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1. Introduction

Many proxy records of sea-level change that cover recent centuries show a distinct positive inflexion in the late 1800s or the early 1900s, marking the transition from late Holocene background rates of sea-level change to the high rates that have been recorded by tide gauges and satellites during the 20th and 21st centuries (Shennan and Woodworth, 1992; Shennan and Horton, 2002; Gehrels et al., 2004; Bindoff et al., 2007; Engelhart et al., 2009; Woodworth et al., 2011a). Using evidence from proxy records, many authors have dated the inflexion, but results have been variable. The following inflexion dates have been suggested: the later half of the 19th century (Connecticut, USA; Donnelly et al., 2004), the period 1900-1920 (Nova Scotia, Canada; Gehrels et al., 2005), the start of the 20th century (southern New Zealand and Tasmania; Gehrels et al., 2008; Gehrels et al., 2012), the period 1880-1920 (northern Spain; Leorri et al., 2008), the period 1879-1915 (North Carolina, USA; Kemp et al., 2009) and the period 1865-1892 (also North Carolina, USA; Kemp et al., 2011). These possible inconsistencies raise the question whether the inflexions could be non-synchronous, which has implications for the interpretation of underlying driving mechanisms. Non-synchroneity would point at a regional cause for rapid sea-level rise, such as ocean dynamical change or thermal expansion, whereas a synchronous inflexion might signal forcing by melt of ice sheets and/or glaciers. Alternatively, the varying dates of the onset of modern rates of sea-level rise could be due to chronological limitations of the proxy records.

Sea-level records spanning several decades to centuries, whether from tide gauges or proxy information, are often parameterised in terms of a linear trend superimposed upon which is variability on interannual and decadal timescales. Relative sea-level trends arise from long term changes in the ocean and/or from vertical land movements and are the subject of great interest by study groups such as the Intergovernmental Panel on Climate Change (e.g. Bindoff et al., 2007). 'Accelerations' in sea level can take the form of a gradual change in linear trends over the period of the entire record. These accelerations are often estimated by including a quadratic term in addition to the linear trend in the parameterisation, and the mean acceleration during the record is thereby calculated by multiplying the determined quadratic coefficient by two (Douglas, 1992). Many studies of tide-gauge time series (e.g. Douglas, 1992; Maul and Martin, 1993; Church and White, 2006; Jevrejeva et al., 2006; Houston and Dean, 2011; Watson, 2011; Woodworth et al., 2011a) have focussed on century-scale accelerations as determined by quadratic regressions or low order polynomials through long datasets.

When a record exhibits an abrupt change of linear trend at some time 't', then instead of using a quadratic term it may be more appropriate to parameterise the time series as an 'inflexion', the record either side of 't' being described adequately by its own linear trend and the two trend lines constrained to have the same value of sea level at 't'. The use of an inflexion parameterisation to characterise acceleration in European tide-gauge records spanning the 19th and 20th centuries was investigated by Woodworth (1990) who focused on a possible inflexion around 1930 in the longest tide-gauge records from northern Europe. Global and regional tide-gauge compilations (such as in Figure 1) have recorded inflexions around 1930 (Church and White, 2006, 2011; Jevrejeva et al., 2008; Woodworth et al., 2009), and around 1850 (Jevrejeva et al., 2008). The aforementioned inflexions identified in proxy records fall roughly between these dates, creating a possible discrepancy between the instrumental and proxy records of recent sea-level change.

Comparisons between proxy and tide-gauge records raise two main questions which are addressed in this review:

(1) why does the timing and magnitude of inflexions appear to differ in proxy and instrumental records?

(2) when did sea-level rise start departing from the long-term slow rate of sea-level rise that was persistent during much of the late Holocene?

The main aim of this paper is to reconcile the proxy and instrumental records of sea-level change during the 19th and 20th centuries. More specifically, we test the hypothesis that instrumental and proxy datasets of sea-level change are actually in agreement and both record similar times when modern rates of sea-level rise were first attained.

2. Instrumental records of sea-level change

The history of systematic sea-level observations is over three centuries long, starting in Amsterdam in 1682. What we now call automatic (or 'self-registering') tide gauges that could record the full tidal curve were developed in the 1830s, with the first often credited to Palmer (1831). These instruments took the form of a stilling well inside which was a float that was connected by a wire run over pulleys to a pen that moved up and down as the tide rose and fell, thereby drawing a tidal curve on a rotating drum of paper. The resulting continuous water-level measurements could then be expressed relative to the height of a benchmark on the nearby land.

By the end of the 19th century similar instruments had been installed at most major ports and, although sea-level measurements are often made nowadays by acoustic, pressure or radar techniques (IOC, 2006), it is important to recognise that the majority of the historical information in the archives of the Permanent Service for Mean Sea Level (PSMSL, Woodworth and Player, 2003, www.psmsl.org) stems from such conventional float and stilling well devices, and that they still constitute a large fraction of the global network.

Reviews of sea-level recording in the 'instrumental era' of the late-18th century onwards can be found in Pugh (1987), Woodworth et al. (2011a,b) and Woodworth (2012) and references therein. There are five locations in northern Europe for which instrumental records exist with lengths of two centuries or more (Amsterdam, Brest, Liverpool, Stockholm and Swinoujscie). These have been presented several times with the most recent versions shown by Woodworth et al. (2011a). Early sections of some of the records were derived from mean high water (MHW) information, rather than mean sea level (MSL), with the data having been obtained as part of operational use of docks at high waters (Woodworth, 1999). The heights and times of high water were obtained from visual observations of water level at what were then called 'tide gauges', graduated markings on the outer stone walls of the dock to indicate water depth over the dock sill. Alternatively, wooden measuring rods called 'tide-poles' or 'tide-staffs' were used. Such visual measurements could have had centimetre-level accuracy in calm weather conditions, but would have been much less accurate in the presence of waves, especially in night-time during winter.

If one applies a simple second-order fit ($a + bt + ct^2$ where t is time) of the type discussed above to the long northern European records, then quadratic coefficients 'c' of order 0.005 mm/yr^2 are obtained (i.e. accelerations of order 0.01 mm/yr^2), providing evidence for a long term acceleration in sea level and suggestive that the 20th century rise started at around the end of the 19th century (Woodworth, 1990; Wöppelmann et al., 2006; Woodworth et al., 2009, 2011a,b).

It is important to emphasise that any reported acceleration (or linear rate of change for that matter) applies only the epoch of the data from which it was computed and any such value cannot be assumed to be the same over another epoch. This is particularly true for the short term accelerations. However, it also applies to discussion at longer timescales. For example, it has been known for at least twenty years that many records from Europe and North America exhibited an overall negative acceleration (deceleration) during the 20th century (Woodworth, 1990; Tsimplis and Baker, 1990), whereas, if earlier data from the 19th century are included, then a small positive acceleration is apparent (Woodworth et al., 2011a). This important fact has been ignored in the headline reporting of some recent studies of accelerations in sea level (e.g. as pointed out in a comment by Rahmstorf and Vermeer (2011) on the study of US sea-level records by Houston and Dean (2011)). It is also clear that if we want to date the start of modern rates of sea-level rise, the lengths of the tide-gauge records are a serious limitation.

The character of shorter term acceleration in sea level has been described many times in previous publications (e.g. Douglas et al., 2000). The ocean is variable on all timescales, but particular on those from years to decades owing to large-scale processes such as El Niño or the North Atlantic Oscillation (Trenberth et al., 2007). Shorter periods of higher or lower linear sea-level trend (i.e. periods of short term ‘acceleration’) can be studied by calculating the linear trends within individual windows of a decade (or similar) throughout the record. For example, Holgate and Woodworth (2004) and Holgate (2007) studied such ‘decadal trends’ in regional and global-average records, demonstrating that the high rates of change observed in the 1990s were not unprecedented earlier in the 20th century, while other authors have investigated the use of windows of 20-30 years (e.g. Church et al., 2008; Jevrejeva et al., 2006). Periods of fairly constant acceleration (or deceleration) can be identified by inspection of such short-term trends, determining whether the linear trends in the windows increase or decrease at a uniform rate. Boon (2012) discusses the use of ‘serial trend’ analysis to provide similar information, identifying possible recent acceleration along the Atlantic coast of North America (see also Sallenger et al., 2012).

