APPENDIX A: SEA-LEVEL INDEX POINTS

The standardized methodology for reconstructing former sea levels from low energy, sedimentary environments has been established during the International Geological Correlation Programs (IGCP) (van de Plassche, 1986; Shennan and Horton, 2002; Edwards, 2006). To be a validated sea-level index point (SLI), a sample must have a location, an age and a known relationship between the sample and a known tidal level and the indicative meaning (Shennan, 1986; van de Plassche, 1986). The indicative meaning is constructed of two parameters, the reference water level (e.g. mean higher high water (MHHW)) and the indicative range (the vertical range over which the sample could occur). To constrain the indicative meaning of the index points in the US Atlantic database, we have used published zonations of modern vegetation (Redfield, 1972; Niering and Warren, 1980; Lefor et al., 1987; Gehrels, 1994) and the distribution of microfossils (Gehrels, 1994) supported by $\delta^{13}C$ values from the radiocarbon-dated sediments (Andrews et al., 1998; Törnqvist et al., 2004). As an example, where we have a floral and/or faunal indication that a sample was formed within a salt marsh environment but cannot be identified as specifically high or low marsh, the index point is conservatively estimated to have formed between MHHW and mean tide level (Törnqvist et al., 2004). For samples where we have a positive identification of plant macrofossil species, we can reduce the indicative range. Where authors have used microfossils to quantitatively assess the relationship between the sample and former sea level, these predictions of the indicative meaning have been retained. In practice, over 70% of the samples in the database can only be identified as salt-marsh deposits.

The relative sea level of the sea-level index points is calculated using the equation:

$$\text{Relative Sea Level} = \text{Elevation}_{\text{sample}} - \text{Reference Water Level}_{\text{sample}}$$

where the elevation and reference water level are expressed in meters relative to the national datum, NAVD 88, and subsequently corrected to local mean sea level (MSL).

For each sample, we calculated the vertical error of the index point from a variety of factors that are inherent to sea-level research (Shennan, 1986). Further errors are incorporated including the type of coring equipment used, techniques of depth measurement and the compaction of the sediment during penetration (Woodroffe, 2006). We also included an error estimate associated with the leveling of the sample with respect to NAVD 88. For high precision leveling using modern techniques, this can be as low as ±0.05 m but can rise as high as ±0.5 m for less precise methods. A further error is included due to the leveling of the sample to local tide levels. This is typically ±0.1 m but may be much larger, particularly when samples are collected offshore (Shennan, 1986). The errors in this study do not include the effects of tidal range change through time; we assume that this influence is minimal (Gehrels et al., 1995). The total error (E_h) for each sample is then calculated from the expression:

$$E_h = (e^2{}_1 + e^2{}_2 \ldots + e^2{}_n)^{1/2}$$

Where $e_1 \ldots e_n$ are the individual sources of error.

A further source of error in sea-level reconstruction is sediment consolidation, that is, compression of a sedimentary package by its own weight or the weight from overlying sediment (Kaye and Barghoorn, 1964). The significance of sediment consolidation was recognized from early studies of North American (Bloom, 1964; Kaye and Barghoorn, 1964) and European (Jelgersma, 1961; Streif, 1971; van de Plassche, 1980) salt marshes. If consolidation is not corrected for, then index points will be lowered from their original elevation and the rate and magnitude of relative sea-level rise will be overestimated. However, correcting for the compaction of sediments is a complex process involving many variables (Pizzuto and Schwendt, 1997). Therefore, we have reduced the influence of compaction by only employing basal peat samples, which are deposited directly on the presumed compaction-free substrate (Kaye and Barghoorn, 1964).

Every SLI in the validated database (Figure DR1, Figure DR2) was radiocarbon dated and calibrated using CALIB 5.0.1 (Stuiver et al., 2005). We used a laboratory multiplier of 1 with 95% confidence limits and employed the dataset IntCal04 (Reimer et al., 2004). The database contains samples that were dated by Accelerator Mass Spectrometry (AMS), Gas Proportional Counting (GPC) and Liquid Scintillation Counting (LSC). Sample material in the database varies from dates on bulk peat to dates on identifiable salt marsh rhizomes.

