

1 **RELATIVE SEA-LEVEL CHANGE IN NEWFOUNDLAND, CANADA DURING THE PAST ~3000 YEARS**

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3 Andrew C. Kemp<sup>1\*</sup>, Alexander J. Wright<sup>2</sup>, Robin J. Edwards<sup>3</sup>, Robert L. Barnett<sup>4,5</sup>, Matthew J.  
4 Brain<sup>6</sup>, Robert E. Kopp<sup>7,8</sup>, Niamh Cahill<sup>9</sup>, Benjamin P. Horton<sup>10,11</sup>, Dan J. Charman<sup>5</sup>, Andrea D.  
5 Hawkes<sup>12</sup>, Troy D. Hill<sup>13</sup>, and Orson van de Plassche<sup>2‡</sup>

- 6  
7 1. *Department of Earth and Ocean Sciences, Tufts University, Medford, MA 02155, USA*  
8 2. *Department of Marine Biogeology, Faculty of Earth & Life Sciences, Vrije Universiteit*  
9 *Amsterdam, 1081 HV Amsterdam, The Netherlands*  
10 3. *School of Natural Sciences, Trinity College Dublin, Dublin 2, Ireland*  
11 4. *Département de Biologie, Chimie et Géographie, Université du Québec à Rimouski,*  
12 *Rimouski, G5L 3A1, Canada*  
13 5. *Geography, College of Life and Environmental Sciences, University of Exeter, Exeter,*  
14 *EX4 4RJ, United Kingdom*  
15 6. *Department of Geography, Durham University, Durham, DH1 3LE, United Kingdom*  
16 7. *Department of Earth & Planetary Sciences, Rutgers University, Piscataway, NJ 08854*  
17 8. *Institute of Earth, Ocean & Atmospheric Sciences, Rutgers University, New Brunswick,*  
18 *NJ 08901*  
19 9. *School of Mathematics and Statistics, University College Dublin, Dublin 4, Ireland*  
20 10. *Asian School of the Environment, Nanyang Technological University, Singapore*  
21 11. *Earth Observatory of Singapore and Asian School of the Environment, Nanyang*  
22 *Technological University, Singapore*  
23 12. *Department of Earth and Ocean Sciences, University of North Carolina Wilmington,*  
24 *Wilmington, NC 28403, USA*  
25 13. *Louisiana Universities Marine Consortium, Chauvin, LA 70344, USA*

26  
27 \* *Corresponding author: andrew.kemp@tufts.edu; +1 617-627-0869*

28 ‡ *Deceased*

29 **ABSTRACT**

30 Several processes contributing to coastal relative sea-level (RSL) change in the North Atlantic  
31 Ocean are observed and/or predicted to have distinctive spatial expressions that vary by latitude.  
32 To expand the latitudinal range of RSL records spanning the past ~3000 years and the likelihood  
33 of recognizing the characteristic fingerprints of these processes, we reconstructed RSL at two  
34 sites (Big River and Placentia) in Newfoundland from salt-marsh sediment. Bayesian transfer  
35 functions established the height of former sea level from preserved assemblages of foraminifera  
36 and testate amoebae. Age-depth models constrained by radiocarbon dates and chronohorizons  
37 estimated the timing of sediment deposition. During the past ~3000 years, RSL rose by ~3.0 m at  
38 Big River and by ~1.5 m at Placentia. A locally calibrated geotechnical model showed that  
39 post-depositional lowering through sediment compaction was minimal. To isolate and quantify  
40 contributions to RSL from global, regional linear, regional non-linear, and local-scale processes,  
41 we decomposed the new reconstructions (and those in an expanded, global database) using a  
42 spatio-temporal statistical model. The global component confirms that 20<sup>th</sup> century sea-level rise  
43 occurred at the fastest, century-scale rate in over 3000 years ( $P > 0.999$ ). Distinguishing the  
44 contributions from local and regional non-linear processes is made challenging by a sparse  
45 network of reconstructions. However, only a small contribution from local-scale processes is  
46 necessary to reconcile RSL reconstructions and modeled RSL trends. We identified three  
47 latitudinally-organized groups of sites that share coherent regional non-linear trends and indicate  
48 that dynamic redistribution of ocean mass by currents and/or winds was likely an important  
49 driver of sea-level change in the North Atlantic Ocean during the past ~3000 years.

## 50 1. INTRODUCTION

51 Relative sea level (RSL) varies across space and through time in response to physical processes  
52 acting on a range of spatial and temporal scales. During the ~3000 years preceding the onset of  
53 historic sea-level rise in the late 19<sup>th</sup> century, global mean sea-level change was on the order of  $\pm$   
54 0.1 m (e.g., Kopp et al., 2016; Lambeck et al., 2014) and the primary driver of RSL change along  
55 the Atlantic coast of North America was glacio-isostatic adjustment (GIA), which varies  
56 regionally and is approximately linear over this time period (e.g., Peltier, 1996). While the  
57 magnitude and spatial pattern of global mean sea level and GIA during the past ~3000 years are  
58 relatively well understood and constrained, there remains considerable uncertainty in the  
59 contributions from regional-scale processes such as the dynamic redistribution of existing ocean  
60 mass by atmospheric and ocean circulation on timescales from years to centuries (e.g., Ezer,  
61 2015; Levermann et al., 2005; McCarthy et al., 2015; Piecuch et al., 2016; Woodworth et al.,  
62 2017) and the changing mass and distribution of land-based ice (e.g., Mitrovica et al., 2009).  
63 These dynamic processes are non-linear during the past ~3000 years and identifying their  
64 contributions to RSL change is important for two reasons. Firstly, they will be important  
65 modulators of future global sea-level rise and therefore regional-scale RSL projections will be  
66 improved through better understanding of these processes (Church et al., 2013). Secondly, if the  
67 contribution from regional-scale processes can be isolated, then RSL reconstructions can provide  
68 insight into paleoclimate trends and events (e.g., phasing of the North Atlantic Oscillation, or  
69 strength of Atlantic Meridional Overturning Circulation) on centennial timescales that are often  
70 poorly resolved in marine sedimentary archives due to low sedimentation rates and uncertainty  
71 introduced by marine reservoir corrections applied in the calibration of radiocarbon ages (e.g.,  
72 Keigwin and Boyle, 2000). However, separating regional, centennial-scale contributions to RSL

73 change from those caused by local processes such as sediment compaction using a sparse spatial  
74 network of RSL reconstructions is challenging.

75

76 Buried salt-marsh sediment is one of the principal sources of geological evidence for RSL  
77 changes that occurred during the past ~3000 years (e.g., Gehrels et al., 2005; Thomas and  
78 Varekamp, 1991). Salt-marsh plants and foraminifera are sea-level proxies because their vertical  
79 distribution is controlled by the frequency and duration of tidal flooding, which is correlated  
80 strongly with elevation (e.g., Edwards and Wright, 2015; Scott and Medioli, 1978). In response  
81 to RSL rise, salt-marshes accumulate sediment (including abundant plant remains that are well  
82 suited to radiocarbon dating) to maintain their position in the tidal frame and the resulting  
83 sequences of salt-marsh sediment become archives from which RSL can be reconstructed (e.g.,  
84 Redfield, 1972). On the Atlantic coast of North America, the long-term RSL rise needed to  
85 create accommodation space is caused by ongoing GIA. Earth-ice models predict that the hinge  
86 line between regions experiencing GIA-driven RSL rise/fall during the past ~3000 years occurs  
87 close to the Gulf of St. Lawrence (Figure 1; e.g., Peltier, 2004). Therefore, Newfoundland is one  
88 of the most northerly regions that is likely to preserve a detailed salt-marsh record of RSL  
89 changes, which makes it an important location for investigating regional, non-linear drivers of  
90 paleo-RSL because many of these processes vary with latitude.

91

92 To expand the latitudinal range of RSL reconstructions that span the past ~3000 years along the  
93 Atlantic coast of North America, we reconstructed RSL at two sites (Big River and Placentia) in  
94 Newfoundland, Canada (Figure 1) using foraminifera and testate amoebae preserved in cores of  
95 salt-marsh sediment that we dated using radiocarbon and recognition of pollution markers of

96 known age. Prior to this study, similar and near-continuous RSL reconstructions were available  
97 from northeastern Florida to Nova Scotia and the Magdalen Islands (Figure 1A). We use these  
98 new reconstructions alongside published datasets to explore the role of global, regional linear,  
99 regional non-linear, and local processes as drivers of sea-level change during the past ~3000  
100 years by decomposing RSL reconstructions using a spatio-temporal statistical model. We further  
101 quantify the role of post-depositional lowering through sediment compaction (a local-scale  
102 process) by calibrating and applying a geotechnical model to the Big River core.

103

## 104 **2. STUDY AREA**

105 The number and area of salt marshes in Newfoundland is small compared to other parts of  
106 Atlantic Canada and New England (Roberts and Robertson, 1986). The sediment preserved  
107 beneath many of Newfoundland's salt marshes was described by previous research (Bell et al.,  
108 2005; Brookes et al., 1985; Daly, 2002; Daly et al., 2007; Wright and van de Plassche, 2001).  
109 Based on this body of literature and an extensive coring program in the late 1990s, we selected  
110 Big River and Placentia (~430 km apart, in western and eastern Newfoundland, respectively;  
111 Figure 1) as sites at which to develop RSL reconstructions because they had sequences of high  
112 salt-marsh peat that we anticipated (from experience and existing publications; e.g., Gehrels et  
113 al., 2005; Kemp et al., 2017a) would yield near-continuous and precise records of RSL change  
114 during the past ~3000 years. Reconstructing RSL at sites in western and eastern Newfoundland  
115 provides a sampling regime to better distinguish regional- and local-scale processes.

116

117 Big River is located on the Port-au-Port Peninsula, and the salt marsh occupies a small, incised  
118 valley that is protected by a beach barrier/spit at its mouth (Figure 1B–D). An adjacent salt

119 marsh (Hynes Brook, ~1 km away) shares a similar geomorphology. Big River and Hynes Brook  
120 have a low salt-marsh floral community of *Spartina alterniflora* (tall form) and a mixed high  
121 salt-marsh plant community comprising variable combinations of *Iris versicolor*, *Schoenoplectus*  
122 spp., *Spartina patens*, *Distichlis spicata*, *Plantago maritima*, *Triglochin maritima*, *Glyceria*  
123 *borealis*, *Glaux maritima*, *Agrostis stolonifera*, and *Spartina alterniflora* (short form). Although  
124 these species are frequently intermixed, some mono-specific stands are present. The great diurnal  
125 tidal range (mean lower low water, MLLW to mean higher high water, MHHW) at Big River and  
126 Hynes Brook is 1.06 m. Up to ~3 m of salt-marsh peat underlies the Big River site (Wright and  
127 van de Plassche, 2001) and previous work at Hynes Brook demonstrated that salt-marsh  
128 sediment at this depth is ~3000 years old (Bell et al., 2005; Brookes et al., 1985; Daly, 2002;  
129 Daly et al., 2007).

130  
131 Placentia is located on the Avalon Peninsula (Figure 1E–F). It is indirectly connected to the  
132 ocean because the “South Arm” is closed by a gravel barrier, meaning that marine water  
133 currently reaches the site circuitously having entered through the open “North Arm”. This  
134 geomorphology results in a great diurnal tidal range of 0.91 m, compared to 1.60 m at the nearby  
135 (~8 km) Argentia tide gauge that is situated on the open coast. The modern salt marsh and  
136 sequences of Holocene organic sediment lie on top of a gravel barrier system (Forbes et al.,  
137 1995). The low, gravel barrier directly in front of the salt marsh is submerged at high tide. Low  
138 salt-marsh environments are absent at the site, although isolated stands of *Spartina alterniflora*  
139 (tall form) are present and rooted in gravel rather than fine-grained sediment. The high  
140 salt-marsh is a diverse, peat-forming community that includes *Distichlis spicata*, *Spartina*  
141 *patens*, *Juncus gerardii*, *Potentilla* sp., *Plantago maritima*, *Glaux maritima*, and *Carex glareosa*.

142 These plants are frequently intermixed in the high salt-marsh zone, although some mono-specific  
143 stands are present. The upland edge of the salt marsh is marked by a near-vertical slope of  
144 weathered bedrock that is sparsely vegetated. Daly (2002) described the sediment underlying the  
145 Placentia site and noted that there was typically less than ~0.5 m of peat except in one small area  
146 where up to ~1.3 m of peat was present (Figure 1H–J). Radiocarbon dates from Forbes et al.  
147 (1995) and Daly et al. (2007) indicate that the bottom of this peat sequence is ~2000 years old.  
148 Daly (2002) interpreted the stratigraphy at Placentia as including a unit of freshwater peat  
149 between two units of salt-marsh peat. Although the possible presence of freshwater peat makes  
150 the stratigraphy sub-optimal for reconstructing RSL because it would cause a temporal gap in  
151 any reconstruction, we chose to work at Placentia because of the scarcity of alternative sites in  
152 eastern Newfoundland and our goal of identifying regional-scale RSL trends was best met by  
153 having a spatial network of reconstructions available for analysis.

154

### 155 **3. METHODS**

#### 156 *3.1 Coring and sample elevations*

157 We used hand-driven cores positioned along transects to describe the sediment beneath the Big  
158 River and Placentia salt marshes (Figure 1). The positioning of these transects was guided by  
159 previous stratigraphic investigations of both sites (Daly et al., 2007; Wright and van de Plassche,  
160 2001). A core from each site was selected for detailed analysis based on the likelihood that it  
161 would produce a near-continuous reconstruction of RSL change. At Big River, core F25 was  
162 chosen because it is representative of the stratigraphy at the site and included the thickest unit of  
163 high salt-marsh that was not underlain by an amorphous organic unit (Figure 1G), which in our  
164 experience can have a relationship to sea level that is difficult to reconstruct (e.g., Kemp et al.,

165 2012b). Different criteria were employed at Placentia, where we selected core C106 as being the  
166 most likely to provide a long and near-continuous reconstruction because it was located in the  
167 relatively small area of the site where the underlying topography allowed a relatively thick  
168 sequence of sediment to accumulate (Figure 1H–J) and also did not (at least visibly) include the  
169 unit of amorphous (potentially freshwater) organic material. Therefore, C106 is representative of  
170 only of a small part of the stratigraphic record at Placentia. We recovered each core in  
171 overlapping, 50-cm long sections using an Eijelkamp Russian-type peat sampler. Each section  
172 was placed in a rigid plastic sleeve, wrapped in plastic, and kept refrigerated until processing in  
173 the laboratory to minimize desiccation and/or decay.

174

175 Core-top elevations were measured in the field by leveling each core site to a temporary  
176 benchmark using a total station. The tidal elevation of the temporary benchmarks was previously  
177 established by deploying water-level loggers at each site and comparing local tides to those at  
178 nearby tide gauge stations (Kemp et al., 2017c; Wright et al., 2011). We analyzed observed water  
179 levels from 1<sup>st</sup> January 1983 to 31<sup>st</sup> December 2001 from the Port-aux-Basques and Argentia tide  
180 gauges (Figure 1) and extracted high and low tides using the “*Tides*” package (Cox, 2016) for R  
181 (R Development Core Team, 2017). The water-level data was downloaded from the Canadian  
182 Hydrographic Service (Argentia: [http://www.isdm-gdsi.gc.ca/isdm-gdsi/twl-mne/inventory-](http://www.isdm-gdsi.gc.ca/isdm-gdsi/twl-mne/inventory-inventaire/sd-ds-eng.asp?no=835&user=isdm-gdsi&region=ATL&ref=maps-cartes)  
183 [inventaire/sd-ds-eng.asp?no=835&user=isdm-gdsi&region=ATL&ref=maps-cartes](http://www.isdm-gdsi.gc.ca/isdm-gdsi/twl-mne/inventory-inventaire/sd-ds-eng.asp?no=835&user=isdm-gdsi&region=ATL&ref=maps-cartes). Port-Aux-  
184 Basques: [http://www.isdm-gdsi.gc.ca/isdm-gdsi/twl-mne/inventory-inventaire/sd-ds-](http://www.isdm-gdsi.gc.ca/isdm-gdsi/twl-mne/inventory-inventaire/sd-ds-eng.asp?no=665&user=isdm-gdsi&region=ATL&ref=maps-cartes)  
185 [eng.asp?no=665&user=isdm-gdsi&region=ATL&ref=maps-cartes](http://www.isdm-gdsi.gc.ca/isdm-gdsi/twl-mne/inventory-inventaire/sd-ds-eng.asp?no=665&user=isdm-gdsi&region=ATL&ref=maps-cartes)). From this observation  
186 period, we calculated tidal datums following the definitions used by the National Oceanic and  
187 Atmospheric Administration (NOAA) for the 1983–2001 tidal epoch.



188

189 *3.2 Dating*

190 The cores from Big River and Placentia were dissected to isolate individual plant macrofossils  
191 found in growth position. Each macrofossil was identified through comparison with modern  
192 specimens collected in the field and published descriptions (e.g., Niering et al., 1977). From this  
193 suite of samples, we selected macrofossils for radiocarbon dating based on size, preservation  
194 state, and their distribution throughout each core. The selected samples were cleaned under a  
195 microscope to remove adhered sediment and ingrowing rootlets, after which they were dried at  
196 ~40 °C and submitted for dating. Samples from Big River were dated at the Utrecht van der  
197 Graaf Laboratorium and samples from Placentia were dated at the National Ocean Sciences  
198 Accelerator Mass Spectrometry (NOSAMS) facility. At both laboratories, each sample  
199 underwent acid-base-acid pre-treatment and conversion to a graphite target.  $\delta^{13}\text{C}$  was measured  
200 on an aliquot of  $\text{CO}_2$  collected during sample combustion. Radiocarbon results (Table 1) from  
201 Utrecht were originally reported without rounding, while those from NOSAMS were rounded  
202 following Stuiver and Polach (1977). To ensure comparability among sites, we rounded the  
203 reported radiocarbon ages and their uncertainties from the Utrecht laboratory prior to any further  
204 analysis following the conventions of Stuiver and Polach (1977). Three samples in the Placentia  
205 core were dated more than once as part of a quality control procedure at NOSAMS (Table 1). At  
206 each depth the replicate measurements were performed on the same individual plant macrofossil  
207 (as opposed to multiple samples from the same depth) and we therefore created a pooled mean  
208 for these depths using the Calib software (Stuiver et al., 2017) prior to any further analysis.

