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Understanding Tide Gauge Mean Sea Level Changes on the East Coast of North America

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UNDERSTANDING TIDE GAUGE MEAN SEA LEVEL CHANGES ON THE
EAST COAST OF NORTH AMERICA

BY
CHRISTOPHER GILBERT PIECUCH

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
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IN
OCEANOGRAPHY

UNIVERSITY OF RHODE ISLAND

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DOCTOR OF PHILOSOPHY DISSERTATION
OF
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2016

ABSTRACT

Sea level (SL) is an informative index of climate and a serious concern for coastal communities. Understanding the observational SL record is important from scientific and societal points of view. We consider the tide gauge SL record, focusing on data along the North American northeast coast, aiming to identify relevant geophysical processes responsible for observed SL changes.

SL changes reflect dynamic and isostatic ocean effects. Recent works have interpreted accelerated and extreme SL changes along the northeast coast of North America primarily in terms of dynamic changes. In manuscript 1, we consider the influence of the ocean’s isostatic response to surface atmospheric pressure loading—the inverted barometer (IB) effect—on annual mean SL from tide gauge records. The IB effect explains $\sim 25\%$ of interannual SL variance and accounts for $\sim 50\%$ of the magnitude of a recent extreme event of SL rise along Atlantic Canada and New England. Estimated IB effects also amount to $\sim 10\text{--}30\%$ of recent multidecadal SL accelerations over the Mid-Atlantic Bight and Southern New England. These findings reiterate the need for careful estimation and removal of isostatic effects for studies of dynamic SL.

In manuscript 2, we continue our investigation of east coast tide gauge SL, seeking to better understand the relation between coastal SL and the variable ocean circulation. Annual SL records (adjusted for the IB effect) from tide gauges along the North American northeast coast over 1980–2010 are compared to a set of data-assimilating “ocean reanalysis” products as well as a global barotropic model solution forced with wind stress and barometric pressure. Correspondence between models and data depends strongly on model and location. At sites north of Cape Hatteras, the barotropic model shows as much (if not more) skill than ocean reanalyses, explaining $\sim 50\%$ of the variance in the adjusted annual tide

gauge SL records. Additional numerical experiments show that annual SL changes along this coast from the barotropic model are driven by local wind stress over the continental shelf and slope. This result is interpreted in the light of a simple dynamic framework, wherein bottom friction balances surface wind stress in the alongshore direction and geostrophy holds in the across-shore direction. Results highlight the importance of barotropic dynamics on coastal SL changes on interannual and decadal time scales; they also have implications for diagnosing errors in ocean reanalysis, using tide gauge records to infer past changes in ocean circulation, and identifying mechanisms responsible for projected regional SL rise.

Finally, in manuscript 3, three global gridded reanalysis products are used alongside a carefully curated set of station records and tide gauges to consider IB changes over the global ocean and their relation to SL changes over the Twentieth Century. Centennial IB trends from reanalysis products show meridional structure consistent with the IB response expected under global warming. Annual IB variations show stronger amplitudes at higher latitudes, as in past studies focusing on higher frequencies or shorter periods. Discrepancies between gridded IB products tend to be smaller (larger) for more recent (earlier) periods and ocean regions with good (poor) historical data coverage. Comparisons between reanalysis products and station data reveal evidence for common errors across reanalyses over a wide range of time scales from annual to centennial. Notwithstanding their errors, the gridded reanalysis products are useful for interpreting tide gauge records: subtracting IB from SL records reduces both temporal variance within tide gauge records and spatial variance across tide gauge sites. Results advocate for making the IB correction to tide gauge SL data using reanalysis products in studies of ocean circulation and climate on centennial time scales.

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It's been a long road, and I've got lots of people to thank—

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Thank you to my wonderful family, especially my darling sister and mother.

And, saving the best for last, to Hannah, my very best friend—

Thank you. So much. This is yours as much as it's mine.

S. D. G.

DEDICATION

For Hannah; another brick in our Shining Barrier.

i carry your heart with me(i carry it in
my heart)

PREFACE

This dissertation is written in manuscript format suitable for publication in a scientific journal. It consists of three chapters in various stages of publication. Manuscript 1 was published in Geophysical Research Letters in 2015. Manuscript 2 was published in Journal of Climate in 2016. Manuscript 3 is in preparation for submission to Journal of Atmospheric and Oceanic Technology.

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MANUSCRIPT 1

**Inverted barometer contributions to recent sea level changes along the
northeast coast of North America**

by

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1.1 Abstract

Regional sea level (SL) changes reflect dynamic and isostatic ocean effects. Recent works have interpreted accelerated and extreme SL changes along the northeast coast of North America primarily in terms of dynamic changes; however, dedicated study of isostatic changes related to surface atmospheric pressure loading—the inverted barometer (IB) effect—has been lacking. This investigation uses five different atmospheric pressure products to analyze the influence of the IB effect on annual mean SL from tide gauge records. The IB effect explains $\sim 25\%$ of interannual SL variance and accounts for $\sim 50\%$ of the magnitude of a recent extreme event of SL rise along Atlantic Canada and New England. Estimated IB effects also amount to $\sim 10\text{--}30\%$ of recent multidecadal SL accelerations over the Mid-Atlantic Bight and Southern New England. These findings reiterate the need for careful estimation and removal of isostatic effects for studies of dynamic SL.

1.2 Introduction

Sea level changes can adversely impact coastal communities, leading to submergence, flooding, and erosion [Cazenave and Le Cozannet, 2014]. Global mean sea level has risen steadily over the last century, with the rate of rise likely having accelerated during the last quarter-century [Church *et al.*, 2013]. Moreover, magnitudes of extreme regional sea levels have likely increased over the last half-century [Rhein *et al.*, 2013]. Given the flood exposure of growing coastal populations and assets [Hallegatte *et al.*, 2013], there is ample motivation to understand physical mechanisms underlying such past and present changes to project future sea levels and facilitate community adaptive planning [e.g., Rubin, 2013].

Regional sea level changes can reflect ocean dynamics, for example, adjustments of the gyre or overturning circulations to wind or buoyancy forcing; but they can also represent processes unrelated to ocean dynamics, such as the isostatic response to surface loading due to barometric pressure (or the inverted barometer effect) [Stammer *et al.*, 2013]. Recent papers identify multidecadal accelerations [Boon, 2012; Ezer and Corlett, 2012; Sallenger *et al.*, 2012; Calafat and Chambers, 2013; Ezer, 2013; Ezer *et al.*, 2013; Kopp, 2013; Ezer and Atkinson, 2014; Haigh *et al.*, 2014; Kenigson and Han, 2014; Sweet and Park, 2014; Park and Sweet, 2015] and extreme interannual anomalies [Sweet and Zervas, 2011; Goddard *et al.*, 2015] in tide gauge sea level along the northeast coast of North America; those works, and follow-on studies [Andres *et al.*, 2013; Yin and Goddard, 2013; Thompson and Mitchum, 2014; Woodworth *et al.*, 2014], interpret these regional sea level changes primarily in terms of ocean dynamics, for instance, changes in the Florida Current, Gulf Stream, meridional overturning circulation, as well as other coastal current regimes.

Given the focus of these papers, the impact of the inverted barometer effect

on such multidecadal accelerations and interannual extremes in tide gauge records remains unclear. Some of these studies acknowledge and remove the inverted barometer prior to analysis, but without providing a sense of its impact, whereas a few others make no mention of this effect. Several of them briefly discuss the inverted barometer effect [e.g., *Boon*, 2012; *Calafat and Chambers*, 2013; *Kopp*, 2013; *Haigh et al.*, 2014; *Woodworth et al.*, 2014; *Goddard et al.*, 2015] and their findings are treated below in the context of our results. Earlier works show that, while not amounting to dominant contributions, inverted barometer impacts on sea level trends and variations nevertheless are not always negligible; for example, *Ponte* [2006] reveals that removing the inverted barometer can alter trends by up to 15% and variance by 20% in annual mean tide gauge sea level from Baltimore to Halifax over the years 1958–2000 [cf. *Kolker and Hameed*, 2007; *Woodworth et al.*, 2009].

Given the aforementioned considerations, a detailed inquiry into the influence of the inverted barometer effect along the northeast coast of North America would seem timely. To this end, we consider coastal locations along the Mid-Atlantic Bight, Gulf of Maine, and Atlantic Canada, determining inverted barometer contributions to recently reported sea level accelerations and extremes using tide gauges and various barometric pressure datasets. We estimate that the inverted barometer effect accounts for $\sim 50\%$ of an extreme event of sea level rise between 2008 and 2010 along New England and Atlantic Canada [*Goddard et al.*, 2015] and $\sim 10\text{--}30\%$ of multidecadal sea level accelerations along the Mid-Atlantic Bight and Southern New England [*Sallenger et al.*, 2012; *Boon*, 2012], reiterating need for careful estimation and removal of isostatic effects for studies of dynamic sea level.

1.3 Datasets

To investigate sea level (η) between Virginia and Newfoundland, we use annual revised local reference time series from 30 tide gauges (Table 1.1; Figure 1.1). Data were extracted from the Permanent Service for Mean Sea Level (PSMSL) database [Holgate *et al.*, 2013; PSMSL, 2015] on 16 February 2015. Motivated by several recent investigations focusing on these same locations [Sallenger *et al.*, 2012; Andres *et al.*, 2013; Thompson and Mitchum, 2014; Woodworth *et al.*, 2014; Goddard *et al.*, 2015], we consider these records from 1950 onwards. Given our interest in multidecadal variations and interannual fluctuations, we use least squares to estimate and remove linear trends from the gauge records, which mainly reflect vertical land motion and global mean η changes [Kopp, 2013]. Consistent with past works [Thompson, 1986; Papadopoulos and Tsimplis, 2006; Boon, 2012; Andres *et al.*, 2013; Thompson and Mitchum, 2014; Woodworth *et al.*, 2014], these tide gauge records evidence interannual and multidecadal η variations that are coherent across the Mid-Atlantic Bight, Gulf of Maine, and Atlantic Canada coast (Figure 1.1). Recent studies attributing such η variations have highlighted the role of nearshore wind stress [Andres *et al.*, 2013; Woodworth *et al.*, 2014], baroclinic Rossby waves [Hong *et al.*, 2000; Miller and Douglas, 2007], overturning circulation changes [Bingham and Hughes, 2009; Sallenger *et al.*, 2012], and divergence of Sverdrup transport [Thompson and Mitchum, 2014].

To estimate the inverted barometer response (η^{ib}), we use monthly sea level pressure (P_a) from five different datasets: Version 2 Forcing for Coordinated Ocean-ice Reference Experiments (COREv2) [Large and Yeager, 2009], the Hadley Centre Sea Level Pressure data (HadSLP2) along with its near-real time updates [Allan and Ansell, 2006], Interim European Centre for Medium-Range Weather Forecasts Reanalysis (ERA-Interim) [Dee *et al.*, 2011], National Centers for Environmental

Prediction (NCEP)-Department of Energy (DOE) Reanalysis II (R2) [*Kanamitsu et al.*, 2002], and National Oceanic and Atmospheric Administration (NOAA) Twentieth-Century Reanalysis (20CR) Version 2 [*Compo et al.*, 2011]. The HadSLP2, ERA-Interim, NCEP-DOE R2, and NOAA 20CR data are defined on regular grids with horizontal spacings of 5° , 0.75° , 2.5° , and 2° , respectively, while COREv2 has $\sim 1.9^\circ$ mean meridional spacing and 1.875° zonal spacing. Time periods covered by COREv2, HadSLP2, ERA-Interim, NCEP-DOE R2, and NOAA 20CR are 1948–2009, 1850–2012, 1979–2014, 1979–2014, and 1871–2012, respectively. We compute spatial grids of η^{ib} based on each of the different P_a products according to [*Ponte*, 2006]

$$\eta^{ib} = -\frac{P_a - \overline{P_a}}{\rho g}, \quad (1.1)$$

where overbar is spatial average over the global ocean surface, g is acceleration due to gravity, and ρ is a constant reference ocean density. Annual η^{ib} values are mapped to tide gauges using nearest neighbor interpolation only considering P_a grid cells over the ocean.

1.4 Multidecadal changes and accelerations

Motivated by *Sallenger et al.* [2012] and others, we study multidecadal changes across 20 tide gauge stations on the Mid-Atlantic Bight and Southern New England (Table 1.1); these locations constitute the “northeast hotspot” of *Sallenger et al.* [2012]. We average the detrended time series of tide gauge η and corresponding estimates of η^{ib} from COREv2, HadSLP2, and NOAA 20CR across these stations; note that we use neither ERA-Interim nor NCEP-DOE R2 at this point because they do not extend far enough back in time. We compute η and η^{ib} accelerations (denoted symbolically as a_η and $a_{\eta^{ib}}$) using least squares. Based on recommendations from *Haigh et al.* [2014], we systematically determine a_η and $a_{\eta^{ib}}$ for all accessible periods with beginning years ≥ 1950 and window lengths ≥ 40

years.

For most time periods, a_η is not significant at the $1\text{-}\sigma$ level (Figure 1.2). (We have adjusted the standard errors based on the autocorrelation of residuals [e.g., *Maul and Martin*, 1993; *Calafat and Chambers*, 2013; *Haigh et al.*, 2014].) However, we find significant a_η values over the most recent periods; for example, for 1950–2009, 1970–2009, and 1969–2011, we compute a_η values of 0.044 ± 0.040 , 0.176 ± 0.131 , and $0.218 \pm 0.110 \text{ mm yr}^{-2}$, respectively, similar to *Sallenger et al.* [2012] (their Figures S2, S4) and *Boon* [2012] (his Figure 5). Averaging over periods with significant values, we find a mean a_η value of 0.109 mm yr^{-2} .