APPENDIX B: LATE HOLOCENE RATES OF RELATIVE SEA-LEVEL RISE

We have used validated geological observations from basal peat over the last 4 ka (the late Holocene) to reconstruct background rates of sea-level rise. We assume that the ice-equivalent meltwater input over the last 4 ka is either zero (Douglas, 1995; Peltier, 1996, 2002) or minimal (Milne et al., 2005). A meltwater input of 1 m during the late Holocene (Church et al., 2008) would reduce the estimate of subsidence by 0.25 mm yr^{-1}. We also assume that the tectonic component is small, except in close proximity to the Cape Fear Arch, North Carolina, which has experienced uplift (Marple and Talwani, 2004).

When calculating the background rate of relative sea-level rise, it is necessary to remove the modern component, as this will overestimate the background rate due to the sea-level rise experienced during the 20th century (~0.2 – 0.3 m along the US Atlantic coast). In this study, we remove this modern sea level rise by using the nearest reliable tide gauge rate to extrapolate to MSL in 1900 AD. We then express all dates with respect to 1900 AD. At all sites the linear regression is run over the last 4 ka and is forced through zero. Regression errors are at the 95% confidence level. This contrasts with previous work (Gornitz, 1995; Peltier, 1996) that reported the error as the standard deviation and not the standard error.

Table DR1. Location of the 19 sites along the US Atlantic Coast and the rate of late Holocene (last 4 ka) relative sea-level rise (RSLR) derived from geological data. The references for the geological data are shown. GPS rates of vertical motion are from (1) Snay et al. (2007) and (2) Sella et al. (2007). Geological and GPS rates are shown with two sigma errors. Positive and negative values from the geological and GPS data refer to subsidence and uplift, respectively.

Site Number	**Site Name**	**Late Holocene RSLR (mm yr^{-1})**	**Rate from Nearest GPS Station (mm yr^{-1})**	**References**
1	Sanborn Cove, Maine	0.7 ± 0.1	1.9 ± 2.0 (1)	Gehrels and Belknap, 1993; Gehrels, 1999
2	Phippsburg, Maine	0.7 ± 0.5	-0.2 ± 3.2 (2)	Gehrels et al., 1996
3	Boston, Massachusetts	0.6 ± 0.1	2.3 ± 1.2 (2)	Newman et al., 1980; Donnelly, 2006
4	Barnstable, Massachusetts	1.2 ± 0.2	N/A	Redfield and Rubin, 1962; Stuiver et al., 1963
5	Clinton, Connecticut	1.1 ± 0.1	N/A	Cinquemani et al., 1982; van de Plassche, 1991; Nydick et al., 1995; van de Plassche et al., 2002
6	Hudson River, New York	1.2 ± 0.1	0.6 ± 3.0 (2)	Newman et al., 1980; Pardi et al., 1984
7	Northern Long Island, New York	0.8 ± 0.3	1.6 ± 3.0	Cinquemani et al., 1982; Pardi et al., 1984
8	Sandy Hook, New Jersey	1.4 ± 0.7	2.2 ± 1.4 (1)	Cinquemani et al., 1982
9	Atlantic City, New Jersey	1.3 ± 0.2	N/A	Stuiver and Daddario, 1963; Cinquemani et al., 1982; Pardi et al., 1984; Psuty, 1986
10	Inner Delaware Estuary, Delaware	1.7 ± 0.2	2.9 ± 2.0 (2)	Belknap, 1975; Belknap and Kraft, 1977; Nikitina et al., 2000
11	Lewes, Delaware	1.2 ± 0.2	1.1 ± 2.3 (1)	Elliot, 1972; Belknap, 1975; Belknap and Kraft, 1977; Fletcher et al., 1993; Ramsey and Baxter, 1996; Nikitina et al., 2000
12	Blackwater, Maryland	1.3 ± 0.2	2.2 ± 2.3 (1)	Cinquemani et al., 1982
13	Eastern Shore, Virginia	0.9 ± 0.3	3.5 ± 1.6 (2)	Engelhart and Kemp (unpublished)
14	Outer Banks, North Carolina	1.0 ± 0.1	N/A	Horton et al.; 2009 Cinquemani et al., 1982
15	Beaufort, North Carolina	0.7 ± 0.1	N/A	Horton et al. 2009; Cinquemani et al., 1982; Spaur and Snyder, 1999
16	Wilmington, North Carolina	0.8 ± 0.3	N/A	Cinquemani et al., 1982
17	Georgetown, South Carolina	0.8 ± 0.1	N/A	Cinquemani et al., 1982
18	Charleston, South Carolina	0.6 ± 0.1	1.6 ± 1.7 (1)	Cinquemani et al., 1982
19	Port Royal, South Carolina	0.6 ± 0.2	N/A	Cinquemani et al., 1982

