SEA-LEVEL CHANGE AND SUBSIDENCE IN THE DELAWARE ESTUARY DURING THE LAST ~2200 YEARS

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ABSTRACT

We produced eight new sea-level index points that reconstruct a ~2.5 m relative sea-level (RSL) rise at Sea Breeze in the Delaware Bay from ~200 BCE to 1800 CE. The precision of our reconstruction improved upon existing data by using high-resolution surveying methods, AMS radiocarbon dating of *in-situ* plant macrofossils collected immediately above the basal contact between pre-Holocene sand and salt-marsh sediments, foraminifera as sea-level indicators, and by accounting for tidal range changes through time. Our new data were combined with a database of 65 sea-level index points available for the Delaware Bay to estimate the rate of RSL rise in the upper (1.26 ± 0.33 mm/yr) and lower bay (1.30 ± 0.36 mm/yr) using a spatial-temporal model. Correction for changes in tidal range through time removed the disparity in rate between the upper and lower Delaware Bay that had previously been postulated. After paleotidal correction, the rates of RSL rise estimated for the Delaware Bay (1.25 ± 0.27 mm/yr) correlate with the ~1.3 mm/yr rate reported for New Jersey, Maryland, and Virginia, and confirm that the maximal ongoing forebulge collapse along the U.S. Atlantic coast is focused on the mid-Atlantic.

1. INTRODUCTION

Proxy reconstructions are important for understanding the driving mechanisms of past relative sea-level (RSL) trends and for constraining predictions of future sea-level change (e.g., Dutton et al., 2015). On the U.S. Atlantic coast the principal cause of prolonged, regional RSL change during the Common Era (last ~2000 years) was glacio-isostatic adjustment (GIA), driven by collapse of the Laurentide Ice Sheet's proglacial forebulge (e.g., Peltier, 1996). This process causes a change in the geoid height and the local radius of the solid Earth relative to Earth's center of mass (Farrell and Clark, 1976). Despite disintegration of the Laurentide ice sheet by ~7000 years before present (BP; e.g., Carlson et al., 2008), GIA on the U.S. Atlantic coast continues to the present day (e.g., Engelhart et al., 2009), because of the slow response time (~4000 years) of the solid Earth to redistribution of mass during deglaciation (Peltier, 1998). RSL also includes contributions from other processes causing vertical land motion (Kopp et al., 2015) such as sediment compaction (e.g., Miller et al., 2013), dynamic topography (e.g., Rowley et al., 2013), and tectonics (e.g., van de Plassche et al., 2014). For convenience, we use the term "land-level change" to describe the net effect of GIA and these other sources of vertical land motion. Accurate estimates of land-level change are important for generating regional sea-level projections (e.g., Kopp et al., In Press), isolating the climate-driven components of RSL trends measured by tide gauges and satellite altimetry (e.g., Church and White, 2011; Nerem et al., 2010), and testing Earth-ice models (e.g., Roy and Peltier, 2015).

In the absence of long-term instrumental measurements, rates of land-level change can be estimated from RSL reconstructions spanning the last 1000 to 4000 years (Engelhart et al., 2009; Shennan and Horton, 2002; Engelhart et al., 2015). This approach assumes that the non-land-level change component of reconstructed RSL over this period was zero or minimal (Bassett et al., 2005; Lambeck et al., 2014; Milne et al., 2005; Peltier, 2002). A standardized database of RSL reconstructions from the U.S. Atlantic Coast demonstrated that spatially-variable rates of land-level change (and RSL rise) during the last 4000 years reflect distance from the former center of the Laurentide Ice Sheet (Engelhart and Horton, 2012; Engelhart et al., 2009). After grouping RSL reconstructions by location, Engelhart et al. (2009) estimated the rate of subsidence for 16 regions from Maine to South Carolina. Maximum subsidence (1.7 ± 0.2 mm/yr) occurred in the upper Delaware Bay. However, the vertical and temporal uncertainties on individual RSL reconstructions from the Delaware Bay were relatively large and were not formally accounted for when estimating rates of RSL change based on linear regression of reconstruction mid points. Furthermore, such estimates did not include the influence of tidal-range change, a process shown to be significant in the Delaware Bay during the Holocene (Belknap, 1975; Belknap and Kraft, 1977; Hall et al., 2013; Leorri et al., 2011). A similar pattern of subsidence was identified by isolating the

regionally-coherent linear component of RSL at tide-gauge stations where rates of 1.8 ± 0.5 mm/yr and 1.5 ± 0.4 mm/yr were estimated for Cape May, NJ and Philadelphia, PA respectively (the closest permanent gauges to our site at Sea Breeze: Figure 1), but these analyses were limited to the short period of available data (Kopp, 2013).

RSL reconstructions comprise discrete sea-level index points that estimate RSL at a particular time and place. The RSL history of a site or region is described by combining sea-level index points that were generated using a standardized approach where the age of each index point was estimated with uncertainty (usually by radiocarbon dating), its geographic location is known, and the vertical position of former sea level was estimated using a proxy called a sea-level indicator. The indicative meaning (van de Plassche, 1986; Shennan, 1986; Horton et al., 2000; Woodroffe and Barlow, 2015) formalizes the relation between an indicator and sea level by establishing the elevation range (termed the indicative range) across which particular indicators are found. The mid-point of this range is called the reference water level.

To redress a scarcity of Common Era RSL reconstructions from the New Jersey side of the Delaware Bay we produced eight new sea-level index points from Sea Breeze, New Jersey using foraminifera preserved in radiocarbon-dated, basal sediment. We combined the new reconstruction with an existing database of standardized sea-level index points and corrected for the effect of tidal-range change. Analysis of the resulting regional-scale RSL dataset using a spatio-temporal models showed that the rate of subsidence in the Upper Delaware Bay was previously overestimated. Subsidence in the Delaware Bay is comparable to rates estimated on the Atlantic coasts of New Jersey, Delaware, Maryland, and Virginia.

2. STUDY REGION

Sea Breeze, NJ is located on the northeastern shore of the Delaware Bay (Figure 1). The modern salt marsh at Sea Breeze displays the characteristic regional pattern of floral zonation. Low salt-marsh areas between mean tidal level (MTL) and mean high water (MHW) on the shore of Delaware Bay and along tidal channels are vegetated by *Spartina alterniflora* (tall form). The high marsh is the largest zone by area and lies between MHW and mean higher high water (MHHW). It is vegetated by *Spartina patens*, *Distichlis spicata*, and the short form of *Spartina alterniflora*. The border between salt marsh and freshwater upland at elevations from MHHW to highest astronomical tide (HAT) is a brackish zone vegetated by *Schoenoplectus* spp. and *Phragmites australis*. Great diurnal tidal range (mean lower low water, MLLW to MHHW) measured at the NOAA tide gauge nearest to Sea Breeze (Ship John Shoal;

Figure 1) is 1.79 m. Marine influence reached the region at ~6000 years BP when tidal wetlands began to establish around the Delaware Bay (e.g., Leorri et al., 2006; Nikitina et al., 2003). Continuous RSL rise since then resulted in the Sea Breeze salt marsh being underlain by 2-4 m of salt-marsh and intertidal sediments (Nikitina et al., 2014).

Modern salt-marsh foraminifera in southern New Jersey form at least seven distinctive assemblages (Kemp et al., 2012; Kemp et al., 2013c). The dominant species of foraminifera in low-marsh floral zones are *Miliammina fusca* and *Ammobaculites* spp. High-marsh floral zones are populated by assemblages of foraminifera in which *Trochammina inflata*, *Arenoparrella mexicana*, and *Tiphotrecha comprimata* are the most abundant species. At brackish sites with strong fluvial influence, *Ammonia* is the characteristic high-marsh species. The transition between salt-marsh and fresh-water ecosystems is dominated by *Haplophragmoides manilaensis*.