These short term accelerations can be considered as contributing to longer term ones which are our main interest. For example, in a review of the evidence for sea-level accelerations, Woodworth et al. (2009) pointed to a positive inflexion at many stations at around 1920-1930 and a negative one around 1960 which have contributed to the overall accelerations reported for the late 19th century onwards or for the 20th century alone that are usually considered the most appropriate for climate studies. These inflexions are also apparent in global compilations of tide-gauge records (Figure 1a).

3. Global compilations of instrumental sea-level records

Many attempts have been made to compute a 'global average' sea-level time series for the instrumental era by making use of the Permanent Service for Mean Sea Level (PSMSL) data set. The main difficulty with such an exercise is that most historical tide-gauge information in the data set is from the northern hemisphere, while there are obviously fewer suitable records as one goes back in time (Woodworth and Player, 2003).

The most elementary method of making a 'global-average' time series is to simply average all the available individual tide-gauge records, with each record corrected for vertical land movement using a geodynamic model of glacial isostatic adjustment (GIA) (e.g. Peltier, 2004). That method clearly biases the resulting average time series towards regions with most records (i.e. Europe, North America and Japan). A second approach involves the combining of regional-average time series into a global-average one (e.g. Douglas, 1991; Holgate and Woodworth, 2004) or averages in latitude bands into global averages (Merrifield et al., 2009).

Inevitably, these methods cannot take into consideration the possible sea-level changes which have occurred either in coastal regions not represented in the PSMSL data set, or across the vast areas of the deep ocean.

Efforts to account at least partly for sea-level variations in a more spatially-representative fashion include the use of low degree and order spherical harmonics to parameterise sea-level changes worldwide (Nakiboglu and Lambeck, 1991); empirical orthogonal functions (EOFs) of known modes of ocean variability since the 1990s when quasi-global sea level coverage became available from satellite altimetry (Church and White, 2011; Ray and Douglas, 2011); EOFs based on modes of variability in ocean circulation models with model runs performed over many decades and therefore, in principle, capable of representing lower-frequency sea-level changes more reliability than the EOFs based on altimetry (Llovel et al., 2009); and cyclo-stationary EOFs to represent progressive motions in sea-level variations instead of the standing waves of conventional EOFs (Hamlington et al., 2011). Each of these methods has drawbacks that are inevitable when using a sparse data set (e.g. see Christiansen et al. (2010), Ray and Douglas (2011) and Meyssignac and Cazenave (2012) for comments on techniques).

Meanwhile, more sophisticated ways have been designed to average individual records in a region, or globally, without consideration of particular modes of variability. These methods include the ‘virtual station’ technique of Jevrejeva et al. (2006) wherein individual records, which may be quite short, are successively combined into regional and global time series. Wenzel and Schröter (2010) used 56 selected records from the PSMSL and combined them using a neural network technique which connects coastal sea level with the regional and global mean via a non-linear empirical relationship.

In the next section we will make use of such ‘global’ and ‘regional averages’. The former is that of Church and White (2011), which is anyway similar to most others as described by

Woodworth et al. (2011b). The latter were obtained from Milne et al. (2009) and Woodworth et al. (2009) based on data from Jevrejeva et al. (2006). However, we would like to take the opportunity to make some general observations about the various global and regional 'reconstruction' exercises. One is that while there is an interesting range of different statistical techniques employed, there is little oceanographic science behind any of them, other than perhaps the various forms of EOFs. For example, we know that the sea-level variations of the central North Atlantic are dominated by the strength of the sub-tropical and sub-polar gyres including the Gulf Stream: how can such complicated patterns of variability be parameterised by a limited number of base functions (EOFs)? Nevertheless, in spite of concern about individual methods, it is interesting that they all tend to result in a similar global sea-level time series, with the 'accelerations' and 'inflexions' discussed above (cf. Figure 3 of Woodworth et al., 2009). (An exception might be the time series of Wenzel and Schröter (2010) which appears more linear with time than the others.) This apparently reassuring conclusion has to be qualified by the realisation that the same data set (PSMSL) has been used in all analyses.

In spite of the approximate agreement for the global time series, there is considerable uncertainty in providing corresponding reliable regional information using spatial parameterisation methods for those coastal regions where few or no historical data exist and for the deep ocean. This inability is hardly unexpected if one uses methods that have a statistical, rather than an oceanographic, basis. Moreover, while satellite altimetry shows that sea level is changing significantly on a regional scale, existing climate models are largely in disagreement about patterns and magnitudes of the observed variability, resulting in uncertainties on how accurate they may be in predicting future regional sea-level change. Concerns over these topics are summarised by Stammer and Gregory (2011).

4. Proxy sea-level records

Proxy records of relative sea-level change are derived from sea-level index points which are sediments, or fossils, with a known age and elevation that contain information about where sea level was in the past (Shennan, 1986). In Figure 2 we illustrate schematically the coastal stratigraphy which is typical for many settings along North Atlantic mid-latitude coastlines. In these settings sea-level index points are usually obtained from samples collected from salt-marsh or estuarine deposits. The samples contain the fossil remains of microfauna (e.g. foraminifera, diatoms) and plants which allow the relationship with former sea level (or the 'indicative meaning') to be established by comparison with the distribution of microfauna and plants on the modern coast (thus following a uniformitarian principle).

Basal samples commonly overlie an uncompressible substrate and therefore are not affected by compaction. When the index points from basal samples are plotted in an age-altitude graph, they provide an estimate for the long-term, millennial scale, rates of relative sea-level rise (Figure 2) which are important to obtain a pre-industrial rate of sea-level change, i.e. a 'baseline' against which modern rates of sea-level rise can be compared. The basal peat methodology was originally developed in the Netherlands to derive a Holocene sea-level curve (Jelgersma, 1961) and has been widely applied in other areas of northwest Europe (e.g. Denys and Baeteman, 1995; Shennan and Horton, 2002) and along the US East Coast (e.g., Redfield and Rubin, 1962; Bloom and Stuiver, 1963; Gehrels et al., 1996; Engelhart et al., 2009, 2011a). In these coastal lowlands, basal peat forms due to the rising groundwater table that is controlled by the Holocene rise in sea level, and the peat growth, although mostly of freshwater origin, is an accurate recorder of sea-level rise. In salt marshes along the east coast of North America basal peat is often formed in salt-marsh environments (Gehrels, 1999), which makes it possible to derive a more precise water-level relationship. While basal peats are widespread in the coastal lowlands of northwestern Europe and the estuaries and salt marshes of eastern north

America, in many other coastal locations organic-rich deposits are absent. In Australia and New Zealand, for example, many Holocene sea-level reconstructions rely on shells preserved in tidal flat deposits (e.g. Gibb, 1986; Sloss et al., 2007) and, in lower latitudes, on dating of mangroves, micro-atolls and other corals (e.g. Woodroffe and Horton, 2005; Woodroffe, 2009).

Sea-level index points derived from basal peat, shells, corals and other indicators do not provide sub-centennial precision which is required to reconstruct the most recent sea-level changes and to link the geological record with the instrumental record. For this reason continuous sequences in the upper sections of salt-marsh deposits are also sampled for sea-level index points (Figure 2). It has been known for over three decades that salt-marsh stratigraphy, and the fossils contained within the sediments, can be used as precise indicators of sea-level change (e.g., Scott et al., 1978; Thomas and Varekamp, 1991; Gehrels, 1994; Horton et al., 1999; Edwards et al., 2004; Donnelly et al., 2004; Gehrels et al., 2005; Kemp et al. 2009), providing data for the centuries immediately preceding the observational period (i.e. last two centuries). When a tide-gauge is located nearby, such proxy records can be directly compared to instrumental observations for the period for which they overlap, thereby providing a useful check on the validity of the proxy reconstructions. The precision and accuracy of salt-marsh proxy records depend on the integrity of the stratigraphy, i.e. low marsh sediments and tidal creeks are to be avoided (Kelley et al. 2001; Gehrels, 2006). Resolution is a function of the sedimentation rates in the marshes and is usually on the order of one data point per decade, obviously lower than observational records. The marsh records therefore do not provide information on interannual sea-level variability, but they are valuable archives of (multi-)decadal relative sea-level trends. The vertical precision of sea-level estimates from salt-marsh sediments is typically ±5-20 cm and is constrained by the types of fossil indicators that are used to reconstruct sea level and by the local tidal range. Along microtidal coasts, the vertical ranges of the sea-level indicators are small and here the most precise sea-level reconstructions are possible (Southall et al. 2006; Callard et al. 2011).

