PRESERVATION OF SUPERISOSTASY IN LARGE LUNAR BASINS. Jeffrey A. Balcerski[1], Steven A. Hauck, II[1], and Andrew. J. Dombard[2],[1]Dept. of Geological Sciences, Case Western Reserve University, Cleveland, OH 44106 (jeffab@case.edu), [2]Dept. of Earth and Environmental Sciences, University of Illinois at Chicago, Chicago, IL 60607.

Introduction: Analysis of lunar gravity and topography data suggest that several large basins currently exist in a state of isostatic disequilibrium [1-4]. That is, assuming a constant density crust and mantle, the topographic relationship of the surface and crust-mantle interface indicate that the Moho is substantially elevated above and beyond an equilibrium position.

It has been widely postulated that this superisostasy is a product of the basin forming event that resulted in mantle material being excessively uplifted below the basin crust, and was subsequently "frozen" in this position.[e.g., 3,4,10] This situation implies that the conditions within the basin rapidly stabilized with thermomechanical parameters that were able to preserve this state of isostatic overcompensation throughout the vast majority of the geologic history of the Moon. Prior studies have concluded that there has been little or no isostatic adjustment or lower crustal flow since the time of basin formation, because many basins retain a high degree of surface relief and a positive Bouguer gravity anomaly [1]. Using a simple analytic model, Namiki et. al [1] have concluded that in order to satisfy this condition, the Moho could have been no warmer than 1000 K with an effective viscosity of no less than 10^{27} Pa s.

By adopting a numerical approach, the effective viscosity and strain rate develop implicitly rather than being explicitly specified, as in the case of the analytic model above. We use the flexibility afforded by this method to test the stability of these basins in a range of thermal and mechanical conditions in order to determine the limiting bounds that effectively preserve the currently observed superisostatic configuration.

Model: Using commercial finite element software, we develop a suite of axisymmetric viscoelastic models representing basins ranging from 400-1000 kilometers in diameter. This range roughly brackets the lunar basins from Mendel-Rydberg to Orientale in size, both of which appear to be superisostatically compensated. [1-4].

The surface topography is generated with a quartic polynomial that is truncated at the surface and extends to a maximum depth at the basin center that is determined by the minimum crustal thickness at that point, and the degree of isostatic compensation. No short-wavelength elevated rim is included in this geometry. We assume a nominal crustal thickness of 60 km and a minimum crustal thickness (in the center of the basin) of 30 km, which is generally representative of many of the basins of the lunar far-side [1,3,4]. The degree of compensation is determined by the ratio:

$$D = \frac{h_m(\rho_m - \rho_c)}{h_c(\rho_c)}$$

where D is the degree of compensation, h_m is the height of the elevation of the Moho above nominal, h_c is the depth of the basin, and ρ_m and ρ_c are the density of the mantle and crust respectively. The degree of isostatic compensation is assumed to be initially uniform throughout the basin, resulting in a crust-mantle topography that proportionally reflects the surface topography. For the purposes of this study, the degree of compensation is chosen to be 50% greater than equilibrium, i.e. $D = 1.5$.

Mare fill in these basins simply adds to an already uncompensated load, which further enhances topographic depression and subsidence. In order to simplify the morphology of our models, we therefore do not explicitly include this effect.

We approximate the thermal environment with a linear gradient throughout the depth of the model, with temperatures prescribed a priori at the surface and crust-mantle interface. This temperature distribution is held constant for the duration of the model, as it represents a maximum of viscous relaxation that may occur for any given initial temperature. For each basin size that is modeled (400-1000 km), we create four thermal scenarios where the surface temperature is fixed at 253 K and the crust-mantle interface is maintained at either 400, 600, 800, or 1000 K. In the hottest case, the Moho exceeds the maximum temperature that allows preservation of topographic overcompensation (800 K as argued in [1]). To test the lower limit of viscosities needed to prevent basin relaxation [1], we allow these model viscosities to range well below the 10^{27} Pa s noted previously. Any decrease in effective viscosity carries a significant increase in computational cost, and since the lowest viscosities are generally found well outside of the region of interest [9], we enforce a lower limit of 10^{21} Pa s to maintain manageable run times for our preliminary results.

Viscous flow is controlled by temperature and stress-dependent dislocation creep flow laws. The models presented here have a two-layer rheology (following [5]) with a diabase crust and olivine mantle.

Results: Figure 1 shows the resulting degree of compensation of the 16 models previously described. Consistent with the well-known wavelength dependent elastic support of topography [e.g. 11], the larger basins maintain the least amount of initial overcompensation over a 1 Gyr timeframe. We also observe that the smaller basins of 400-600 km in diameter can preserve a significant fraction of the initial overcompensation, though all basins modeled relax at least partially.

