3. Constructing models of Mars' internal structure

The method that we use to construct interior-structure models is based on our previous work Khan and Connolly (2008) and the work of Nimmo and Faul (2013). For brevity, only a cursory description is presented here. We rely on a unified description of the elasticity and phase equilibria of multicomponent, multiphase assemblages from which mineralogical and seismic wave velocity models as functions of pressure (depth) and temperature are constructed. Specifically, we use the free-energy minimization strategy described by Connolly (2009) to predict rock mineralogy, elastic moduli, and density along self-consistently computed mantle adiabats for a given bulk composition. For this purpose we employ the thermodynamic formulation of Stixrude and Lithgow-Bertelloni (2005a) with parameters as in Stixrude and Lithgow-Bertelloni (2011). Bulk rock elastic moduli are estimated by Voigt-Reuss-Hill (VRH) averaging. The pressure profile is obtained by integrating the load from the surface. Possible mantle compositions are explored within the Na_2O-CaO-FeO-MgO-Al_2O_3-SiO_2 (NCFMAS) system, which accounts for more than 98% of the mass of the mantle of the experimental Martian model of Bertka and Fei (1997).

Estimates for the Martian mantle composition derive from geochemical studies (e.g., Dreibus and Wänke, 1985; Treiman, 1986; McSween, 1994; Taylor, 2013) of a set of basaltic achondrite meteorites, collectively designated the SNC's (Shergotty, Nakhla, and Chassigny), that are thought to have originated from Mars. Based on the analysis of Dreibus and Wänke, the Martian mantle contains about 17 wt% FeO compared to Earth's upper mantle budget of 8 wt% (e.g., McDonough and s. Sun, 1995; Lyubetskaya and Korenaga, 2007). This implies a Martian mantle Mg# of 75 (100×molar Mg/Mg+Fe), in comparison to the magnesian-rich terrestrial upper mantle value of ∼90. There is little information that bears directly on the thermal state of the Martian mantle as a result of which the areotherm has proved more difficult to constrain (e.g., Verhoeven et al., 2005; Khan and Connolly, 2008). For the purposes of the computations here, we rely on the "hot" geotherm of Verhoeven et al. (2005) (figure 3).

For crustal structure, we rely on a physical parameterization, i.e., P- and S-wave speed, density, and Moho depth as model parameters, rather than thermo-chemical parameters employed in modeling properties. Average crustal thickness is taken from the study of Wieczorek and Zuber (2004) and density, P- and S-wave speed are modeled as increasing linearly from 2 to 3 g/cm^3, 4 to 6.5 km/s, and 2 km/s to 3.5 km/s, respectively, between surface and the base of the crust.

As seismic waves propagate in the interior of Mars they are expected to be attenuated with distance much as on Earth. This is a manifestation of an anelastic medium. Another property of a dissipative medium is dispersion, which manifests itself in seismic waves of different frequencies traveling at different speeds. As a consequence, the elastic moduli become complex and frequency-dependent, which provides an appropriate start for the description of viscoelastic dissipation (e.g., Anderson, 1989).

The dissipation model adopted here (for details we refer the reader to Nimmo and Faul (2013)) is based on laboratory experiments of torsional forced oscillation data on melt-free polycrystalline olivine and is described in detail in Jackson and Faul (2010). In the absence of melting, dissipation has been observed in the Earth, Moon, and Mars to follow a frequency-dependence of the form $1/Q \sim \omega^{-\alpha}$, where ω is angular frequency and α is a constant (e.g., Lognonné and Mosser, 1993; Williams et al., 2001; Benjamin et al., 2006; Efroimsky, 2012). α has been determined from seismic and geodetic studies to lie in the range 0.1–0.4 (e.g., Minster and Anderson, 1981; Benjamin et al., 2006). The failure of Maxwellian viscoelasticity to reproduce this frequency-dependence has led to other rheological models such as the Burgers model (e.g., Jackson and Faul, 2010). The Burgers model of Jackson and Faul (2010) is preferred over other rheological models because of its ability to describe the transition from (anharmonic) elasticity to grainsize-sensitive viscoelastic behaviour as a means of explaining the observed dissipation in the forced torsional oscillation experiments on olivine.

For present purposes, computations were conducted employing a single shear-wave attenuation (Q) model at seismic frequencies (1 s) and a grain-size of 1 cm in accordance with Nimmo and Faul (2013). For the Martian crust and lithosphere, we fixed shear-wave Q to 600 after PREM (Dziewonski and Anderson, 1981) and to 100 in the core. Dissipation in bulk is neglected and we assume Q_κ=10^4 in line with terrestrial applications (e.g., Durek and Ekström, 1996). Anelastic P- and S-wave speeds ($\mathrm{V_{P/S}}$) as a function of pressure (p), temperature (T), composition (c), and frequency (ω) are estimated from the expressions for the visco-elastically computed temperature-, pressure-, and frequency-dependent moduli (further details may be found in Nimmo and Faul (2013)).

The physical properties (isotropic anelastic P- and S-wave speeds, density, and attenuation) so computed are shown in figure 3 to a depth of 1700 km.

For comparison with sampled seismic wave-speed and density profiles, we are also showing a set of models (DWref and T13) that are constructed in the same manner as the "Input" model, but using a different areotherm to highlight the influence of mantle thermal structure on physical properties. Model DWref is based on the bulk mantle composition of Dreibus and Wänke (1985) and the "Hot" areotherm of Verhoeven et al. (2005) (figure 3), whereas model "T13" is computed using the bulk mantle composition of Taylor (2013) and the self-consistently computed adiabat ("Adiabat" in figure 3).

These profiles contain prominent features above 400 km and around 1000–1100 km depth. The wave-speed decrease above 400 km depth (DWref only) is due to the steep increase in temperature in the lithosphere that results in a strong low-velocity zone (LVZ) (e.g., Nimmo and Faul, 2013; Zheng et al., 2015). The LVZ zone is not present in the other two models because of a smoother transition between the conductive lithosphere and the mantle adiabat. As shown elsewhere (e.g., Bertka and Fei, 1998; Khan and Connolly, 2008), the discontinuity at $\sim$1100 km depth is linked to the mineral phase transformation olivine$\rightarrow$wadsleyite (see also Mocquet et al. (1996) and Verhoeven et al. (2005)), which in Earth is responsible for the "410-km" seismic discontinuity. The associated shear-wave attenuation structure is also shown in figure 3. In the case of "DWref", the Q-structure is based on PREM, whereas for "Input" and "T13", we use the viscoelastic approach described above. Generally, shear-wave attenuation structure is observed to be fairly constant throughout most of the mantle in overall agreement with expectations based on PREM and existing Martian models (e.g., Lognonné and Mosser, 1993; Zharkov and Gudkova, 1997; Nimmo and Faul, 2013). Theoretical predictions for the attenuation in the Martian mantle have been discussed by Lognonné and Mosser (1993) and Zharkov and Gudkova (1997). Based on a Q value of 50–150 at the tidal period of Phobos (5h 32 min) and assuming the absorption band model of Anderson and Given (1982) to hold over the entire frequency range (seismic to tidal), these studies find Q values in the range ∼150–400 (at 1 s). For comparison, current estimates of Q at the period of Phobos are 80–105 (e.g., Bills et al., 2005; Lainey et al., 2007; Nimmo and Faul, 2013).