209

210 Due to a plateau in the calibration curve, radiocarbon dates on material younger than ~1600 CE  
211 often return multiple possible ages and large uncertainty. To develop a chronology with the  
212 desired precision for the past ~400 years, we recognized pollution horizons of known age in  
213 downcore profiles of elemental and isotopic abundance. <sup>137</sup>Cs activity was measured by gamma  
214 counting at Yale University using the methods and instrument described in Hill and Anisfeld  
215 (2015). We used 1-cm thick samples spaced at 2-cm intervals in the topmost parts of the Big  
216 River and Placentia cores to measure elemental abundance and lead-isotope ratios using mass  
217 spectrometry. The samples were oven dried at 40 °C and ground to a fine, homogenized powder  
218 that was analyzed by inductively coupled plasma mass spectrometry at a commercial laboratory  
219 (SGS Mineral Services).

220

221 We generated an age-depth model for each core using Bchron (Haslett and Parnell, 2008; Parnell  
222 et al., 2008). Since we obtained radiocarbon dates on rhizomes of salt-marsh plants, an  
223 uncertainty and depth correction was added to the Bchron input to reflect that these macrofossils  
224 likely grew at 2 cm ± 2 cm below the paleo marsh surface that we sought to date. Radiocarbon  
225 dates were calibrated individually by Bchron using the IntCal13 calibration curve (Reimer et al.,  
226 2013). Marker horizons identified in downcore profiles of elemental abundance, isotopic activity,  
227 and isotopic ratios were assumed to have a normally-distributed temporal uncertainty. We also  
228 assigned a thickness uncertainty to each horizon to reflect that the chronological marker could  
229 occur in more than one sample. From this input Bchron generated a large suite of  
230 equally-probable accumulation histories that were summarized by a probability density function  
231 to estimate the age of every 1-cm thick sample in the core with uncertainty. Throughout the text,  
232 chronological uncertainties from Bchron age-depth models are reported as the 95% credible

233 interval. We also used Bchron to estimate annual sediment accumulation with uncertainty  
234 (reported as the 90% credible interval; 5<sup>th</sup>–95<sup>th</sup> percentiles).

235

### 236 3.3 Foraminifera and Testate Amoebae

237 Every other 1-cm thick sample from each core was prepared for foraminiferal analysis by sieving  
238 the sediment under running water to retain the 63–500  $\mu\text{m}$  fraction, from which we counted a  
239 minimum of 100 foraminifera. If fewer than 100 tests were present, then the entire sample was  
240 counted. Taxonomy followed Wright et al. (2011) and Kemp et al. (2017c); most notably, counts  
241 of *Jadammina macrescens* and *Balticammina pseudomacrescens* were combined. Since these  
242 species often co-exist and have similar relationships to tidal elevation this decision is unlikely to  
243 affect the resulting RSL reconstruction, but does minimize the likelihood of bias being  
244 introduced by taxonomic inconsistencies among workers.

245

246 Following preliminary analysis of the Placentia core, we identified an interval (~60–16 cm;  
247 1330–1870 CE) with few or no foraminifera due to their absence or poor preservation. To avoid  
248 a gap in the RSL reconstruction from Placentia, we analyzed testate amoebae from this interval  
249 following Barnett et al. (2013; after Charman et al., 2000). A subsample of 1  $\text{cm}^3$  sediment was  
250 combined with a *Lycopodium clavatum* L. tablet (to calculate concentration; Stockmarr, 1971) in  
251 100 ml of deionized water and heated at 80 °C for one hour to aid disaggregation. The solution  
252 soaked for ~12 hours prior to wet sieving to retain the 15–300  $\mu\text{m}$  fraction. Retained sediment  
253 was mounted on glass slides with a water-glycerol solution for counting. We aimed to count 100  
254 tests per sample, but where test-concentration was low (<2000 tests/ $\text{cm}^3$ ), a minimum of 60 tests

255 were counted (e.g., Charman et al., 2010). Identifications followed the taxonomy of Barnett et al.  
256 (2017b), which was developed for intertidal salt-marsh testate amoebae of the North Atlantic.

257

### 258 *3.4 Compaction*

259 Salt-marsh sediments are prone to compaction (physical compression and biodegradation) that  
260 can cause post-depositional lowering of samples used to reconstruct RSL (Bloom, 1964; Brain,  
261 2016; Long et al., 2006). Post-depositional lowering results in an overestimation of the  
262 magnitude and rate of reconstructed RSL rise (e.g., Shennan and Horton, 2002). We estimated  
263 post-depositional lowering of samples in the Big River core using the geotechnical modelling  
264 approach of Brain et al. (2011, 2012, 2015). Post-depositional lowering was not estimated for the  
265 Placentia core because it is considerably shorter (and therefore less prone to compaction through  
266 physical compression) than the Big River core. As such, Big River likely represents an upper  
267 limit for the contribution from compaction to RSL reconstructions derived from unbroken  
268 sequences of high salt-marsh sediment in Newfoundland. To provide the empirical data that are  
269 necessary to calibrate the geotechnical model, we collected 11 undisturbed surface sediment  
270 samples from Big River, Hynes Brook, and Placentia, ensuring that samples were obtained from  
271 the range of eco-sedimentary zones encountered (Figure 1; Table 2). For each sample we  
272 measured (1) organic content by loss-on-ignition (LOI; three determinations per sample); (2)  
273 specific gravity ( $G_s$ ) using gas pycnometry (one determination per sample; Head, 2008); and (3)  
274 compression behaviour using automated oedometer testing (Head and Epps, 2011; Rees, 2014).  
275 From these measurements we established empirical relationships between LOI and specific  
276 gravity, and between LOI and the compression properties of modern salt-marsh sediments. For  
277 contiguous, 2-cm thick samples in the Big River core, we measured LOI and dry bulk density

278 and then used the empirical relationships established from modern, analogous sediments to  
279 estimate the compression properties of each sample. In each modelled layer in each model run,  
280 we assigned values of the compressive yield stress,  $\sigma'_y$ , by sampling from a continuous,  
281 triangular probability distribution, defined by the modal value and range of  $\sigma'_y$  observed in  
282 surface sediment (Table 2). We applied the geotechnical model using a Monte Carlo approach  
283 (5000 runs) to estimate depth-specific post-depositional lowering with uncertainty ( $\pm 1$  standard  
284 deviation). To assess the predictive capacity of the model, we compared measured and model-  
285 derived estimates of dry bulk density.

286

### 287 *3.5 Reconstructing relative sea level*

288 We reconstructed paleommarsh elevation (the tidal elevation at which a sample originally  
289 accumulated) using two independent Bayesian transfer functions. The foraminifera-based  
290 transfer function was calibrated using an empirical modern training set of 134 surface sediment  
291 samples collected along intertidal transects at Big River, Hynes Brook, and Placentia (Kemp et  
292 al., 2017c; Wright et al., 2011; Figure 1). Sample elevations were expressed as a standardized  
293 water level index because great diurnal tidal range varied among sites. Throughout the  
294 manuscript we use F-SWLI in reference to the standardized water level index used for  
295 foraminifera. A value of 100 equates to local mean tide level and a value of 200 is the highest  
296 occurrence of foraminifera at each site. Wright et al. (2011) proposed that using the highest  
297 occurrence of foraminifera as a datum resulted in more reliable paleommarsh elevation  
298 reconstructions than using a tidal datum such as MHHW. Foraminiferal abundance was  
299 expressed as raw counts and core samples that yielded fewer than 50 foraminifera were excluded  
300 from the final RSL reconstruction.

301  
302 The testate amoebae transfer function was calibrated on the North Atlantic dataset described in  
303 Barnett et al. (2017b) with additional data from Big River, Hynes Brook, and Placentia (Kemp et  
304 al., 2017c). The resulting dataset comprised 308 samples from 17 sites. Sample elevations were  
305 expressed as a standardized water level index, where zero is the lowest occurrence of testate  
306 amoebae and 100 is highest astronomical tide. Throughout the manuscript, we use T-SWLI in  
307 reference to the standardized water level index used for testate amoebae. The training set of  
308 testate amoebae was modified by condensing and reducing the number of taxa in the dataset to  
309 improve the Bayesian transfer function performance and to reduce computation time (which also  
310 depends on the number of samples). In the BTF it is more difficult to achieve convergence of  
311 model parameters if there are many rare species and/or if the number of species approaches the  
312 number of samples available compared to when there are few rare taxa and considerably more  
313 samples than species. Furthermore, computation time is reduced by analyzing fewer taxa,  
314 although even with all taxa the analysis is completed on a timescale of hours (for example,  
315 building the BTF for foraminifera took ~27 hours of computational time). Taxa and groups of  
316 taxa with invariant ecological tolerance ranges (established by weighted averaging regression of  
317 the species data) were combined to reduce the number of taxa and groups from 53 to 38 without  
318 reducing the number of samples in the modern dataset. Low occurrence taxa were removed  
319 following Legendre and Birks (2012). Taxa occurring in only one sample (then 2, 3, ...  $n$   
320 samples...) were removed from the dataset until taxa variation (canonical correspondence  
321 analysis axis 1 eigenvalue as a percentage of total inertia) explained by changes in elevation  
322 became negatively affected by the removal. This resulted in the removal of 16 low-occurring

323 taxa/groups that were encountered in 15 or fewer samples, with no loss in the amount of species  
324 variation explained by elevation.

325

326 The two Bayesian transfer functions were developed following the approach described in Cahill  
327 et al. (2016). This approach assumes a multinomial model for foraminiferal/testate amoebae  
328 assemblages, and uses a set of penalized spline smoothing functions (Lang and Brezger, 2004) to  
329 describe the non-linear relationship between each taxa and tidal elevation. The parameters that  
330 describe each taxon response curve were estimated empirically from the modern training set.

331 Performance of the Bayesian transfer functions was measured by 10-fold cross validation. When  
332 applied to fossil assemblages of foraminifera/testate amoebae preserved in the Big River and  
333 Placentia cores, the BTFs returned a paleommarsh elevation reconstruction (in F-SWLI/T-SWLI  
334 units) with a sample-specific 95% uncertainty interval. Paleommarsh elevations were converted to  
335 absolute tidal elevations using the modern relationship between local tidal datums at each site.

336 This approach therefore assumes a constant tidal range through time.

337

338 RSL was reconstructed by subtracting paleommarsh elevation estimated by the Bayesian transfer  
339 functions from sample elevation, where both values are expressed relative to the same tidal  
340 datum. Sample elevation was measured as depth in core, where the core-top elevation was  
341 measured in the field (as described in Kemp et al., 2017c) and this term is considered to be a  
342 fixed and known value. The corresponding age of each sample (with uncertainty) was provided  
343 by the Bchron age-depth model.

344

345 If an assemblage of foraminifera and/or testate amoebae preserved in core sediment lacks an  
346 appropriate modern analog, then the reconstructed paleomorph elevation may not be ecologically  
347 plausible (e.g., Jackson and Williams, 2004; Simpson, 2012). To quantify the degree of analogy,  
348 we measured dissimilarity between each core sample and its closest analog in the modern  
349 training set using the Bray-Curtis metric. If this value exceeded the 20<sup>th</sup> percentile of  
350 dissimilarity measured in all possible pairings of modern samples, then we determined that the  
351 core sampled lacked a modern analog and it was excluded from the final RSL reconstruction.  
352 The choice of the 20<sup>th</sup> percentile as a threshold for dissimilarity is common for populations such  
353 as foraminifera that form low-diversity assemblages, particularly in instances where the  
354 distribution of modern samples is uneven (i.e. concentration at particular elevations such as mean  
355 tide level to MHHW for foraminifera; Figure 2) since this causes dissimilarity measured among  
356 pairs of modern samples to be biased low (e.g., Kemp and Telford, 2015). For higher diversity  
357 populations, it may be appropriate to use a lower percentile of dissimilarity in determining the  
358 threshold for whether or not a core sample has a modern analog (Jackson and Williams, 2004;  
359 Watcham et al., 2013).

360

### 361 *3.6 Statistical analysis of relative sea-level reconstructions*

362 To self-consistently estimate (with uncertainty) differences in RSL among locations, rates of  
363 RSL change, and to distinguish among the global, regional linear, regional non-linear, and local  
364 components of RSL change, we employed a spatio-temporal empirical hierarchical model (e.g.,  
365 Kopp et al., 2016) . The input data for the model are the global database of RSL reconstructions  
366 in Kopp et al. (2016) after it was updated to include the Big River and Placentia reconstructions  
367 (this study; uncorrected for post-depositional lowering) and recently-published records (Figure



368 1A) from North Carolina (Kemp et al., 2017b), New York City (Kemp et al., 2017a), the Gulf  
369 Coast of Florida (Gerlach et al., 2017), and the Magdalen Islands (Barnett et al., 2017a).  
370 Tide-gauge records were updated from the Permanent Service for Mean Sea Level (Holgate et  
371 al., 2013) to include data through 2017. As in Kopp et al. (2016), the input data also include the  
372 global mean sea-level reconstruction of Hay et al. (2015), that was generated from tide-gauge  
373 records. Posterior estimates generated by the spatio-temporal model for sites in eastern North  
374 America are provided in the supporting information.

375  
376 The model has (1) a process level that characterizes the behavior of RSL over space and time, (2)  
377 a data level that links RSL observations (reconstructions) to the RSL process, and (3) a  
378 hyperparameter level that characterizes prior expectations regarding dominant spatial and  
379 temporal scales of RSL variability.

380  
381 At the process level, we model the RSL field  $f(\mathbf{x}, t)$  as the sum of seven components:

$$382 \quad f(\mathbf{x}, t) = g_f(t) + g_s(t) + m(\mathbf{x})(t - t_0) + r_s(\mathbf{x}, t) + r_f(\mathbf{x}, t) + l_s(\mathbf{x}, t) + l_f(\mathbf{x}, t) \quad (1)$$

383 where  $\mathbf{x}$  represents geographic location,  $t$  represents time, and  $t_0$  is a reference time point (2000  
384 CE). The seven components are: fast and slow common (global) terms,  $g_f(t)$  and  $g_s(t)$ ,  
385 representing the time-variable signal shared by all sites that are included in the analysis; a  
386 regionally varying, temporally linear (on the timescale under examination) term,  $m(\mathbf{x})(t - t_0)$ ,  
387 which represents slowly changing processes such as GIA and/or tectonics (Engelhart et al.,  
388 2009); regionally varying, fast and slow temporally non-linear fields,  $r_f(\mathbf{x}, t)$  and  $r_s(\mathbf{x}, t)$ , which  
389 primarily represent factors such as dynamic sea level and the fingerprints of land-ice melt, and  
390 fast and slow local fields,  $l_f(\mathbf{x}, t)$  and  $l_s(\mathbf{x}, t)$ , which represent site-specific processes such as

391 sediment compaction. As in Kopp et al. (2016), the common global signal is assumed to average  
392 0 mm/yr over 0–1700 CE.

393

394 The data level of the spatio-temporal model represents the inherently noisy RSL reconstructions  
395 (comprised of multiple data points termed observations;  $y_i$ ) as:

$$396 \quad y_i = f(\mathbf{x}_i, t_i) + \varepsilon_i + w(\mathbf{x}_i, t_i) + y_0(\mathbf{x}_i) \quad (2)$$

$$397 \quad t_i = \hat{t}_i + \delta_i \quad (3)$$

398 where  $f(\mathbf{x}_i, t_i)$  is the true RSL value at location  $\mathbf{x}$  and time  $t$ ;  $\varepsilon_i$  is the vertical uncertainty of the  
399 RSL observation  $i$ ;  $w(\mathbf{x}_i, t_i)$  is supplemental white noise added to capture incompleteness of the  
400 process model and under-estimation of vertical and temporal measurement error; and  $y_0(\mathbf{x}_i)$  is a  
401 site-specific vertical datum correction that is applied to ensure that the independent RSL  
402 reconstructions are directly comparable to one another. The true age of a RSL observation ( $t_i$ )  
403 includes the mean estimate of the true age ( $\hat{t}_i$ ) and its chronological error ( $\delta_i$ ), which is  
404 approximated using the noisy-input Gaussian process method (McHutchon and Rasmussen,  
405 2011).