3. MATERIALS AND METHODS

3.1 CORE SELECTION, RECOVERY, AND ELEVATION

A detailed lithostratigraphic investigation of the Sea Breeze site was completed from 59 exploratory gouge cores. The site is unsuitable for continuous reconstructions of Common Era RSL change because the stratigraphy beneath the modern salt marsh reveals peats that are intercalated by inorganic units that likely resulted in sediment compaction. Furthermore, the site was subjected to repeated erosion during hurricanes or large storms resulting in a fragmentary and incomplete sedimentary record (Nikitina et al., 2014). Therefore, we selected eight cores that contained basal peats to generate eight new sea-level index points between 3.21 m and 0.59 m below North American Vertical Datum 1988 (NAVD88). In each case, a 0.5 m depth interval was sampled to capture the transition from incompressible, sandy, pre-Holocene substrate to overlying organic sediment. Samples were collected using a Russian-type hand auger to minimize compaction and/or contamination of sediment during recovery. Cores were transferred to labeled PVC pipes, sealed in plastic wrap, and held in refrigerated storage until processing.

Core top elevations were referenced to NAVD88 by leveling to nearby benchmarks with a total station. Elevations were converted to tidal elevations using VDatum (v2.3.5) and the New Jersey Coastal Embayment regional dataset. At Sea Breeze, NAVD88 is 0.06 m above MTL.

3.2 RADIOCARBON DATING

The cores recovered from Sea Breeze were processed to isolate plant macrofossils in growth position from the sediment matrix. Identification of the macrofossils was made by comparison with modern equivalents and published descriptions (Niering et al., 1977). To minimize the contribution of sediment compaction to RSL reconstruction, samples for radiocarbon dating were selected from as close to the basal contact as possible. Samples were first cleaned using distilled water under a binocular microscope to remove contaminating material including adhered organic sediment and in-growing rootlets and then oven dried at <45 °C. The dry samples were weighed and submitted to the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) radiocarbon dating laboratory. All samples underwent acid-base-acid pretreatment to remove carbonate and humic acids. A portion of CO₂ collected during sample combustion provided $\delta^{13}\text{C}$ values and was used to correct the sample for natural fractionation during its formation. Individual radiocarbon dates were calibrated using the IntCal13 (Reimer et al., 2013) dataset and are presented with 2 σ calibrated uncertainties where zero is 1950 CE (Table 1).

3.3 FORAMINIFERA AND THE INDICATIVE MEANING

Foraminifera were enumerated from 1-cm thick slices of core material. These samples were sieved under running water to disaggregate the sediment and retain the fraction in the 63-500 μm size range. A minimum of 100 individuals were counted wet from each sample under a binocular microscope, which is sufficient to accurately describe low-diversity assemblages that are typical of salt-marsh sediment (e.g. Horton and Edwards, 2006). Species identifications were confirmed by comparison with type slides of individuals collected from modern salt marshes in the study area (Kemp et al., 2012; Kemp et al., 2013c).

We estimated the indicative meaning of dated samples at Sea Breeze (Table 2) and in the regional Delaware Bay database (Table 3) based on sediment character, preserved plant macrofossils, and foraminifera. We classified samples as having been deposited in either a high salt-marsh, low salt-marsh or undifferentiated salt-marsh environment. Following the definitions of Engelhart and Horton (2012), we applied an indicative range of MTL to MHW and MHW to HAT for low- and high-marsh samples, respectively. For undifferentiated salt-marsh samples an indicative range of MTL to HAT was applied.

3.4 PALEOTIDAL MODELING

Tidal data for Delaware Bay were predicted using the approach from Hall et al. (2013). To briefly summarize their methods, a high-resolution finite-element mesh, with horizontal spacing of 80-175 m,

was created for the Delaware Bay. This high-resolution mesh was combined with the coarser Western North Atlantic grid (Mukai et al., 2002) to create an overall model grid with regional scope, but high resolution locally in the study area. The baseline grid, created with present-day bathymetric data sets, was combined with RSL predictions from the ICE-6G VM5b model (Engelhart et al., 2011b) in order to produce paleobathymetries from 0 to 7000 years BP at 1000 year intervals. Leorri et al. (2011) and Hall et al. (2013) both noted that the bathymetric change from the late Holocene to present day exhibited considerable spatial variability across the Delaware Bay. Offshore open boundary forcing was obtained from the TPXO 7.2 model (Egbert et al., 1994). This model provides the amplitudes and phases of the eight principal tidal constituents. For the paleo-tidal simulations, the present-day open boundary forcing was used. This was justified since previous work by Hill et al. (2011) showed that tides in the northern Atlantic did not change appreciably from 0 to 7000 years BP.

The ADCIRC model (Luetich and Westerink, 1991) was used to simulate the tides in the domain. Wetting and drying was enabled in order to properly model the intertidal zone and finite amplitude effects, and nonlinear bottom friction were included in order to capture the generation of overtides in shallow water. The tidal model did not include models of sediment infilling of either the estuaries or coastal lowlands or freshwater discharge from the catchments (Shennan et al., 2000). For each time slice of interest, a 90-day simulation was carried out. Harmonic analysis of the results allowed for the determination of the tidal amplitudes and phases in the interior of the domain. These results were converted to tidal datums, such as MHHW and mean low water (MLW), using the method of Mofjeld et al. (2004). HAT at each model node was computed as the sum of the tidal amplitudes at that node. The paleotidal model was used to modify reference water levels and indicative ranges for all index points (Sea Breeze and the Delaware Bay database). Following Horton et al. (2013), data at locations other than the model grid points (e.g., a field sampling site) were obtained through interpolation. We compared the model predictions of tidal datums with the three tide gauges in Delaware Bay. The root mean squared error of the misfit between predicted and observed datums is ± 0.14 m (i.e. the tidal model error to reconstruct current tides, after Shennan et al., 2000).

3.5 RECONSTRUCTING RELATIVE SEA LEVEL

RSL was reconstructed by subtracting the estimated indicative meaning from measured sample altitude (Shennan and Horton, 2002):

$$RSL_i = A_i - RWL_i \quad [1]$$

where A_i is the altitude of sample i established by measuring depth in a core with a known surface elevation. RWL_i is the reference water level assigned to sample i . A_i and RWL_i are expressed relative to the same tidal datum (e.g., MTL) so that for a modern (surface) sample the two terms are equal and RSL is 0 m. The vertical position of RSL for each index point is expressed in relation to this value where negative values indicate RSL below present.

The total vertical error (TE_i) for each sample was calculated from the expression (Shennan and Horton, 2002):

$$TE_i = (IR^2 + e_1^2 + e_n^2)^{1/2} \quad [2]$$

where IR is the indicative range and $e_1 \dots e_n$ are the individual sources of sampling error for sample i . For the Sea Breeze index points this includes: (1) benchmark error of ± 0.10 m; (2) leveling error of ± 0.05 m; (3) sample thickness error of ± 0.01 m; (4) core angle error of $\pm 1\%$ of depth; (5) tidal modeling error of ± 0.14 m; and (6) Russian core sampling error of ± 0.01 m.