We determine η^{ib} contributions to the above η accelerations by computing the quantity

$$\mathcal{R} = 100\% \times (a_{\eta^{ib}}/a_\eta), \quad (1.2)$$

for all accessible combinations of P_a dataset and time period with significant a_η value. For a particular time period, we compute $\mathcal{R}$ based on a given P_a product only if that product covers the entirety of that period. Contributions from η^{ib} to a_η are sensitive to choice of P_a product and time period (Figure 1.2 inset); for example, for periods ending in 2009 we compute average $\mathcal{R}$ values of 36%, 19%, and 4% using COREv2, HadSLP2, and NOAA 20CR, respectively, whereas we find mean $\mathcal{R}$ values of 26% and 15% for periods ending in 2011 based on HadSLP2 and NOAA 20CR, respectively. More generally, averaging over periods with significant a_η values (Figure 1.2), we find average $\mathcal{R}$ values of 33%, 22%, and 10% using COREv2, HadSLP2, and NOAA 20CR, respectively. Additionally, we estimate standard errors for the $\mathcal{R}$ values using usual procedures for propagation of uncertainty [e.g., *Emery and Thomson*, 2001, section 3.16.2]. Considering these three P_a products and all time periods with a_η values significant at the $1\text{-}\sigma$ level (Figure 1.2), we find that $\mathcal{R}$ values are larger than their standard errors two-thirds

of the time (not shown). These results attest to the overall meaningfulness of the η^{ib} contributions to η the accelerations.

1.5 Interannual variations and extremes

Prompted by *Goddard et al.* [2015] among others, we investigate interannual variations over 1979–2013 at 18 tide gauges on coastal New England and Atlantic Canada (Table 1.1); these sites comprise the “northeast composite” of *Goddard et al.* [2015] and this period is chosen as it is the longest one mostly covered by the various P_a datasets. After removing quadratic fits to each time series over this period, we average tide gauge η records and the corresponding η^{ib} estimates from the different P_a products across the sites (Figure 1.3a). Large positive η anomalies during 1983, 1996–1998, and 2010 coincide with strong El Niño–Southern Oscillation values [*Sweet and Zervas*, 2011]. The striking η event in 2010 also occurred alongside other ocean circulation and climate anomalies elsewhere in and around the North Atlantic [e.g., *Landerer and Volkov*, 2013; *Tsimplis et al.*, 2013; *Piecuch and Ponte*, 2014] linked to extreme North Atlantic and Arctic Oscillation phases [e.g., *Cattiaux et al.*, 2010; *Cohen et al.*, 2010; *L’Heureux et al.*, 2010; *Taws et al.*, 2011].

The various η^{ib} curves from the different P_a products show very good agreement; they are highly correlated with one another and residual variances between pairs of time series are much smaller than variances of the curves themselves (Figure 1.3a). On average η^{ib} explains 25% of the interannual η variance (Figure 1.3b), where we have defined the percentage of variance $\mathcal{V}$ in a time series x explained by another series y as (using σ^2 to denote variance)

$$\mathcal{V} = 100\% \times \left[1 - \frac{\sigma^2(x - y)}{\sigma^2(x)} \right]. \quad (1.3)$$

Considering the largest η anomalies, we notice that η^{ib} contributes even more importantly; restricting focus to η anomalies with magnitudes larger than the

sample standard deviation, the interannual η variance explained by η^{ib} increases to 42% on average (Figure 1.3b).

The COREv2 η^{ib} estimate explains noticeably less η variance than η^{ib} from other P_a datasets (Figure 1.3b). This discrepancy reflects the η^{ib} contribution to the η anomaly in 2010 that is not captured by the former dataset which terminates in 2009 (Figure 1.3a). Indeed, between 2008 and 2010, annual mean η increased by 81 mm (Figure 1.3a); considering time series of centered two-year differences in η from Figure 1.3a (not shown), we see that this η change was 2.4 standard deviations above the 1980–2012 average difference value. (These numbers are slightly smaller than the corresponding values given by *Goddard et al.* [2015] using the same records mainly because we consider a shorter period and remove the background trend that mostly reflects global mean changes and vertical land motion.) During the 2008–2010 period, annual mean η^{ib} rose on average by 41 mm (Figure 1.3a), so we estimate that η^{ib} increase represented 51% of this η rise event on average (Figure 1.3c). Moreover, this pronounced η^{ib} change was largely responsible for the unique nature of this particular η event. Based on two-year differences of the η and η^{ib} time series (not shown), we find that the increase in η^{ib} between 2008 and 2010 was on average 2.9 standard deviations higher than the mean over 1980–2012 across P_a datasets, while the corresponding change in *dynamic* η (i.e., $\eta - \eta^{ib}$; 40 mm on average across the different P_a datasets) was only 1.4 standard deviations higher than its mean (Figure 1.3d).

1.6 Discussion

We used tide gauge records and barometric pressure datasets to assess the influence of the inverted barometer on sea level on the northeast coast of North America. Considering recent 40- to 64-year periods falling between 1950 and 2013, we found that, depending on the choice of barometric pressure dataset, the in-

verted barometer effect on average accounts for $\sim 10\text{--}30\%$ of significant sea level accelerations on the Mid-Atlantic Bight and Southern New England (Figure 1.2). This finding suggests that, while there are considerable uncertainties in available pressure datasets, the inverted barometer is not always negligible on multidecadal time scales, in broad agreement with *Ponte* [2006]. We also saw that the inverted barometer makes important interannual contributions, for example, amounting to $\sim 50\%$ of an extreme sea level rise over 2008–2010 and explaining $\sim 25\%$ of the interannual sea level variance during 1979–2013 at tide gauges along New England and Atlantic Canada (Figure 1.3). While not precluding a role for ocean dynamics, our results clarify the influence of static processes on sea level, informing efforts to reconstruct past ocean current transports [*Ezer*, 2015] and forecast future sea levels [*Slangen et al.*, 2014].

This work complements recent studies of sea level at tide gauges along the northeast coast of North America that largely focus on dynamic changes rather than static effects. Although *Calafat and Chambers* [2013] and *Haigh et al.* [2014] regress Boston and New York sea level onto barometric pressure, winds are also included in their regression, so it is unclear how much the inverted barometer contributes to sea level at these cities. *Kopp* [2013] notes that significant anti-correlation between the North Atlantic Oscillation and “nonlinear regional sea level anomaly” at Portland and Halifax is consistent with an inverted barometer, but he does not elaborate on how much this effect contributes at these sites. Our study builds on the foundation of these works, quantifying the inverted barometer’s influence on accelerated multidecadal sea level rise along the northeast coast.

Woodworth et al. [2014] find that quadratic fits to tide gauge records along the North American northeast coast “are largely unaffected by applying the inverse barometer correction or not” over 1950–2009 and 1960–2009. Similarly, *Boon*

[2012] states that removing the inverted barometer from the Boston tide gauge “had minimal effect on the regression coefficients” of a quadratic fit to the data over 1969–2011. Our findings do not contradict those of *Boon* [2012] and *Woodworth et al.* [2014], since one of our atmospheric reanalysis products also suggests a lesser role for the inverted barometer during these periods (Figure 1.2), but they do reveal that different pressure data can lead to distinct conclusions. Therefore, an important direction of future work, especially for interpretation of apparent sea level accelerations, will be better understanding of discrepancies between air pressure datasets at low frequencies, which could be due to differences in spatial resolution between products or other issues affecting reanalyses more generally (systematic errors in models, changes in the observing systems, etc.) [e.g., *Dee and Uppala*, 2009; *Simmons et al.*, 2010].

Sweet and Zervas [2011] interpret anomalously high sea levels along the northeast coast of the United States during the 2009/10 cool season without reference to the inverted barometer effect; they reconcile this event in terms of anomalous northeasterly winds over the Gulf of Maine connected to sea level pressure changes over the southeastern United States and eastern Canada linked to El Niño. Similarly, *Goddard et al.* [2015] reason that extreme sea level rise along Atlantic Canada and New England between 2008 and 2010 was largely connected to changes in the overturning circulation and wind stress anomalies related to exceptional values of the North Atlantic Oscillation; these authors posit that the inverted barometer contributed only $\sim 15\%$ to the extreme rise. Our findings, based on four different pressure datasets, give inverted barometer contributions around 50% (Figure 1.3), or roughly three times larger than those found by *Goddard et al.* [2015], and indicate that isostatic effects were of central importance to this unique sea level rise event.

Our results underscore the need for careful consideration of the inverted barometer effect and any associated uncertainties for interpretation of tide gauge sea level records, lest the influence of isostatic effects be misinterpreted as a reflection of ocean dynamics.

Acknowledgements

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Figure 1.1. Annual revised local reference sea level (in dm) over 1950–2013 from 30 tide gauges along the northeast coast of North America (Table 1.1). Linear trends have been removed from the records. Time series are offset in the vertical by an amount corresponding to the latitude of the respective tide gauge location. The colors correspond to the locations indicated in the inset.

Figure 1.2. Average sea level acceleration a_η (in mm yr^{-2}) along the Mid-Atlantic Bight and Southern New England for different recent periods as a function of period starting and ending years. Shading indicates values larger than their standard error while contouring denotes values smaller than their standard error (contour interval is 0.025 mm yr^{-2}). Inset shows, for different P_a datasets (HadSLP2 red; COREv2 blue; NOAA 20CR yellow) and period ending years (2009 solid; 2011 dashed), percentage contribution $\mathcal{R}$ of η^{ib} to a_η as a function of period starting year; note that the $\mathcal{R}$ values are only shown for time periods with a_η values significant at the $1\text{-}\sigma$ level.

Figure 1.3. **(a)** Interannual η from tide gauges (black) and η^{ib} from different P_a datasets (COREv2 blue; HadSLP2 red; NOAA 20CR yellow; NCEP-DOE R2 purple; ERA-Interim green) along New England and Atlantic Canada over 1979–2013 (in mm). Also indicated are magnitudes of the change over 2008–2010 (vertical lines) and η anomalies larger than the η standard deviation (dots). **(b)** Total interannual η variance explained by η^{ib} (bars) and η variance explained by η^{ib} when considering only the largest η anomalies as described in text (whiskers); **(c)** percent η^{ib} contribution to the η change over 2008–2010; and **(d)** magnitude of the 2008–2010 change in η^{ib} (whiskers) and $\eta - \eta^{ib}$ (bars) shown as standard deviations from the means of the respective centered difference time series over 1980–2012. The color coding in **(b)**, **(c)**, and **(d)** is as in **(a)**. In **(c)** and **(d)**, there are no estimates presented for COREv2 because this pressure dataset does not cover 2010.

Station Name	PSMSL ID	Longitude	Latitude	Years (Completeness)
Rimouski [†]	1597	−68.5167	48.4833	1985–2013 (0.83)
Port-aux-Basques [†]	392	−59.1333	47.5667	1959–2010 (0.77)
Argentia [†]	1321	−53.9833	47.3	1972–2011 (0.70)
Charlottetown [†]	427	−63.1167	46.2333	1950–2012 (0.83)
North Sydney [†]	1299	−60.25	46.2167	1971–2011 (0.88)
Eastport [†]	332	−66.9817	44.9033	1950–2013 (0.83)
Halifax [†]	96	−63.5833	44.6667	1950–2011 (0.87)
Bar Harbor [†]	525	−68.205	44.3917	1950–2013 (0.84)
Yarmouth [†]	1158	−66.1333	43.8333	1967–2011 (0.71)
Portland [†]	183	−70.2467	43.6567	1950–2013 (0.98)
Boston ^{†,‡}	235	−71.0533	42.3533	1950–2013 (0.97)
Providence ^{†,‡}	430	−71.4	41.8067	1957–2013 (0.95)
Woods Hole ^{†,‡}	367	−70.6717	41.5233	1950–2013 (0.88)
Newport ^{†,‡}	351	−71.3267	41.505	1950–2013 (0.97)
New London ^{†,‡}	429	−72.09	41.36	1950–2013 (0.92)
Nantucket Island ^{†,‡}	1111	−70.0967	41.285	1965–2013 (0.92)
Bridgeport ^{†,‡}	1068	−73.1817	41.1733	1965–2013 (0.92)
Montauk ^{†,‡}	519	−71.96	41.0483	1950–2013 (0.83)
New York [‡]	12	−74.0133	40.7	1950–2013 (0.95)
Sandy Hook [‡]	366	−74.0083	40.4667	1950–2013 (0.94)
Philadelphia [‡]	135	−75.1417	39.9333	1950–2013 (0.95)
Atlantic City [‡]	180	−74.4183	39.355	1950–2013 (0.83)
Baltimore [‡]	148	−76.5783	39.2667	1950–2013 (0.98)
Annapolis [‡]	311	−76.48	38.9833	1950–2013 (0.92)
Cape May [‡]	1153	−74.96	38.9683	1966–2013 (0.90)
Washington DC [‡]	360	−77.0217	38.8733	1950–2013 (0.97)
Lewes [‡]	224	−75.12	38.7817	1953–2013 (0.97)
Solomon’s Island [‡]	412	−76.4517	38.3167	1950–2012 (0.94)
Kiptopeke Beach [‡]	636	−75.9883	37.165	1952–2013 (0.97)
Sewells Point [‡]	299	−76.33	36.9467	1950–2013 (1.00)

Table 1.1. Tide gauges used here. Completeness is fraction of years for which data are available. Single daggers denote stations considered in the “northeast composite” of *Goddard et al.* [2015] that were used to produce the results in Figure 1.3. Double daggers denote stations considered in the “northeast hotspot” of *Sallenger et al.* [2012] that were used to produce results in Figure 1.2.

MANUSCRIPT 2

**Annual Sea Level Changes on the North American Northeast Coast:
Influence of Local Winds and Barotropic Motions**

by

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2.1 Abstract

Understanding the relation between coastal sea level and the variable ocean circulation is crucial for interpreting tide gauge records and projecting sea level rise. In this study, annual sea level records (adjusted for the inverted barometer effect) from tide gauges along the North American northeast coast over 1980–2010 are compared to a set of data-assimilating “ocean reanalysis” products as well as a global barotropic model solution forced with wind stress and barometric pressure. Correspondence between models and data depends strongly on model and location. At sites north of Cape Hatteras, the barotropic model shows as much (if not more) skill than ocean reanalyses, explaining $\sim 50\%$ of the variance in the adjusted annual tide gauge sea level records. Additional numerical experiments show that annual sea level changes along this coast from the barotropic model are driven by local wind stress over the continental shelf and slope. This result is interpreted in the light of a simple dynamic framework, wherein bottom friction balances surface wind stress in the alongshore direction and geostrophy holds in the across-shore direction. Results highlight the importance of barotropic dynamics on coastal sea level changes on interannual and decadal time scales; they also have implications for diagnosing the uncertainties in current ocean reanalysis, using tide gauge records to infer past changes in ocean circulation, as well as identifying the physical mechanisms responsible for projected future regional sea level rise.

