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Tidal Dissipation in the Early Eocene and Implications for Ocean Mixing

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Tidal dissipation in the early Eocene and implications for ocean mixing

J. A. M. Green¹ and M. Huber²

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1. Introduction

[1] The tidally driven vertical diffusivity in the abyssal ocean during the early Eocene (55 Ma) is investigated using an established tidal model. A weak tide is predicted in the Eocene ocean, except in the Pacific. Consequently, the integrated global tidal dissipation rate is a mere 1.44 TW, of which 40% dissipate in the Pacific. However, due to a stronger abyssal vertical stratification the predicted Eocene vertical diffusivities are consistently larger than at present. The results support the hypothesis that altered tidal dissipation may play a role in explaining the maintenance of past climate regimes, especially the anomalously warm temperatures in the southwest Pacific in the Eocene, and the low dissipation rates may be important for lunar evolution history. Citation: Green, J. A. M., and M. Huber (2013), Tidal dissipation in the early Eocene and implications for ocean mixing, Geophys. Res. Lett., 40, 2707–2713, doi:10.1002/grl.50510.

2. Modeling Eocene Tides

2.1. Oregon State University Tidal Inversion System

[6] The Oregon State University Tidal Inversion System has been used in several previous investigations to simulate global (barotropic) tides in the past and present oceans [e.g., Egbert et al., 2004; Green, 2010]. It provides a numerical solution to the linearized shallow water equations,
The (a) PD and (b) EE bathymetries (depth in meters with land grey).

\[
\frac{\partial U}{\partial t} + \mathbf{f} \times \mathbf{U} = -g\nabla \left( \eta - \eta_{\text{SAL}} - \eta_{\text{EQ}} \right) - \mathbf{F}
\]  

(1)

\[
\frac{\partial \eta}{\partial t} = -\nabla \cdot \mathbf{U}
\]  

(2)

[7] Here \( \mathbf{U} = \mathbf{u}/H \) is the depth-integrated volume transport given by the velocity \( \mathbf{u} \) multiplied by the water depth \( H \), \( \mathbf{f} \) is the Coriolis parameter, \( \eta \) the tidal elevation, \( \eta_{\text{SAL}} \) the self-attraction and loading elevation, \( \eta_{\text{EQ}} \) the equilibrium tidal elevation (which is not changed here—it should be about 1% larger than at present), and \( \mathbf{F} \) the dissipative term. The latter term is split into two parts, related to bed-friction and tidal conversion, respectively. We thus have \( \mathbf{F} = \mathbf{F}_b + \mathbf{F}_w \), where \( \mathbf{F}_b = C_d \mathbf{u} / H \) \((C_d \) is a drag coefficient, and \( \mathbf{u} \) is the total velocity vector for all the tidal constituents\) represents bed friction and \( \mathbf{F}_w = \mathbf{C} \mathbf{U} \) is a vector representing energy losses due to tidal conversion. There are several formulations for the conversion coefficient \( \mathbf{C} \), and here we use

\[
C(x, y) = \gamma (\nabla H)^2 \frac{N_b N}{\pi \alpha}
\]  

(3)

in which \( \gamma \approx 100 \) is a scaling factor, \( N_b \) is the observed buoyancy frequency at the seabed, \( N \) is the vertical average of the observed buoyancy frequency, and \( \alpha \) is the frequency of the tidal constituent under evaluation. Equation (3) is a modified version of the parameterization in Zaron and Egbert [2006], where the wave number \( k \) in the original expression has been replaced by \( k = \pi \omega / (\nabla H) \) [e.g., Kundu, 1990].

2.2 Bathymetry and Stratification

[8] The PD bathymetry (see Figure 1a) comes from a combination of v.14 of the Smith and Sandwell database [Smith and Sandwell, 1997], with Arctic (northward of 79°N) and Antarctic (southward of 79°S) bathymetries from the International Bathymetric Chart of the Arctic Ocean [Jakobsson et al., 2012] and Padman et al. [2002] data sets, respectively. All data are averaged to 1/4° in both latitude and longitude.

