



RESEARCH LETTER

10.1002/2017GL072845

Key Points:

- Understanding sea level acceleration requires consideration of multiple spatially varying physical processes
- Recent acceleration is well modeled using Greenland and Antarctic ice loss and ocean dynamics
- Acceleration from ice loss alone translates to regional rise of 0.2–0.75 m in one century, depending on location

Supporting Information:

- Supporting Information S1

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Citation:

Davis, J. L., and N. T. Vinogradova (2017), Causes of accelerating sea level on the East Coast of North America, *Geophys. Res. Lett.*, *44*, 5133–5141, doi:10.1002/2017GL072845.

Received 27 JAN 2017

Accepted 10 MAY 2017

Accepted article online 15 MAY 2017

Published online 29 MAY 2017

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Causes of accelerating sea level on the East Coast of North America

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Abstract The tide-gauge record from the North American East Coast reveals significant accelerations in sea level starting in the late twentieth century. The estimated post-1990 accelerations range from near zero to $\sim 0.3 \text{ mm yr}^{-2}$. We find that the observed sea level acceleration is well modeled using several processes: mass change in Greenland and Antarctica as measured by the Gravity Recovery and Climate Experiment satellites; ocean dynamic and steric variability provided by the GECCO2 ocean synthesis; and the inverted barometer effect. However, to achieve this fit requires estimation of an admittance for the dynamical and steric contribution, possibly due to the coarse resolution of this analysis or to simplifications associated with parameterization of bottom friction in the shallow coastal areas. The acceleration from ice loss alone is equivalent to a regional sea level rise in one century of 0.2 m in the north and 0.75 m in the south of this region.

Plain Language Summary We describe a new analysis of sea level accelerations derived from tide-gauge data along the East Coast of North America. Previous analyses of acceleration in this region have focused on ocean dynamics as the cause of recent rapid sea level changes. We have included a number of sources of sea level acceleration, including not only ocean dynamics but also ice-mass loss from Greenland and Antarctica and atmospheric pressure. By focusing on accelerations we are able to remove one of the greatest sources of uncertainty in the sea level budget, postglacial rebound. We have devised an approach wherein data from previous decades (during which the long-term variation is assumed linear) are included, thereby improving the accuracy of estimated post-1990 accelerations by a factor of 3. We conclude that the spatial variability of sea level acceleration is well modeled using these multiple processes. The results indicate that multiple physical processes must be considered to understand changing sea level. We also conclude that the acceleration from Antarctic and Greenland ice loss alone is equivalent to a sea level rise in one century of 0.2 m in the north and 0.75 m in the south of this region.

1. Introduction

Many studies now indicate that global mean sea level (GMSL) is rising. Several reconstructions of GMSL in the twentieth century show a rapid increase starting in approximately 1990, although various approaches to reconstruction disagree both in the GMSL rate throughout most of the twentieth century and the change in GMSL rate post-1990 [e.g., Church and White, 2006; Jevrejeva et al., 2008, 2014; Hay et al., 2015].

Despite the focus on GMSL, the contributions to sea level change are spatially variable. Changes in the height of the sea surface from steric perturbations associated with temperature and salinity, as well as from ocean circulation changes from a variety of forcings, vary significantly over the ocean [e.g., Forget and Ponte, 2015; Yin and Goddard, 2013; Stammer et al., 2013; Calafat et al., 2012; Cazenave and Llovel, 2010; Häkkinen and Rhines, 2004; Cummins and Lagerloef, 2004]. At high latitudes and over shallow-shelf seas, regional changes in sea level due to ocean mass redistribution can also be important [Vinogradova et al., 2007]. Glacial isostatic adjustment (GIA) includes sea level change from gravitational self-attraction and viscoelastic loading associated with the ongoing redistribution of mass in the mantle and ocean from previous cycles of glaciation and deglaciation in the last 2 million years [e.g., Tamisiea and Mitrovica, 2011]. Exchange of mass between the ocean and glaciers/ice sheets not only changes the volume of water in the ocean but also results in self-attraction and elastic loading [e.g., Mitrovica et al., 2001].

