HYPERVELOCITY COLLISIONS OF PROJECTILES AND TARGETS: SHOCK WAVE PROPAGATION IN HETEROGENEOUS ROCKS INFERRED FROM MICROSTRUCTURES T. Kenkmann[1], [1]Museum für Naturkunde – Mineralogie, Humboldt-Universität Berlin, Invalidenstrasse 43, 10115 Berlin, Germany, thomas.kenkmann@museum.hu-berlin.de

Introduction: What do we know about the very early stage of impact cratering from the observational point of view? When projectile and target collide, the kinetic energy of the projectile splits up into four components: The internal energies of projectile and target, the residual kinetic energy of the projectile, and the kinetic energy transferred to the target. The fractions of each component depend on the material properties of projectile and target, e.g. the fraction of internal energy increases with increasing compressibility of the material. However, the partitioning between these components is also strongly time dependent and the energy is finally almost completely transferred to the target [1]. The structure of shock waves that are generated at contact planes depends on the rise time to the final peak pressure which is influenced by the dynamic elastic limit and phase transitions. Shock magnitude and propagation velocity are controlled by the projectile and target material and the impact energy. The duration of the shock corresponds to the size of the projectile [2].

Considering large impact craters, rocks involved in the initial contact and compression stage may get completely vaporized upon pressure release from shocks exceeding hundreds of GPa (except for thin spall plates of the rear of the projectile). The vaporized material subsequently may condensate to fine-grained spherules and is disseminated in ejecta plume deposits (suevites). Trace element analysis [3] of these materials suggests that mixing of projectile and target occurs. Jetting by oblique convergence of the projectile-target contact plane is also a very early time phenomenon observed in experiments. It accelerates material to velocities higher than the initial impact speed [4]. Likewise tektite formation is linked to the early cratering [5]. More information on the contact and compression stage is provided from impact events where the entropy increase from shock compression is not sufficient to entirely vaporize or melt the projectile and target at ground zero. Likely candidates to analyse the early shock propagation are (a) shock experiments, (b) meteorites, and (c) small natural impact craters.

Heterogeneities in rocks: Rocks are principally heterogeneous. Any type of interface between different constituents of a rock such as lithological interfaces, voids, fluid inclusions, grain- and phase boundaries are characterized by a discontinuity in density and, hence, shock wave propagation velocity and cause impedance contrasts [6]. Shock impedance is defined as the product of the density of the material times the shock wave velocity in this material. Impedance contrasts disorganize shock waves and lead to shock wave reverberations. This causes the development of localized concentrations of stress, temperature, and deformation, which can ultimately lead to the formation of thin melt veins [7],[8]. The propagating shock wave reverberates on many interfaces and will increase or decrease the pressure stepwise with respect to the first shock wave. After a short time the grains achieve a mean shock state that is often misleadingly called "equilibrium shock state", e.g. [9],[10]. Shock microstructures used for shock barometry are calibrated against this average pressure. They are in contrast to highly localized shock features like melt veins that can contain high pressure polymorphs, which are sometimes called "disequilibrium shock features". While they are not suited as pressure gauges to determine the average shock level in projectile or target, they are important for the understanding of shock propagation in different materials. The formation of networks of melt veins in the central portions of impact structures and projectiles is an important energy sink and may leave the rock substantially hotter than predicted from Hugoniot data [11]. The formation and persistence of melt veins produced as a result of rock heterogeneities may be a contributor to the loss of strength on pressure release [11].

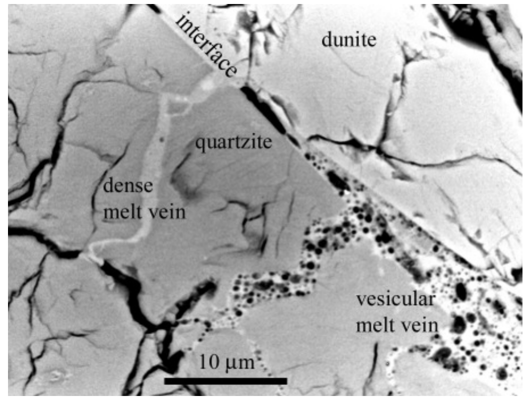

Fig. 1 Experimentally produced dense and vesicular melt vein at the interface between quartzite and dunite [6]

(a) Observation in shock experiments: The shock wave plateau in shock recovery experiments (impedance and reverberation techniques) is always limited to microseconds, which usually prevents the formation of high-pressure phases [12]. Beside the extensive pressure calibration of "equilibrium" shock features by shock experiments, e.g. [12],[13], shock veins were successfully reproduced along lithological interfaces

(Fig. 1)[6], tabular surfaces [8], and within single crystals [14]. Along with the effect of porosity [15] they document the importance of these features for absorbing shock wave energy and for the attenuation of the shock. Formation mechanisms derived from these experiments include shock melting by pressure-temperature excursions (hot spots) plus shear-induced frictional melting. Recently, target and projectiles interpenetration and mixing could be demonstrated experimentally (Fig.2) [16].