[9] The EE bathymetry (Figure 1b) is derived from the digital elevation models produced by Müller et al. [2008], and is also averaged to a 1/4° resolution and interpolated linearly to the same grid as the PD bathymetry. For computational reasons, both final bathymetries effectively ran from 90°S to 88.5°N in latitude due to the introduction of a vertical wall covering the North Pole. Note that global tidal simulations are relatively insensitive to small-scale topographic changes [cf. Egbert et al., 2004].

[10] Stratification for the tidal conversion scheme is taken from the World Ocean Circulation Experiment database for the PD case [Gouretski and Koltermann, 2004] and from the end of long equilibrated NCAR CCSM3 integrations for the EE cases. The latter match well with most early Eocene paleoclimatic proxy data with the exception of the southwest Pacific near New Zealand [Huber and Caballero, 2011; Hollis et al., 2012]. WOCE has a resolution of 1° whereas the EE output is on a variable resolution with 122 by 100 gridpoints in latitude and longitude, respectively. Both data sets are consequently interpolated to their respective bathymetric grids using linear interpolation.

2.3 Simulations and Computations

[11] A series of sensitivity simulations are conducted for both the PD and EE oceans for a number of constituents, although focus here is solely on the M2 because it dominates the tide. For the PD we evaluate four different simulations. The first is the “PD control” run, which used the modern tuned model setup without any other modifications. This is followed by the “+80 m SLR” simulation (with SLR denoting sea level rise), which uses the same setup as the control run but with PD sea-level raised 80 m, with flooding of land allowed. Note that there is no tidal conversion in these new cells because there is an obvious lack of knowledge about the stratification in them, and shallow water is usually well mixed. The “EE stratification” simulations include adjusting the PD buoyancy frequency, computed from the WOCE data, by the ratio between PD global averages and the EE global averages. This is carried out for \( N_b \) and \( N \), and shows significantly larger values of both variables for the EE: some 5.3 for \( N_b \) and 2.4 for \( N \). The final PD simulation is “stratification + SLR”, where the SLR and EE stratification cases are combined.

[12] For the EE we evaluate similar cases, starting with the “EE control”, which uses the same parameters as the PD control, but Eocene stratification and bathymetry. The “-80 m SLR” simulation uses the same setup as the EE control run but with the sea-level reduced with 80 m over the entire domain, whereas the “PD stratification” simulation is the equivalent to the EE stratification run described above, but with the EE stratification modified by the averaged strength of the PD stratification.

[13] The PD control simulation is compared to the TPXO7.2 database, an inverse tidal solution for both
### Table 1. Summary of the Results for the Different Simulations\(^a\)

<table>
<thead>
<tr>
<th>Run</th>
<th>Average Amplitude</th>
<th>Total Dissipation</th>
<th>Deep Dissipation</th>
<th>Pacific Total Dissipation</th>
<th>Pacific Deep Dissipation</th>
<th>Ratio Deep/Total Dissipation</th>
</tr>
</thead>
<tbody>
<tr>
<td>PD control</td>
<td>0.33</td>
<td>2.78</td>
<td>0.87</td>
<td>0.98</td>
<td>0.44</td>
<td>0.31</td>
</tr>
<tr>
<td>80 m SLR</td>
<td>0.29</td>
<td>1.78</td>
<td>0.08</td>
<td>0.51</td>
<td>&lt;0.01</td>
<td>0.04</td>
</tr>
<tr>
<td>EE stratification</td>
<td>0.14</td>
<td>1.90</td>
<td>1.64</td>
<td>0.89</td>
<td>0.80</td>
<td>0.86</td>
</tr>
<tr>
<td>80 m SLR, EE stratification</td>
<td>0.13</td>
<td>1.77</td>
<td>1.51</td>
<td>0.88</td>
<td>0.79</td>
<td>0.85</td>
</tr>
<tr>
<td>EE control</td>
<td>0.27</td>
<td>1.44</td>
<td>1.19</td>
<td>0.59</td>
<td>0.56</td>
<td>0.82</td>
</tr>
<tr>
<td>–80 m SLR</td>
<td>0.30</td>
<td>1.38</td>
<td>1.29</td>
<td>0.56</td>
<td>0.55</td>
<td>0.93</td>
</tr>
<tr>
<td>PD stratification</td>
<td>0.30</td>
<td>1.38</td>
<td>1.29</td>
<td>0.56</td>
<td>0.55</td>
<td>0.93</td>
</tr>
</tbody>
</table>