The East Coast of North America is a particularly interesting and significant place to study the spatial variability of sea level change, for several reasons. A relatively high density of high-quality long tide-gauge records is available for this region throughout much of the 20th and early 21st centuries. The East Coast is highly developed, and those areas vulnerable to storm flooding, inundation, and erosion could experience a significant economic impact [e.g., Gornitz *et al.*, 2002; Bin *et al.*, 2011; Wdowski *et al.*, 2016]. Ongoing sea level rise on the East Coast is exacerbated by the large region of land subsidence associated with GIA [Davis and Mitrovica, 1996; Wake *et al.*, 2006; Love *et al.*, 2016]. Along some parts of the coast—roughly from Chesapeake Bay to New York, as well as areas north of Maine—GIA dominates the twentieth century local sea level rate, reaching a magnitude of $\sim 1.5\text{--}2\text{ mm yr}^{-1}$ [e.g., Davis and Mitrovica, 1996; Love *et al.*, 2016]. Recent observations in sea level have also pointed to a potential “hot spot” in sea level change along the northern East Coast, which has been attributed to ocean dynamic effects [Sallenger *et al.*, 2012; Kopp, 2013]. Several mechanisms have been mentioned as potential drivers of recent increase in dynamic sea level along the East Coast, including weakening of the overturning circulation [Sallenger *et al.*, 2012; Bingham and Hughes, 2009]; changes in the wind stress [Woodworth *et al.*, 2014; Andres *et al.*, 2013]; divergence of Sverdrup transport [Thompson and Mitchum, 2014]; baroclinic Rossby waves [Miller and Douglas, 2007]; changes in the Florida Current [Park and Sweet, 2015] or the Gulf Stream [Ezer and Atkinson, 2014]; and other coastal processes [Goddard *et al.*, 2015; Sweet and Park, 2014; Ezer, 2013; Kopp, 2013].

Mass loss from the nearby Greenland ice sheet (GIS) produces a geographically variable sea level signal, due to the combined effects of self-attraction and gravitational loading that result in sea level “fingerprints,” which have been used to estimate present-day mass loss from GIS and other ice complexes [e.g., Mitrovica *et al.*, 2001], more recently in combination with simultaneous constraints on GIA Earth models [Wake *et al.*, 2006; Hay *et al.*, 2015]. Because models for GIA rely on (simplified) Earth models and ice history, this process remains a significant source of uncertainty in reconstructions of long-term sea level rates on the North American East Coast [e.g., Love *et al.*, 2016]. Since 2002, estimates of mass change from GIS, the Antarctic ice sheet (AIS), and other areas are available from the Gravity Recovery and Climate Experiment (GRACE) mission. GRACE measurements of GIS and AIS show significant acceleration of mass loss [Velicogna *et al.*, 2014], which should be reflected in East Coast sea level.

While an increasing number of studies highlighted multidecadal acceleration in sea level along the northeast coast of North America [Park and Sweet, 2015; Sweet and Park, 2014; Kenigson and Han, 2014; Ezer, 2013; Calafat *et al.*, 2012; Kopp, 2013; Boon, 2012; Sallenger *et al.*, 2012], those works interpret the changes in sea level in terms of ocean dynamics only. Here we present the first model for the observed geographic variability of eastern North American sea level acceleration that combines ocean dynamics (a term we will take to include steric changes) with accelerated mass loss from Greenland and Antarctica, and the inverted barometer (IB) effect. Studying accelerations has the advantage of being (relatively) insensitive to GIA, which has exponential relaxation timescales in the thousands of years [e.g., McConnell, 1968]. Understanding the sources of sea level acceleration is an important step in improving the accuracy of predictions of future sea level rise.

2. Observations and Estimated Sea Level Acceleration

We use a subset of annual East Coast tide-gauge records obtained from the Permanent Service for Mean Sea Level (PSMSL) [Holgate *et al.*, 2013] that has few data gaps and avoids river locations. The names and locations of the tide gauges are shown in Figure 1. To maximize overlap of the tide-gauge records, we use a common time frame of 1955–2014 for the analysis. For each tide gauge, we express the observed relative sea level (RSL) using a model that has a constant linear rate over the entire period, plus an acceleration that is zero prior to 1990. Including a long time span of data prior to the onset of acceleration enables more accurate (factor of ~ 3) determination of the acceleration compared to using the very short time span during which acceleration has been occurring, by decreasing the formal correlation among the estimated parameters. Our basic model for nontidal RSL change $\Delta\zeta$ at latitude ϕ , longitude λ , and epoch t for all time series is