3.6. STATISTICAL FRAMEWORK TO ASSESS SPATIAL VARIABILITY OF RSL

We assessed the spatial variability of RSL in Delaware Bay using a spatio-temporal empirical hierarchical model (Cressie and Wikle, 2011) applied to the Sea Breeze reconstruction and database of sea-level index points ($n = 65$). Spatio-temporal hierarchical models have been previously used to analyze paleo-sea level data (e.g., Kopp et al., In Press), as well as spatio-temporal fields ranging from surface winds (e.g., Wikle et al., 2001) to species populations (e.g., Wikle, 2003) to paleo-temperature (e.g., Tingey and Huybers, 2010). By explicitly modeling the spatio-temporal covariance structure, they enable the testing of hypotheses about site-to-site similarities and differences, which cannot be done self-consistently by comparing analyses of individual sites (see discussion in Kopp, 2013).

RSL at each space point was modeled as the sum of basin-wide sea-level change and a more localized linear trend:

$$f(\mathbf{x}, t) = l(\mathbf{x})(t - t_0) + s(t) \quad [3]$$

where t_0 is a reference time period (here, 1800 CE). The spatial covariance of the linear trend and the temporal covariance of the basinal sea-level signal were both modeled with mean-zero Gaussian process priors (Rasmussen and Williams, 2005):

$$l(x) \sim GP\{0, \sigma_l^2 \gamma(x, x'; \lambda)\} \quad [4]$$

$$s(t) \sim GP\{0, \sigma_m^2 \rho(t, t'; \tau)\} \quad [5]$$

where $\gamma(x, x'; \lambda)$ is an exponential covariance function and $\rho(t, t'; \tau)$ a Matérn covariance function with smoothness parameter 3/2. The observed sea-level indicators were modeled as the sum of the true sea level and a white noise term with variance based on the total error estimated using equation (4). For samples that were not basal peats, an additional error equal to an optimized hyperparameter κ times sample depth was added to account for compaction-related uncertainty. Geochronological uncertainty was incorporated through the noisy-input Gaussian process approximation of McHutchon and Rasmussen (2011):

$$y_i = f(\mathbf{x}_i, \hat{t}_i + \delta_i) + \varepsilon_i \approx f(\mathbf{x}_i, \hat{t}_i) + f'(\mathbf{x}_i, \hat{t}_i)\delta_i + \varepsilon_i \quad [6]$$

Here, $\hat{t}_i$ is the mean estimate of the calibrated age of indicator i , δ_i is the error in this estimate (approximated as normally distributed), and ε_i is the vertical error. We employ simulated annealing followed by local optimization to find the hyperparameters that maximize the likelihood of the model, conditional upon all the index points in the data set: $\sigma_1 = 2.5$ mm/y, $\lambda = 4.5^\circ$, $\sigma_m = 620$ mm, $\tau = 760$ y, and $\kappa = 0.013$.

4. RESULTS

4.1 LITHO, BIO AND CHRONOSTRATIGRAPHY OF SEA BREEZE SEA-LEVEL INDEX POINTS

The Sea Breeze salt marsh is underlain by fluvial and tidal sediments consisting of five lithologic units (Figure 2). Gray, fine-to-medium compacted fluvial sand was deposited in the area prior to the Holocene transgression. The lowest stratigraphic unit above the pre-Holocene surface is an amorphous, black, humic, sandy mud that we interpreted as a paleosol. A gray-brown to dark-brown fibrous peat with fragments of *Schoenoplectus* sp., *Spartina patens*, and *Distichlis spicata* macrofossils overlies the paleosol. The macrofossils present within the peat indicate that it accumulated in brackish and high salt-marsh environments. A gray mud described at different stratigraphic levels above high-marsh peat represents tidal-flat deposits. A gray-brown, muddy peat with abundant *Spartina alterniflora* fragments was likely deposited in a low-marsh environment.

Core SB30 had a surface elevation of 1.04 m MTL (Figure 3A). A fragment of a *Schoenoplectus* sp. plant recovered at 3.40 m depth in amorphous black sandy organic-rich sediment yielded an age of 1760 ± 25 radiocarbon years BP (Table 1). The calibrated age range for the date was 1736 to 1569 years BP.

Dominant foraminifera between 3.36 m and 3.44 m were *Arenoparrella mexicana* with *Jadammina*

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4 *macrescens* and *Tiphotrocha comprimata*. Foraminifera and plant macrofossils indicated that the sample
5 formed in a high salt-marsh floral environment and was assigned a reference water level of the midpoint
6 between MHW and HAT and an indicative range of MHW to HAT. The resulting sea-level index point
7 was produced directly at the contact with the incompressible sand substrate.
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11 Core SB32 had a surface elevation of 1.07 m MTL (Figure 3B). An *in-situ* *Distichlis spicata* rhizome was
12 recovered at 4.21 m depth in a unit of dark, compacted, brown peat with woody fragments and the
13 remains of *Phragmites australis* plants. It yielded an age of 2210 ± 30 radiocarbon years BP (Table 1)
14 that calibrated to 2325-2149 years BP. Dominant foraminifera in the interval between 4.18 m and 4.21 m
15 were *Jadammina macrescens* and *Tiphotrocha comprimata*. Samples at 4.22 m and 4.23 m included a
16 small number of *Jadammina macrescens*. Foraminifera and plant macrofossils indicated that the sample
17 formed in a high salt-marsh floral environment and was assigned a reference water level of the midpoint
18 between MHW and HAT and an indicative range of MHW to HAT. The resulting sea-level index point
19 was 0.01 m above the basal contact.
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23 Core SB76 had a surface elevation of 0.86 m MTL (Figure 3C). An *in-situ* vertical stem and attached
24 rhizome of *Schoenoplectus sp.* was recovered at 1.57-1.63 m depth in high salt-marsh peat and yielded an
25 age of 960 ± 25 radiocarbon years BP (Table 1). The calibrated age range for the date was 929-795 years
26 BP. The assemblage of foraminifera at the dated level included *Jadammina macrescens* and *Tiphotrocha*
27 *comprimata*. Three deeper samples at 1.57 m, 1.58 m and 1.59 m included a small number of *Tiphotrocha*
28 *comprimata*. Foraminifera and plant macrofossils indicated that the sample formed in a high salt-marsh
29 floral environment and was assigned a reference water level of the midpoint between MHW and HAT and
30 an indicative range of MHW to HAT. The resulting sea-level index point was at the basal contact as the
31 rhizome was growing into the sandy substrate.
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35 Core SB81 had a surface elevation of 0.74 m MTL (Figure 3D). A fragment of a *Phragmites australis*
36 stem recovered at 3.20 m depth in amorphous black sediment yielded an age of 1840 ± 30 radiocarbon
37 years BP (Table 1). The associated calibrated age range was 1865-1708 years BP. Dominant foraminifera
38 between 3.16 m and 3.22 m were *Arenoparrella mexicana* with *Ammonoastuta inepta*. A sample at 3.24 m
39 included just three individual foraminifera (two *Trochammina inflata* and one *Jadammina macrescens*).
40 Foraminifera and plant macrofossils indicated that the sample formed in a high salt-marsh floral
41 environment and was assigned a reference water level of the midpoint between MHW and HAT and an
42 indicative range of MHW to HAT. The resulting sea-level index point was 0.1 m above the basal contact.
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46 Core SB83 had a surface elevation of 0.84 m MTL (Figure 3E). An *in-situ* *Spartina patens* rhizome
47 recovered at 2.77 m depth in high salt-marsh peat sediment yielded an age of 1740 ± 30 radiocarbon years
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BP (Table 1). The calibrated age range for the sample was 1715-1562 years BP. Characteristic foraminifera between 2.73 m and 2.77 m were *Ammoastuta inepta*, *Arenoparrella mexicana* and *Tiphotrecha comprimata*. A sample at 2.78 m was dominated by *Jadammina macrescens* with foraminifera absent from deeper samples. Foraminifera and plant macrofossils indicated that the sample formed in a high salt-marsh floral environment and was assigned a reference water level of the midpoint between MHW and HAT and an indicative range of MHW to HAT. The resulting sea-level index point was 0.13 m above the basal contact.

