GIANT IMPACTS AND THERMOCHEMICAL MARTIAN MANTLE CONVECTION: IMPLICATIONS FOR THARSIS. C. C. Reese, *Univ. of Minnesota, Morris, Morris MN 56267 (reesec@mrs.umn.edu)*, V. S. Solomatov, *New Mexico State Univ., Las Cruces NM 88003*, J. R. Baumgardner, *Los Alamos Natl. Lab., Los Alamos NM 87545*, D. R. Stegman, *Univ. of California, Berkeley, Berkeley CA 94720*.

Introduction. The initial thermal and compositional state of Mars was likely heterogeneous as a result of giant (Moon-sized) impacts during late stage planet formation which can locally shock heat the deep interior to supersolidus temperatures. The subsequent evolution of Mars is governed by the nature of mantle convection. Early thermochemical mantle heterogeneity is efficiently removed only if convection is sufficiently vigorous. If convection is sluggish, early heterogeneity persists throughout planetary history.

The fact that Mars has only one large volcanic complex, Tharsis, is an important constraint on Martian evolution. Stagnant lid scaling relationships suggest that for subsolidus Martian mantle temperatures, thermal convection is sluggish and mantle heat transfer and homogenization inefficient (Solomatov and Moresi, 2000). In this case, mantle heating during early evolution makes early plume formation and Tharsis development difficult. An alternative mechanism for Tharsis formation with the potential for early and long-lived magma generation is an impact induced thermal anomaly preserved because of sluggish convection (Reese et al., 2002). These models, however, did not address compositional variations resulting from impact heating, mantle melting, and crust formation.

Model. *Mantle viscosity.* Newtonian viscosity is considered and the Frank-Kamenetskii approximation is adopted, $\eta = be^{-\gamma T}$, where b and γ are constant. A rigid upper surface and $\gamma = 4 \times 10^{-3}$ ensures stagnant lid convection. For the preliminary results shown here, the interior viscosity is $\eta \sim 10^{23}$ Pa s.

Initial conditions. The nominal initial mantle temperature is assumed to be 1500 K with a fixed surface temperature of 220 K. Due to numerical resolution restrictions a cold boundary layer with thickness $\sim$ 500 km is considered. If the energy of core formation is dissipated thermally in the iron, the core can initially be $\sim$ few hundred degrees hotter than the mantle (Stevenson, 2001). A fixed core-mantle boundary (CMB) temperature of 1800 K is assumed. Of course, the CMB temperature should evolve with time in order to satisfy the core energy budget (Stevenson et al., 1983).

Internal heating. Parameterized convection models suggest early, widespread mantle melting. and possible extensive differentiation of heat producing elements into the crust (Reese et al., 2002). Geochemical analysis also suggests early separation of a radiogenic isotope enriched crust (McLennan, 2001). An end-member model of complete mantle differentiation is considered, i.e., there is no internal heating.

Impact heating Giant impacts produce a shock wave which generates an intact volume of melt (Tonks and Melosh, 1993). Subsequent isostatic adjustment and iron/silicate segregation can produce further localized heating. Concurrent with these processes is crystallization of the melt region and formation of a stable crust (Solomatov, 2000). While interaction of these mechanisms undoubtedly results in complicated early planetary dynamics, it seems likely that the latest, largest impacts produce mantle regions that are hot and depleted in incompatible elements.

A simple paramterization of impact heating, melting, and differentiation is considered. If the release adiabat is a Hugoniot, the radial distance R heated by a temperature ΔT can be calculated for a vertically incident impactor of radius a (Melosh, 1989). Temperatures near the impact site are limited by melting and vaporization while the heating drops off rapidly in amplitude away from the impact site. For the region with $\Delta T \geq 300$ K a constant temperature increase of 300 K is assumed. The same regions is assumed to be differentiated into depleted mantle and basaltic crust (although the crust is not modelled). The density difference between depleted and undepleted mantle is 2 %. The impact inventory includes Hellas, Utopia, Isidis and a large hypothetical impact in the western hemisphere Tharsis region.

Preliminary results. Figure 1 shows the evolution of the temperature and composition fields along a cross section through 90° W - 270° W. The large compositional buoyancy associated with the hypothetical Tharsis impact results in a mantle upwelling. This brings hot material close to the surface where decompression melting occurs accompanied by surface volcanism. Depleted mantle material spreads out along the bottom of the viscous lid. The compositional upwelling also focuses the hot thermal bounday layer at the CMB into a large thermal plume beneath the western hemisphere which survives throughout Martian evolution.

Discussion. These results suggest that impact induced compositional differentiation may play an important role in Martian evolution. In particular, regions subject to giant impact melting might be compositionally buoyant. The associated mantle upwelling can focus magmatism and induce a stable thermal plume if there is a hot boundary layer at the CMB. The model can be modified to include a parameterized core thermal evolution and compositional effects of decompression melting. Finally, it should be pointed out that the pre-impact spherical symmetry is a zero order approximation. Mantle temperature and composition is affected by earlier impacts, core formation, and previous mantle dynamics. Generation of the initial condition should be treated in a statistical fashion by specifying a plausible inventory of impactors.