$$\Delta\zeta(\phi, \lambda, t) = \Delta\zeta_{\text{ice}}(\phi, \lambda, t) + \Delta\zeta_{\text{GIA}}(\phi, \lambda, t) + \Delta\zeta_{\text{dyn}}(\phi, \lambda, t) + \Delta\zeta_{\text{IB}}(\phi, \lambda, t) \quad (1)$$

where the subscript “ice” refers to RSL change associated with ice-mass changes, “GIA” to glacial isostatic adjustment, “dyn” to dynamic sea level changes, and “IB” to the inverted barometer effect.

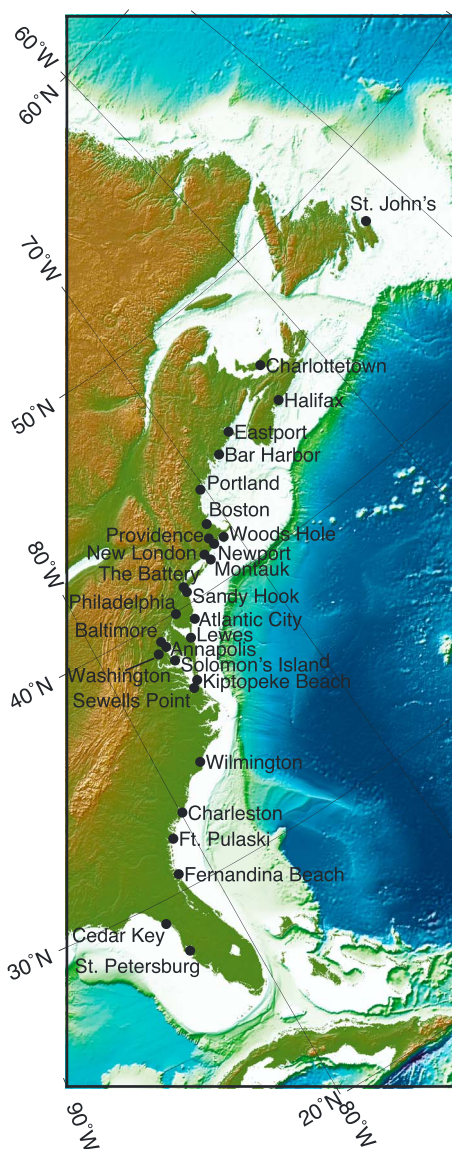


Figure 1. Names and locations of tide gauges used in this study.

latitudes along the East Coast in Figures 2a–2c. The estimated sea level accelerations as a function of site latitude are shown in Figure 3a and are provided in Table S1 in the supporting information.

We do not account either for temporal or spatial correlations in the data or models. *Doran et al.* [2015] found that this could lead to an overestimation of the acceleration uncertainty. We found an average lag-1 correlation for our annual RSL residuals of ~ 0.3 , which could lead to an underestimation of the uncertainties of $\sim 25\%$. However, our fit to equation (3), described in section 4, indicates that, if anything, our uncertainties are overestimated. This result is possibly a result of spatial correlation of unmodeled interannual oceanographic variability, which leads to smaller-than-predicted uncertainty in the spatial variability of estimated parameters [e.g., *Davis et al.*, 1999].

3. Contributions to Sea Level Acceleration

We consider four contributions to sea level acceleration: mass change from (1) AIS and (2) Greenland, (3) ocean dynamics including steric changes, and (4) the inverted barometer (IB) response of the ocean due to surface pressure variations. Below, we describe the calculation of each of these contributions.