Limited precision of dating techniques provide additional uncertainties, but in recent years some important advances have been made to improve chronologies of salt-marsh based sea-level reconstructions (e.g. Marshall et al., 2007). Dating methods include analyses of radioactive isotopes, such as ^{14}C and ^{210}Pb. The former has recently been applied to young sediments (e.g. Marshall et al., 2007; Gehrels et al., 2012) using high-precision Accelerator Mass Spectrometry (AMS) ^{14}C dating of multiple samples. Errors are typically reduced to less than 10 radiocarbon years by matching a stratigraphically constrained set of radiocarbon ages to the calibration curve (Marshall et al., 2007). The ^{210}Pb method can only be used to date sediments younger than ~120 years and results depend on the particular dating model that is applied, the selection of which can be aided by additional dating methods such as ^{137}Cs which identifies the 1965 level (when nuclear bomb testing was globally at its peak) or local nuclear spill events. In addition to ^{137}Cs, which can be diluted and transported within the sediment column, bomb-spike AMS^{14}C dating can give very precise dates for the period after 1950. This method measures ^{14}C activity in fossil samples and matches these to the known atmospheric bomb-spike curve, providing monthly precision (Marshall et al., 2007). Stratigraphic marker techniques, such as Pb isotopes and metal concentrations can usefully fill in some dating gaps in the 19th century, for example by matching levels in cores with archives of hemispheric atmospheric pollution, such as ice cores, and by comparisons with historical regional pollution records, such as mining histories (Gehrels et al., 2006, 2008, 2012; Marshall et al., 2007). Along the North American east coast, and in Tasmania and New Zealand, pollen markers and charcoal records provide additional chronological markers by revealing distinct changes in vegetation resulting from deforestation, land clearing, forest fires and agricultural activities by European settlers (Gehrels et al., 2005, 2008, 2012; Kemp et al., 2009).

In this paper we re-analyse some key proxy records by plotting the sea-level index points from stratigraphic levels that have been directly dated by one of the methods described above. This approach is different from some of the published records (Gehrels et al., 2005; Leorri et al., 2008; Kemp et al., 2009, 2011) which are based on age-depth models and also include sea-level index points for which an age is derived by interpolation. In our re-evaluation we ignore these 'synthetic' data, because we take the view that they obscure the true age uncertainties of the records. We compare recent proxy and instrumental sea-level records with the late Holocene rate of relative sea-level change and determine the start of modern rates sea-level rise by identifying by visual inspection the time when sea-level rise departed from the long-term background rate. We assume a linear rate of late Holocene relative sea-level change which is consistent with our data (see next section). An important point to note is that this method does not require any corrections of the relative sea-level records for land movements, as subtracting the background rate from the modern rate essentially eliminates all millennium-scale relative sea-level processes from the record, including those resulting from GIA.

5. Identifying start of modern rates of sea-level rise

Establishing the timing of the onset of modern rates of sea-level rise can be achieved reliably in sites where two main criteria are fulfilled: (1) the available recent sea-level record, based on either instrumental or proxy data, is of sufficient length (~200 years); and (2) the 'background' rate of relative sea-level rise is known at the same location. These restrictions limit our analyses to the seven sites that are discussed below. It is perhaps surprising that there are only a small number of sites where this type of analysis can be carried out. There are many coastal sites around the world where late Holocene sea-level trends can be determined, but the recent sea-level record, from either proxy or instrumental data, is often too short to determine when the modern sea-level rise departed from this background trend. For the seven sites where we are able to address this aim of our study we show the late Holocene and the recent sea-level data in Figures 3 and 4, respectively. Data that underlie our analyses are given in Tables 1 and 2. The late Holocene and recent sea-level records from the seven sites are briefly discussed in the following sections, but for additional details we refer to the original studies.

5.1. Nova Scotia, Canada

The reconstruction from Chezzetcook (Scott et al., 1995; Gehrels et al., 2004, 2005) is from a salt marsh on the central Atlantic coast of Nova Scotia (Figure 3a, 4a). Foraminifera were used as sea-level indicators. The tide-gauge record from nearby Halifax agrees well with the reconstruction for the overlapping period. The 20th century part of the proxy record is dated by ^{210}Pb ages, but there are no dates for the 19th century. The earlier part of the record contains the introduction of ragweed following European settlement dated at 1760-1800, and a radiocarbon measurement in the early 18th century. The background late Holocene rate is based on seven basal sea-level index points that are younger than 4000 years (Scott et al. 1995; Gehrels et al., 2004). A linear regression through these dates gives a long-term rate of relative sea-level rise of 2.17 mm/yr (Figure 3a). The lower error bars of the proxy data overlap with the late Holocene trend until the 1970s, but the departure of the tide-gauge record from the background trend occurs in the period 1930-1940. This is 20 to 30 years later than the inflexion identified by Gehrels et al. (2005) who compared 19th and 20th century rates.

5.2. Connecticut, USA

The Barn Island record by Donnelly et al. (2004) includes proxy data up to ~1900 AD (Figure 3b, 4b) that are coupled with tide-gauge data from New York starting in 1856, and New London starting in 1939. The proxy data are 11 basal sea-level index points, all younger than 700 years, collected from the base of a salt-marsh section where it overlies a sloping glacial erratic. The indicative meaning of these samples was determined by analyses of fossil plant remains. The radiocarbon ages were calibrated, but some of the calibrated results were rejected on stratigraphic grounds, producing seemingly small age errors (Table 1). Two ages were stratigraphic pollen markers and one was a pollution marker. Donnelly et al. (2004) suggest that “the nearly three-fold increase in the regional rate of sea-level rise to modern levels likely occurred in the later half of the 19th century”. However, it can be seen in Figure 4b that the late Holocene background trend of 1.2 mm/yr continues until at least 1920-1930.

5.3. North Carolina, USA

The proxy record from Sand Point (Kemp et al., 2009, 2011; Engelhart et al. 20011a) contains a good chronology for the 20th century, based on bomb-spike ^{14}C, ^{137}Cs and ^{210}Pb. The preceding 200 years are covered by two dates, one high-precision ^{14}C and one pollen marker (ragweed introduction). The sea-level indicators that are used are foraminifera. Kemp et al. (2009, 2011) also discuss a second site (Tump Point), about 120 km to the southwest, but it is ignored in this analysis because the basal dates there show large scatter (Horton et al., 2009; Engelhart et al., 2011a) and the late Holocene background relative sea-level rise cannot be established with satisfactory precision. Twenty late Holocene basal sea-level index points from Sand Point (Engelhart et al., 2011a) are all younger than 2700 years and provide a background relative sea-level rise of 0.9 mm/yr (Figure 3c). The modern trend (Figure 4c) agrees well with tide-gauge data and diverges between 1925 and 1935, significantly later than determined by Kemp et al. (2011) using change-point analysis on age-modelled data.

5.4 Southwest England

Gehrels et al. (2011) published 10 late Holocene sea-level index points from a basal salt-marsh section in Thurlestone, southwest England, using foraminifera as sea-level indicators. There is some clustering in the ages of the index points, but overall they provide a late Holocene trend of sea-level rise of 0.9 mm/yr (Figure 3d). There are no proxy data for the last 200 years, but the western English Channel contains the longest continuous tide-gauge record in the world at Brest (Wöppelmann et al., 2006). We compare the background late Holocene rate with the Brest record, and also with the tide-gauge record at Newlyn, which is closer to Thurlestone and which shows a similar trend as Brest (Figure 4d). Despite the interannual variability in the tide-gauge records it appears that the most recent trend of sea-level rise has exceeded the background trend for most of the 20th century.