\(^a\)All dissipation rates are in TW, the average amplitudes are given in meters.

elevation and velocity based on satellite altimetry and the shallow water equations [see *Egbert and Erofeeva*, 2002]. Based on satellite altimetry and the shallow water equations [see *Egbert and Erofeeva*, 2002]. The RMS difference between the modeled and observed elevations is computed, along with the percentage of sea surface elevation variance captured, given by \(P_{var} = 100(1 - (S/RMS)^2)\), where RMS is the RMS discrepancy between the modeled elevations and the TPXO elevations, and \(S\) is the RMS of the TPXO elevations.

[14] The tidal dissipation, \(D\), is computed using [see *Egbert and Ray*, 2001, for details]

\[
D = W - \nabla \cdot P
\]  

[15] \(W\) is the work conducted by the tide-producing force and \(P\) is the energy flux, defined as

\[
W = g \rho \left( U \cdot \nabla \left( \eta_E \cdot \eta_Q + \eta_{\text{SAL}} \right) \right)
\]  

\[
P = g \rho (U \eta)
\]

in which the angular brackets mark time averages. Both the accuracy and the dissipation computations are presented as total values, i.e., using the entire data set, and deep water values, i.e., using only data where \(H > 1000\) m.

### 3. Tidal Dynamics

#### 3.1. Present-Day Tides

[16] The PD control simulation captures 91% of the tidal amplitude variance in the TPXO database, and the RMS difference between the modeled amplitudes and the TPXO-database is 12 cm (see Figure 2a and Table 1). However, it is known that the TPXO (and coarse-resolution tidal models) do not perform well near coastlines, and if we compute the RMS in deep water only we arrive at an RMS difference of a mere 7 cm—a decent result for this type of model. Also, the dissipation levels shown in Figure 3a are comparable to those calculated from the altimetry and presented by *Egbert and Ray* [2001]. Our model results give a total dissipation rate of 2.78TW for M2 of which 0.87GW dissipate in deep waters, whereas *Egbert and Ray* [2001], suggest 2.44TW in total and 0.8TW in deep ocean dissipation rates. Considering that *Egbert and Ray* [2001] only look at dissipation equatorward of 60°, we argue that the model reproduces both M2 elevations and dissipation rates comparable to those obtained from altimetry data.

#### 3.2. Eocene Tides

[17] The simulated Eocene tides show a very different picture compared to the present (Figures 2e–2g). The tide has almost vanished in the North Atlantic, around the coast of Africa and in the Indian Ocean. Consequently, the EE Pacific Ocean is the only major basin with a significant tide. Tides are qualitatively different in the Eocene compared with present, with a double amphidromic system in the North Pacific at present, but a single point in the Western part of the basin and another amphidrome in the equatorial east Pacific. As a consequence there is a significantly enhanced tide in the EE case in the southwest Pacific. The global mean tidal dissipation rate in the EE case is about half of those in the PD case (Figure 3 and Table 1)—1.44TW against 2.78TW in the model. A far larger fraction of this energy dissipates in the abyssal ocean in the EE case (1.18TW vs. 0.87TW, or 81% vs. 31%), and 41% of the total energy dissipates in the Pacific in the EE case, compared to 35% at present. In the EE case, 94% of the Pacific tidal energy is lost in the deep part of the basin (the corresponding PD value is 45%). Interestingly, the contribution by the Pacific to the overall dissipation rates is quite similar between the two cases but the surface area of the EE Pacific was almost 30% larger than at present, indicating weaker dissipation rates on average in the EE case.