The temporal model for all sea level time series is taken to be a linear variation with an acceleration that begins in 1990 [e.g., *Doran et al.*, 2015]:

$$\begin{aligned}\zeta(\phi, \lambda, t) &= \zeta_o(\phi, \lambda) + \Delta\zeta(\phi, \lambda, t) \\ &= \zeta_o(\phi, \lambda) + v(\phi, \lambda)(t - t_o) \\ &\quad + \frac{1}{2}a(\phi, \lambda)(t - t_{\text{start}})^2 \mathcal{H}(t - t_{\text{start}})\end{aligned}\quad (2)$$

where $\zeta_o = \zeta(t_o)$ is a reference value, with $t_o = 1985$, v is the sea level rate, a is the sea level acceleration, $t_{\text{start}} = 1990$, and $\mathcal{H}(t)$ is the Heaviside step function. In performing the least squares fits to equation (2) for all time series, we assume unit weighting on each point. The standard deviation of the estimated parameters is found by scaling the standard error of unit weight by the postfit RMS residual.

Because of the linearity of equations (1) and (2), the acceleration of sea level is therefore modeled prior to inclusion of the ocean-model admittance (see below), as

$$\begin{aligned}a(\phi, \lambda) &= a_{\text{ice}}(\phi, \lambda) + a_{\text{GIA}}(\phi, \lambda) \\ &\quad + a_{\text{dyn}}(\phi, \lambda) + a_{\text{IB}}(\phi, \lambda)\end{aligned}\quad (3)$$

with $a_{\text{GIA}} \approx 0$.

An acceleration onset epoch of 1990 is reasonable, but it is difficult to assert positively that there is a single epoch at which acceleration began. Data from GRACE are available only since 2002, but the accelerations in mass loss found by *Velicogna et al.* [2014], particularly for southwest Greenland, are so large that they could not have been ongoing for very long before ~ 2000 without requiring a long period of mass gain prior to the 1990–2000 time frame. [see *Velicogna et al.*, 2014, Figure 2e]. *Kopp* [2013] found a brief period of negative sea level rate anomalies for the period 1975–1980, followed by a more sustained period of positive sea level rate anomalies starting ~ 1990 . *Boon* [2012] reported an onset epoch of 1987. We show some sample time series along with model fit for a range of

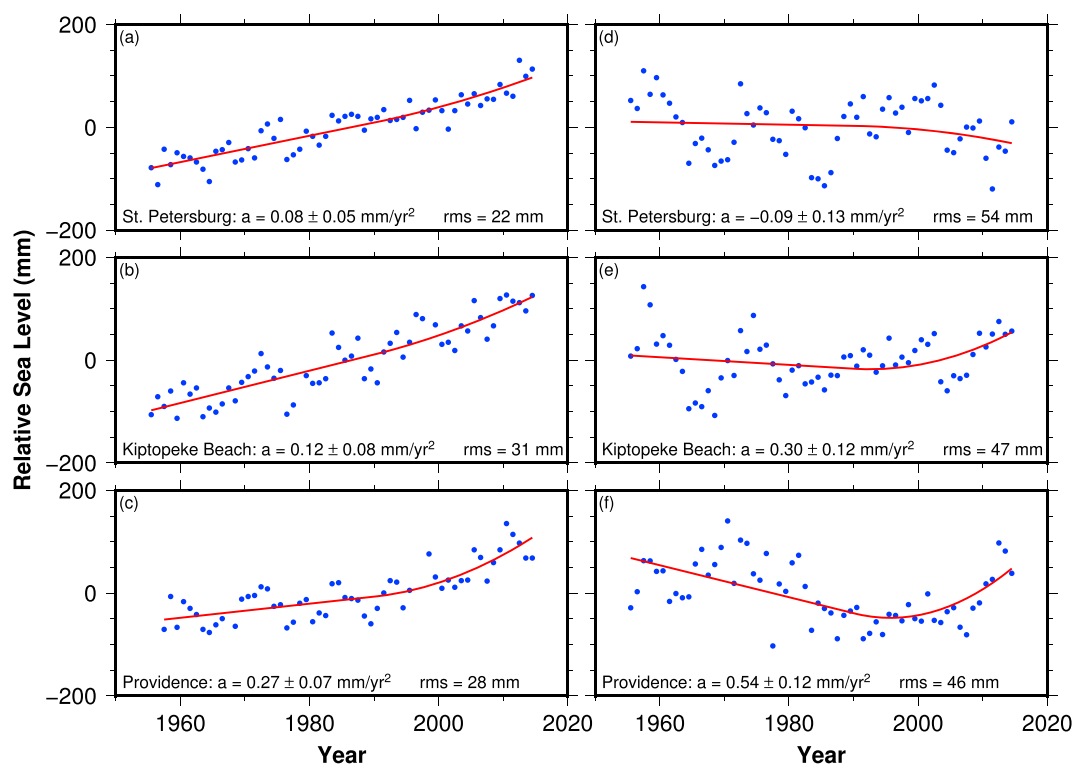


Figure 2. (a–c) Examples of tide-gauge time series showing model fit, with “RMS” indicating the root-mean-square residual. (d–f) Same for the GECCO2 East Coast time series.