406

407 The priors for each term in the model are Gaussian processes (Rasmussen and Williams, 2005)  
408 with Matérn covariance functions (see Kopp et al., 2016 for more details). All priors have zero  
409 mean except for the regional-scale linear component,  $m(\mathbf{x})$ , which uses predictions from the  
410 ICE5G–VM2–90 Earth-ice model (Peltier, 2004) as prior means. This does not significantly  
411 affect the final (posterior) estimate of the regional-scale linear component, which the model  
412 estimates based on the observations. Hyperparameters defining prior expectations for the  
413 amplitudes and spatio-temporal scales of variability were set through a maximum-likelihood

414 optimization. The non-linear terms were characterized by three spatial scales (global, regional,  
415 and local) and two temporal scales (fast and slow; Table 3). These different spatial and temporal  
416 scales enable RSL to be decomposed and can be thought of as akin to filters of different spatial  
417 and temporal bandwidths. Optimized values indicate that the dominant temporal scales are 326  
418 years for the slow terms and 12 years for the fast terms, the dominant regional spatial scale is  
419  $6.1^\circ$  angular distance, and the local spatial scale is  $0.05^\circ$ . The largest non-linear signal comes  
420 from the slow, global term (prior standard deviation of 16 cm), followed by the slow, regional  
421 term (prior standard deviation of 6 cm). The fast, regional term and the two local terms make  
422 comparable contributions (prior standard deviations of 2–3 cm). The supplemental white noise  
423 and datum correction terms are negligible, which indicates that the stated measurement errors are  
424 adequate to explain the differences between the process model and the observations.

425

## 426 **4. RESULTS**

### 427 *4.1 Performance of Bayesian transfer functions*

428 The Bayesian transfer function developed for foraminifera utilized a modern training set (134  
429 samples from three sites in Newfoundland) that spanned an elevational range of 84–200 F-SWLI.  
430 Under 10-fold cross validation there is a strong ( $r^2=0.81$ ) relationship between actual (measured  
431 at the time of sample collection) and predicted elevation (Figure 2). Within their uncertainty, the  
432 predicted elevation of 96% (128 of 134) of samples included the actual elevation. Absolute  
433 residuals (difference between actual and predicted elevation) averaged 3.3 F-SWLI, with a  
434 standard deviation of 3.2 F-SWLI, and a maximum of 46.8 F-SWLI. The lack of any systematic  
435 relationship between residuals and actual elevation ( $r^2=0.05$ ) indicates that the foraminiferal  
436 Bayesian transfer function should produce unbiased reconstructions of paleomarch elevation.

437

438 The Bayesian transfer function developed for testate amoebae utilized a modern training set (308  
439 samples from sites in the North Atlantic Ocean, including three sites in Newfoundland) that  
440 spanned an elevational range of 0–435 T-SWLI. Under 10-fold cross validation there is a strong  
441 ( $r^2=0.67$ ) relationship between actual and predicted elevation (Figure 2). Within their  
442 uncertainty, the predicted elevation of 80% (247 of 308) of samples included the actual  
443 elevation. Of the 61 samples for which the Bayesian transfer function did not accurately predict  
444 elevation, a disproportionate number came from Sillons (14 samples) in the Magdalen Islands  
445 (Canada) and Scheldt Estuary (16 samples) in Belgium, which may indicate local-scale  
446 variability in the distribution of testate amoebae. Absolute residuals averaged 33 T-SWLI, with a  
447 standard deviation of 29 T-SWLI, and a maximum of 157 T-SWLI. The lack of any systematic  
448 relationship between residuals and actual elevation ( $r^2=0.001$ ) indicates that the testate amoebae  
449 Bayesian transfer function should produce unbiased reconstructions of paleomarch elevation.

450

#### 451 *4.2 Relative sea level at Big River*

452 The stratigraphy beneath Big River is comprised primarily of high salt-marsh peat with abundant  
453 plant remains overlying an incompressible substrate of consolidated, grey-blue silt and clay of  
454 glacial origin (Figure 1G). Within the high salt-marsh peat are thin (< 5 cm) and discontinuous  
455 horizons containing increased fine-grained clastic sediment. The discontinuous nature of these  
456 units likely indicates that they are the result of a process operating intermittently and on  
457 meter-scales, such as delivery of ice-rafted sediment (Richard, 1978). Pools (diameters on the  
458 order of a few meters and lifespans of years) are also a common feature of salt marshes in the

459 study region and are associated with deposition of fine-grained minerogenic sediment that could  
460 leave thin clastic layers in the sedimentary record (e.g., Wilson et al., 2014; Wilson et al., 2010).  
461 A total of 42 radiocarbon ages obtained on plant macrofossils in core F25 (Table 1; Figure 3)  
462 reveal that the uppermost 3.05 m of sediment accumulated during the past ~3000 years.

463

464 We assigned the lowest depth in the Big River core with detectable  $^{137}\text{Cs}$  activity ( $21 \pm 2$  cm) an  
465 age of  $1954 \text{ CE} \pm 2$  years and the depth with maximum activity ( $13 \pm 2$  cm) was recognized as  
466  $1963 \text{ CE} \pm 1$  year (Figure 3A). These horizons represent the initiation and peak of above-ground  
467 testing of nuclear weapons respectively (see discussion in Wright et al., 1999 for example).

468 Prevailing westerly winds carried pollution from the major industrial centers in the United States  
469 (including the Upper Mississippi Valley and northeastern states) to Newfoundland and we  
470 therefore used historic emissions from the United States to assign ages to downcore trends in  
471 elemental and isotopic abundance (Figure 3A). Our interpretation relies on trends rather than  
472 absolute values since emissions per unit of production likely changed through time. We assume  
473 that pollution was deposited principally onto the high salt-marsh surface from the atmosphere,  
474 shortly after emission and without further isotopic fractionation. The ratio of lead isotopes  
475 ( $^{206}\text{Pb}:^{207}\text{Pb}$ ) measured in bulk core sediment reflects the evolving sources of lead pollution that  
476 began in North America with processing of galena ore with a distinctive isotopic signature in the  
477 Upper Mississippi Valley (Doe and Delevaux, 1972; Heyl et al., 1966; Kemp et al., 2012a; Lima  
478 et al., 2005). The initial increase in  $^{206}\text{Pb}:^{207}\text{Pb}$  ( $35 \pm 2$  cm) is interpreted as reflecting the start of  
479 emissions from the Upper Mississippi Valley and correspondingly assigned an age of  $1827 \text{ CE} \pm$   
480  $5$  years. A peak in  $^{206}\text{Pb}:^{207}\text{Pb}$  ( $29 \pm 2$  cm) occurred in  $1858 \pm 5$  years when the quantity of lead  
481 produced in the Upper Mississippi Valley reached its maximum proportion of U.S. national

482 output. The decline of lead production in the Upper Mississippi Valley coupled with growing  
483 production elsewhere from sources with less distinctive isotopic ratios caused  $^{206}\text{Pb}$ : $^{207}\text{Pb}$  to  
484 reach a minimum in 1880 CE  $\pm$  20 years ( $25 \pm 2$  cm). The onset of widespread industrialization  
485 at 1875 CE  $\pm$  5 years caused a corresponding increase in the quantity of lead deposited at Big  
486 River ( $25 \pm 2$  cm), while the introduction of the Clean Air Act reduced lead pollution (1974 CE  
487  $\pm$  5 years) at Big River ( $13 \pm 2$  cm). We attribute the second rise in lead (at depths above 13 cm  
488 in the Big River core) to local trends of unknown provenance since similar features are not  
489 observed elsewhere in the northeastern United States or eastern Canada (e.g., Chillrud et al.,  
490 1999; Gobeil et al., 2013; Graney et al., 1995; Kemp et al., 2015). Assimilation of the  
491 radiocarbon dates and pollution markers into a Bchron age-depth model allowed us to  
492 reconstruct the history of sediment accumulation at Big River (Figure 3B). The chronological  
493 uncertainty for a 1-cm thick sample in this core (measured by the mean width of the 95%  
494 credible interval) was  $\pm$  53 years. Prior to  $\sim$ 1850 CE the mean annual rate of sediment  
495 accumulation was 0.65 mm/yr (95% credible interval of 0.20–2.8 mm/yr), after which the rate  
496 increased to 0.9 mm/yr (95% credible interval of 0.28–3.8 mm/yr), including a current rate of  
497  $\sim$ 2.0 mm/yr (95% credible interval of 0.6–10.0 mm/yr; Figure 3C).

498

499 Foraminifera were counted in 151, 1-cm thick samples from the Big River core. In these  
500 samples, *Jadammina macrescens/Balticammina pseudomacrescens* was the single most abundant  
501 taxon at most depths (60% of all counted individuals; Figure 4). Foraminifera were present at  
502 depths below 288 cm, but these samples yielded fewer than 50 individual tests (almost  
503 exclusively *Jadammina macrescens/Balticammina pseudomacrescens*). From 288 cm to 222 cm,  
504 there was elevated relative abundance of *Miliammina fusca* (average 38%). Similarly, the

505 intervals at depths of 178–164 cm and 126–124 cm were characterized by assemblages that  
506 included more than 25% *Miliammina fusca*. At 138–38 cm assemblages of foraminifera included  
507 an average of 20% *Haplophragmoides* spp. *Tiphotrecha comprimata* was present throughout the  
508 core, but most abundant (average 27%) above 56 cm. Only one sample (at 40 cm) lacked an  
509 appropriate analog in the modern training set (at the 20<sup>th</sup> percentile of dissimilarity) and two  
510 samples above 288 cm contained fewer than 50 foraminifera (Figure 4). These samples (along  
511 with the seven samples below 288 cm with low counts) were excluded from the final RSL  
512 reconstruction. When the threshold for dissimilarity was lowered to the 10<sup>th</sup>/5<sup>th</sup> percentile, the  
513 number of foraminiferal assemblages lacking a modern analogue was 10/85 out of a total of 151  
514 samples. Application of the Bayesian transfer function to assemblages of foraminifera preserved  
515 in the Big River core generated paleommarsh elevation reconstructions with sample-specific  
516 uncertainties. For samples with <20% *Miliammina fusca*, the average reconstructed paleommarsh  
517 elevation was 162 F-SWLI. In comparison, samples with >20% *Miliammina fusca* produced  
518 paleommarsh elevation reconstructions with an average of 133 F-SWLI. The average uncertainty  
519 in paleommarsh elevation reconstructions (95% credible interval) was  $\pm 25$  F-SWLI, which is  $\pm$   
520 0.23 m under our assumption of a stationary tidal regime at Big River.

521

522 At Big River, RSL rose continuously from approximately  $-2.64$  m at  $-800$  CE (Figure 5). The  
523 rate of RSL rise estimated by the spatio-temporal model was  $0.60 \pm 0.15$  mm/yr (95% credible  
524 interval) until 0 CE. Over the period 0–1700 CE, RSL rose at  $0.98 \pm 0.05$  (95% credible interval)  
525 and during the 20<sup>th</sup> century the rate of rise was  $2.05 \pm 0.61$  mm/yr (95 credible interval). In  
526 comparison, the tide gauge at Port-aux-Basques recorded a linear rate of rise of  $\sim 2.2$  mm/yr over  
527 the period 1959–1999 CE. Therefore, within uncertainty the tide gauge and RSL reconstruction

528 record the same rate of historic rise. The data used to reconstruct RSL at Big River are tabulated  
529 in the supporting information.

530

#### 531 *4.3 Relative sea level at Placentia*

532 The stratigraphy underlying the Placentia site is comprised of up to ~1.75 m of organic sediment  
533 overlying basal sand (Figure 1 H–J). A unit of red-hued, highest salt-marsh peat lies above the  
534 basal sand. At most coring locations, this unit was replaced upcore by an amorphous, black  
535 organic unit that outcrops at the surface in a few locations and which Daly (2002) interpreted as  
536 a detrital, freshwater peat. A unit of brown-coloured, high salt-marsh peat is present at most  
537 locations on the marsh surface. We selected core C106 for detailed analysis because it included  
538 the thickest highest salt-marsh unit and the freshwater peat appeared to be absent. Visual  
539 inspection of cores in the field and of C106 in the laboratory did not reveal any clear evidence  
540 for a hiatus or disturbance events. Radiocarbon dates from 19 depths in the core (including three  
541 depths where pooled means were generated from replicate radiocarbon measurements; Table 1)  
542 reveal that this sediment accumulated during the past ~2750 years (Figure 6B). Our  
543 interpretation of isotopic and elemental profiles in the upper ~0.3 m of the core followed our  
544 interpretation of similar measurements on the Big River core (Figure 3A). In addition to the  
545 chronological horizons recognized at Big River, we identified the onset of copper pollution  
546 (1900 CE  $\pm$  10 years) at 13  $\pm$  4 cm in the Placentia core (Figure 6A), alongside a peak in lead  
547 deposition following World War One (1925 CE  $\pm$  5 years at 11  $\pm$  4 cm), a reduction in lead  
548 deposition during the Great Depression (1935 CE  $\pm$  6 years at 9  $\pm$  6 cm), a minimum in the  
549  $^{206}\text{Pb}:^{207}\text{Pb}$  ratio caused by the changing source of lead being added to vehicle fuel (1965 CE  $\pm$  5  
550 years at 8  $\pm$  2 cm) and a peak in vanadium concentration (1974  $\pm$  5 years at 5  $\pm$  4 cm) that we



551 interpreted as confirmation that peak pollution was curtailed by the introduction of the Clean Air  
552 Act in the United States. Assimilation of these dates using Bchron, generated an age-depth model  
553 with an average uncertainty (95% credible interval) of  $\pm 75$  years for 1-cm thick samples,  
554 although we note that below depths of 137 cm, the uncertainty is greater (Figure 6B). The history  
555 of sedimentation at Placentia includes distinctive pulses of sedimentation and a prolonged period  
556 of no/slow deposition (Figure 6C). Faster sedimentation occurred at approximately -500 CE to -  
557 250 CE (average 0.58 mm/yr; 0.17–2.4 mm/yr, 95% credible interval) and at ~1300 CE to ~1500  
558 CE (average 0.94 mm/yr; 0.24–4.2 mm/yr, 95% credible interval), when the mean annual rate  
559 reached a maximum of 3.3 mm/yr. From 100 CE to 1000 CE the average mean annual rate of  
560 sedimentation was positive, but slow (average 0.11 mm/yr; 0.03–0.39 mm/yr, 95% credible  
561 interval).

562

563 Assemblages of foraminifera in the Placentia core were characterized at all depths by high  
564 abundances of *Jadammina macrescens*/*Balticammina pseudomacrescens* (minimum of 81% in  
565 all samples that yielded representative counts below 4 cm), although in the uppermost 4 cm,  
566 *Trochammina inflata* comprised ~50% of the assemblage (Figure 7). Foraminifera were scarce  
567 (1–16 tests in 1-cm thick samples), or absent from 60 cm to 16 cm. Application of the Bayesian  
568 transfer function to these assemblages indicates that the sediment above 16 cm and below 60 cm  
569 was deposited in the uppermost part of the intertidal zone close to highest astronomical tide and  
570 the highest occurrence of foraminifera. The average (95% credible interval) uncertainty for  
571 reconstructed paleommarsh elevation was  $\pm 15$  F-SWLI (equivalent to  $\pm 0.12$  m at Placentia). All  
572 assemblages of foraminifera in the core had analogues in the modern training set when the 20<sup>th</sup>  
573 percentile of dissimilarity was employed as a threshold. When the threshold was lowered to the

574 10<sup>th</sup>/5<sup>th</sup> percentile, the number of foraminiferal assemblages lacking a modern analogue was 0/6  
575 out of a total of 61.

576

577 The top 60 cm of the Placentia core was analysed for testate amoebae because of the scarcity or  
578 absence of foraminifera (Figure 7). Testate amoebae assemblages were typical of salt-marsh  
579 environments around the North Atlantic Ocean, including Newfoundland (Figure 7). From ~60  
580 cm to ~20 cm the most common taxa were *Centropyxis aculeata* type (15–39%), *Centropyxis*  
581 *cassis* type (14–47%), and *Tracheleuglypha dentata* type (up to 17%). Above ~20 cm, the  
582 abundance of *Pseudocorythion* type increased (up to 68%), while there was a corresponding  
583 reduction in the abundance of *Centropyxis aculeata* type (less than 38%) and *Centropyxis cassis*  
584 type (less than 36%). Application of the Bayesian transfer function to these assemblages  
585 generated paleommarsh elevation reconstructions that indicate deposition in a high salt-marsh  
586 environment below highest astronomical tide (and notably below the highest occurrence of  
587 foraminifera in the modern training set; Figure 7). The average uncertainty (95% credible  
588 interval) for the paleommarsh elevation reconstructions was  $\pm 65$  T-SWLI (equivalent to  $\pm 0.28$  m  
589 at Placentia), although we note that two samples (at 28 cm and 24 cm) had particularly large  
590 uncertainties of  $\pm 172$  T-SWLI and  $\pm 177$  T-SWLI respectively. All assemblages of testate  
591 amoebae had an analogue in the modern training set from the North Atlantic Ocean when the  
592 20<sup>th</sup> percentile was used as a cut-off value. Using the 10<sup>th</sup>/5<sup>th</sup> percentile instead, resulted in 1/20  
593 samples lacking a modern analogue.

594

595 At Placentia, paleommarsh elevation exhibited little variability until ~1970 CE (~6 cm depth;  
596 Figure 7). Therefore, reconstructed RSL trends prior to ~1970 CE arise primarily from changes  
597 in the rate of sediment accumulation. The reconstructed mean annual sedimentation rate (Figure  
598 6C) only exceeded 1 mm/yr in 137 years out of a total record length of more than 2500 years  
599 (mean annual sedimentation exceeding 2/3/4 mm/yr occurred in 18/5/0 years). These rates of  
600 accumulation and the corresponding rates of time-averaged RSL rise are well below proposed  
601 upper limits of salt-marsh sedimentation (e.g, Kirwan et al., 2010; Kirwan et al., 2016; Morris et  
602 al., 2002), indicating that the consistency of reconstructed paleommarsh elevation is ecologically  
603 plausible. Importantly, the switch to using testate amoebae as sea-level indicators in the absence  
604 of foraminifera did not cause a marked change in paleommarsh elevation (Figure 7). Foraminifera  
605 below 60 cm indicated that paleommarsh elevation was above local MHHW and below highest  
606 astronomical tide, while testate amoebae from 60 cm to ~16 cm also show paleommarsh elevation  
607 in this range.