Core SB84 had a surface elevation of 0.72 m MTL (Figure 3F). An *in-situ* macrofossil of *Distichlis spicata* rhizome recovered at 2.22 m depth in amorphous black sediment yielded a radiocarbon age of 1260 ± 25 years BP (Table 1). The calibrated age range for the sample was 1280-1091 years BP. The transition to high salt-marsh peat sediment was documented in the core at the depth of 2.21 m. Dominant foraminifera between 2.21 m and 2.24 m were *Jadammina macrescens* and *Tiphotrecha comprimata*. Foraminifera and plant macrofossils indicated that the sample formed in a high salt-marsh floral environment and was assigned a reference water level of the midpoint between MHW and HAT and an indicative range of MHW to HAT. The resulting sea-level index point was 0.15 m above the contact with sand.

Core SB98 had a surface elevation of 0.78 m (Figure 3G). An *in-situ* macrofossil of *Spartina patens* recovered at 2.03 m depth in amorphous black sediment yielded an age of 1010 ± 25 radiocarbon years BP (Table 1). The calibrated age of the sample was 969-905 years BP. The stratigraphy changed to mud at a depth of 2.00 m. A sample at 1.99 m was dominated by *Tiphotrecha comprimata*, while samples between 2.01 m and 2.07 m were characterized by high abundances of *Jadammina macrescens*. Foraminifera and plant macrofossils indicated that the sample formed in a high salt-marsh floral environment and was assigned a reference water level of the midpoint between MHW and HAT and an indicative range of MHW to HAT. The resulting sea-level index point was 0.07 m above the basal contact.

Core SB110 had a surface elevation of 0.83 m MTL (Figure 3H). A rhizome of *Spartina patens* rooted in an underlying palesol sedimentary unit at 1.35 m depth yielded an age of 830 ± 25 radiocarbon years BP (Table 1). Its calibrated age was 784-690 years BP. A sample at 1.31 m was dominated by *Jadammina macrescens* (93%), but samples at 1.32 m, 1.33 m and 1.35 m had few foraminifera. Foraminifera and plant macrofossils indicated that the sample formed in a high salt-marsh floral environment and was assigned a reference water level of the midpoint between MHW and HAT and an indicative range of MHW to HAT. The resulting sea-level index point was 0.15 m above the basal contact.

4.2. PALEOTIDES IN DELAWARE BAY

While regional tidal model results showed minimal changes to deep-ocean Great Diurnal Range (GT) during the Common Era (Hill et al., 2011), results in the Delaware Bay showed significant spatial and temporal variability. The variability within the Delaware Bay can be seen in Figure 4, which shows contours of GT at 2000 and 1000 years BP, along with present day. Present day GT is largest (~1.8 m) in the Delaware River and upper reaches of the Delaware Bay (Figure 4A). GT at the river mouth in lower reaches of the bay is ~1.4 m and decreases further with distance beyond the bay and onto the continental shelf. At 2000 years BP, GT in the upper Delaware Bay was <1 m, while in the lower reaches of the bay GT was similar to present, indicating a sharp spatial gradient in the temporal rate of change in GT (Figure 4C). The paleogeographic evolution of Delaware Bay from a narrow river at ~ 4000 years BP to the familiar funnel shaped bay of today (Hall et al., 2013) is the most likely explanation for the large changes in upper bay GT, compared to the lower bay. Changes in tidal range do not affect all types of sea-level index points (high marsh, low marsh, undifferentiated marsh) evenly. The greatest changes in reference water level and indicative ranges are for high-marsh deposits that are located in the upper Bay. The reference water level of the four index points from Sea Breeze dated between 1500 and 2200 years BP decreased from 0.98 m to 0.60 m. The indicative range was reduced from 0.20 m to 0.15 m.

5. DISCUSSION

5.1 RSL CHANGE AT SEA BREEZE AND ASSESSMENT OF ERRORS

The eight new sea-level index points from Sea Breeze reconstruct RSL change on the New Jersey coast of the Delaware Bay. After correction for tidal-range changes, the new reconstruction shows that RSL rose from approximately -3.8 m at 2200 years BP to -1.4 m at 700 years BP (Figure 5A). The average rate of RSL rise from 2200 to 150 years BP, estimated using the spatio-temporal empirical hierarchical model, was 1.12 ± 0.22 mm/yr (2σ uncertainty).

Comparison of the Sea Breeze RSL record with published Delaware Bay sea-level index points shows that application of modern techniques results in improved precision compared to older reconstructions (Figure 5B) that were generated by studies published as early as 1972 (index point #29, Table 2). It is important to note that there is no systematic difference in the RSL reconstructed by older and more recent studies, indicating that all index points in the regional database accurately reconstruct former RSL

irrespective of where the data were produced (e.g., Engelhart and Horton, 2012; Horton et al., 2013; Hijma et al., 2015). RSL reconstructions require that sample altitude (part of term A_i in equation 1) is known. In some of the studies from which sea level index points were compiled and standardized by Engelhart and Horton (2012), the core top elevation was either not measured or not explicitly reported and could only be estimated from a description of the surface coring environment such as “salt marsh” or “high salt marsh”. In part this was because the original studies were not necessarily seeking to reconstruct RSL. Since these environments span a range of elevations, some sample altitudes in the database have an uncertainty of up to ± 0.9 m (e.g., index point #28). Improved surveying instruments resulted in smaller sampling uncertainties, this is particularly true where real time kinematic satellite navigations and digital total stations enabled core tops to be leveled accurately and precisely to a network of survey markers since these are rarely located on or close to salt marshes. The indicative range of the samples was not reduced by using foraminifera at Sea Breeze, because we used the foraminiferal assemblages to support the clear vertical division of plants into high and low salt-marsh floral zones that reflect the varied tolerances of plant species to tidal flooding (Gehrels, 1994; Redfield, 1972; van de Plassche, 1991). This methodology allows direct comparisons between sea-level reconstructions that were produced from different sea-level indicators.

