Provided for non-commercial research and education use.
Not for reproduction, distribution or commercial use.

Volume 271, Issues 1–4
15 July 2008
ISSN 0012-821X

EARTH
& PLANETARY
SCIENCE LETTERS

Editors
Rick Carlson, *Washington, DC*
Peggy Delaney, *Santa Cruz, Calif.*
Claude Jaupart, *Paris*
Peter B. deMenocal, *Palisades, NY*
Tilman Spohn, *Berlin*
Lars Stixrude, *London*
Rob van der Hilst, *Cambridge, Mass.*

Frontiers Editor
Albrecht W. Hofmann, *Mainz*

This article appeared in a journal published by Elsevier. The attached copy is furnished to the author for internal non-commercial research and education use, including for instruction at the authors institution and sharing with colleagues.

Other uses, including reproduction and distribution, or selling or licensing copies, or posting to personal, institutional or third party websites are prohibited.

In most cases authors are permitted to post their version of the article (e.g. in Word or Tex form) to their personal website or institutional repository. Authors requiring further information regarding Elsevier's archiving and manuscript policies are encouraged to visit:

http://www.elsevier.com/copyright

Earth and Planetary Science Letters 271 (2008) 101–108

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

journal homepage: www.elsevier.com/locate/epsl

Dynamic topography and long-term sea-level variations: There is no such thing as a stable continental platform

Robert Moucha [a,*], Alessandro M. Forte [a], Jerry X. Mitrovica [b], David B. Rowley [c], Sandrine Quéré [a,d], Nathan A. Simmons [e,f], Stephen P. Grand [f]

[a] *GEOTOP, Université du Québec à Montréal, Montréal, QC, CANADA H3C 3P8*
[b] *Department of Physics, University of Toronto, ON, Canada M5S 1A7*
[c] *The Department of the Geophysical Sciences, University of Chicago, Chicago, IL, 60637, USA*
[d] *Now at: UFR des Sciences et Techniques Laboratoire de Planétologie et Géodynamique, Université de Nantes, France*
[e] *Now at: Atmospheric, Earth & Energy Division Lawrence Livermore National Laboratory, Livermore, California 94551, USA*
[f] *Jackson School of Geological Sciences, University of Texas at Austin, Austin, Texas 78712, USA*

ARTICLE INFO

Article history:
Received 4 December 2007
Received in revised form 24 March 2008
Accepted 25 March 2008
Available online 10 April 2008

Edited by: M.L. Delaney

Keywords:
eustasy
sea level
epeirogeny
mantle convection
seismic tomography
dynamic topography
uplift rates

ABSTRACT

Globally uniform (i.e., eustatic) sea-level trends with time scales approaching 100 Myr have been inferred from both seismic and backstripping stratigraphic analysis at a small set of geographic sites that are presumed to lie on stable continental platforms and passive continental margins characterized by simple thermal subsidence histories (e.g., the New Jersey margin, the western African margin). We demonstrate, using mantle flow simulations based on high resolution seismic tomography, that both the New Jersey margin and the conjugate western African margin have been subject to orders of 100 m of dynamic (i.e., flow induced) topography change over the last 30 Myr. We also show that the changing pattern of downwelling mantle flow associated with plate subduction is a significant contributor to the background eustatic sea-level trend, which is also of order 100 m during the 30 Myr time window. Therefore, Late-Cenozoic variations of dynamic topography on these passive margin sites are comparable to the eustatic sea-level changes and can partially mask the latter. Furthermore, even if the observed trend could be accurately corrected for local dynamic topography variation, the residual eustatic signal does not merely reflect changes in mean spreading rates at mid-ocean ridges. We conclude that the observed long term sea-level variations at so-called “stable” sites cannot be interpreted as eustatic. Moreover, previous analyses that have used long-term sea-level trends as a proxy for spreading rates and geochemical fluxes must be revisited.

© 2008 Elsevier B.V. All rights reserved.

1. Introduction

Stratigraphic analyses of seismic data from continental platforms and passive continental margins reveal global-scale sea-level variations spanning a wide range of time scales (e.g., Vail et al., 1977b; Hallam, 1978; Haq et al., 1987). The origins of some aspects of this variation are well known, for example the sea-level changes that accompanied the growth and ablation of grounded ice sheets during the Pleistocene (Milne et al., 2002) and older times (e.g. Miller et al., 2005). In contrast, the origin of longer time scale sea-level trends, the first and second order sea-level cycles (e.g., Vail et al., 1977b), which we define here as lasting 10–100 million years, is the source of increasing debate (e.g. Carter, 1998; Miall and Miall, 2001).

Under the widespread assumption that these long-term trends are globally synchronous, that is, eustatic, in nature, they are generally interpreted to reflect variations in the mean rate of spreading (production of oceanic crust) at mid-ocean ridges (Hays and Pitman, 1973). According to this view, an increase (decrease) in rate of spreading leads to a higher (lower) mean topography of ocean basins and a displacement upwards (downwards) of sea-level at coastlines (Pitman, 1978). Advocates of this connection cite the apparent correlation between the time history of plate spreading, as inferred from magnetic reversal records, and the variation in continental flooding from the Cretaceous to Present.

Arguments countering various elements of this “textbook” view of long-term sea-level trends have appeared within the literature. First, a recent reanalysis of an updated ocean age distribution (Rowley, 2002) does not require a change in the mean spreading rate over the last 180 Ma, though other analyses question this conclusion (Demicco, 2004; Conrad and Lithgow-Bertelloni, 2007). Second, various simulations of mantle convective flow indicate that the spreading rate-sea level connection neglects a crucial element of the plate tectonic system, namely subduction (Hager, 1980; Mitrovica et al., 1989; Gurnis, 1990). Mitrovica et al. (1989) and Gurnis (1990) have shown

---

* Corresponding author.
E-mail address: rmoucha@gmail.com (R. Moucha).

0012-821X/$ – see front matter © 2008 Elsevier B.V. All rights reserved.
doi:10.1016/j.epsl.2008.03.056

that viscous stresses driven by plate subduction can drive topographic (and hence apparent sea-level) variations in the overriding lithosphere. An increase in subduction rate, or the subduction of progressively older (colder) lithosphere, will lead to a dynamic subsidence of the overriding lithosphere and the associated infill of water to this area may, at least in part, compensate for the sea-level rise associated with an increase in the spreading rate. There may be important phase lags between the sea-level signal from spreading and subduction rate changes (Husson and Conrad, 2006); however, the link between plate subduction and tilting of continental margins (i.e., regional transgressions and regressions of sea level) is widely established (Mitrovica et al., 1989; Gurnis, 1990).