APPENDIX C: UNCERTAINTY OF SEA-LEVEL TRENDS FROM TIDE GAUGE DATA

We identified 10 suitable tide gauge records along the US Atlantic Coast from the Permanent Service for Mean Sea Level (Woodworth and Player, 2003) that are at least 50 years in length and where the influence of non-GIA subsidence, such as groundwater withdrawal, is minimal. The tide gauge record at The Battery, New York, is truncated to only include data from the 20th century.

Formal uncertainties of trends of relative sea-level (RSL) obtained from tide gauge data are usually a few tenths of a mm per year for records longer than about 50 years. These formal uncertainties are optimistic, since tide gauge records do not satisfy the criteria for a linear regression, i.e., that the data consist of a trend plus Gaussian random noise. The records also contain interannual and longer variations of high amplitude that can negate the underlying trend of sea level for even many decades in some cases (Douglas, 2001).

As glacial isostatic adjustment (GIA) is considered to be the dominant control on the variation in the tide gauge records, we can assess the appropriate error term by running a linear regression through the rates from long-term tide-gauge records, going from areas of isostatic uplift in Canada to the proposed peak of GIA in the mid-Atlantic (Figure DR3). It is apparent that these rates lie along a straight line with little variation. Therefore, we can run a linear regression through these rates to produce a single estimate of the error for the tide gauges along the US Atlantic Coast of ± 0.3 mm yr^{-1}.

