian age for the northern lowlands crust. A plot of the cumulative frequency of the MARSIS basins shows that most of the basins lie above the Early Noachian

MARSIS: Mars Advanced Radar for Subsurface and Ionospheric Sounding

SNC: Shergottites,
Nakhlites, and Chassigny

boundary determined by Tanaka (1986) based on N(16) crater densities extrapolated to larger diameters with a -2 power law, indicating an Early Noachian age for the northern lowlands crust. The MARSIS-based crater density for the lowlands crust is a lower limit because it is likely that not all the buried basins in the area surveyed have been detected (see Watters et al. 2006).

A comparison of the MARSIS basins detected thus far with the locations of mapped QCDs with diameters greater than 200 km shows that some QCDs coincide with MARSIS basins (**Figure 4**). The agreement supports the interpretation that QCDs are the surface expression of buried basins. However, the small number of QCDs that correlate with subsurface echoes may indicate that not all QCDs are related to buried basins (Watters et al. 2006). In spite of this, QCD- and MARSIS-based crater density ages both suggest that the northern lowlands crust is Early Noachian in age. Thus, the consensus from the two data sets is that the dichotomy formed early in Martian geologic history, before the formation of the Utopia basin. What is not well known is the time interval between the formation of the Martian crust and the formation of the dichotomy. As discussed above, the crater density of the highlands, which includes exposed basins and QCDs, suggests that the highlands crust is somewhat older than the lowlands crust (Frey 2006a,b). Based on a model of absolute age that includes proposed lowland-making basins, Frey (2006a) suggests that the age difference between the highlands crust and lowlands crust may be as little as ~ 100 Myr. This age difference, however, is not well constrained because only a few mapped QCDs have been independently verified with MARSIS data (Watters et al. 2006) or with other geophysical data. The age difference between the highlands and lowlands crusts can also be evaluated by comparing the MARSIS-based crater density for the lowlands basement to the crater density of the oldest exposed highlands basement. In the Tharsis region, near Claritas Fossae, an area of well-preserved highlands basement is exposed ($\sim 28^\circ\text{S}$, 259°E). Tanaka (1986) determined the N(16) crater density of this area of highlands to be 294 ± 81 . Based on extrapolation to larger diameters with a -2 power law, this corresponds to an N(200) density of $\sim 1.9 \pm 0.5$. Thus, it can be concluded from the MARSIS data that the lowlands crust is at least as old as the oldest exposed highlands crust (Watters et al. 2006).

Origin of the Crustal Dichotomy

The age of the northern lowlands basement crust is a critical constraint on the origin of the crustal dichotomy. The population of impact basins in the northern lowlands, buried by volcanic plains and sediments, indicates that the basement crust is old, perhaps as old as the oldest exposed highlands crust (see above). Planetary accretion models and isotope data from SNC (Shergottites, Nakhlites, and Chassigny) meteorites suggest that the crust of Mars formed very early in the geologic history of the planet (see Halliday et al. 2001). Isotope systematic relations for Pb, Sr, and Os in the SNCs are consistent with large-scale silicate fractionation ~ 50 Myr after the formation of the Solar System with no subsequent appreciable remixing of the mantle or crust (see Solomon et al. 2005). The early, rapid formation of the Martian crust may have been accomplished by differentiation of a global silicate magma ocean

driven by accretional heating (Elkins-Tanton et al. 2003, 2005, Solomon et al. 2005). Early, rapid formation of the Martian crust may present a challenge to endogenic mechanisms for the origin of the dichotomy.

The earliest endogenic models for the dichotomy involved a primordial difference in crustal thickness (Mutch et al. 1976) or thinning of the northern lowlands crust by mantle convection (Lingenfelter & Schubert 1973, Wise et al. 1979). Mutch et al. (1976) postulated that to maintain isostatic equilibrium, a thin northern lowlands crust and a thick highlands crust would result in a topographically low northern hemisphere and topographically high southern hemisphere. Wise et al. (1979) suggested that early convective upwelling resulted in subcrustal erosion and fracturing of the northern lowlands crust. The crust foundered, was resurfaced by volcanic materials, and sank to maintain isostatic equilibrium. They suggested that the crustal material was transported to the southern hemisphere and initiated the formation of the Tharsis rise. McGill & Dimitriou (1990) proposed a similar mechanism, suggesting that the early northern crust was thinned and subsided by mantle convection through crustal delamination. They argued that the eroded crustal material was globally dispersed rather than concentrated beneath Tharsis. However, McGill & Dimitriou (1990) cited Late Noachian to Early Hesperian fracturing and faulting in the northern lowlands and along the dichotomy as evidence of a younger age of formation. A Late Noachian to Early Hesperian age for the dichotomy is ruled out by the apparent Early Noachian age of the northern lowlands crust (see above).

Recent numerical modeling of early degree-1 mantle convection on Mars assumes a mantle layered in viscosity (Zhong & Zuber 2001, Zuber 2001, Zhong et al. 2004). For a viscosity contrast between the upper and lower mantle of ≥ 100 , the predicted convection pattern results in preferential heating, upwelling, and crustal thinning in one hemisphere and downwelling and crustal thickening in the other hemisphere (Zhong & Zuber 2001, Zuber 2001, Roberts & Zhong 2006). The modeling indicates that degree-1 mantle convection develops in timescales ranging from several hundred million years (Zhong & Zuber 2001, Zuber 2001) to on the order of 100 Myr (Roberts & Zhong 2006), depending on mantle viscosity. The smaller the viscosity, the faster the degree-1 mantle convection develops. For reasonable mantle viscosity for early Mars, degree-1 mantle convection can ensue in less than 100 Myr (Roberts & Zhong 2006). Widespread crustal extension (McGill & Dimitriou 1990) and volcanism (Zhong & Zuber 2001, Zuber 2001) in the northern lowlands likely accompanied the formation of the crustal dichotomy. A somewhat related mechanism to mantle convection for the formation of the dichotomy involves an early transient superplume (Ke & Solomatov 2006). Modeling suggests that the high thermal gradient at the core and the mantle expected at the end of planetary accretion results in a low viscosity layer at the base of the mantle that gives rise to a single large-scale mantle plume. A superplume may have resulted in global-scale melting that contributed to the formation of the crustal dichotomy (Ke & Solomatov 2006).