It is now also widely recognized that mantle flow drives significant vertical motions of the crust, or “dynamic topography”, across the entire surface of the planet (e.g., Hager et al., 1985; Forte et al., 1993; Lithgow-Bertelloni and Silver, 1998; Gurnis et al., 1998; Forte and Perry, 2000; Daradich et al., 2003). This general coupling raises a fundamental question regarding the nature of long-term sea-level trends; namely, are these trends, as has long been assumed, eustatic?

Methods have been proposed for separating eustasy from the signal associated with dynamic topography (epeirogeny) that are based on either the geometry of the associated stratigraphy (Mitrovica et al., 1989) or a comparison of the flooding histories of different continents (Bond, 1979; Harrison, 1990). However, to date, long-term (first and second order cycles) sea level variations are based on inferences derived from continental flooding (e.g. Sloss, 1963; Vail et al., 1977a) or from modeling of subsidence histories at passive continental margins, such as New Jersey (e.g. Watts and Steckler, 1979; Miller et al., 2005) or northwest Africa (Hardenbol et al., 1981). In the case of the Exxon global sea level curves (Vail et al., 1977b; Haq et al., 1987), the higher frequency (third and higher order cycles) inferred sea level variations, are based on seismic stratigraphy of coastal margins of which a majority is restricted to the northern hemisphere which include margins near tectonically active zones (e.g. Alaska, California, Sumatra, Philippines, and Japan).

The first-order variations in the Exxon curves were scaled to match 1-D backstripping estimates of a well on a northwest-African margin by Hardenbol et al. (1981). Whereas, the short-term history and amplitude variations incorporated Mesozoic and Cenozoic sequences derived from marine outcrops located in the western interior of the United States and the Gulf of Mexico. As discussed above, these areas were subject to large scale dynamic topography variations that would have contributed significantly to the inferred sea-level trends. The absolute height of the long-term Exxon curve representing the late-Cretaceous high sea-level stand was scaled to match the effects due to change in ridge spreading rates, among other factors based on Harrison (1990).

Backstripping of one or more wells to derive long-term sea level variations have, to date, been restricted to a small number of sites (New Jersey, Nova Scotia, and northwest Africa). The presumption that the sea-level trends represent eustatic signals (after correction for effects such as sediment compaction, loading and thermal tectonic subsidence) is based on the assumption that these sites, which are distant from the closest plate boundary, are located on stable (passive) continental platforms unaffected by other, unaccounted for, sources of vertical components of motion.

In light of the results of global mantle flow calculations (e.g., Forte et al., 1993; Forte and Perry, 2000), it is logical to question whether there exists, anywhere on Earth, a continental platform that is stable on times scales of 10–100 Myr. Or, more specifically, whether the key sites used to constrain long-term sea-level changes are located on platforms in which eustasy is expected to dominate dynamic topography (epeirogeny) on these time scales. In this paper we explore this issue using mantle convection simulations initiated by 3-D mantle structure based on seismic tomographic imaging.

1.1. Mantle convection and dynamic topography

Dynamic topography is defined as the response of the Earth's surface to convectively-maintained vertical stresses (Mitrovica et al., 1989). These vertical stresses originate from buoyancy residing in both the lithosphere and the mantle (e.g. Forte et al., 1993), including those associated with the cooling of the oceanic lithosphere (e.g. Parsons, 1982). We note that the cratonic lithosphere is chemically distinct from the oceanic lithosphere, and is neutrally buoyant (Forte and Perry, 2000; Simmons et al., 2007).

To estimate the present-day dynamic topography induced by mantle convection, and its evolution, we employ a spherical, 3-D, time-dependent, compressible mantle convection model. In our simulations, the mantle flow is dynamically coupled with surface tectonic plates (i.e. the plate motions are not used to drive the flow; rather they are predicted by the flow [Forte and Peltier, 1994]) and its time-dependent evolution is solved using a pseudo-spectral numerical method (Espesset and Forte, 2001). We incorporate Newtonian rheology with a viscosity profile that is constrained by global joint inversions (Mitrovica and Forte, 2004) of convection-related surface observables (surface gravity anomalies, residual topography, divergence of tectonic plate motions, excess ellipticity of the core–mantle boundary) and data associated with the response of the Earth to ice-age surface mass loading (decay times inferred from post-glacial sea-level histories in Hudson Bay and Fennoscandia, and the Fennoscandian relaxation spectrum). This profile, shown in Fig. 1, is characterized by a three order of magnitude increase in viscosity from the base of the lithosphere to a depth of ~2000 km, followed by a decrease toward the core–mantle boundary; we term this profile ‘V1’. Additionally, because dynamic topography and its rate of change are sensitive to both buoyancy forces used to drive mantle convection and to the adopted viscosity profile which controls mantle flow velocity, we consider a second viscosity profile in our study termed ‘V2’ in Fig. 1. The V2 viscosity profile is also consistent with the aforementioned geodynamic constraints, though it has a reduced low viscosity notch at the base of the upper mantle and is overall slightly stiffer than the V1 viscosity profile, specifically in the lower mantle.

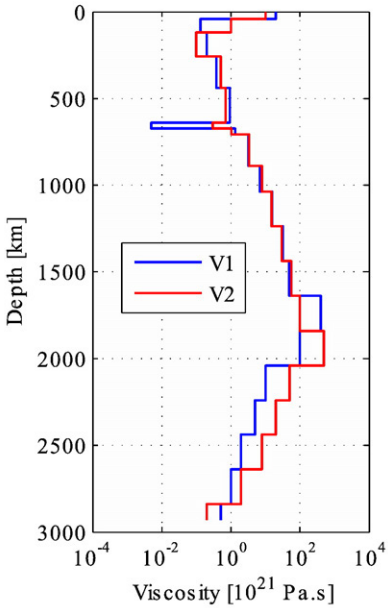

Fig. 1. Mantle viscosity as a function of depth used in the simulations. The viscosity profile labeled V1 is from Mitrovica and Forte (2004). The model V2 is a slightly modified version of V1; it is characterized by a moderate increase in viscosity of the lithosphere and lower mantle with a less pronounced low viscosity notch at the base of the upper mantle.