3.1. AIS and GIS Mass Loss

The contribution to sea level from AIS and GIS mass loss is assumed to be solely due to surface mass redistribution and the gravitational self-attraction and loading effects on the solid Earth and sea surface [e.g., *Mitrovica et al.*, 2001]. No coupling to ocean dynamics is accounted for (see section 3.2). The model for accelerated mass change for AIS and GIS was based on the determination of regional mass loss from these ice sheets from GRACE observations by *Velicogna et al.* [2014]. We assume the accelerations due to other major ice sheets and glaciers are zero. Using the GRACE-based estimated ice-mass acceleration and its geographical distribution provided by *Velicogna et al.* [2014], we then solved the sea level equation [e.g., *Tamisiea et al.*, 2010; *Adhikari et al.*, 2016, and references therein] that accounts for the combined effects of self-attraction and elastic loading, with Love numbers based on the Preliminary Reference Earth Model [*Dziewonski and Anderson*, 1981], to yield the spatial variability of sea level acceleration in the regions of the east coast. The combined contribution to sea level acceleration in the offshore Atlantic due to AIS and GIS is shown in Figure S1. The contributions at the tide-gauge locations are shown in Figure 3b.

The AIS contribution is smaller than the mean GIS contribution, reflecting the smaller total ice-mass accelerations from *Velicogna et al.* [2014], -10.6 Gt yr^{-2} for Antarctica compared to -25.4 Gt yr^{-2} for Greenland (with the negative sign indicating mass loss). Because the North American East Coast is so far from AIS, there is nearly no geographic variability associated with the predicted sea level acceleration. In contrast, the sea level acceleration for GIS varies significantly, with GIS mass loss causing a small negative sea level acceleration of $\sim 0.03 \text{ mm yr}^{-2}$ for the northern sites, changing to a positive sea level acceleration of $\sim 0.1 \text{ mm yr}^{-2}$ for sites at the latitude of Florida. The negative values of acceleration are caused by the effects of self-attraction and loading associated with mass loss of GIS [*Mitrovica et al.*, 2001] overwhelming the added ocean volume from the conversion of ice to water from AIS and GIS.

3.2. Ocean Dynamics

For the dynamic contribution to sea level, we used the publicly available GECCO2 dynamically consistent ocean state estimation [*Köhl*, 2015]. One of the German (hence “G”) contributions to the Estimating the Circulation and Climate of the Ocean (ECCO) system of models [e.g., *Wunsch and Heimbach*, 2006],