An alternative endogenic process proposed to account for the hemispheric dichotomy is the same mechanism responsible for the contrast in crustal thickness between continental and oceanic crust on Earth—plate tectonics. Sleep (1994) proposed

the northern crust was removed by late Noachian hemispheric subduction, the younger lowlands crust resulted from seafloor spreading, and the present dichotomy boundary marks relic plate margins. In this view, the dichotomy boundary extending from the Tyrrenia region, through Amenthes and northern Terra Cimmeria to Memnonia, is a passive margin. The proposed Boreal-Austral plate would have been subducted along the region extending from Arabia Terra through Tempe Terra to the northern part of the Tharsis Montes volcanic line and eventually would have extended as far south as Daedalia (Sleep 1994). Plate tectonics would have ended when melting at spreading centers ceased or when the lithosphere became too buoyant. This is an intriguing model because the topography of the dichotomy boundary in much of the eastern hemisphere is similar to that of terrestrial passive margins (Frey et al. 1998). A Late Noachian formation of the dichotomy, however, is inconsistent with the apparent age of the northern lowlands crust (see above). It is possible that a phase of plate tectonics occurred much earlier in the Early Noachian (Lenardic et al. 2004). Plate recycling may have occurred when the southern highlands crust was forming. Modeling by Lenardic et al. (2004) suggests that plate tectonics ceased when the surface areas of positively buoyant lithosphere grew to $\sim 50\%$ of the planet's surface. Other factors, however, argue against plate tectonics on Mars. First, there is no topographic evidence of a relic subduction zone anywhere along the dichotomy boundary. In Arabia Terra, there is no geologic evidence of subduction (McGill 2000), and in the northern lowlands there are Noachian outcrops and conflicting structural patterns (Pruis & Tanaka 1995).

Perhaps the most fundamental challenge for subcrustal transport from mantle convection and plate recycling models is the timescales on which these mechanisms operate (Solomon et al. 2004, 2005). If the entire crust of Mars formed by 4.5 Ga and plate recycling and subcrustal transport require timescales on the order of several hundred million years, these processes are too slow to form the crustal dichotomy (Solomon et al. 2004, 2005). The timing constraint can be satisfied if the endogenic process was able to produce a significant difference in crustal thickness on a timescales of less than 100 Myr, as may be the case in rapidly developing mantle convection (Roberts & Zhong 2006), or if the process is related to the formation of the earliest crust. Spatially heterogeneous fractionation of a magma ocean could account for an early-rapid formation of the crustal dichotomy (Solomon et al. 2004, 2005). It is not clear, however, why heterogeneous fractionation would occur on a hemispheric scale. Another event that is a direct consequence of the formation of a magma ocean could account for the hemispheric dichotomy. Crystallization of an early magma ocean may have resulted in a gravitationally unstable mantle that rapidly overturned (Elkins-Tanton et al. 2003, 2005a). Fractional crystallization results in a gravitationally unstable mantle because late-stage cumulates formed near the base of the magma ocean may be denser than the upper mantle. This density inversion drives a Rayleigh-Taylor instability and mantle overturn. Overturn could have occurred as rapidly as ≤ 1 to 10 Myr (Elkins-Tanton et al. 2005b). The overturn may result in degree-1 mantle flow that triggered the formation of the crustal dichotomy by thinning the crust over convective upwelling in the lowlands and thickened the crust over downwelling in the highlands (Solomon et al. 2005). Widespread

early volcanism may have accompanied the formation of the martian crust from a magma ocean process through decompression mantle melting (Elkins-Tanton et al. 2005a). Modeling suggests that basaltic to andesitic magmas could have been emplaced 30 to 50 Myr after planetary formation and after the magma ocean solidified and mantle overturn ceased. The predicted composition of the basalts and andesites is roughly consistent with the broad basalt-like (type 1) composition of the highlands and the andesite-like (type 2) composition of the lowlands determined by MGS Thermal Emission Spectrometer (TES) data (Bandfield et al. 2000). However, it is likely that the TES data has revealed andesite-like weathering products in the lowlands, possibly derived from the highlands (Wyatt et al. 2004, Nimmo & Tanaka 2005).

TES: Thermal Emission Spectrometer

The magma ocean-mantle overturn model for the origin of the dichotomy faces several challenges. First, it is not clear that mantle overturn would result in a degree-1 convective pattern. Second, how is such a primordial, first-order crustal feature preserved over Martian geologic time? If the dichotomy formed very early when the heat flux was high, long wavelength features would be expected to relax over timescales of 100 Myr (Nimmo & Stevenson 2001, Roberts & Zhong 2006). Roberts & Zhong (2006) suggest that long-lived degree-1 mantle convection, not achieved by mantle overturn, is needed to maintain the crustal dichotomy after it has formed. Also, the observed isotopic heterogeneities argue against a well-mixed global magma ocean (see Halliday et al. 2001, Foley et al. 2005). An alternative to the global magma ocean model is that the formation of the dichotomy is related to a local magma ocean (Reese & Solomatov 2006). A local magma ocean is formed by a very large impact event during late-stage accretion that induces regional melting. Modeling suggests that rapid crystallization and isostatic adjustment of a local magma ocean can create a global layer of partially molten silicates that result in a degree-1 asymmetry in crustal thickness (Reese & Solomatov 2006). Although a local magma ocean satisfies the timing constraint and could account for evidence of early and surviving, geochemical distinct mantle reservoirs, the model is highly dependent on the size of the impactor and the impact location.

