

**CONSTRAINING THE COMPOSITION AND THERMAL STATE OF MARS FROM INVERSION OF GEOPHYSICAL DATA.** A. Khan<sup>1</sup>, J. A. D. Connolly<sup>2</sup>, <sup>1</sup>Niels Bohr Institute, University of Copenhagen, Denmark (amir@gfy.ku.dk), <sup>2</sup>Earth Sciences Department, Swiss Federal Institute of Technology (james.connolly@erdw.ethz.ch).

**Introduction.** The internal structure of a planet provides essential clues about its origin and evolution. At present we can only claim insight into the internal structure of the Earth and Moon. Seismology has provided by far the most specific information on the interior of these two bodies. However, because no seismic data are available for Mars, it is necessary to look elsewhere for constraints on the Martian interior.

Using the available geophysical data, i.e. mean moment of inertia in combination with mean martian density, have been used to place constraints on the mantle density profile for an assumed core composition and size [1-3]. The study by [4] employed compositions of martian meteorites in combination with models of the planet's radial density distribution to infer the physical structure of the martian mantle. Other attempts, exemplified by [6], started off with a model of the martian mantle and core composition, which had been derived independently of any geophysical constraints [7], to experimentally determine modal mineralogy along a model pressure-temperature profile and then to use it to calculate a mantle density profile. This could then be used to constrain the size of the core. Concerning the latter, recent attempts at retrieving information on its state and size using the second degree tidal Love number, suggested it to be fluid [8,9]. However, the large observed value of  $k_2$  can also be interpreted as the mantle being softer than the assumed elastic solid model because of the presence of partial melt at depth [8,9].

**Purpose.** The aforementioned studies are essentially forward modeling approaches and therefore provide no information on the range of physical models that are actually consistent with the known geophysical parameters for Mars. In view of this limitation, we employ an inverse method as described in [10-13] to constrain martian composition and thermal state directly from geophysical observations. The method allows composition and temperature to be transformed directly to parameters such as mineralogy, Mg# ( $\text{MgO}/(\text{MgO}+\text{FeO}) \times 100$ ), bulk mantle and core physical properties, which are all fundamental to our understanding of Mars.

The data used in the inversion are, mean moment of inertia ( $I$ ), mean density ( $\bar{\rho}$ ), second degree tidal Love number ( $k_2$ ), tidal dissipation factor ( $\bar{Q}$ ) and of course mean radius ( $R$ ). The observed values of these parameters are, respectively,  $0.3635 \pm 0.0012$ ,  $3935 \pm 0.4 \text{ kg/m}^3$ ,  $0.163 \pm 0.017$ ,  $92 \pm 11$  and  $3389.5 \text{ km}$  [8, 9, 14].

**Method of Analysis.** Our model of Mars is assumed spherically symmetric and divided into three layers of variable thickness, corresponding to crust, mantle and core. Crust and mantle layers are parameterized using composition  $c$ , thickness  $d$  and temperature  $T$ , whereas the core is modeled using the parameters size and density. We model the chemical composition of Mars using the model system  $\text{CaO-FeO-MgO-Al}_2\text{O}_3\text{-SiO}_2$ . The core is physically described by its radius and density, because of a lack of thermodynamic data necessary for modeling

metallic phases. For a given model configuration that specifies mantle composition and thermal state, the inversion procedure consists of the following steps:

- Gibbs free energy minimization is used to compute equilibrium mineral modes at the pressure and temperature conditions of interest [15].
- Physical properties in the form of density,  $P$  and  $S$ -wave velocity are then estimated from the computed mineralogy as function of depth [15, 16].
- Mass, moment of inertia, second degree tidal Love number and global tidal dissipation are subsequently computed from the density,  $P$  and  $S$ -wave velocity profiles

The inverse problem is formulated in terms of probability density functions (*pdf*'s) and samples from the outcome which are characterized by the posterior *pdf*. These samples are obtained by employing a Markov Chain Monte Carlo method (MCMC) as described in [17, 18, 19].