2.2 Introduction

Physical oceanographers have long sought to understand the relation between sea level on the northeast coast of North America and ocean dynamics in the North Atlantic. Appealing to simple models of the coastal response (Csanady 1982), earlier studies considered the connection between sea level fluctuations and local atmospheric forcing over the shallow continental shelf. Using two years of data, Sandstrom (1980) reveals a strong link between adjusted sea level (that is, sea level corrected for the ocean's isostatic adjustment to barometric pressure changes) on the Nova Scotia shoreline and alongshore wind on the Scotian Shelf at periods > 20 days. This result is interpreted in light of a barotropic model, wherein the momentum balance is between wind stress and bottom drag in the alongshore direction, and geostrophic in the across-shore direction. Thompson (1986) investigates sea level changes from long tide gauge records on the western boundary of the North Atlantic north of Cape Hatteras. The author hypothesizes that, while they are partly effected by local air pressure and wind stress, mean sea level anomalies along this coastline are also influenced changes in a wind-driven, coastally trapped boundary current. Greatbatch et al. (1996) contrast simulations from a homogeneous ocean model forced with air pressure and wind stress to tide gauge data on the North Atlantic western boundary. These authors discern that the model faithfully reproduces the observed adjusted sea level behavior on synoptic time scales (periods of 3–10 days).

This topic has also enjoyed renewed interest over the last decade, owing to concerns over global climate change, and the possibility that the ocean circulation will change and coastal sea level will rise (e.g., Levermann et al. 2005; Landerer et al. 2007; Vellinga and Wood 2008; Yin et al. 2009). Based on geostrophic considerations and freshwater hosing experiments performed with a coarse resolution

model, Levermann et al. (2005) reason that a 1 Sverdrup ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) decline in the strength of the overturning streamfunction would be accompanied by a 4–5 cm rise in sea level on the North American east coast. Studying an eddy permitting ocean model, Bingham and Hughes (2009) find a qualitatively similar connection between ocean circulation and coastal sea level, such that a 1 Sv decline in the northward volume transport of the upper (100–1300 m) North Atlantic at 50°N is associated with a 2 cm increase in sea level along the northeast coast of North America. In their study of dynamic sea level projections from coupled climate models, Yin et al. (2009) warn that the United States northeast coast may experience rapid sea level rise over the next century in connection with a potential slowing of the Atlantic meridional overturning circulation.

Motivated by such modeling investigations, more recent studies have taken to the tide gauge record to see whether such mean sea level signatures of ocean circulation changes can be inferred, across a variety of time scales. Sallenger et al. (2012) identify a “hotspot” of accelerated sea level rise on the Atlantic coast of North America—a stretch of coastline from Virginia to Massachusetts along which the rate of sea level rise over the last few decades has been increasing $\sim 3\text{--}4\times$ faster than the global average rate. Comparing to previous climate model simulations, those authors suggest that the hotspot is consistent with a downturn in the meridional overturning circulation. Examining solutions from an Earth system model, Yin and Goddard (2013) make the argument that there was an overall northward shift in the Gulf Stream position over the last century, which contributed to coastal tide gauge sea level rise observed along the Mid-Atlantic Bight. Using tide gauge records between New York and Newfoundland, Goddard et al. (2015) determine that there was a striking interannual sea level rise event that recently occurred on the northeast coast of North America, which they partly

ascribe to a contemporaneous downturn in the overturning circulation.

These analyses have prompted contemporary investigations to consider in more detail what are the dynamical mechanisms underlying the interannual and decadal sea level changes observed along this shoreline. Andres et al. (2013) determine a significant correlation between a composite of annual coastal sea level anomaly (from tide gauges averaged over the Mid-Atlantic Bight, Gulf of Maine, and Scotian Shelf) during the period 1970–2012 and (1) alongshore wind stress locally over the continental shelf and (2) wind stress curl remotely over the Labrador Sea. They interpret their findings qualitatively in light of the barotropic model due to Sandstrom (1980). Woodworth et al. (2014) consider the tide gauge record along the northeast American Atlantic coast between Capes Hatteras and Breton Island over 1950–2009, showing a relationship between annual sea level from the data and solutions from the Liverpool/Hadley Centre ocean model driven by winds and thermohaline forcing.¹ In discussing their results, they appeal to simple linear models for the response of stratified, frictional flows on the continental shelf to large-scale, low-frequency wind variations (e.g., Csanady 1982; Clarke and Brink 1985), pointing to the importance of baroclinic signals trapped at the coast. Thompson and Mitchum (2014) show significant correlations between interannual sea level from tide gauges on the North Atlantic western boundary over 1952–2001 and contemporaneous time series from the German Estimating the Circulation and Climate of the Ocean (GECCO) state estimate. Those authors argue that a coherent mode of interannual sea level variability in this region is ultimately owing to Sverdrup flows over the interior of the ocean basin.

While their findings are not necessarily contradictory, and may pertain strictly to particular time periods and frequency bands, the authors of these more recent

¹To avoid any confusion (cf. Wunsch 2002), we use “thermohaline forcing” to mean the combination of surface heat and freshwater exchanges.

dynamical studies are highlighting very different mechanisms in their interpretations of the tide gauge records. Yet for reconstructing past shifts in the ocean’s general circulation (Bingham and Hughes 2009; McCarthy et al. 2015) and anticipating future coastal sea level rise (Landerer et al. 2007; Yin et al. 2009), it is important to distinguish between the relative contributions of different ocean processes to sea level changes observed in the tide gauge record. With the goal of better understanding coastal sea level behavior, and partly motivated by Andres et al. (2013), who suggest the importance of barotropic dynamics, we study tide gauges, ocean reanalyses, and a barotropic model to address the following questions:

- How well are year-to-year changes in sea level observed by tide gauges along the North American northeast coast reproduced by different ocean circulation models?
- Do barotropic processes contribute importantly to these observed sea level changes?
- What are the relative influences of local wind stress forcing over the shallow continental shelf versus remote wind driving over the deep open ocean?

The rest of this paper is structured as follows: in section 2, we describe methods and materials, namely the tide gauge data, ocean reanalysis products, and barotropic model solution; in section 3, we assess the skill of the ocean reanalyses and barotropic model in reproducing the tide gauge data; in section 4, model experiments are performed using the barotropic model to determine the roles of local and remote winds; finally, we conclude in section 5 with a discussion of our findings.

2.3 Methods and materials

2.3.1 Tide gauge records

To study sea level on the northeast coast of North America, we use annual revised local reference (RLR) records from 27 tide gauges (Table 1). Data were extracted from the Permanent Service for Mean Sea Level (PSMSL) database (Holgate et al. 2013; PSMSL 2015) on 16 February 2015. For reasons explained below, we study the sea level records over the 31-year period 1980–2010. The selection criteria satisfied by these records are that the tide gauges are situated along the eastern coast of North America, contain at least 20 years of valid annual sea level values over the study period, and are largely exposed to the open ocean (i.e., and not sheltered within large inland estuarine systems such as Chesapeake Bay, Delaware Bay, or the St Lawrence River). Other recent papers have used very similar subsets of the PSMSL RLR data (Sallenger et al. 2012; Andres et al. 2013; Thompson and Mitchum 2014; Woodworth et al. 2014; McCarthy et al. 2015). We note that, while our main focus will be on tide gauges along the northeast coast, we have also included some tide gauges along the southeast coast of North America for purposes of comparison (Fig. 2.1).

Here we focus on changes in dynamic sea level (ζ), hence we adjust the records for isostatic ocean response to barometric pressure (the inverted barometer effect), which can have an important impact on annual sea level changes in this area. For example, Piecuch and Ponte (2015) find that such air-pressure effects explain $\sim 25\%$ of the interannual variance over 1979–2013 and $\sim 50\%$ of the magnitude of an extreme event during 2009–2010 in tide gauge records along the northeastern coastline. To estimate the inverted barometer effect, we use annual sea level pressure (P_a) from the Hadley Centre Sea Level Pressure dataset (Allan and Ansell 2006). We use these P_a data because the PSMSL recommends them as “the most

suitable gridded data set ... for sea level studies”² [but note that different P_a datasets are very similar in this area over this period and give almost identical results (cf. Fig. 3 in Piecuch and Ponte 2015)]. Data are defined on a regular grid with horizontal resolution of 5° over 1850–2012. We assess the inverted barometer effect (ζ^{ib}) as (cf. Ponte 2006),

$$\zeta^{\text{ib}} \doteq -\frac{P_a - \overline{P_a}}{\rho g}, \quad (2.1)$$

where overbar is spatial average over the ocean, g is gravity, and ρ is ocean density. Values are mapped to gauge sites using nearest neighbor interpolation. Given our focus on ocean dynamics, we also remove estimated global mean sea level changes over the period (Church and White 2011).

Similar to recent works by Andres et al. (2013) and Thompson and Mitchum (2014), we restrict our focus to interannual and decadal changes. To isolate these time scales, we remove a linear trend from each of the annual tide gauge records. This serves to filter out changes over longer periods due to global sea level rise and local vertical land motion (Kopp 2013) and possibly also changes in thermohaline forcing and the Atlantic meridional overturning circulation (Yin and Goddard 2013). Consistent with previous studies (e.g., Bingham and Hughes 2009; Thompson and Mitchum 2014; Woodworth et al. 2014), we observe that the coastal ζ anomalies “cluster” into two distinct groups, which are demarcated by Cape Hatteras (Fig. 2.1). Pairs of tide gauges either north or south of Cape Hatteras are mostly significantly correlated with one another, whereas northern tide gauges do not show statistically significant correlation coefficients with the southern tide gauges (Fig. 2.2). [Critical values of the correlation coefficient are determined for all pairs of time series based on the autocorrelation properties of the records, after section 12.4.2 of von Storch and Zwiers (1999).] In what follows, we seek to

²For example, see <http://www.psmsl.org/train.and.info/geo-signals/atm.php>.

elucidate the dynamical mechanisms underlying these ζ fluctuations.

2.3.2 Ocean reanalysis products

To interpret the observed ζ anomalies (Fig. 2.1), we investigate output from four ocean reanalyses: National Centers for Environmental Prediction (NCEP) Global Ocean Data Assimilation System (GODAS; Behringer and Xue 2004; Xue et al. 2011), Simple Ocean Data Assimilation (SODA) Version 2.2.4 (Giese and Ray 2011; Chepurin et al. 2014), the recent synthesis from the German Estimating the Circulation and Climate of the Ocean (GECCO2) consortium (Köhl 2015), and the operational ocean reanalysis system (ORAS4) taken from the European Centre for Medium Range Weather Forecasting (ECMWF; Balmaseda et al. 2013). Reanalyses were chosen largely based on their availability and temporal coverage. While each solution assimilates some ocean observations, two of them (GECCO2 and ORAS4) bring in altimetry data away from the coast, and none incorporate tide gauge data. A detailed description of the products is given below.

We take annual-mean ζ time series from the reanalyses. Since some models may not be faithful right at the coast, especially where the shelf is narrow compared to the model resolution, for each reanalysis and tide gauge, we map the model to the data by selecting the reanalysis ζ time series from the grid cell within a 300-km radius around the gauge site that explains the most variance in the tide gauge record. Analogous methods have been used in recent studies that compare modeled and observational coastal sea level time series (e.g., Calafat et al. 2014; Dangendorf et al. 2014; Chepurin et al. 2014). [While our choice for the radius around the tide gauge is motivated by Chepurin et al. (2014), who use a similar value, we admit that 300 km is somewhat broader than the width of the continental shelf along this coastline ($\mathcal{O} \sim 100\text{--}200$ km). Note, however, that our findings are insensitive to this particular radius choice, and different choices lead us to effectively identical

conclusions.] Given the temporal overlaps of the reanalysis products, we study ζ over the common interval 1980–2010. As with the tide gauge records, linear trends have been subtracted from all the reanalysis time series and respective global mean time series have also been removed. (As none of the reanalyses includes pressure forcing, no inverted barometer adjustment is needed.)

2.3.3 Barotropic model solution

To complement our study of tide gauge ζ based on ocean reanalyses, we also use a barotropic³ model solution generated by the Massachusetts Institute of Technology general circulation model (Marshall et al. 1997). We configure the global ocean model to solve the Navier-Stokes equations for a homogeneous ocean driven by P_a and wind stress at the sea surface. The model grid has a nominal horizontal spacing of 1° using the same topology and bathymetry files as in the Estimating the Circulation and Climate of the Ocean (ECCO) version 4 ocean state estimate (Forget et al. 2015). Since this horizontal resolution is comparable to the width of the shelf in this region, this model cannot be expected to resolve the details of flows near the coast that are strongly constrained by fine topographic features. However, determining the skill of such a model (e.g., in reproducing tide gauge records) is still of interest, as Intergovernmental Panel on Climate Change-class models, used for sea level projections (e.g., Little et al. 2015), employ comparable horizontal grid spacings.

We force the model with surface fields from the ECMWF Reanalysis Interim (ERA-Interim; Dee et al. 2011), which covers 1979–2015 with a 0.75° horizontal grid spacing. A single layer is used in the vertical with variable ocean depths

³The word “barotropic” has been used variously (and sometimes confusingly) in the physical oceanography and sea level literatures. Generally speaking, a barotropic fluid is one in which the pressure and density surfaces align (e.g., Holton 1992), for example, so that ocean pressure gradients do not generate vorticity (e.g., Pedlosky 1992). Here we use the term in a more restrictive sense to mean a homogeneous ocean with constant density.

implemented using partial cells (Adcroft et al. 1997). The model uses a linear free surface, no-slip boundary conditions at the bottom and along the sides, a vertical eddy viscosity of $1 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$, quadratic bottom drag, and a horizontal eddy viscosity that varies with grid cell size. Observe that, because the model uses only one level in the vertical, the surface wind stress and frictional bottom boundary conditions are cast as body forces that act over the whole fluid column. The barotropic model setup makes use of a 900-s time stepping for the momentum equations along with a 3600-s time step for the free surface condition.

The model is started from rest using a 5-year spin-up period. During that time, it is driven with climatological P_a and wind stress, whereafter it is forced with monthly reanalysis fields. While the model uses low-frequency (monthly) forcing, we also performed runs using high-frequency (daily) forcing fields, but they yielded nearly identical annual ζ solutions (not shown) and so are not discussed any further. To be consistent with the tide gauge records and ocean reanalyses, we remove the inverted barometer effect from the barotropic model solution. As with the reanalyses, we match model and data annual ζ fields by taking the nearby model ζ time series that explains the most variance in the tide gauge record. We remove a linear trend during the 1980–2010 period.

