

PETROGRAPHIC CONSTRAINTS ON SHOCK INDUCED P/T CONDITIONS IN SHERGOTTITES. Jörg Fritz and Ansgar Greshake, Museum für Naturkunde an der Humboldt-Universität zu Berlin, Invalidenstraße 43, 10115 Berlin, Germany (e-mail: joerg.fritz.1@gmx.net.)

Introduction: Martian meteorites (MM) comprise rocks shocked in the range of 5-55 GPa during natural impact events [1]. Together with the suite of experimentally shocked rocks, they are valuable assets for the study of the progressive stages of shock metamorphism. Petrographic investigations of MM, in combination with shock and post shock temperature calculations are presented. This data set is compared with the shock pressures determined by applying the classification schema, which is based on shock recovery experiments and observations in naturally shocked chondritic and terrestrial rocks [1-4].

Method: The approximation (about $\pm 30\%$) of the shock induced temperature increase in different rock types follows the method described in [1, 5]. The calculations are based on the linear relation of the particle velocity U to shock wave velocity D : $D = c + s * U$, with c and s being experimentally determined constants [7-8]. The amount of energy deposited in the decompressed rock was calculated and then divided by the specific heat capacity for mafic rocks, C_p , approximated with 1000 J/kg/K.

Results: Shock temperature increase (ΔT_s) describes the temperature increase during maximum compression, and post shock temperature increase (ΔT_{ps}) describes the temperature increase after passage of the shock wave. ΔT_{ps} for plagioclase dominated (anorthosite), olivine dominated (dunite) and pyroxene dominated (pyroxenite) rocks shocked to 0-55 GPa are shown in Fig. 1. ΔT_{ps} rises with shock pressure to a similar degree in olivine and pyroxene, but a steeper increase is observed for plagioclase shocked above 30-35 GPa. This is due to the fact that plagioclase undergoes substantial phase transitions from plagioclase to maskelynite and melt, while the crystal structure of olivine and pyroxene remains unchanged in this pressure range. For gabbro the ΔT_{ps} is intermediate and depends on the proportions of olivine, pyroxene, and plagioclase in the rock sample.

The temperature-time path for a gabbro-like MM (i.e. shergottites) shocked to ~ 45 GPa is shown in Fig. 2. At the shock front the temperature discontinuously jumps from the initial temperature of the rock ($T_{\text{surface-Mars}}$) to T_s of the individual mineral phases (t_1). MM are accelerated by the steep pressure gradient in the "spallation zone" building up by interactions of the shock wave emerging spherically from the impact site and the release wave reflected from the "free surface" of the planet. Thus, the pulse lengths in MM depend on

their burial depth below the Martian surface and the size of the impacting projectile [5-6]. Mineralogical arguments support a pulse length of $\sim 10^{-2}$ s for Zagami [9]. After adiabatic release the temperatures drop to the T_{ps} ($\Delta T_{ps} + 233^\circ\text{K}$ initial temperature) of the individual mineral phases (t_2). The thermal equilibrium between mm sized grains is assumed to be on the order of seconds (t_3). A shock heated 40 cm \varnothing MM takes ~ 0.5 h for cooling to ambient temperature in space [1].

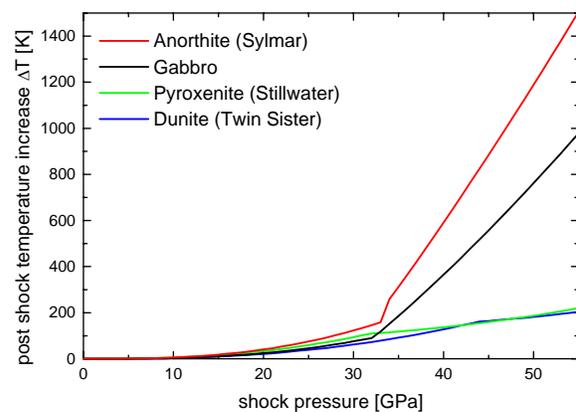


Fig. 1. Calculated post shock temperature increase (ΔT_{ps}).

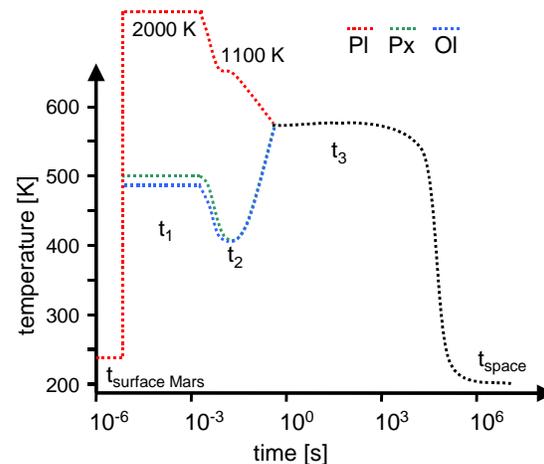


Fig. 2. Temperature-time path of MM; PI = plagioclase, Px = pyroxene, Ol = olivine; t_{1-4} see text; $T_{\text{surface Mars}} = 233^\circ\text{K}$.

Certain petrologic observations in MM can be explained by considering that for a given shock pressure, different ΔT_{ps} are produced in plagioclase compared to pyroxene and olivine. First, during crystallization of the Martian rocks Shergotty and Los Angeles thin

pyroxene lamellas in plagioclase were formed. Later, the plagioclase in both meteorites was transformed to maskelynite by shock. In Shergotty, shocked to ~32 GPa [1], the thin pyroxene lamellas are perfectly preserved (Fig. 3). In Los Angeles, shocked to ~45 GPa [1], the thin pyroxene lamellas show signs of melting and their surface is decorated with vesicles (Fig. 4) although plagioclase was not shock melted. The melting temperatures at $P=1$ bar are 1730°K for plagioclase (An_{50}), 1630°K for pyroxene (diopside) and eutectic melting of the two phases starts at ~1530°K [10]. Eutectic melting could occur in regions where the cold pyroxene has a high surface/volume ratio to the hot enclosing plagioclase. For plagioclase, the required T_{ps} of 1530 K was calculated for shock pressures of ~50 GPa (Figs. 1, 4).