[18] The obvious questions to ask are why the EE tidal amplitude is so much lower and why the total/deep dissipation ratios are so different? A clue can be found in the PD and EE perturbation simulations. The 80 m SLR case shows a large reduction of the tidal amplitudes (Figure 2b), and a subsequent reduction of the total dissipation to 1.78TW (Figure 3b). Furthermore, the rates in the abyssal ocean are less than an order of magnitude of the PD control ones: 0.08TW compared to 0.87TW—a physically important result. The mechanism is straightforward, and supports the results in *Egbert et al.* [2004] and *Green* [2010]: the extensive flooding of land leads to changed resonant properties of the basins, especially in the Atlantic, and reduces tidal amplitude and dissipation rates. Additionally, because of the significantly larger shallow area—nearly twice as much ocean is shallower than 200 m in the SLR run than in the control—an even larger fraction of energy than at present is lost there, draining the abyssal ocean of its energy. This also implies that the sea-level during the Eocene acted to reduce tidal mixing if one neglects the continental location changes. The opposite picture emerges from the simulations with Eocene stratification implemented in PD bathymetry (Figures 2c and 3c). This is not surprising because
the tidal conversion will then be on average 12.7 times larger than at present due to the enhanced EE values of the buoyancy frequency. The increase in the conversion coefficient leads to a dissipation of 1.9TW in total and 1.64TW in the abyssal ocean in the PD case with EE stratification. Instead of the energy being lost at the energetic shelf seas, it is extracted in the abyssal ocean due to an enhanced tidal conversion, and there is simply no energy left as the tides approach the continental shelf in the EE case simulations.

Combining 80 m SLR and the EE stratification leads to a total dissipation rate of $M_2$ energy of 1.77TW, of which 1.51TW is lost in the abyssal ocean (Figures 2d and 3d). This strongly supports an enhanced tidally driven abyssal mixing during the Eocene due to the stronger vertical stratification, although the largely increased sea level would have led to a reduced overall tidal dissipation. Simulating the EE using PD stratification supports the claims made earlier: the overall dissipation is relatively small because of the large-
scale configuration of the continents, but the abyssal dissipation is quite large because of the vertical stratification.

3.3. Vertical Diffusivities

An estimate of the vertical diffusivity $k_z$ can be computed following St Laurent et al. [2002]:

$$k_z = \frac{\Gamma \rho D(x,y)F(z)}{\rho N^2}$$  \hspace{1cm} (7)

Here $\Gamma$ is a mixing efficiency usually taken to be 0.2, $q = 0.3$ is the fraction of energy dissipating locally, $D$ is the dissipation rates calculated using equations (4)–(6), $\rho = 1040$ kg m$^{-3}$ is a reference density, and $F(z)$ is a vertical redistribution function of the energy given by

$$F(z) = \frac{e^{-(H+z)/\zeta}}{\zeta(1-e^{-H/\zeta})}$$  \hspace{1cm} (8)

In equation (8) $\zeta = 500$ m is a vertical decay scale for the dissipation. Note that when used in an ocean model, a background diffusivity is added to equation (7) to account for other processes and for energy propagating in from other grid cells [e.g., Simmons et al., 2004; see also Polzin, 2009, and Declaedd and Luther 2012, for discussions].

The associated vertical diffusivities (Figures 3c and 3d) show a significantly enhanced $k_z$ in the EE simulations at most depths, especially in the Pacific. This is because of the enhanced tidal dissipation rates in the abyssal Pacific, which dominates over the changes in buoyancy frequency in equation (7). This enhanced dissipation in the Eocene Pacific may partially explain some enigmatic features of this interval: for example, why the Eocene abyss has less organic carbon burial than any other greenhouse interval (unlike the similarly warm Cretaceous characterized by ocean anoxic events) as described in Lyle [1997]. It may also explain the anomalously weak temperature gradients found in the region near New Zealand in the Eocene, which could be a consequence of an enhanced Pacific MOC or poleward penetration of an East Australian Current driven by strong dissipation-induced watermass transformation in this region [Bijl et al., 2009; Hollis et al., 2012].

