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The deep lunar interior with a low-viscosity zone: Revised constraints from recent geodetic parameters on the tidal response of the Moon

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ABSTRACT

We revisit the constraints on the deep lunar interior with a possible low-viscosity zone at the core-mantle boundary obtained from our previous forward modeling of the tidal response of the Moon, by comparing a numerical model with several tidal parameters (i.e., k_2, k_3, h_2, and Q) that have been improved or are newly determined by recent geodetic observations and analyses from GRAIL (gravity), LRO (shape), and LLR (rotation). Our results are in principle consistent with these data and suggest a low-viscosity layer (with an outer radius of about 540–560 km) which possibly extends inside the region where deep moonquakes occur.

1. Introduction

The tidal response of the Moon has been observationally and theoretically investigated, with the goal to elucidate its internal structure. Observationally, lunar laser ranging (LLR) to the reflectors on the lunar surface and precision orbit determination (POD) for the lunar orbiters (e.g., Lunar Prospector, Selenological and Engineering Explorer, Chang'e-1, etc.) have provided basic information on the rotational (e.g., Dickey et al., 1994) and gravitational (e.g., Konopliv et al., 2001) variations of the Moon with time, respectively, and thus its tidal response. Both the rotational and gravitational variations are connected with tidal deformation because a tidal force exerted on a celestial body induces periodic potential and inertial perturbations (e.g., Williams et al., 2001; 2014; Williams and Boggs, 2015). Tidal deformation, as well as other geophysical information like seismic (e.g., Nakamura et al., 1973) and electromagnetic (e.g., Hood et al., 1999) data, generally provides useful clues to constrain a model of the deep interior of a planet. Theoretically, the lunar internal structure has been revealed by its tidal parameters as follows.

Previous research on the lunar tidal deformation discussed the possible existence of a soft part in the lunar deep interior, determined through stochastic inversion from observational tidal parameters. These studies performed a Markov chain Monte Carlo method and explored a large parameter space, describing a plethora of realistic internal structure models for the Moon (Khan et al., 2004; Khan and Mosegaard, 2005). While the first attempt (Khan et al., 2004) successfully detected its possible liquid core as a drastic reduction in the seismic velocity profile, the second trial (Khan and Mosegaard, 2005) additionally discovered one more slight reduction in the deepest mantle. This low-velocity zone was interpreted as a melt-bearing layer (Khan and Mosegaard, 2005), similar to their inversion of lunar free oscillation periods (Khan and Mosegaard, 2001). This possible detection of partial melt (Khan and Mosegaard, 2001; 2005) supports evidence from Apollo's lunar seismic experiment (e.g., Nakamura et al., 1973) for the presence of partial melt in the deep lunar interior.

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More recently, several studies have further suggested a low-velocity and/or low-viscosity zone in the lowermost mantle by means of forward or inverse approaches. Concerning the forward approach, our own last study (Harada et al., 2014) computed the viscoelastic tidal deformation of the Moon by assuming that the seismic attenuation zone in the deep interior (e.g., Nakamura et al., 1973; Nakamura, 2005) reflects a strong viscosity contrast. By comparing the numerical values of the tidal parameters with their respective observational ranges, we obtained solutions for the outer radius and viscosity of this layer. Concerning the inverse approach, one more stochastic study (Khan et al., 2014) similar to those cited above estimated the boundary radius of the low-velocity layer at the base of the mantle. Their outer radius is relatively close but somewhat larger than that in our last study (Harada et al., 2014). A larger outer radius of the soft layer that agrees with the recently obtained value (Khan et al., 2014) was also determined by using another Bayesian analysis (Matsumoto et al., 2015) although they dealt with the viscosity structure unlike the other inverse approaches cited above, in the same way as our forward modeling. Their viscosity value is very close but slightly larger than that in our above-mentioned study. In any case, both of the above forward and inverse approaches on the tidal parameters match well with observational values as long as such a specific layer exists in the lowermost mantle.

