MARTIAN RIFTING IN THE ABSENCE OF PLATE TECTONICS?. M. Grott, E. Hauber, *Institute of Planetary Research, German Aerospace Center (DLR), Berlin, Germany (matthias.grott@dlr.de, ernst.hauber@dlr.de)*, P. Kronberg, *Institute of Geology, Technical University of Clausthal, Clausthal-Zellerfeld, Germany*.

Introduction: The forces driving continental rifting on Earth include frictional forces caused by the convecting asthenosphere and tensional plate-boundary forces [1]. Two end-member formation scenarios of continental rifting are discussed as hypothesis for the initiation of rifting. In the active scenario, rifting is driven by a hot mantle plume and extensional stresses originate from prerift uplift, whereas in the passive mechanism rifting is driven by an extensional regional stress field, which is usually assumed to result from remote plate boundary forces [2].

At the onset of rifting, the rheological structure of the lithosphere and its ability to localize strain controls the location, structural style and width of the evolving system [1]. The duration of the rifting stage and whether a rift succeeds in crustal separation or aborts at an earlier stage is then determined to a large extent by the interaction of the lithospheric plates [1].

Recently, Martian candidate analogues to terrestrial continental rifts have been investigated in detail and it has been shown that the Tempe Fossae [3], Acheron Fossae [4] and Thaumasia Highland Rifts [5][6] bear many structural similarities to continental rifts on Earth. However, the question of the rift formation process has so far not been addressed and an active mechanism involving mantle plumes and local doming has usually been assumed. Furthermore, the connection of Martian rifts to plate tectonic forces has so far not been discussed, although rifts are sometimes thought to be at least indirect evidence for plate tectonics [7].

On Mars, we lack geological evidence for plate tectonics, although magnetic field data might support this hypothesis [8]. Also, an early episode of plate tectonics could explain the existence of an early Martian dynamo [9] and the existence of rift structures could be interpreted as further support for the plate tectonics hypothesis [7]. In this paper we will investigate whether forces connected to plate movement are indeed necessary to initiate rifting and show that lithosphere scale faulting at the Thaumasia Highland Rift is feasible even in the absence of mantle plumes or tensional stresses originating from plate-boundary forces. Rather, stresses originating from horizontal differences of the gravitational potential energy will be shown to be almost sufficient to induce rifting, supporting the hypothesis of a passive rifting mechanism in the Thaumasia Highlands [6].

Observations: The Thaumasia Highland rifts are two large extensional structures in the southern Thaumasia region, which are several hundred kilometers long and several tens of kilometers wide [5][6]. The rifts are arranged 90-130 km apart in a nearly subparallel setting, trend approximately SW-NE and cut the Thaumasia highland belt (Fig. 1). Their complex morphology, segmented border faults arranged in en echelon patterns, curvilinear trend and fractured graben floors suggest that they are rifts [10][11]. Like many terrestrial continental rifts the Thaumasia Highland Rifts are associated with several large mountains and topographic rises, which are interpreted as volcanoes [11][6].

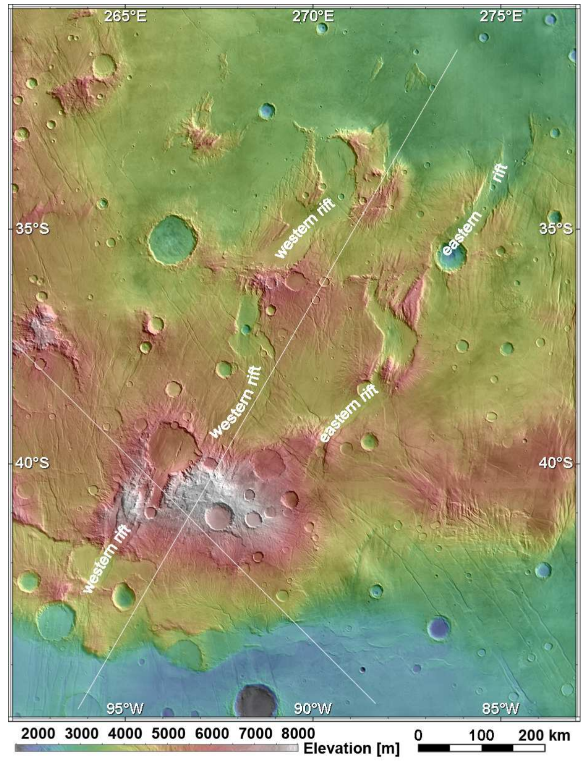

Figure 1: Topographic image map of study area (image data merged with MOLA elevation data). The double rift system is clearly visible as a series of segmented topographic grabens bounded by faults. The locations of the topographic profiles shown in Fig. 2 are also indicated.