608

609 Placentia offers an opportunity to consider similarities and differences between  
610 paleoenvironmental inferences drawn from foraminifera and testate amoebae. At depths of 16–60  
611 cm, 13 samples contained no foraminifera and 10 yielded 1–16 tests. We consider three possible  
612 explanations for the absence/scarcity of foraminifera in this interval. Firstly, the rate of sediment  
613 accumulation was sufficiently high to dilute the concentration of foraminifera and produce low  
614 counts. This seems unlikely, particularly in light of age-depth results, which indicate that the rate  
615 of sediment accumulation was not unusual for a salt-marsh experiencing RSL rise (Figure 6C).  
616 Other studies, show that foraminifera are abundant in salt-marsh sediment accumulating at these  
617 rates (e.g., Kemp et al., 2013). Secondly, this section of core was deposited at, or above, the

618 highest occurrence of foraminifera and therefore no or few tests were ever present. The  
619 absence/scarcity of foraminifera does coincide with faster sediment accumulation (Figure 6) that  
620 was sufficiently high perhaps to cause emergence. However, results from the testate amoebae  
621 transfer function (Figure 7) indicate that the sediment accumulated close to MHHW, in which  
622 case it is reasonable to assume that a healthy population of agglutinated foraminifera existed on  
623 the paleo salt-marsh surfaces if the paleoenvironmental interpretation of testate amoebae is  
624 accurate. Furthermore, core stratigraphy does not indicate a change in depositional environment.  
625 Cross validation of the testate amoebae Bayesian transfer function (Figure 2) does not indicate  
626 that is likely to generate systematically biased paleommarsh elevation reconstructions. Thirdly,  
627 foraminifera were present on the paleommarsh surfaces, but poorly preserved (hence only a  
628 handful of tests being found in core samples). Other studies (e.g., Kemp et al., 2014; Kemp et al.,  
629 2013) also noted the absence/scarcity of foraminifera in sections of cores where supporting  
630 evidence such as plant macrofossils or bulk-sediment geochemistry strongly suggest deposition  
631 in a high-salt marsh environment that probably supported agglutinated foraminifera. In the upper  
632 6 cm of the Placentia core, testate amoebae recorded an increase in paleommarsh elevation  
633 (emergence), while an increase in *Trochammina inflata* drove a modest decrease in paleommarsh  
634 elevation (drowning) estimated using foraminifera. Although these trends are contradictory, they  
635 are not significant within the reconstruction uncertainties.

636

637 At Placentia, RSL rose continuously from  $-1.40$  m at  $-760$  CE. The rate of RSL rise estimated  
638 by the spatio-temporal model was  $0.48 \pm 0.06$  (95% credible interval) over the period 0–1700  
639 CE, but during this interval the rate of RSL included phases of slower ( $0.12 \pm 0.27$  for 400–800  
640 CE; 95% credible interval) and faster rise ( $1.07 \pm 0.23$  for 1200–1600 CE; 95% credible

641 interval). During the 20<sup>th</sup> century the rate of rise was  $1.43 \pm 0.62$  mm/yr (95% credible interval).  
642 For comparison, the Argentia tide gauge measured a linear rate of RSL rise of 1.33 mm/yr  
643 between 1971 CE (first year of measurements) to 2000 CE. Therefore, within uncertainty the tide  
644 gauge and RSL reconstruction record the same rate of historic rise (albeit over a limited time  
645 interval). It is important to recognize that the short duration (29 years) of this instrumental record  
646 may not be suitable for accurately representing sustained RSL trends because of contamination  
647 by decadal-scale variability (e.g., Douglas, 1997). Furthermore, a 1-cm thick sample of sediment  
648 from the uppermost part of the Placentia core represents ~20 years, which inhibits efforts to  
649 reliably estimate RSL trends in the reconstruction on the same timescale as the tide gauge. The  
650 data used to reconstruct RSL at Placentia are tabulated in the supporting information.

651

#### 652 *4.4 Post-depositional lowering at Big River*

653 The physical and geotechnical properties of modern salt-marsh sediment were measured on 11  
654 surface samples from Big River, Hynes Brook, and Placentia (Figure 8A–D; Table 2). This data  
655 set provided empirical constraints on the relationship between organic content (measured by  
656 LOI) and the parameters (voids ratio, recompression index, compression index, and specific  
657 gravity) needed to run the decompaction model. High organic content is associated with  
658 correspondingly high voids ratios, recompression index, and compression index, but lower  
659 specific gravity. The nature of these relationships is similar to those observed on salt-marshes in  
660 North Carolina (Brain et al., 2015), Connecticut (Brain et al., 2017), and the United Kingdom  
661 (Brain et al., 2011; Brain et al., 2012) despite marked difference among these regions in climate,  
662 salt-marsh vegetation, coastal geomorphology, and organic content of salt-marsh sediment.

664 In the Big River core, LOI varied from 13% to 70% (Figure 8F) and these values fall within the  
665 range measured on modern salt-marsh sediment in Newfoundland (12–82%). Samples with low  
666 LOI were typically comprised of thin and localized silt lenses. Measured dry bulk density was  
667 0.10–0.47 g/cm<sup>3</sup>, where the highest values occurred in the silt lenses (Figure 8G). Comparison of  
668 these values with those predicted by the decompaction model indicates a tendency for the model  
669 to modestly over predict the dry bulk density of the silt lenses (estimates of post-depositional  
670 lowering may subsequently be overestimated) in underlying sediments. However, given the low  
671 density and thin nature of these discontinuous lenses, we do not consider this effect to be  
672 significant in our post-depositional lowering estimates throughout the core. For all samples in the  
673 core, the relationship between measured and predicted dry density has an  $r^2$  value of 0.52, and  
674 88% of measured values lie within the  $2\sigma$  predictions of the model (Figure 8E). Application of  
675 the geotechnical model to the Big River core shows that effective stress increases with depth  
676 from zero at the surface to ~3 kPa at 3.20 m, which yields estimates of post-depositional  
677 lowering that increase from zero at the top of the core to a characteristic peak in the middle of  
678 the core (less than 0.02 m), before decreasing once again to zero at the base of the core (Figure  
679 8I). The magnitude of estimated post-depositional lowering through mechanical compression is  
680 similar to estimates generated elsewhere by applying geotechnical models to near-continuous  
681 sequences of high salt-marsh peat (Brain et al., 2015; Zoccarato and Teatini, 2017). Comparison  
682 of the Big River RSL reconstruction with a regional-scale database of basal (compaction-free)  
683 sea-level index points compiled by Love et al. (2016; Figure 5A) further suggests that the  
684 contribution of compaction was small. Based on these results, we contend that post-depositional  
685 lowering through physical compression is unlikely to have exceeded 0.02 m in the samples used

686 to reconstruct RSL at Placentia because the thicker sequence of peat at Big River is more prone  
687 to this effect than the thinner sequence (~1.4 m) at Placentia. Compaction data are tabulated in  
688 the supporting information for Big River.

689

## 690 **5. DISCUSSION**

### 691 *5.1 Decomposition of relative sea-level trends*

692 Although any single RSL reconstruction is necessarily and unavoidably representative of trends  
693 at a specific location, regional- and global-scale trends can be investigated by analyzing a  
694 multi-site network of reconstructions. We used the spatio-temporal empirical hierarchical model  
695 (section 3.6) to decompose a global dataset of RSL reconstructions and co-located tide-gauge  
696 measurements into contributions from processes characterized by different spatial and temporal  
697 scales. The spatio-temporal model simultaneously analyzes each reconstruction and shares  
698 information among sites to generate a posterior RSL history for each site. In agreement with the  
699 original reconstructions, RSL estimated by the spatio-temporal model rose by ~3.0 m at Big  
700 River over the past ~3000 years compared to ~1.5 m at Placentia (Figure 9). The  
701 model-estimated RSL curves fall within the uncertainties of the original RSL reconstruction,  
702 indicating that the underlying data is accurately represented by the model. Implicitly, the  
703 spatio-temporal model assumes that each of the datasets provided as input are accurate  
704 reconstructions of RSL. Therefore, variability within and among reconstructions is considered to  
705 primarily be the result of physical processes rather than the choices (e.g., definitions of  
706 standardized water level indexes, type of transfer function, and type of age-depth model) that  
707 were made to produce the reconstructions.

708

709 At Big River and Placentia, most RSL rise during the past ~3000 years is attributed to  
710 regional-scale linear processes (Figure 9) that we interpret primarily as GIA (e.g., Love et al.,  
711 2016; Peltier, 2004) at a rate of  $0.98 \pm 0.05$  mm/yr (95% credible interval) at Big River and  $0.51$   
712  $\pm 0.05$  mm/yr (95% credible interval) at Placentia. The rate of reconstructed, regional-scale  
713 linear sea-level rise at Big River and Placentia is typically less than the rate inferred from RSL  
714 predictions generated by the widely-used ICE-6G VM5 (Peltier et al., 2015) and ICE-5G VM2  
715 (Peltier, 2004) Earth-ice models (Figure 5). Similar differences were noted in studies that  
716 compared sea-level index points derived from salt-marsh sediment to RSL predictions from  
717 Earth-ice models (e.g., Roy and Peltier, 2015).

718

719 The global-scale component of reconstructed RSL is largely consistent with the results of Kopp  
720 et al. (2016). It varies by ~0.2 m and includes a gradual rise over 0–600 CE that was followed by  
721 a marked fall from 1000 CE to 1200 CE (Figures 9 and 10A). Over the 20<sup>th</sup> century, global sea  
722 level rose by  $13.6 \pm 1.4$  mm/yr (95% credible interval), in agreement with global tide-gauge  
723 compilations. However, this agreement is to be expected because the input data analyzed by the  
724 spatio-temporal model included the global tide gauge record of Hay et al. (2015). This is the  
725 fastest centennial-scale rate of rise in the past 3000 years ( $P > 0.999$ ), which is a stronger  
726 conclusion than the equivalent analysis of Kopp et al. (2016), who found that the 20<sup>th</sup> century  
727 rise was extremely likely ( $P = 0.95$ ) faster than during any century since at least –800 CE.

728

## 729 *5.2 Local-scale contribution*

730 The estimated contribution from local-scale processes to RSL trends along the Atlantic coast of  
731 North America is typically less than  $\pm 5$  cm (Figure 10B). This interpretation is also partly



732 supported by geotechnical modeling, which showed that post-depositional lowering through  
733 sediment compaction (a commonly invoked local-scale process) was minimal at Big River  
734 (Figure 8) and probably also at Placentia. A comparison of basal (compaction-free) sea-level  
735 index points from western Newfoundland (Love et al., 2016) with the Big River RSL  
736 reconstruction confirms that sediment compaction was not a major driver of reconstructed RSL  
737 change in Newfoundland (Figure 5). Work in Connecticut and North Carolina also concluded  
738 that compaction made a minimal contribution to RSL reconstructions generated from continuous  
739 sequences of high salt-marsh peat (Brain et al., 2017; Brain et al., 2015; Kemp et al., 2015;  
740 Zoccarato and Teatini, 2017).

741

742 Salt marshes on the open coast with simple geomorphologies and that continuously accumulated  
743 sediment in (near-)equilibrium with rising sea-level are ideal sites for reconstructing RSL. In  
744 practice sediment accumulation and preservation often favors sheltered, low-energy salt marshes.  
745 This is the case in eastern Newfoundland (e.g., Placentia), where salt marshes are rare and  
746 restricted to back-barrier settings. Placentia's geomorphology makes it potentially vulnerable to  
747 local-scale RSL variability (Figure 1E). For example, closure of the North Arm would isolate the  
748 site from astronomical tides and RSL trends in the ocean, while opening of the South Arm could  
749 increase marine influence and tidal range. However, the estimated contribution from local-scale  
750 processes to RSL at Placentia is not markedly larger than at Big River (Figure 9), or indeed other  
751 sites along the Atlantic coast of North America (Figure 10B). This result does not mean that  
752 local-scale processes were unimportant, only that they are not necessary to explain the  
753 observations (i.e. the original RSL reconstruction) within their uncertainty. For example, at  
754 Placentia the period ~1500–1800 CE (Figure 9) is characterized by variable and uncertain

755 reconstructions that could readily accommodate additional local-scale contributions with a  
756 magnitude of centimeters to tens of centimeters and a duration of decades to centuries. Such  
757 large values do not appear within the uncertainty envelope of the modeled, local-scale estimate at  
758 Placentia because there is no site in the database that requires such large local-scale contributions  
759 to explain the reconstructions within their uncertainty, and so the trained model does not find  
760 them parsimonious.

761

762 The Atlantic coast of North America has the greatest concentration of detailed RSL  
763 reconstructions spanning the past ~3000 years of any coastline in the world, but even here the  
764 network of available data remains sparse with individual records often separated by 100s of  
765 kilometers (Figure 1A). Due to the tendency for research to explore RSL variability among  
766 rather than within coastal regions, a single region (i.e., a geographic area with a shared climate  
767 and oceanographic regime) may have only a handful of reconstructions and often only a single  
768 record. This geographic scarcity of data is a major challenge for efforts to separate regional-scale  
769 contributions that could reflect key climate-driven processes, such as changes in ocean or  
770 atmospheric circulation, from those that are restricted to a single core or site (local-scale  
771 processes). The characteristic scales of the regional non-linear and local contributions in the  
772 spatio-temporal model are  $6.1^\circ$  (~700 km) and  $0.05^\circ$  (<6 km) respectively (Table 3). Given the  
773 current distribution of neighboring RSL reconstructions compared to these scales (e.g., ~430 km  
774 from Big River to Placentia; ~700 km from Nassau Landing in Florida to Tump Point in North  
775 Carolina; Figure 1A), we acknowledge that robustly distinguishing between regional- and  
776 local-scale processes is a considerable challenge and is likely to be refined as additional  
777 reconstructions become available. Increased reconstruction precision will also help to

778 differentiate between contributions from regional- and local-scale processes because larger errors  
779 allow the spatial temporal model to produce a posterior RSL history that is consistent with the  
780 underlying data with only a modest contribution from local-scale processes.

781

### 782 *5.3 Regional-scale non-linear contribution*

783 The contrasting accumulation histories of Big River and Placentia (Figures 3 and 5) indicate that  
784 the sites may have different fidelity in recording RSL changes. Big River is an example of an  
785 ideal, continuously accumulating sequence, while the more complex history of sedimentation at  
786 Placentia includes a ~1100-year period of little accumulation (Figure 5B). During this interval  
787 assemblages of foraminifera (Figure 7; section 4.3) indicate a stable (and uniformly high)  
788 paleo-marsh elevation, meaning that the RSL reconstruction is essentially equal to the  
789 accumulation (age-depth) curve. At face value (and given only a small contribution from  
790 local-scale processes), this implies that modest (~0.5 mm/yr), GIA-driven RSL rise was offset by  
791 a regional, non-linear sea-level fall from ~0 to ~1100 CE. The resulting period of stable RSL  
792 would cause the Placentia salt marsh to cease accumulating sediment, but continued tidal  
793 inundation supported an *in-situ* population of foraminifera and prevented test loss through  
794 desiccation.

795

796 A similar regional non-linear sea-level fall occurred at Big River, but ended several centuries  
797 earlier than at Placentia (Figure 11A). The difference between the two reconstructions is greatest  
798 at ~100–400 CE, perhaps because the lower rate of GIA at Placentia did not create sufficient  
799 accommodation space to fully record the regional non-linear fall. In contrast, regional non-linear  
800 trends at Big River and Placentia are essentially identical when Placentia is unambiguously

801 accumulating sediment. Notably the sites in the northeastern United States where background  
802 rates of GIA are relatively high (Figure 11B) display more coherence in the regional non-linear  
803 signal than sites in the southeastern United States (Figure 11C) where rates of GIA are relatively  
804 low, which further suggests that locations where accommodation space is created more rapidly  
805 have greater fidelity in recording regional non-linear sea-level changes.

806

807 Comparison of the regional non-linear component of RSL change during the past ~3000 years  
808 along the Atlantic coast of North America reveals three latitudinal groups of sites that share  
809 common trends (Figure 11). Sites in Atlantic Canada experienced a declining contribution from  
810 ~7 cm at -750 CE to a minimum of approximately -10 cm at 1000 CE, which was followed by  
811 an increase until ~1850 CE, after which the regional non-linear contribution declined slightly  
812 (Figure 1A). In the northeastern United States, regional non-linear sea level fell from ~8 cm at -  
813 1000 CE to a minimum of approximately -8 cm 400 CE (Figure 11B). It then remained stable  
814 until ~1200 CE, declined modestly until 1600 CE, and rose until present. In the southeastern  
815 United States (Figure 11C), regional non-linear sea level fell also fell from -1000 CE until ~500  
816 CE when a period of stability began. The period 1000-1500 CE is characterized by a ~10 cm  
817 oscillation after which regional non-linear sea level rose to present. Studies of tide-gauge data  
818 noted similar spatial structure in sea-level trends, at least on annual to decadal timescales (e.g.,  
819 McCarthy et al., 2015; Meade and Emery, 1971; Sallenger et al., 2012). We examine physical  
820 processes that could contribute to the coherent pattern of regional, non-linear sea-level trends  
821 presented in Figure 11.