Advancements in radiocarbon dating significantly improved the vertical and temporal precision of RSL reconstructions derived from salt-marsh sediment. Many of the radiocarbon dates reported in older studies were produced by the so-called conventional or radiometric method (measurement of β emissions by liquid scintillation or gas proportional counting). In contrast, newer reconstructions (including the one from Sea Breeze) used Accelerator Mass Spectrometry (AMS) to radiocarbon date individual plant remains by measuring the relative number of ^{14}C and ^{13}C atoms. AMS dating does not necessarily yield a smaller analytical uncertainty than the conventional method (Cook and van der Plicht, 2007). However, the shift to AMS dating has two important implications for the uncertainties of RSL reconstructions. Firstly, sampling error is reduced, because it allows small and well-defined samples such as individual plant macrofossils to be accurately dated. In contrast, conventional radiocarbon dating of salt-marsh material usually required a much larger sample and undifferentiated bulk sediment was commonly used. Short-lived plant macrofossils contain carbon that accumulated during just a few years. Although bulk salt-marsh sediment commonly consists of *in situ* material, it can also include transported macrofossils and unidentifiable carbon in the surrounding sediment matrix from varied sources and of varied ages (Dicken et al., 2004; Grimm et al., 2009; Hu, 2010; Kemp et al., 2013b; Törnqvist et al., 1990; 2015; van de Plassche, 1982). This difference is reflected in a small sample uncertainty for individual plant macrofossils and a large sample uncertainty for bulk sediment, irrespective of the analytical method used

to measure radiocarbon (Tornqvist et al., 2015). For example index point #58 used AMS to radiocarbon date a single plant macrofossil that yielded an uncertainty of ± 25 radiocarbon years. In contrast, index point #16 was based on a bulk-sediment radiocarbon date that resulted in an uncertainty of ± 245 radiocarbon years. In high salt-marsh ecosystems along the U.S. Atlantic coast the primary source of organic material is deposition of in situ plant material (Chmura and Aharon, 1995; Kemp et al., 2012b) and it is therefore unlikely that conventional radiocarbon dating will yield inaccurate results. This is demonstrated by agreement among the age of sea-level index points from the Delaware Bay using both conventional radiocarbon dating of bulk sediment and AMS dating of plant macrofossils. From a comparison of ^{14}C ages of macrofossils and the peat in which they were located (Hu, 2010), Tornqvist et al. (2015) suggested an additional error of ± 100 ^{14}C yr should be applied to bulk peat samples prior to calibration. Our results here suggest, at least for older conventional bulk radiocarbon dates on thick sections of salt-marsh peat, that such corrections are not required due to the agreement between these dates and our new high-precision index points.

Secondly, radiocarbon dating of bulk sediment collected using narrow cores required a thick sample, while AMS commonly utilizes a plant macrofossil that was situated at a single, specific depth in the core. This affects the vertical uncertainty of a sea-level index point because sample thickness is part of the total error calculation (equation 2). For example the bulk-sediment dated for index point #1 resulted in a sample thickness error of ± 0.15 m compared to ± 0.01 m for index point #58 that used a discrete plant macrofossil. Radiocarbon dating of single plant macrofossils (usually by AMS) in place of radiocarbon dating of bulk sediment (usually by the conventional method) reduces the vertical uncertainty of RSL reconstructions because sample thickness is reduced from a section of core to a discrete depth.

5.2 RATE OF RSL AND SUBSIDENCE IN DELAWARE BAY

During the last ~2200 years, eustatic sea-level change was zero (e.g., Bassett et al., 2005; Milne et al., 2005; Peltier, 2002) or minimal (0.1-0.2 mm/yr; Lambeck, 2002; Lambeck et al., 2014), making GIA the predominant control on RSL change in eastern North America (e.g., Davis and Mitrovica, 1996). For this reason, we assume that linear Common Era RSL trends approximate rates of land-level change, which we defined as the net effect of GIA, tectonics, and sediment compaction.

Engelhart et al. (2009) suggested that the rate of RSL rise during the late Holocene in the upper Delaware Bay (1.7 ± 0.2 mm/yr) was the fastest rate along the U.S. Atlantic coast and greater than in the lower

Delaware Bay (1.2 ± 0.2 mm/yr). Similarly Leorri et al. (2006, 2011) estimated rates of 1.7 ± 0.2 mm/yr for the northern area and 1.1 to 1.2 ± 0.2 mm/yr in the southern part. These analyses (like many others, e.g., Gehrels and Woodworth, 2012) were based on linear regression of the mid point of all sea-level index points from a single region. This approach does not account for the vertical or temporal error associated with any RSL reconstruction. We addressed this limitation by applying the spatio-temporal empirical hierarchical model to the expanded Delaware Bay database (Engelhart and Horton, 2012 with the new Sea Breeze reconstructions) to account for the vertical and temporal uncertainties of each index point.

The use of a temporal Gaussian process prior for the sea-level signal allows for non-linear variations in this term, while the spatial model for land-level change allows information to be shared between adjacent sites. Importantly, the spatial model also allows for fully consistent comparisons of differences in rates of RSL between sites that allow for the presence of both common signals and site-specific changes. Such models have been previously applied to a variety of historic (Kopp, 2013) and late Holocene (Kemp et al., 2014, van de Plasshe et al., 2014; Engelhart et al., 2015, Kopp et al., In Press) sea-level data sets. By fitting the spatio-temporal field as a whole and thus accounting for the correlation of sea-level change between sites, they allow statistically robust estimation of multi-site averages and differences.

Prior to tidal correction, the average rate of RSL change at sites in the upper bay was 1.45 ± 0.34 mm/yr, and the average rate in the lower bay was 1.11 ± 0.30 mm/yr. The probability that sea-level rose faster in the upper bay was 97%. Following correction for tidal-range change rates of RSL change within the Delaware Bay during the last 2200 years do not show a significant difference between the upper (1.26 ± 0.33 mm/yr) and lower Bay (1.15 ± 0.29 mm/yr) and the probability of faster sea-level rise in the upper bay falls to 73% (Figure 6; Table 3). The paleotidal model indicates that corrections of RSL reconstructions for tidal-range change are greatest (up to 0.4 m) in the upper bay. This convergence is consistent with predictions from Earth-ice models (Engelhart et al., 2011a; Roy and Peltier, 2015) that do not show different rates of Common Era RSL rise between the upper and lower Bay and likely explains why agreement was observed between models and data only in the upper Bay (Engelhart et al., 2011). A similar study by Shennan and Horton (2002) and Horton and Shennan (2009) in eastern England found that the changes in tidal range at the open coast are relatively small, but within the large estuaries of the Humber and the Wash modeled tidal ranges were significantly less in the late Holocene. The modeled changes for both the Humber inner and outer estuary offered an explanation for the apparent differences between the RSL reconstructions produced from these location. Shennan and Horton (2002) concluded that the rate of Common Era land subsidence within estuaries is prone to being overestimated unless tidal-range changes are quantified.

The slightly lower mean estimate of the rates in the lower bay result from the influence of two neighboring sites: Cape Henlopen (1.00 ± 0.35 mm/yr) and Great Marsh/Broadkill Beach (0.81 ± 0.42 mm/yr). Excluding these two sites, the average rate in the lower bay after tidal correction is increased to 1.30 ± 0.36 mm/yr and is indistinguishable from the rate in the upper bay (probability of a higher rate in the upper bay than the lower bay = 41%). The Cape Henlopen spit currently curves from the south to the north-west around a breakwater built early in the 19th century (Kraft et al., 1978). The spit previously had a cusped foreland that recurved over the past 300 to 2,000 years (Kraft et al., 1978). Such intricate paleogeography is not included in the current paleotidal model and, furthermore, such depositional environments have a complex relation with sea level (Belknap and Kraft, 1977; Long et al., 2014).

The highest mean estimates of the rates of RSL rise in Delaware Bay occur at Jake's Landing, NJ (1.49 ± 0.87 mm/yr). The high rates are supported by Varekamp and Thomas (1998), who estimated a rate of RSL rise of 6.9 mm/yr since the 1650 CE, using ²¹⁰Pb and AMS radiocarbon dating salt-marsh sediment at the site. This historic rates is much higher than a comparable study from Cape May, NJ (Kemp et al., 2013a). This region is susceptible to compaction due to the thick muds created by the Dennis Creek drainage system that underlie the salt-marsh peat (Meyerson, 1972). The compaction of salt-marsh stratigraphies cause post-depositional lowering of salt-marsh samples used to reconstruct sea level, because of compaction creates an estimation of former sea level that is too low and a rate of rise that is too great (Belknap and Kraft, 1977; Bloom, 1964; Brain et al., 2012; Horton and Shennan, 2009; Kaye and Barghoorn, 1964). We propose that the anomalously high rates of RSL rise at Jake's Landing are the result of rapid and localized sediment compaction.