Impact excavation models for the origin of the dichotomy do not share the timing problem of the endogenic models. Both the giant impact (Wilhelms & Squyres 1984) and multiple impacts (Frey & Schultz 1988) models require that the dichotomy is very old (Early Noachian), forming well before the end of the period of heavy bombardment. In these models, the crust is thinned by ballistic transport of material out of the northern lowlands, and the dichotomy boundary (marked by the relatively steep scarp) is a relic rim of one or more basins (Wilhelms & Squyres 1984, Frey & Schultz 1988). The impact origin for the dichotomy, however, has other challenges. The outline of the northern lowlands, as expressed by the topography or the crustal thickness, is not circular (Zuber et al. 2000, Zuber 2001). There are no other Utopia-scale circular zones of crustal thinning in the northern lowlands that would be expected if multiple impacts formed the dichotomy (Zuber et al. 2000, Zuber 2001, Solomon et al. 2005). Also, there is a low probability of multiple Utopia-scale impacts in the same hemisphere (McGill & Squyres 1991, see Nimmo & Tanaka 2005).

Clues from the Magnetic Field

Another expression of the hemispheric dichotomy is found in the crustal magnetization determined by vector magnetic-field measurements from MGS (Acuna et al. 1999; Connerney et al. 1999, 2001, 2005; Purucker et al. 2000; Arkani-Hamed 2002; Cain et al. 2003). The strong remnant magnetic field preserved in the southern highlands does not extend far beyond the dichotomy boundary into the northern lowlands (**Figure 2e**). There are, however, regions of the northern lowlands with a relatively weak remnant magnetic field (Lillis et al. 2004, Connerney et al. 2005). Some of these weak magnetic features cover broad regions of the northern plains and suggest that the underlying crust is magnetized (Connerney et al. 2005). The large lowlands impact basins Isidis and Utopia and the highlands basins Hellas and Argyre are conspicuous by the absence of any detectable remnant magnetic field (**Figure 2e**) (Acuna et al. 1999; Connerney et al. 1999, 2001, 2005; Purucker et al. 2000; Arkani-Hamed 2002; Cain et al. 2003). Also devoid of magnetic anomalies are Chryse Planitia and the Tharsis and Elysium volcanic centers. The correlation between the large impact basins in both hemispheres and areas devoid of magnetic anomalies suggests that the Martian dynamo had failed prior to the formation of these basins (Acuna et al. 1999), and any preexisting remnant magnetic field would have been lost as a result of shock and heating owing to basin formation (Acuna et al. 1999, Hood et al. 2003). Several mechanisms have been proposed to account for the relatively weak to absent remnant magnetic field in the northern lowlands. A uniformly magnetized northern lowlands crust may have been demagnetized by hydrothermal alteration of magnetic carriers (Solomon et al. 2005). This would require that large volumes of water penetrate the crust to great depths on the order of 25 km. Alternatively, demagnetization of the northern lowlands crust may have been accomplished by multiple, large impacts on the scale of the Utopia basin. However, as discussed in the previous section, the lack of supporting evidence and low probability of multiple Utopia-scale impacts in the same hemisphere argue against this mechanism (see Nimmo & Tanaka 2005). The widespread volcanic resurfacing of the northern lowlands after the dynamo ceased might also account for the weakening and loss of the remnant magnetic field in the northern lowlands (Connerney et al. 2004, 2005). Volcanic resurfacing of the northern lowlands, however, may only make a small contribution to the loss of the remnant field because heat from rapidly cooling lava flows does not result in significant reheating of the crust (Solomon et al. 2005).

The initiation and demise of the early dynamo may be directly related to the formation of the crustal dichotomy. An early episode of plate tectonics could have resulted in a high core-mantle boundary heat flux necessary to drive a core dynamo (Nimmo & Stevenson 2000). If the dichotomy formed as a result of mantle overturn, cold cumulates placed against the core-mantle boundary could create a heat flux that might have maintained a dynamo for ~ 15 to 150 Myr (Elkins-Tanton et al. 2005b). Likewise, an early transient superplume would be expected to result in a high heat flow at the core-mantle boundary (Ke & Solomatov 2006).

THE DICHOTOMY BOUNDARY

The dichotomy boundary is one of the most intriguing physiographic features of Mars. The distinct contrast in the topographic expression of the dichotomy boundary in western Arabia Terra and much of the eastern hemisphere is suggestive of two different evolutionary paths. The present-day dichotomy boundary could be one of the most ancient surviving crustal features on Mars, evolving from an ancient transition zone between the hemispheres. If this is the case, the dichotomy boundary, where not obscured by Tharsis, holds important clues to the mechanisms that formed the hemispheric dichotomy.

Tectonics of the Boundary

Whereas the dichotomy boundary in the eastern hemisphere is marked by the most dramatic change in elevation, it is also marked by tectonic features in the highlands and lowlands. The tectonic features found along the dichotomy boundary express both crustal extension and compression (**Figure 5**) (McGill & Dimitriou 1990; Watters 1993, 2003a,b; Watters & Robinson 1999). Lobate scarps, landforms interpreted to be the surface expression of thrust faults, are found in the highlands of northern Terra Cimmeria, Amenthes, and northern Arabia Terra (Watters 1993, 2003b; Watters & Robinson 1999). They occur ~100–500 km from the dichotomy boundary and roughly parallel the dichotomy boundary scarp. Estimates of the cumulative slip on lobate scarp thrust faults are generally less than 1 km (Watters & Robinson 1999, Watters 2003b). The largest of the lobate scarps along the dichotomy boundary is Amenthes Rupes. With a maximum relief of >1200 m and a length of ~380 km, the estimated maximum cumulative slip or displacement on the underlying thrust fault is on the order of 3 km (Watters 2003b). The compressional