5.5 The Netherlands

The Amsterdam instrumental record is one of the longest in Europe commencing in 1682 but terminating in 1925 prior to the closure of the Zuiderzee. Van Veen (1945) documented the series starting in 1700; we use a slightly amended version provided by the Rijkswaterstaat available from the PSMSL (www.psmsl.org/data/longrecords/ancill_rep.htm). The Amsterdam time series was extended by Woodworth et al. (2011a) with the use of modern MSL information from Den Helder, located on the open North Sea and some 65 km north of Amsterdam, with the two time series constrained to have the same values of sea level in their period of overlap 1865-1925. This composite record is shown in Figure 4e.

The late Holocene rate of relative sea-level rise is estimated from the dataset of van de Plassche et al. (2005). This dataset (Figure 3e) is from Schokland, about 100 km to the northeast of Amsterdam. There are 31 basal sea-level index points, but only four of these are younger than 4000 cal yr BP. If we include a further three dates that are between 4000 and 5000 years old, we obtain a background relative sea-level rise of 1.2 mm/yr for the central Netherlands between 5000 and 1500 cal yr BP. However, these relatively old sea-level index points are likely to produce an overestimate, as relative sea-level rise during the middle to late Holocene gradually slowed down. We therefore include in our calculation of the background trend the sea-level position known from the Amsterdam tide gauge around 1700, which is about -0.15 m. This yields a background rate of -0.7 mm/yr for index points that are younger than 4000 cal yr BP and -0.8 mm/yr if we also include the three points between 4000 and 5000 cal yr BP. We use the more conservative rate (-0.7 mm/yr) as a best estimate of the late Holocene relative sea-level trend in the central Netherlands (Figure 3e). The instrumental sea-level record and the late Holocene background trend diverge after ca. 1910.

5.6 Tasmania, Australia

The proxy record from Little Swanport in southeastern Tasmania is based on foraminifera and contains 31 sea-level index points (Figure 4f), including 6 that provide age control for the 19th century (Gehrels et al., 2012). Dating methods include ^{210}Pb, ^{137}Cs, bomb-spike ^{14}C, high-precision ^{14}C, stable Pb isotopes, chemostratigraphy and pollen markers. The background rate of late Holocene sea-level change is less certain. Two middle Holocene sea-level index points from shells, one of which is basal, and a GIA model prediction (Lambeck et al. 2002) suggest that the background rate is close to zero (Figure 3f). Nearby tide gauges were only installed in recent decades and do not provide suitable records for comparison. There are short-term (ca. 2 years) historical observations from the 1840s (Hunter et al., 2003), but they do not agree with the proxy reconstruction, for reasons that are not clear (Gehrels et al., 2012). The record departs from the zero trend between 1895 and 1920, but the poorly constrained late Holocene sea-level index points, as well as the lack of dates in the early 20th century, hinders the exact dating of the divergence. Also note that sea level has not risen by much in the second half of the 20th century. This deceleration can also be seen in the regional compilations of tide-gauge data from the western Pacific and the Indian Ocean (Figure 1b).

5.7 Pounawea, New Zealand

The proxy record from New Zealand (Gehrels et al., 2008) relies on ^{137}Cs, stable Pb isotopes, chemostratigraphy, pollen and charcoal analyses. Sea-level estimates are derived from foraminiferal analyses and for the 20^{th} century they are in good agreement with the tide-gauge record from Lyttelton (Figure 4g) and Bluff (Gehrels et al., 2008). Similar to Tasmania, basal sea-level index points are not found in this setting. As in Tasmania, the flat background rate (-0.1 mm/yr) is poorly constrained. It is calculated from middle and late Holocene sea-level index points (Figure 3g) derived from shells from Pounawea (Gehrels et al., 2008) and nearby Blueskin Bay (Gibb, 1986), and is confirmed by a GIA model prediction (Lambeck et al., 2002; Gehrels et al., 2012). There are no dates in the proxy record between 1895 and 1935, but the departure from the late Holocene background trend of relative sea-level rise occurs in this period.

6. Discussion

In Table 3 we have compiled the data from the seven sites that are investigated in this study. We also include an entry for the 'typical' global reconstruction mentioned above (Church and White, 2011) and a compilation of regional-averaged tide-gauge records for those ocean basins which are relevant to the sites (those of Figure 1b). For each of the global and regional entries we have estimated an inflexion timing in a similar way to the seven individual records above (i.e. by visual inspection). It is clear that, in both proxy and instrumental records, the inflexions are recorded in the early part of the 20^{th} century, roughly between 1905 and 1945 (the average calculated from the inflexion ranges shown in Table 3, excluding the global average, is 1925).

An unexplained exception is the regional subset for the NE Atlantic for which an inflexion is less clear in Figure 1b, unlike the situation for individual long instrumental and proxy records (Woodworth et al., 2011a). The difference may be due to the inclusion of shorter records in the

Jevrejeva et al. (2006) analysis although that remains unconfirmed. As mentioned above, the longest European instrumental records all show an acceleration on the order of 0.01 mm/yr^2 (e.g., Woodworth, 1999). The 20th century rates of sea-level change are significantly higher than the 19th century rates at these locations, and also at four additional German stations (Warnemünde, Wismar, Travemünde and Cuxhaven) where records are available from the middle of the 19th century (Woodworth et al., 2011b).

When combining instrumental sea-level data with proxy data it is important to be aware of limitations and (dis)advantages of both datasets. Tide gauges have advantages over the proxy methods in two main ways. First, modern tide gauges can sample sea-level change at any desired frequency. For example, many of the modern gauges in the Global Sea Level Observing System (GLOSS, Merrifield et al., 2009) also contribute data to tsunami warning networks, with measurements integrated over a minute or even more frequent sampling. Earlier, the paper charts of the pre-electronic era were digitised to provide hourly (or similar) values of sea level for the determination of tidal parameters and storm-surge statistics. Second, gauges provide values of sea level relative to benchmarks on the nearby land to sub-centimetre accuracy. As a consequence, tide gauges can provide accurate values of annual mean sea level (MSL) relative to a benchmark to centimetre accuracy or better, which contrasts with proxy information which has poorer temporal resolution (typically decadal, but depending on the particular situation) and often less accurate datum control.

Conversely, tide gauges have two main disadvantages relative to the proxy records. First, the lengths of most of their records are limited to a century or less, with only a small number of longer records from northern Europe (Woodworth et al., 2011a). This compares to the several centuries or even millennia of some salt-marsh records. Second, it is well known that most of the historical instrumental information stems from the northern hemisphere (see Figure 1 of Woodworth et al., 2011b), while proxy techniques are in principle applicable worldwide (although the type of proxy, e.g. salt-marsh indicators or micro-atolls, are latitude dependent). It is therefore necessary to compare information from the two methods where available, and then to make maximum use of the proxy methods especially in parts of the world where little historical instrumental information exists.

The use of age-depth models in proxy studies helps to explain seemingly incompatible age ranges for the onset of modern rates of sea-level rise. Construction of a best fit age-depth plot from the available dates and interpolation to estimate ages of intermediate samples result in many ‘synthetic’ data points. This approach essentially violates the established methodology of using sea-level index points (Shennan, 1986), because many samples are not directly dated. Age-depth models have been adopted from other palaeoenvironmental studies, in particular those that are concerned with environments where sedimentation rates are relatively constant, such as peatlands and the deep sea. However, salt-marsh sedimentation is notoriously episodic so that resolving changes in accretion rates would require a large number of dates. Sedimentation rates in salt marshes are, in fact, coupled with sea-level changes. When sea-level rise slows, the frequency of submergence and sedimentation rates decrease. Dating resolution will be adversely affected. The converse is true for accelerating sea-level rise.

