IMPACTS INTO THIN CRUST OVERLYING A MAGMA OCEAN

A.P. Jackson[1], V. Perera[2], T.S.J. Gabriel[2], L.T. Elkins-Tanton[2], E. Asphaug[2]. [1]Centre for Planetary Sciences, University of Toronto, Toronto, ON, Canada, [2]School of Earth and Space Exploration, Arizona State University, Tempe, AZ, USA.
Email: ajackson@cita.utoronto.ca

Introduction: Magma oceans were common in the early solar system, whether produced by accretion energy [1,2], metal-silicate differentiation [3] or radiogenic heating [4,5]. As the number of small bodies, and so the frequency of impacts, was much higher than today, impacts onto bodies hosting a magma ocean would also have been common. How these impacts interact with the target body and influence its evolution depend on the nature of the target and bring energy and disruption to the solidifying crust of the magma ocean.

A magma ocean with a free surface presents a very different scenario to one with an overlying lid. On large planets like Earth magma oceans freeze from the bottom upwards, maintaining a molten free surface until the magma ocean has fully solidified. On smaller planets like the Moon, and possibly Mercury, buoyant phases that rise to the surface can form a flotation crust [6]. In the case of asteroids that undergo internal melting due to radiogenic heating a primitive crust that is never melted may be maintained [7].

Magma oceans on the terrestrial planets that are formed as a result of giant impacts will be subject to especially intense bombardment since the giant impact will release large quantities of debris into the solar system on heliocentric orbits (~1.3 lunar masses for the, rather gentle, canonical moon forming scenario [8]). This debris will have high probabilities for re-impacting the originating body. Such returning material will also have lower typical impact velocities than impactors originating from elsewhere in the solar system. This sets the stage for an immediate cataclysmic epoch that has been little studied, and that has important, testable, thermal and collisional consequences for the Moon and planets.

Crusts as thin as a few tens of metres can serve as an insulating blanket over a cooling magma ocean. Re-impacting debris can therefore substantially influence the thermal evolution of the body by disrupting this blanket. In particular, we focus on the magma ocean that existed on the early Moon.

Methods: We use the well-established iSALE shock physics code [9,10,11] to simulate impacts into the flotation crust over the lunar magma ocean. We examine a range of impactor sizes, impact velocities and crustal thicknesses. Impacts by objects 10 km and smaller are simulated in a half-space setup, while those larger than 30 km use a 2D spherical setup. Impacts by objects between 10 and 30 km are simulated in both geometries to ensure continuity between regimes. Our Moon is constructed with a dunite mantle and a granite crust using the ANEOS equation of state [12]. These material choices satisfy our two primary criteria; the solid crust is less dense than the liquid mantle, and the crust has a higher melting temperature than the mantle. The depth of the liquid magma layer is determined by the crossing point of the mantle adiabat and melting temperature-pressure curve. This depth is matched to that in [13] for the appropriate crustal thickness by adjusting the temperature at the base of the crust.

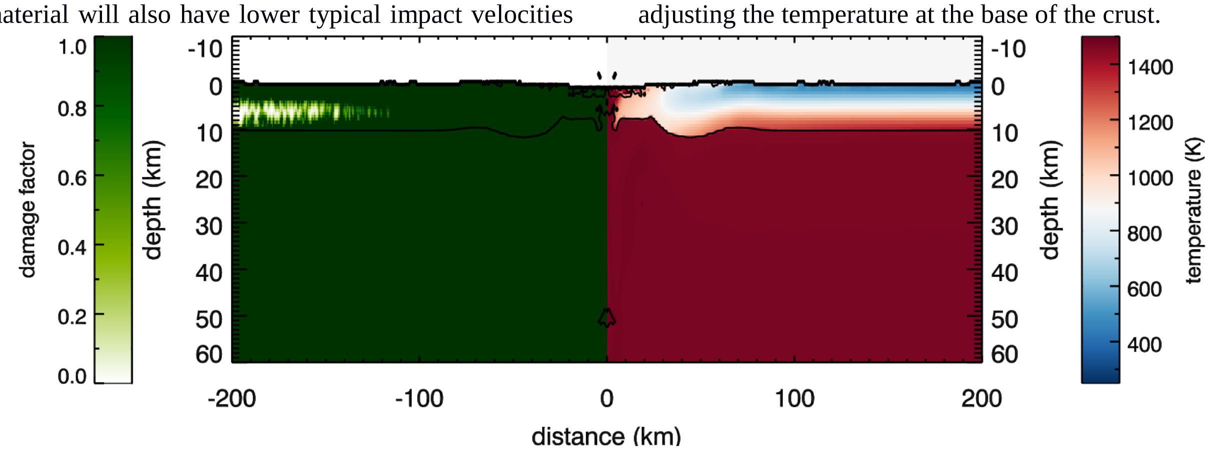

Figure 1: iSALE simulation of 10 km object impacting 10 km thick crust at 5 km/s, snapshot at 4000s. At this time the simulation volume is very well settled into a post-impact equilibrium. The liquid magma ocean is 90 km deep. The left-hand pane shows the damage factor (the liquid magma is strengthless and so completely damaged), the right-hand pane shows the temperature. The black contours indicate the boundary between crust and mantle materials. Thermal profiles in the punctured crust are plotted in figure 2.