Excluding Cape Henlopen, Great Marsh, and Jake's Landing, the tidal-range-corrected rate of RSL rise in Delaware Bay over the last ~2200 yr (1.25 ± 0.27 mm/yr) is indistinguishable from the 1.3 ± 0.1 mm/yr estimated for the New Jersey Atlantic coast (Horton et al., 2013), which was also corrected for tidal range changes (Figure 5C). This rate is also comparable to rates estimate for the Delaware Atlantic coast (Horse Island, 1.39 ± 0.48 mm/yr; and Rehoboth Bay, 1.29 ± 0.78 mm/yr). This similarity among the Atlantic coast of Delaware, New Jersey and the inner and outer parts of the Delaware estuary support inferences from Earth-ice models, tide-gauge records (e.g., Kopp, 2013), and global positioning systems (e.g., Sella et al., 2007; Snay et al., 2007) that maximum rates of land-level changes along the U.S. Atlantic coast occur in the mid-Atlantic region.

6. CONCLUSIONS

We have produced a new RSL record covering the last ~2200 years from a series of eight basal salt-marsh peat samples. We estimated the elevation at which these samples formed (the indicative meaning) by using plant macrofossils supported by foraminiferal assemblages. Our new sea-level index points have significantly improved vertical and age precision compared to previously available data due primarily to high-resolution leveling methods and the use of AMS radiocarbon dating of *in-situ* plant macrofossils. We utilized a previously developed high-resolution tidal model for the Delaware Bay (Hall et al., 2013) to estimate the effects of changes in tidal range during the late Holocene and to correct our RSL record. Not considering the effects of tidal range would result in an over-estimation of the rate of rise at Sea Breeze of 0.2 mm/yr. We estimated a rate of RSL rise from 2200 to 150 years BP at Sea Breeze of 1.12 ± 0.22 mm/yr using a spatio-temporal empirical hierarchical model.

Utilizing the paleotidal correction removes the difference previously observed between RSL change in the upper and lower Delaware Bay (e.g., Engelhart et al., 2009; Leorri et al., 2013). We estimate the rates of RSL are 1.26 ± 0.33 mm/yr and 1.30 ± 0.36 mm/yr for the upper and lower Bay, respectively. The rates of rise now estimated for Delaware Bay (1.25 ± 0.27 mm/yr) using the tidal corrections are similar to the ~1.3 mm/yr observed in sites less prone to the influence of tidal range in New Jersey, Maryland, and Virginia, confirming earlier research that the maximal ongoing forebulge collapse is focused on the mid-Atlantic. Our improvement in the precision of index points and the application of a spatio-temporal empirical hierarchical model will enable further advances in the understanding of the regional (eustatic, glacio-isostatic and hydro-isostatic) and the local (coastal morphology, sediment supply, tidal regime and terrestrial and fluvial input) processes that control RSL.

7. ACKNOWLEDGMENTS

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8. TABLES

TABLE 1: RADIOCARBON DATES

Core (Depth)	Latitude	Longitude	Altitude (m MTL)	Lab Identifier	^{14}C Age $\pm 1\sigma$	$\delta^{13}\text{C}$ (‰, PDB)	Dated Material
SB30 (340cm)	39.31634	-75.3150	-2.36	OS-84497	1760 ± 25	-24.63	<i>Schoenoplectus</i> sp. fragment
SB32 (421cm)	39.3147	-75.315	-3.16	OS-88743	2210 ± 30	-14.72	<i>Distichlis spicata</i> rhizome
SB76 (163m)	39.32461	-75.317	-0.77	OS-88662	960 ± 25	-22.73	<i>Schoenoplectus</i> sp. stem and rhizome
SB81 (320cm)	39.3217	-75.319	-2.46	OS-88661	1840 ± 30	-25.94	Fragment of <i>Phragmites australis</i> stem
SB83 (277cm)	39.3217	-75.318	-1.93	OS-80666	1740 ± 30	-13.83	<i>Spartina patens</i> rhizome
SB84 (222cm)	39.3220	-75.317	-1.5	OS-88663	1260 ± 25	-13.60	<i>Distichlis spicata</i> rhizome
SB98 (203cm)	39.3229	-75.3159	-1.25	OS-80667	1010 ± 25	-12.76	<i>Spartina patens</i>
SB110 (135cm)	39.3238	-75.315	-0.52	OS-80668	830 ± 25	-12.55	<i>Spartina patens</i> rhizome

All age determinations were made at the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) facility (OS lab identifiers).

Reported ages follow rounding conventions and are expressed in radiocarbon years before present. Samples underwent acid-base-acid pretreatment. $\delta^{13}\text{C}$ (relative to Pee Dee Belemnite, PDB) was measured in an aliquot of CO_2 gas collected during sample combustion and was used to correct ^{14}C measurements. The reported $\delta^{13}\text{C}$ values are for the dated macrofossil and do not represent bulk sediment.

TABLE 2: SUMMARY OF SEA-LEVEL INDEX POINTS FROM THE DELAWARE BAY DATABASE AFTER CORRECTION FOR TIDAL RANGE CHANGE

Index point	Site Name	Latitude (°N, decimal degrees)	Longitude (°E, decimal degrees)	Lab Identifier	¹⁴ C Age ± 1σ	Calibrated age range (2σ years before present, 1950 CE)	Tidal range correction (m)	Altitude (m MTL)	Reference water level (m MTL)	RSL (m)	Type	Citation
1	Bowers	39.0486	-75.3883	P-1687	1952±45	2002-1745	-0.11	-1.77	0.37	-2.14±0.96	Intercalated	Belknap (1975)
2	Bowers	39.0519	-75.3903	P-1686	1950±55	2036-1736	-0.11	-4.06	0.37	-4.43±0.95	Basal	Belknap (1975)
3	Broadkill Beach	38.8019	-75.2044	I-4353	1990±100	2300-1706	-0.02	-3.55	0.77	-4.32±0.82	Basal	Belknap (1975)
4	Cape Henlopen	38.7639	-75.0972	TEM-158	280±60	496-0	0.00	-0.58	0.40	-0.98±0.95	Basal	DGS (unpublished)
5	Cape Henlopen	38.7653	-75.0986	TEM-164	690±100	892-513	-0.02	-0.97	0.37	-1.34±0.95	Basal	DGS (unpublished)
6	Cape Henlopen	38.7642	-75.0975	TEM-163	750±70	896-555	-0.02	-1.37	0.37	-1.74±0.95	Basal	DGS (unpublished)
7	Cape Henlopen	38.7681	-75.1056	TEM-165	760±70	900-558	-0.02	-1.43	0.38	-1.81±0.95	Basal	DGS (unpublished)
8	Cape Henlopen	38.7600	-75.1000	I-8119	920±90	980-677	-0.02	-1.54	0.37	-1.91±0.95	Basal	Belknap (1975)
9	Cape Henlopen	38.7642	-75.0975	TEM-162	930±90	1046-680	-0.02	-0.79	0.37	-1.16±0.95	Basal	DGS (unpublished)