2.4 Comparing models and data

A number of recent papers compare tide gauges to sea level from ocean models in different areas (Dangendorf et al. 2014; Calafat et al. 2014; Chepurin et al. 2014; Thompson and Mitchum 2014; Woodworth et al. 2014). In order to gain deeper physical insight, we revisit this important topic, examining the tide gauge records and ocean model solutions along the North American northeast coast. To infer how well models reproduce the data, we compute two quantities: (1) the correlation coefficient (r) and (2) the relative root-mean-square deviation (δ) between the

model and the data,

$$\delta \doteq \frac{\sigma(m - d)}{\sigma(d)}, \quad (2.2)$$

where m and d represent model and data ζ time series, respectively, while σ is standard deviation.

The relationship between the models and the data varies from place to place and from model to model. There are no tide gauge sites at which the ζ data are significantly correlated with the modeled record from GECCO2, ORAS4, or GODAS (Fig. 2.3b, c, e). Root-mean-square deviations between the data and either GECCO2 or GODAS are relatively large (greater than roughly 0.9; Fig. 2.3g, j). ORAS4 performs only slightly better in this regard, for example, yielding $\delta \sim 0.7$ at Fernandina Beach (Fig. 2.3h). These results are consistent with Köhl (2015), who shows that GECCO2 has little skill in reproducing altimetric ζ data in this area over 1993–2011. Such poor correlations are surprising, since an earlier GECCO solution shows good correlation over 1952–2001 with tide gauges in this region (Thompson and Mitchum 2014). These findings also accord with Chepurin et al. (2014), who reveal poor correlation between tide gauges and ORAS4 along this coastline over 1950–2008.

The barotropic model and SODA solution show better correspondence to the data along the northeast coast of North America. At most sites north of Cape Hatteras, SODA and the barotropic model both manifest statistically significant correlation coefficients with the tide gauge records (Fig. 2.3a, d). Additionally, these two solutions give relative root-mean-square deviations with the data that are considerably smaller than δ values based on the three other model products (Fig. 2.3f, i). However, despite their skill at sites north of Cape Hatteras, neither SODA nor the barotropic model compares well with the tide gauge data along the South Atlantic Bight, evidenced by insignificant correlation coefficients

(Fig. 2.3a, d) and elevated root-mean-square deviations (Fig. 2.3f, i). Calafat et al. (2014) and Dangendorf et al. (2014) present similar findings, demonstrating that the SODA model captures the annual tide gauge records better north of Cape Hatteras than south of this point.

Due to the alongshore coherence of the tide gauge records (Fig. 2.2), very similar conclusions regarding model performance follow from comparison of the models and data on larger scales. Figure 4 shows ζ time series from the different model and observational records averaged over the sites either north or south of Cape Hatteras, whereas the correspondence between models and data is summarized by the Taylor diagram (Taylor 2001) shown in Fig. 2.5. North of Cape Hatteras, SODA and the barotropic model both show significant correlations with the data; however, while the barotropic model underestimates the amplitude of the observed signal, SODA overestimates the observed signal's amplitude. GODAS similarly overestimates the observed magnitude along the northeast coastline, but this model solution shows poor correlation with the observational time series. South of Cape Hatteras, SODA and GODAS capture the observed signal amplitude, but neither of them is significantly correlated with the observations. Whereas ORAS4 and GECCO2 strongly underestimate the amplitude of the composite tide gauge record on the southeast coast, the barotropic model drastically underestimates the magnitude of this tide gauge ζ record (Figs. 2.4, 2.5).

The good correlation between tide gauges and the barotropic model along the northeast coast is consistent with previous studies. Based on a regression analysis, Andres et al. (2013) hypothesize that local winds and barotropic response are important to annual ζ changes along this shoreline. Similarly, Calafat and Chambers (2013) demonstrate that a multiple linear regression involving local wind and sea-level pressure can explain a substantial portion of the annual ζ variance

at the Boston and New York tide gauges. Moreover, the barotropic model’s poor performance south of Cape Hatteras is also in agreement with past works. Based on linear dynamics, Hong et al. (2000) reason that baroclinic response to open-ocean wind curl by means of Rossby waves is an important contributor to decadal ζ variability along the South Atlantic Bight. Bingham and Hughes (2012), using a high-resolution global ocean circulation model, show that interannual variations in seafloor density along the continental slope and deep ocean have more of an influence on coastal ζ changes south of Cape Hatteras, hence suggesting that there is a stronger decoupling between coastal ζ and deep steric signals to the north of Cape Hatteras. Moreover, numerical experiments considered by Woodworth et al. (2014) hint that thermohaline forcing effects ζ changes south of Cape Hatteras.

In summary, our results show that ocean models differ in their ability to reproduce annual ζ changes observed on the North American east coast. They also suggest that barotropic processes contribute appreciably to interannual and decadal ζ variance on the coast north of Cape Hatteras. To elucidate the relevant barotropic dynamics, in the section that follows we report on results from additional numerical forcing simulations that were performed based on the barotropic model setup.

2.5 Forcing experiments and dynamical interpretation

Our simple barotropic model solution performs as well as, if not better than, other more complete (and data-assimilating) ocean general circulation model frameworks with regard to reproducing annual tide gauge observations along the northeast coast of North America. This demonstrates that more complex models do not necessarily produce more realistic solutions. In the most general terms, the ζ signals from the barotropic model can reflect dynamic ocean response to barometric pressure and wind stress locally as well as remotely. To reveal the roles of

local and remote wind and pressure, we conduct the following experiments based on the barotropic model configuration:

- **PRES:** In this experiment, we again run forward the barotropic model as described previously, but we “turn off” the wind stress surface forcing. Hence, once corrected for the inverted barometer effect, this solution represents the dynamic ocean response to barometric pressure.
- **SHAL:** For this run, we set to zero barometric pressure and wind stress over the deep ocean, leaving the wind stress over the shelf and slope (< 1000 m) as the only driver of ζ variability.
- **DEEP:** Similar to SHAL, we remove pressure and wind forcing over the shallow ocean from this simulation, allowing only wind stress over the deep ocean (> 1000 m) to force the model.

In all other respects (e.g., initial conditions), these perturbation runs are identical to the original barotropic ocean model simulation, which hereafter we refer to as the **BASE** experiment for clarity.

The outcomes of the experiments are summarized in Fig. 2.6, which compares ζ time series from the BASE, PRES, DEEP, and SHAL simulations averaged over the tide gauge sites north of Cape Hatteras. [Due to the strong spatial coherence of the signals (Fig. 2.2), analogous conclusions follow from comparing the different barotropic model experiments at the various individual tide gauges (not shown).] The PRES experiment evidences no appreciable dynamic behavior in this region and explains none of the ζ variance from the BASE simulation (Fig. 2.6a). This result is not surprising, as the barotropic oceanic adjustment to pressure loading at these space and time scales is expected to be mainly isostatic and mostly explained by the inverted barometer response (e.g., Ponte 1993).

In sharp contrast, the ζ time series from the SHAL and BASE experiments are nearly identical—the correlation coefficient between them is 0.99 (Fig. 2.6b). This suggests that annual barotropic ζ fluctuations along the coast are driven by wind stress over the shelf and slope. The ζ fluctuations from the SHAL experiment are almost perfectly anticorrelated (coefficient of -0.99) with the local alongshore wind stress over the Mid-Atlantic Bight, Gulf of Maine, and Scotian Shelf (Fig. 2.7a). Andres et al. (2013) also find strong anticorrelation between alongshore wind stress and coastal sea level, but the relation shown in Fig. 2.7a is much stronger than the one they see (cf. their Fig. 4b), likely because, as we use the barotropic component from the model rather than tide gauge data, we have effectively removed the influence of wind stress over the deep ocean and barometric pressure.

Sandstrom (1980) provides a physical framework for interpreting this antiphase relationship between sea level and alongshore wind stress. Consider a shelf of width W and depth H along the coast. Suppose that the momentum balance in the alongshore direction (here $\hat{y}$) is between wind stress and bottom friction, and say that geostrophy holds in the across-shore direction ($\hat{x}$), whence,

$$-fv = -g \frac{\partial \zeta}{\partial x}, \quad (2.3)$$

$$\frac{\tau_y}{\rho H} = 2 \frac{A_v}{H} v, \quad (2.4)$$

where f is the Coriolis parameter, g is the gravitational acceleration, v and τ_y are the alongshore (i.e., meridional) velocity and wind stress, respectively, and A_v is a vertical eddy viscosity.⁴ If we assume that alongshore wind stress is constant, integrate across the shelf, and make substitutions with the equations, we obtain

⁴This form of vertical dissipation (i.e., with the pre-factor of 2 and inverse dependence on depth) is chosen to be consistent with the formulation of the no-slip bottom condition in the model (e.g., see section 2.14.6 in Adcroft et al. 2015), where we ignore quadratic bottom drag for simplicity.

the following relation between sea level and alongshore wind stress,

$$\Delta\zeta^+ = \frac{fW}{2A_v\rho g}\tau_y, \quad (2.5)$$

where $\Delta\zeta^+$ is the difference between coastal and offshore (i.e., at the edge of the shelf) sea level. Choosing values representative for the shelf along the North American northeast coast in the model ($f \sim 10^{-4} \text{ s}^{-1}$, $W \sim 200 \text{ km}$, $A_v \sim 10^{-3} \text{ m}^2 \text{ s}^{-1}$, $\rho \sim 10^3 \text{ kg m}^{-3}$, $g \sim 10 \text{ m s}^{-2}$), and supposing that sea level vanishes at the oceanward edge of the shelf, we find that Eq. (2.5) gives us a constant of proportionality between coastal sea level and alongshore wind stress of roughly $-1 \text{ m}^3 \text{ N}^{-1}$. This is very close to what we actually find in the SHAL experiment (Fig. 2.7a), and moreover it is consistent with the range given by Andres et al. (2013), which suggests that the barotropic mechanism described by Sandstrom (1980) and appealed to by Andres et al. (2013) is in fact an important contributor to interannual and decadal ζ change on the North American northeast coast.

Consistent with these findings, barotropic response to wind driving over the deep ocean has only a small influence, with ζ along the northeast coast from the DEEP experiment amounting to just $\sim 15\%$ of the coastal ζ variance from the BASE simulation (Fig. 2.6c). (The ζ signals from the SHAL and DEEP experiments covary, so their variances are not additive.) The ζ changes on coast from the DEEP simulation are correlated (coefficient of ~ -0.9) with wind stress curl forcing integrated zonally over the deep basin (Fig. 2.7b). Such a relationship between the coastal sea level and wind stress curl variations is anticipated in case of a barotropic Sverdrup balance; specifically,

$$\Delta\zeta^- = -\frac{f}{gD\beta\rho} \int \nabla \times \tau \, dx, \quad (2.6)$$

where $\Delta\zeta^-$ is the zonal difference in sea level across the ocean basin, β is the meridional derivative of f , D represents the depth of the deep ocean, and $\nabla \times \tau$

is the vertical component of the wind stress curl. (Here we have also assumed a β -plane ocean with a flat bottom.) Now supposing that ζ vanishes at the eastern boundary of the basin and using order-of-magnitude parameter values ($f \sim 10^{-4} \text{ s}^{-1}$, $D \sim 4000 \text{ m}$, $\beta \sim 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$, $\rho \sim 10^3 \text{ kg m}^{-3}$, $g \sim 10 \text{ m s}^{-2}$), we obtain a constant of proportionality between northeast coast sea level and the zonally integrated wind stress curl of about $0.25 \text{ m}^3 \text{ N}^{-1}$, which is on the order of what we see in the DEEP simulation (Fig. 2.7b), suggesting that barotropic Sverdrup balance is a plausible mechanism explaining this relationship.

2.6 Discussion

Previous investigations have studied the relation between coastal sea level and ocean circulation changes in observations of the past as well as projections of the future (e.g., Landerer et al. 2007; Bingham and Hughes 2009; Yin et al. 2009; Andres et al. 2013; McCarthy et al. 2015). Motivated by such works, we considered annual tide gauge sea level records along the North American east coast over the 1980–2010 period (Figs. 2.1, 2.2); these records were interpreted using different ocean circulation model solutions. We found that the correspondence between the data and models depends strongly on region and model—none of the models faithfully reproduce the coastal sea level changes observed south of Cape Hatteras, and only some models skillfully capture coastal sea level behavior measured north of Cape Hatteras (Figs. 2.3–2.5). Interestingly, we saw that a simple barotropic ocean model performed as well as (if not better than) more complex ocean reanalyses, which incorporate effects of buoyancy forcing and ocean stratification; this was apparent at tide gauge locations north of Cape Hatteras, where the barotropic model generally explains $\sim 50\%$ of the variance in the observational sea level records (Figs. 2.3, 2.5). Using this same barotropic ocean model framework, we also performed additional numerical simulations, variously driving the model with

wind stress or barometric pressure over different ocean regions (Figs. 2.6, 2.7). Based on those experiments, we reasoned that anomalous alongshore wind stress is the dominant driver of barotropic sea level variations along the North American northeast coast on these time scales (Figs. 2.6b, 2.7a); less relevant in this instance is wind curl forcing over the deep open ocean (Fig. 2.6c).

These findings improve our understanding of coastal sea level behavior and generally accord with previous works. Based on correlation and regression analyses, Andres et al. (2013) argue that a considerable portion of annual sea level variance in this region is controlled by local alongshore wind stress, consistent with what we found here (Figs. 2.4a, 2.6b). The numerical model experiments performed by Woodworth et al. (2014) hint that wind forcing contributes more to the coastal sea level variance north of Cape Hatteras than it does to the south (see their Fig. 2.6). This is in rough agreement with our results, suggesting that coastal sea level dynamics are distinct north and south of Cape Hatteras, with barotropic processes being more influential at locations north of this site than they are to the south (e.g., Fig. 2.4). However, we note that our results on this point contrast with the conclusions drawn by Yin and Goddard (2013) that baroclinic processes control dynamic sea level changes to the north of Cape Hatteras and barotropic effects dominate south of this point.

More generally, our conclusions corroborate previous global ocean modeling efforts suggesting that sea level and bottom pressure can be strongly coupled on shallow shelf sea regions even on interannual and longer time scales (Vinogradova et al. 2007; Bingham and Hughes 2008). But, we emphasize that the local barotropic mechanisms highlighted in this study account for roughly one half of the dynamic sea level variance along the northeast coast of North America (Figs. 2.3, 2.5), leaving a substantial fraction of the adjusted tide gauge variance to be explained.

