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Explicitly modelled deep-time tidal dissipation and its implication for Lunar history

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ABSTRACT

Dissipation of tidal energy causes the Moon to recede from the Earth. The currently measured rate of recession implies that the age of the Lunar orbit is 1500 My old, but the Moon is known to be 4500 My old. Consequently, it has been proposed that tidal energy dissipation was weaker in the Earth's past, but explicit numerical calculations are missing for such long time intervals. Here, for the first time, numerical tidal model simulations linked to climate model output are conducted for a range of paleogeographic configurations over the last 252 My. We find that the present is a poor guide to the past in terms of tidal dissipation: the total dissipation rates for most of the past 252 My were far below present levels. This allows us to quantify the reduced tidal dissipation rates over the most resent fraction of lunar history, and the lower dissipation allows refinement of orbitally-derived age models by inserting a complete additional precession cycle.

1. Introduction

Tidally induced energy dissipation in the earth and ocean gradually slows the Earth's rotation rate, changes Earth and lunar orbital parameters, and increases the Earth–Moon separation (Darwin, 1899; Munk, 1968). A long-standing conundrum exists in the evolution of the Earth–Moon system relating to the present recession rate of the moon and its age: if present day observed dissipation rates are representative of the past, the moon must be younger than 1500 Ma (Hansen, 1982; Sonett, 1996). This does not fit the age model of the solar system, putting the age of the moon around 4500 Ma (Hansen, 1982; Sonett, 1996; Walker and Zahnle, 1986; Canup and Asphaug, 2001; Waltham, 2004), and the possibility that the tidal dissipation rates have changed significantly over long time periods has been proposed (Hansen, 1982; Ooe, 1989; Poliakov, 2005; Green and Huber, 2013; Williams et al., 2014). A weaker tidal dissipation must be associated with a lower recession rate of the moon. Consequently, it can be argued that prolonged periods of weak tidal dissipation must have existed in the past (Webb, 1982; Bills and Ray, 1999; Williams, 2000). There is support for this in the literature using quite coarse resolution simulations driven by highly stylized, rather than historically accurate, boundary conditions (Munk, 1968; Kagan and Sundermann, 1996). However, with the present knowledge of the sensitivity of tidal models to resolution and boundary conditions, e.g., the oceans density structure (Egbert et al., 2004), the results of prior work should be revisited with state-of-the-art knowledge and numerical tools.

It was recently shown through numerical tidal model simulations with higher resolution than in previous studies that the tidal dissipation during the early Eocene (50 Ma) was just under half of that at present (Green and Huber, 2013). This is in stark contrast to the Last Glacial Maximum (LGM, around 20 ka) when simulated tidal dissipation rates were significantly higher than at present due to changes in the resonant properties of the ocean (Green, 2010; Wilmes and Green, 2014; Schmittner et al., 2015). However, the surprisingly large tides during the LGM are due to a quite specific combination of continental scale bathymetry and low sea-level, in which the Atlantic is close to resonance when the continental shelf seas were exposed due to the formation of extensive continental ice sheets (Platzman et al., 1981; Egbert et al., 2004; Green, 2010). It is therefore reasonable to assume — and proxies support this — that the Earth has only experienced very large tides during the glacial cycles over the last 1–2 Ma and that the rates have been lower than at present during the Cenozoic (Palike and Shackleton, 2000; Lourens and Brumsack, 2001; Lourens et al., 2001). Such (generally) low tidal dissipation rates

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may have led to reduced levels of ocean mixing, with potential consequences for the large scale ocean circulation, including the Meridional Overturning Circulation (Munk, 1966; Wunsch and Ferrari, 2004).

The tidally induced lunar recession and increased day length also act to reduce the precession rate of Earth's axis and, as a result, produce falling rates of climatic precession and obliquity oscillation through time (Berger et al., 1992). As a direct consequence, cyclostratigraphy may be severely compromised because many important Milankovitch cycle periods are directly affected by the Earth–Moon separation. Nevertheless, Milankovitch frequencies have been estimated assuming either a constant lunar-recession rate or a constant tidal dissipation rate (Berger et al., 1992; Laskar et al., 2004). Based on the literature related to tidal evolution mentioned above, neither assumption is valid. For example, it was recently suggested that the tidal dissipation between 11.5–12.3 Ma was either at least 90% of the Present Day (PD) rate or 40% of the present rate, with the lower estimate obtained by shifting the precession a whole cycle (Zeeden et al., 2014). Constraining the tidal dissipation rates on geological time scales is consequently important. Investigating the tidal dynamics for select time slices over the Cenozoic era will shed light on the changes of tidal dissipation and hence on Earth–Moon system evolution.

Our aim in this paper is to answer the basic question: when considering the past, should our null hypothesis be that tidal dissipation was near modern values (the most common approach), much higher (suggested by LGM), or much lower (such as found for the Eocene)? We use the same tidal model as Green and Huber (2013), and we present results from simulations of the tidal dynamics for the PD, LGM (21 ka, Green, 2010), Pliocene (3 Ma), Miocene (25 Ma), Eocene (50 Ma, Green and Huber, 2013), Cretaceous (116 Ma, Wells et al., 2010), and for the Permian–Triassic (252 Ma). We explore dissipation changes across a wide cross-section of ocean states and palegeographic configurations, from the nearly modern to a world with one global ocean basin, and we investigate sensitivity to substantial imposed changes in ocean stratification. Consequently, this encompasses the likely range of continental and paleoclimate configurations over much of Earth's history.