10	Cape Henlopen	38.7611	-75.0958	TEM-157	940±120	1172-664	-0.02	-1.19	0.37	-1.56±0.95	Basal	DGS (unpublished)
11	Cape Henlopen	38.7681	-75.1056	TEM-166	980±120	1173-684	-0.02	-0.91	0.38	-1.29±0.95	Basal	DGS (unpublished)
12	Cape Henlopen	38.7639	-75.0972	TEM-161	1100±90	1258-799	-0.02	-1.12	0.37	-1.49±0.96	Basal	DGS (unpublished)
13	Cape Henlopen	38.7639	-75.0972	TEM-160	1150±80	1261-930	-0.02	-1.58	0.37	-1.95±0.95	Basal	DGS (unpublished)
14	Cape Henlopen	38.7528	-75.1192	GX- 16214	1775±150	2038-1353	-0.01	-3.66	0.71	-4.37±0.58	Intercala ted	Fletcher et al. (1993)
15	Cape Henlopen	38.7539	-75.1203	GX- 16221	1885±170	2302-1414	-0.01	-3.38	0.71	-4.09±0.58	Intercala ted	Fletcher et al. (1993)
16	Cape Henlopen	38.7539	-75.1203	GX- 16220	1910±245	2425-1302	-0.01	-1.24	0.39	-1.63±0.76	Intercala ted	Fletcher et al. (1993)
17	Cape Henlopen	38.7542	-75.1156	GX- 15835	2095±205	2699-1572	-0.01	-3.31	0.71	-4.02±0.58	Intercala ted	Fletcher et al. (1993)
18	Great Marsh	38.7861	-75.1719	Beta- 14681	80±60	274-9	0.00	-0.50	0.79	-1.29±0.54	Basal	Stedman (1990)
19	Great Marsh	38.7856	-75.1714	Beta- 14683	670±70	726-537	-0.05	-0.90	0.74	-1.64±0.54	Basal	Stedman (1990)
20	Great Marsh	38.7853	-75.1714	Beta- 14684	930±80	968-690	-0.05	-1.10	0.74	-1.84±0.54	Basal	Stedman (1990)
21	Great Marsh	38.7858	-75.1717	Beta 14682	950±90	1051-690	-0.05	-0.78	0.75	-1.53±0.54	Basal	Stedman (1990)

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22	Great Marsh	38.7853	-75.1711	Beta-14685	1150±80	1261-930	-0.05	-1.00	0.74	-1.74±0.54	Basal	Stedman (1990)
23	Great Marsh	38.7850	-75.1708	Beta-14686	1370±60	1385-1180	-0.05	-1.20	0.74	-1.94±0.54	Basal	Stedman (1990)
24	Great Marsh	38.7847	-75.1706	Beta-14687	1650±70	1711-1389	-0.01	-1.45	0.40	-1.85±0.75	Basal	Stedman (1990)
25	Great Marsh	38.7781	-75.1736	I-4625	2330±100	2714-2135	-0.01	-4.71	0.40	-5.11±0.97	Basal	Belknap (1975)
26	Horse Island Marsh	38.6717	-75.1342	Beta-118808	170±80	426-0	0.00	-0.73	0.70	-1.43±0.49	Basal	Nikitina et al. (2000)
27	Horse Island Marsh	38.6717	-75.1342	Beta-118807	960±50	960-764	-0.03	-1.45	0.67	-2.12±0.49	Basal	Nikitina et al. (2000)
28	Horse Island Marsh	38.6700	-75.1300	Beta-118811	1810±60	1876-1572	-0.01	-3.29	0.38	-3.67±1.07	Intercalated	DGS (unpublished)
29	Jakes Landing	39.1700	-74.8600	Teledyne	1335±95	1413-999	-0.09	-2.35	0.78	-3.13±0.94	Intercalated	Meyerson (1972)
30	Leipscic River	39.2497	-75.4695	Beta-118804	80±60	274-9	0.00	-0.56	0.90	-1.46±0.49	Intercalated	Nikitina (2003)
31	Leipscic River	39.2486	-75.4583	Beta-118801	300±60	502-0	0.00	-0.61	0.90	-1.51±0.49	Intercalated	Nikitina (2003)
32	Leipscic	39.2525	-75.4597	Beta-	970±80	1055-709	-0.16	-1.22	0.74	-1.96±0.48	Basal	Nikitina et al.

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33	Leipsic River	39.2464	-75.4106	Beta-117238	1130±100	1281-803	-0.16	-1.46	0.75	-2.21±0.48	Intercalated	Nikitina (2003)
34	Leipsic River	39.2511	-75.4686	GrN-18995	1160±50	1229-961	-0.16	-2.52	0.74	-3.26±0.48	Basal	Nikitina et al. (2000)
35	Leipsic River	39.2489	-75.4569	Beta-118800	1770±60	1857-1552	-0.35	-2.84	0.55	-3.39±0.47	Basal	Nikitina et al. (2000)
36	Leipsic River	39.2506	-75.4689	GrN-18994	2030±80	2301-1818	-0.34	-2.85	0.55	-3.40±0.47	Basal	Nikitina et al. (2000)
37	Leipsic River	39.2494	-75.4689	Beta-118803	2070±80	2306-1870	-0.34	-2.50	0.55	-3.05±0.47	Basal	Nikitina et al. (2000)
38	Marsh Island	38.6700	-75.1300	I-8118	690±85	780-530	-0.02	-0.90	0.37	-1.27±0.94	Basal	Belknap (1975)
39	Rehoboth Bay	38.6694	-75.0683	R-4101_a	250±140	501-0	0.00	-0.77	0.38	-1.15±0.96	Intercalated	Belknap (1975)
40	Rehoboth Bay	38.6694	-75.0700	R-4100_a	350±130	629-0	0.00	-0.38	0.68	-1.06±0.83	Intercalated	Belknap (1975)
41	Sea Breeze	39.3162	-75.3153	OS-84498	165±25	285-0	0.00	-0.25	0.93	-1.18±0.28	Intercalated	Nikitina et al. (2014)
42	Sea Breeze	39.3191	-75.3179	OS-80662	365±25	499-318	0.00	-0.83	0.93	-1.76±0.28	Intercalated	Nikitina et al. (2014)
43	Sea Breeze	39.3204	-75.3177	OS-84495	425±25	521-340	0.00	-1.08	0.30	-1.38±0.46	Intercalated	Nikitina et al. (2014)

44	Sea Breeze	39.3191	-75.3179	OS-80692	535±30	632-512	-0.07	-1.03	0.23	-1.26±0.40	Intercalated	Nikitina et al. (2014)
45	Sea Breeze	39.3146	-75.3158	Beta-240760	810±40	789-674	-0.06	-1.30	0.23	-1.53±0.40	Intercalated	Nikitina et al. (2014)
46	Sea Breeze	39.3191	-75.3180	Beta-252034	810±40	789-674	-0.06	-1.68	0.23	-1.91±0.40	Intercalated	Nikitina et al. (2014)
47	Sea Breeze	39.3238	-75.3153	OS-80668	830±25	784-692	-0.16	-0.65	0.77	-1.42±0.26	Basal	This paper
48	Sea Breeze	39.3191	-75.3179	OS-88662	960±25	929-796	-0.17	-0.87	0.77	-1.64±0.26	Basal	This paper
49	Sea Breeze	39.3229	-75.3160	OS-80667	1010±25	969-804	-0.16	-1.37	0.77	-2.14±0.26	Basal	This paper
50	Sea Breeze	39.3191	-75.3179	OS-80663	1010±25	969-804	-0.06	-1.71	0.23	-1.94±0.40	Intercalated	Nikitina et al. (2014)
51	Sea Breeze	39.3191	-75.3179	Beta-252035	1070±40	1059-927	-0.06	-1.89	0.23	-2.12±0.40	Intercalated	Nikitina et al. (2014)
52	Sea Breeze	39.3162	-75.3153	OS-84499	1090±25	1057-938	-0.16	-1.53	0.77	-2.30±0.26	Intercalated	Nikitina et al. (2014)
53	Sea Breeze	39.3191	-75.3179	OS-80664	1170±30	1179-986	-0.07	-1.95	0.23	-2.18±0.40	Intercalated	Nikitina et al. (2014)
54	Sea Breeze	39.3220	-75.3173	OS-88663	1260±25	1280-1091	-0.16	-1.61	0.77	-2.38±0.26	Basal	This paper
55	Sea Breeze	39.3191	-75.3179	Beta-252036	1460±40	1413-1292	-0.06	-2.65	0.23	-2.88±0.40	Intercalated	Nikitina et al. (2014)