Indeed, similar to the adjusted tide gauge records (Fig. 2.2), the residual time series (i.e., adjusted tide gauges minus barotropic model solution) evidence broad spatial coherence along the coast (not shown); these residual time series show significant correlation with the adjusted tide gauge records, but are not significantly correlated with the barotropic model solutions (not shown); these results possibly implicate mechanisms emphasized in other studies, e.g., zonal flows across the 65°W meridian (Thompson and Mitchum 2014) or baroclinic signals trapped at the coast (Woodworth et al. 2014).

We also performed various analyses (wavelet coherence, spectral analysis, etc.) in the frequency domain (not shown). The tide gauge and barotropic model sea level time series north of Cape Hatteras show stronger coherence at higher (inter-annual) frequencies and weaker coherence at lower (decadal) frequencies. Indeed, although removing the barotropic model solution reduces the spectral power of the tide gauge data at all frequencies, the residual difference between them is slightly red. These findings accord with basic theory of the oceanic response (e.g., Gill and Niiler 1973; Frankignoul et al. 1997), which says that ocean stratification effects become more important with decreasing frequency. Additionally, the relationship between tide gauge and barotropic model sea level north of Cape Hatteras seems not to be stationary. For example, the correlation coefficient between these two time series is 0.91 for the decade 1983–1993 but 0.43 for the decade 1994–2004. Somewhat similarly, Andres et al. (2013) find that the correspondence between northeast coast sea level and the North Atlantic Oscillation was stronger during 1987–2012 than during 1970–1986. This emphasizes that results here only apply to the time periods and frequency bands considered.

It is disconcerting that some ocean reanalysis products perform so poorly on this coastline (Figs. 2.3, 2.5). For them to yield meaningful projections of

future coastal sea level change, models must be able to represent processes at the boundaries and capture the coupling between sea level over the deep ocean and the shallow shelf (cf. Higginson et al. 2015; Hughes et al. 2015; Saba et al. 2016). To that end, understanding the reasons for the dispersion in model performance (Fig. 2.5) is imperative. Based on our findings (Figs. 2.6, 2.7), good estimates of local alongshore wind stress seem to be crucial for accurate simulations of sea level changes on the North American northeast coast. This suggests that the observed dispersion in model skill (Figs. 2.3–2.5) might be partly due to the different wind stress forcing fields used by the various models over this region. To assess this suggestion, we took alongshore wind stress time series over the North American northeast shelf from different atmospheric reanalysis products—including all those used as surface forcing in the ocean models considered here (see below)—comparing to the annual tide gauge sea level records averaged over this coastline (Fig. 2.8). We found that all alongshore wind stress products are significantly anticorrelated with the tide gauge records; after multiplying by the $-1 \text{ m}^3 \text{ N}^{-1}$ scale factor determined in the last section, the reanalysis wind stress time series explain 44–55% of the annual variance in the tide gauge sea level record, depending on the choice of atmospheric reanalysis. This suggests that uncertainties in alongshore wind stress and local barotropic response are probably not responsible for the discrepancies in the skills of the different ocean models in this region (Fig. 2.5); rather, these discrepancies must be owing to inaccurate representation of some other forcing or process (e.g., thermohaline forcing, ocean stratification, baroclinic response, etc.).

Based on global analyses, Hernandez et al. (2014) and Balmaseda et al. (2015) find that models that assimilate altimetric data and have finer resolution generally reproduce tide gauge records better than solutions that either are more coarse

or do not utilize altimetry. Thus, it might appear strange that the two models studied here that do incorporate altimetry (i.e., ORAS4 and GECCO2) perform poorly compared to other models that do not bring in this dataset (e.g., SODA).⁵ However, it must be kept in mind (see below) that neither ORAS4 nor GECCO2 uses altimetric data near land. Notwithstanding concerns over potentially degraded quality of satellite altimetry data near the coast, the correspondence between standard altimetric products and tide gauge records can be good in some coastal regions (e.g., Vinogradov and Ponte 2011), and so it could be that the assimilation methods are discarding valuable data at the coast. Indeed, as specially tailored coastal altimetry products (e.g., Passaro et al. 2015) come online and become more readily available, it will be important to bring them into ocean reanalyses for better representation of the coastal ocean.

Another consideration is that representation of bathymetry could affect the model performance. This point might be especially relevant south of Cape Hatteras, where the coupling of the deep sea and coastal ocean appears to be stronger, and where accurate representation of bathymetric gradients could be very important for communicating the influence of deep steric signals on coastal sea level (cf. Bingham and Hughes 2012). However, this issue might not be such a critical factor north of Cape Hatteras, seeing as GECCO2 (which performs poorly along this region) and our barotropic model (which does well in this area) use the same coastline and bathymetry input files. In any case, definitive determination of underlying causes for model discrepancies is beyond our scope; future works should focus in more detail on understanding such poor model performances.

Our results have other implications for interpreting past sea level changes and projecting future sea level rise. We have interpreted the coastal sea level

⁵The performances of these reanalyses that assimilate altimetry are not made any better if only the period 1993–2010 is considered (cf. Fig. 2.4).

behavior from the barotropic model in light of a framework similar to Sandstrom (1980)—bottom friction balances the wind stress in the alongshore direction, and geostrophy holds in the across-shore direction. This reasoning implies that these tide gauge records can be partly interpreted in terms of alongshore flow. For example, coastal sea level anomalies of 1–2 cm over a 200-km wide shelf would correspond to variations of 0.5–1.0 cm s⁻¹ in barotropic alongshore geostrophic currents, which amounts to 4–14% of mean flows observed along the southwest Nova Scotian Shelf (e.g., Hannah et al. 2001; Li et al. 2014).

Previous works consider projected overturning circulation changes and their bearing on coastal sea level rise (e.g., Landerer et al. 2007; Yin et al. 2009). Our results hint that future alongshore wind behavior should also be factored into such sea level rise scenarios. With this in mind, we considered projections of alongshore wind stress averaged over the North American northeast continental shelf from 1% yr⁻¹ CO₂ increase experiments from 29 coupled climate models as part of the fifth phase of the Climate Model Intercomparison Project (CMIP5; Taylor et al. 2012). We found that, while projected alongshore wind stress trends are mostly not statistically significant, some models do give significant positive trends, amounting to an increase of 0.01–0.02 N m⁻² over 140 years (not shown). Based on reasoning in the preceding section [Eq. (2.5)], this corresponds to a sea level drop of 1–2 cm along this stretch of coastline, which is small compared to the regional sea level rise anticipated during this coming century (e.g., Kopp et al. 2014; Slangen et al. 2014). We also found that, for a great majority (93%) of models considered, there is no significant change in the interannual alongshore wind stress variance over the duration of the simulation (not shown).

Goddard et al. (2015) examine tide gauge records on the northeast coast of North America and reveal an extraordinary rise in annual sea level between 2008

and 2010. Considering transport data, climate models, and an ocean data assimilation product, those authors conclude that this extreme sea level fluctuation was related to a contemporaneous downturn in the overturning circulation and wind stress anomalies associated with strong values of the North Atlantic Oscillation. Taken together with the findings of Piecuch and Ponte (2015), our barotropic model runs (Figs. 2.4a, 2.6b, 2.7a) suggest that this sea level rise event can be understood almost entirely in terms of the dynamic and isostatic ocean responses to local meteorological conditions over the shelf. This emphasizes that, while sea level and ocean circulation are correlated (e.g., Bingham and Hughes 2009), the Atlantic meridional overturning circulation is not directly coupled to observed sea level changes along the North American northeast coast over these time scales. However, as suggested by one reviewer, this does not preclude a more indirect link to the overturning circulation. For instance, Bryden et al. (2014) argue that the sharp reduction in the overturning circulation (and associated meridional heat transport) during 2009–2010 lead to an anomalous atmospheric state over the North Atlantic sector, whose influence was subsequently felt at the coast (cf. Goddard et al. 2015). Anyways, the extent to which overturning circulation and coastal sea level changes share common forcing, result from distinct (but still simultaneous) mechanisms, or are intimately coupled through complex ocean-atmosphere interactions should be explored in more detail in future investigations.

We have focused on sea level along the northeast coast of North America on interannual and decadal time scales. However, other studies point to interesting sea level behavior on this shoreline on multidecadal periods. For example, Chambers et al. (2012) reveal a prominent multidecadal fluctuation in the New York and Baltimore tide gauge records; these authors generally suggest that redistribution by oceanic Rossby or Kelvin waves may contribute to such regional sea level signals.

Analogously, based on a lagged correlation analysis considering European tide gauges, Miller and Douglas (2007) suggest that westward wave propagation could result in multidecadal sea level oscillations at tide gauges between Halifax and Baltimore. However, it remains to determine how important variations in more local meteorological conditions are to multidecadal sea level changes along the coast. These important questions are beyond our current scope, and left for future study.

2.7 Description of ocean reanalysis products

The SODA solution spans 1871–2010 and is defined on a grid with a $0.4^\circ \times 0.25^\circ$ horizontal spacing and 40 vertical levels. (Fields are provided interpolated onto a regular 0.5° horizontal grid.) Observations of ocean temperature and salinity from the World Ocean Database 2009 (Boyer et al. 2009) and sea surface temperature from the International Comprehensive Ocean-Atmosphere Dataset Release 2.5 (Woodruff et al. 2011) are assimilated using the sequential scheme described by Carton and Giese (2008). Forcing fields are based on NOAA 20CR (Compo et al. 2011) and the ocean model is based on the Parallel Ocean Program (POP) Version 2.0.1 (Smith et al. 1992).

The GODAS product covers 1980–2015. It is defined on a quasi-global (75°S – 65°N) ocean grid with a nominal lateral resolution of 1° (but reducing to $1/3^\circ$ in the tropics) and 40 levels in the vertical. Using a three-dimensional variational (3DVAR) method, this solution incorporates Reynolds sea surface temperature, and *in situ* temperature from expendable bathythermographs, profiling floats, as well as moorings from the Tropical Atmosphere Ocean (TAO) project, but not altimetry. The basic forcing fields are surface fluxes of momentum, heat, and freshwater from the NCEP Reanalysis 2 (Kanamitsu et al. 2002), and the baseline ocean general circulation model is the Geophysical Fluid Dynamics Laboratory’s

(GFDL) Modular Ocean Model (MOM) Version 3.

The ORAS4 solution spans 1958–2014 and is defined on a tripolar spatial grid, which has a nominal horizontal spacing of 1° , telescoping to 0.3° near the equator, with 42 vertical levels. It is generated using the Nucleus for European Modelling of the Ocean (NEMO) model (Madec 2008) and assimilates Reynolds surface temperature, satellite ζ , and temperature and salinity data from the EN3 biased-corrected database (Ingleby and Huddleston 2007) using the NEMOVAR method described by Mogensen et al. (2012) and with a 10-day assimilation window; a noteworthy aspect of this methodology is that the influence of observational data (including altimetry) on the solution is deemphasized in more coastal ocean regions (Mogensen et al. 2012). Surface temperature and sea-ice information are used along with a Newtonian relaxation scheme to constrain the upper levels. The atmospheric forcing until 1989 is from the ECMWF Reanalysis 40 (ERA40; Uppala et al. 2005), over 1989–2010 from the ECMWF Reanalysis Interim (ERA-Interim; Dee et al. 2011), while from 2010 onwards it is based on the ECMWF operational archive (Balmaseda et al. 2013).

The GECCO2 product is a global ocean state estimate over the period 1948–2011. It is defined on a spatial grid with nominal 1° spacing but reducing to $1/3^\circ$ close to the equator and effectively 40 km in the Arctic. (Interpolated solutions are provided on a regular 1° grid.) This solution is generated using the Massachusetts Institute of Technology general circulation model (MITgcm; Marshall et al. 1997). It employs the adjoint (or 4D-VAR) method to incorporate various satellite and *in situ* measurements including AVISO along-track ζ , mean dynamic topography, sea surface temperature from the AMSR-E satellite mission and the Hadley Centre Sea Ice and Sea Surface Temperature data set (Rayner et al. 2003), as well as subsurface temperature and salinity from the EN3 database (Ingleby and Huddle-

ston 2007). Note that altimetric ζ fields are assimilated into the estimate only over regions deeper than 130 m. Bulk formulae are used for the adjusted surface forcing fields, which are based on the NCEP Reanalysis 1 (Kalnay et al. 1996; Kistler et al. 2001).

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Figure 2.1. **(a)** Filled colored circles show the locations of the 27 PSMSL RLR (Holgate et al. 2013) tide gauges used in this study. The white star denotes Cape Hatteras and the grey contour delineates the 100-m depth contour. Annual sea level records from those tide gauges **(b)** north and **(c)** south of Cape Hatteras, with the colors corresponding between panels. Inverted barometer and linear trend have been removed from the records.

Figure 2.2. Correlation coefficient between pairs of annual mean sea level time series. Site numbers correspond to the values given in Table 1. Filled circles are correlation coefficients statistically significant at the 95% confidence level. Critical correlation coefficient values, determined for each pair of time series (von Storch and Zwiers 1999), are usually on the order 0.6–0.7. The black dashes separate sites north and south of Cape Hatteras.

Figure 2.3. Correlation coefficient r (top row) and relative root-mean-square deviation δ (bottom row) between annual tide gauge records and sea level time series from the barotropic model (first column), GECCO2 (second column), ORAS4 (third column), SODA (fourth column), and GODAS (fifth column). Correlation values in the top row with filled circles are statistically significant at the 95% confidence level (von Storch and Zwiers 1999).

Figure 2.4. Observed and modeled sea level averaged over tide gauges north (left column) or south (right column) of Cape Hatteras. (See Fig. 2.1a for locations.) The black curves are the tide gauge time series while the colored curves in the different rows indicate the various model solutions: the barotropic model (blue; first row); GECCO2 (orange; second row); ORAS4 (yellow; third row); SODA (purple; fourth row); and GODAS (green; fifth row).

Figure 2.5. Taylor diagram summarizing the correspondence between tide gauge records averaged north (circles) and south (squares) of Cape Hatteras and the corresponding sea level time series from the barotropic model (blue), GECCO2 (orange), ORAS4 (yellow), SODA (purple), and GODAS (green). Along the radial coordinate of the diagram is shown the standard deviation of the simulated ζ record divided by the standard deviation of the corresponding observational time series; along the azimuthal coordinate is shown the correlation coefficient r between the modeled and observed time series; and emanating from the reference point [i.e., the coordinate pair (1,1) starred on the diagram] is the relative root-mean-square deviation δ between the model and gauge records. The only significant correlation values are those from SODA and the barotropic model north of Cape Hatteras. (Note that the orange circle, corresponding to the performance of the GECCO2 product north of Cape Hatteras, is not missing from the figure, but rather falls outside the axis limits, owing to a negative correlation coefficient.)