**Results.** The resulting compositions for the martian mantle (figure 1), excepting FeO, are remarkably similar to the pyrolite model [20], which is thought to be representative of the Earth's mantle. In this regard, FeO is slightly enriched relative to pyrolite, a result that is consistent with inferences from the SNC meteorites [21]. From the obtained compositions we calculated mantle Mg#, which are close (sampled peak values are  $\sim 0.86\text{-}0.87$ ) to the accepted terrestrial value of 0.89 [20], but differ significantly from the value (0.75) obtained for the SNC derived bulk model of [7]. Our results therefore suggest a closer kinship of the material from which the Earth and Mars accreted than previously thought [22]. The sampled thermal profiles indicate temperatures at the core-mantle boundary (CMB) of 2350 K, significantly higher than previously suggested [4, 23]. The martian mineralogy (figure 3) is generally seen to be similar to that of the Earth's. The upper mantle is made up principally of olivine and clinopyroxene, with orthopyroxene and garnet making up between 20 and 30 vol% and spinel about 15 vol%. The major lower mantle phases are garnet and wadsleyite, with ringwoodite and magnesiowustite constituting smaller amounts. Other lower mantle minerals are akimotoite and ca-perovskite. There has been some discussion about whether physical conditions deep within Mars are sufficient to stabilise perovskite, the major terrestrial lower mantle mineral [e.g. 21]. Although not established, previous analyses [e.g. 4,6] generally assumed the presence of a perovskite layer just above the CMB. However, the results presented here suggest that perovskite is not stable in the martian mantle.

Figure 4 core properties and most probable core mantle boundary temperatures, radii and densities are found to be around 2100 °C,  $\sim 1540\text{-}1620 \text{ km}$ , and  $\sim 7.6 \text{ g/cm}^3$ , respectively. This is most consistent with a metallic composition containing  $\sim 12 \text{ wt}\% \text{S}$ , in agreement with earlier assessments

[e.g. 24]. Moreover, combining the CMB temperatures found here with experimental melting temperatures of possible core compositions [25], strongly suggests that the core of Mars is fluid or at least that its outer part is.

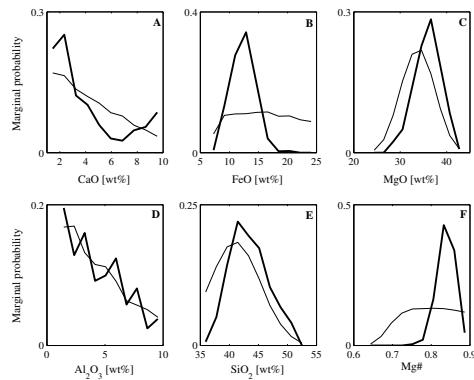


Figure 1: Sampled mantle compositions and Mg#. Thin lines denote prior probability density functions (*pdf*'s), while heavy lines depict the posterior *pdf*'s.

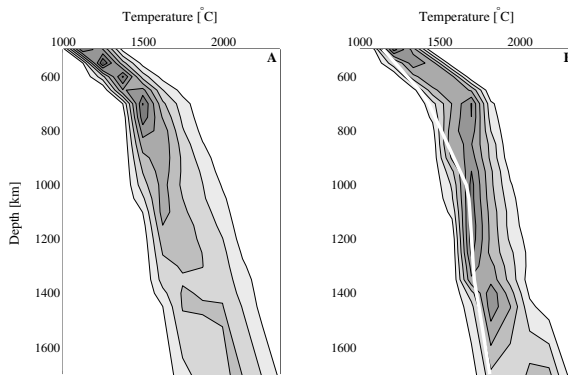


Figure 2: Sampled prior and posterior thermal profiles. The contours directly reflect the probability of occurrence of the temperature at a given depth, with white corresponding to least probable and black to most probable. White line denotes maximum likelihood model.