Results: In figure 1 we show an example of the state of the crust at the end of one of our simulations, here for a 10 km impactor at 5 km/s into crust 10 km deep (corresponding to a magma ocean depth of 90 km). There are two particular features of interest:

1) The central hot hole in which the crust has been substantially thinned and strongly raised in temperature, bringing the heat of the magma ocean in direct contact with the surface.
2) The extensive fracture zone throughout which the crust is heavily damaged.

Figure 2 shows profiles of temperature vs. radial distance from the centre of the impact site for three depths in the crust. A direct pipe between the surface and the magma ocean is only present in the innermost 5 km, however out to 30 km the crust is thinned and hotter at all depths. Beyond 30 km the picture is more complicated, rather than directly bringing magmatic heat to the surface the crust is churned and folded, bringing hotter material to the surface but forcing cooler material downwards. This is partly due to splashing of hot magma onto the surface.

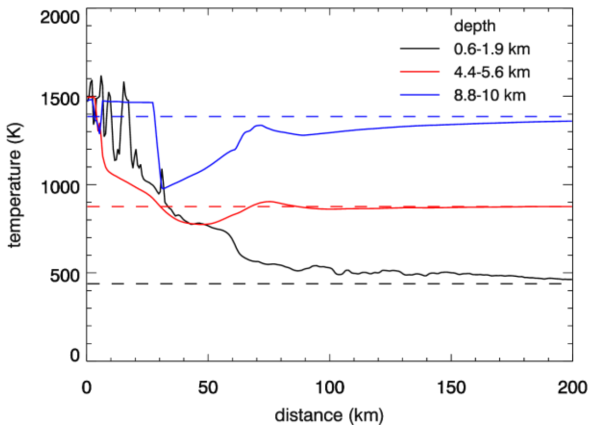

Figure 2: Horizontal temperature profiles at three depths, at 4000s after the impact (solid) and before the impact (dashed).

Damage reaches unity once the tensile stress in the material exceeds a scalar threshold (we use the implementation of [14] for material damage). Previous studies of lithosphere rupture [15] show that damage can extend to several crater radii. Here the 10 km impactor damages surface material well beyond 100 km in response to shock and large-scale flexure.

Implications: The combined effects of the creation of thermal holes and extensive fracturing of the crust have profound consequences for the evolution of the lunar magma ocean and crust. As demonstrated in the LPSC 2017 abstract of Perera et al. even a covering fraction <1% of hot holes substantially decreases the solidification time of the magma ocean since radiative cooling is proportional to T^4. The extensive fracturing will weaken the crust to subsequent impacts and may indicate that the lunar crust has always had a substantial mega-regolith component.

Acknowledgements: The authors gratefully acknowledge support for this project through NASA grant NNX16AI31G (Stop hitting yourself). This work has made use of the iSALE shock physics code (Dellen release) [4,5,6] and we gratefully acknowledge the developers, including G. Collins, K. Wunnemann, D. Elbeshausen, B. Ivanov & H.J. Melosh.

References:

[1] Wetherill, G.W. (1980) *AREPS*, 18, 77
[2] Tonks, W.B., Melosh, H.J. (1993) *JGR*, 98, 5319
[3] Monteaux, J., Ricard, Y., Coltice, N., Dubuffet, F., Ulrova, M. (2009) *Earth Planet. Sci. Lett.*, 287, 353
[4] Miyamoto, M., Fujii, N., Takeda, H. (1981), *Proc. Lunar Planet. Sci. Conf.*, 12, 1145
[5] Merk, R., Breuer, D., Spohn, T. (2002) *Icarus*, 159, 183
[6] Elkins-Tanton, L.T. (2012) *AREPS*, 40, 113-139
[7] Elkins-Tanton, L.T., Weiss, B.P., Zuber, M.T. (2011) *Earth Planet. Sci. Lett.*, 305, 1
[8] Jackson, A.P., Wyatt, M.C, (2012) *MNRAS*, 425, 657
[9] Amsden, A., Ruppel, H., Hirt, C. (1980) *Los Alamos Nat. Lab. Rep.*, LA-8095:101p
[10] Collins, G.S., Melosh, H.J., Ivanov, B.A., *Meteoritics Planet Sci.*, 39, 217
[11] Wunnemann, K., Collins, G., Melosh, H. (2006) *Icarus*, 180, 514
[12] Melosh, H.J. (2007) *Meteoritics Planet. Sci.*, 42, 2079
[13] Elkins-Tanton, L.T. (2011) *Earth Planet. Sci. Lett.*, 304, 326
[14] Ivanov, B.A., Deniem, D., Neukum, G. (1997) *Int. J. Impact Eng.*, 20, 411
[15] Turtle, E.P., Pierazzo, E. (2001), *Science*, 294, 1326