Figure 2.6. Annual sea level averaged over 20 tide gauges north of Cape Hatteras from the different barotropic model runs. Black curves in each panel are identical and represent the sea level time series from the original simulation (BASE). Grey curves in the different panels are the sea level changes averaged over the sites from the different forcing experiments—(a) PRES, (b) SHAL, and (c) DEEP. The dependent axes are all in units of mm.

Figure 2.7. **(a)** Sea level from SHAL averaged over the 20 tide gauge sites north of Cape Hatteras (black) versus minus the average alongshore wind stress (here denoted $\tau_{\parallel}$) over the northeast continental shelf (grey). Here we define the alongshore wind stress as the inner product between wind stress vector $\vec{\tau} = (\tau_x, \tau_y)$ and an alongshore unit vector $\hat{n} = (\cos \vartheta, \sin \vartheta)$, where we have chosen $\vartheta = 30^\circ$. We define the extent of the northeast continental shelf as the region between $53\text{--}100^\circ\text{W}$ and $35\text{--}45^\circ\text{N}$ where the ocean depth is < 1000 m. **(b)** Sea level from DEEP averaged over the 20 tide gauge sites north of Cape Hatteras (black) versus wind stress curl integrated zonally across the deep ocean (> 1000 m) and averaged over $35\text{--}45^\circ\text{N}$ (grey). All the time series are detrended.

Figure 2.8. Sea level and alongshore wind on the northeast coast. Black curve in each panel is observed sea level record averaged over the 20 tide gauges north of Cape Hatteras (cf. Fig. 2.1a). Various colored curves in the different panels are the sea level predicted by averaging detrended annual alongshore wind stress anomalies over the shelf from various atmospheric reanalyses and scaling by $-1 \text{ m}^3 \text{ N}^{-1}$ (see the text for more details): (a) NOAA-20CR (blue; Compo et al. 2011), (b) ERA-20C (orange; Poli et al. 2013), (c) ERA-Interim (yellow; Dee et al. 2011), (d) NCEP-NCAR (purple; Kalnay et al. 1996), and (e) NCEP-DOE (green; Kanamitsu et al. 2002). We define alongshore wind stress and shelf extent as in Fig. 2.6 caption. Dependent axes have units of mm.

No.	Station Name	PSMSL ID	Longitude	Latitude	Completeness
1	Mayport	316	−81.4317	30.3933	65%
2	Fernandina Beach	112	−81.465	30.6717	81%
3	Fort Pulaski	395	−80.9017	32.0333	97%
4	Charleston I	234	−79.925	32.7817	100%
5	Springmaid Pier	1444	−78.9183	33.655	68%
6	Wilmington	396	−77.9533	34.2267	97%
7	Duck Pier	1636	−75.7467	36.1833	77%
8	Lewes	224	−75.12	38.7817	97%
9	Cape May	1153	−74.96	38.9683	100%
10	Atlantic City	180	−74.4183	39.355	74%
11	Sandy Hook	366	−74.0083	40.4667	90%
12	Bergen Point	1637	−74.1417	40.6367	65%
13	New York	12	−74.0133	40.7	90%
14	Montauk	519	−71.96	41.0483	71%
15	Bridgeport	1068	−73.1817	41.1733	94%
16	Nantucket Island	1111	−70.0967	41.285	90%
17	New London	429	−72.09	41.36	94%
18	Newport	351	−71.3267	41.505	100%
19	Woods Hole	367	−70.6717	41.5233	87%
20	Providence	430	−71.4	41.8067	94%
21	Boston	235	−71.0533	42.3533	94%
22	Portland	183	−70.2467	43.6567	97%
23	Yarmouth	1158	−66.1333	43.8333	65%
24	Bar Harbor	525	−68.205	44.3917	77%
25	Cutler II	1524	−67.2967	44.6417	77%
26	Halifax	96	−63.5833	44.6667	74%
27	Eastport	332	−66.9817	44.9033	84%

Table 2.1. Tide gauge records used here. The completeness is the percentage of years over 1980–2010 for which valid records are available. Tide gauges 1–6 (7–27) are located south (north) of Cape Hatteras (see Fig. 2.1).

MANUSCRIPT 3

**Air-Pressure Effects on Sea-Level Changes during the Twentieth
Century**

by

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3.1 Abstract

Interpretation of tide gauge data in terms sea level (η) and ocean dynamics requires estimates of air pressure (p_a) to determine the ocean’s isostatic response—the inverted barometer effect (η_{ib}). Three gridded p_a estimates (HadSLP2, NOAA-20CRv2, ERA-20C) are used alongside meteorological station p_a and tide gauge η records to evaluate the contribution of η_{ib} to η changes over the Twentieth Century. Agreement between gridded estimates is better during more recent periods and over regions with good historical data coverage, whereas it is worse for earlier time periods or in ocean areas with poor observational data coverage. Comparison against station data reveals the presence of systematic errors in the gridded estimates, for example, such that uncertainties estimated through differencing the gridded estimates underestimate the true errors by roughly 40% on interannual and decadal time scales. Notwithstanding such correlated errors, gridded estimates are still *useful* for interpretation of tide gauge data. Removing η_{ib} estimates from η records reduces spatial variance in centennial trends across tide gauges by 10–30%, formal errors in centennial trends from individual gauges by $\sim 5\%$, and the temporal variance in detrended records by 10–15% on average (depending on choice of gridded estimate). Results here advocate for making the η_{ib} correction to tide gauge records in studies of ocean circulation and global η over long, multidecadal and centennial time scales using an ensemble mean taken across several gridded η_{ib} estimates.

3.2 Introduction

Tide gauges furnish some of the longest instrumental records of the ocean. These records contain valuable signals related to global sea level and the ocean general circulation (Douglas 1991, 1992). However, tide gauge records must be interpreted carefully because they also measure the influences of processes unrelated to changes in the ocean’s circulation or its volume (Tamisiea et al. 2014). Surface loading owing to barometric pressure is one such process. Over periods longer than a few weeks, the oceanic adjustment to barometric pressure is more or less isostatic (Ponte 1992, 1993). The sea level responds like an “inverted barometer” in these cases (Wunsch and Stammer 1997), with water volume being redistributed in such a way that gradients in sea level cancel air pressure gradients, so that there is no signature in subsurface pressure, and no change in oceanic circulation.

To isolate the “signal” of changing ocean circulation or rising sea level, it is advisable to remove the “noise” related to the inverted barometer effect from tide gauge records. However, in numerous studies of global sea level (Church and White 2011; Ray and Douglas 2011; Jevrejeva et al. 2014) or the ocean general circulation (Sallenger et al. 2012; Goddard et al. 2015; McCarthy et al. 2015) over the last century based on tide gauges, no correction is made for the inverted barometer effect. Some authors reason that this correction is small on century time scales (Woodworth et al. 2009), while others argue that, due to increased scarcity of reliable air pressure data earlier in time, “no pressure adjustment is preferable to a partial or error ridden adjustment” (Ray and Douglas 2011).

Notwithstanding such concerns, the effects of air pressure on changes in sea level over very long (multidecadal and centennial) time periods, and the uncertainties associated with estimating them, have not been explored in any real systematic fashion. For example, some studies comment on the consistency between air pres-

sure datasets or the impact of air pressure on sea level at a single tide gauge site (Woodworth et al. 2010; Sturges and Douglas 2011; Dangendorf et al. 2014) or a small handful of locations (Kolker and Hameed 2007; Miller and Douglas 2007; Thompson et al. 2014). Since they are based on a limited number of tide gauge records, or a single pressure dataset, it is difficult to generalize their findings. The quality of available pressure datasets and the relevance of the inverted barometer correction more generally as a function of space and time remains unclear.

Ponte (2006) studies the inverted barometer effect over the ocean during the period 1958–2000. He shows that pressure effects explain up to 40% of the variance in tide gauge records on monthly time scales, and that correcting for the inverted barometer reduces formal errors in sea level trends. Ponte recommends that, as more air pressure datasets become available, the impact of the inverted barometer correction on tide gauge records should be revisited for longer, centennial time periods. Piecuch and Ponte (2015) find that air pressure effects made important contributions to recent sea level changes on the North American northeast coast (Sallenger et al. 2012; Goddard et al. 2015), underscoring that such isostatic effects should be carefully removed in dynamical sea level studies. They also reveal that estimated inverted barometer contributions to sea level changes in this region on multidecadal time scales can be sensitive to the choice of air pressure dataset, which advocates for future investigations to better characterize uncertainties in these datasets over long time scales.

In light of the burgeoning number of pressure datasets covering the last century, it is timely to revisit the topic of sea level and the inverted barometer effect. We address the following questions:

1. How consistent are air pressure reconstructions and atmospheric reanalysis products?

2. Are these gridded air pressure estimates characterized by systematic (i.e., correlated) errors?
3. How much does the inverted barometer effect contribute to changes in sea level?
4. What is the best strategy for adjusting tide gauges for the inverted barometer effect?

The remainder of this study is structured as follows: in section 2, we describe the main datasets (i.e., meteorological station data, tide gauge records, and gridded reconstructions and atmospheric reanalyses); in section 3, we compare gridded air pressure estimates to evaluate their consistency; in section 4; we contrast those gridded estimates to a few rigorously vetted long meteorological station records to test for the presence of any systematic errors; in section 5, we evaluate inverted barometer effects on a carefully selected set of long tide gauge records; in section 6, we summarize the results, discussing the best practices for correcting tide gauge records for effects of air pressure.

3.3 Materials and methods

We investigate barometric pressure (p_a) changes and their contribution to sea level (η) changes. To this end, we use p_a from reconstructions, atmospheric reanalyses, and meteorological stations, as well as η from tide gauges. We assess annual mean values over the common period 1900–2000.

3.3.1 Gridded p_a products

We use three gridded p_a products: (1.) Hadley Centre monthly historical mean sea level pressure reconstruction (HadSLP2) from Allan and Ansell (2006); (2.) National Atmospheric and Oceanic Administration Twentieth Century Reanalysis Project Version 2 (NOAA-20CRv2) from Compo et al. (2011); and (3.)

European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis of the twentieth century (ERA-20C) from Poli et al. (2013). These gridded estimates are described in more detail in section 3.8. In the broadest strokes, these products are generated by blending together gappy, sparse, discontinuous p_a observational data based on advanced statistical methods. These products incorporate common datasets, and are not truly independent, but can differ in terms of analysis method, temporal coverage, horizontal resolution, quality control, bias correction, and detection of outliers. We evaluate the inverted barometer adjustment (η_{ib}) following Ponte (2006),

$$\eta_{\text{ib}} = -\frac{p_a - \bar{p}_a}{\rho g}, \quad (3.1)$$

where overbar is mean over the global ocean, ρ is ocean density, and g is gravitational acceleration.

3.3.2 Station data

We consider long p_a time series from a few meteorological stations: Darwin, Chennai, Nagasaki, Azores, Gibraltar, Iceland, and Tahiti (Figure 3.1). These data have undergone intense scrutiny and quality control, which involves deleting dubious values, adjusting for step changes, comparing to paleoclimate data, and concatenating station records (e.g., Allan et al. 2002; Können et al. 2003). These records are taken from a NOAA-hosted website for the Working Group on Surface Pressure (http://www.esrl.noaa.gov/psd/gcos_wgsp/Timeseries/) given with precision of 0.1 hPa.

3.3.3 Tide gauges

We also consider some long η records from coastal tide gauges (Figure 3.1) from the Permanent Service for Mean Sea Level [PSMSL (e.g., Holgate et al. 2013)] Revised Local Reference (RLR) annual database (www.psmsl.org/data/). To

avoid studying short, unrepresentative records, or ones that are overly influenced by solid earth processes, we applied stringent selection criteria to the tide gauges. The 21 records used here: (1.) have at least 70 annual values during 1900–2000; (2.) are from sites where vertical land motion trend due to postglacial rebound is less than half the trend shown by the tide gauge record; and (3.) are from sites where important vertical land motion unrelated to glacial isostatic adjustment has not been reported. The only correction applied to the records is that postglacial rebound trends are subtracted based on values from Peltier et al. (2015).

3.4 Changes in η_{ib} from gridded products

In this section, we compare η_{ib} changes over the global ocean during the Twentieth Century from the gridded reconstruction and reanalysis products to determine the consistency of these datasets. We find that discrepancies between gridded estimates, in terms of trends and variations, are smaller for more recent periods and over regions of the ocean having good historical data coverage, while they are larger for earlier time periods or areas with poorer historical data coverage (Figures 3.2, 3.3).

3.4.1 Centennial trends

We compute η_{ib} trends from the p_a estimates using least squares. Let the η_{ib} trend from estimate $A \in \{H, N, E\}$ (where H , N , and E represent HadSLP2, NOAA-20CRv2, and ERA-20C) be T_A . Suppose that T_A represents the sum of the signal (true trend T_S) plus noise (error trend T'_A), whence,

$$T_A = T_S + T'_A. \quad (3.2)$$

The sample variance in trends across the estimates s_T^2 then relates to the error trends according to,

$$s_T^2 \doteq \underbrace{\frac{\sum_A (T_A - \bar{T})^2}{n-1}}_{\chi_a} = \underbrace{\frac{T_H'^2 + T_N'^2 + T_E'^2}{3}}_{\chi_b} - \frac{T_H' T_N' + T_H' T_E' + T_N' T_E'}{3}. \quad (3.3)$$

where $n = 3$ is the number of estimates and $\bar{T} \doteq n^{-1} \sum_A T_A$ is the sample mean trend over estimates. In case of random errors, the sample variance χ_a approaches the mean squared error trend χ_b ; for common (positively correlated) errors across estimates, χ_a represents a lower bound on χ_b .