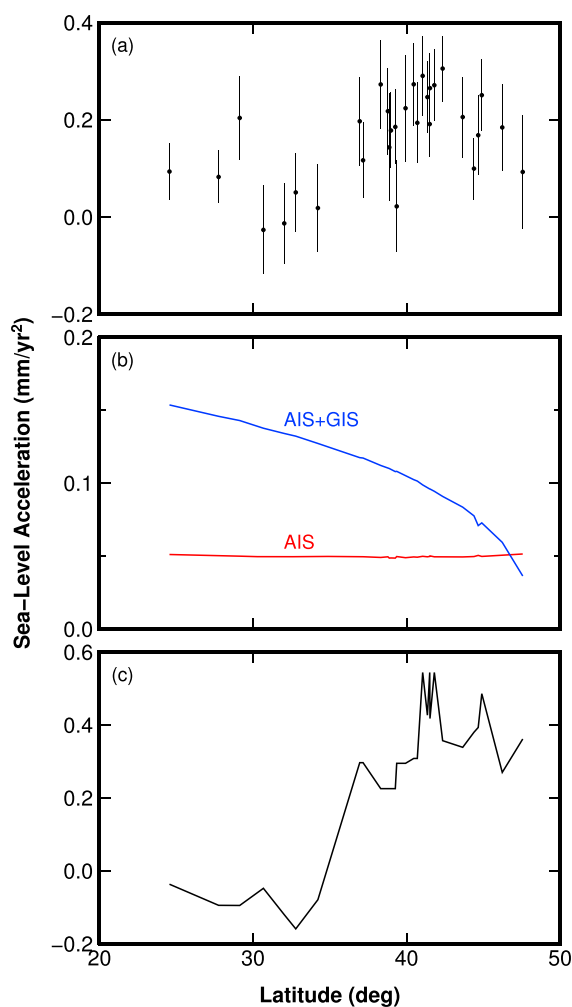


Figure 3. (a) Estimated East Coast sea level acceleration (1990–2014) at TG locations versus latitude. (b) Sea level acceleration for the AIS (red) and AIS plus GIS (blue) based on models for AIS and GIS mass loss and solution of the sea level equation. (c) Sea level acceleration from the GECCO2 ocean estimate (see text). Note the use of different vertical scales.

GECCO2 provides an optimal and dynamically consistent synthesis of most available oceanic data (in situ and satellite, including altimetry and sea surface temperature) along with surface fluxes derived from atmospheric reanalyses within their prescribed uncertainties. An advantage in using GECCO2 for our purposes is the long time span of estimates it provides. This choice necessitates using estimates prior to the period for which we have altimetry, but it increases the accuracy of acceleration derived from these estimates as discussed above.

Within the context of this study, the contribution to sea level from ocean dynamics includes variations in surface height arising from vertical expansion or contraction of the water column due to changes in temperature and freshwater contents [e.g., Johnson and Wijffels, 2011], as well as from changes related to redistribution of mass within the ocean and mass input from precipitation, evaporation, and river runoff. Input from land ice to the ocean dynamic contribution is not considered. Global mean mass variations are not included in GECCO2, and all changes in mass occur only because of mass redistribution within the oceans. For this study, we used monthly estimates of dynamic sea level change calculated prognostically within GECCO2, which integrates the continuity equation in the Boussinesq approximation [e.g., Wunsch et al., 2007; Forget and Ponte, 2015]. This approach is reasonably accu-

rate since ocean density varies within 1% of a mean value [Gill and Niiler, 1973]. In such formulation, the GECCO2 continuity equation is volume conserving, implying that, although relative horizontal gradients are correct, the global mean sea level is constant and zero.

To account for non-Boussinesq volume changes, a spatially uniform but time-varying correction can be applied [e.g., Greatbatch, 1994; Ponte, 1999; Vinogradova et al., 2007] but is generally not included due to known imbalances in global mean freshwater fluxes in the GECCO2 solution [Köhl and Stammer, 2008], which affect the accuracy of the net steric changes. We calculate that the impact of non-Boussinesq processes on the global mean sea level acceleration over the 60 year period is $\sim 0.03 \text{ mm yr}^{-1}$.

We used monthly solutions during the period 1955–2014 on a 1° horizontal grid. The fit to the time series assumed equal weights on the sea level estimates. Although it might be assumed that the later estimates are more accurate due to the assimilation of altimeter data, no uncertainties are provided for GECCO2 products. To calculate the acceleration contribution at TG site, we used nearest GECCO2 grid point (Figure S2). Sample time series and fits are shown in Figures 2d–2f.

The GECCO2 ocean estimates yield negative sea level rates for some East Coast sites, consistent with Wake et al. [2006] and Yin and Goddard [2013]. Comparison of Figures 2a–2e and 2d–2f indicates clearly that ocean

dynamics, as predicted by GECCO2, cannot alone explain many of the features observed in the tide-gauge record, including not only the acceleration but also the rates and interannual variability.

The contributions to the sea level acceleration at the tide-gauge locations are shown in Figure 3c. The pattern of accelerations for the dynamic sea level contribution has a strong geographic variability, with sites south of latitude 35° experiencing negative sea level accelerations, and a rapid increase in sea level acceleration north of this latitude. North of latitude 40°, the sea level accelerations become large, exceeding 0.5 mm yr⁻² in the Montauk-Providence area.

