Seismotectonics of the Amenthes Rupes thrust fault population, Mars

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[1] Thrust faults within the Amenthes Rupes population in eastern Mars accommodated contractual strains of ~0.06% with total cumulative moment (energy) release of ~2 × 10^{24} \text{MJ} during Late Noachian time. “Small” faults, that scale three-dimensionally, accommodate only 12~17% of the strain and 6~9% of the moment release, indicating that inventories of Martian faults need not be complete down to the smallest lengths for meaningful estimates of strain and moment to be obtained. Frictional stability relations for Martian lithosphere suggest that most (83%) of the largest fault, Amenthes Rupes, was seismogenic during the duration of faulting and its potential for generating Marsquakes during Noachian to Hesperian time.

[2] Thirteen major faults in the array define the thrust fault population [e.g., see Watters et al., 2000] that accommodates primarily northeast to southwest directed contraction within this domain. Map lengths \(L\) range from ~68 to ~421 km along the gently curved surface breaks. The area containing the fault population is shown in Watters and Robinson [1999], with fault coordinates and lengths listed in AGU’s web supplement to Watters et al. [2000]. Average displacements are taken to be 0.67 of the maximum displacement \(D_{\text{max}}\) along the fault [e.g., Wilkins et al., 2002]; \(D_{\text{max}}\) is obtained from forward modeling of topography at the fault break [Schultz and Watters, 2001]. For this population, \(D_{\text{max}}/L = 6 \times 10^{-3}\) [Watters et al., 2000].

[3] Faults typically have elliptical shapes, with aspect ratios (\(L/H\)) of 2~3 [Nicoll et al., 1996]. The down-dip fault height \(H\) is limited by the thickness \(T\) of the faulted domain (normally taken to be the seismogenic depth [e.g., Scholz, 1997]), with “small” faults defined by \(H = L/2 < H_{\text{max}}\) and “large” faults by \(H = H_{\text{max}}\), where \(H_{\text{max}} = T \sin \delta\). \(H\) is evaluated for each fault in the population by using the maximum value obtained for \(T\), from the largest fault (Amenthes Rupes), of 30 km (using bounded dip angles of \(\delta = 30^\circ\) [Schultz and Watters, 2001]), so that the maximum down-dip fault height \(H_{\text{max}} = 60\) km.

[4] The total geometric moment for the thrust fault population (for \(L/H = 2\)) is \(9.3 \times 10^{13}\) \text{m}^2. The contribution of each fault to the total geometric moment, shown in Figure 1, reveals that ~9% of the total moment is borne by “small” faults (assuming \(L/H = 2\); i.e., with lengths <124 km (Figure 1a). A smaller percentage (7%) of moment is contained in “small” faults for \(L/H = 3\) (\(L < 185\) km; \(M_g = 8.7 \times 10^{13}\) m^2). These values demonstrate that reliable estimates of brittle moment and strain (see below) from Martian fault populations can be obtained even if the smallest faults (and still smaller blind faults that do not break the surface) are not included (see parallel and independent conclusions for Martian normal fault populations in Tempe Terra by Wilkins et al. [2002]).

[5] Total geometric moment is converted into cumulative geologic moment \(M_0\) by using \(M_0 = M_g G\), where \(G\) is the
Taking $G = 20$ GPa gives a value of $M_0 = 1.86 \times 10^{24}$ N-m (or MJ) that represents the energy that was consumed during the growth of the thrust fault population (assuming $L/H = 2$). Typical uncertainties in $G$ contribute perhaps a factor of two in $M_0$.

Brittle strain normal to the fault array (i.e., in the vergence or contraction direction) is obtained from the fault population data by using

$$
\varepsilon = \frac{1}{V} \sum_{k=1}^{N} \left[ (D_{av}) L^4 H^4 \sin \delta \cos \delta \right] \text{ for small faults}
$$

$$
\varepsilon = \frac{1}{A} \sum_{k=1}^{N} \left[ (D_{av}) L^4 \cos \delta \right] \text{ for large faults}
$$

for small faults and large faults, respectively, where $D_{av}$ is the average displacement, $V$ is the volume of the faulted domain, and $A$ is the area of the faulted domain. The total brittle strain accommodated by the thrust fault population, accounting for both “small” and “large” faults by using (2), is $\sim 0.06\%$ (Figure 1b). Using $L/H = 3$, the strain is reduced by 6% to 0.058%. This value (0.06%) is a spatial average over the entire faulted domain; local strains near the individual thrust faults would be larger [e.g., Schultz and Lin, 2001]. Amenthes Rupes itself accommodates almost half (44%) the strain across the entire domain.

