CONSTRAINTS ON THE DEPTH AND TEMPERATURE OF MELTING IN THE MARTIAN MANTLE.

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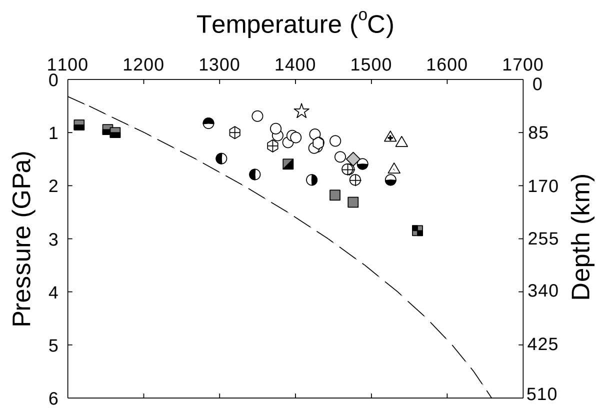
Introduction: Investigations of the Martian crust by rovers, and analyses of recent meteorite falls, have expanded our knowledge of the composition and age of Martian rocks. The rocks examined on the surface at Gale Crater, Gusev Crater, and Merdinani Planum are older and have significantly different chemistry than the younger shergottites [1-3]. The ancient basaltic regolith breccia, NWA 7034 (and pairs), provides a potential link between the martian meteorites and the surface investigations, but each clast with the breccias has a unique mineralogy and composition expanding our knowledge of Martian rocks [4-6]. Further, investigations of the surface chemistry (from orbit) have shown distinct difference in the bulk chemistry of volcanic providences with different ages [7]. The difference in age and chemistry between the surface rocks, the clasts in NWA 7034, and the shergottites may reflect different conditions of melting within the martian interior. Therefore, we can combine estimates for basalt formation for rocks from the Noachian (Gale Crater, Gusev Crater, Bounce Rock in Meridiani Planum and a clast in NWA 7034), Hesperian (surface volcanics), and Amazonian (surface volcanics and shergottites).

Here, we build upon our previous models of basalt formation based on rocks analyzed in Gusev Crater and Meridiani Planum [8, 9]. We compare our new results on rocks from Gale Crater and a clast in meteorite NWA 7034 [10] to the similarly ancient rocks from Gusev Crater and Meridiani Planum to calculate an average mantle potential temperature for ancient Mars. We then compare the results for Noachian Mars to estimates for formation conditions of younger rocks (Hesperian and Amazonian surface regions; [7] and Amazonian Martian meteorites; [11-13]) to investigate how the mantle potential temperature of Mars may have changed through time.

Samples and Estimates of Melting Conditions: *Meteorites:* Estimates for melting conditions for the shergottite source region comes from inverse experimental results and modeling for Yamato 980459 (Y98) [11], NWA 5789 [13], and NWA 1068 [12]. Y98 and NWA 5789 are thought to be 472 Ma, have calculated conditions of formation of 1.2 GPa and 1540 °C and 1.1 GPa and 1525 °C respectively, and a mantle potential temperature (T_P) of 1550 °C. NWA 1068 is 185 Ma, has a *P-T* of formation of 1.7 GPa and 1550 °C, and a T_P of 1480 °C.



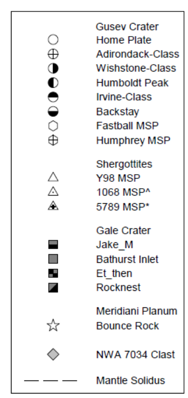


Figure 1. Temperatures and pressures calculated for primitive magma compositions estimated for Gale Crater rocks (gray squares) and a clast in NWA 7034 (gray diamond) compared with basalts from Gusev Crater (white circles), Bounce rock in Meridiani Planum (white star), and ol-phyric shergottites (white triangles). Experimental estimates for the Gusev Crater rocks Humphrey and Fastball (white hexagons) are shown as an internal calibration between experimental and modeling results [9, 14, 15]. Also shown for reference (dashed line) is the solidus of nominally volatile-free Martian mantle [8].

Gusev Crater: Rocks in Gusev crater are thought to be 3.65 Ga based on crater counting [16]. Experimental and modeling results for melting conditions in their source region are: 1.1 GPa and 1410 °C for Barnhill-class basalts and ~2 GPa and 1490 °C for Adirondack-class basalts, with an average T_P of 1445 ± 85 °C [8].

Meridiani Planum: The only rock analyzed in Merdinai Planum that has a calculated melting condition is Bounce Rock. Bounce is thought to be Noachian in age [17], has a calculated melting condition of 0.6 GPa and 1410 °C, and a T_P of 1475 ± 15 °C [8].

Gale Crater: Our new results for Gale Crater come mainly from the Bradbury Rise, and specifically Bathhurst Inlet. Our calculated melting conditions are 1.6 to 2.8 GPa and 1490 to 1560 °C, giving a T_P of 1450 ± 30 °C [10].

NWA 7034: Our estimate for a vitrophyric clast in NWA 7034 [6] yields P-T of 1.2 GPa and 1419 °C and a T_P of 1430 ± 15 °C [10]. The bulk rock is 4.4 Ga [5].

Mantle Potential Temperature Comparison: The rocks in Gale Crater, Gusev Crater, Bounce rock, and the clast in NWA 7034 are all thought to be Noachian in age based on geomorphology, crater counting, the single age date for Gale Crater, and age dating of NWA 7533 which is paired with NWA 7034 [5, 16-18], and give us the same range in T_P [8, 9]. This suggests that the calculated average T_P (1450 ± 70 °C) may represent an average global mantle temperature during the Noachian.