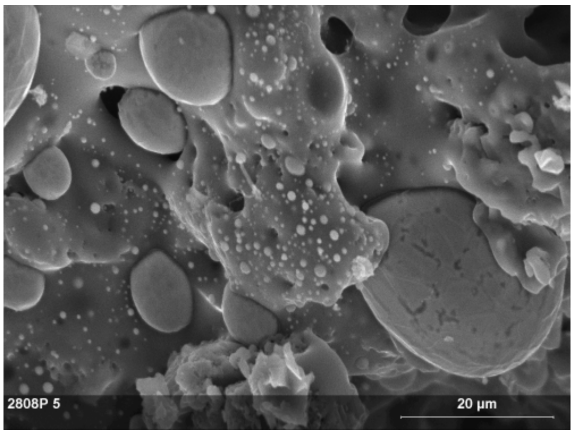

Fig. 2 Surface of a spalled steel projectile after impacting a sandstone target at 5.3 kms^{-1} [16]. The surface is coated by SiO_2 and iron oxide, the bubbles indicate boiling.

(b) Observations in meteorites: Highly shocked meteorites may give some clues to early shock propagation although the observed shock features may result from multiple impacts [10]. The presence of metals in some meteorites lead to impedance contrasts that are substantially higher than those characterizing most terrestrial rocks [9]. Meteorites are often fractured and porous. The strongly different material properties and heterogeneities in meteorites result in shock veins with high pressure silicates [17], melt pockets, melt dikes, and troilite/metal deposits in fractures [9][18]. The effects are most pronounced at metal-silicate and metal-pore space interfaces.

(c) Observation in small natural impact craters: Impact craters can contain residues of the projectile. Their emplacement most likely occurs in a very early stage of the crater forming process. The mechanism by which the emplacement occurs is a matter of debate. It could be emplaced in a vapor phase, as melt droplets or as fine grained solid material. Rocks closest to ground zero undergo the largest amount of strain [11], and frictional heating may contribute to an unknown magnitude to the total heat budget. Like projectiles, rocks located near the point of impact initially contain open pore space that can be filled with volatiles such as water. The strongest impedance contrasts within a rock occur at free surfaces. Open fractures, cleavage planes, or porosity effectively absorb the shock wave energy and heat the target [19]. Consequently, the volume of material shocked during an impact is commonly smaller in porous rocks than in dense rocks as a significant part of energy is used for pore space collapse and localized melting. The depth down to which cavities remain open is inversely proportional to gravity of a planetary body. On Earth, open fractures and pore space may occur down to depths of 1 km and more. Simple impact craters may develop completely within this upper target layer where cavities are present. The larger the impact crater, the less prominent is the effect of open pores or fractures for the cratering process. Interestingly, the simple-to-complex transition of impact craters on Earth (~3-5 km diameter) seems to correlate with depth down to which open cavities may exist. Is the simple-to-complex transition of impact craters even influenced by this critical depth?

Cooperation of observers and modelers: To better understand shock microstructures, more micro- and mesoscale numerical simulations are desired to fully understand their dynamic formation. These micromechanical models may also give valuable input algorithms for macro-scale modeling. Among other aspects observers should try to quantify how much shock energy is absorbed at heterogeneities.

Acknowledgement : I am grateful to the Workshop organizers for their kind invitation. The research was made possible by Deutsche Forschungsgemeinschaft.

References: [1] O´Keefe, J. D., Ahrens, T. J. (1982). GSA SP 190, 103-120. [2] Melosh, H. J. (1989) Impact cratering, 245 p. Oxford. [3] Koeberl, C. (1994) GSA-SP 293, 133-154. [4] Gault, D. E. et al. (1968) in Shock metamorphism of Natural Materials, 87-99. [5] Artemieva, N. A. (2002) in Impacts in Precambrian shields, 257-276, Springer. [6] Kenkmann, T. et al. (2000) MAPS 35, 1275-1290. [7] Kieffer, S. W. (1977) in Impact and Explosion cratering, 751-769. [8] Heider, N. Kenkmann, T. (2003). MAPS, 38, 151-1460. [9] Stöffler, D. et al. (1991) Geochim. Cosmo. Acta, 55, 3845-3867. [10] Stöffler, D., Grieve, R. A. F. (2007) IUGS Subcommission on Metamorphic rocks, Chapter 11, Blackwell. [11] Fiske, P. S. et al. (1995) Science, 270, 281-283. [12] Langenhorst, F., Deutsch, A. (1994) EPSL 125, 407-420. [13] Reimold, W. U., Stöffler, D. (1978) Proc. LPSC, 9, 2805-2824. [14] Langenhorst, F. et al. (2002) MAPS, 37, 1541-1553. [15] Stöffler, D. et al. (1975) JGR, 80, 4062-4077. [16] Kenkmann, T. et al. (2007). LPSC 38,CD-Rom#1831 [17] Sharp et al. (1997) Science, 277, 352-355. [18] Spray, J. G. (1999). Geology, 27, 695-698. [19] Kieffer, S. W. (1971) JGR 76, 5449-5473.