Three-dimensional density perturbations in the mantle are derived from a high resolution global seismic shear-wave model that is parameterized into spatial blocks defined by 22 layers extending from the core–mantle boundary to the surface, with thicknesses ranging from 75 km to 150 km, and with lateral dimensions of ~250 km (Simmons et al., 2007). This model is the first to be obtained through a joint inversion of global seismic (travel times) and geodynamic surface observables (see above) in which both thermal and compositional contributions to density are simultaneously considered. These compositional contributions are crucial for an adequate representation of the intrinsic buoyancy in the sub-cratonic root of North America (Forte et al., 2007). The resulting model of mantle heterogeneity satisfies both seismic constraints and geodynamic constraints equally well. Specifically, a mantle convection simulation driven by this model of mantle heterogeneity fits the observed free-air gravity, residual topography, and plate divergence (up to spherical harmonic degree 16) to a variance reduction of 90%, 94%, and 76%, respectively. The seismic travel time constraints are satisfied to within 96%.

The resolution inherent to this model enables us to perform flow calculations with much greater spatial resolution than previously possible. All mantle flow results presented below are based on spherical harmonic representations up to maximum degree 128 (~156 km half-wavelength at surface), thereby allowing us to fully resolve the fine-scale details in the tomography model. The radial variation in the mantle flow field is parameterized in terms of Chebyshev polynomials up to order 65, corresponding to a maximum radial flow resolution of ~70 km at mid-mantle depth.

A time-dependent reconstruction of dynamic topography in the past is obtained via "backward" mantle convection simulations. We adopt the approach used by Forte and Mitrovica (1997) and by others (e.g., Steinberger and O'Connell, 1997; Conrad and Gurnis, 2003) where the direction of buoyancy induced flow in the mantle is numerically reversed by using negative time. Naturally, the advective term in the heat-energy equation is a time-reversible process, whereas the diffusive portion is not. However, for a relatively short simulation period of 30 Myr, the diffusion of heat in the mantle is very small compared to advection and therefore, to first order, the diffusive term in the heat-energy equation can be explicitly set to zero. To this end, the initial (present-day) heterogeneity is advected backwards using boundary conditions that are consistent with a new Indo-Atlantic plate reconstruction model in the no-net-rotation reference frame (Quéré et al., 2007).

To be consistent with geological markers of sea-level change, sea level at a specific geographic point is defined as the height of the sea-surface, or geoid, relative to the solid surface of the planet. Thus local changes in sea level arise from variations in the radial position of either of these two bounding surfaces. Thus, in our backward advection scheme we keep track of fluctuations in both the solid surface (i.e., dynamic topography variations) and the geoid.

Geoid height changes have two contributions. The first is associated with gravitational potential perturbations arising from both the advection of density heterogeneity and deflections of boundaries (including the surface) in the solid Earth. This is the usual definition of the geoid in the mantle convection literature. The second contribution is a uniform shift in the geoid that reflects the fact that the equipotential defining the sea surface may change in time (Dahlen, 1976), and this term may be computed by invoking conservation of mass within the oceans. Specifically, if we assume a constant volume of water in the oceans, and ignore any sedimentation, the uniform shift in the geoid will be equal to the change in the mean dynamic topography within the ocean basins. That is, the mean geoid must rise (or fall) to account for a net increase (decrease) in the volume of ocean basins driven by convective flow.

This shift in the geoid will be termed the eustatic sea-level change. It is important to note that this change does not imply a variation in the volume of ocean water. The eustatic sea-level change, as we have defined it, does, however, have a clear physical explanation; it is the sea level rise (or fall) that would be measured by an observer on a fixed edge of the ocean basin (e.g. on a perfectly passive margin), subject to no local geoid perturbations.

Following Moucha et al. (2006), we estimate the change in the volume of ocean basins as a function of time by integrating the dynamic topography over the geometry of these basins. This surface integration is performed via a 1st order numerical integration on a very fine (sub-degree) unstructured spherical triangular mesh where each ocean basin is defined by its own polygon. The surface area coverage of ocean basins as a function of time was obtained by backward-rotation of present-day continental blocks using the same Indo-Atlantic plate reconstruction model as used in the backward mantle convection simulations.

2. Dynamic topography, eustasy and relative sea level

Two examples of predicted dynamic topography, at 0 Ma and 30 Ma, are shown in Fig. 2 for a simulation adopting the V2 viscosity profile. The change in the volume of the ocean basins over the last 30 Myr appears to be dominated by a change in the volume of the east Pacific ridge system and an uplift of the lithosphere in vicinity of various subduction zones (e.g., Sumatra, Indonesia, and southwestern South America). To quantify these contributions, we have computed the change in eustatic sea level over this time period for simulations based on both the V1 and V2 viscosity profiles by integrating the dynamic topography over the ocean basins at each 5 Myr instance back to 30 Ma. The change in the area of the ocean basins over this time period is negligible. These results are presented in Fig. 3(a). In both cases, our convection simulation predicts a eustatic sea level rise in the last 30 Myr. As one would expect, the viscosity profile controls the mean rate of uplift; in particular, the stiffer viscosity profile V2 corresponds to a sea level rise of ~100 m in the last 30 Myr, whereas the viscosity V1 profile predicts ~140 m of sea level rise. (We emphasize that we are computing sea level change driven by mantle convection alone; we are not including effects such as changes in sedimentation flux into the ocean basins or the formation of ice sheets in Antarctica and Greenland. The formation of permanent ice sheets began at about 30 Ma, and is estimated to be responsible for about a 40 m eustatic sea level fall over this time period [Harrison, 1990]).

To investigate the effect that thermal diffusion in the lithosphere has on our inferences of eustatic sea level change we remove the upper 100 km of heterogeneity in the mantle and compute the resulting dynamic topography. In essence, we are then assuming that the thermal structure of the lithosphere, which is controlled by a balance between horizontal advection and vertical conduction (Turcotte and Oxburgh 1967), is effectively unchanging or in steady-state over this 30 Myr time interval. On the basis of this assumption, any change in sea level due to variations in dynamic topography of the ocean floor will be driven entirely by convective flow below the lithosphere. This result is labeled with the suffix "100 km" in Fig. 3(a) and compared to results that include the upper 100 km of (lithospheric) mantle heterogeneity. We thereby obtain an approximate estimate of the uncertainty associated with the neglect of thermal diffusion in the lithosphere that is about 15%.