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SUPPLEMENTARY FIGURES

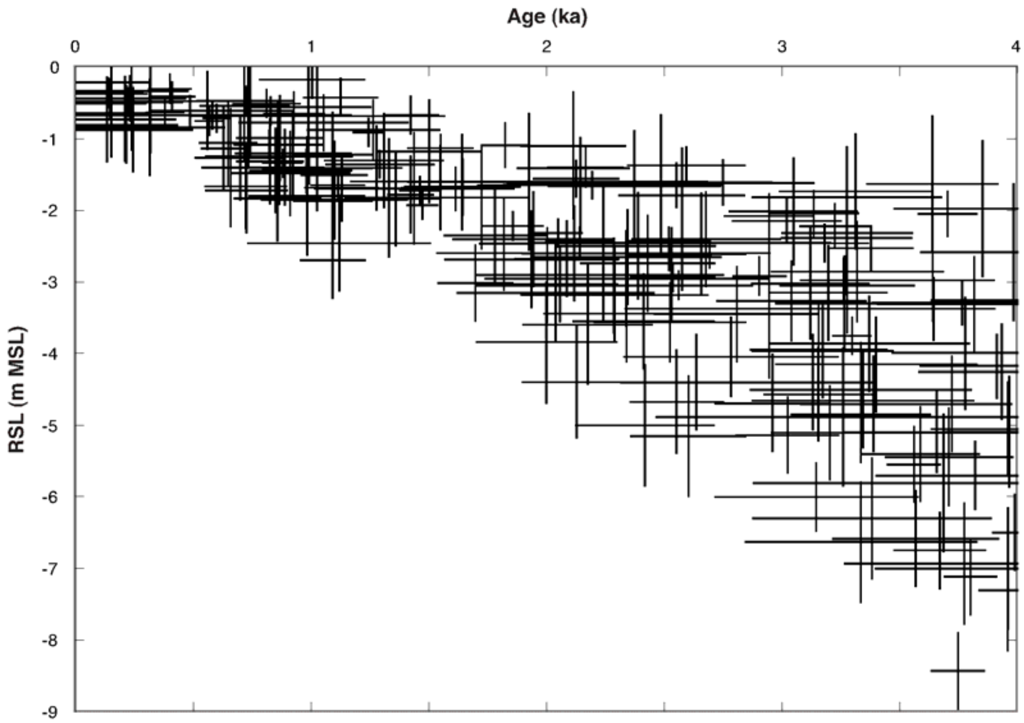
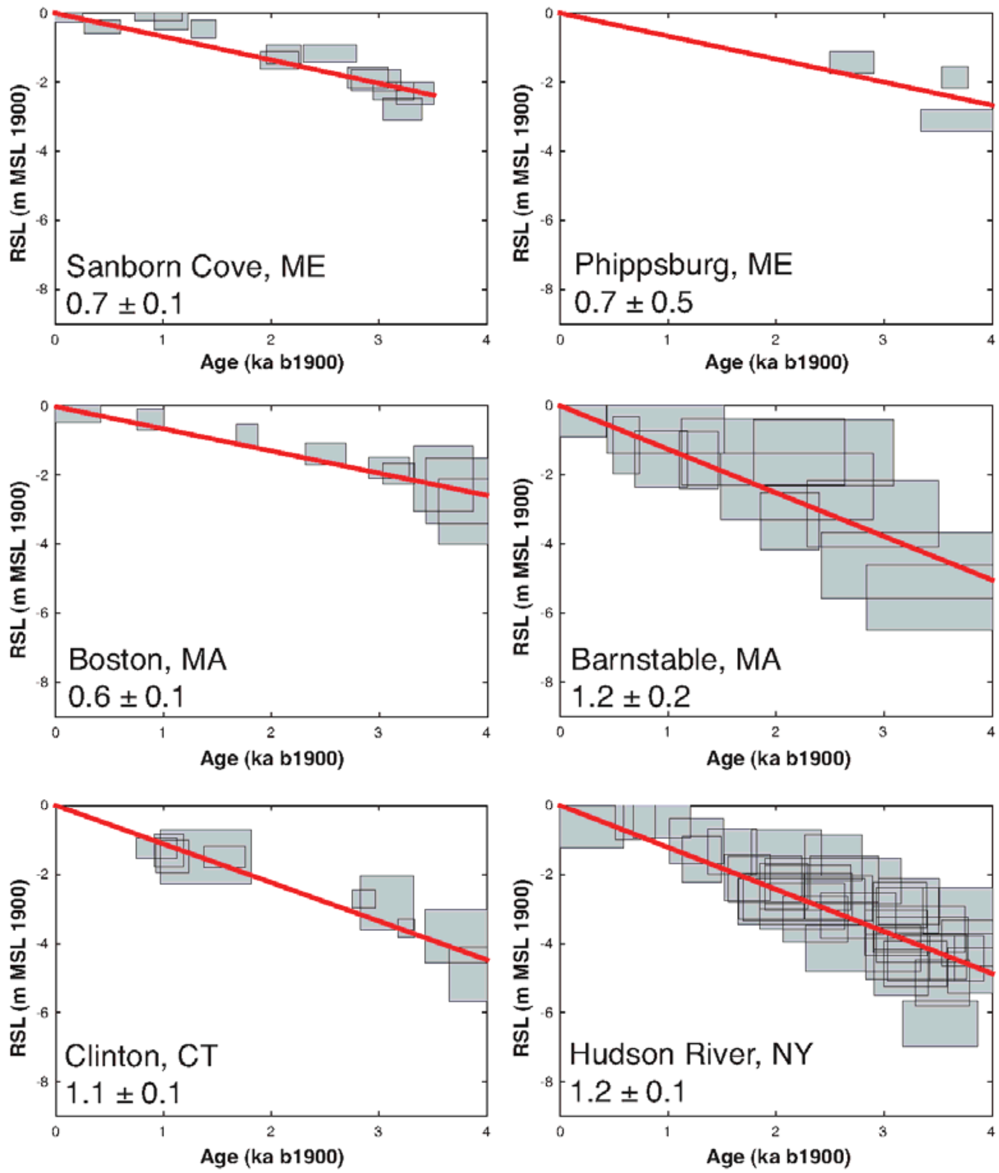