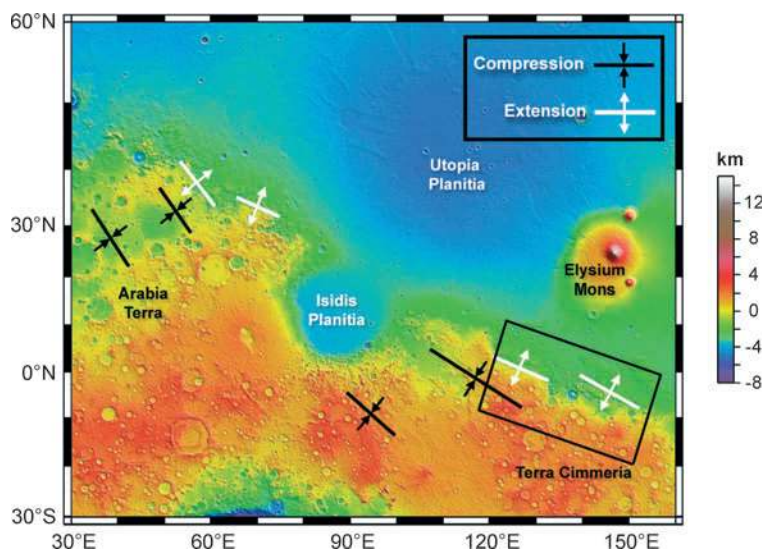


Figure 5

Generalized location of lobate scarp thrust faults and extensional troughs along dichotomy boundary in the eastern hemisphere shown on MOLA color-coded shaded relief. Black box shows the location of the area in northern Terra Cimmeria where the long wavelength topography of the dichotomy boundary is examined (**Figure 6**).

strain in the highlands near the dichotomy boundary, estimated using the maximum displacement to fault length ratio γ of the thrust fault population ($\gamma \cong 6.2 \times 10^{-3}$), is $\sim 0.2\%$ (Watters & Robinson 1999, Watters 2003b).

In the lowlands and along the dichotomy boundary, evidence of fracturing and normal faulting is expressed in fault-controlled fretted valleys and extensional troughs (McGill & Dimitriou 1990, Smrekar et al. 2004). In some locations, these fractures appear to form the dichotomy boundary scarp (McGill & Dimitriou 1990). Evidence also points to buried normal faults in the lowlands along the dichotomy boundary. The disappearance of knobs ~ 300 to 500 km from the dichotomy boundary in northern Arabia Terra (Ismenius area) suggests that faulted highlands crust was down dropped and formed a lowland bench (Smrekar et al. 2004). The extensional strain across the lowland bench is estimated to be $\sim 3.5\%$ (Smrekar et al. 2004) with the regional strain across a ~ 450 km boundary zone of extension on the order of 0.4% .

The proximity and parallel orientation of lobate scarp thrust faults, fractures, and extensional troughs suggest that tectonism was involved in the formation or modification of the dichotomy boundary in the eastern hemisphere. The age of the tectonic events is significant in determining the role of tectonism in the formation of the crustal dichotomy. McGill & Dimitriou (1990) estimated that fracturing took place in the Late Noachian to Early Hesperian from areal crater densities in knobby terrain based on the total crater population and craters with unfractured rims. Smrekar et al. (2004) concluded that extension in the Ismenius area could be no older than Middle Noachian because faults cut Middle Noachian plateau material. The timing of the compressional event that formed the highland lobate scarps has been estimated using crosscutting relationships. The fact that lobate scarps such as Amenthes Rupes crosscut the walls and floor material of heavily degraded Noachian impact craters and valley networks and have very few superimposed craters suggests that the faults are young relative to the Early Noachian highlands material they deform (Watters & Robinson 1999, Watters 2003b). Lobate scarps in northern Terra Cimmeria crosscut Middle Noachian valley network terrain (Wilhelms & Baldwin 1989), suggesting that the faults formed no earlier than the Late Noachian. Maxwell & McGill (1988) suggest that the lobate scarps near the dichotomy boundary in the Amenthes region formed during the Late Noachian. Also, some of the lobate scarps in northern Terra Cimmeria deform intercrater plains estimated to be Early Hesperian in age (Greeley & Guest 1987), suggesting that the formation of the scarps may have continued into the Early Hesperian (Watters & Robinson 1999). The extensional and compressional tectonism along the dichotomy boundary may have been roughly syntectonic, occurring in the highlands and lowlands at a distance of up to approximately ± 500 km from the dichotomy boundary (McGill & Dimitriou 1990; Watters & Robinson 1999; Watters 2003a,b; Smrekar et al. 2004).

The dominant tectonic features in the northern lowlands are wrinkle ridges partially buried by VBF sediments (Withers & Neumann 2001, Head et al. 2002). Like wrinkle ridges in ridged plains material in the highlands (Watters 1991), those in the northern lowlands have subparallel trends and are regularly spaced (Head et al. 2002). Many of the northern lowlands wrinkle ridges appear to be related to the Tharsis-circumferential ridge system, whereas others reflect regional stress patterns

related to the Isidis and Utopia basins. The number of modified and partially buried impact craters in the lowland ridged plains suggests they are Early Hesperian in age (Head et al. 2002). It is important to note that the emplacement and subsequent deformation of the lowland and highland ridged plains occur in the same timeframe as the deformation along the dichotomy boundary (Late Noachian to Early Hesperian) (e.g., Watters 1993, Head et al. 2002).

Long-Wavelength Topography and Flexure and Relaxation

The tectonic features associated with the dichotomy boundary in the eastern hemisphere are generally found where the elevation change from the highlands to lowlands is the greatest (**Figure 1, 2a**) (McGill & Dimitriou 1990; Watters 2003a,b). The long wavelength topography along this segment of the dichotomy boundary is distinctive, generally consisting of a broad rise and an arching ramp that slopes down into the lowlands and often terminates in an escarpment (Watters 2003a, Watters & McGovern 2006). The cross-sectional profile of the broad rise is often asymmetric with gentle slopes on the southern flanks that slope away from the dichotomy boundary. On the northern flanks, the broad rise transitions into a relatively steep ramp that slopes down into the lowlands. The maximum slopes are reached on the ramp and the boundary scarp (**Figure 6**).