Moreover, we have decomposed the V2 results in Fig. 3(a) into contributions from both uplift and subsidence of the solid surface in the ocean basins. Specifically, the net contribution to the eustatic sea-level change over the last 30 Myr from all regions showing a dynamic topography high (i.e., yellows and reds) in Fig. 2 (bottom) is a rise of 40 m, while the contribution from regions that are dynamically depressed in this same figure is a rise of 60 m. We conclude that variations in eustatic sea level are not, as is generally assumed, solely driven by changes in the volume of mid-ocean ridge structures; that is, as our qualitative inspection of Fig. 2 suggested, the contribution from

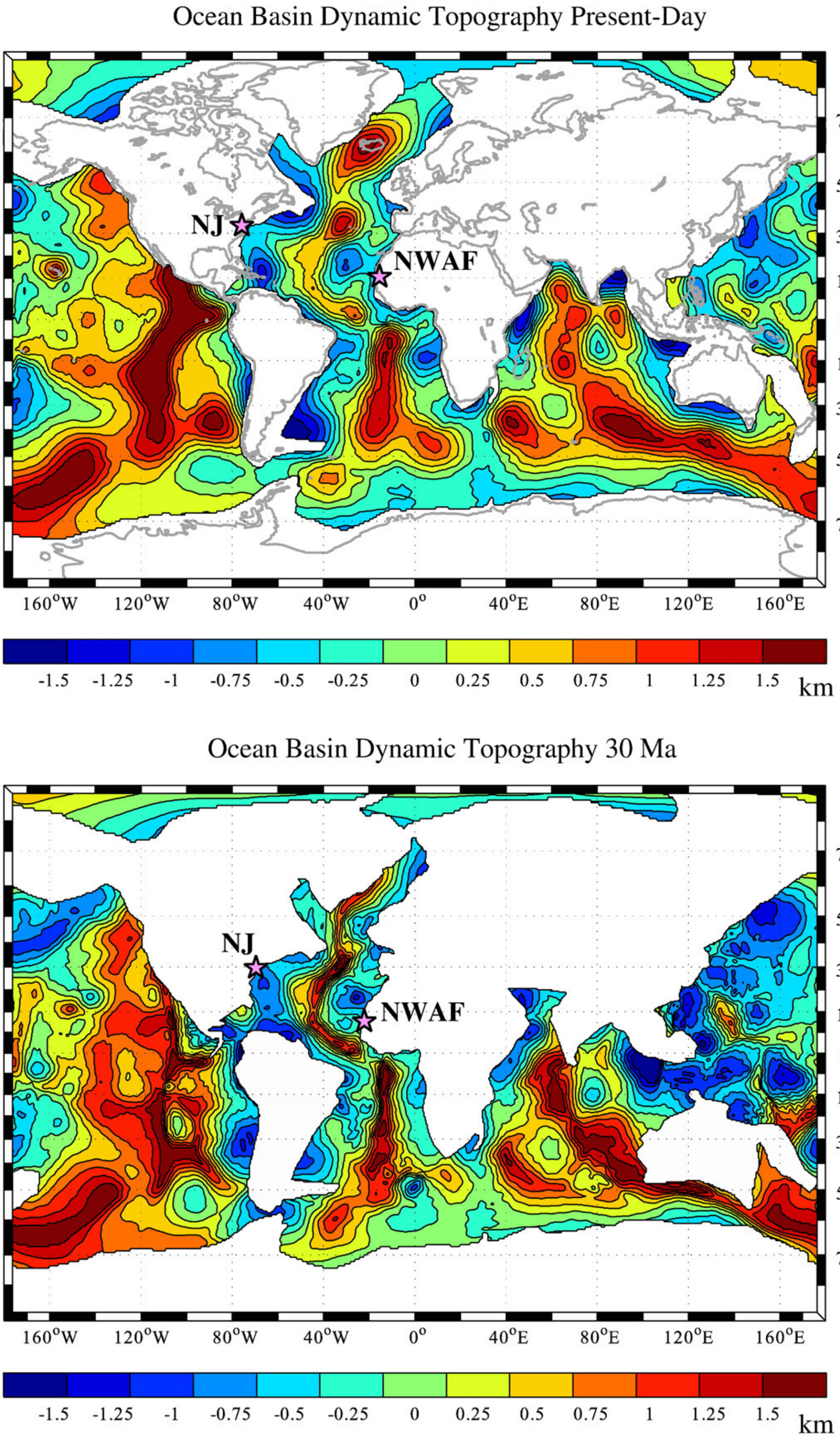
Ocean Basin Dynamic Topography Present-Day
NJ
NWAF
72°N
54°N
36°N
18°N
0°
18°S
36°S
54°S
72°S
160°W
120°W
80°W
40°W
0°
40°E
80°E
120°E
160°E
-1.5
-1.25
-1
-0.75
-0.5
-0.25
0
0.25
0.5
0.75
1
1.25
1.5
km
Ocean Basin Dynamic Topography 30 Ma
NJ
NWAF
72°N
54°N
36°N
18°N
0°
18°S
36°S
54°S
72°S
160°W
120°W
80°W
40°W
0°
40°E
80°E
120°E
160°E
-1.5
-1.25
-1
-0.75
-0.5
-0.25
0
0.25
0.5
0.75
1
1.25
1.5
km

Fig. 2. Modeled ocean basin dynamic topography for viscosity profile V2 (Fig. 1) at present day (top) and 30 Ma (bottom); the latter with reconstructed continental shelf positions on a present-day graticule. The continental blocks were backward rotated using an Indo-Atlantic no-net-rotation plate reconstruction model (Quéré et al., 2007). Also shown are the locations of regional relative sea level sites discussed in the text: the New Jersey coastal plain (NJ) and the northwest African margin (NWAF).