To establish with confidence the timing of the onset of modern rates of sea-level rise in proxy records, it is undesirable to use age-depth models, because they can create inflexions that will bias the sea-level reconstruction. Information about the true age estimates and their uncertainties may be lost. For example, if microfossil assemblages are similar throughout a core the sea-level inflexions would be entirely driven by the age-depth model, so that a lack of dates in a section of the core becomes a critical limitation. Some studies (Kemp et al., 2009, 2011) have used Bayesian change-point linear regression (Carlin et al., 1992) on sea-level reconstructions, but these analyses do not produce objective results when used in combination with age-depth models, again because inflexions in age-depth models strongly depend on the number of available dates (i.e. where there are gaps in the chronology an inflexion is produced midway between two dates). Our study demonstrates that when the interpolated sea-level index points are removed from analyses, the inflexions seen in proxy data are in good agreement with the instrumental datasets.

Another issue related to chronology is the use of separate dating methods that each define a part of the chronology. An example is the use of ^{14}C chronology for the older part of the sea-level record in combination with ^{210}Pb dating for the more recent part. An inflexion is often observed where the ^{14}C and ^{210}Pb-based reconstructions meet (e.g. Gehrels et al., 2005; Kemp et al., 2009), raising the possibility that the inflexion is an artefact of issues and limitations intrinsic to the dating techniques, such as selection and uncertainties of ^{210}Pb models and/or ^{14}C calibration precision. For more robust estimates of inflexion ages in proxy sea-level records it is desirable that multiple dating methods are used, and that at least some techniques date the parts of the reconstructions before and after the inflexion. Examples of such complementary dating methods are stable Pb isotopic markers, Pb pollution markers and high-precision ^{14}C dating (e.g., Marshall et al., 2007; Gehrels et al., 2008, 2012). All these methods can be applied to 19th century and early 20th century sediments and can therefore capture the inflexion more accurately than the combination of ^{210}Pb and conventional ^{14}C dating alone.

Compaction is an important issue when dealing with sea-level reconstructions from intertidal sediments (Brain et al., 2011). Compaction can lower the altitude of sea-level index points in age-depth plots relative to the level of original deposition. The use of basal peats is an established method that provides sea-level index points that are immune to compaction (Jelgersma, 1961; Gehrels, 1999; Donnelly et al., 2004), resulting in reliable late Holocene sea-level trends for our North Atlantic sites. The stiff tidal flat sediments in Tasmania and New Zealand from which late Holocene sea-level trends are derived can be considered as ‘over-consolidated’ due to drying out and slow deposition rates during the stable sea levels that prevailed for many millennia during the middle and late Holocene. These sediments also provide a stable substrate for the overlying salt-marsh stratigraphies from which the recent sea-level trends are reconstructed. However, it is possible that compaction within salt-marsh stratigraphies can lower the surface of the marsh and produce a transgressive signal that can be erroneously interpreted as a rise of sea level (Mörner, 2010). This issue has recently been studied by Brain et al. (2012) using a numerical modelling approach based on physical sediment properties and geotechnical theory. Importantly, they specifically scrutinised the recent proxy reconstructions that are discussed in this paper. Brain et al. (2012) conclude that the stratigraphies of the salt marshes in Tasmania and New Zealand are not conducive to compaction. The Connecticut record is derived from basal sea-level index points and is therefore also immune. In the other North Atlantic marshes, records with shallow (<0.5 m) uniform lithologies experience negligible compaction. Deeper sequences that contain transgressive stratigraphies, resulting in increases in the amount of minerogenic sediment towards the top, can add up to 0.4 mm/yr of local sea-level rise to the record (for a 3 metre long sequence). The Nova Scotia sequence is about 2 m long, but the lithology is relatively uniform. The North Carolina sequence is about 1 m long, and there is an increase in the minerogenic component in the upper part, including a sand layer at the top. Compaction in both the Nova Scotia and North Carolina records cannot be entirely ruled out, but two further observations make it unlikely that compaction is significant (Brain et al., 2012). Firstly, compaction processes are time dependent and do not produce abrupt inflexions as seen in the sea-level reconstructions. If compaction were important, the reconstructed curves would be gradual with continuous curvatures. Secondly, the reconstructed 20th century sea-level trends in Nova Scotia and North Carolina are similar to the trends observed in nearby tide-gauge records. Therefore, we conclude that compaction processes have little bearing on both the timing and the magnitude of inflexions observed in the proxy sea-level reconstructions.

We cannot establish with great certainty the cause of the early 20th century inflexion, but we speculate that melting of northern hemisphere ice masses may have been an important contributor. The mass balance of Greenland changed in the 1920s to a more negative state according to several modelling studies (Wake et al., 2009; Hanna et al., 2011) as a result of considerable warming over Greenland (Chylek et al., 2006). These studies only deal with changes in surface mass balance of the Greenland Ice Sheet, but Box (in review) shows a model reconstruction that also includes rapid ice discharge and that suggests a contribution to global sea-level rise of 5.4 mm per decade for the period 1922-1932 (Figure 5a). Reconstructions based on the length of glaciers (LeClercq et al., 2011) show that glaciers started to contribute significantly to sea-level rise in the middle of the 19th century, with highest contributions to sea-level rise achieved in the 1930s (Figure 5b; Gregory et al., in revision). Given the warming that has been recorded in high northern latitudes in the 1920s and 1930s (Johannessen et al., 2004), it is likely that Arctic glaciers contributed significantly to the signal shown in Figure 5b (Gregory et al., in revision). This is confirmed by the modelling study of Marzeion et al. (2012). According to their reconstructions, the overall contributions of glaciers to global sea-level rise were higher during the 20th century than calculated by LeClercq et al. (2011). Marzeion et al. (2012) suggest that this discrepancy could be due to a lack of records from the Canadian and Russian Arctic in the LeClercq et al. (2011) study, or to the inability of their own model to capture ice dynamics and distinguish between floating and land-based ice. They calculate a contribution of Arctic glaciers of up to 2 mm/yr in the 1920s/1930s (Figure 5c).

Sea-level fingerprinting is an indirect method which can be used to constrain past contributions of ice melt. The method takes advantage of the observation that sea-level change caused by melting ice sheets and glaciers is not globally uniform but results in distinct spatial patterns, or fingerprints, whose geometries depend on the location of the melt source and result from the diminishing gravitational attraction which the ice mass exerts on the ocean surface while it is melting (Mitrovica et al., 2001). Water will migrate away from the ice sheet and the net effect is that the most rapid sea-level rise occurs up to 1000s of kilometres away from the ice mass, whilst nearby the melting ice mass sea level may actually be falling (Tamisiea et al., 2003). Using sea-level fingerprinting, Mitrovica et al. (2001) estimated that Greenland has contributed ~0.6 mm/yr to global sea-level rise during the 20th century, whereas Nakada and Inoue (2005) suggested a Greenland melt contribution of ~1 mm/yr. Neither of these studies, however, provided a temporal framework for the melt, only overall 20th century estimates. These are high compared to direct measurements of Greenland Ice Sheet mass-balance changes, which show that, during the relatively warm 1960s, Greenland melted at a rate equivalent to a sea-level rise of 0.3±0.2 mm/yr (Rignot et al., 2008). There are no reliable direct observations from before the 1960s, only the modelled reconstructions discussed above, but if Greenland and high-latitude glaciers contributed significantly to the global rise of sea level in the 1920s, one would expect to observe a spatial pattern that shows higher rates of sea-level rise in the southern hemisphere. The general pattern of Greenland mass loss (Mitrovica et al., 2001) is, at least qualitatively, in agreement with our data (Figure 6) which show that the differences between late Holocene and 20th century rates in the Tasmania and New Zealand records are of a greater magnitude than in the North Atlantic records. The same pattern for pre- and post-inflexion rates of sea-level change was discussed by Gehrels et al. (2012). There is also a suggestion of a latitudinal trend along the Atlantic coast of North America, also observed by Engelhart et al. (2009). As discussed above, Brain et al. (2012) rule out compaction problems in southern hemisphere sites, but they do point out that if North Atlantic salt-marsh sequences are corrected for compaction the existing contrast between northern and southern hemisphere sites would increase, thereby potentially amplifying the Northern Hemisphere melt signal that could be interpreted from the spatial fingerprint. The modelling study by Kopp et al. (2010) suggests that the southeastern Indian and the western Atlantic oceans are areas where ocean dynamics may obscure a gravitational sea-level fingerprint signal. It is therefore desirable that future studies using the fingerprinting technique explore the Greenland/Arctic melt hypothesis further, especially in other southern hemisphere locations (e.g. the southern Atlantic) where, according to Kopp et al. (2010), the Greenland mass-loss signal is predicted to be large and detectable.