2. Methods

2.1. Tidal modelling

The simulations of the global tides were done using the Oregon State University Tidal Inversion Software (OTIS, Egbert et al., 1994). OTIS has been used in several previous investigations to simulate global tides in the past and present oceans (Egbert et al., 2004; Green, 2010; Green and Huber, 2013; Wilmes and Green, 2014). It provides a numerical solution to the linearized shallow water equations,

$$\frac{\partial \mathbf{U}}{\partial t} + \mathbf{f} \times \mathbf{U} = -gH\nabla(\eta - \eta_{SAL} - \eta_{EQ}) - \mathbf{F} \quad (1)$$

$$\frac{\partial \eta}{\partial t} - \nabla \cdot \mathbf{U} = 0 \quad (2)$$

Here $U = uH$ is the volume transport given by the velocity u multiplied by the water depth H, f is the Coriolis parameter, η the tidal elevation, η_{SAL} the self-attraction and loading elevation, η_{EQ} the equilibrium tidal elevation, and $\mathbf{F}$ the dissipative term. Self-attraction and loading was introduced by doing 5 iterations following the methodology in Egbert et al. (2004). The dissipative term is split into two parts: $\mathbf{F} = \mathbf{F}_B + \mathbf{F}_W$. The first of these represents bed friction and is written as

$$\mathbf{F}_B = C_d \mathbf{u}|\mathbf{u}| \quad (3)$$

where C_d is a drag coefficient, and $\mathbf{u}$ is the total velocity vector for all the tidal constituents. We used $C_d = 0.003$ in the simulations described below, but for all time slices simulations were done where C_d was increased or decreased by a factor 3 to estimate the sensitivity of the model to bed roughness. This only introduced minor changes in the results (within a few percent of the control), and we opted to use the value which provided the best fir to observations for the present. The second part of the dissipative term, $\mathbf{F}_w = C\mathbf{U}$, is a vector describing energy losses due to tidal conversion. The conversion coefficient C is here defined as Green and Huber (2013)

$$C(x, y) = \gamma \frac{(\nabla H)^2 N_b \bar{N}}{8\pi\omega} \quad (4)$$

in which $\gamma = 100$ is a scaling factor, N_b is the buoyancy frequency at the sea-bed (taken from coupled climate model outputs), $\bar{N}$ is the vertical average of the buoyancy frequency, and ω is the frequency of the tidal constituent under evaluation. We did simulations with varying scaling factors (with $50 < \gamma < 200$) to cover the possible ranges of N, with only minor quantitative changes to the overall dissipation rates. This means that errors and uncertainties in the estimates of the buoyancy frequency from the climate model simulations will only change the quantitative results less than 10%.

The PD bathymetry is a combination of v.14 of the Smith and Sandwell database (Smith and Sandwell, 1997) with data for the Arctic (Jakobsson et al., 2012), northwards of 79°N, and Antarctic (Padman et al., 2002), southwards of 79°S. All data were then averaged to 1/4° in both latitude and longitude.

The PD control simulation is compared to the TPXO8 database, an inverse tidal solution for both elevation and velocity based on satellite altimetry and the shallow water equations (see Egbert and Erofeeva, 2002, and http://volkov.oce.orst.edu/tides/tpxo8_atlas.html for details). The root-mean-square (RMS) difference between the modelled and observed elevations is computed, along with the percentage of sea surface elevation variance captured, given by $V = 100[1 - (S/RMS)^2]$, where RMS is the RMS discrepancy between the modelled elevations and the TPXO elevations, and S is the RMS of the TPXO elevations.

The tidal dissipation, D, is computed using (Egbert and Ray, 2001):

$$D = W - \nabla \cdot P \quad (5)$$

in which W is the work done by the tide-producing force and P is the energy flux. They are defined as

$$W = g\rho\langle \mathbf{U} \cdot \nabla(\eta_{SAL} + \eta_{EQ})\rangle \quad (6)$$

$$P = g\rho\langle \eta \mathbf{U}\rangle \quad (7)$$

in which the angular brackets mark time-averages. When we discuss the accuracy and the energy dissipation rates we use a cutoff between deep and shallow water at 1000 m depth.

2.2. Earth–Moon separation

The tidal dissipation rate, D, should be (Murray and Dermott, 2010)

$$D = 0.5m'na(\Omega - n)\frac{\partial a}{\partial t} \quad (8)$$

where $m' = mM/(m + M)$, m is Moon-mass, M is Earth-mass, a is the Earth–Moon separation, Ω is the Earth's rotation rate and n is the lunar mean motion. The next step is to note that lunar recession is well approximated using (Lambeck, 1980; Bills and Ray, 1999; Waltham, 2015)