The time of rifting has been constrained by evaluating the crater size-frequency distribution of key surface units associated with the rifting and rifting ages of 3.5 to 3.9 Gyr have been obtained [5]. This is consistent with detailed mapping by [11], who assign the faulting of the Coracis Fossae to the Noachian and Early Hesperian periods.

Estimates of the elastic lithosphere thickness at the time of rifting have been obtained from modeling the flank uplift, a result of the isostatic lithospheric rebound driven by the unloading associated with extension. Thus, the elastic thickness was constrained to 10.3 - 12.5 km [5]. Using the strength envelope formalism [12], the corresponding mechanical thickness and thermal gradient were also determined, yielding 15 - 19 km and 27 - 33 K km^{-1}, respectively [5].

The initial ambient conditions at the Thaumasia Highland

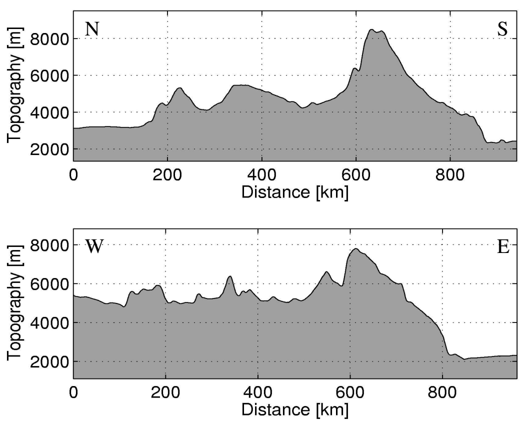

Figure 2: Topographic profiles taken along the lines indicated in Fig. 1. Top: NNE-SSW trending profile. Bottom: NW-SE trending profile.

Rifts are only compatible with the observed narrow rift mode of extension, if the crustal thickness at 3.5-3.9 Gyr was smaller than ∼50 km. As this is significantly lower than the estimates for the present day crustal thickness in that region (∼70 km), this is interpreted as an indication for continued crustal production at that time [13].

It has also been shown that strain localization by volcanic weak zones is a major factor for fault-pattern formation at the Thaumasia Highland rifts, indicating widespread pre- and synrift volcanic activity in the region [6]. Geological evidence indicates that faults initiate at the volcanoes and propagate away from them before interconnecting [6] and the fault patterns are best compatible with extensional stresses acting in the NW-SE to NNW-SSE directions.

Stress State: The Thaumasia Highlands are part of the Tharsis main bulge and the stresses leading to the formation of Tharsis need to be considered as possible sources for the formation of the Thaumasia Highland rifts. Lithospheric loading, induced isostatic compensation and lithospheric flexure models have been proposed (see [14] for a review), but the lithospheric flexure model is generally ruled out on the grounds that it does not predict extension compatible with Valles Marineris and the lithospheric loading and isostatic compensation models predict extension circumferential to Tharsis, contrary to the extension directions derived by [6].

A recent study [15] has investigated the stress field associated with horizontal gradients of the gravitational potential energy (GPE) and predict extension and normal faulting in elevated areas such as the Thaumasia Highland belt. Although larger extension is predicted for the NE-SW direction, extension in the NW-SE direction as required by the observed fault pattern is also present. This has led [6] to propose that the stress state in the Thaumasia Highland Belt may be comparable to that in the high Himalayas and Tibet [16]. In this analogy the Thaumasia Highland rifts would correspond to the north-south trending Quaternary grabens there. This interpretation is

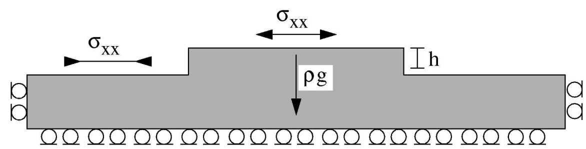

Figure 3: Model setup for the calculation of stresses caused by topography. The left and right boundaries are restricted to movement in the vertical direction, the bottom boundary is restricted to horizontal movement. The body force ρg is applied and the resulting stresses are calculated.

also supported by recent work of [7], who have shown that the Thaumasia Highland Belt is structurally similar to orogenic belts on Earth, and E-W trending lobate scarps indicate a N-S directed compressional environment [e.g., 17].