Figure DR1. All 212 radiocarbon dated basal index points, covering the last 4 kyr. The data demonstrates the considerable scatter caused by the differential GIA along the Atlantic Coast.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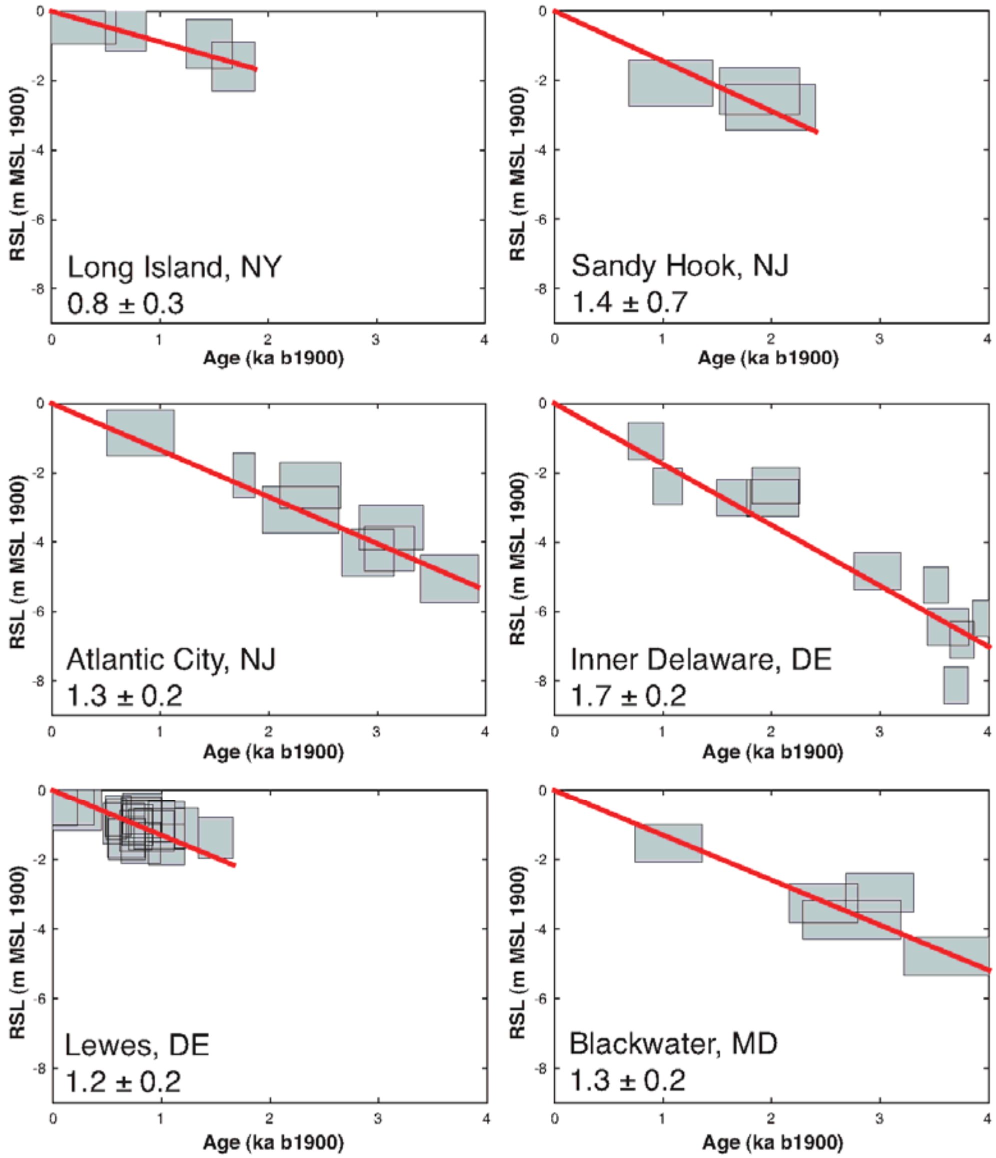
0.7 ± 0.1
Phippsburg, ME
0.7 ± 0.5
Boston, MA
0.6 ± 0.1
Barnstable, MA
1.2 ± 0.2
Clinton, CT
1.1 ± 0.1
Hudson River, NY
1.2 ± 0.1
RSL (m MSL 1900)
Age (ka b1900)