References. Melosh, *Impact cratering: A geolocial process*, Oxford Univ. Press, 1989; McLennan, Geophys. Res. Lett., **28**, 4019., 2001; Reese et al., J. Geophys. Res., **107**, 10.1029/2000JE001474, 2002; Solomatov and Moresi, J. Geophys. Res., **105**, 21795, 2000; Solomatov, in Origin of the Earth and Moon, Univ. Ariz. Press, 2000; Stevenson et al., Icarus, **54**, 466, 1983; Stevenson, Nature, **412**, 214, 2001; Tonks and Melosh, Icarus, **100**, 326, 1993

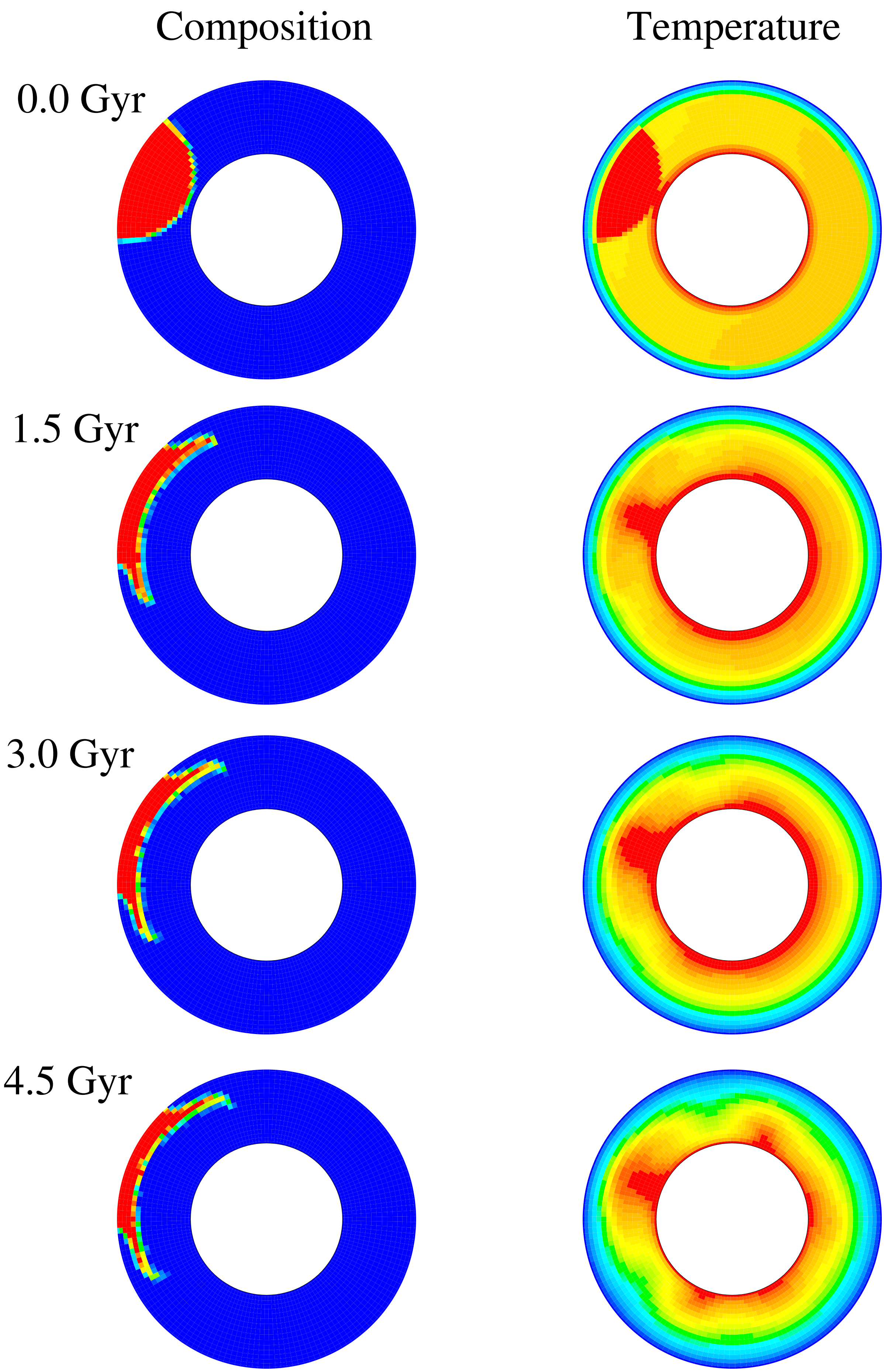

Figure 1: Evolution of the temperature and comosition for impact initial conditions. The maximum temperature (red) is 1800 K, the minimum (blue) 220 K. For the composition field, blue is depleted mantle while red is undepleted.