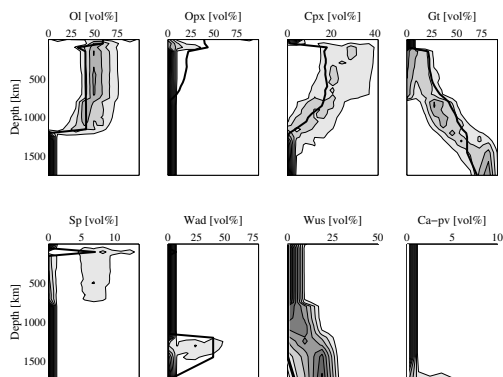


Figure 3: Mineralogy as a function of depth. All mineral abundances are in vol%. Shades of gray as in figure 2. White line denotes maximum likelihood model.

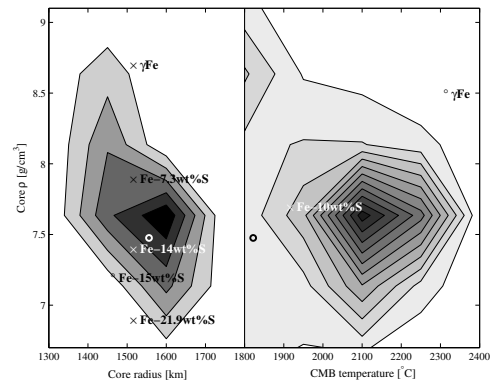


Figure 4: Summary of derived core properties. The left panel depicts a 2 dimensional marginal posterior *pdf* showing the correlation between core  $\rho$  and radius, whereas the right panel highlights the correlation between CMB temperatures and core  $\rho$ . Maximum likelihood values are indicated by the symbols (o). Experimental density measurements of FeS carried out at high pressures and temperatures are represented by (x) and (o) are experimental determinations of the melting relations in the Fe-FeS system. Note that the CMB temperature axis extends to the left of 1800 °C.

**References.** [1] J. Wood et al., in *Basaltic Volcanism on the Terrestrial Planets*, 634, 1981. [2] B. Bills, *J. Geophys. Res.*, 15, 14131, 1990. [3] W. Folkner et al., *Science*, 278, 1749, 1997. [4] F. Sohl & T. Spohn, *J. Geophys. Res.*, 102, 1613, 1998. [5] D. Stevenson, *Nature*, 412, 214, 2001. [6] C. Bertka & Y. Fei, *J. Geophys. Res.*, 102, 5251, 1997. [7] G. Dreibus & H. Wänke, *Meteoritics*, 20, 367, 1985. [8] C. Yoder et al., *Science*, 300, 299, 2003. [9] B. Bills et al., *J. Geophys. Res.*, 110, E07004, 2005. [10] A. Khan et al., *Geophys. J. Int.*, in press, 2006. [11] A. Khan et al., *J. Geophys. Res.*, 111, B10102, 2006. [12] A. Khan et al., *J. Geophys. Res.*, 111, E05005, 2006. [13] A. Khan et al., *Earth Planet. Sci. Lett.*, 248, 2006. [14] F. Sohl et al., *J. Geophys. Res.*, in press, 2005. [15] J. Connolly, *Am. J. Sci.*, 666, 290, 1990. [16] J. Connolly & D. Kerrick, *Earth Planet. Sci. Lett.*, 204, 61, 2002. [17] A. Tarantola & B. Valette, *J. Geophys.*, 50, 159, 1982. [18] A. Tarantola, *Inverse Problem Theory*, Elsevier, Amsterdam, 1987. [19] K. Mosegaard & A. Tarantola, *J. Geophys. Res.*, 100, 12431, 1995. [20] W. McDonough & S. S. Sun, *Chem. Geol.*, 120, (3-4), 223, 1995. [21] J. Longhi et al., *Mars*, 184, Univ. of Ariz. Press, Tucson, 1992. [22] G. Wetherill, *Geochim. Cosmochim. Acta*, 58, 4513, 1994. [23] J. P. Williams & F. Nimmo, *Geology*, 32, 97, 2004. [24] C. Bertka & Y. Fei, *Science*, 281, 1838, 1998. [25] A. Kavner et al., *Earth Planet. Sci. Lett.*, 185, 25, 2001.