Two approaches have been taken to explain the long-wavelength topography of the dichotomy boundary, crustal relaxation (Nimmo 2005, Guest & Smrekar 2005), and lithospheric flexure (Watters 2003a,b; Watters & McGovern 2006). Lithospheric flexure in response to vertical loading results in long-wavelength topography with a distinct deflection profile characterized by a flanking upwarp or bulge (Turcotte 1979, Turcotte & Schubert 2002, Watts 2001). Modeling an elastic plate overlying an incompressible fluid subjected to an end load (Watters 2003a) or a line load (Watters & McGovern 2006), using both a semi-infinite or broken lithosphere and an infinite or continuous lithosphere, shows that the best-fit deflection profiles are for a broken lithosphere with an elastic thickness T_e of ~ 30 to 36 km. This suggests the dichotomy boundary is the result of flexure of the southern highland lithosphere in response

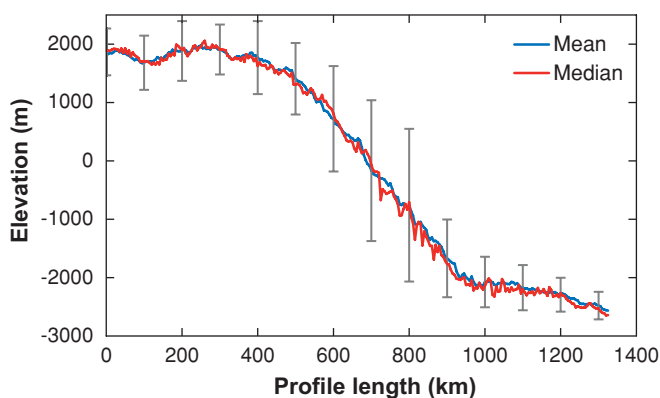


Figure 6

Long wavelength topography of the dichotomy boundary in northern Terra Cimmeria. Topographic profiles are the mean and median of 41 profiles spaced at approximately 50 km intervals covering the area of the boundary shown in **Figure 5**.

The error bars are ± 1 standard deviation. Vertical exaggeration is $\sim 160:1$.

to vertical loading (Watters 2003a, Watters & McGovern 2006). This value of T_e is in general agreement with independently derived estimates using MGS gravity and topography (Zuber et al. 2000; McGovern et al. 2002, 2004) and estimates of the seismogenic thickness from elastic dislocation modeling of the Amenthes Rupes thrust fault (Schultz & Watters 2001), but is near the lower limit of range of T_e (37 to 89 km) obtained by Nimmo (2002).

Extensional bending stresses induced by lithospheric flexure reach a maximum in the lowlands of >400 MPa, several hundred kilometers from the dichotomy boundary scarp for $T_e \sim 30$ km (Watters 2003a). Maximum compressional stresses of ~ 20 MPa occur in the highlands approximately 500 km from the base of the boundary scarp. The location of the predicted stresses agrees with the observed tectonic features along the dichotomy boundary. However, the magnitude of the predicted compressional stresses is low relative to the stress necessary to initiate thrust faulting at even relatively shallow depths (Watters et al. 2000). This suggests that other compressional stresses, possibly related to global contraction, must have contributed to the formation of the lobate scarp thrust faults (Watters 2003a,b; Watters & McGovern 2006).

Alternatively, the long-wavelength topography of the dichotomy boundary may be the result of relaxation of the highlands through lateral flow of the lower crust. Variations in the lateral crustal thickness result in pressure gradients that can drive flow in the lower crust even if the crust is isostatically compensated (Nimmo & Stevenson 2001, Nimmo 2005). Lateral crustal flow results in a reduction in relief of the thick crust and increases the relief of the thin crust. Nimmo (2005) used diffusion-like equations, where the diffusion coefficient depends on the crustal temperature structure and rheology, to describe the change in crustal thickness (see Nimmo & Stevenson 2001) with the temperature structure determined by the decay of radioactive materials within and below the crust. Guest & Smrekar (2005) also model relaxation of preexisting dichotomy boundary topography using semianalytical and finite element modeling. They use finite elements to model a visco-elastic-plastic rheology that accounts for the possible development of plastic zones, and they employ the thermal cooling model of Hauck & Phillips (2002). These relaxation models produce generally consistent results, predicting a similar topographic expression of a relaxed dichotomy boundary and similar zones of extension and compression (Guest & Smrekar 2005, Nimmo 2005). A topographic feature of a relaxed dichotomy boundary predicted by the models is a lowland bench, an elevated bench below the boundary escarpment. In the Ismenius region of northern Arabia Terra, this lowland bench correlates with knobby terrain and topographic feature interpreted to be a down-dropped block of highlands crust (Smrekar et al. 2004). Elsewhere along the dichotomy boundary, the bench corresponds to fretted terrain. However, the knobby and fretted terrain of Aeolis Mensae, in lowlands north of Terra Cimmeria, may have formed in Late Noachian to Early Hesperian aeolian sediments deposited along the dichotomy boundary (Irwin et al. 2004). If this is the case, the topography in the lowlands contributed by knobby and fretted terrain is unrelated to either flexure or relaxation of the dichotomy boundary.

