Tidal dissipation in Mars: where and how?

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Introduction: The orbit of Phobos is secularly decaying at a rate which implies that Mars is highly dissipative; even more so than Earth's mantle. Most tidal dissipation on Earth occurs within the oceans. Where and how the tidal energy dissipation occurs within Mars is not presently well constrained. We examine several alternatives, including a uniformly warm mantle, a partial melt zone within the mantle, atmospheric tides, and dissipation within a crustal aquifer system. We also discuss how future observations may better constrain this process.

Tidal observations: The primary tidal signatures seen by spacecraft in orbit about Mars are solar gravitational tides, with both semi-diurnal and semi-annual components visible. The imposed tidal potential causes deformation of the body, which induces further gravitational change. The ratio of imposed to induced potential is the tidal Love number [1] and is diagnostic of density and rigidity within the body, with softer and/or less dense bodies being more pliable. For Mars the solar tidal Love number is quite large [2] and this has been interpreted to imply existence of a fluid core [3], though it could equally well imply a soft zone within the mantle [4].

Another tidal signal on Mars is that raised by Phobos. It is most clearly seen via a small phase lag, which causes the tidal bulge to lag behind the current position of Phobos and results in a net secular torque, which causes the orbit of Phobos to evolve inward toward Mars [4-7]. The rate of tidal evolution gives a measure of dissipation within the mantle of Mars, and it is surprisingly high. The rate of change in the orbital energy of Phobos is 3.3 MW [4], and the tidal quality factor is Q = 80 ± 1 [7]. For comparison, the tidal quality factor of Earth's mantle is 280 [8], implying that Mars is substantially more dissipative than the solid parts of Earth. Of course, most of the tidal dissipation within the Earth-Moon system occurs in the oceans [9,10]. Where and how it occurs on Mars is still largely unknown, and is the question we attempt to address.

Warm mantle models: The simplest model of dissipation within Mars assumes a homogenous, incompressible Maxwell visco-elastic structure. That model has just 3 parameters, a density, rigidity μ, and viscosity η. The density is well known, and requiring the model to match the observed Love number and the secular acceleration of Phobos yields estimates of $\mu = (4.6 \pm 2.0)\ 10^{10}$ Pa and $\eta = (8.7 \pm 0.6)\ 10^{14}$ Pa s [4]. A viscosity that low would not support topography for long. We thus need to consider models in which there is a strong upper layer, and also a region of high dissipation.

Most models of martian thermal evolution include a magma ocean [11,12,13]. Whether or not such a feature persists to the present is less obvious, but cannot be easily ruled out [14]. Much of the uncertainty derives from lack of definitive constraints on the elemental and mineralogical composition of the martian interior [15,16], and equally profound ignorance of the depth variation of temperature and heat sources [16,17].

If a partial melt region exists, it would have a lower viscosity than the surrounding layers, but may be more dissipative than that effect alone would suggest [18, 19], since the flow is not that of a homogeneous visco-elastic medium, but is more akin to squeezing water from a sponge (see below).

Atmospheric tides: There is ample evidence for atmospheric tidal effects on Mars, though the most important such effects are thermal, rather than gravitational, in origin [20, 21]. The possibility of tidal dissipation associated with the gravitational tide raised by Phobos has been examined [22], but found wanting. The energy in the dominant tidal mode excited by Phobos would be 20-30 10^3 J, and the damping time required to provide the observed 3.3 MW is thus 100-200 s. This is much shorter than the inferred radiative damping time of 10^5 s [23].

Global aquifer: The possibility of regional to global scale aquifer systems on Mars has been considered in a number of contexts [24-27] but apparently not in the context of tidal energy dissipation. It is known from terrestrial experience that water levels in wells fluctuate with the tides, even far from oceans or lakes. As tidal stress compresses the rock, any pore water will be forced out. The volumetric rate of energy dissipation can be roughly estimated as follows.

Consider a cylindrical fluid reservoir of radius L, surrounded by an aquifer of permeability k and thickness h. Tidal strains will squeeze water in and out of the aquifer (like a sponge). Fluid pressure gradients will extend into the aquifer a distance

$$\delta \sim \sqrt{\kappa \, \Delta t}$$

where κ is hydraulic diffusivity and Δt is the forcing period. The amplitude ΔP of the pressure variations depends on the solid body tides, but the distance over which they extend, and thus the pressure gradient, depends upon the diffusion length scale δ. Assuming κ = 1 $m^2 \, s^{-1}$, and using the 11.1 hour orbital period of Phobos, as seen from the surface of Mars, we estimate δ = 200 m.