The maximum depth of faulting $T$ is defined approximately by the 300°C isotherm, which is associated with the lower stability transition between unstable (seismogenic) frictional sliding above and stable sliding (creep) below [e.g., Tse and Rice, 1986; Scholz, 1998]. Using $T = 30$ km [Schultz and Watters, 2001] for the faulted domain, the paleogeothermal gradient during Martian thrust faulting was approximately $10^6 \degree C$ m$^{-1}$ (assuming a surface temperature of $\sim 0\degree C$ and an approximately linear gradient). Maximum (limiting) values of the paleogeotherm (e.g., within a factor of 2 [Schultz and Watters, 2001]) can be obtained for the

Figure 1. (a) Geometric moment and (b) brittle strain for thrust faults within the Amenthes Rupes population.

Figure 2. Frictional stability of Martian faults showing depth ranges for seismic slip.
same depth of faulting (30 km) by using an approach based on lithospheric strength envelopes [e.g., Maggi et al., 2000]. Down-dip portions of the Martian thrust faults, deeper than 30 km, would tend to slip stably but would contribute only small components to the surface topography, given their greater depth below the surface [e.g., Cohen, 1999].

[12] The upper (shallow) limit of seismogenic slip is related to the upper stability transition [Marone, 1998; Scholz, 1998], above which fault zone material (gouge) is velocity strengthening [Marone and Scholz, 1988]. This upper stability transition is pressure dependent and independent of fault type [Scholz, 1998]. By scaling the values used for terrestrial faults (3–4 km [Cowie et al., 1993; Scholz, 1998]), hydrostatic pore-fluid conditions) to Martian conditions (g = 3.7 m s⁻²), frictional sliding along Martian faults should be conditionally stable (barrel large perturbations, such as Marsquakes on subjacent fault segments, or rapid healing processes; see Scholz [1998]) at depths shallower than ~8–10 km for a “wet” Noachian crust (hydrostatic pore-fluid pressure) or ~5–7 km for a “dry” Noachian crust (Figure 2). An active hydrologic system (“wet” crust), along with slow slip rates along the faults, would promote healing of the fault zone, leading to decreasing depth for the upper stability transition. Given the lack of constraints on the pore-pressure state of the Martian crust at the time of thrust faulting, however, a value of 8 km is used in this paper.

[13] Seismogenic (unstable) frictional sliding along the largest thrust fault in the population, Amenthes Rupes, should occur primarily between depths of 8 and 30 km (corresponding to the likely Marsquake nucleation depth; Figures 2 and 3). Using the depth range obtained above for unstable frictional sliding, ~82% of the total moment release and strain associated with the Martian thrust fault population should be seismogenic (80% for L/H = 2). The fraction of seismogenic strain will decrease for smaller and less deeply penetrating surface-breaking faults from other areas on Mars given that the upper ~8 km should remain largely devoid of nucleating Marsquakes.

[14] Although slip rates for Martian faults are unknown, plausible ranges can be obtained from the slip rates of terrestrial thrust faults. Slip rates of plate-bounding thrust faults typically exceed 0.2–1 mm yr⁻¹, with comparable [e.g., Champion et al., 2001] or slower rates of 0.01–1 mm yr⁻¹ being common for intraplate thrust faults [e.g., Wescousky, 1999]. Using the intraplate slip rates, the amount of time needed for 1.5 km of thrust displacement on the largest Martian fault in the population, Amenthes Rupes [Schultz and Watters, 2001], to accumulate would be t = 1.5–150 m.y. The average strain rate for the population (assuming slip rates of 0.01–1 mm yr⁻¹) is /C0 = 10⁻¹⁷ to 10⁻¹⁹ s⁻¹, which is the same order as the values for terrestrial plate interiors [Gordon, 1998]. A constant value of regional strain rate implies that slip rates must vary along individual faults in the population [Marrett, 1994], with the longest faults such as Amenthes Rupes having the slowest average slip rates.

3. Conclusions

[15] The magnitude of brittle strain accommodated by thrust faults in the Amenthes Rupes population is quantified for the first time by calculating their total geometric moment. As in other studies [e.g., Cowie et al., 1993], “small” faults contribute only a small fraction of the total moment and strain. Fault segments deeper than the ~8 km upper stability transition, and Shallower than the ~300°C isotherm, should generate Marsquakes, leading to a seismogenic strain that is less than the total brittle strain.

[16] Slow slip rates inferred for the Martian thrust faults imply that seismogenic rupture along the faults would promote larger earthquakes and attendant ground motions [Anderson et al., 1996] than would be expected from standard compilations that assume faster (and constant) slip rates (~1 mm/yr; Wells and Coppersmith [1994]) unless the fast slip rates inferred for some terrestrial intraplate thrust faults are characteristic of the Martian faults. Observations of morphologic changes near Martian faults, such as slope failures, may test whether fault slip was episodic (for slower slip rates and longer recurrence intervals [e.g., Marrett, 1994]) or more continuous (for faster slip rates), within and between tectonic regions. 

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References


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