822

823 Each of the three regions exhibits a sustained sea-level fall in the period preceeding ~500 CE  
824 although the magnitude and rate of this trend varies among regions (Figure 11) and is  
825 greatest/fastest in the southeastern United States. Over increasingly long periods of time, the  
826 assumption that GIA is a linear process becomes weaker because the GIA process is slowing  
827 down over multi-millennial timescales during the Holocene (e.g., Clark et al., 1978). The falling  
828 sea-level trend, however, is not an effect of this non-linearity of GIA because this would produce  
829 an early regional-scale sea-level rise that is larger at locations where the rate of GIA is faster. We  
830 therefore, hypothesize that long-term cooling from the early to mid-Holocene maxima (e.g.,  
831 Marcott et al., 2013; Marsicek et al., 2018) caused a corresponding (falling) steric sea-level  
832 change. Observational data indicate that steric effects can produce a complex pattern of  
833 regional-scale sea-level trends (e.g., Roemmich and Gilson, 2009; Willis et al., 2004). Therefore,  
834 it is physically plausible that the trend is considered a regional rather than global trend by the  
835 spatio-temporal model.

836

837 Since ~500 CE, the three regions show different regional non-linear sea-level trends (Figure 11).  
838 Changes in the volume of the Greenland Ice Sheet would produce a characteristic fingerprint of  
839 sea-level change (e.g., Clark and Lingle, 1977; Hay et al., 2014; Mitrovica et al., 2011) where  
840 distal locations experience a larger sea-level change than locations closer to the ice sheet. For  
841 example, the ratio of RSL change to global mean sea-level change caused by a uniform loss of  
842 mass from the Greenland Ice Sheet is 0.7 at Fernandina Beach in northeastern Florida, 0.4 at  
843 New London, Connecticut, and 0.0 at Argentia, Newfoundland (Kopp et al., 2014). The  
844 latitudinal distribution of available RSL reconstructions is therefore well suited to identifying the  
845 contribution (if sufficiently large) to RSL changes from the evolving mass of the Greenland Ice

846 Sheet. However, the characteristic and tell-tale fingerprint was not detected in our analysis  
847 (Figure 11D). Although the behavior of the Greenland Ice Sheet during the past ~3000 years is  
848 not well constrained (Alley et al., 2010), it likely advanced in response to cooling temperatures  
849 (e.g., Marcott et al., 2013; PAGES2k, 2017) to reach a peak in mass and size at the end of the  
850 Little Ice Age (Weidick et al., 2004). For example, Long et al. (2012) interpreted RSL rise  
851 reconstructed in central west Greenland as evidence for growth of the Greenland Ice Sheet from  
852 1300 CE to 1600 CE (these RSL reconstructions are included in the dataset analyzed using the  
853 spatio-temporal model). Melting of the Greenland Ice Sheet since the end of the Little Ice Age  
854 (~1850 CE) returned mass to the global ocean, although Long et al. (2012) propose that a slower  
855 rate of RSL rise since 1600–1750 CE could indicate that mass loss began earlier. The fingerprint  
856 of recent and accelerating mass loss is detectable on annual timescales in recent GRACE data  
857 (Hsu and Velicogna, 2017). However, detailed analysis of 20<sup>th</sup> century tide-gauge data along the  
858 U.S. Atlantic coast concluded that no fingerprint of Greenland Ice Sheet melt could be detected  
859 (Piecuch et al., Accepted) as did studies that compared long-term rates of RSL rise to those  
860 measured by nearby tide gauges (Engelhart et al., 2011). We therefore conclude that  
861 regional-scale, non-linear sea-level trends during the past ~3000 years in the Western North  
862 Atlantic Ocean cannot be adequately attributed to changes in the mass of the Greenland Ice Sheet  
863 because the changes were too small and/or overprinted by contributions from other processes.  
864  
865 Redistribution of existing ocean mass by changes in the position and or/strength of prevailing  
866 winds and ocean currents can cause dynamic sea-level trends that are regional in scale and act  
867 over timescales from seasons to at least decades. For example, RSL observations from tide  
868 gauges coupled with measurements of northward ocean transport at 26 °N from the RAPID array

869 (e.g., Bryden et al., 2005; Cunningham et al., 2007; McCarthy et al., 2012) indicated that a 30%  
870 reduction of ocean transport (from 18.5 Sv to 12.8 Sv) caused by wind forcing and stronger  
871 geostrophic flow over the winter of 2009–2010 led to sea-level rise north of Cape Hatteras  
872 (Goddard et al., 2015). On decadal timescales, McCarthy et al. (2015) used the difference in  
873 measured sea-level trends at sites to the north and south of Cape Hatteras (Figure 1A) as an  
874 index of the strength of ocean circulation during the 20<sup>th</sup> century. These short timeseries of direct  
875 observations and modeling studies indicate a relatively large range of coastal sea-level sensitivity  
876 along the Atlantic coast of North America to changes in meridional ocean circulation (cm/Sv;  
877 e.g., Bingham and Hughes, 2009; Ezer, 2016; Kuhlbrodt et al., 2007; Levermann et al., 2005).  
878 Our results show that there was little difference in regional non-linear sea-level trends to the  
879 north and south of Cape Hatteras for the period since ~1600 CE, which suggests that changes in  
880 ocean circulation were too short-lived and/or small to cause a corresponding change in sea level  
881 that was recorded by coastal sediment. In contrast, sea-level trends during the period from ~1000  
882 CE to ~1500 CE do exhibit a notable difference between the northeastern and southeastern  
883 United States (Figure 11). At sites south of Cape Hatteras sea level rose at 1000–1200 CE and  
884 then fell to ~1500 CE. This oscillation was not recorded at locations north of Cape Hatteras,  
885 which is suggestive of a contribution from ocean dynamics. However, recent proxy,  
886 instrumental, and modeling studies also show that AMOC strength was stable until pronounced  
887 weakening in the 20<sup>th</sup> century (Caesar et al., 2018; Thornalley et al., 2018). Furthermore, the  
888 magnitude (~0.1 m) and duration (multiple centuries) of the sea-level difference would require a  
889 correspondingly pronounced change in AMOC strength and may not be physically plausible,  
890 unless it was associated (cause or effect) with an amplifying process such as shifts in prevailing  
891 winds. There is some evidence that prevailing patterns of atmospheric circulation varied during

892 the past ~2000 years from proxy reconstructions of the North Atlantic Oscillation (e.g., Franke et  
893 al., 2017; Ortega et al., 2015; Trouet et al., 2009). Although these reconstructions contradict one  
894 another at certain times, there is also some agreement among records that a positive phase of the  
895 North Atlantic Oscillation was sustained over the period of approximately 1100–1400 CE, which  
896 broadly coincides with the noted differences in sea-level trends among regions on the Atlantic  
897 coast of North America. However, similar spatial patterns of sea-level change are not present at  
898 other times of sustained positive phases in the North Atlantic Oscillation (e.g., 450–750 CE).  
899 Importantly, the relationship between the state of the North Atlantic Oscillation and RSL on the  
900 Atlantic coast of North America (particularly as expressed by the difference between sites  
901 located to the north and south of Cape Hatteras) is observed to vary depending on the timescale  
902 and specific time window under consideration (Kenigson et al., 2018; McCarthy et al., 2015;  
903 Piecuch et al., 2016; Woodworth et al., 2017). Furthermore, Woodworth et al. (2017) highlighted  
904 differences in the response of the open, shelf, and coastal oceans to changing North Atlantic  
905 Oscillation, which complicates understanding of how variability in the North Atlantic Ocean is  
906 (or indeed is not) translated physically into sea-level variability. Deciphering the role of ocean  
907 and atmospheric circulation as drivers of sea-level change (particularly during the past ~1000  
908 years) is an avenue for future work that will likely require a higher density network of RSL  
909 reconstructions than is currently available.

910

## 911 **6. CONCLUSIONS**

912 Newfoundland is likely one of the most northerly locations in the western Atlantic Ocean where  
913 salt-marsh sediment records RSL rise during the past ~3000 years. It is therefore an important  
914 region for reconstructing RSL to better understand the role of regional-scale, non-linear



915 processes as drivers of RSL change. We reconstructed RSL using foraminifera and testate  
916 amoebae preserved in cores of salt-marsh sediment at Big River and Placentia that were dated  
917 using radiocarbon and recognition of pollution horizons of known age and origin. Since  
918 approximately –1000 CE, RSL rose by ~3.0 m at Big River and ~1.5 m at Placentia. Results  
919 from a geotechnical model indicate that sediment compaction was minimal and therefore not a  
920 primary driver of RSL trends. We used a spatio-temporal model to decompose these RSL  
921 reconstructions (alongside those in an expanded Common Era database) into contributions from  
922 global, regional linear, regional non-linear, and local processes. The global contribution confirms  
923 that the 20<sup>th</sup> century rate of rise was the fastest century-scale trend of the past 3000 years. Most  
924 of the reconstructed RSL rise in Newfoundland is attributed to GIA. Only small local-scale  
925 contributions are needed to reconcile RSL reconstructions and denoised RSL trends, although  
926 larger contributions (particularly of short duration) can be accommodated within the RSL records  
927 and Placentia in particular may have experienced larger local-scale trends than we estimate.  
928 Regional non-linear sea-level in the western North Atlantic Ocean displays spatially- and  
929 temporally-coherent trends that enabled us to recognize three groups of sites (Canada, North of  
930 Cape Hatteras, and South of Cape Hatteras). Trends within and among these groups indicate that  
931 redistribution of existing ocean mass by ocean currents and/or winds likely contributed to  
932 reconstructed RSL trends.

933

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**TABLE 1:** Reported radiocarbon ages from the Big River (F25) and Placentia (C106) cores.

Core	Depth (cm)	Radiocarbon Age	Radiocarbon Age Error	$\delta^{13}\text{C}$ (‰, PDB)	Laboratory Number	Dated Material
Big River F25	16	76	29	-12.3	UtC-10372	Sp
Big River F25	24	143	41	-12.3	UtC-10373	Sp
Big River F25	32	176	31	-12.8	UtC-10374	Sp
Big River F25	40	125	39	-26.0	UtC-10375	Cr
Big River F25	48	125	35	-26.4	UtC-10376	Cr
Big River F25	54	233	26	-13.2	UtC-10517	Sp
Big River F25	60	286	33	-26.5	UtC-10516	Cr
Big River F25	66	332	34	-26.8	UtC-10515	Cr
Big River F25	74	367	33	-12.7	UtC-10470	Sp
Big River F25	80	415	40	-12.1	UtC-10518	Sp
Big River F25	86	530	34	-13.1	UtC-10519	Sp
Big River F25	94	546	42	-13.6	UtC-10520	Sp
Big River F25	102	263	29	-28.1	UtC-10521	Tm
Big River F25	107	846	38	-26.2	UtC-10522	Cr
Big River F25	108	837	36	-27.8	UtC-10523	Tm
Big River F25	119	1039	38	-13.8	UtC-10524	Sp
Big River F25	125	1012	38	-27.2	UtC-10525	Tm
Big River F25	133	1111	30	-27.1	UtC-10526	Tm
Big River F25	141	1243	39	-12.1	UtC-10527	Sp
Big River F25	149	1323	36	-25.6	UtC-10528	Tm
Big River F25	157	1321	31	-27.7	UtC-10529	Sr
Big River F25	166	1508	31	-27.9	UtC-10471	Sr
Big River F25	174	1643	29	-11.8	UtC-10472	Sp
Big River F25	174	1597	28	-15.0	UtC-10473	Sp
Big River F25	182	1653	29	-12.5	UtC-10474	Sp
Big River F25	189	1684	30	-26.6	UtC-10475	Sp
Big River F25	190	1722	31	-26.2	UtC-10476	Cr
Big River F25	195	1782	29	-13.1	UtC-10477	Sp
Big River F25	203	1842	28	-14.1	UtC-10478	Sp
Big River F25	208	1892	36	-27.8	UtC-10479	Sr
Big River F25	214	1944	30	-26.5	UtC-10480	Cr
Big River F25	220	1909	32	-27.5	UtC-10481	Tm
Big River F25	229	2017	30	-27.2	UtC-10482	Sr
Big River F25	235	2168	31	-26.5	UtC-10483	Tm
Big River F25	245	2234	33	-26.8	UtC-10484	Tm
Big River F25	254	2359	39	-27.7	UtC-10485	Tm
Big River F25	261	2422	35	-28.6	UtC-10486	Sr
Big River F25	267	2492	38	-12.9	UtC-10441	Sr
Big River F25	271	2470	43	-28.5	UtC-10442	Sr
Big River F25	277	2481	38	-26.3	UtC-10502	Tm
Big River F25	285	2578	40	-27.7	UtC-10398	Sr
Big River F25	291	2705	41	-27.5	UtC-10399	Sr
Placentia C106	30	295	20	-25.77	OS-123338	Woody rhizome
Placentia C106	30	155	20	-25.83	OS-123569	Woody rhizome
Placentia C106	30	215	15	-25.45	OS-123880	Woody rhizome
Placentia C106	39	495	15	-28.07	OS-122235	Woody rhizome
Placentia C106	45	565	20	-28.94	OS-123342	Woody rhizome
Placentia C106	45	550	20	-29.07	OS-123570	Woody rhizome
Placentia C106	51	520	15	-28.55	OS-120939	Woody rhizome
Placentia C106	60	605	20	-26.80	OS-122236	Woody rhizome
Placentia C106	64	625	25	-27.92	OS-123341	Woody rhizome
Placentia C106	77	890	20	-27.00	OS-122237	Woody rhizome
Placentia C106	80	1740	20	-24.95	OS-120940	Woody rhizome
Placentia C106	83	950	25	-27.77	OS-123340	Wood fragments
Placentia C106	89	1500	20	-26.27	OS-123339	Wood fragments

Placentia C106	94	1730	20	-26.83	OS-120941	Woody rhizome
Placentia C106	95	1910	20	-13.23	OS-122238	Ds
Placentia C106	102	2060	20	-12.65	OS-122239	Ds
Placentia C106	109	2200	20	-12.84	OS-120942	Ds
Placentia C106	118	2240	20	-25.02	OS-122240	Woody rhizome
Placentia C106	127	2220	20	-13.40	OS-120981	Ds
Placentia C106	132	2330	20	-12.11	OS-123306	Ds
Placentia C106	136	2430	20	-12.44	OS-120982	Ds
Placentia C106	141	2480	20	-23.91	OS-123337	Woody rhizome
Placentia C106	141	2470	20	-24.01	OS-123571	Woody rhizome

950  
951 Reported radiocarbon ages for samples in the Big River and Placentia cores. Samples identified  
952 by the prefix “UtC” were dated at the Utrecht van der Graaf Laboratorium. The reported  
953 radiocarbon ages and errors ( $1\sigma$ ) from this laboratory were not rounded following the  
954 conventions of Stuiver and Polach (1977). Samples identified by the prefix “OS” were dated at  
955 the U.S. National Ocean Sciences Accelerator Mass Spectrometry laboratory and reported ages  
956 and ( $1\sigma$ ) errors were rounded following the conventions of Stuiver and Polach (1977). Sample  
957  $\delta^{13}\text{C}$  values are provided relative to the Pee Dee Belemnite (PDB) standard. Sample depths listed  
958 in table include a 2-cm adjustment for dating of plant remains that lived beneath the paleo  
959 salt-marsh surface. In the Placentia core, the samples at depths of 30 cm, 45 cm, and 141 cm  
960 were dated more than once as part of a quality control procedure at the NOSAMS facility. The  
961 ages represented repeated measurements on the same plant sample and we therefore incorporated  
962 these dates into the age-depth model after creating a pooled mean for each of the three depths  
963 with replicate radiocarbon measurements. Samples described as woody rhizomes were not  
964 identified to the species level. Sp = *Spartina patens*, Cr = *Carex rostrata*, Tm = *Triglochin*  
965 *maritima*, Sr = *Scirpus robustus*, Ds = *Distichlis spicata*.

966 **Table 2:** Physical and geotechnical properties of modern salt-marsh sediment in Newfoundland.  
 967

Sample ID	Loss-on-ignition, LOI (%)	Particle density, $G_s$	Voids ratio at 1 kPa, $e_1$	Recompression index, $C_r$	Compression index, $C_c$	Yield stress, $\sigma'_y$ (kPa)
BRB15/ST06	24.05	2.42	5.69	0.07	0.83	10.5
BRM15/ST09	24.35	2.33	7.57	0.09	1.78	4
HBM15/ST02	13.88	2.75	3.23	0.04	0.93	3
HBM15/ST05	24.61	2.53	4.96	0.06	1.06	5
HBM15/ST06	32.52	2.13	5.52	0.044	1.64	4
HBM15/ST07	37.33	2.29	9.82	0.076	3.26	3
HBM15/ST09	40.00	2.08	10.08	0.12	1.32	11.5
HBM15/ST011	23.16	2.34	6.37	0.09	1.9	4
HBM15/ST013	12.32	2.57	4.15	0.02	1.56	6
PAL15/C/ST08	68.28	2.06	12.21	0.17	4.19	7
PAL15/B/ST12	81.72	1.98	16.93	0.32	2.97	8

968 The recompression index,  $C_r$ , describes the compressibility of the sample in its pre-yield, reduced  
 969 compressibility condition. The compression index,  $C_c$ , describes the compressibility of the  
 970 sample its post-yield, increased-compressibility condition. The yield stress,  $\sigma'_y$ , defines the  
 971 transition from reduced- to increased-compressibility states.  
 972  
 973

974  
975

**Table 3. Optimized hyperparameters for spatio-temporal empirical hierarchical model**

<b>Term</b>	<b>Prior SD</b>	<b>Characteristic Timescale (years)</b>	<b>Characteristic Length Scale (Degrees)</b>
$g_f(t)$ [fast global]	$\pm 0.2$ cm	12	–
$g_s(t)$ [slow global]	$\pm 16$ cm	326	
$m(\mathbf{x})$ [linear]	$\pm 1.1$ mm/yr	–	4.7
$r_f(\mathbf{x},t)$ [fast regional]	$\pm 2$ cm	12	6.1
$r_s(\mathbf{x},t)$ [slow regional]	$\pm 6$ cm	326	6.1
$l_f(\mathbf{x},t)$ [fast local]	$\pm 2$ cm	12	0.05
$l_s(\mathbf{x},t)$ [slow local]	$\pm 3$ cm	326	0.05
$w(\mathbf{x},t)$ [additional error]	$\pm 0.02$ mm	–	–
$y_0(\mathbf{x})$ [datum offset]	$\pm 0.1$ mm	–	–

976

977 **FIGURE CAPTIONS**

978

979 **Figure 1:** (A) Location of study sites in Newfoundland, Canada and existing, high-resolution  
980 relative sea-level reconstructions from the Atlantic coast of North America. Dashed line marks  
981 the approximate hinge line separating locations experiencing recent uplift/subsidence (RSL  
982 fall/rise) caused by glacio-isostatic adjustment. Location of the Port-aux-Basques and Argentia  
983 tide gauges are shown. (B–F) Site maps showing locations of coring transects and surface  
984 transects that provided samples to characterize the modern distribution of foraminifera, testate  
985 amoebae (Kemp et al., 2017c; Wright et al., 2011), and the geotechnical properties of modern  
986 sediment (this study). (G–J) Simplified stratigraphy beneath the salt marshes at Big River and  
987 Placentia. At Big River, core F25 was selected for detailed analysis. At Placentia, core 106 was  
988 selected for detailed analysis.

