Water, Melting, and Convection in the Martian Mantle

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Model design

We present models of mantle convection focused on the situation in early Mars in which particular attention is paid to water and its effects on partial melting, melt extraction, and viscosity. We have combined the mantle convection program STAGYY [7] with a parameterized thermodynamic model of martian mantle mineralogy [5] and carried out calculations of convection in a two-dimensional compressible model of the planet's mantle. As the depth to the base of the mantle is known only to within a few hundred kilometers, models with core–mantle boundary depths of 1700 km (LC) and 2000 km (SC), which essentially bracket the range of possible values, are considered. The models are heated from below by a cooling core and from within by the radioactive decay of ^{40}K, ^{232}Th, ^{235}U, and ^{238}U. At the top, a rigid boundary with a temperature corresponding to the average surface on Mars is imposed as a boundary condition. The model domain is a full spherical annulus [2].

Near-fractional hydrous melting is included in a parameterized form by modifying the dry solidus of martian peridotite [1, 6] by means of a simple method to include the solidus-lowering effect of water in low concentrations [3]; the exhaustion of phases is also taken into account in a simple form. Melt deeper than the density crossover depth is retained. Melting reduces the concentration and changes the distribution of both the water and the heat-producing radionuclides in the solid mantle. Different initial water contents are assumed, and water content is allowed to evolve with time as a consequence of melting and dehydration; the content of the trace components is tracked with tracer particles, and half of the water carried with erupting melts is removed from the model permanently by outgassing. The viscosity of the mantle is dependent on temperature, pressure, water content, the amount of retained melt, and the presence of high-pressure polymorphs of olivine or of basalt/eclogite and therefore also undergoes a secular change as the mantle cools and becomes depleted in some parts and accumulates basaltic crustal material in others.

Several aspects of Mars' internal structure are not well known, so that we have to probe the effects of several model parameters within a certain range; among these, core size, bulk Mg#, initial radionuclide and water content and distribution, and initial potential temperature are particularly important. In this presentation, we show a subset of this survey that centers on two water-free (D) models and two with an initially homogeneous water content of 200 wppm (W), all with a bulk Mg# of 0.75; radionuclide contents [8] are the same in all four models, and are also initially homogeneous. In either pair, one model has a large core and the other has a small one, so that in the latter case, a dense layer of perovskite (pv)+ferropericlase (fp) exists at the bottom.

Results

As in the models without melting of *Ruedas and Tackley* [5], these models with melting show strong, if not always as long-lived, layering. Comparison with models where the dynamical effects of phase transitions have been suppressed while maintaining the density structure and mineralogy show no such layering and confirm the importance of the thermodynamical effects of phase transitions in establishing layered convection, although an increase of viscosity due to the appearance of high-pressure polymorphs is found to promote layering as well in some circumstances. The lower layer is largely isolated for some time, heats up due to radioactive decay, and melts

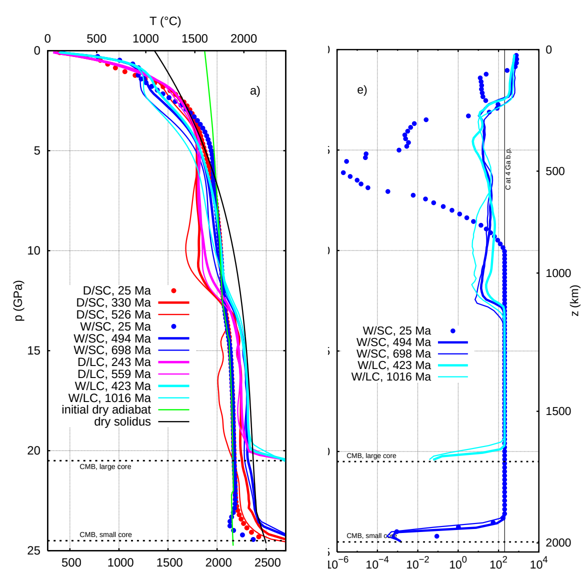

Figure 1: Horizontally averaged depth profiles of temperature (left) and water concentration (right) in some dry (D) and water-bearing (W) models with large (LC) and small (SC) core.

only marginally, if at all, whereas the upper layer loses its trace components by melting and is cooled significantly by remixing of downwelling lithospheric material; a basaltic crust enriched in radionuclides and water forms at the top. In the models with a small core, there is an additional basal layer of pv+fp, which largely coincides with the thermal boundary layer and is also largely isolated from the upper ones.