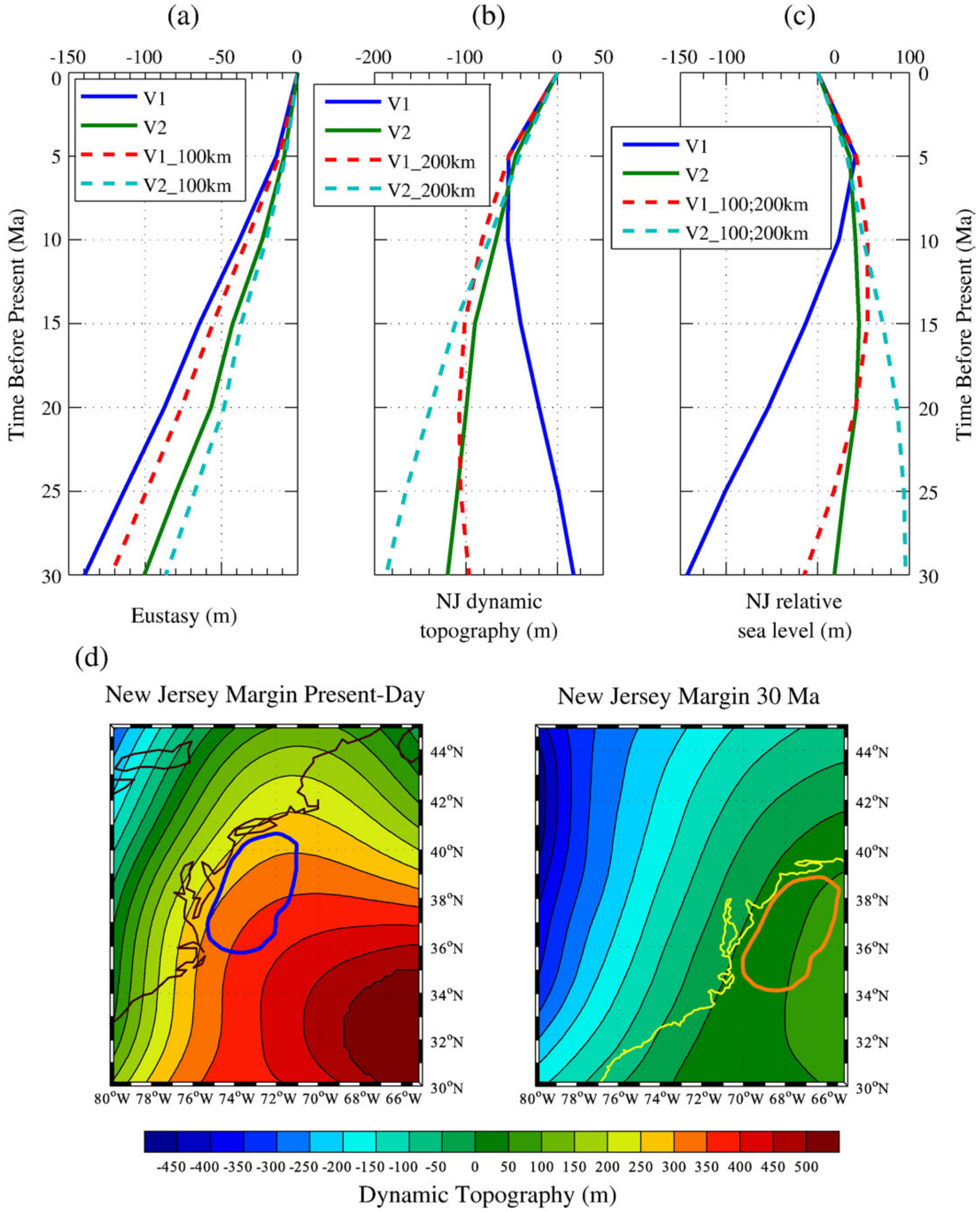

Fig. 3. (a) Eustatic sea level (relative to the present-day value) over the last 30 Myr obtained by integrating the dynamic topography over the full ocean basin geometry (water–mantle density contrast 2.2 g/cm^3). The two curves refer to simulations based on the viscosity profiles V1 and V2, as shown in Fig. 1. The curves denoted by the suffix "100 km" refers to results obtained with the upper 100 km heterogeneity set to zero (see text). (b) Dynamic topography (i.e., convection-induced radial displacement of the solid surface) at a site on the New Jersey (NJ) coastal plain (see Fig. 2) relative to present day (air–mantle density contrast 3.2 g/cm^3). Results are shown for both the V1 and V2 viscosity profiles, as well as cases in which all lithosphere density heterogeneity within the upper 200 km is set to zero (denoted by suffix "200 km"). (c) Modeled relative sea level variation at the New Jersey coastal plain. RSL is computed from the predicted eustatic sea level trend (frame a) and the predicted difference between changes in the radial position of the local geoid and solid surfaces (see text). The suffix "100;200 km" corresponds to the combination of eustatic sea level change with suffix "100 km" in (a) and solid surface change with suffix "200 km" in (b). (d) Present-day dynamic topography for model V2_200 km at the central east margin of USA compared to dynamic topography at 30 Ma. Also shown is the NJ area of integration used for results in (b). The coast line and area of integration were backward rotated using an Indo-Atlantic no-net-rotation plate reconstruction model (Quéré et al., 2007).

dynamic topography fluctuations in the vicinity of subduction zones cannot be ignored.

Of course, the results in Fig. 2 also indicate that the notion that there exist perfectly stable continental margins that serve as natural markers for eustatic sea-level change may be overly simplistic. Our goal in the following is to quantify the impact of local dynamic topography fluctuations on estimates of sea-level change at putatively stable sites that have played a role in past efforts to constrain long-term eustatic sea-level change.

Detailed backstripping inferences of global sea level change from the presumed passive margin off eastern North America, including sites on the coastal plains of New Jersey and Nova Scotia, were carried out by Watts and Steckler (1979). Analyses of boreholes on the New Jersey coastal plain have been the most prominent focus of attention over the past three decades (Steckler and Watts 1978, Van Sickel et al., 2004; Miller et al., 2005) and they have been interpreted as indicating a fall in global eustatic sea level since the Cretaceous. Our prediction of the dynamic topography over the past 30 Myr, relative to the present day, at this site is shown in Fig. 3(b) for simulations using both viscosity profiles V1 and V2. We may also explore the impact of the stability of the continental lithosphere on dynamic topography by removing all heterogeneity above a depth of 200 km and repeating the simulations (results denoted by the term “200 km”). These results provide a measure of the variability in the predictions associated with uncertainties in lithospheric density within continental regions.

We compare the present-day dynamic topography of the New Jersey margin with the dynamic topography at 30 Ma in map form in Fig. 3(d). This comparison illustrates the effect of the progressive descent of the Farallon slab which is located at shallower depth at 30 Ma relative to present-day and hence produced a more pronounced dynamic topography low. Between 30 Ma and today, the Farallon slab inclination steepened and descended deeper into the mantle thereby driving a sublithospheric influx of warmer suboceanic mantle under the east coast of North America from the Atlantic. This increase in positive buoyancy in the asthenospheric mantle produces the uplift evident in Figs. 3(b) and (d). The mantle flow geometry produced by Farallon slab subduction under eastern North America is described in greater detail in the recent study by Forte et al. (2007).