Our forward modeling study, however, did not include the latest geodetic parameters on the tidal deformation derived from new observations and analyses. For example, our previous study only referred to the tidal quality factor (i.e., Q) published previously in Williams et al. (2001) but did not include the recent Q determination of Williams and Boggs (2015). This redetermination affects the monthly and annual periods considered in Harada et al. (2014). Although the frequency-dependence in the old Q solution by Williams et al. (2001) is relatively weak, that in the new solution by Williams and Boggs (2015) is even weaker, indicating that the observational Q ranges in the monthly and annual periods are overlapped with each other. In addition to the Q values for each of the tidal frequencies, the latest POD analyses provided the potential (i.e., k_2 and k_3, for the spherical harmonics degree two and three, respectively) and displacement (i.e., h_2) Love numbers. As to the tidal potential, both Konopliv et al. (2013, 2014) and also Lemoine et al. (2013, 2014) independently derived extremely precise gravity field models by using the twin satellites of the Gravity Recovery and Interior Laboratory (GRAIL) mission. They determined k_2 with relatively high precision from the time-varying gravity field, and also provided a first satellite-based estimate of k_3. As to the tidal displacement, and apart from the LLR studies of Williams et al. (2001) and Williams and Boggs (2015), Mazarico et al. (2014) recently detected the body tide by the analysis of altimetry crossovers from the Lunar Orbiter Laser Altimeter (LOLA) instrument on board the Lunar Reconnaissance Orbiter (LRO) mission and estimated h_2. Nevertheless, our previous study referred to the pre-GRAIL k_2 values given by Goossens et al. (2011) and Yan et al. (2012), but did not include GRAIL's k_2 value, nor did it include any values for k_3 and h_2.

Because the previous inverse approaches already took some (though not all) of the above-mentioned up-to-date observables into consideration, any comparable forward approach ought to include them as well. Although a parallel approach in terms of not only new observations but also of viscoelastic modeling has already been done by Matsumoto et al. (2015) based on the in-depth inversion as mentioned above, it is still worth reinvestigating the same issue in here by means of a forward approach instead. These approaches are complementary in general. While a stochastic inversion method like the one used in Matsumoto et al. (2015) enables us to handle a large parameter space more easily, a forward one can help us to gain concise insights into model responses more intuitively by virtue of being able to make quick connections between input and output parameters. This reconsideration is potentially important to see the hidden parts inside our natural satellite more clearly. It is of great importance especially if the soft layer implies a specific physical state. Such a constraint further allows us to learn the physical behavior at the deep interior of the Moon.

Therefore, we revisit the constraints on the deep lunar interior derived from our earlier investigation in order to reconsider several tidal parameters that have been improved or are newly determined from recent geodetic observations and analyses. First, we compute the tidal response (namely, Q, k_2, k_3, and h_2) of our model Moon with a deep low-viscosity layer for the monthly, annual, triennial, and sexennial tidal periods. Second, we compare these theoretical values with observational ranges from GRAIL, LRO, and LLR (Konopliv et al., 2013; Lemoine et al., 2013; 2014; Mazarico et al., 2014; Williams and Boggs, 2015), and finally discuss possible implications from the revised constraints. Furthermore, we attach Supplementary material to this main article in order to demonstrate parameter sensitivity with regard to structure dependence of the tidal response in more detail, especially on the viscosity profile of the model Moon.

2. Parameters and method

We investigate the viscoelastic tidal deformation of the Moon in order to compare the resultant theoretical values with the recent observational ranges. We choose the well-known scheme to numerically estimate the Love numbers, that is, the y method (e.g., Alterman et al., 1959; Takeuchi and Saito, 1972). We skip the details because this method is widely adopted for the computation of global deformation of a solid planetary body in general, and is already explained in our previous study (Harada et al., 2014, Supplementary information).

We input density and elasticity profiles, and also a simplified viscosity profile that contains a low-viscosity zone in the deep lunar interior, so as to evaluate the possible effects of viscous relaxation. We set the same density, elasticity, and viscosity structure as in our last study (Harada et al., 2014). The reference model for the density and elasticity structure employed here was originally illustrated in Weber et al. (2011). One exception is the parameter range for the outer radius of the soft layer. Although we defined the outer radius from 450 to 500 km corresponding to its uncertainty in Weber et al. (2011) in our previous estimate, we redefine it from 500 to 600 km instead in the present estimate. Moreover, we demonstrate structure dependence on the model parameters of the viscosity profile other than the viscosity value and outer radius of the low-viscosity layer in Supplementary material. Whereas a few previous analyses have considered the possible effect of non-Maxwellian rheology (e.g., Efroimsky, 2012; Nimmo et al., 2012; Williams and Boggs, 2015), we follow the constitutive relations of a Maxwell body to create the complex elasticity as done earlier (e.g., Harada et al., 2014; Matsumoto et al., 2015). The use of a Maxwellian rheology to model the tidal response is discussed further in Nakada and Karato (2012).

We calculate the tidal response for different tidal periods so as to clarify the frequency-dependence of the viscoelastic deformation. In particular, we compute the tidal parameters for the monthly, annual, triennial, and sexennial periods. Following this, concerning Q, we compare each of the numerical predictions to the observational ranges obtained from LLR (Williams and Boggs, 2015). Concerning the frequency-dependence of k_2, k_3, and h_2, we assume that the observations represent the monthly tidal response, as in Williams and Boggs (2015) and Matsumoto et al. (2015) for k_2. This treatment may be justified at least for the GRAIL values by its mission duration, which was shorter than a year. Because of