3.3. IB Response

The IB response was calculated using the National Oceanic and Atmospheric Administration (NOAA) Twentieth-Century Reanalysis Version 2c [Compo *et al.*, 2011]. This model was selected on the basis of a comparison by Piecuch and Ponte [2015] who compared several models and concluded that this model explained the most sea level variance for East Coast sites. Monthly time series of surface pressure was used to estimate sea level acceleration assuming a perfect IB response. Despite the IB response being a significant contribution to interannual sea level variation in this region [Piecuch and Ponte, 2015], the overall contribution to sea level acceleration during the period 1990–2014 was small, ranging from zero in the south to 0.05 mm yr⁻² for sites in the north. Because the IB is small, we did not test other atmospheric pressure models, and we do not have a separate plot for this contribution, although it is included in the model.

4. Analysis and Implications for Future Sea Level Rise

Comparing observed sea level accelerations (Figure 3a) and ice-mass contributions (Figure 3b), it is clear that the ice-mass contribution is a significant fraction of the observed acceleration only south of ~40°, north of which ocean dynamics (Figure 3c) dominate the observed signal. However, it is visually evident from Figure 3 that the sea level acceleration predicted by combining the ice-mass and ocean dynamic contributions would exceed the observed acceleration in some parts by nearly a factor of 2. Since this is the region in which the dynamic contributions dominates, it seems reasonable to assume that the accelerations predicted by the GECCO2 model are too large, although the pattern of geographic variability seems reasonable. We therefore model the observed sea level acceleration by allowing for a scaling of the ocean dynamic contribution:

$$a_{TG}(\phi, \lambda) = \beta a_{OM}(\phi, \lambda) + a_{AIS}(\phi, \lambda) + a_{GIS}(\phi, \lambda) + a_{IB}(\phi, \lambda) + \epsilon(\phi, \lambda) \quad (4)$$

where β is an unknown ocean dynamic scaling, or admittance, parameter, and the a are the accelerations observed from the tide-gauge data (subscript TG), the GECCO2 ocean estimate (OM), the inverted barometer signal (IB), and from the GRACE observations for the AIS and GIS. The term $\epsilon(\phi, \lambda)$ is unknown random error. In the least squares fit for the ocean-model admittance parameter, we assume that the $\epsilon(\phi, \lambda)$ are Gaussian, zero-mean, and uncorrelated, and that their standard deviation is provided by the acceleration uncertainty from the fit of the tide-gauge time series to the time series model (equation (2)).

Using a standard least squares inversion, we found $\beta = 0.28 \pm 0.04$ yields the best fit to data. The final model and data as well as the postfit residuals are shown in Figure 4. The weighted RMS postfit residual between observed and model accelerations of 0.07 mm yr⁻², and the reduced χ^2 difference is 0.75, indicating a good fit relative to the assumed standard deviations for the error, which as mentioned above accounts only for errors associated with the tide-gauge estimates based on the fit to equation (2). No account has been taken of random (or random-like) errors in the model terms on the right-hand side of equation 4. The fact that we then obtained a better-than-expected fit implies that the errors in the estimated tide-gauge accelerations are overestimated, our assumption of uncorrelated errors is incorrect, or both. This result could also imply, though, that the combination of errors in the OM, AIS, GIS, and IB terms have a standard deviation significantly less than 0.07 mm yr⁻².

The admittance $\beta < 1$ indicates that the observed sea level accelerations are significantly less than the accelerations calculated from the GECCO2 time series. However, GECCO2 did not provide sea level predictions at the locations of the tide gauges, but for a 1° horizontal grid that mostly avoided coastal locations. Hill *et al.* [2007], in comparing seasonal variability of sea level measured by tide gauges and predicted by an ocean model, found that for the North American East Coast, an admittance of ~0.7 was required (for ECCO), which they hypothesized may be due to limitations in coastal geometry in the ocean model and to insufficient dissipation associated with parameterization of bottom friction. In particular, along shallow coastal regions such as the continental shelf (Figure 1), velocities resulting from the wind stress are largely balanced by bottom