7. Conclusions

This paper has addressed two main questions:

(1) why does the timing and magnitude of inflexions appear to differ in proxy and instrumental records?

(2) when did sea-level rise start departing from the long-term slow rate of sea-level rise persistent during much of the late Holocene?

The timing of inflexions in published proxy records appears to be affected by the use of age-depth modelling. When the sea-level data for which no firm age information is available are removed from the proxy datasets a consistent picture emerges. In the seven sites analysed in this study, two from the eastern North Atlantic, three from the western North Atlantic, one from the Tasman Sea and one from the southwest Pacific, it appears that modern rates of sea-level rise started between 1905 and 1945. This change produced inflexions that are also seen in many compilations of tide-gauge records, and we therefore suggest that the proxy and instrumental sea-level datasets are compatible. The consistent timing across the globe points mainly at a glacio-eustatic origin of the inflexion, although some variability is to be expected due to oceanographic dynamical processes and steric overprints. It appears that the magnitude of the inflexion is larger in the southern hemisphere, which is consistent with a northern hemisphere melt source as suggested by sea-level fingerprinting theory. This hypothesis is supported by reconstructions of the contributions to global sea-level rise by the Greenland Ice Sheet and Arctic glaciers and requires further testing in other sites, especially in the southern hemisphere.

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9. References

Bindoff, N.L., Willebrand, J., Artale, V., Cazenave, A., Gregory, J., Gulev, S., Hanawa, K., Le Quéré, C., Levitus, S., Nojiri, Y., Shum, C.K., Talley, L.D., Unnikrishnan, A., 2007. Observations: Oceanic Climate Change and Sea Level. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.), Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 385-432.

Bloom, A.L., Stuiver, M., 1963. Submergence of the Connecticut coast. Science 139, 332-334, doi:10.1126/science.139.3552.332.

Boon, J.D., 2012. Evidence of sea level acceleration at U.S. and Canadian tide stations, Atlantic coast, North America. Journal of Coastal Research 28, 1437-1445, doi: 10.2112/JCOASTRES-D-12-00102.1.

Box, J., in review. Greenland ice sheet mass balance reconstruction. Part III: Marine ice loss and total mass balance (1840-2010). Journal of Climate, submitted 31 July, 2012.

Brain, M.J., Long, A.J., Petley, D.N., Horton, B.P., Allison, R.J., 2011. Compression behaviour of minerogenic low energy intertidal sediments. Sedimentary Geology 233, 28-41, doi:10.1016/j.sedgeo.2010.10.005.

Brain, M.J., Long, A.J., Woodroffe, S.A., Petley, D.N., Milledge, D.G., Parnell, A.C., 2012. Modelling the effects of sediment compaction on salt marsh reconstructions of recent sea-level rise. Earth and Planetary Science Letters 345-348, 180-193, doi:10.1016/ j.epsl.2012.06.045.

Callard, S.L., Gehrels, W.R., Morrison, B.V. Grenfell, H.R., 2011. Suitability of salt-marsh foraminifera as proxy indicators of sea level in Tasmania. Marine Micropaleontology 79, 121-131, doi:10.1016/j.marmicro.2011.03.001.

Carlin, B.P., Gelfand, A.E., Smith, A.F.M., 1992. Hierarchical Bayesian analysis of changepoint problems. Applied Statistics 41, 389–405, doi:10.2307/2347570.

Chylek, P., Dubey, M.K., Lesins, G., 2006. Greenland warming of 1920-1930 and 1995-2005. Geophysical Research Letters 33, L11707, doi:10.1029/2006GL026510.

Church, J.A., White, N.J., 2006. A 20th century acceleration in global sea-level rise. Geophysical Research Letters 33, L01602, doi: 10.1029/2005GL024826.

Church, J.A., White, N.J., 2011. Sea-level rise from the late 19th to the early 21st century. Surveys in Geophysics 32, 585-602, doi: 10.1007/s10712-011-9119-1.

Church, J.A., White, N.J., Aarup, T., Wilson, W.S., Woodworth, P.L., Domingues, C.M., Hunter, J.R., Lambeck, K., 2008. Understanding global sea levels: past, present and future. Sustainability Science 3, 9-22, doi:10.1007/s11625-008-0042-4.

Christiansen, B., Schmith, T., Thejll, P., 2010. A surrogate ensemble study of sea level reconstructions. Journal of Climate, 23, 4306-4326, doi:10.1175/2010JCLI3014.1.

Denys, L., Baeteman, C., 1995. Holocene evolution of relative sea level and local mean high water spring tides in Belgium – a first assessment. Marine Geology 124, 1-19, doi:10.1016/0025-3227(95)00029-X.

Donnelly, J.P., Cleary, P., Newby, P., Ettinger, R., 2004. Coupling instrumental and geological records of sea-level change: evidence from southern New England of an increase in the rate of sea-level rise in the 19th century. Geophysical Research Letters 30, L05203, doi:10.1029/2003GL017801.

Douglas, B.C., 1991. Global sea level rise. Journal of Geophysical Research 96, 6981-6992, doi:10.1029/91JC00064.

Douglas, B.C., 1992. Global sea level acceleration. Journal of Geophysical Research 97, 12699-12706, doi:10.1029/92JC01133.

Douglas, B.C., Kearney, M.S., Leatherman, S.P. (Eds.), 2000. Sea-level rise: history and consequences. Academic Press, San Diego. 232pp.

Edwards, R. J., van de Plassche, O., Gehrels, W. R., Wright, A. J., 2004. Assessing sea-level data from Connecticut, USA, using a foraminiferal transfer function for tide level. Marine Micropalaeontology 51, 239-255, doi: 10.1016/j.marmicro.2003.11.003.

Engelhart, S.E., Horton, B.P., Douglas, B.C., Peltier, W.R., Törnqvist, T., 2009. Spatial variability of late Holocene and 20th century sea-level rise along the Atlantic coast of the United States. Geology 37, 1115-1118, doi: 10.1130/G30360A.1.

Engelhart, S.E., Peltier, W.R., Horton, B.P., 2011a. Holocene relative sea-level changes and glacial isostatic adjustment of the U.S. Atlantic coast. Geology 39, 751-754, doi:10.1130/G31857.1.

Engelhart, S.E., Horton, B.P., Kemp, A.C., 2011b. Holocene sea-level changes along the United States' Atlantic Coast. Oceanography 24, 70-79, doi:10.5670/oceanog.2011.28.

Gehrels, W.R., 1994. Determining relative sea-level change from salt-marsh foraminifera and plant zones on the coast of Maine, U.S.A. Journal of Coastal Research 10, 990–1009.

Gehrels, W.R., 1999. Middle and late Holocene sea-level changes in eastern Maine reconstructed from foraminiferal saltmarsh stratigraphy and AMS ^{14}C dates on basal peat. Quaternary Research, 52, 350-359, 10.1006/qres.1999.2076.

Gehrels, W.R., 2006. Sea level studies: Microfossil reconstructions. In: Elias, S. (Ed.), Encyclopedia of Quaternary Sciences, Elsevier, 3015-3023.

Gehrels, W.R., Belknap, D.F., Kelley, J.T., 1996. Integrated high-precision analyses of Holocene relative sea-level changes: Lessons from the coast of Maine. Geological Society of America Bulletin 108, 1073-1088, doi:10.1130/0016-7606(1996)108<1073:IHPAOH>2.3.CO;2.