The results in Fig. 3(b) indicate that the variations in dynamic topography are comparable to eustatic sea-level changes over the last 30 Myr, and thus, at least in this regard, the New Jersey coastal plain cannot be considered a stable continental platform. This implies that analyses of geological markers of sea-level change at this site will not provide a direct estimate of the eustatic sea-level trend. This is reinforced in Fig. 3(c), where we plot predicted relative sea-level (RSL) curves for this site generated from the same four numerical

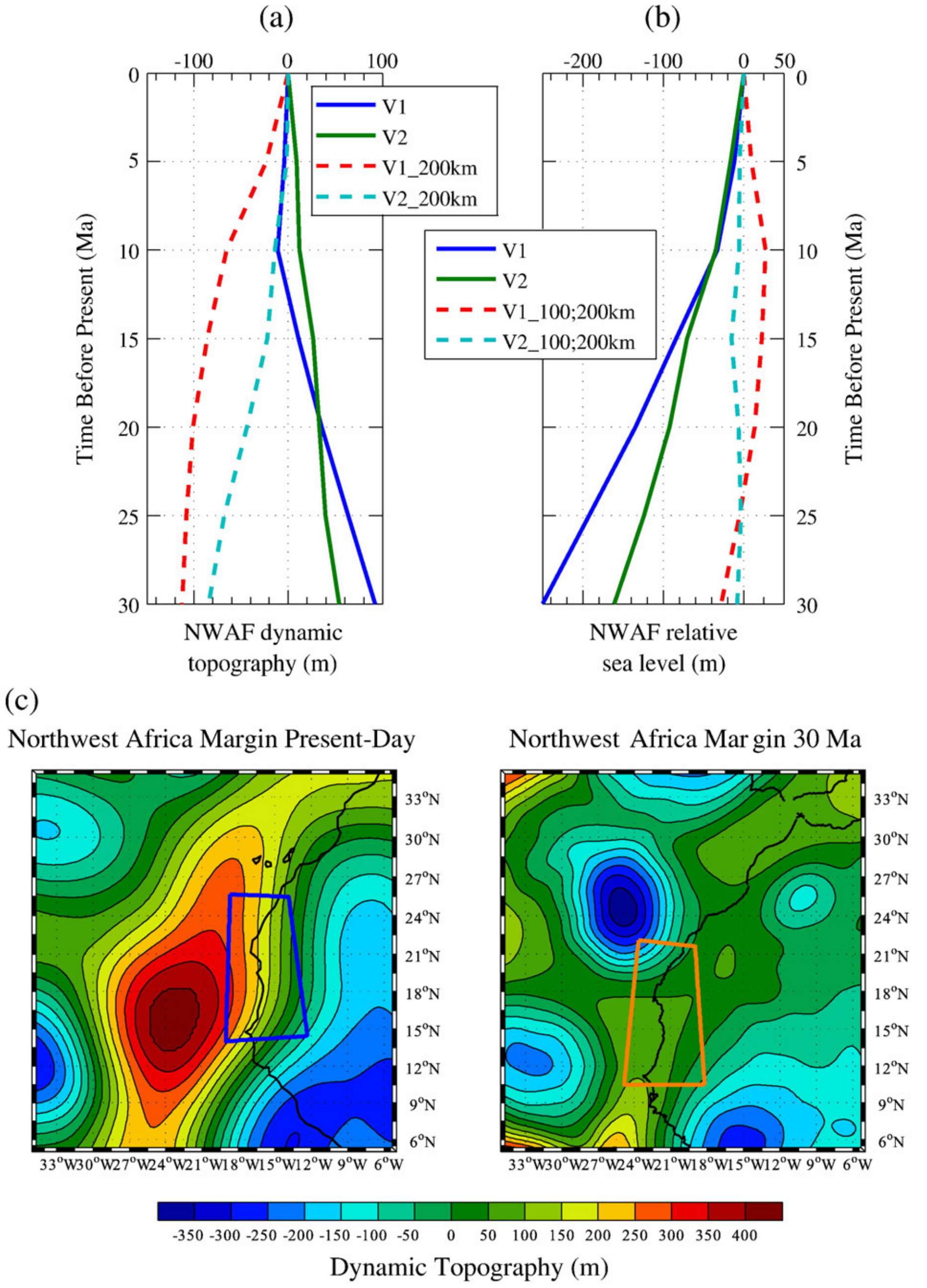

Fig. 4. Dynamic topography (solid surface) variations, relative to the present-day value, predicted for the site on the coasts of (a) northwest Africa (NWAF) and (location shown in Fig. 2) (air–mantle density contrast 3.2 g/cm^3). b) Modeled relative sea level at the NWAF margin. Labels in the inset are defined in Fig. 3. (c) Present-day dynamic topography for model V2_200 km at the NWAF margin compared to dynamic topography at 30 Ma. Also shown is the NWAF area of integration used for results in (b). The coast line and area of integration were backward rotated using an Indo-Atlantic no-net-rotation plate reconstruction model (Quéré et al., 2007).

simulations. RSL refers to predictions of sea-level relative to the present day value, and these predictions include not only eustasy (Fig. 3a) and dynamic topography (Fig. 3b), but also local changes in geoid height (i.e., RSL equals eustasy plus local geoid height changes minus dynamic topography).

The net change in the predicted RSL over the last 30 Myr depends on the adopted viscosity model, although a majority of the results show a sea-level fall in the last 15 Myr of order 50 m. The viscosity profile V2 appears to provide the closest match to observational constraints from the New Jersey coastal plain since the early Oligocene. We note that excluding density heterogeneity within the oceanic and continental lithosphere for a simulation with this viscosity model V2 (denoted by "V2_100;200 km" in Fig. 3c) yields a marked increase in the magnitude of the predicted RSL fall to ~100 m. This value is comparable to the 150 m fall that characterizes the global sea-level curve of Haq et al. (1987) and also fall within the range of observational constraints from backstripping analysis (e.g. Van Sickel et al., 2004; Miller et al., 2005). The main conclusion from Fig. 3 is that local dynamic topography fluctuations cannot be ignored in interpreting sea-level constraints from this site; indeed, these fluctuations are large enough to yield a RSL trend with a different sign than the eustatic trend.

We have also explored the implications of dynamic topography for the Nova Scotian margin analyzed by Watts and Steckler (1979) and we found that for viscosity profile V2 the predicted convection induced RSL over the past 15 Ma was, within model uncertainties, essentially null. The prediction of constant sea level over this time interval is in accord, within observational uncertainties, with the Nova Scotia margin RSL interpretation by Watts and Steckler (1979).