Long Island, NY
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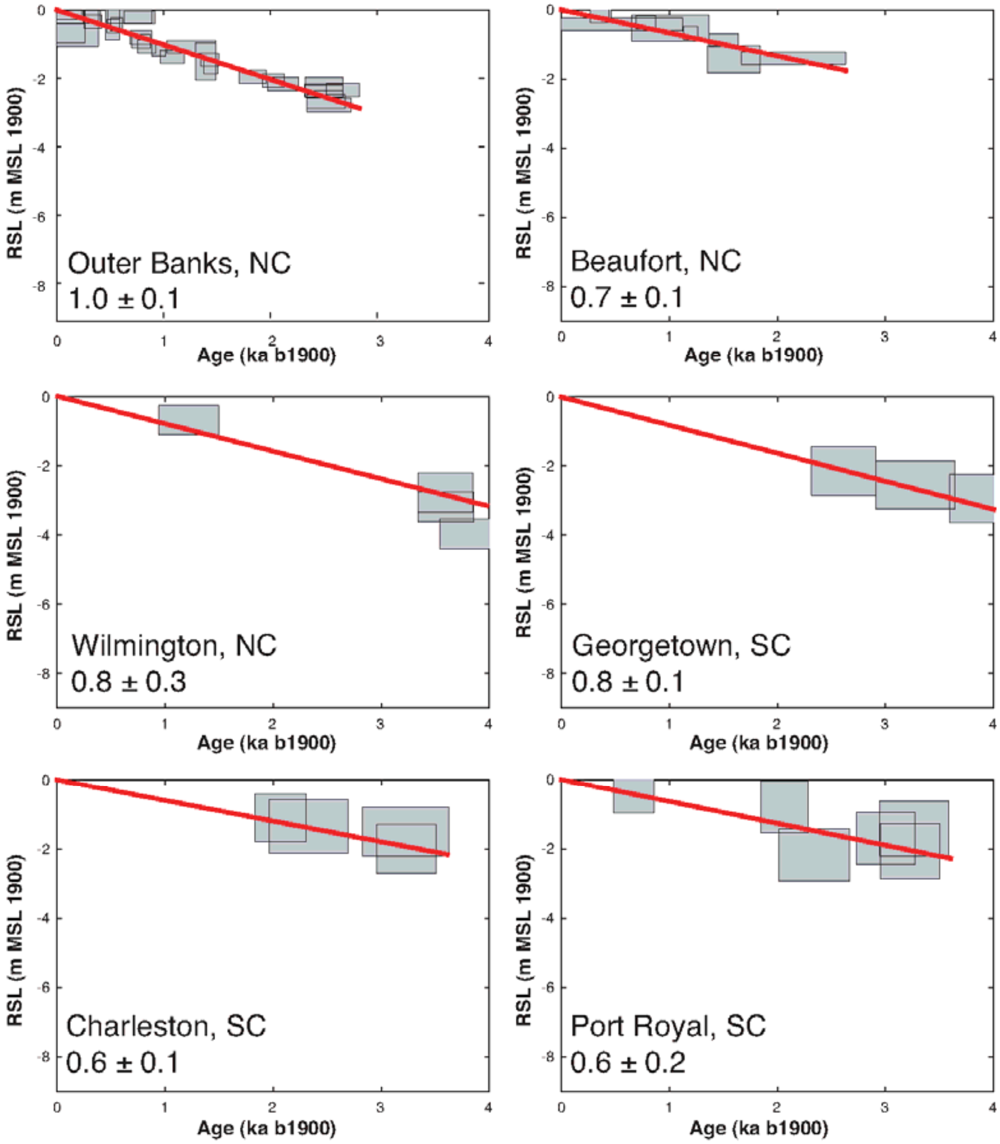
Sandy Hook, NJ
1.4 ± 0.7
Atlantic City, NJ
1.3 ± 0.2
Inner Delaware, DE
1.7 ± 0.2
Lewes, DE
1.2 ± 0.2
Blackwater, MD
1.3 ± 0.2
RSL (m MSL 1900)
Age (ka b1900)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1.0 ± 0.1
Beaufort, NC
0.7 ± 0.1
Wilmington, NC
0.8 ± 0.3
Georgetown, SC
0.8 ± 0.1
Charleston, SC
0.6 ± 0.1
Port Royal, SC
0.6 ± 0.2
RSL (m MSL 1900)
Age (ka b1900)