The overall contribution of the deep ocean to heat transport in the modern ocean may be modest [e.g., Boccaletti et al., 2005; Ferrari and Ferreira, 2011], but it may have played a greater role in the Eocene because of differences in basin geography. Today, the global circulation and structure is largely set up by the presence of the Antarctic Circumpolar Current system, and there may only be a limited need for mechanical energy inputs to maintain the relatively weak abyssal stratification [e.g., Toggweiler and Samuels, 1998; Vallis, 2000; Gnanadesikan et al., 2005]. In the Eocene ocean, however, the basics of the ocean circulation are different, with far weaker upper ocean temperature gradients from 60°N to 60°S latitude than at the present and very different thermal structure due to the lack of a circumpolar current. Because the present-day high-latitude upwelling and isopycnal (vertical) mixing systems are missing in the Eocene ocean, the abyssal stratification is very sensitive to the assumed abyssal dissipation because there is no other means to remove dense water from the abyss [e.g., Munk, 1966]. In the absence of diapycnal abyssal mixing, there should be a
weak abyssal stratification [Toggweiler and Samuels, 1998], but this is not what is observed in the Eocene oceans [e.g., Bijl et al., 2009; Hollis et al., 2012]. Consequently, Eocene ocean heat transport estimates are likely to have been much more sensitive to the abyssal stratification, which is a direct function of abyssal mixing rates. Also, enhanced mixing drives a stronger meridional overturning circulation and this has huge importance for understanding past oceanography, especially in the area of biogeochemical cycling.

4. Summary

The present paper aims to describe the tidal dynamics, with a focus on the dissipation rates of the semidiurnal tide, in simulations representing the early Eocene. It is shown that the tides are significantly reduced in the EE simulations due to large-scale changes in bathymetry. This is especially true for the North Atlantic, in which only minor tides are found, and can be explained by the near-resonant properties of the PD Atlantic. It further supports previous results that increasing the horizontal area of the shelf seas in the North Atlantic leads to a decreased dissipation there [e.g., Green, 2010]. The EE abyssal dissipation rates, however, are twice that of the PD case due to a stronger abyssal vertical stratification in the Eocene case. Furthermore, the abyssal Pacific is especially energetic in the EE case, with potentially significant implications for the global overturning circulation and its relevance for climate [see also Thomas et al., 2008; Hague et al., 2012].

The study here is made at relatively coarse resolution, but the PD simulations still show a decent capacity in recreating the (deep water) tides (see section 3.1). The diffusivity scheme is also a quite coarse approximation, but the purpose here is to illustrate that there was an enhanced diffusivity during the EE, rather than providing exact values. If these results are to be used in an actual climate simulation, it may well be worth using another scheme, or a more careful investigation of the parameter values [e.g., Saenko et al., 2012].

The current generation of Earth System models can include an explicit tidal dissipation mixing calculation using the parameterization of Jayne [2009], but the tide-induced diffusivity must be calculated ahead of time. This study represents a first effort to do that in pre-Quaternary paleoclimates. The present results indicate the importance of the paleo-Pacific in the climate system, and they have again pointed to the potential contribution of abyssal tidal dissipation to the Earth System. The results agree with Green et al. [2009] and Lund et al. [2011] in stressing the importance of understanding abyssal mixing rates in past climates. Furthermore, a major mystery in lunar evolution is that past tidal dissipation rates overall must have been much weaker than at present [e.g., Hansen, 1982; Sonett et al., 1996]: with the present tidal dissipation rates the Moon must be younger than 1500 Ma, which does not fit the age model of the Earth-Moon system. Because a weaker tidal dissipation is associated with a lower recession rate of the Moon, prolonged periods of weak tidal dissipation must therefore have existed. Here concrete evidence is provided of a far weaker tidal dissipation rate in the past and further support for the notion that the present dissipation rates are anomalously high [e.g., Bills and Ray, 1999].

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