989

990 **Figure 2:** 10-fold cross validation of the Bayesian transfer functions. Left column shows results  
991 for foraminifera, symbol shading denotes site. Right column shows results for testate amoebae;  
992 samples from Newfoundland are represented by circles (Kemp et al., 2017c) shaded to denote  
993 sites; samples from the North Atlantic dataset of Barnett et al. (2017b) are represented by squares  
994 shaded to denote region. Top row shows actual elevation (measured at the time of sample  
995 collection) and elevation predicted under cross validation by the transfer function. Vertical error  
996 bars show the 95% credible interval of predicted elevation. Dashed 1:1 lines are provided for  
997 reference. Bottom row is the residual (difference) between actual and predicted elevations. All  
998 elevations are expressed as a standardized water level index. For foraminifera, a standardized  
999 water level index (F-SWLI) of 100 is local mean tide level (MTL) and a value of 200

1000 corresponds to the highest occurrence of foraminifera (HOF). For testate amoebae a standardized  
1001 water level index (T-SWLI) of 0 is the lowest occurrence of testate amoebae (LOT) and a value  
1002 of 100 corresponds to highest astronomical tide (HAT).

1003

1004 **Figure 3:** Big River chronology. **(A)** Downcore profiles of  $^{137}\text{Cs}$ , Pb isotopes ( $^{206}\text{Pb}$ : $^{207}\text{Pb}$ ), and  
1005 Pb concentration used to recognize marker horizons of known age. Shaded depths show the  
1006 interval in the core that corresponds to each horizon and were included in the age-depth model as  
1007 a sampling uncertainty. Labeled ages list the age assigned to each horizon and the uncertainty  
1008 was included in the age-depth model with a uniform probability distribution. **(B)** Age-depth  
1009 model developed for the Big River core using Bchron. Shaded envelope is the 95% credible  
1010 interval summarized from a suite of equi-probable accumulation histories. Symbols represent  
1011 marker horizons (uncertainties are smaller than symbols). Radiocarbon ages are calibrated ages  
1012 with bar thickness proportional to probability (note that a single radiocarbon age often yields  
1013 multiple calibrated ranges; thicker bar denotes higher probability). **(C)** Mean annual rate of  
1014 sediment accumulation estimated using the Bchron age-depth model, solid line represents mean  
1015 rate and shaded envelope captures the 5<sup>th</sup>–95<sup>th</sup> percentiles.

1016

1017 **Figure 4:** Assemblages of foraminifera in the Big River core. Abundance of the five most  
1018 common taxa are presented and together these taxa made up 99.6% of all counted individuals.  
1019 *Jm* = *Jadammina macrescens*; *BP* = *Balticammina pseudomacrescens*; *Tc* = *Tiphotrocha*  
1020 *comprimata*; *Hs* = *Haplophragmoides* spp.; *Ti* = *Trochammina inflata*; *Mf* = *Miliammina fusca*.  
1021 Samples yielding fewer than 50 individual foraminifera (shaded circles) were excluded from the  
1022 final relative sea-level reconstruction. The dissimilarity between each sample in the core and its



1023 closest modern analog was measured using the Bray-Curtis distance metric. If measured  
1024 dissimilarity exceeded the 20<sup>th</sup> percentile of dissimilarity among all possible pairs of modern  
1025 samples (labeled and dashed vertical line; the 10<sup>th</sup> and 5<sup>th</sup> percentiles are also shown for  
1026 reference) the sample was excluded from the final relative sea-level reconstruction. Symbol  
1027 shading denotes the site in the modern training set that provided the closest modern analog.  
1028 Application of the Bayesian transfer function yielded reconstructions of paleommarsh elevation in  
1029 standardized water level index (F-SWLI) units, where a value of 100 corresponds to mean tide  
1030 level (MTL) and 200 is the highest occurrence of foraminifera (HOF) in the modern training set.  
1031 Sample-specific uncertainties are the 95% credible interval.

1032

1033 **Figure 5:** Relative sea-level reconstructions from Big River and Placentia. Each box represents  
1034 the vertical and chronological uncertainty of a single data point. Predictions from two widely  
1035 used Earth-ice models (ICE-6G VM5; Argus et al., 2014 and ICE-5G VM2 ; Peltier, 2004) are  
1036 shown for comparison. For Big River, discrete, independent, and previously published relative  
1037 sea-level reconstructions from sites in western Newfoundland are show as blue boxes (Love et  
1038 al., 2016) that represent vertical and chronological uncertainty.

1039

1040 **Figure 6:** Placentia chronology. **(A)** Downcore profiles of <sup>137</sup>Cs, Pb isotopes (<sup>206</sup>Pb:<sup>207</sup>Pb), and  
1041 Pb, V and Cu concentrations used to recognize marker horizons of known age. Shaded depths  
1042 show the interval in the core that corresponds to each horizon and were included in the age-depth  
1043 model as a sampling uncertainty. Labeled ages list the age assigned to each horizon and the  
1044 uncertainty was included in the age-depth model with a uniform probability distribution. **(B)**  
1045 Age-depth model developed for the Placentia core using Bchron. Shaded envelope is the 95%

1046 credible interval summarized from a suite of equi-probable accumulation histories. Symbols  
1047 represent marker horizons (uncertainties are smaller than symbols). Radiocarbon ages are  
1048 calibrated ages with bar thickness proportional to probability (note that a single radiocarbon age  
1049 often yields multiple calibrated ranges). For reference, the depths at which foraminifera were  
1050 sparse or absent are highlighted. (C) Mean annual rate of sediment accumulation estimated using  
1051 the Bchron age-depth model, solid line represents mean rate and shaded envelope captures the  
1052 5<sup>th</sup>–95<sup>th</sup> percentiles.

1053

1054 **Figure 7:** Assemblages of foraminifera (red) and testate amoebae (blue) in the Placentia core.  
1055 Abundance of the two most common taxa of foraminifera are presented and together comprise  
1056 98.6% of individuals across all samples. Jm = *Jadammina macrescens*; Bp = *Balticammina*  
1057 *pseudomacrescens*; Ti = *Trochammina inflata*. Paleomarch elevation (PME) was reconstructed  
1058 from the assemblages of foraminifera using a Bayesian transfer function (error bars represent the  
1059 95% credible interval) and is expressed as a standardized water level index (F-SWLI) where a  
1060 value of 200 is equal to the highest occurrence of foraminifera (HOF). Abundance of the three  
1061 most common taxa of testate amoebae are presented and together comprise 58.2% of individuals  
1062 across all samples. Ce.ac.t = *Centropyxis aculeata* type; Ce.ca.t = *Centropyxis cassis* type; Ps.t =  
1063 *Pseudocorythion* type. PME was reconstructed from assemblages of testate amoebae using a  
1064 Bayesian transfer function and is expressed as a standardized water level index (T-SWLI) where  
1065 a value of 0 is equal to the lowest occurrence of testate amoebae (LOT) and a value of 200 is  
1066 highest astronomical tide (HAT). In this panel HOF is also shown for comparison following  
1067 correction for the difference in standardization between datasets. Dissimilarity between each core  
1068 sample and its closest modern analog was measured using the Bray Curtis distance metric.

1069 Samples with a minimum dissimilarity less than the 20<sup>th</sup> percentile measured among all possible  
1070 pairings of samples (dashed and labeled vertical lines; the 10<sup>th</sup> and 5<sup>th</sup> percentiles are also shown  
1071 for comparison) in the modern training sets are deemed to have an appropriate analogue.

1072

1073 **Figure 8:** Estimation of post-depositional lowering caused by physical compression of core  
1074 sediment. **(A–D)** Observed relationships between organic content (measured using  
1075 loss-on-ignition; LOI) and geotechnical/physical properties of modern salt-marsh sediments  
1076 collected from Big River, Hynes Brook, and Placentia (distinguished by symbol shape and  
1077 shading). Dashed lines (with equations) represent best fit, except in panel D where the equation  
1078 and line are from Hobbs (1986). **(E)** Comparison of model-predicted and measured bulk density  
1079 for sediment samples in the Big River core. Modeled values are the mean and standard deviation  
1080 of 5000 model runs. **(F–I)** Measured (blue) and modeled (orange, mean and 95% credible  
1081 interval) geotechnical/physical properties (including post-depositional lowering) of sediment  
1082 samples in the Big River core. **(J)** Lithostratigraphy of the Big River core.

1083

1084 **Figure 9:** Decomposition of relative sea-level reconstructions from Newfoundland using a  
1085 spatio-temporal model. In all cases, the solid line and shaded envelope are the model mean and  
1086  $1\sigma$  uncertainty. The left column presents results from Big River and the right column presents  
1087 results from Placentia. Rows are the estimated contribution to relative sea level from specific  
1088 groups of physical processes. Axes are standardized across columns to aid comparison of Big  
1089 River and Placentia. The vertical scale for the panels depicting relative sea level and the  
1090 regional-scale linear component are the same. Likewise, the panels presenting global sea level

1091 share a vertical scale with those presenting the combined contribution from regional, non-linear  
1092 and local processes.

1093

1094 **Figure 10: (A)** Reconstructed global mean sea level (mean with  $1\sigma$  uncertainty). **(B)**

1095 Contribution of local-scale trends to reconstructed relative sea level at sites along the Atlantic  
1096 coast of North America. For ease of comparison, only select sites (colored in latitudinal order)

1097 and model means are shown. Geotechnical models estimate little or no contribution from

1098 sediment compaction at Big River, Newfoundland and the record from East River, Connecticut

1099 was generated entirely from compaction-free basal sediment.

1100

1101 **Figure 11: (A–C)** The regional-scale non-linear contribution to reconstructed relative sea level  
1102 for three, latitudinally-organized regions along the Atlantic coast of North America. For ease of

1103 comparison, only select sites are shown. Each series is represented by a mean (solid line) and

1104 uncertainty ( $\pm 1\sigma$ ; shaded envelope). **(D–E)** Spatial pattern of mean, non-linear sea-level change

1105 estimated by the spatio-temporal model (units of mm/yr) for three time intervals. Note that scale

1106 differs among panels. CH = location of Cape Hatteras.

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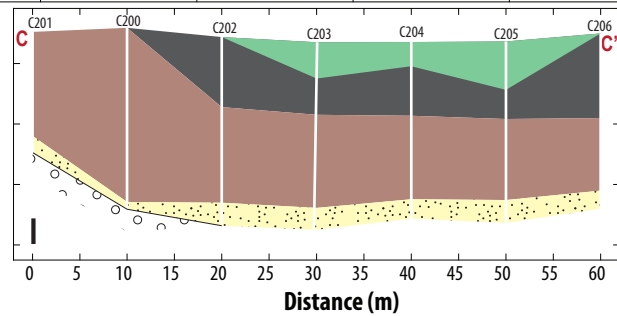
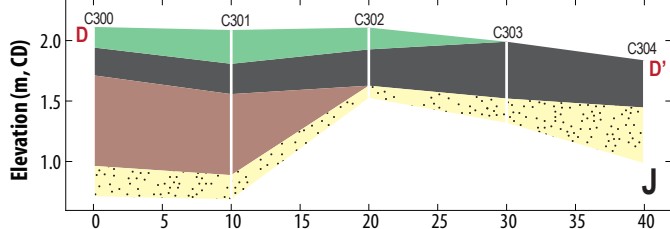
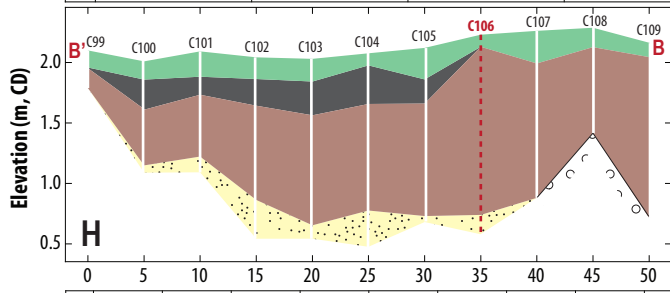
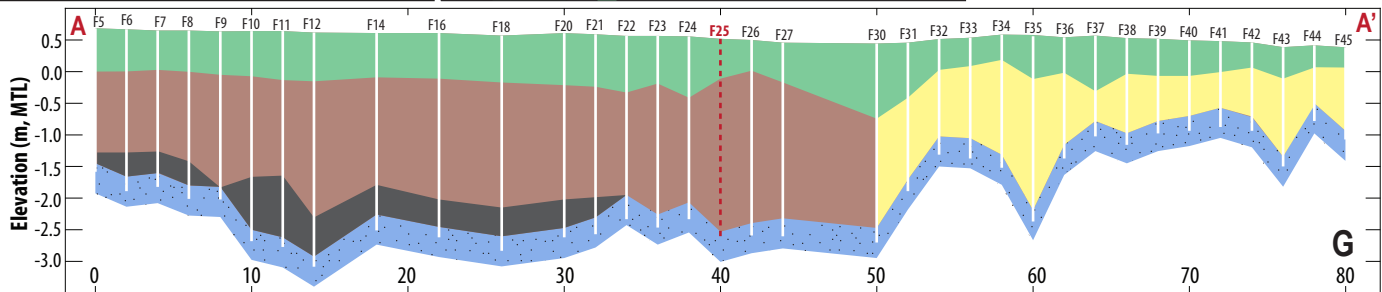
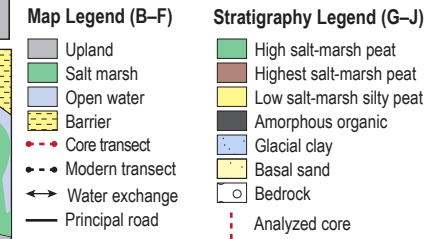
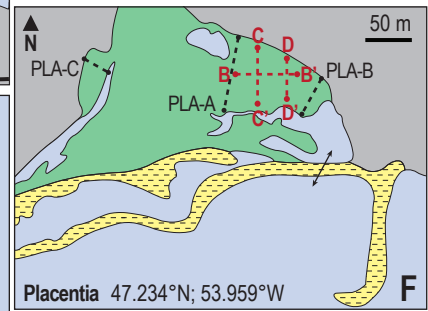
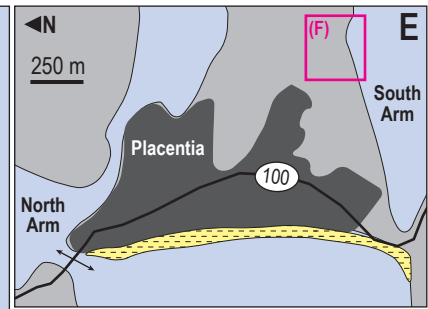
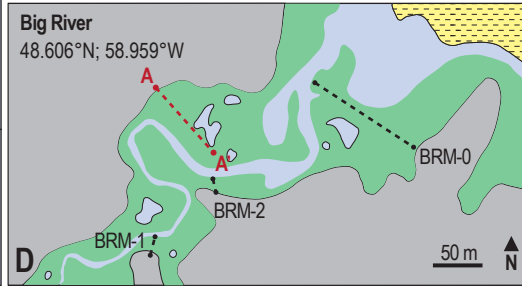
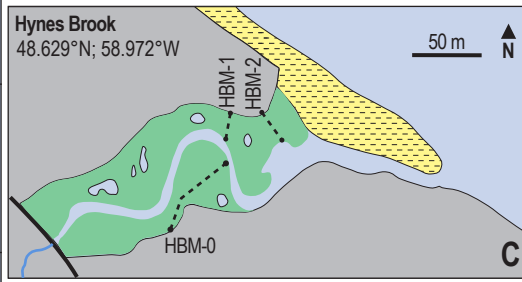
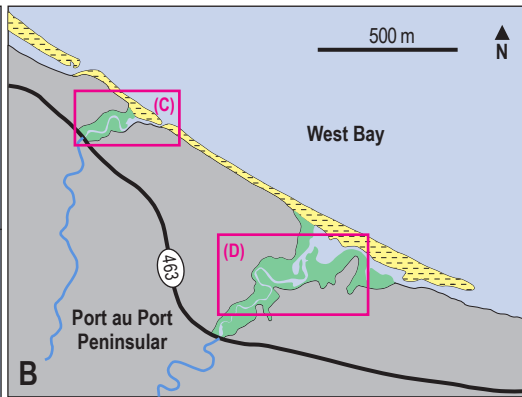
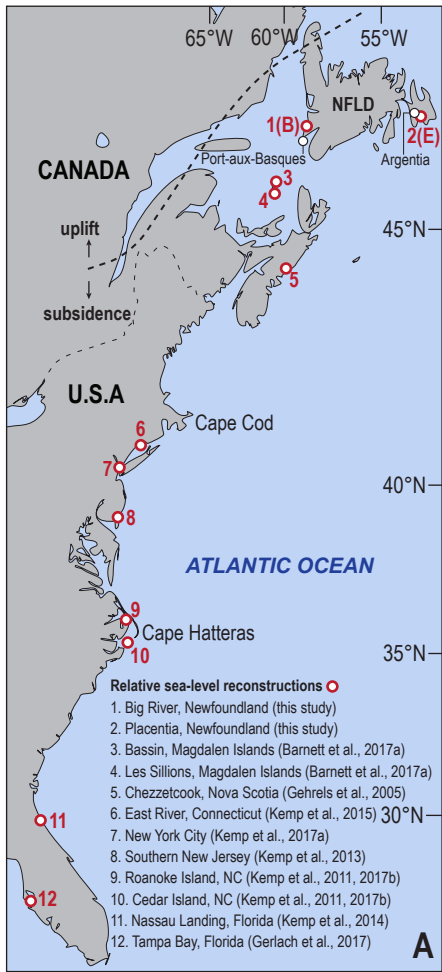