In the following we will consider if the stresses derived from horizontal gradients of GPE are sufficient to initiate lithosphere scale faulting in the Thaumasia Highland Belt. The stresses may be estimated by considering the vertically averaged stress of a column of lithosphere and the GPE per square meter is then given by

$$E(h) = \int_0^D \rho g z' dz' \tag{1}$$

where ρ is density, g is gravitational acceleration and D the lithospheric thickness. Horizontal gradients of $E(D)$ will be balanced by horizontal stress gradients in the lithosphere and the magnitude of horizontal stresses $\bar{\sigma}_{xx}$ may be estimated for a simple two-dimensional model by

$$\bar{\sigma}^1_{xx} = \frac{E(D_1) - E(D_2)}{D_1} \tag{2}$$

Fig. 2 shows topographic profiles of the Thaumasia Highland Belt taken along the lines indicated in Fig. 1. The Belt stands up to 6000 m above Icaria Planum to the south and 5000 m above Solis Planum to the north, resulting in extension of 50 MPa. This is of the same order of magnitude as the average strength of the lithosphere, which is 50-70 MPa for the thermal gradients derived from rift flank uplift [5]. In the following, we will present a more detailed analysis, using finite element modelling.

Modeling: We use a two dimensional plane-strain finite element model to calculate the stresses resulting from the differences in GPE associated with topography, which is sketched in Fig. 3. The lithospheric thickness is taken to be equal to the mechanical thickness of the lithosphere, as derived from rift flank uplift (∼20 km) [5]. The left and right boundaries are restricted to movement in the vertical direction, the lower boundary is restricted to horizontal movement and assumed to overly an inviscid substratum. Topography of height h is added to the setup and the structure is loaded by the body force ρg. The lithosphere is assumed to be in a lithostatic state of stress and thus we choose the Poisson's ratio to be $\nu = 0.5$. Young's modulus is taken to be $E = 100$ GPa, density ρ is 3000 kg m^{-3} and the gravitational acceleration is 3.72 m s^{-2}.

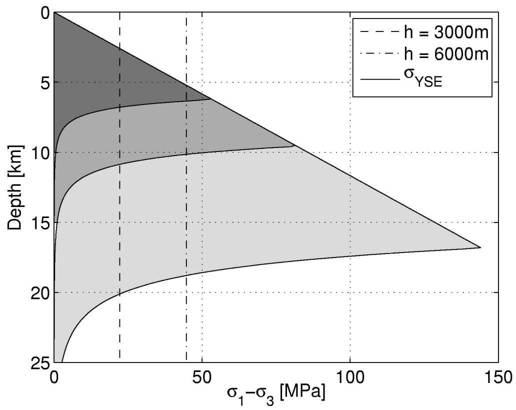

Figure 4: Yield strength envelope for a diabase crust and thermal gradients of 27 (light), 50 (medium) and 80 K km^{-1} (dark). The deviatoric shear stresses resulting from 3000 and 6000 m of topography are also indicated.

We have run two models for topographic loads of 3000 and 6000 m, respectively. The results are summarized in Fig. 4, where the resulting deviatoric shear stress $\sigma_1 - \sigma_3$ is given for the two models. As the lithosphere is assumed to be in a lithostatic state of stress, the deviatoric shear stress is depth independent. For $h = 3000$ and 6000 m, the resulting stresses are ∼22 MPa and 44 MPa, respectively.

As a comparison, the lithospheric strength with respect to extension is also given in Fig. 4, where brittle and ductile yielding are calculated from Byerlee's law and a flow law appropriate for wet diabase [18], respectively. Strength envelopes are given for thermal gradients of 27, 50 and 80 K km^{-1}. For $dT/dz = 27\,\mathrm{K\,km^{-1}}$, the thermal gradient derived from flexure analysis of the rift flank uplift, the lithosphere will yield in the shallow upper and deep lower layers, retaining an elastic core of significant thickness. This implies that although the Thaumasia Highland Belt is in a state of significant extension, lithosphere scale yielding is unlikely for the background thermal gradient in the region during the late Noachian.

