THE STRESS STATE OF A COOLING MAGMA OCEAN. R. S. Scott and L. Wilson, Planetary Science Research Group, Environmental Science Dept., Lancaster University, Lancaster LA1 4YQ, UK. (email: r.s.scott@lancaster.ac.uk).

Summary: We address the early thermal and mechanical evolution of a lunar crust forming from a magma ocean to assess the conditions under which the earliest forms of volcanism may have occurred. Potentially large compressive stresses may have been temporally and spatially common.

Introduction: Understanding the Moon's thermal and magmatic evolution entails understanding the origin of the earliest crust, in particular whether it formed from a magma ocean [1]. The energy of late stage planet-building impacts would be at least sufficient to form magma oceans and possibly enough to melt whole planets [2].

The solidification history of the magma ocean will place constraints on the tectonic evolution of the Moon. Lunar igneous activity starts from the formation of the original crust that developed in response to the radiative cooling effects of a body devoid of an atmosphere. There is still some uncertainty as to the precise age of the Moon and consequently the time span during which the crust formed. However, there is common agreement that the Moon formed between 4.55 and 4.5 Ga ago [3, 4, 5], approximately 50 Ma after the formation of the Solar System [6], [7], and that the magma ocean existed for between 40 and 200 Ma [3, 8, 9,10, 11, 12].

The Moon's crust initially contracted in response to the cooling of the magma ocean. Little attention has been given to the stresses built up within the developing crust as the magma ocean cooled during the Moon's early history. These stresses are likely to have been very large and should have manifested themselves as compressive surface features similar to those evident on Mercury [13, 14]. Unfortunately, tectonic features may have been obliterated by the tail end of heavy impact bombardment [15].

The surface of a magma ocean will quickly cool and a crust would form [9]. It has been suggested [8] that a crust approximately 10 km thick would act as an effective thermal insulator. This 10 km thick crust represents one of the initial conditions for the following model; it ignores the effect bombardment of the surface will have on the thermal properties of the crust. It is likely that the bombardment and the subsequent brecciation will enhance the insulating effects and only serve to slow the cooling process, although in the early stages of the development of this crust the bombardment may in fact increase the rate of cooling [9, 16].

The cooling magma ocean thermal stress model: The order of magnitude of stresses applicable to the cooling of a magma ocean has been considered in previous work [17]; however the stress-related consequences of a cooling magma ocean are likely to be complex. Several factors need to be considered and taken into account if a realistic mathematical model is to be developed. In the simple thermal stress model [17] it was assumed that the thermal and mechanical properties of the developing crust were constant and that the temperature gradient was linear. It is very unlikely that this will be the case. In our modified model, the effects of variations in thermal conductivity, thermal diffusivity, coefficient of expansion and non-linear cooling of the crust are introduced. Introducing these variables takes into account the more realistic situation of a non-linear rate of change in deviatoric stresses and should lead to a better understanding of how these same stresses might have accumulated or relaxed whilst the crust was growing. The rate of change of absolute temperature T with depth y below the surface and time t since the onset of cooling is determined by representing the crust as a semi-infinite half-space. The variation of temperature with depth within the crust can then be solved by the use of the similarity variable, η, and the complimentary error function, erfc η [18].

If the absolute temperatures of the surface and of the base of the thermal boundary layer are defined as T_S and T_B, respectively, then:

$$\left[\frac{(T-T_B)}{(T_S-T_B)}\right] = erfc\left[\frac{y}{2(\kappa t)^{0.5}}\right] \quad (1)$$

where κ is the mean thermal diffusivity of the cooling layer and erfc is the complementary error function [18], values of which can be obtained from standard tables.

Data on the temperature dependence of thermal diffusivity for rocks and minerals are scarce [19]. However, thermal diffusivity and thermal conductivity are related by [18, 19]

$$\kappa = \frac{k}{\rho C_p} \quad (2)$$

where k, ρ, and C_p are the thermal conductivity, density and specific heat capacity at constant pressure, respectively. Thermal expansion, conductivity, density and the specific heat capacity are all temperature dependant and these changes, whilst small, have been accounted for in the model.

Deviatoric mean thermal stresses for several representative depths and a variety of different crustal thicknesses, each subject to a temperature gradient spanning from 250 K at the surface [8, 20] to 1450 K at the crust/magmasphere interface [21], are shown for the simple model [17] and the modified model in Fig.1 and 2.

Both the simple and the modified models appear to predict similar values of deviatoric stresses, 140, 370 and

600 MPa for crustal temperatures of 1100 K, 800 K and 500 K respectively. However, they reflect totally different treatments of the problem. In the simple model the relationship between the depth of crust and a corresponding temperature is based on the assumption that the geotherm is linear. In the modified model the relationship between the depth in the crust and temperature is determined by solving the Stefan Problem. A comparison will show that the temperature-related stresses are at different depths in the crust. In the simple model the stresses associated with the 1100 K, 800 K, and 500 K isotherms are at significantly greater depths than those applicable to the modified model. An important consequence of this significant difference in depth is that the 1100 K isotherm, i.e. the isotherm that represents the base of the lithosphere and the top of the viscoelastic zone, is much higher in the crust in the modified model. This shallower depth of the viscoelastic zone means that viscous relaxation of stresses can occur much closer to the surface than would be apparent from the simple model.

Conclusion: This model serves to illustrate the magnitude of compressive stresses that could exist in a totally homogeneous crust. The magnitude of these stresses is sufficient to cause mechanical failure, and the magnitude and orientation of the stresses must have had a major control on the extent, location and nature of pre-mare volcanism.

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Fig. 1 and Fig. 2 show the deviatoric mean thermal stresses for several representative depths and a variety of different crustal thicknesses, each subject to a temperature gradient spanning from 250 K at the surface to 1450 K at the crust/magmasphere interface, for the simple model and the modified model respectively.

Fig. 1

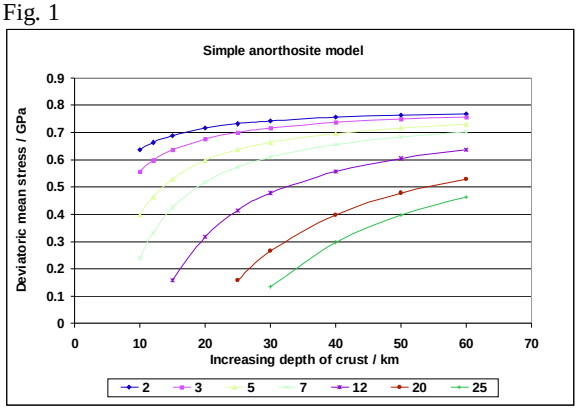

Fig. 2