In Fig. 4, we consider a similar set of predictions for the northwest margin of Africa (NWAF in Fig. 2), an area that was once joined to the New Jersey coastal plain prior to the opening of the North Atlantic. For this site we model a RSL rise of about 150 m over the past 30 Myr using the viscosity profile V2 that best-fits the sea-level history along the New Jersey coast. This rise in sea-level has contributions of ~100 m from eustasy (Fig. 3a, green line) and ~50 m from a dynamically driven subsidence of the crust (Fig. 4a, green line). If we exclude the lithosphere, we predict an uplift of the crust of ~80 m (Fig. 4a, cyan dashed line). We note that Bond (1979) estimated a 90 m uplift of Africa for this same time period in order to bring it into line with the mean global sea level variations. The model, denoted by "V2_100;200 km", in Fig. 4 (b) yields a negligible change in RSL at NWAF margin. This is in stark contrast to the RSL results for conjugate NJ margin, but further demonstrates how local epeirogeny can mask the "true" eustatic sea level.

3. Conclusions

We have demonstrated that so-called passive continental margins can be subject to significant, mantle-flow induced vertical motions, or dynamic topography, on the order of 100 m over geologically short time intervals (30 Myr). This variation is comparable in magnitude to the eustatic (i.e., globally uniform) sea-surface height change computed over the same time interval; we thus conclude that long term sea-level variability at these locations inferred from either seismic sequence stratigraphy or backstripping analysis of borehole data cannot be interpreted as solely reflecting global eustatic sea level change. As an illustration of this mantle dynamic contamination, we predict a sea-level fall of 50–100 m for a site in the New Jersey coastal plain for viscosity model V2, with contributions of ~100 m (rise) from eustasy and ~150–200 m (fall) from dynamic topography. The sensitivity of the time-dependent dynamic topography predictions to viscosity opens the possibility of using these RSL data as an additional geodynamic constraint in joint inversions for mantle viscosity (e.g. Mitrovica and Forte, 2004).

We have deconstructed our ~100 m prediction of eustatic sea level rise into contributions from regions that are dynamically elevated (e.g., areas in the vicinity of mid-ocean ridges) and depressed (e.g., area close to subduction zones). We find these two contributions to be of comparable magnitude and we conclude that eustatic sea-level variations are not, as is commonly assumed, related to changes in mid-ocean ridge spreading rates (i.e., ocean ridge volume) alone. Thus, even if one were able to decontaminate local sea-level records for the non-negligible signal from dynamic topography, the residual, eustatic signal would not serve as a robust measure of spreading rate changes.

Models of long-term geochemical fluxes often assume the following chain of proxies (e.g., Hardie, 1996; Stanley and Hardie, 1998; Lowenstein et al., 2001; Horita et al., 2002): long-term trends in sea level at sites on passive continental margins serve as a measure of eustasy, which in turn serve as a measure of the history of mean plate spreading rates, which in turn serve as a measure of geochemical ridge fluxes. Our results indicate two flaws in this chain of logic, related to both the impact of dynamic topography at stable margins and the origin of eustatic sea-level trends. We conclude that the conclusions based on such analysis require reappraisal.

Acknowledgements

We thank K.G. Miller, C.G. Harrison, and an anonymous reviewer, for their comments which helped to clarify the arguments and results presented in this paper. Support for RM was provided by the Earth System Evolution Program of the Canadian Institute for Advanced Research in the form of a postdoctoral fellowship. AMF acknowledges funding provided by the Canada Research Chair program and by the Natural Sciences and Engineering Research Council of Canada. SPG acknowledges NSF grant EAR0309189. This is GEOTOP-UQAM-McGill contribution number 2008-0017.

References

Bond, G.C., 1979. Evidence for some uplifts of large magnitude in continental platforms. Tectonophysics 61, 285–305.

Carter, R.M., 1998. Two models: global sea-level change and sequence stratigraphic architecture. Sediment. Geol. 22, 23–36.

Conrad, C.P., Gurnis, M., 2003. Seismic tomography, surface uplift, and the breakup of Gondwanaland: Integrating mantle convection backwards in time. Geochem. Geophys. Geosys. 4. doi:10.1029/2001GC000299.

Conrad, C.P., Lithgow-Bertelloni, C., 2007. Faster seafloor spreading and lithosphere production during the mid-Cenozoic. Geology 35, 29–32.

Daradich, A., Mitrovica, J.X., Pysklywec, R.N., Willett, S., Forte, A.M., 2003. Mantle convection, dynamic topography and rift-flank uplift of Arabia. Geology 31, 901–904.

Dahlen, F.A., 1976. The passive influence of the oceans upon the rotation of the Earth. Geophys. J. R. astr. Soc. 46, 363–406.

Demicco, R.V., 2004. Modeling seafloor-spreading rates through time. Geology 32, 485–488.

Espesset, A., Forte, A.M., 2001. The importance of a high viscosity peak at 2000 km depth for time-dependent convection dynamics: An explanation for the dominance of degree-2 Structure in the lower-most mantle. Fall Meet. Suppl., Abstract T51D-01. Eos Trans. AGU, vol. 82(47).

Forte, A.M., Peltier, W.R., 1994. The kinematics and dynamics of poloidal–toroidal coupling in mantle flow: The importance of surface plates and lateral viscosity variations. Adv. Geophys. 36, 1–119.

Forte, A.M., Mitrovica, J.X., 1997. A resonance in the Earth's obliquity and precession over the past 20 Myr driven by mantle convection. Nature 390, 676–680.

Forte, A.M., Perry, H.K.C., 2000. Geodynamic evidence for a chemically depleted continental tectosphere. Science 290, 1940–1944.

Forte, A.M., Peltier, W.R., Dziewonski, A.M., Woodward, R.L., 1993. Dynamic surface topography — a new interpretation based upon mantle flow models derived from seismic tomography. Geophys. Res. Lett. 29, 225–228.

Forte, A.M., Mitrovica, J.X., Moucha, R., Simmons, N.A., Grand, S.P., 2007. Descent of the ancient Farallon slab drives localized mantle flow below the New Madrid seismic zone. Geophys. Res. Lett. 34, L04308. doi:10.1029/2006GL027895.

Gurnis, M., 1990. Ridge spreading, subduction and sea level fluctuations. Science 250, 970–972.

Gurnis, M., Müller, R.D., Moresi, L., 1998. Dynamics of Cretaceous to the present vertical motion of Australia and the origin of the Australian–Antarctic discordance. Science 279, 1499–1504.

Hager, B.H., 1980. Eustatic sea level and spreading rate are not simply related. EOS Trans. AGU 61, 374.