The sample mean trend $\bar{T}$ shows clear meridional structure (Figure 3.2a). There are stronger positive trends at high southern latitudes ($> 0.1 \text{ mm yr}^{-1}$), more moderate negative trends at middle latitudes ($\sim -0.1 \text{ mm yr}^{-1}$), weaker negative trends ($> -0.1 \text{ mm yr}^{-1}$) at low latitudes, and weaker positive trends ($< 0.1 \text{ mm yr}^{-1}$) at high northern latitudes. Noteworthy features are also apparent more regionally about centers of action like the Mascarene High and Aleutian Low. By virtue of Eq. (3.1), these trends can be interpreted in terms of near-surface winds. Positive η_{ib} trends at southern high latitudes and negative η_{ib} trends at southern middle latitudes (Figure 3.2a) correspond to p_a decrease over the former region and p_a increase over the latter region. These patterns are consistent with the observed strengthening of westerly circumpolar flows linked to depletion of stratospheric ozone (e.g., Thompson and Solomon 2002; Gillett and Thompson 2003).

Stammer and Hüttemann (2008) study the η_{ib} adjustment in climate change model simulations. In response to quadrupled atmospheric CO_2 , they observe weak decreases in η_{ib} (-0.2 mm yr^{-1}) at low and middle latitudes and notice strong η_{ib} increases ($0.4\text{--}0.6 \text{ mm yr}^{-1}$) at high latitudes. They relate the η_{ib} changes at low latitudes to increases in vertically integrated water vapor linked to atmospheric

warming and increasing moisture content in their model, while they ascribe the η_{ib} trends at middle and high latitudes to atmospheric circulation changes and air mass redistribution. The similarity between our trends (Figure 3.2a) and theirs (cf. their Figure 4) suggests that global warming has influenced η_{ib} trends over the Twentieth Century (cf. Stammer and Hüttemann 2008).

To assess uncertainties in the trends, we compute the square roots of the trend sample variances [i.e., $\sqrt{\chi_a}$] (Figure 3.2b). Values show correspondence to patterns of historical data availability. Comparatively low values ($< 0.05 \text{ mm yr}^{-1}$) appear over midlatitude North Atlantic and Pacific Ocean areas, and in the tropical Indian Ocean, along shipping routes where data are more dense. Relatively high ($> 0.3 \text{ mm yr}^{-1}$) values are apparent over the Southern and Arctic Oceans, where the data are more sparse, even for more recent times (cf. Figure 1 from Allan and Ansell 2006). A meridional gradient is also observed, such that values are generally larger at higher latitudes. Averaging values in Figure 3.2b over the ocean gives a representative error value of $\sim 0.1 \text{ mm yr}^{-1}$.

Figure 3.2c shows signal-to-noise ratios, which we define as magnitude of sample mean divided by the square root of the sample variance [i.e., $|\bar{T}|/\sqrt{\chi_a}$]. Larger values are apparent in regions with strong mean trends (large signal) and weak trend variance (small noise), while smaller values are observed in regions with weak mean trends (small signal) or strong trend variance (large noise).

3.4.2 Detrended fluctuations

Next, we examine the fluctuations in η_{ib} that remain after removing the trends studied previously (Figure 3.2a-c). Similar to the trend analysis, we interpret each time series as signal plus noise, *viz.*,

$$A(t) = S(t) + A'(t), \quad (3.4)$$

where $A(t)$ represents the time series detrended fluctuations from any particular gridded estimate, $A'(t)$ the time series of error fluctuations, and $S(t)$ the time series of the true detrended fluctuations. Variances in the differences between products can be related to error variances and covariances as,

$$\underbrace{\frac{\langle (H - N)^2 \rangle + \langle (H - E)^2 \rangle + \langle (N - E)^2 \rangle}{6}}_{\psi_a} = \underbrace{\frac{\langle H'^2 \rangle + \langle N'^2 \rangle + \langle E'^2 \rangle}{3}}_{\psi_b} - \frac{\langle H'N' \rangle + \langle H'E' \rangle + \langle N'E' \rangle}{3}. \quad (3.5)$$

where brackets $\langle \cdot \rangle$ are time mean. For uncorrelated errors, the left-hand side ψ_a will tend toward the mean error variance ψ_b . In the case of systematic errors, ψ_a will be a lower bound on ψ_b .

To garner a sense of the patterns of the variability, we show a map of root mean variances (RMV) [i.e., $\sqrt{n^{-1} \sum_A A(t)^2}$] over the ocean in Figure 3.2d. There is a broad meridional structure, such that values are lower (< 10 mm) at low latitudes and higher (> 15 mm) at high latitudes. Largest values (~ 20 mm) occur over the Arctic Ocean and Pacific sector of the Southern Ocean, while smallest values (< 5 mm) occur over the central equatorial Pacific Ocean. This increase in variability with latitude has been seen in previous intraseasonal or interannual η_{ib} or p_a studies (e.g., Ponte 1993; Fu and Pihos 1994; Ponte 2006). Such behavior is expected in case of quasi-geostrophic balance: given the change in Coriolis parameter with latitude, the magnitude of p_a variation required to support a given wind fluctuation will increase towards the poles and decrease towards the equator.

We evaluate $\sqrt{\psi_a}$ over the ocean to see how well the estimates agree (Figure 3.2e). A latitudinal gradient is once again apparent. Differences between products are lower (< 5 mm) at low latitudes and higher (> 10 mm) at high latitudes. The smallest values (< 2 mm) are observed along the equatorial Indian Ocean, whereas the largest values (> 18 mm) are observed over the Pacific sector

of the Southern Ocean. Allan and Ansell (2006) point to the southeastern Pacific Ocean as a region of especially pronounced uncertainty in the HadSLP2 product due to sparse historical data coverage. Ponte (2006) shows similar patterns based on a comparison between annual fields from two reanalysis products over 1958–2000, highlighting the Pacific sector of the Southern Ocean as the area of largest differences between products. Compared to Ponte (2006), we find comparatively larger differences between products (cf. our Figure 3.2e and his Figure 4b). This could be due to the fact that we compare more reanalysis products (cf. Chaudhuri et al. 2013). Seeing as we consider a longer period extending more into the past, it could also reflect uncertainties earlier in the record or at lower frequencies. Averaging over the global ocean suggests a typical error value of ~ 5 mm.

Figure 3.2f shows signal-to-noise ratios, defined here as the values in Figure 3.2d divided by those from Figure 3.2e $[\text{RMV}/\sqrt{\psi_a}]$. Values are highest (~ 4) in the midlatitude North Atlantic, made manifest by a “tongue” emanating southwest from western Europe. Comparatively higher values here are not surprising, given the relatively abundant historical data in this region. Modest values (> 2) are seen in the northern North Atlantic (e.g., around Iceland), the midlatitude North Pacific, the subtropical South Pacific, and the tropical Indian Ocean. Lowest values (< 1) are seen along the Southern Ocean and the Beaufort Sea where noise is large (Figure 3.2d), and over the central equatorial Pacific Ocean and the west coast of Central American where signal is small (Figure 3.2e).

To give a more concrete sense of agreement between products, Figure 3.3 shows detrended time series at a few sites. Apparent are prominent interannual fluctuations and subtle decadal variations. Estimates show strong correspondence and significant correlations in the midlatitude North Pacific and tropical Indian Oceans (Figure 3.3a, b), consistent with the higher signal-to-noise in these places

(Figure 3.2f). In the South Pacific and Indian sectors of the Southern Ocean, the products can differ wildly from one another (Figure 3.3c, d), consistent with the lower signal-to-noise in these regions (Figure 3.2f). Discrepancies are also more acute earlier in time. In the Southern Ocean Indian sector, estimates agree fairly well from the late 1970s, however before that point time series exhibit strong divergence (Figure 3.3d), consistent with there being fewer data in this region earlier in the century.

3.5 Comparisons with station data

In this section, we compare the gridded reconstruction and reanalysis products to long records from meteorological stations to test for the presence of systematic errors in the gridded estimates. We find that errors can be correlated across gridded estimates and that some estimates perform better than others with respect to the data. In terms of trends, errors likely are correlated across the estimates, and HadSLP2 as likely as not performs better than either NOAA-20CRv2 or ERA-20C. In terms of fluctuations, it is extremely likely that errors are correlated across the gridded estimates, and likely that HadSLP2 performs better than either NOAA-20CRv2 or ERA-20C (Figures 3.4–3.6).

3.5.1 Preliminaries

In the foregoing, it is difficult to partition errors between the estimates. Moreover, since products use similar data streams, they might share common errors. A more thoroughgoing analysis would partition and separate errors in individual products (e.g., $\langle H'^2 \rangle$, $\langle N'^2 \rangle$, $\langle E'^2 \rangle$) from covariance terms (e.g., $\langle H'N' \rangle$, $\langle H'E' \rangle$, $\langle N'E' \rangle$). One way to approach this problem (e.g., Ponte and Dorandeu 2003) is to bring in a data source D with relatively small errors compared to the estimates,

such that, e.g.,

$$(T_D - T_H)(T_D - T_N) \approx T'_H T'_N, \quad (3.6)$$

and,

$$\langle (D - H)(D - N) \rangle \approx \langle H' N' \rangle. \quad (3.7)$$

Our strategy here is to use data from a small number of meteorological stations that have undergone intense individual scrutiny (section 2.b), diagnosing all the terms in (3.3) and (3.5) at station locations.

Figure 3.4a-d shows time series at some sites. Estimates are generally well correlated with the data (average correlation between estimates and the data is ~ 0.7). This is expected, since the estimates incorporate the data to varying degrees. But, upon closer examination, interesting discrepancies become apparent. At Tahiti, NOAA-20CRv2 and ERA-20C exhibit anomalously positive values relative to the data earlier in the record, while later in time, NOAA-20CRv2 shows anomalously negative values (Figure 3.4b). These results show that there are systematic errors in the estimates. Motivated by this cursory inspection, we now consider errors and their partitioning in more detail.

3.5.2 Centennial trends

Are error trends $[T'_H, T'_N, \text{ and } T'_E \text{ in (3.3)}]$ correlated or uncorrelated across the various products? For a detailed investigation of the error trends, we evaluate all terms in (3.3) at the sites (Figure 3.5). At four of the seven stations (Darwin, Chennai, Azores, and Iceland), HadSLP2 trends are closer to the data trends than are trends from either NOAA-20CRv2 or ERA-20C. At three of the locations (Nagasaki, Azores, and Tahiti), error trends from the three gridded estimates are all of the same sign. And at all locations the error trends from NOAA-20CRv2 and ERA-20C have the same sign.

To see if these results are consistent with random normal errors (about a

zero mean), we perform 100,000 trials of 21 draws ($3 \text{ products} \times 7 \text{ sites}$) from the standard normal distribution (not shown). We interpret each of these random draws as the error from a particular product at a particular site. The probability that one product shows the smallest errors at four of the seven sites is $P = 38\%$. Similarly, we determine that one product showing smallest errors at more than three locations has a probability $P = 52\%$. So, HadSLP2 showing smaller errors than both NOAA-20CRv2 and ERA-20C at more than half the stations is about as likely as not in the case of random errors. We also see that the probability that all three products show the same sign at three locations (more than two locations) is $P = 17\%$ ($P = 24\%$). Therefore, we reason that the three products showing error trends of same sign at the Nagasaki, Azores, and Tahiti stations is unlikely given errors normally distributed about a zero mean. Finally, the probability of two products having errors of the same sign at all seven sites is $P = 2.4\%$. Hence, we conclude that NOAA-20CRv2 and ERA-20C having error trends of the same sign at all stations is very unlikely given chance. Since these two estimates incorporate similar data (section 3.8), it is unsurprising that they show signs of common errors.

3.5.3 Detrended fluctuations

Are the error fluctuations [H' , N' , and E' in (3.5)] correlated or uncorrelated across the estimates? To answer this question, we diagnose all the terms in the error budget (3.5) at the stations (Figure 3.6). We contrast those evaluations, based on comparisons between gridded products and station data, to an analogous exercise using 100,000 trials of 21 randomly simulated time series (not shown). Each series has 101 entries (“1900–2000”) randomly drawn from the standard normal distribution. Each time series is also scaled by a random factor drawn from the standard lognormal distribution. We interpret each of these series as representing the error fluctuations from a product at a station.

Strikingly, error fluctuations are positively correlated across the estimates at all sites (Figure 3.6). Such an occurrence is not seen in any of the 100,000 simulations of randomly generated series, and is therefore exceptionally unlikely in the case of random errors. At five of the seven stations, HadSLP2 shows the smallest error variance and NOAA-20CRv2 the largest error variance with respect to the station data. In our experiments with random series, the probability that one product has the smallest or largest variance at five (more than four) locations is $P = 11\%$ ($P = 14\%$). Thus, HadSLP2 having the lowest error variance, or NOAA-20CRv2 having the highest error variance, is unlikely given chance. At six of seven sites, HadSLP2 and ERA-20C have the lowest covariance and ERA-20C and NOAA-20CRv2 have the highest covariance. For random series, two products having lowest or highest covariance at more than five sites has a probability of $P = 2.0\%$. Thus, HadSLP2 and ERA-20C showing lowest covariance, or ERA-20C and NOAA-20CRv2 showing the highest covariance, is very unlikely in the case of errors randomly distributed across products.

These findings demonstrate that error fluctuations covary across products. To visualize the point, error fluctuations (estimate minus data) are shown for four sites in Figure 3.4e-h. Consistent with Figure 3.6, error fluctuations from HadSLP2 are smaller than from NOAA-20CRv2 or ERA-20C at Darwin and Tahiti (Figure 3.4e, f), but are larger in the Azores (Figure 3.4g). Error fluctuations from the different estimates are comparable in Iceland (Figure 3.4h). Strong correlations between error fluctuations from NOAA-20CRv2 and ERA-20C are apparent, namely during the first quarter century at Darwin and Tahiti (Figure 3.4e, f). Errors are also larger earlier in the record (Figure 3.4e-f).

The correlated errors strongly impact uncertainties estimated by comparing products (Figure 3.2e). For example, while there is a significant correlation (0.97)

between ψ_a and ψ_b evaluated across the stations, the former underestimates the latter by a factor of ~ 2 (Figure 3.6). This suggests that increasing the values in Figure 3.2e by $\sim 40\%$ or so would paint a truer portrait of the average errors.

3.6 Impacts on tide gauge η records

In this section, we compare the gridded reconstruction and reanalysis η_{ib} products to η records at a few tide gauge sites to determine how much η_{ib} contributes to η during the Twentieth Century. We find that removing estimates of η_{ib} from η records reduces spatial variance in centennial trends across tide gauges by 10–30%, formal errors in centennial trends from individual gauge records by $\sim 5\%$, and temporal variance in detrended tide gauge records by 10–15% on average (Figures 3.7, 3.8).