Gehrels, W.R., Kirby, J.R., Prokoph, A., Newnham, R.M., Achterberg, E.P., Evans, E.H., Black, S., Scott, D.B., 2005. Onset of recent rapid sea-level rise in the western Atlantic Ocean. Quaternary Science Reviews 24, 2083-2100, doi: 10.1016/j.quascirev.2004.11.016.

Gehrels, W.R., Marshall, W.A., Gehrels, M.J., Larsen, G., Kirby, J.R., Eiriksson, J., Heinemeier, J., Shimmield, T., 2006. Rapid sea-level rise in the North Atlantic Ocean since the first half of the 19th century. The Holocene 16, 948-964, doi:10.1177/0959683606hl986rp.

Gehrels, W.R., Milne, G.A., Kirby, J.R., Patterson, R.T., Belknap, D.F., 2004. Late Holocene sea-level changes and isostatic crustal movements in Atlantic Canada. Quaternary International 120, 79-89, doi:10.1016/j.quaint.2004.01.008.

Gehrels, W.R., Hayward, B.W., Newnham, R.M., Southall, K.E., 2008. A 20th century sea-level acceleration in New Zealand. Geophysical Research Letters 35, L02717, doi:10.1029/2007GL032632.

Gehrels, W.R., Callard, S.L., Moss, P.T., Marshall, W.A., Blaauw, M., Hunter, J., Milton, J.A., Garnett, M.H., 2012. Nineteenth and twentieth century sea-level changes in Tasmania and New Zealand. Earth and Planetary Science Letters 315-316, 94-102, doi:10.1016/j.epsl.2011.08.046.

Gibb, J.G., 1986, A New Zealand regional Holocene eustatic sea-level curve and its application to determination of vertical tectonic movement, Royal Society of New Zealand Bulletin 24, 377-395.

Gregory, J. J.A., Church, N.J., White, M.F., Bierkens, P., Box, J E., van den Broeke, M., Cogley, J.G., Fettweis, X., Hanna, E., Huybrechts, P., Leclercq, P.W., Marzeion, B., Oerlemans, J., Wada, Y., Wake, L.M., van de Wal, R.S.W. (in revision). Twentieth-century global-mean sea-level rise: is the whole greater than the sum of the parts? Journal of Climate.

Hamlington, B.D., Leben, R.R., Nerem, R.S., Han, W., Kim, K-Y. 2011. Reconstructing sea level using cyclostationary empirical orthogonal functions. Journal of Geophysical Research 116, C12015, doi:10.1029/2011JC007529.

Hanna, E., Huybrechts, P., Cappelen, J., Steffen, K., Bales, R.C., Burgess, E., McConnell, J.R., Steffensen, J.P., van den Broeke, M., Wake, L., Bigg, G., Griffiths, M., Savas, D., 2011. Greenland Ice Sheet surface mass balance 1870 to 2010 based on twentieth century reanalysis, and links with global climate forcing. Journal of Geophysical Research 116, D24121, doi:10.1029/2011JD016387.

Holgate, S.J., 2007. On the decadal rates of sea level change during the twentieth century. Geophysical Research Letters 34, L01602, doi:10.1029/2006GL028492.

Horton, B. P., Edwards, R. J., Lloyd, J. M., 1999. A foraminiferal-based transfer function: Implications for sea- level studies. Journal of Foraminiferal Research 29, 117-129.

Horton, B.P., Peltier, W.R., Culver, S.J., Drummond, R., Engelhart, S.E., Kemp, A.C., Mallinson, D., Thieler, E.R., Riggs, S.R., Ames, D.V., Thomson, K.H., 2009. Holocene sea-level changes along the North Carolina coastline and their implications for glacial isostatic adjustment models. Quaternary Science Reviews 28, 1725-1736, doi:10.1016/j.quascirev.2009.02.002.

Holgate, S.J., Woodworth, P.L. 2004. Evidence for enhanced coastal sea level rise during the 1990s. Geophysical Research Letters, 31, L07305, doi:10.1029/2004GL019626.

Houston, J., Dean, R., 2011. Sea-level acceleration based on U.S. tide gauges and extensions of previous global-gauge analysis. Journal of Coastal Research 27, 409–417, doi:10.2112/JCOASTRES-D-10-00157.1.

Hunter, J., Coleman, R., Pugh, D., 2003. The sea level at Port Arthur, Tasmania, from 1841 to the present. Geophysical Research Letters 30, doi:10.1029/2002GL016813.

IOC (Intergovernmental Oceanographic Commission), 2006. Manual on sea-level measurement and interpretation. Volume 4 - An update to 2006. Aarup, T., Merrifield, M., Perez, B., Vassie, I., Woodworth, P. (Eds.). IOC Manuals and Guides No. 14. Paris: Intergovernmental Oceanographic Commission, 80 pp.

Jansen, E., Overpeck, J., Briffa, K.R., Duplessy, J.-C., Joos, F., Masson-Delmotte, V., Olago, D., Otto-Bliesner, B., Peltier, W.R., Rahmstorf, S., Ramesh, R., Raynaud, D., Rind, D., Solomina, O., Villalba, R., Zhang, D., 2007. Palaeoclimate. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.), Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press: Cambridge, United Kingdom and New York, NY, USA, pp. 433-497.

Johannessen, O.M., Bengtsson, L., Miles, M.W., Kuzmina, S.I., Semenov, V.A., Alekseev, G.V., Nagurny, A.P., Zakharov, V.F., Bobylev, L.P., Pettersson, L.H., Hasselmann, K., Cattle, H.P., 2004. Arctic Climate Change – Observed and Modelled Temperature and Sea Ice. Tellus 56A, 328-341, doi:10.1111/j.1600-0870.2004.00060.x.

Jelgersma, S., 1961. Holocene sea level changes in The Netherlands. PhD Thesis, Rijks Universiteit Leiden, The Netherlands. Mededelingen van de Geologische Stichting C-IV, 7, 1-100.

Jevrejeva, S., Grinsted, A., Moore, J.C., Holgate, S. J., 2006. Nonlinear trends and multiyear cycles in sea level records, Journal of Geophysical Research 111, C09012, doi:10.1029/2005JC003229.

Jevrejeva, S., Moore, J.C., Grinsted, A., Woodworth, P.L., 2008. Recent global sea-level acceleration started over 200 years ago? Geophysical Research Letters 35, L08715, doi:10.1029/2008GL033611.

Kelley, J.T., Belknap, D.F., Daly, J.F., 2011. Comment on "North Atlantic climate-ocean variations and sea level in Long Island Sound, Connecticut, since 500 cal yr A.D.". Quaternary Research 55, 105-107, doi:10.1006/qres.2000.2194.

Kemp, A.C., Horton, B.P., Culver, S.J., Corbett, D.R., van de Plassche, O., Gehrels, W.R., Douglas, B.C., Parnell, A.C., 2009. Timing and magnitude of recent accelerated sea-level rise (North Carolina, United States). Geology 37, 1035-1038, doi:10.1130/G30352A.1.

Kemp, A.C., Horton, B.P., Donnelly, J.P., Mann, M.E., Vermeer, M., Rahmstorf, S., 2011. Climate related sea-level variations over the past two millennia. Proceedings of the National Academy of Sciences 108, 11017-11022, doi:10.1073/pnas.1015619108.

Kopp, R.E., Mitrovica, J.X., Griffies, S.M., Yin, J., Hay, C.C., Stouffer, R.J., 2010. The impact of Greenland melt on local sea levels: a partially coupled analysis of dynamic and static equilibrium effects in idealized water-hosing experiments. Climatic Change 103, 619-625, doi:10.1007/s10584-010-9935-1.

Lambeck, K., Yokoyama, Y., Purcell, T., 2002. Into and out of the Last Glacial Maximum: sea-level change during Oxygen Isotope Stages 3 and 2. Quaternary Science Reviews 21, 343-360, doi:10.1016/S0277-3791(01)00071-3.