Relaxation models predict four zones of stress, two compressional and two extensional (Guest & Smrekar 2005, Nimmo 2005). Nimmo (2005) predicts peak

compression and extension in the highlands and lowlands that generally agree with the location of the observed tectonic features. The maximum compressional stress predicted by relaxation is higher (60 MPa) than obtained through flexure, but the maximum of extensional stress is considerably lower (80 MPa) (Nimmo 2005). The zones of extension predicted by Guest & Smrekar (2005) agree well with the observed fractures and extensional troughs along the dichotomy boundary and the inferred down dropped highlands block in the Ismenius region. However, one of the two predicted zones of compression is located in the knobby terrain of the lowland bench where no compressional tectonic features are observed. An outer zone of compression predicted by the relaxation models may contribute to the formation of the northern lowlands wrinkle ridges; however, the magnitude of the stresses is relatively small. Also, the orientations of northern lowlands wrinkle ridges near the dichotomy boundary in the eastern hemisphere are roughly perpendicular to the boundary escarpment, suggesting that regional stresses possibly related to the Utopia basin dominated (Head et al. 2002) (**Figure 1e**).

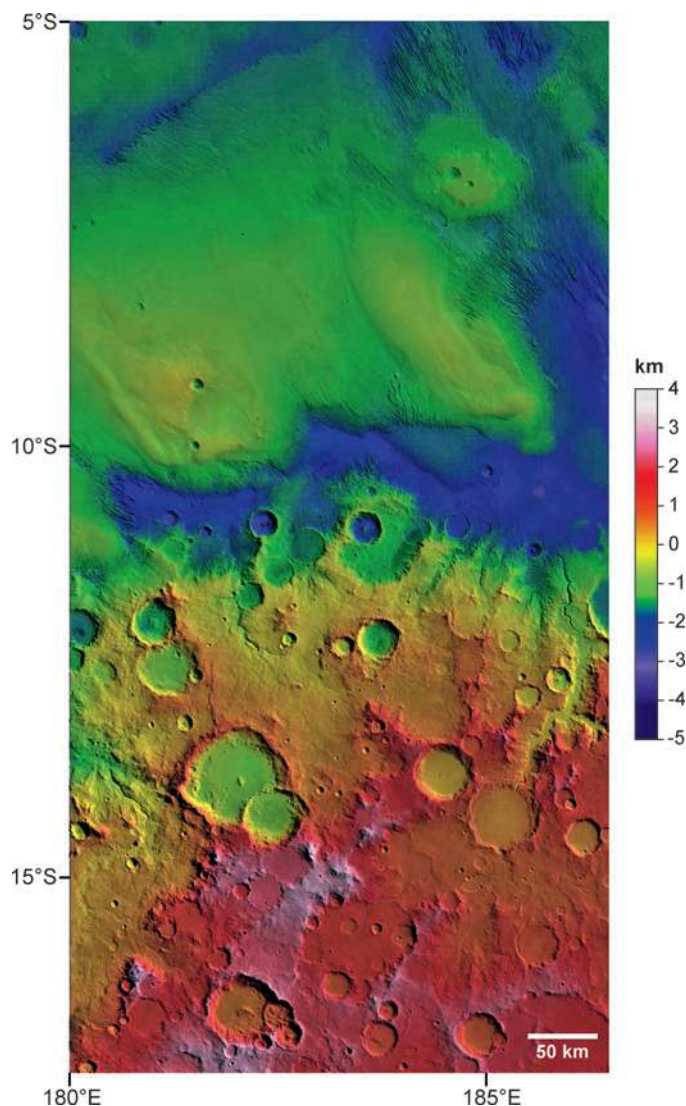
Timing of Tectonic Events and Age of the Boundary Slope

The tectonic features observed along the dichotomy boundary in the eastern hemisphere and within the northern lowlands are likely Late Noachian to Early Hesperian in age, with the onset of tectonic activity no earlier than the Middle Noachian. There are no surviving tectonic landforms that might be associated with the formation of the dichotomy in the very early Noachian age. In the models of Watters (2003a) and Watters & McGovern (2006), flexure of the southern highlands is in response to vertical loading of the northern lowlands from the emplacement of Early Hesperian ridged plains volcanic material. In the model of Nimmo (2005), relaxation of the dichotomy boundary occurs during the Late Noachian to Early Hesperian, consistent with the timing of the formation of the associated tectonic features. Relaxation at this time might have been triggered by the global heating event responsible for widespread ridged plains volcanism (see Watters 1993, Head et al. 2002). Guest & Smrekar (2005) suggest relaxation occurred at 4 Ga (during the Early Noachian) when the crust may have been wet and the thermal gradient was ~ 15 to 20 K/km. However, none of the tectonic features associated with the dichotomy boundary are Early Noachian in age.

Another important constraint on the relaxation and flexure models is when the present-day long-wavelength topographic expression of the dichotomy boundary in the eastern hemisphere (i.e., broad rise and an arching ramp that slopes northward) was established. The age of the sloping ramp can be constrained by the Noachian valley network and modified impact crater morphologies (**Figure 7**) (Irwin & Watters 2007). Areal crater densities indicate that the highlands along the dichotomy boundary in the eastern hemisphere are Early to Middle Noachian in age (Tanaka 1986; Tanaka et al. 2003, 2005). Evidence suggests that the heavily cratered sloping ramp influenced post-Noachian fresh crater morphometry, Late Noachian valley network planform, and the degradation patterns of Middle to Late Noachian impact craters (Irwin & Watters 2007). The plane defined by the rim crest of fresh and slightly modified craters

Figure 7

The cratered slope of the dichotomy boundary in Terra Sirenum. Neither the large, degraded Noachian impact craters nor fresher-appearing, younger craters have experienced tilting or extensional faulting since the time of their emplacement in the Middle to Late Noachian Epochs. Late Noachian valley networks are incised into the slope but show no evidence of tilting or changes in flow direction. These observations suggest that the slope existed before the impact craters formed and was slowly eroded during the Middle and Late Noachian epochs. The ridged surface to the north is the younger superimposed Medusae Fossae Formation. The color-coded MOLA topography is overlaid on a Viking Orbiter mosaic (from the Mars Digital Image Mosaic 2.1).