One way to reduce the yield strength of the lithosphere is by a thermal disturbance, which could, e.g., be caused by the emplacement of a magma-body in the upper crust. Also, the buoyancy of hot magma would further act to reduce lithospheric strength. Concerning the temperature effect only, a lifting of the 700 K isotherm, corresponding to the brittle-ductile transition in wet basaltic rock (e.g, [18]), to a depth of 7 km would be sufficient to weaken the entire lithosphere to the point of yielding. This scenario would correspond to a geothermal gradient of 80 K km^{-1} or a surface heat flow of ∼200 mW m^{-2}. Although large, these heat flows are not unreached in magmatically active regions on Earth.

Discussion: We have shown that due to its high-standing topography the Thaumasia Highland is in a state of extension with extensional stresses of the order of 40 MPa. Although these stresses are sufficient to induce thin-skinned deformation to a depth of ∼5 km, they cannot account for the observed lithosphere scale faulting. Therefore, other sources of stress and/or weakening of the lithosphere need to be considered.

The Solis Planum, situated north of the study area, is topographically elevated by 1-1.5 km with respect to the Icaria Planum to the south. This topographic gradient should result in N-S directed compressional stresses, which are expected to be of the order of a few tens of MPa. This compression will increase the deviatoric stresses, thus weakening the lithosphere. Although probably not a key mechanism, the compression could facilitate the observed extension.

Furthermore, the emplacement of magma bodies in the upper crust could sufficiently weaken the lithosphere to initiate lithosphere scale faulting and thus induce rifting. This hypothesis is in good agreement with the observation of rift related volcanism [5][6] as well as the fact that faults seem to initiate at the volcanoes and propagate away from them before interconnecting [6].

We conclude that rifting on Mars is feasible even if key factors connected to continental rifting on Earth, i.e. plate boundary forces and convection induced drag on the lower lithosphere, are absent. Rather, stresses originating from horizontal gradients in GPE are almost sufficient to induce lithosphere scale faulting of the young Martian lithosphere. The rifting process could then be initiated by the emplacement of magma in the upper crust. The absence of forces connected to plate tectonics is also consistent with the observed moderate extension of only a few kilometers. These values are typical for young terrestrial rifts (e.g., the Kenya-rift) and failed arms and suggest that large scale plate movement and subduction did not play a role in Martian rifting.

Although we have investigated the environment of only one rift system, the ambient conditions at other Martian rifts share similarities to those reported here. In particular, all rifts are found in topographically elevated areas and are connected to magmatic activity [3][4]. They are Noachian to early Hesperian in age and the lithosphere at the time of rift formation was probably very thin [4]. It therefore seems debatable if the observed rifts bear witness to an early episode of plate tectonics on Mars.

References: [1] Ziegler and Cloetingh (2004), *Earth Sci. Rev.*, 64, 1-50. [2] Corti et al. (2003), *Earth Sci. Rev.*, 63, 169-247. [3] Hauber and Kronberg (2001), *J. Geophys. Res.*, 106, E9, 20587-20602. [4] Kronberg et al. (2007),*J. Geophys. Res.*, in press. [5] Grott et al. (2005), *Geophys. Res. Lett.*, 32, L21201. [6] Grott et al. (2007), *J. Geophys. Res.*, in press. [7] Anguita et al. (2006), *Icarus*, 185, 2, 331-357. [8] Connerney et al. (2004), *PNAS*, 102, 42, 14970-14975. [9] Breuer and Spohn (2003), *J. Geophys. Res.*, 108, E7, 5072. [10] Dohm and Tanaka (1999), *Plant. Space Sci.*, 47, 411-431. [11] Dohm et al. (2001), *Geologic Investigations Series I-2650*. [12] McNutt (1984), *J. Geophys. Res.*, 89, 11180-11194. [13] Grott (2005), *Geophys. Res. Lett.*, 32, L23201. [14] Mège and Masson (1996), *Plan. Space Sci.*, 44, 1471-1497. [15] Dimitrova et al. (2006), *Geophys. Res. Lett.*, 33, 8, L08202. [16] Flesch et al. (2001) *J. Geophys. Res.*, 106, B8, 16435-16460. [17] Grott et al. (2006), *Icarus*, 186, 517-526. [18] Karato et al. (1986), *J. Geophys. Res.*, 91, 8151-8176.