3.6.1 Centennial trends

Centennial trends in η and $\eta - \eta_{ib}$ from gauge data and gridded products are shown in Figure 3.7a. On average, removing estimated η_{ib} from η data reduces the variance in the linear trends across tide gauges. Reduction in spatial variance of the trend field ranges from $\sim 10\%$ to 30% depending on choice of gridded product. Moreover, formal error bars on trends in $\eta - \eta_{ib}$ generally tend to be somewhat smaller than formal errors in η trends. While formal trend error reduction can be $> 10\%$ at some locations (e.g., Brest, San Francisco, Delfzijl, Newlyn), more generally it is $\sim 5\%$.

Given the limited number of records, these inferences could be very sensitive to this particular choice of tide gauges. To assess the sensitivity of results to tide gauge selection (“sampling error”), we perform 100,000 iterations of randomly choosing half of the records and recomputing values. In $> 95\%$ of cases, subtracting the ensemble mean η_{ib} trend (averaged across the three estimates) from each gauge

η record results in a reduction in the variance in the trends across the tide gauges. Similarly, $> 99\%$ of the time, removing η_{ib} from η reduces the average relative formal trend error. Thus, we conclude that results are robust and not sensitive to the particulars of tide gauge selection.

3.6.2 Detrended fluctuations

In Figure 3.7b, we show the percentage variance in each η record explained by the estimated η_{ib} , where we define the percentage variance V in a time series $x(t)$ explained by a time series $y(t)$ as,

$$V = 100\% \times \left[1 - \frac{\sigma^2(x - y)}{\sigma^2(x)} \right], \quad (3.8)$$

where σ^2 denotes variance. The variance explained in η depends strongly on tide gauge location. Generally speaking, η_{ib} tends to explain less η variance at lower-latitude sites (e.g., Fernandina, Honolulu, Key West) and more variance at higher-latitude locations (e.g., Brest, Trieste, Newlyn). On average, η_{ib} explains $\sim 10\text{--}15\%$ of the η variance, depending on the choice of gridded product.

According to (3.8), η and η_{ib} must agree in terms of amplitude and phase for η_{ib} to explain high η variance, while low η variance explained by η_{ib} result from differences in amplitude or phase. Figure 3.8a shows time series of η and η_{ib} at a few example sites to elucidate values in Figure 3.7b. In San Francisco (Figure 3.8a), η_{ib} is small compared to η , but the two time series are significantly correlated. At Delfzijl (Figure 3.8b), the η_{ib} fluctuations are not negligible relative to η variations, but the records can show differences in phase. As a result, η_{ib} explains nearly one-fifth of the η variance at these two locations. In Honolulu (Figure 3.8c), the η_{ib} changes are small compared to the η changes, and the two time series are generally not correlated. Consequently, η_{ib} effectively explains none of the η variance here. At Newlyn (Figure 3.8d), η and η_{ib} have similar amplitude

and phase, leading to almost half of the η variance being explained by η_{ib} changes at this location.

3.6.3 Wavelet coherence

There are interesting differences between our results and the findings of previous investigations. For example, Calafat et al. (2012) find that η_{ib} explains 22% of annual η variance at the Newlyn tide gauge over 1950–2009, whereas we determine that 42–45% of the annual Newlyn η variance over 1916–2000 is explained by η_{ib} (depending on choice of gridded product). Whereas Piecuch and Ponte (2015) show that η_{ib} accounts for $\sim 25\%$ of annual η variance during 1979–2013 on the North American northeast coast, we see that η_{ib} explains little variance ($< 10\%$) in annual η records along the east coast of the United States during 1900–2000. Such contrasts between our results here and the findings of previous studies hint that the relation between η and η_{ib} might not be stationary. Indeed, considering Newlyn (Figure 3.8d), for example, it appears that the correspondence between η and η_{ib} time series is stronger over 1950–1970 than during 1980–2000.

To shine brighter light on the relationship between η and η_{ib} as a function of time period and frequency band, we assess wavelet coherences (Grinsted et al. 2004) at some sites (Figure 3.8e-h). At some sites, η changes are generally not significantly coherent with η_{ib} changes. At Delfzijl (Figure 3.8f) and Honolulu (Figure 3.8g), significant coherence between η and η_{ib} is apparent only for fluctuations with periods of 4–8 years centered on the 1960s and 1970s. At other locations, η_{ib} changes are significantly coherent with η changes across a range of temporal epochs and frequency bands. At Newlyn (Figure 3.8h), strong coherence between η and η_{ib} is seen in terms of interannual variations over the middle of the century and decadal fluctuations more generally. Similarly, at San Francisco (Figure 3.8e), η and η_{ib} signals with periods ~ 16 years are significantly coherent

with one another over the entire study period, whereas changes with periods ~ 8 years are coherent more during the middle of the century. While wavelet coherence plots are similar for the two sites (cf. Figure 3.8e, h), η_{ib} explains more η variance at Newlyn than at San Francisco since η_{ib} amplitudes are more comparable to η magnitudes at the former than the latter place (Figure 3.8a, d).

3.7 Discussion

Tide gauges are one of the oldest observing systems of the ocean. For quantitative interpretation in terms of ocean dynamics or sea level, it is best to remove from tide gauge records the influence of isostatic volume redistribution related to surface pressure loading (the inverted barometer effect). We used gridded estimates and station data of surface air pressure (p_a) to investigate the inverted barometer effect (η_{ib}) and its impact on sea level (η) over the ocean during the Twentieth Century.

Centennial η_{ib} trends (Figure 3.2a) evidence a meridional structure consistent with the observed strengthening of westerly circumpolar flows in the Southern Hemisphere that have been tied to stratospheric ozone depletion (e.g., Thompson and Solomon 2002; Gillett and Thompson 2003). Detrended η_{ib} fluctuations (Figure 3.2d) show stronger variability at higher latitudes and weaker variability at lower latitudes, as in previous intraseasonal and interannual studies (Ponte 1993; Fu and Pihos 1994; Ponte 2006), and as expected for quasi-geostrophic motions, given the change in the Coriolis parameter with latitude. The differences between the gridded η_{ib} estimates are smaller for more recent periods and ocean regions with good historical data coverages, and are larger for earlier time periods or regions of the ocean with poor historical data coverage (Figures 3.2b, 3.2e, 3.3).

Comparisons against data records at meteorological stations show that errors can be correlated across gridded estimates and that some estimates perform better

than others with respect to data. In terms of centennial trends, experiments with simulated random time series suggest that error trends likely are correlated across the gridded estimates considered here and also the HadSLP2 reconstruction as likely as not performs better with respect to the data than the NOAA-20CRv2 and ERA-20C reanalyses (Figure 5). In terms of detrended fluctuations, it is extremely likely that error fluctuations are correlated across gridded estimates and likely that HadSLP2 performs better than NOAA-20CRv2 and ERA-20C (Figure 6). Errors in NOAA-20CRv2 and ERA-20C are more strongly correlated with one another than with HadSLP2 (Figure 3.4e-h, 3.5, 3.6), perhaps due to the fact that the former ingest such similar data streams, while the latter brings in many additional datasets not employed by the other products (section 3.8). These systematic errors can have a strong influence on uncertainties estimated through comparisons of gridded estimates. For example, in case of detrended fluctuations, variances in differences between gridded estimates underestimate the true error variances on those gridded estimates by roughly a factor of 2 on average (Figure 6).

Notwithstanding these errors, gridded estimates are still useful for interpretation of tide gauges. Removing η_{ib} estimates from η records reduces the spatial variance in centennial trends across tide gauges by 10–30%, the formal errors in centennial trends from individual tide gauges by $\sim 5\%$, and the temporal variance in the detrended tide gauge records by 10–15% on average (Figure 3.7). Therefore, while η_{ib} contributions to η changes are not dominant on time scales considered here, consistent with past studies based on individual tide gauge records or single air pressure estimates (e.g., Kolker and Hameed 2007; Miller and Douglas 2007), neither are they completely negligible. Interestingly, the influence of η on η_{ib} is a function not only of geographic location and frequency band, as noted in earlier

works (e.g., Ponte 2006), but also of time period (Figure 3.8). This finding elicits questions (beyond our scope) as to the reasons for the apparently non-stationary relationship between η and η_{ib} .

Our results give guidance on how (and whether) to adjust tide gauge η records for the η_{ib} effect. We advocate for making the η_{ib} correction to tide gauge records in ocean circulation and global η studies on long, multidecadal and centennial time scales. While they can be characterized by (sometimes systematic) errors, gridded η_{ib} estimates are of sufficiently good quality that adjusted tide gauge records typically have lower spatiotemporal variance than unadjusted tide gauge records. Although uncertainties in η_{ib} estimates can be large over some regions where historical p_a data are scarce, long tide gauge η records tend to be found in regions where historical p_a data are more plentiful, and so η_{ib} estimates will tend to be relatively more well constrained at tide gauge sites. As a general “rule of thumb” for making the adjustment, we recommend using an ensemble mean across several gridded estimates. This approach, in addition to being objective, reduces variance within and across tide gauge records slightly more on average than individual estimates (Figure 3.7).

3.8 Gridded p_a products

NOAA-20CRv2 covers the period 1850–2014 and is defined on a regular $2^\circ \times 2^\circ$ horizontal grid. Observations of p_a from the International Surface Pressure Databank [ISPD (Cram et al. 2015)] and the International Comprehensive Ocean-Atmosphere Data Set [ICOADS (Worley et al. 2005)] (and other sources) are assimilated with an ensemble Kalman filter. Boundary conditions are taken from Hadley Centre Sea Ice and Sea-Surface Temperature dataset [HadISST (Rayner et al. 2003)]. The fields were downloaded from NOAA (<http://www.esrl.noaa.gov/psd/data/gridded/>).

The ERA-20C estimate covers the years 1900–2011 with a nominal lateral resolution of 125 km. Using a four-dimensional variational (4D-Var) method, p_a data from ISPD and ICOADS, as well as surface winds from ICOADS are incorporated. Similar to NOAA-20CRv2, boundary conditions in ERA-20C also consist of sea ice and sea-surface temperatures taken from the HadISST dataset. The fields are downloaded from ECWMF (apps.ecmwf.int/datasets/data/era20c-mnth/).

The HadSLP2 reconstruction is defined on a $5^\circ \times 5^\circ$ lateral grid. It covers the period 1850–2004. Marine p_a data come from ICOADS while terrestrial p_a records come from various compilations, for example, the Global Historical Climatology Network [GHCN (Peterson and Vose 1997)]. This reconstruction is achieved based on reduced-space optimal interpolation (e.g., Kaplan et al. 2000). Fields were provided by the Met Office (<http://www.metoffice.gov.uk/hadobs/hadslp2/>). The Hadley Centre also makes available a near-real time product (HadSLP2r), which covers from 2005 to present. However, the Met Office notes that variances from the HadSLP2r dataset are not homogeneous with variances from the HadSLP2 dataset, and so the former is not considered here.

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Figure 3.1. Locations of tide gauges (green circles) and meteorological stations (blue squares) used here.

Figure 3.2. (a) Sample mean η_{ib} linear trends $\bar{T}$ during 1900–2000 over the ocean across estimates (mm yr^{-1}). Black lines are mean p_a contours (10-hPa increments). Black squares and circles mark meteorological stations and tide gauges, respectively (cf. Fig. 3.1). (b) Square roots of trend sample variances $\sqrt{\chi_a}$ (mm yr^{-1}). (c) Trend signal-to-noise ratios $|\bar{T}|/\sqrt{\chi_a}$ (Fig. 3.2a/Fig. 3.2b). (d) Root mean η_{ib} variance (RMV) over the ocean during 1900–2000 across products (mm). (e) Values of $\sqrt{\psi_a}$ (mm). (f) Signal-to-noise ratios $\text{RMV}/\sqrt{\psi_a}$ for detrended fluctuations (Fig. 3.2d/Fig. 3.2e). Four white stars mark the locations of the η_{ib} time series in Fig. 3.3.

Figure 3.3. Gridded η_{ib} estimates [H (blue), N (orange), and E (yellow)] at sites starred in Fig. 3.2f.

Figure 3.4. (Left) Time series of η_{ib} based on station data (black) and gridded estimates [H (blue), N (orange), and E (yellow)] at meteorological stations. (Right) Time series of η_{ib} errors (i.e., station data minus gridded estimate) at those same meteorological stations. For this figure, we set $\bar{p}_a = 0$ in Eq. (3.1) in computing all series for consistency, as there are no observations of p_a averaged over the ocean with which to adjust the station data.

Figure 3.5. Trend error budget [Eq. (3.3)] at meteorological stations. All values are in units of $(\text{mm yr}^{-1})^2 \times 10^4$. For this figure, we set $\bar{p}_a = 0$ in Eq. (3.1) in computing all time series (and related quantities) for consistency, because there are no observations of p_a averaged over the global ocean with which to adjust station data values.

Figure 3.6. Fluctuation error budget [Eq. (3.5)] at various meteorological stations. All values are in units of mm^2 . For this figure, we set $\bar{p}_a = 0$ in Eq. (3.1) in computing all time series (and related quantities) for consistency, because there are no observations of p_a averaged over the global ocean with which to adjust station data values.

Figure 3.7. **(a)** Dots are centennial trends (mm yr^{-1}) in tide gauge η (black) and $\eta - \eta_{ib}$ from tide gauges and gridded estimates [H (blue), N (orange), E (yellow), and mean of estimates (red)] at the tide gauge sites. Horizontal bars are $\pm$ one standard error of the fit accounting for autocorrelation of residuals. Colored σ^2 values on the left are the variances in the η (black) or $\eta - \eta_{ib}$ (various colors) trends across sites [units $(\text{mm yr}^{-1})^2$]. **(b)** Squares are percentages variance explained in tide gauge η by gridded estimate η_{ib} across tide gauge sites. Colored μ values along the right are the average variances explained in tide gauge η by the gridded η_{ib} estimates.

Figure 3.8. (Left) Time series of tide gauge η (black) and gridded η_{ib} estimates (colors) at different tide gauges. (Right) Wavelet coherences between η and η_{ib} (ensemble average across gridded products) at the gauge sites. Thin solid white contouring indicates statistically significant values (5% level against red noise). Thick dashed white contours denote a “cone of influence” (Grinsted et al. 2004) outside which edge effects may impact results.

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