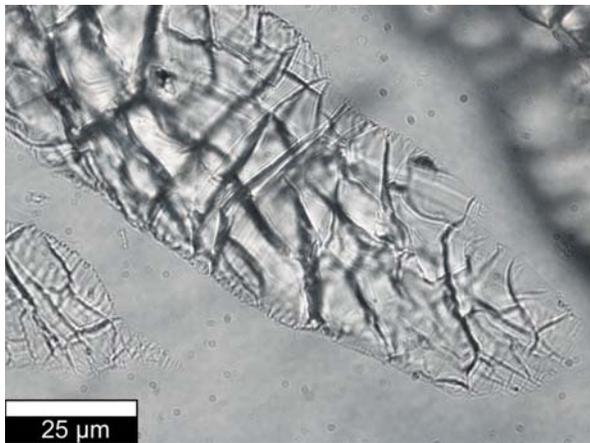


Fig. 3. Transmitted light; thin pyroxene lamella enclosed by maskelynite in the moderately shocked Shergotty meteorite.

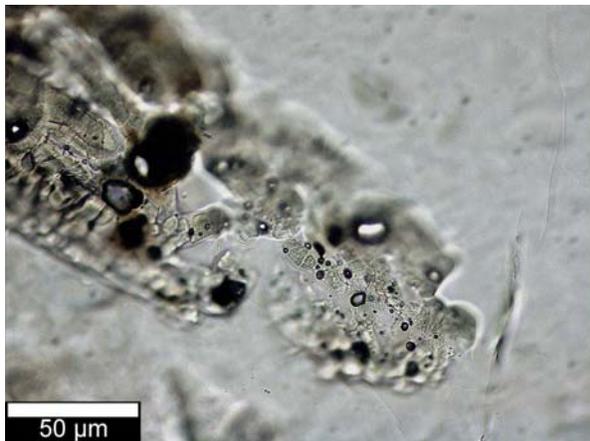


Fig. 4. Transmitted light; thin pyroxene lamella enclosed by maskelynite in the highly shocked Los Angeles meteorite.

Second, in ALH77005 (~55 GPa [1]) vesiculated plagioclase glass formed. This indicates $T_{ps} > 1730^{\circ}\text{K}$, as vesicles form only after pressure release. This tem-

peratures in the plagioclase are sufficient to melt the adjacent pyroxene (~1630°K), but not the adjacent olivine (~1800°K for Fe_{70} [10]). Resulting chemical heterogeneities at the contact of plagioclase to pyroxene may explain that the small rims of crystallized plagioclase are mostly restricted to the boundaries with olivine and not with pyroxene (Fig 5). A thermal equilibrium (gabbro ~55 GPa; Fig. 1) between the very high temperatures in plagioclase and the cooler olivine and pyroxene was reached quickly. The equilibration temperature in the whole rock prevailed long enough to allow partial crystallization of the plagioclase melt.

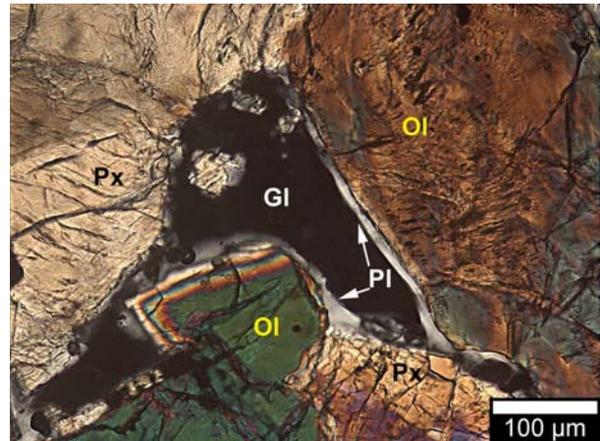


Fig. 5. Transmitted light with crossed polarizers; pyroxene (Px) and olivine (Ol) enclosing isotropic (and vesiculated [1]) plagioclase glass (Gl) in ALH77005. Pl = plagioclase.

Conclusions: Two independent data sets are compared: (1) Shock pressures deduced from the shock metamorphic overprint observed in the rock forming minerals of shergottites [1]. (2) Post shock temperatures calculated with experimentally determined equations of state (EOS) for relevant terrestrial analogue rocks. The deduced shock pressures in combination with post shock temperature calculations are in good agreement with the petrographic observations in shergottites.

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References: [1] Fritz et al. (2005) *MAPS* **40**, 1393–1411 [2] Stöffler et al. (1991) *GCA* **55**, 3845–3867 [3] Stöffler et al. (1986) *GCA* **50**, 889–903 [4] Schmitt R.T. (2000) *MAPS* **35**, 545–560 [5] Artemieva and Ivanov (2004) *Icarus* **171**, 84–101 [6] Melosh H.J. (1989) Oxford University Press. 246 p. [7] Stöffler (1982) *Geophysics & Space Research* Vol 1, Springer, pp. 120-183 [8] Trunin et al. (2001) Sarov: Russian Federal Nuclear Center 446 p. [9] Beck et al. 2005 *Nature* **435**, 1071-1074 [10] Okrusch and Matthes (2005) Springer Berlin Heidelberg, ISSN 0937-7433.