Hager, B.H., Clayton, R.W., Richards, M.A., Comer, R.P., Dziewonski, A.M., 1985. Lower mantle heterogeneity, dynamic topography and the geoid. Nature 313, 541–545.

Hallam, A., 1978. Eustatic cycles in the Jurassic. Palaeogeogr. Palaeoclimatol. Palaeoecol 23, 1–32.

Haq, B.U., Hardenbol, J., Vail, P.R., 1987. Chronology of fluctuating sea levels since the Triassic. Science 235, 1156–1167.

Hardenbol, J., Vail, P.R., Ferrer, J., 1981. Interpreting Paleoenvironments, Subsidence History and Sea Level Changes of Passive Margins from Seismic and Biostratigraphy. Proceedings of International Geologic Congress Meetings, Paris, 1980. Oceanol. Acta, pp. 33–44.

Hardie, L.A., 1996. Secular variation in seawater chemistry: An explanation for the coupled secular variation in the mineralogies of marine limestones and potash evaporites over the past 600 m.y. Geology 24, 279–283.

Harrison, C.G.A., 1990. Long term eustasy and epeirogeny in continents. In: Revelle, Roger (Ed.), Sea-Level Change. National Academy of Sciences, Washington, DC, pp. 141–158.

Horita, J., Zimmermann, H., Holland, H.D., 2002. Chemical evolutions of seawater during the Phanerozoic: Implications from the record of marine evaporites. Geochim. Cosmochim. Acta 66, 3733–3756.

Hays, J.D., Pitman, W.C., 1973. Lithospheric plate motion, sea level changes and climatic and ecological consequences. Nature 246, 18–22.

Husson, L., Conrad, C.P., 2006. Tectonic velocities, dynamic topography, and relative sea level. Geophys. Res. Lett. 33, L18303. doi:10.1029/2006GL026834.

Lithgow-Bertelloni, C., Silver, P.G., 1998. Dynamic topography, plate driving forces and the African Superswell. Nature 395, 269–272.

Lowenstein, T.K., Timofeeff, M.N., Brennan, S.T., Hardie, L.A., Demicco, R.V., 2001. Oscillations in Phanerozoic seawater chemistry: evidence from fluid inclusions. Science 294, 1086–1088.

Miall, A.D., Miall, C.E., 2001. Sequence stratigraphy as a scientific enterprise: the evolution and persistence of conflicting paradigms. Earth Sci. Rev. 54, 321–348.

Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N., Pekar, S.F., 2005. The Phanerozoic record of global sea-level change. Science 310, 1293–1298.

Milne, G.A., Mitrovica, J.X., Schrag, D.P., 2002. Estimating past continental ice volume from sea level data. Quat. Sci. Rev. 21, 361–376.

Mitrovica, J.X., Forte, A.M., 2004. A new inference of mantle viscosity based upon a joint inversion of convection and glacial isostatic adjustment data. Earth Planet. Sci. Lett. 225, 177–189.

Mitrovica, J.X., Beaumont, C., Jarvis, G.T., 1989. Tilting of Continental Interiors by the Dynamical Effects of Subduction. Tectonics 5, 1078–1094.

Moucha, R., Quéré, S., Forte, A.M., Mitrovica, J.X., Rowley, D., 2006. Implications of mantle convection for present-day rates of global sea level change. Fall Meet. Suppl., Abstract T53D-1643. AGU, vol. 87(52).

Parsons, B., 1982. Causes and consequences of the relation between area and age of the ocean floor. J. Geophys. Res. 87, 289–302.

Pitman, W.C., 1978. The relationship between eustasy and stratigraphic sequences of passive margins. Geological Society of America Bulletin 89, 1389–1403.

Quéré, S., Rowley, D.B., Forte, A.M., Moucha, R., 2007. No-net-rotation and Indo-Atlantic hotspot reference frames: towards a new view of tectonic plate motions and Earth dynamics. Fall Meet. Suppl., Abstract U34A-03. Eos Trans. AGU, vol. 88(52).

Rowley, D.B., 2002. Rate of plate creation and destruction: 180 Ma to Present. Geol. Soc. of Am. Bull. 114, 927–933.

Simmons, N.A., Forte, A.M., Grand, S.P., 2007. Thermochemical structure and dynamics of the African super-plume. Geophys. Res. Lett. 34, L02301.

Sloss, L.L., 1963. Sequences in the cratonic interior of North America. Geol. Soc. Amer. Bull. 74 (1963), 93–113.

Stanley, S.M., Hardie, L.A., 1998. Secular oscillations in the carbonate mineralogy of reef-building and sediment-producing organisms driven by tectonically forced shifts in seawater chemistry. Palaeogeogr. Palaeoclimatol. Palaeoecol. 144, 3–19.

Steckler, M.S., Watts, A.B., 1978. Subsidence of the Atlantic-type continental margin off New York. Earth Planet. Sci. Lett. 41, 1–13.

Steinberger, B., O'Connell, R.J., 1997. Changes of the Earth's rotation axis owing to advection of mantle density heterogeneities. Nature 387, 169–173. doi:10.1038/387169a0.

Turcotte, D.L., Oxburgh, E.R., 1967. Finite amplitude convection cells and continental drift. J. Fluid Mech. 28, 29–42.

Vail, P.R., Mitchum, R.M., Thompson, S., 1977a. Seismic stratigraphy and global changes of sea level, part 3: relative changes of sea level from coastal onlap. In: Payton, C. (Ed.), Seismic Stratigraphy — Applications to Hydrocarbon Exploration. Am. Assoc. Petrol. Geol. Memoir, vol. 26, pp. 63–81.

Vail, P.R., Mitchum, R.M., Thompson, S., 1977b. Seismic stratigraphy and global changes of sea level, part 4: global cycles of relative changes of sea level. In: Payton, C. (Ed.), Seismic Stratigraphy — Applications to Hydrocarbon Exploration. Am. Assoc. Petrol. Geol. Memoir, vol. 26, pp. 83–97.

Van Sickel, W.A., Kominz, M.A., Miller, K.G., Browning, J.V., 2004. Late Cretaceous and Cenozoic sea-level estimates: backstripping analysis of borehole data, onshore New Jersey. Basin Res. 16 (4), 451–465. doi:10.1111/j.1365-2117.2004.00242.x.

Watts, A.B., Steckler, M., 1979. Subsidence and eustasy at the continental margin of eastern North America. In: Talwani, M., Hay, W.W., Ryan, W.B.F. (Eds.), Deep drilling results in the Atlantic Ocean: Continental Margins and Paleoenvironment. Ewing Series, vol. 3, pp. 218–234.