In all of these models, the crust and mantle remain strongly coupled with negligible changes in crustal thickness. Correspondingly, we observe on the order of less than 1% horizontal creep strain throughout the crust, indicating a dramatically different time scale for vertical isostatic adjustment than for horizontal topographic relaxation.

Discussion: The thermal model that we have chosen to employ yields a remarkably similar viscosity depth profile, in the crust and upper mantle, to those created by conductive radiogenic heating models (e.g., [6]). To completely describe the thermal environment, however, the effect of secular cooling and parent radioisotope depletion should be included. Given the conclusions of other studies (e.g., [6]), there may have been regions of the Moon that remained anomalously warm for periods of time exceeding the simulated time of our model, which reduces the impact of cooling for these areas.

Other rheological models may have additional effects upon basin relaxation and may be more appropriate for the lunar far side. We continue to evaluate the effect of rheologies of other materials, such as anorthite [8] in this environment.

Lastly, due to the planar geometry of these models, the effect of membrane support of topographic loads, which dominates over elastic support for lunar basins larger than ~600 km in diameter [7], is not included. However, preliminary analysis of axisymmetric hemispheric models suggests that this effect is minimal, yielding only slight additional resistance to isostatic adjustment at lower model temperatures, and having a limited effect at higher temperatures. Thus, a planar approximation for this investigation is both appropriate and considerably less demanding on computational resources.

Conclusions: In contrast to the analytic model of [1], our results suggest that topography associated with isostatic overcompensation may be maintained throughout a wide range of thermal conditions, including those with Moho temperatures well below 1000 K. This is especially pronounced for the smaller basin sizes, like many of the lunar far side basins. In addition, we note that even the larger basins are able to retain a substantial degree of initial overcompensation if they are located within an environment of cooler temperatures.

References: [1] Namiki et. al. (2009) *Science, 323,* 900-904. [2] Ishihara et. al. (2009) *GRL, 36, L19202.* [3] Neumann et. al. (1996) *JGR*, 101, 16841–16843. [4] Wieczorek, M.A. and Phillips, R.J. (1999) *Icarus, 139,* 246-259. [5] Mohit, P.S. and Phillips, R.J. (2006) *JGR*, 111, E12001. [6] Wieczorek, M.A. and Phillips, R.J (2000) *JGR*, 105, 20417-20430. [7] Turcotte et. al. (1981) *JGR* 86, 3951-395. [8] Rybacki, E. and Dresen, G. (2000) *JGR,* 105, 26017-26026 [9] Dombard, A. and McKinnon, W. (2006) *JGR,* 111, E01001. [10] Stewart, S. (2010) *LPS XXXXI*, Abstract #2722. [11] Turcotte, D. and Schubert, G. (2002) *Geodynamics,* Camb. Univ. Press.

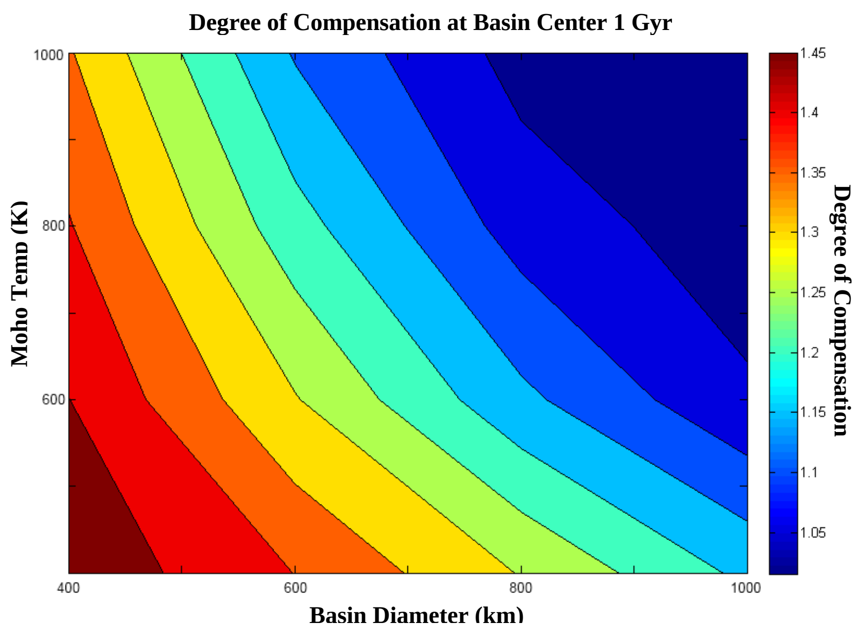

Figure 1 – Final degree of compensation of centers of basin models after 10^9 years of viscoelastic evolution. The initial degree of compensation in all cases is 1.5.