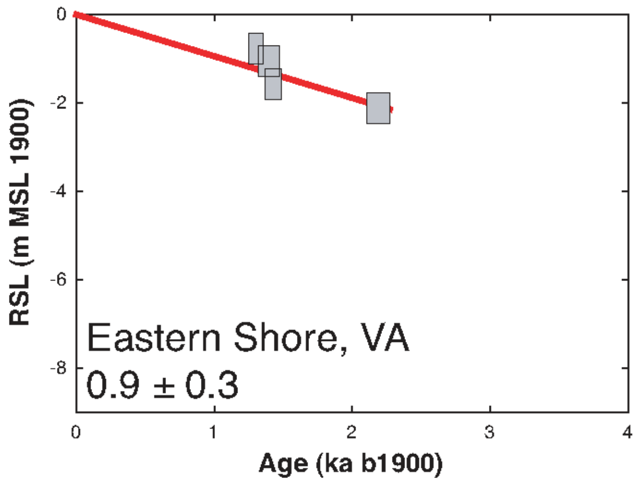

Figure DR2a-d. 19 locations along the US Atlantic Coast with 3 or more basal sea-level index points and the late Holocene rates of RSL rise. Sea-level index points are plotted as calibrated age versus change in RSL relative to MSL in AD 1900 (m). The red line is the linear regression for each site. Rates and errors shown to 1 d.p. Data sources for sea level index points are referenced in Table DR1.

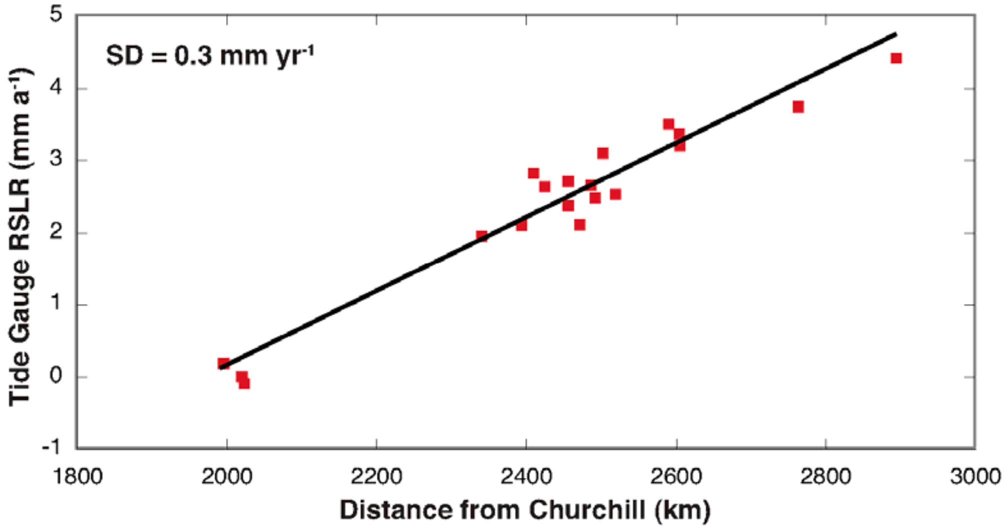

Figure 3. Long-term tide gauge records from Canada to Virginia, USA, plotted against distance from Churchill, Canada. The regression line demonstrates the methodology used to ascertain an appropriate error for the tide gauges.