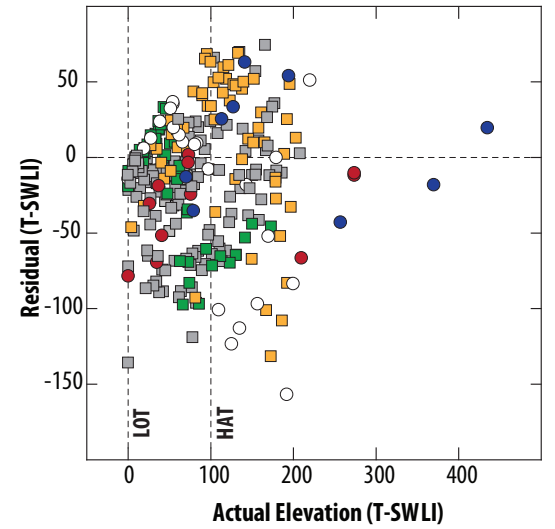
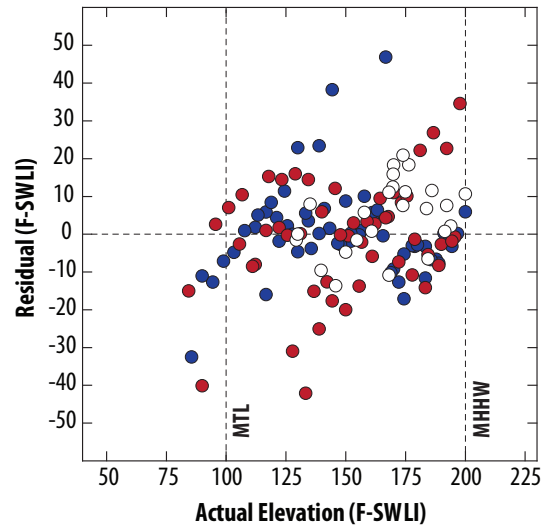
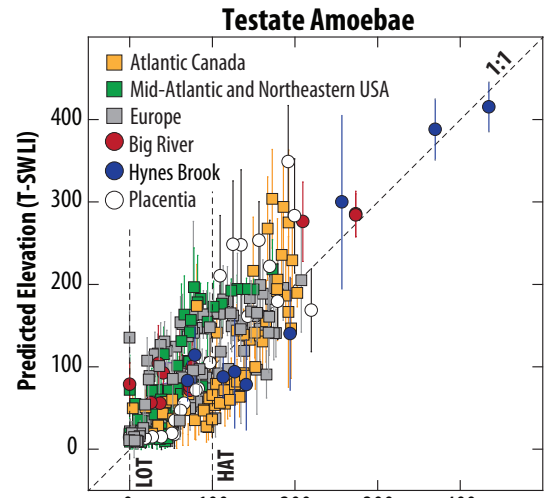
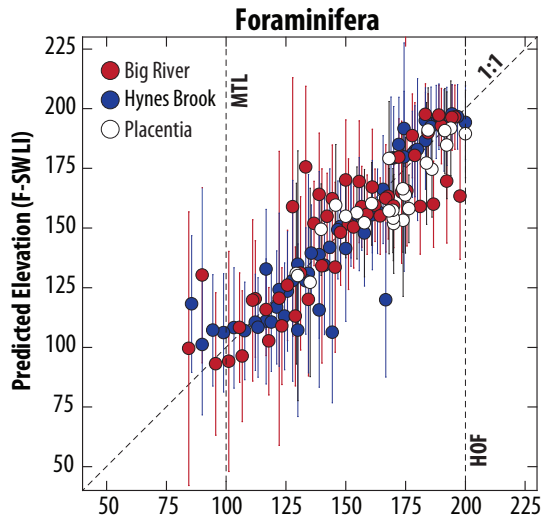
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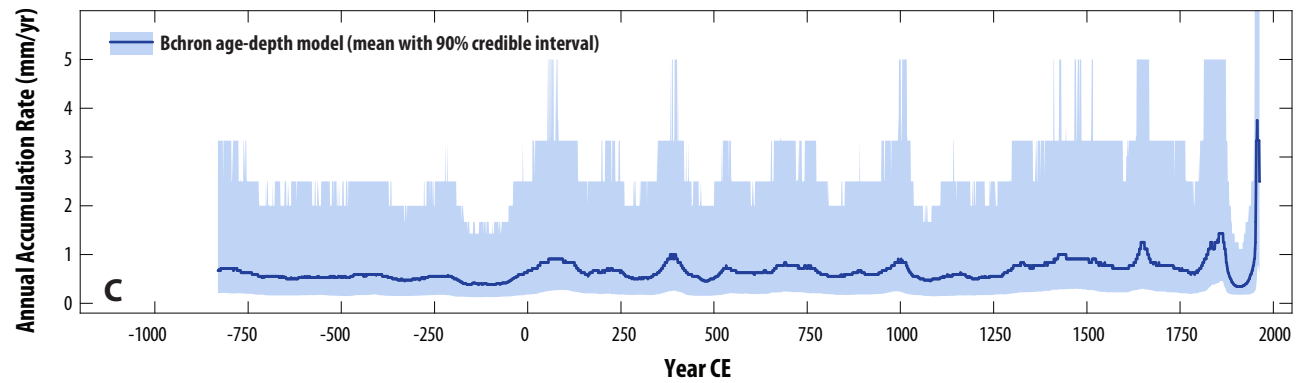
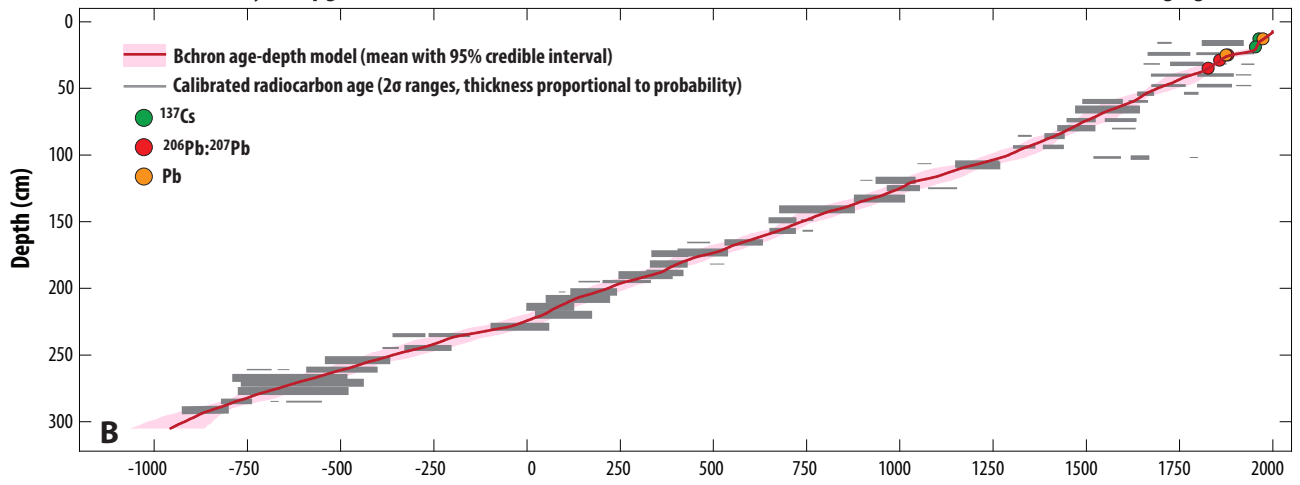
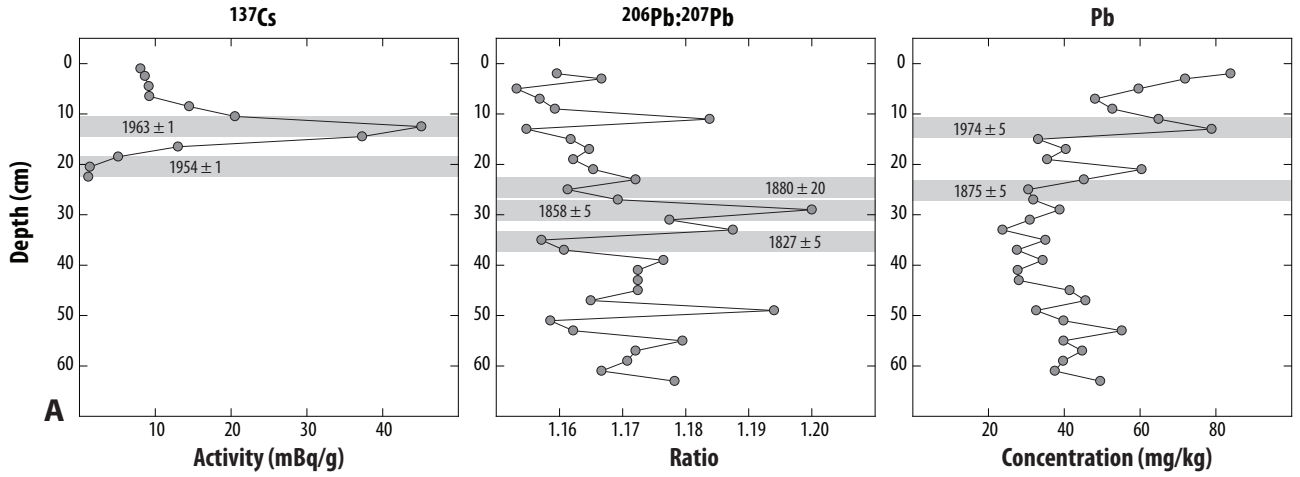
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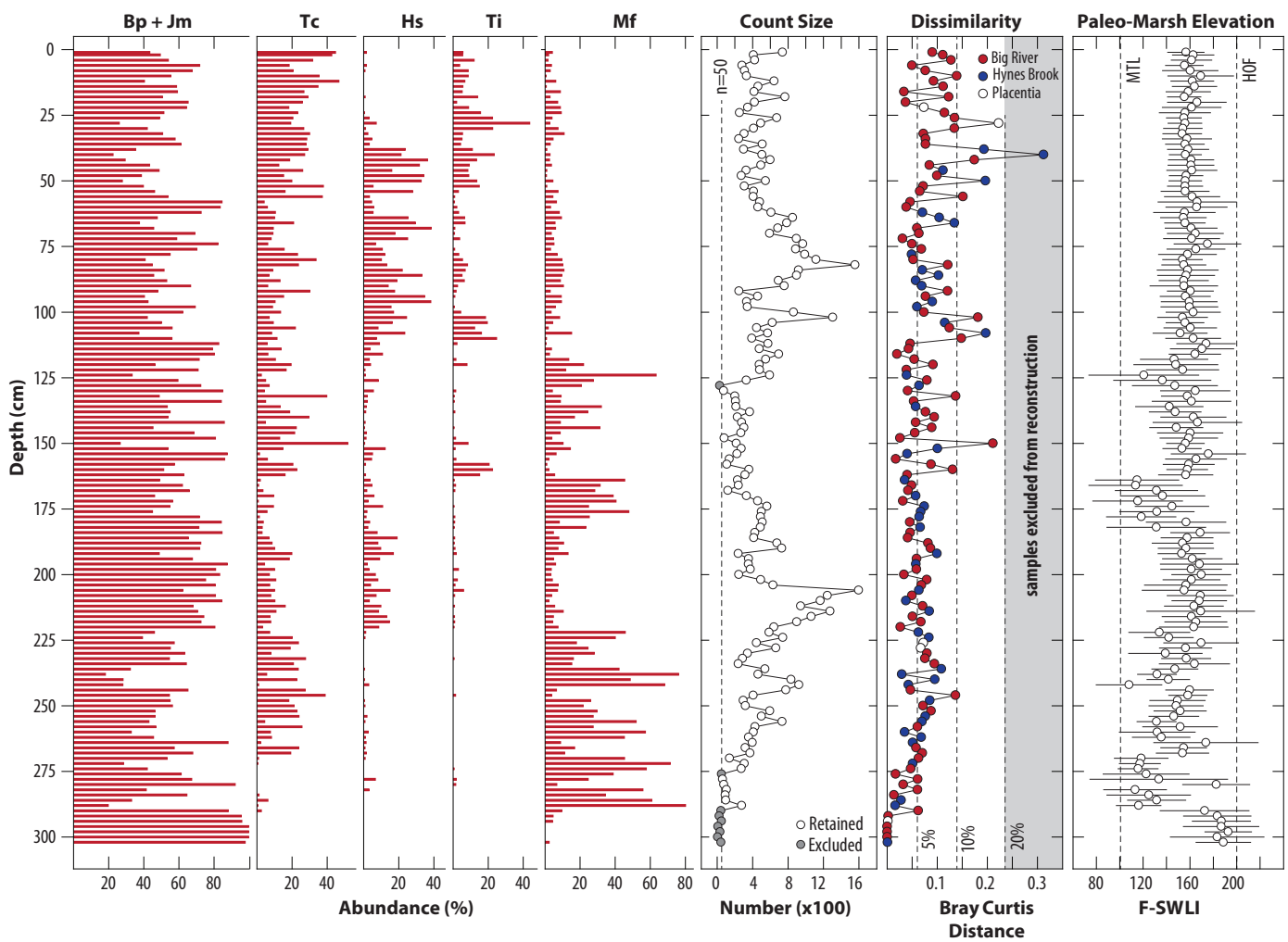
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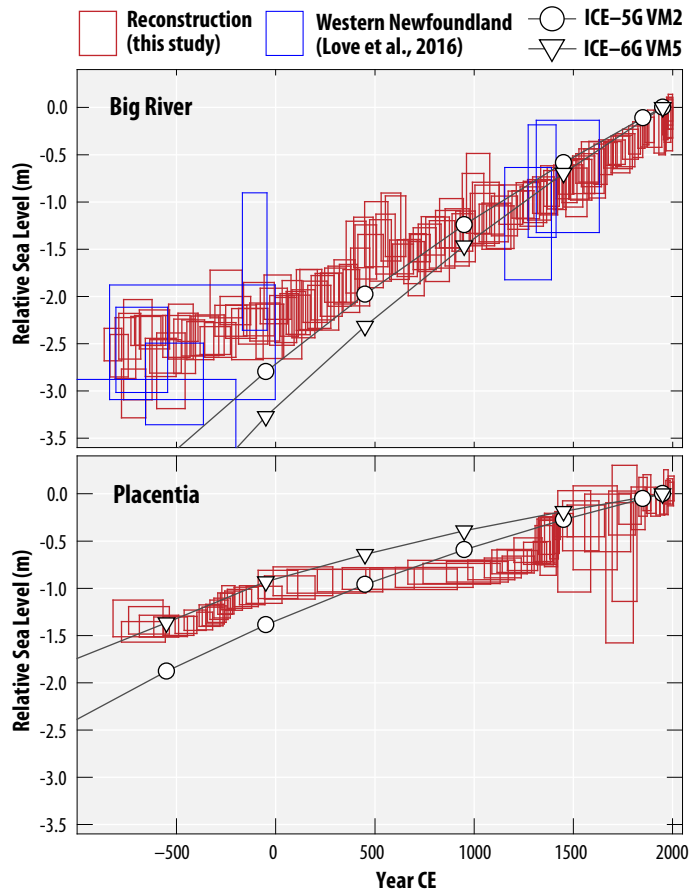