along the dichotomy boundary is inclined subparallel to the regional slope while the interior cavity is oriented vertically (see figure 6 in Irwin et al. 2005). This suggests the impact craters formed on a preexisting regional slope. Late Noachian valley network development appears to have been controlled by a preexisting boundary slope. Valley network tributaries converge down modern intercrater slopes, and valleys are incised into the low points of many enclosed basin divides (Irwin & Watters 2007). A lack of entrenchment of northward-flowing valleys and infilling of south-flowing valleys, and no evidence of stream piracy among valley networks, suggests the boundary slope did not develop during fluvial activity or a hiatus between two discrete episodes

of activity. There is also no evidence that valley networks on the sloping boundary ramp deviate from topographic gradients or expected basin overflow points (Irwin & Watters 2007). In addition, the floor deposits of modified craters on the boundary slope, emplaced from the Middle Noachian to the Early Hesperian, are flat or slightly concave. If the boundary slope had developed after infilling of these craters, the floors would be expected to slope northward. Also, flat-floored amphitheaters that open to the north are found on the boundary ramp, consistent with lowering and infilling of a crater rim and floor formed on a slope (Irwin & Howard 2002). Thus, the evidence indicates that the slope on the boundary ramp has existed since the Early Noachian. This suggests the present-day slope on the boundary ramp in the eastern hemisphere is a relic slope of the primordial dichotomy boundary or that it is the result of early modification of the primordial boundary by flexure or relaxation.

An ancient boundary slope has important implications for the tectonic features along the present-day dichotomy boundary. The Late Noachian to Early Hesperian extension and compression associated with the dichotomy boundary may not be related to its formation or early modification. The surface expression of any tectonic features resulting from Early Noachian formation or modification of the dichotomy boundary would not be expected to be preserved. One possibility is that the tectonic features associated with the dichotomy boundary formed as a result of late-stage boundary modification owing to relaxation or flexure. Later modification of the boundary might have been triggered by Late Noachian to Early Hesperian ridged plains volcanism that emplaced kilometers of basalt-like material in the northern lowlands (Head et al. 2002) and portions of the highlands in response to a global heating event (see Watters 1993). However, if much of the present-day slope on the boundary ramp developed in the Early Noachian, the amount of Late Noachian–Early Hesperian relaxation or flexure of the boundary must be less than suggested by existing models (Nimmo 2005, Watters 2003a, Watters & McGovern 2006).

Gravity Anomalies and Edge-Driven Convection

Some of the most pronounced free-air and Bouguer gravity anomalies in the northern lowlands are associated with the Utopia and Isidis basins (**Figure 2c**) (Zuber et al. 2000, Yuan et al. 2001, Neumann et al. 2004). The anomalies related to these basins are similar to lunar mascons, attributed in part to the uncompensated mare basalt fill (see Konopliv et al. 1998). In contrast to the roughly circular basin-related anomalies is a distinct positive linear anomaly in the lowlands along the dichotomy boundary in the eastern hemisphere (**Figure 2c**) (Yuan et al. 2001, Neumann et al. 2004). This gravity anomaly extends along a 3000-km-long segment of the dichotomy boundary between 105°E and 160°E (Kiefer 2005). Based on admittance modeling, Nimmo (2002) attributed the anomaly to a subsurface structure exerting a load that might be related to erosion or sedimentation. The anomaly has also been attributed to a thick accumulation of volcanic material along the dichotomy boundary causing a vertical load and inducing flexure of the highlands (Watters 2003a). The results of a spatial domain analysis suggest that the anomaly is due to buried, high-density material along the boundary (Kiefer 2005). The source of the high-density structure may be mantle

material that has replaced crust owing to localized crustal thinning (Kiefer 2005). A possible mechanism for localized crustal thinning along the dichotomy boundary is edge-driven convection (Kiefer 2005).

Edge-driven mantle convection may occur when there are large lateral gradients in lithospheric thickness and thermal structure (King & Redmond 2004). This form of small-scale convection may be localized at terrestrial passive margins and could be associated with intraplate volcanism (King & Ritsema 2000). As discussed previously, the topography of the dichotomy boundary in the eastern hemisphere is reminiscent of a terrestrial continental margin (Frey et al. 1998). Edge-driven convection requires rapid development of lateral variations in lithospheric thermal gradients on the order of a few tens of millions of years to prevent conductive re-equilibration of the lithosphere (Kiefer 2005). Kiefer (2005) suggests that this time constraint is met by an impact formation mechanism for the crustal dichotomy. Rapid endogenic formation mechanisms for the crustal dichotomy, however, may also satisfy this time constraint.

Edge-driven convection associated with the dichotomy boundary may have other important consequences. Modeling suggests that edge-driven convection may have initiated the formation of the Tharsis volcanic province (King & Redmond 2004). King & Redmond (2004) observe that the Tharsis rise is located in the interior of a concave arc of the dichotomy boundary. They suggest the long-lived Tharsis volcanism, often attributed to a mantle plume, could be the result of edge-driven convection along this concave arc. If this is the case, the formation of the hemispheric dichotomy set the stage for the development of the Tharsis rise. Another consequence of edge-driven convection could be early flexure of southern highlands. The replacement of lowlands crust with mantle material along the dichotomy boundary owing to crustal thinning from edge-driven convection (Kiefer 2005) would likely have resulted in a significant vertical load. This vertical load may have induced early lithospheric flexure of the highlands that would account for the long wavelength topography of the dichotomy boundary and the age of the boundary slope.