The fluid velocity u depends on the pressure gradient, the permeability and the fluid viscosity η. The energy dissipation rate per unit rock volume due to viscous friction as the pore fluid is forced backwards and forwards is just the product of stress times strain rate, and for laminar flow can be estimated from

$$H \sim \frac{C\,k}{\eta}\left(\frac{\Delta P}{\delta}\right)^2$$

where C is a constant in the permeability-porosity relationship which is typically in the range 10-10^3 [28], and we are assuming cylindrical pore geometry [29]. If the reservoirs are uniformly distributed across Mars, the rock volume containing fluid in motion will be

$$V \sim 4 R^2 \, h \, \delta \, / \, L$$

where R is the radius of Mars. The total rate of dissipation is thus

$$\frac{dE}{dt} = H\,V \sim \frac{4\,h\,C\,k}{\delta\,L\,\eta}\left(R\,\Delta P\right)^2$$

If we take h = 3 km [25], $\eta = 10^{-3}$ Pa s, R = 3400 km, C = 30 and ΔP = 10 Pa (for a solid-body tide raised by Phobos), we estimate a global aquifer-related tidal dissipation rate of

$$\frac{dE}{dt} \sim 2\ MW\left(\frac{k}{10^{-11}\,m^2}\right)\left(\frac{10\,km}{L}\right)$$

This estimate is proportionately reduced if the reservoirs only span a fraction of the martian subsurface. The permeability k of the martian crust is poorly known, but estimates as high as 10^{-9} m^2 have been proposed [30, 31]. The length scale L represents the characteristic distance over which the mechanical properties of the martian aquifer vary, and is unknown. Nonetheless, this simple calculation suggests that tidally-driven fluxing of water through subsurface aquifers could result in substantial dissipation.

If even half of the observed 3.3 MW of dissipation comes from tidally-driven groundwater motion, it would make the requirements on the deep interior much less stringent.

Future prospects: When a global network of seismic monitoring stations is deployed on Mars, we should obtain a much better view of the deep subsurface structure. If the seismometers have sufficiently good performance at low frequency, they can plausibly perform as gravimeters, and monitor the tidal response at the surface of Mars. In that case, the amplitude and phase of the tidal response could be directly monitored, at several locations.

References: [1] Munk WH and MacDonald GJF (1960) Rotation of the Earth, Camb. Univ. Press. [2] Yuan DN et al. (2001) JGR 106, 23377-23401. [3] Yoder CF et al. (2003) Science, 300, 299-303. [4] Bills BG et al. (2005) JGR, 110, E07004. [5] Sinclair AT (1989) A&A, 220, 321-328. [6] Bell JM et al. (2005) Nature, 436, 55-57. [7] Lainey V et al. (2007) A&A, 465, 1075-1084. [8] Ray RD et al. (2001) GJI, 144, 471-480. [9] Egbert GD and Ray RD (2001) JGR, 106 22475-22502. [10] Dickey JO et al. (1994) Science 265, 482-490. [11] Righter K and Drake MJ (1996) Icarus 124, 513-529. [12] Elkins-Tanton LT (2008) EPSL 271, 181-191. [13] Reese CC and Solomatov VS (2006) Icarus 184, 102-120. [14] Reese CC et al. (2002) JGR 107 5082. [15] Fey YW et al. (1995) Science 268, 1892-1894. [16] Bertka CM and Fei YW (1997) JGR 102, 5251-5264 [17] Khan A and Connolly JAD (2008) JGR 113, E07003 [18] Green DH and Cooper RF (1993) JGR, 98, 19,807-19,817. [19] Jackson I et al. (2006) Mat. Sci. Eng. 442, 170-174 [20] Zurek, RW (1976) J. Atmos. Sci. 33, 321-337. [21] Wilson RJ and Hamilton K (1996) J. Atmos. Sci. 53, 1290-1326. [22] Rondanelli, R et al. (2006) GRL, 33, L15201. [23] Zurek RW et al (1992) Mars, Univ. Ariz. Press. [24] Carr MH (1979) JGR 84, 2995-3007 [25] Clifford SM (1993) JGR, 98, 10973-11016. [26] Andrews-Hanna JC and Phillips RJ (2007) JGR 112, E08001. [27] Harrison KP and Grimm RE (2008) JGR 113, E02002. [28] McKenzie D (1989) EPSL 95, 53-72. [29] Turcotte D and Schubert G (2002) Geodynamics, Camb. Univ. Press. [30] Hanna JC and Phillips RJ (2005) JGR 110, E101004. [31] Manga M (2004) GRL 31, L02702.