THERMAL EVOLUTION OF LARGE LUNAR IMPACT BASINS: IMPLICATIONS FOR BASIN COMPENSATION AND THE DURATION OF THE LUNAR CATACLYSM

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Introduction

Observations of the gravity and topography of lunar impact basins show a range of patterns [1-5]. Some, such as Imbrium and Orientale, are deep basins with super-isostatic uplift of the mantle (often termed "mascons"). Others, such as Australe and Fecundatatis, are shallow, isostatically compensated structures. These properties are correlated with basin age, suggesting that these basins may have evolved with time by visco-elastic flow of the crust and mantle [6-7].

The rate at which basins relax depends on the temperature and viscosity of the crust and mantle. Highly relaxed basins imply that the Moon experienced a hot early state, whereas unrelaxed basins require a cold, stiff Moon that is capable of supporting large stresses. Because the cooling of the Moon took a finite period of time, it is possible to use these geophysical observations to test the lunar cataclysm hypothesis. If the cooling period for stabilizing unrelaxed basins is much longer than the duration of an impact cataclysm, then all basins would have a similar relaxation state, which is not observed. The observed range of basin relaxation states implies that any impact cataclysm had a duration that was at least comparable in duration to the cooling time. If most of the Moon's impact basins formed in a 100 million year period [8, 9], then the Moon's thermal evolution must have been very rapid. Alternatively, Baldwin [10, 11] and Schmitt [12] suggested that the basin-forming epoch lasted 400-500 million years. We are modeling basin thermal evolution using finite element methods to set a geophysical lower bound on the duration of the lunar cataclysm.

Methods

Large impact basins can impose significant thermal perturbations on a planetary mantle, and the effects of such perturbations have been explored in a number of recent studies [13-17]. The focus of the current study is understanding how these thermal anomalies affect the lithospheric structure of the Moon and thus the compensation state of impact basin structures such as mascons and basin rings. The initial thermal states of our models are based on the output of hydrocode simulations of the formation of the South Pole-Aitken and Orientale impact basins [18]. Large amounts of melt are produced in these simulations. We assume that convection in the liquid magma chamber will rapidly cool the melt to the solidus and thus begin our modeling with the maximum temperature set to the pressure-dependent solidus temperature. The background thermal gradient is assumed to be in the range 10-50 K/km, appropriate for the early stages of lunar history [13, 19-23].

Because of the approximate circular symmetry of many large lunar impact basins, we model the thermal evolution in spherical axisymmetric geometry using a finite element model that has been previously used to study mantle plume volcanism on Mars [24, 25]. Because of the high thermal gradients used in some of these models, very fine spatial resolution (~2 km per element) is used in the calculations. The initial results shown here focus on the conductive thermal evolution of the crust and uppermost lithosphere during the first 100 million years after a basin forming impact. The longer term thermal evolution of the mantle, including the possible role of thermal or thermal-chemical convection is a matter of on-going study (whether or not the early lunar mantle is actually convectively unstable will depend in part on the nature of density gradients associated with solidification of the lunar magma ocean [26]). The models include radioactive heating at a magnitude appropriate for early lunar history.

The time evolution of the elastic lithosphere is calculated using a standard strength envelope approach [25] (Figure 1). The upper crust, lower crust, and mantle are represented by dry anorthite, dry diabase (as a proxy for norite), and dry olivine respectively. Brittle portions of the strength envelope are modeled using Byerlee's law.

Results

Based on the amount of crustal thinning and the surface distribution of crust and mantle rocks, the most successful hydrocode simulations for the South Pole-Aitken basin assume a pre-impact thermal gradient of 50 K/km [18]. Given this high thermal gradient as well as the large volume of shock melting that forms near the center of the basin, the calculated lithospheric thickness immediately after the impact is only 2-3 km out to a distance of 500 km from the basin center. Such a thin elastic lithosphere is unable to support significant loads, so it is not surprising that the SP-A basin is observed to be essentially isostatic [e.g., 5].

At larger distances from the basin center, the level of shock heating is much less, allowing the lithosphere to be thicker. At 1000 km radius, which is close to the SP-A basin rim, the lithosphere thickness immediately after the impact is ~11 km (Figure 1). Although not large, such a lithosphere is sufficient to provide some degree of flexural support for short-wavelength topographic features, such as mountain rings in the basin rim. The amount of shock heating in the SP-A impact is sufficiently large that the lithosphere at the center of the basin is only a few km thick even if the pre-impact thermal gradient is only 10 K/km. However, the lithospheric thickness at the basin rim is very strongly dependent on the pre-impact thermal gradient. Moreover, the lithospheric structure near the basin rim is essentially unperturbed from its pre-impact state. Thus, gravity and topography observations of the basin rim can be used to estimate the pre-impact lithospheric thickness and thermal gradient. This would provide useful new constraints on the Moon's overall thermal history. Moreover, constraints on the pre-impact thermal gradient would also be useful in constraining one of the major variables for future hydrocode simulations of basin impacts.

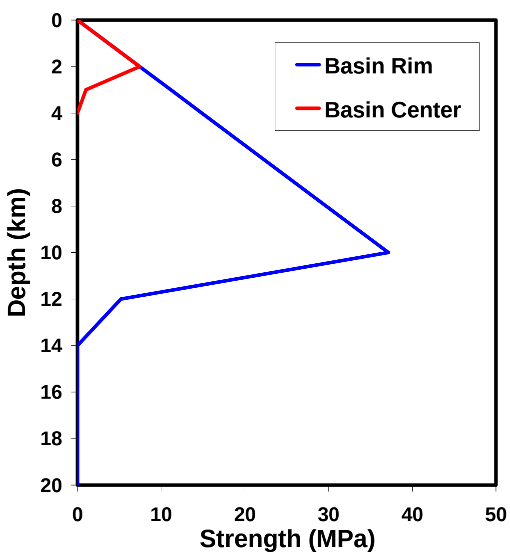

FIGURE 1. Lithospheric strength envelope at the basin center (red) and near the basin rim (blue) immediately after the impact.

The lithosphere in these models thickens gradually with time due to conductive cooling. Because of the very high post-impact thermal gradient, the rate of thickening is greatest near the basin center. For the nominal SP-A hydrocode model and a 50 K/km pre-impact thermal gradient, the lithosphere thickens from 2 km at time t=0 to 13 km at 100 Ma after the impact. On the periphery of the basin, the thermal gradients are smaller and thus the rate of thickening is also smaller. For the nominal model, the lithosphere at 1000 km from the basin center thickens from 11 km at t=0 to 17 km at t=100 Ma. The rate of thickening is also sensitive to the amount and distribution of radioactive heating and thus depends both on the location of the basin with respect to the high radioactivity Procellarum KREEP Terrane [27] and to the to the thickness of the crust [3]. The sensitivity of the model results to these parameters and to the possible presence of convection in the lunar mantle is currently being assessed. These thermal evolution results can be used as input to improved models of the visco-elastic relaxation of impact basin topography.

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