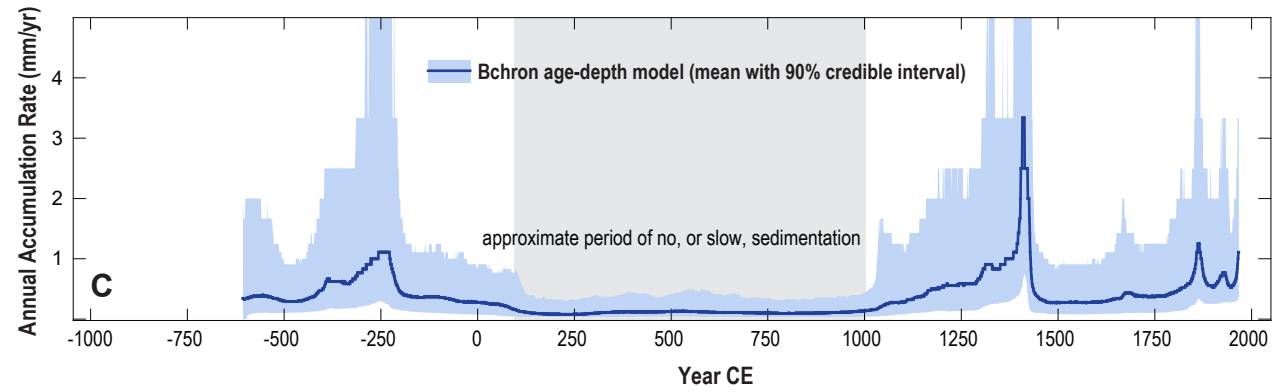
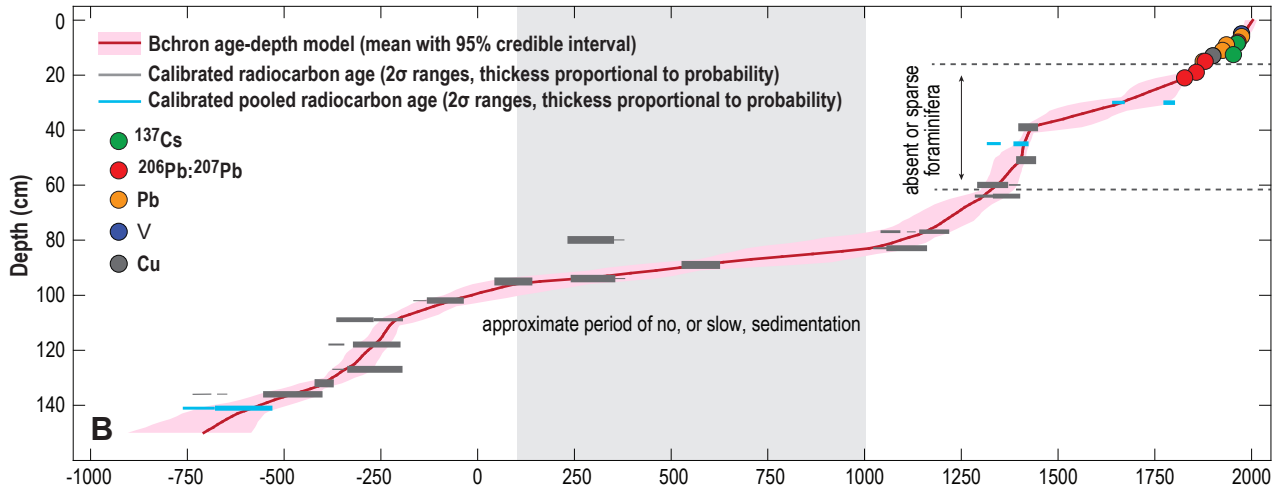
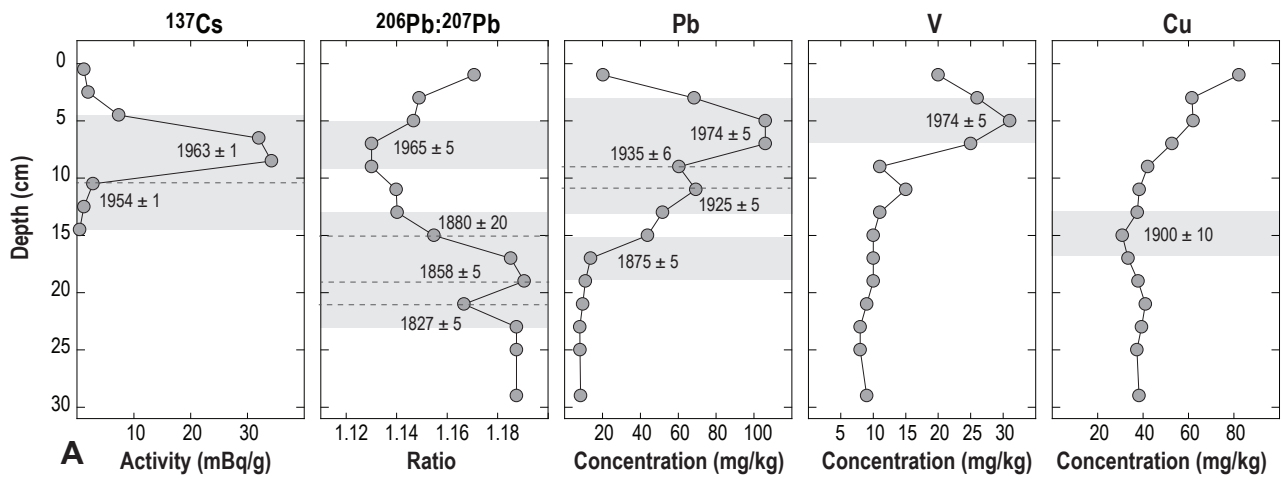


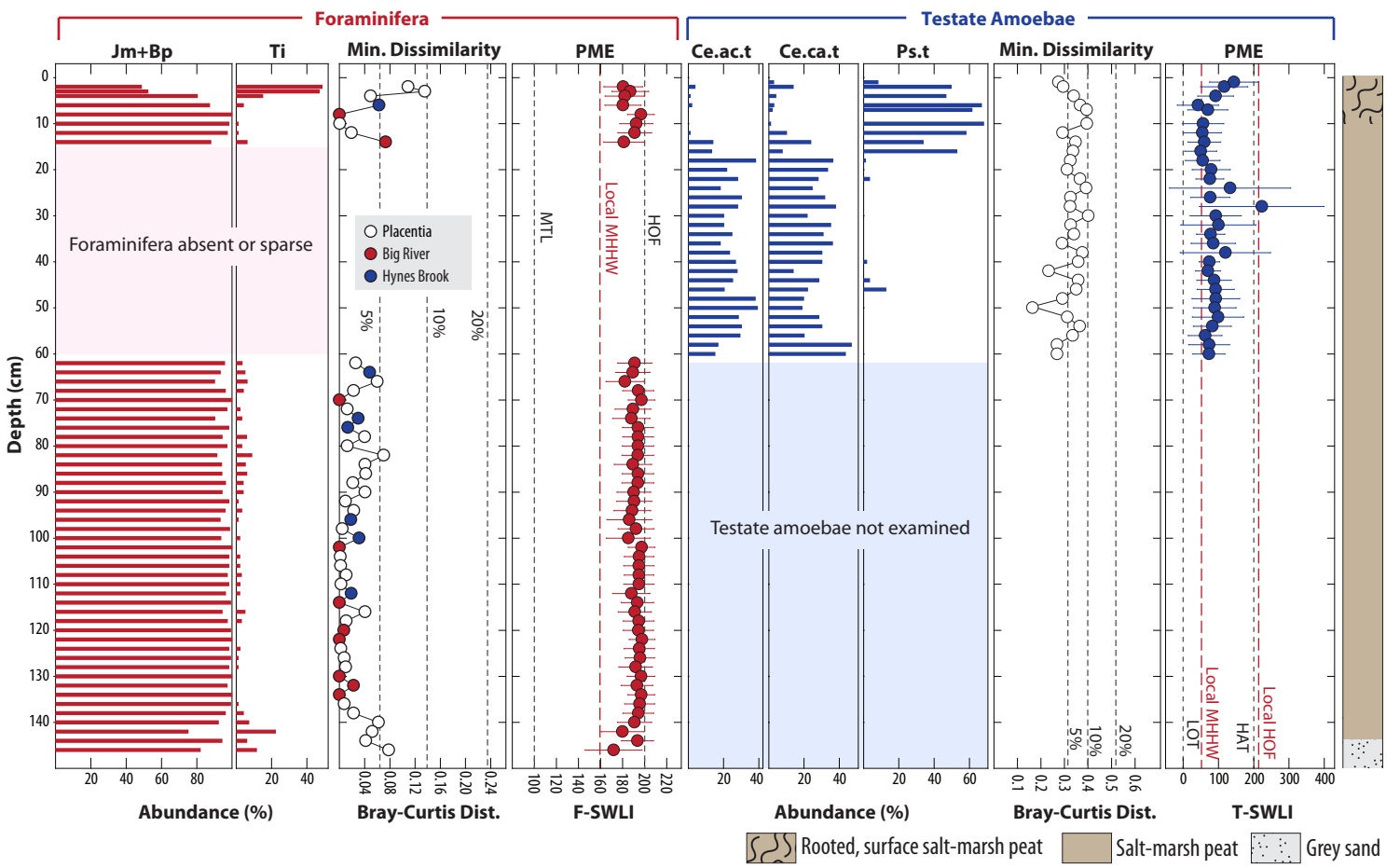


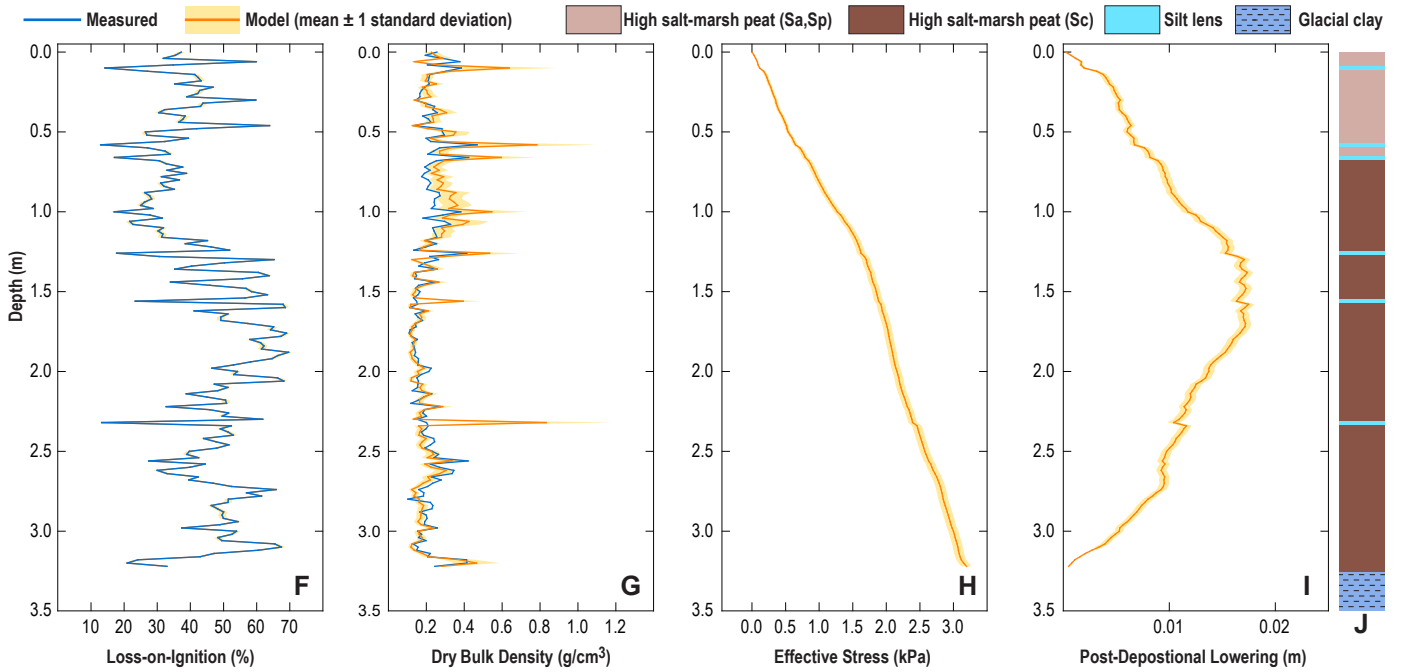
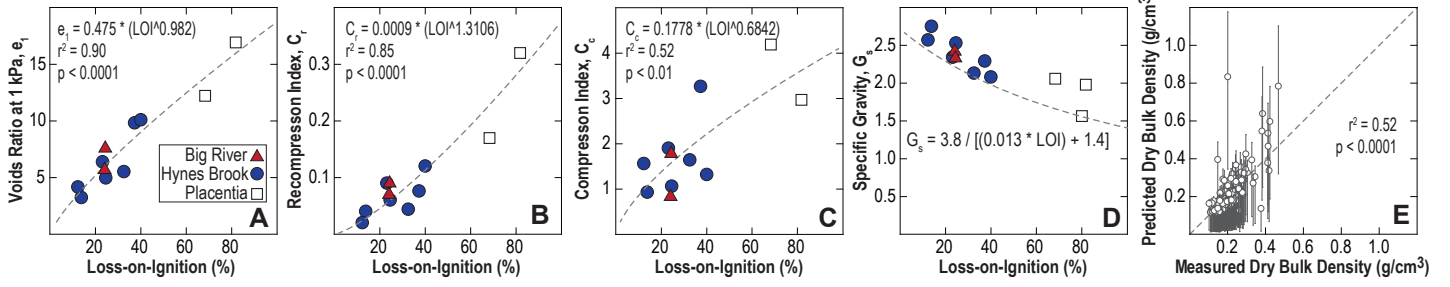


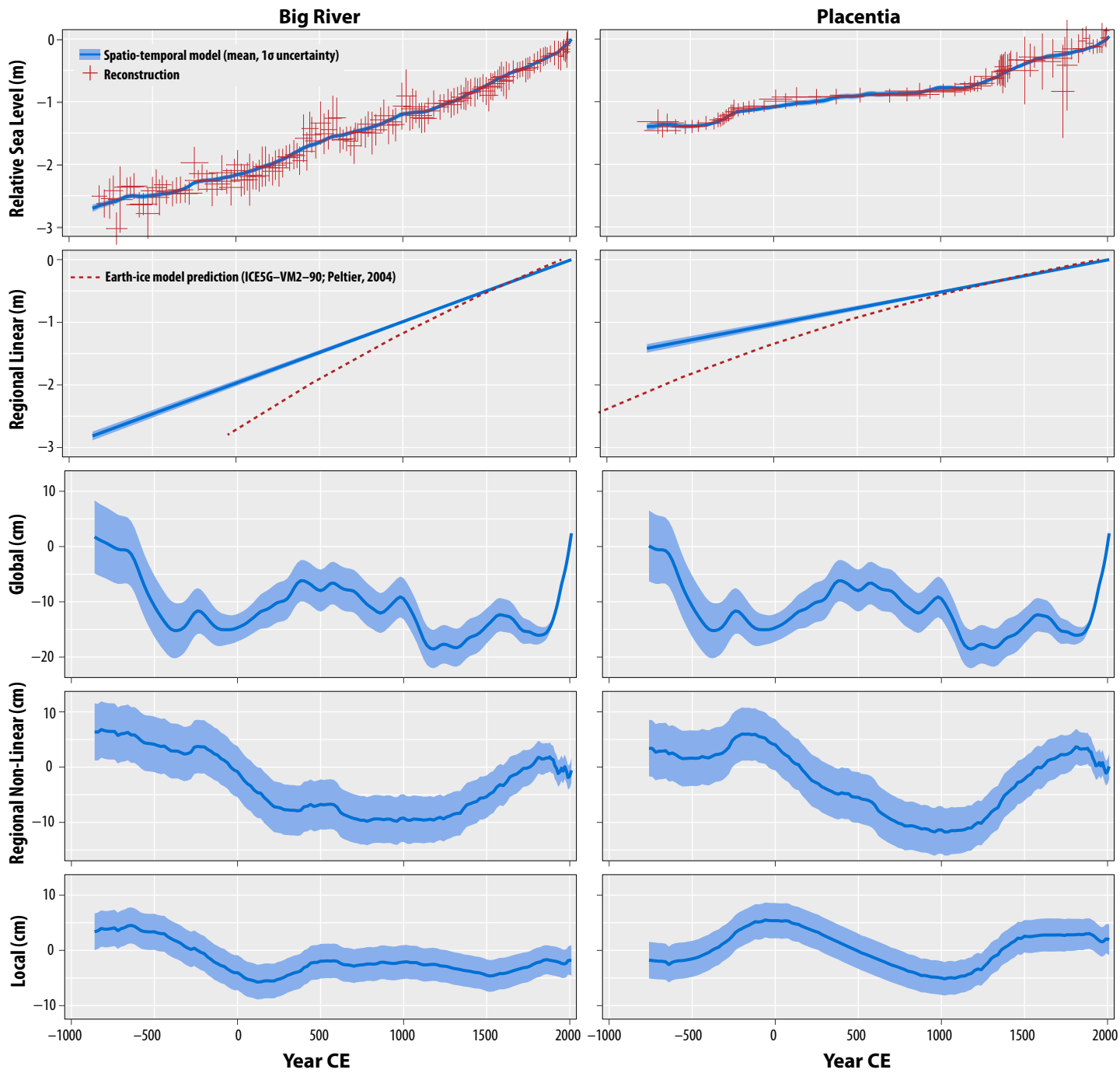


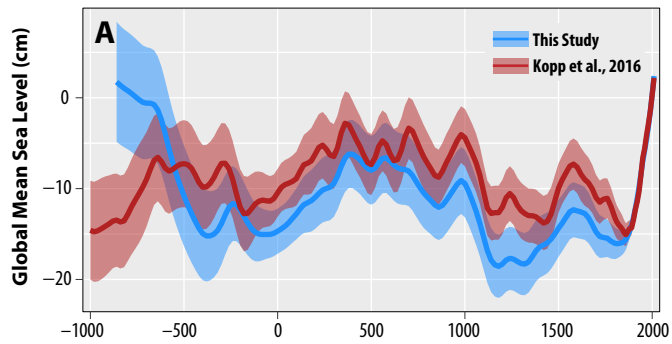












- Big River, NFLD    — Placentia, NFLD    — Les Sillons, Magdelen Islands
- East River, CT    — Pelham Bay, NY    — Cape May, NJ    - - Sand Point, NC
- Nassau Landing, FL    — Little Manatee River, FL