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56	Sea Breeze	39.3204	-75.3177	OS-84496	1550±25	1524-1385	-0.17	-2.31	0.77	-3.08±0.26	Intercalated	Nikitina et al. (2014)
57	Sea Breeze	39.3217	-75.3181	OS-80666	1740±30	1713-1566	-0.40	-2.05	0.54	-2.59±0.23	Basal	This paper
58	Sea Breeze	39.3162	-75.3153	OS-84497	1760±25	1734-1571	-0.40	-2.49	0.54	-3.03±0.24	Basal	This paper
59	Sea Breeze	39.3217	-75.3193	OS-88661	1840±30	1864-1708	-0.40	-2.59	0.54	-3.13±0.23	Basal	This paper
60	Sea Breeze	39.3146	-75.3158	OS-88743	2210±30	2319-2148	-0.39	-3.27	0.54	-3.81±0.24	Base of Basal	This paper
61	Sea Breeze	39.3146	-75.3158	Beta-240761	2240±40	2341-2153	-0.39	-3.03	0.54	-3.57±0.24	Basal	Nikitina et al. (2014)
62	Slaughter Beach	38.9047	-75.2964	I-9230	720±80	792-540	-0.05	-0.97	0.40	-1.37±0.97	Intercalated	Kraft (1976)
63	Slaughter Beach	38.9261	-75.3222	I-9228	1690±85	1813-1409	-0.05	-1.98	0.40	-2.38±0.97	Intercalated	Kraft (1976)
64	Slaughter Beach	38.9317	-75.3175	TEM-172	2020±110	2306-1721	-0.08	-4.32	0.73	-5.05±0.82	Basal	Marx (1981)
65	Smyrna	39.3017	-75.5983	DC-3_c	1370±110	1524-1058	-0.10	-2.32	0.38	-2.70±0.68	Intercalated	Rogers & Pizzuto (1994)

TABLE 3: RATES OF RELATIVE SEA-LEVEL RISE (2200-150 YEARS BP) WITH AND WITHOUT PALEOTIDAL RANGE CORRECTION

Location	Rate with tidal correction (mm/yr; $\pm 2\sigma$)	Rate without tidal correction (mm/yr; $\pm 2\sigma$)
<i>Delaware Bay</i>	<i>1.19 $\pm$ 0.26</i>	<i>1.25 $\pm$ 0.26</i>
<i>Delaware Bay ex. Cape Henlopen, Great Marsh and Jake Landing</i>	<i>1.25 $\pm$ 0.27</i>	<i>1.33 $\pm$ 0.28</i>
<i>Upper Bay</i>	<i>1.26 $\pm$ 0.33</i>	<i>1.45 $\pm$ 0.34</i>
Sea Breeze	1.12 $\pm$ 0.22	1.33 $\pm$ 0.23
Smyrna	1.15 $\pm$ 0.63	1.33 $\pm$ 0.65
Leipsic River	1.27 $\pm$ 0.30	1.48 $\pm$ 0.31
Jakes Landing	1.49 $\pm$ 0.87	1.64 $\pm$ 0.90
<i>Lower Bay</i>	<i>1.15 $\pm$ 0.29</i>	<i>1.11 $\pm$ 0.30</i>
<i>Lower Bay ex. Cape Henlopen, Great Marsh</i>	<i>1.30 $\pm$ 0.36</i>	<i>1.21 $\pm$ 0.36</i>
Bowers	1.19 $\pm$ 0.45	1.24 $\pm$ 0.45
Slaughter Beach	1.33 $\pm$ 0.54	1.34 $\pm$ 0.54
Great Marsh	0.69 $\pm$ 0.33	0.78 $\pm$ 0.34
Cape Henlopen	1.00 $\pm$ 0.40	1.05 $\pm$ 0.40
Horse Island Marsh	1.39 $\pm$ 0.48	1.14 $\pm$ 0.49
Rehoboth Bay	1.29 $\pm$ 0.78	1.10 $\pm$ 0.78

9. FIGURE CAPTIONS

Figure 1: (A) Map of the Delaware Bay and coast of southern New Jersey, USA showing the location of sites referred to in the main text. (B) Location of the Sea Breeze, NJ study site on the north coast of the Delaware Bay. The location of cores used to reconstruct relative sea level are shown and the transects used to describe sediment beneath the Sea Breeze salt marsh.

Figure 2: Stratigraphy underlying Sea Breeze described from a series of cores collected along transects. These transects are a subset of (and modified from) those used by Nikitina et al. (2014). The basal sand-peat contact was targeted to produce compaction-free relative sea-level reconstructions across a range of elevations. Ages shown for radiocarbon dates are in radiocarbon years.

Figure 3: Microfossil diagram for sea-level index point from cores (A) SB30, (B) SB32, (C) SB76, (D) SB81, (E) SB83, (F) SB84, (G) SB98, (H) SB110. Foraminifera expressed as % as relative abundances. Red bars indicate assemblage (expressed as percentage) at radiocarbon-dated level, grey bars represent other counts. Samples where few individuals of the indicated species were present are represented by '+' symbols. Depth (m) down-core shown to the left of the lithology column. Core-top altitudes (m MTL) are listed above each lithology column.

Figure 4: Contours of Great Diurnal Range (GT) calculated from the tidal model of Hall et al. (2013) for Delaware Bay and the Western North Atlantic illustrating temporal changes at (A) present day, (B) 1000 years BP, (C) 2000 years BP, and (D) 3000 years BP.

Figure 5: Relative sea level (RSL) reconstructions. Sea level index points are represented by boxes with a width that includes 2σ calibrated radiocarbon ages and height includes sea level uncertainty. (A) RSL reconstructions from Sea Breeze before and after correction for tidal-range change. (B) Basal sea level index points from Delaware (Engelhart and Horton, 2012), and Sea Breeze (this study) corrected for tidal range. (C) Basal sea level index points from New Jersey (Engelhart and Horton, 2012), and Sea Breeze (this study) corrected for tidal range.

Figure 6: Average relative sea-level rise field (2200-150 BP) estimated for the entire study area (A) without and (C) with paleotidal range correction. Uncolored areas exhibit a posterior variance $> 10\%$ of the prior variance. (B, D) This field is sub-sampled to obtain rates for individual sites. (1) Sea Breeze; (2) Smyrna; (3) Leipsic River; (4) Bowers; (5) Slaughter Beach; (6) Great Marsh/Broadkill Beach; (7) Cape Henlopen; (8) Horse Island Marsh/Marsh Island; (9) Rehoboth Bay; (10) Jake's Landing.

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Figures

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6