MARTIAN SEISMICITY

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Introduction. The design and ultimate success of network seismology experiments on Mars depends on the present level of martian seismicity. Volcanic and tectonic landforms observed from imaging experiments show that Mars must have been a seismically active planet in the past and there is no reason to discount the notion that Mars is seismically active today but at a lower level of activity. Here we explore models for present-day seismicity.

Method. The frequency of occurrence, N_f, of marsquakes as a function of surface wave magnitude, M_s, is given by the relationship

$$dN_f/dM_s = A_f \exp(-BM_s) \tag{1}$$

where the B-value adopted is 2.07 [1]. The maximum rate of shear strain over a time interval t can be related to the sum of the seismic moments, ΣM_0, through [2]

$$d\varepsilon_{max}/dt = \Sigma M_0/\mu Vt \tag{2}$$

where V is the seismogenic volume and μ is the shear modulus (here taken at $7\text{x}10^{10}$ Pa). The sum of the seismic moments can be expressed as

$$\Sigma M_0 = \int_{-\infty}^{M_s(\mathrm{Max})} M_0(Ms)[dN_f/dMs]dMs \tag{3}$$

where N_f is the rate of occurrence of marsquakes (a function of Ms) and M_s(Max) is the magnitude of the largest marsquake, which we guess by comparison to the Earth and Moon. We use a relationship between seismic moment and shear wave magnitude for intraplate earthquakes in the terrestrial oceanic lithosphere [3] (these events are probably thermoelastic in origin):

$$M_0(M_s) = 4.68\text{x}10^{11}\exp(2.58M_s) \tag{4}$$

where the units are N-m. Substituting (1), (2), and (4) into (3) yields in units of (yr-magnitude)$^{-1}$

$$A_f = 2.40\text{x}10^6[(d\varepsilon_{max}/dt)V]\exp[-0.509M_s(\mathrm{Max})] \tag{5}$$

Substitution of (5) into (1) and integrating over the interval (M_s, Ms + ΔM_s) yields the rate of occurrence of marsquakes in that interval as a function of M_s(Max), strain rate, and seismogenic volume. We proceed to estimate strain rates and the attendant seismogenic volumes for various processes operating in the martian environment.

Sources of Global Seismicity. Sources of martian seismicity include thermoelastic cooling, regional loading of the surface including Tharsis and the polar caps, changes in the principal moments of inertia, obliquity changes, daily and annual solar tides, and atmospheric coupling. Seismicity is estimated for the first three mechanisms.

We performed a simple parameterized convection calculation [4], which shows that the mantle of Mars has cooled about 3° C and the lithosphere has cooled about 10° C (and thickened) in the last 10^8 years. In the oceanic lithosphere, earthquakes occur as deep as the 800° C isotherm [5]. In the martian thermal model this corresponds to a depth of about 150 km, which defines the seismogenic thickness, Z_L. The cooling rate in the lithosphere can be related to strain rate through the coefficient of thermal expansion, α. If $Z_L << R_0$, the martian radius, then

$$d\varepsilon_{max}/dt \approx \alpha(dT/dt)Z_L/R_0 \qquad (6)$$

Using $M_s(Max) = 6$, strain rate as calculated from equation (6) ($= -6\times10^{-21}\ s^{-1}$), and a spherical shell of thickness Z_L for seismogenic volume yields the frequency distribution for marsquakes shown in the figure below. From this distribution we see that thermoelastic cooling of the lithosphere would be expected to yield 17 events greater than magnitude 4 in a decade. This result is probably a lower bound in that much higher cooling rates without a proportional decrease in seismogenic volume might be expected in the Tharsis region. Whether such events are detected in a network depends, *inter alia*, on the seismic attenuation and scattering properties of the martian interior [6].

If the Tharsis load is supported in part by thermal buoyancy associated with a magmatic source region, then loss of buoyancy due to cooling leads to a strain rate:

$$d\varepsilon/dt \approx -\rho\alpha(dT/dt)g_0h/E \qquad (7)$$

where g_0 is planetary gravity, h is the vertical dimension of the cooling region, and E is Young's modulus. Using h = 200 km and the same cooling rate as above leads to $d\varepsilon/dt = -1\times10^{-21}$ with $E = 2\times10^{11}$ Pa. Combined with a smaller seismogenic volume, this process appears to be less important than thermoelastic cooling. The cooling and contraction of Tharsis will also change the difference in the principal moments of inertia, but the accompanying strain rate is also small relative to thermoelastic effects *per se*.

Conclusions. Depending on the sensitivity and geometry of a seismic network and the attenuation and scattering properties of the interior, it appears that a reasonable number of martian seismic events would be detected over the period of a decade. The thermoelastic cooling mechanism as estimated is surely a lower bound, and a more refined estimate would take into account specifically the regional cooling of Tharsis and lead to a higher frequency of seismic events. This analysis should be repeated using the moment-magnitude relationship for body waves, and other mechanisms, as listed above, should also be evaluated.

References. [1] F.D. Stacey, *Physics of the Earth*, 113 (John Wiley, New York, 1977); [2] S.R. Bratt, E.A. Bergman, and S.C. Solomon, *J. Geophys. Res.*, **90**, 10249 (1985); [3] E.A. Bergman, *Tectonophysics*, **132**, 1 (1986); [4] R.J. Phillips and M.C. Malin, in *Venus*, 159-214 (University of Arizona Press, Tucson, 1983); [5] D.A. Wiens and S. Stein, *J. Geophys. Res.*, **90**, 6455 (1983); [6] N.R. Goins and A.R. Lazarewicz, *Geophys. Res. Lett.*, **6**, 368 (1979).