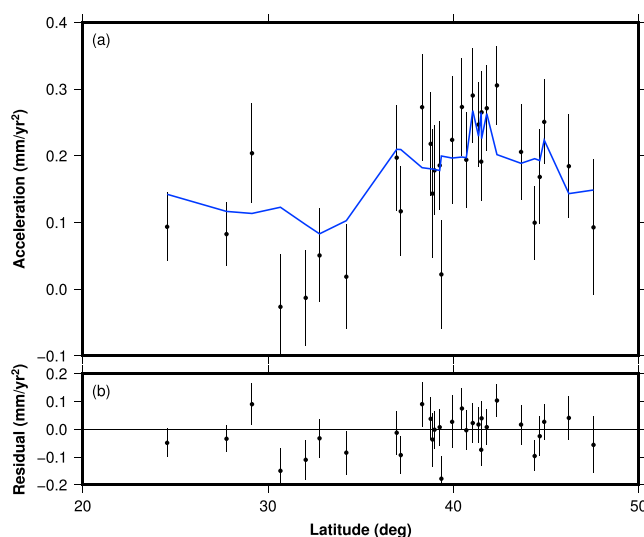


Figure 4. Observed sea level acceleration (points with error bars) and postfit model (blue line). The weighted RMS difference between the data and model is 0.07 mm yr⁻², and the reduced χ^2 difference is 0.75, indicating a good fit. The uncertainties have been scaled to reflect this fit.

friction [e.g., *Strub et al.*, 2013], making sea level estimates sensitive to friction parameterization. The large variability in sea level suggests that the bottom friction in the GECCO2 framework might be too low. In addition, the coarse resolution of GECCO2 does not allow for the resolution of mesoscale eddies. Our result hints at potential difficulties in the approach that uses coarse-resolution ocean models to predict sea level change at the coast and may imply that care should be taken in this regard.

Estimates of sea level acceleration due to mass loss in Greenland and Antarctica should be robust, their being based on the estimated global gravity from the GRACE satellites over the period 2002–2015, but the extrapolation back to 1990 may be a source of error. The sea level equation used

to calculate sea level change based on mass change includes the well-known theories of gravitational attraction and deformation on an elastic planet [Farrell, 1972]. Performing the fit with the ice-mass contributions omitted produces a poor fit; the χ^2 fit nearly doubles (from 0.75 to 1.40) and the residuals have an obvious north-south systematic trend (Figure S3).

Understanding the contributions to sea level acceleration improves our ability to make predictions of future sea level rise in this region [e.g., *Love et al.*, 2016] and may help to constrain climate models used to predict future sea level changes. Studies focusing on statistical tests based on ocean dynamical variability [e.g., *Kopp*, 2013] conclude that the observed accelerations are within reasonable bounds for twentieth century. Our finding a reliable model for the observed spatial variability in acceleration (having only one free parameter) gives us confidence not only that the observed sea level acceleration is “real” but also that we better understand its causes.

In addition, by explicitly accounting for contributions from Antarctica and Greenland accelerated melting, we find that there is a significant component of the observed sea level acceleration that will likely not fluctuate, (that is, increase and decrease) on decadal timescales, but will if anything increase systematically and perhaps nonlinearly with time [e.g., *Hansen et al.*, 2016, and references therein]. We conclude that the North American East Coast has recently experienced sea level acceleration due to combined AIS and GIS ice-mass loss only (that is, without any ocean dynamic component) of 0.04–0.15 mm yr⁻², depending on location. This acceleration is equivalent—in a purely quantitative sense, i.e., with no implication for a specific future projection—to an increase in sea level in one century of 0.2–0.75 m.

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Acknowledgments

This work was supported by National Aeronautics and Space Administration grants NNX14AP33G (Sea-Level Change Team) and NNX12AF34G (GRACE Science Team). Figures were produced using Generic Mapping Tools [Wessel et al., 2013]. The GECCO2 solution is available at icdc.zmaw.de. PSMSL tide-gauge data are available at psmsl.org. Support for the Twentieth Century Reanalysis Project version 2c data set is provided by the U.S. Department of Energy, Office of Science Biological and Environmental Research (BER) and by the National Oceanic and Atmospheric Administration Climate Program Office and is available at www.esrl.noaa.gov. We thank two anonymous reviewers for their valuable comments on the manuscript. This is LDEO contribution number 8117.

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