Erosion and Glacial Modification of the Dichotomy Boundary

The extent that erosion and backwasting of the highlands have modified the dichotomy boundary is important to understanding its evolution. Geologic mapping suggests that the present-day dichotomy boundary has a long and complex history, evolving from an ancient, poorly understood transition zone between the highlands and lowlands (Skinner et al. 2004; Tanaka et al. 2003, 2005). Tanaka et al. (2001) estimated that the eroded material from the highlands might have formed a wedge several hundred kilometers wide and as much as 2 to 4 km high. This constitutes a large volume of highlands material, up to 2×10^7 km³, removed from the boundary region. Erosion of such a large volume of highlands material is a source of compressional stress in the highlands at the dichotomy boundary, and may have contributed to the formation of the lobate scarp thrust faults. The maximum compressional stress from erosion of a 3-km-high and 200-km-wide area of highlands along the dichotomy boundary, approximated by a rectangular finite-width load, is ~ 140 MPa

(Watters 2003a). This is likely an upper limit, however, because such a large zone of erosion would probably have removed the topographic expression of flexure (Watters 2003b) or relaxation. Relaxation modeling suggests that erosion of several kilometers from the highlands is difficult to reconcile with the pattern of tectonic features observed along the dichotomy boundary unless the elastic thickness was large (Nimmo 2005). Also, evidence suggests that the location of the dichotomy boundary may have been largely fixed since its formation. Knob rings in knobby terrain are interpreted to be the erosional remnants of crater rims that suggest a Noachian crater population in the lowlands along the boundary (McGill & Dimitriou 1990). Based on the depth of erosion necessary to remove all impact craters superposed on the lowlands near the dichotomy boundary, it does not appear that large-scale erosion of the boundary could have occurred since the Early Noachian (McGill & Dimitriou 1990, Smrekar et al. 2004). Thus, the dichotomy boundary may have experienced only relatively minor postfaulting erosion. A more or less fixed dichotomy boundary is also supported by recent studies of the origin of fretted terrain. As described above, fretted terrain has been interpreted to be the erosional remnants of highland materials. Evidence suggests that fretted terrain is not formed in highland materials but formed in aeolian sediments deposited along the dichotomy boundary in the Late Noachian to Early Hesperian (Irwin et al. 2004).

Another process that may have contributed to the evolution of the dichotomy boundary is glaciation. Evidence of widespread ice-rich mantle deposits superposed on older terrain, and debris aprons and concentrically ridged lobate features indicative of glacier-like viscous flow, suggest recent ice ages and glacial activity on Mars (Head et al. 2003, 2005). Episodes of glaciation may have been triggered by variations in the planet's obliquity (Head et al. 2003, 2005). Fretted terrain along Deuteronilus-Portonilus Mensae in northern Arabia Terra, a region where the dichotomy boundary extends to the highest northern latitudes, shows evidence of glacial processes in the fretted valleys (Lucchitta 1984, Head et al. 2006). The convergence and merging of lineated valley fill material in the down-valley direction and landforms such as lobe-shaped flows that terminate where valley fill material emerges into the northern lowlands are interpreted to be the result of glacial-like flow. The broad occurrence of glacial-like features in the fretted valleys suggests that glaciation may have been a major erosional process, possibly causing hundreds of kilometers of retreat of the dichotomy boundary in the Amazonian (Head et al. 2006). If, however, the fretted terrain material of Deuteronilus-Portonilus Mensae is aeolian sediments deposited along the dichotomy boundary, as suggested for the fretted terrain in Aeolis Mensae (Irwin et al. 2004), and the location of the boundary has remained largely unchanged since its formation, the dominant role of glaciation in the lowlands may have been in shaping the fretted terrain.

SUMMARY

The new view that has emerged from the unprecedented amount of new data returned from orbiting spacecraft and rovers on the surface is that the hemispheric dichotomy is an ancient and fundamental crustal feature of Mars. It is expressed

by a physiographic divide between the heavily cratered highlands in the southern hemisphere and relatively smooth lowland plains in the northern hemisphere. The dichotomy is manifested in the geology, tectonics, cratering record, gravitational and magnetic fields, and crustal structure. Despite the wealth of new data, the origin of the hemispheric dichotomy, which may have set the course for most of the subsequent geologic evolution of Mars, including the Tharsis volcanic and tectonic province, remains unclear. Endogenic or internally driven models for the dichotomy form the lowlands by mechanisms that involve subcrustal transport through mantle convection or by a superplume, the generation of thinner crust by plate tectonics, rapid mantle overturn after formation of a global magma ocean, or an impact induced, local magma ocean. Exogenic or externally driven models ballistically remove crust from the northern lowlands by either one giant impact or multiple impacts. The time of formation of the dichotomy is an important constraint on these models. Two windows on the age of the lowlands crust are quasi-circular depressions expressed in topographic data and parabolic-shaped echoes in radar sounding data. Both are interpreted to be evidence for a population of buried impact features in the lowlands basement. Areal densities of buried basins suggest that the dichotomy formed early in the geologic evolution of Mars. Lobate scarps and extensional troughs along the dichotomy boundary suggest late-stage tectonic modification of an ancient dichotomy boundary by flexure or relaxation of the highlands after resurfacing of the northern lowlands by ridged plains volcanism. Relatively recent deposition and erosion of material by fluvial, aeolian, and glacial processes have modified, and may continue to shape, the present-day dichotomy boundary.

The plethora of new data for Mars, both currently available and to be obtained in the foreseeable future, has and will continue to increase our understanding of the origin and evolution of the hemispheric dichotomy and the dichotomy boundary. Radar sounder data from MARSIS and now SHARAD (Shallow Subsurface Radar) along with continued investigation of QCDs may allow the age of northern lowlands crust to be much better constrained. Further modeling of early, rapid-developing mantle processes may help narrow the range of possible internal mechanisms for the formation of the crustal dichotomy and clarify the relationship between the dichotomy, the demise of the core dynamo, and the formation of Tharsis. Future quantitative modeling of basin-scale impacts may help determine if multiple impacts can explain the difference in crustal thickness. Higher resolution imaging and spectral, topographic, and geophysical data will also greatly improve our understanding of the evolution of the dichotomy boundary. Additional modeling may help determine the role of flexure and relaxation in reshaping the ancient dichotomy boundary and their relationship to late-stage volcanism and tectonism.

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