Mantle Potential Temperature Through Time: Since we now have estimates for T_P from the Noachian through the Amazonian (Fig. 2), we can investigate how the mantle temperature may have changed through time and investigate if there were changes in the style of melting during these Martian eons. The results from NWA 7034, Gale Crater, Gusev Crater, and Meridiani Planum are consistent with an average Noachian T_P of 1450 ± 70 °C. The T_P estimates for the Hesperian and Amazonian are based on orbital analyses of the crust and are lower in temperature than the estimates for the Noachian (1400 ± 40 °C and 1370 ± 30 °C for surface flows) [7]. This is consistent with simple convective cooling of the interior of Mars and may reflect the fact that that the conditions of mantle melting capture secular cooling of the whole mantle [7, 8, 19].

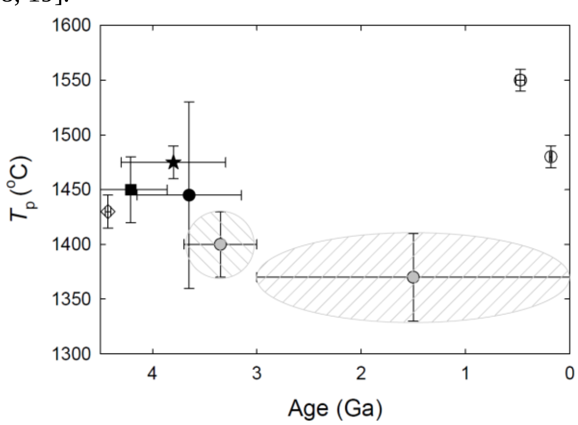


Figure 2. T_P versus age for surface rocks from Gale Crater (black square) [10], the clast in NWA 7034 (white diamond) [10], Gusev Crater (black circle) [8, 9], and Bounce rock in Meridiani (black star) [8] compared with similar estimates for surface volcanics (gray circles) [7] and ol-phyric shergottites (white circles) [11-13].

However, estimates for the SNC meteorites [11-13], which are for the youngest rocks [20, 21], are significantly higher than those for equally young rocks viewed from orbit [7] and are similar to our estimates for rocks from the Noachian. This could suggest that the Martian mantle is actually heating up, as has recently been suggested based on paleo heat flow estimates [22]; however, perhaps a more realistic notion is that the Martian meteorites represent products of localized hot spot magmatism. Estimates for the T_P of terrestrial intraplate ocean island basalts on Earth are often as much as 100-200 °C hotter than those for mid-ocean ridge basalts [23-26]. Assuming that this 100-200 °C temperature difference is appropriate for background mantle melting conditions versus thermal-boundary layer-driven active upwelling and melting scenarios in the Martian interior, the temperature estimates for the Martian meteorites compared with surface rocks is consistent with the former being representative of localized melting product.

Conclusions: Estimates for the average global mantle temperature during the Noachian based on analyses in Gale Crater, Gusev Crater, Bounce Rock in Meridiani Planum, and a clast in NWA 7034 are 1450 ± 70 °C. The T_P estimates for the Hesperian and Amazonian, based on orbital analyses of the crust [7], are lower in temperature than the estimates for the Noachian. Hotter temperatures derived from young shergottites suggest the possible presence of localized thermal anomalies. Our analyses are consistent with simple convective cooling of the Martian interior.

References: [1] Schmidt M.E. et al. (2014) *JGR-planets*, 119, doi:2013JE004481. [2] McSween H.Y. et al. (2009) *Science*, 324, 736-739. [3] McSween H.Y. et al. (2003) *JGR-Planets*, 108, (E12). [4]. Agee C.B. et al. (2013) *Science*, 339, 780-785. [5] Humayun M. et al. (2013) *Nature*, 503, 513-516. [6] Udry A. et al. (2014) *GCA*, 141, 281-293. [7] Baratoux D. et al. (2011) *Nature*, 472, 338–341. [8] Filiberto J. and Dasgupta R. (2011) *EPSL*, 304, 527-537. [9] Filiberto J. et al. (2010) *GRL*, 37, doi:10.1029/2010GL043999. [10] Filiberto J. and Dasgupta R. (2015) *JGR-planets*. DOI: 10.1002/2014JE004745. [11] Musselwhite D.S. et al. (2006) *MaPS*, 41, 1271-1290. [12] Filiberto J. et al. (2010) *MaPS*, 45, 1258-1270. [13] Gross J. et al. (2011) *MaPS*, 46, 116-133. [14] Monders A.G. et al. (2007) *MaPS*, 42, 131-148. [15] Filiberto J. et al. (2008) *MaPS*, 43, 1137-1146. [16] Greeley R. et al. (2005) *JGR-planets*, 110, doi:10.1029/2005JE002401. [17] Zipfel J. et al. (2011) *MaPS*, 46, 1-20. [18] Farley K.A. et al. (2014) *Science*, 343, 6169. [19] Breuer D. and Spohn T. (2006) *Planet.& Spac. Sci.*, 54, 153-169. [20] Jones J.H. (1986) *GCA*, 50, 969-977. [21] Moser D. et al. (2013) *Nature*, 499, 454-457. [22] Ruiz J. (2014) *Sci. Rep*, 4. [23] Putirka K.D. (2005) *G-cubed*, 6,. 10.1029/2005gc000915. [24] Putirka K.D. et al. (2007) *Chem Geo*, 241, 177-206. [25] Courtier A.M. et al. (2007) *EPSL*, 264, 308-316. [26] Herzberg C. et al. (2010) *EPSL*, 292, 79-88.