In the dry model with a large core (D/LC), the lower layer is in contact with the core, whereas in the model with a small core (D/SC), the lower layer is separated from the core by a thin basal layer corresponding to the pv+fp stability field. Several tiny plumes form in this basal layer but stall at the endothermic phase boundary, so that the large lower layer is essentially plume-free; this seems to have a destabilizing effect, as eventually several large, cold downwellings from the upper layer penetrate it and cool it substantially, sweeping together the small plumes in the base layer into one patch. Depending on the viscosity structure, a similar configuration can arise in the D/LC model, with the mini-plumes being confined to the ringwoodite stability field, or else some thin plumes develop in the D/LC model that rise through the lower layer. In large-core models, cold downwellings seem to be weaker and more episodic.

Models with a certain amount of water also show layering with a heating lower layer, but nonetheless develop differently; no fundamental difference was observed between models with 100 and with 200 wppm water. The upper layer cools less quickly, because the dehydration by melting (cf. fig. 1, right) increases the viscosity of a major part of it strongly (fig. 2, lower half of lower panel). A small number of very thick and sluggish thermal upwellings develops from the olivine–olivine+ringwoodite transition. This kind of model seems to hold more promise to explain the thick, stiff lithosphere inferred from SHARAD observations [4] and the existence of a few, long-lived volcanic centers.

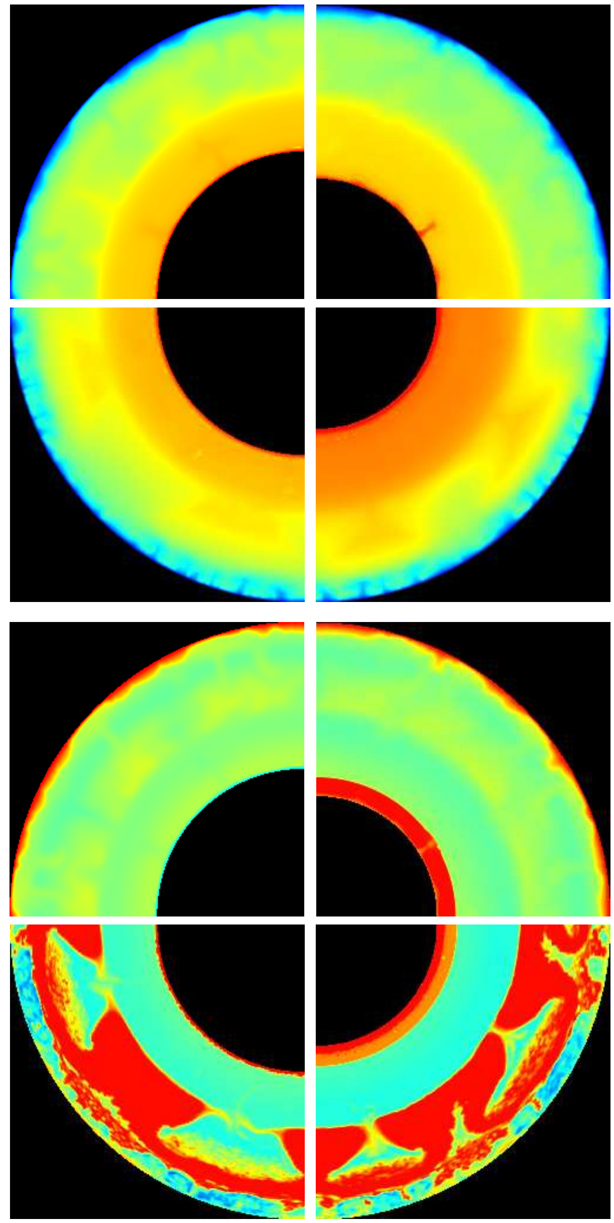

Figure 2: Temperature (upper panel) and decadic logarithm of viscosity in Pa s (lower panel) at times between ca. 240 and 500 Ma, in one quarter of the full model domain. Upper halves of panels show dry, lower halves wet models. T ranges from 217 to ca. 2900±100 K, η from $2 \cdot 10^{20}$ to $4 \cdot 10^{22}$ Pa s in the dry and ca. 10^{19} to 10^{24} Pa s in the wet models.

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