Leclercq, P.W., Oerlemans, J. , Cogley, J.G., 2011. Estimating the glacier contribution to sea-level rise for the period 1800-2005. Surveys in Geophysics 32, 519-535, doi:10.1007/s10712-011-9121-7.

Leorri, E., Horton, B.P., Cearreta, A., 2008. Development of a foraminifera-based transfer function in the Basque marshes, N. Spain: Implications for sea-level studies in the Bay of Biscay. Marine Geology 251, 60-74, doi:10.1016/j.margeo.2008.02.005.

Llovel, W., Cazenave, A., Rogel, P., Lombard, A., Nguyen, M.B. 2009. Two-dimensional reconstruction of past sea level (1950–2003) from tide gauge data and an Ocean General Circulation Model. Climate of the Past 5, 217-227, doi:10.5194/cp-5-217-2009.

Marshall, W.A., Gehrels, W.R., Garnett, M.H., Freeman, S.P.H.T., Maden, C., Xu, S., 2007. The use of ‘bomb spike’ calibration and high-precision AMS ^{14}C analyses to date salt-marsh sediments deposited during the past three centuries. Quaternary Research 68, 325-337, doi:10.1016/j.yqres.2007.07.005.

Marzeion, B., Jarosch, A.H., Hofer, M., 2012. Past and future sea-level change from the surface mass balance of glaciers. The Cryosphere Discussions 6, 3177-3241, doi:10.5194/tcd-6-3177-2012.

Maul, G.A., Martin, D.G., 1993. Sea level rise at Key West, Florida, 1846-1992: America's longest instrument record? Geophysical Research Letters 20, 1955-1958, doi:10.1029/93GL02371.

Merrifield, M., Aarup, T., Allen, A., Aman, A., Caldwell, P., Bradshaw, E., Fernandes, R.M.S., Hayashibara, H., Hernandez, F., Kilonsky, B., Martin Miguez, B., Mitchum, G., Pérez Gómez, B., Rickards, L., Rosen, D., Schöne, T., Szabados, M., Testut, L., Woodworth, P., Wöppelmann, G. and Zavala, J., 2009. The Global Sea Level Observing System (GLOSS). In: Hall, J., Harrison, D.E., Stammer, D. (Eds.), Proceedings of OceanObs'09: Sustained Ocean Observations and Information for Society, Vol. 2, Venice, Italy, 21-25 September 2009, European Space Agency Publication WPP-306, doi:10.5270/OceanObs09.cwp.63. Downloadable from http://www.oceanobs09.net/cwp/.

Merrifield, M.A., Merrifield, S.T., Mitchum, G.T., 2009. An anomalous recent acceleration of global sea level rise. Journal of Climate 22, 5772-5781, doi:10.1175/2009JCLI2985.1.

Meyssignac, B., Cazenave, A., 2012. Sea level: a review of present-day and recent-past changes and variability. Journal of Geodynamics 58, 96-109, doi:10.1016/j.jog.2012.03.005.

Mitrovica, J.X., Tamisiea, M.E., Davis, J.L., Milne, G.A., 2001. Recent mass balance of polar ice sheets inferred from patterns of global sea-level change. Nature 409, 1026-1029, doi:10.1038/35059054.

Mörner, N.-A., 2010. Some problems in the reconstruction of mean sea level and its changes with time. Quaternary International 221, 3-8, doi: 10.1016/j.quaint.2009.10.044.

Nakada, M., Inoue, H., 2005. Rates and causes of recent global sea-level rise inferred from long tide gauge data records. Quaternary Science Reviews 24, 1217-1222, doi:10.1016/j.quascirev.2004.11.006

Nakiboglu, S.M., Lambeck, K., 1991. Secular sea-level change. In: Sabadini et al. (Eds.), Glacial Isostasy, Sea-level and Mantle Rheology. Kluwer Academic Publishers: The Netherlands, pp. 237-258.

Palmer, H.R. 1831. Description of a graphical register of tides and winds. Philosophical Transactions of the Royal Society of London 121, 209-213.

Peltier, W.R. 2004. Global glacial isostasy and the surface of the ice-age Earth: the ICE-5G (VM2) model and GRACE. Annual Review of Earth and Planetary Sciences 32, 111-149, doi:10.1146/annurev.earth.32.082503.144359.

Pugh, D.T., 1987. Tides, surges and mean sea-level: a handbook for engineers and scientists. Wiley, Chichester, 472pp.

Rahmstorf, S., Vermeer, M., 2011. Discussion of: Houston, J.R. and Dean, R.G., 2011. Sea-Level acceleration based on U.S. tide gauges and extensions of previous global-gauge analyses. Journal of Coastal Research 27, 747-787, doi:10.2112/JCOASTRES-D-11-00082.1.

Ray, R.D., Douglas, B.C., 2011. Experiments in reconstructing twentieth-century sea levels. Progress in Oceanography 91, 496-515, doi:10.1016/j.pocean.2011.07.021.

Redfield, A.C., Rubin, M., 1962. The age of salt marsh peat and its relation to recent changes in sea level at Barnstable, Massachusetts. Proceedings of the National Academy of Sciences 48, 1728-1735, doi:10.1073/pnas.48.10.1728.

Rignot, E., Box, J.E., Burgess, E., Hanna, E., 2008. Mass balance of the Greenland ice sheet from 1958 to 2007. Geophysical Research Letters 35, L20502, doi:10.1029/2008GL035417.

Sallenger, A.H. Jr, Doran, K.S., Howd, P.A., 2012. Hotspot of accelerated sea-level rise on the Atlantic coast of North America. Nature Climate Change, doi:10.1038/NCLIMATE1597.

Scott, D.B., Brown, K., Collins, E.S., Medioli, F.S., 1995. A new sea-level curve from Nova Scotia: evidence for a rapid acceleration of sea-level rise in the late mid-Holocene. Canadian Journal of Earth Sciences 32, 2071-2080, doi:10.1139/e95-160.

Scott, D. B., Medioli, F. S., 1978. Vertical zonations of marsh foraminifera as accurate indicators of former sea-levels. Nature 272**,** 528–531, doi:10.1038/272528a0.

Shennan, I., 1986. Flandrian sea-level changes in the Fenland, II: tendencies of sea level movement, altitudinal changes and local and regional factors. Journal of Quaternary Science 1, 155–179, doi:10.1002/jqs.3390010205.

Shennan, I., Horton, B.P., 2002. Relative sea-level changes and crustal movements of the UK. Journal of Quaternary Science 16, 511-526, 10.1002/jqs.3390040109.

Shennan, I., Woodworth, P.L., 1992. A comparison of late Holocene and twentieth-century sea-level trends from the UK and North Sea region. Geophysical Journal International 109, 96–105, doi:10.1111/j.1365-246X.1992.tb00081.x.

Sloss, C.R., Murray-Wallace, C.V., Jones, B.G., 2007. Holocene sea-level change on the southeast coast of Australia: a review. The Holocene 17, 999-1014, doi:10.1177/0959683607082415.

Southall, K.E., Gehrels, W.R., Hayward, B.C., 2006. Foraminifera in a New Zealand salt marsh and their suitability as sea-level indicators. Marine Micropaleontology 60, 167-179, doi:10.1016/j.marmicro.2011.03.001.

Stammer, D., Gregory, J., 2011. Understanding processes contributing to regional sea level change. EOS, Transactions American Geophysical Union 92, 328, doi:10.1029/2011EO390004. (This is a Meeting Report of a World Climate Research Programme/Intergovernmental Oceanographic Commission conference on regional sea level change. A fuller report is available from http://www.wcrp-climate.org/documents/Regional_SeaLevel_workshop.pdf).

Tamisiea, M.E., Mitrovica, J.X., Milne, G.A., Davis, J.L., 2003. Long wavelength sea level and solid surface perturbations driven by polar ice mass variations: fingerprinting Greenland

and Antarctic Ice Sheet flux. Space Science Reviews 108, 81-93, doi:10.1023/A:1026178014950.

Thomas, E., Varekamp, J.C., 1991. Paleo-environmental analyses of marsh sequences (Clinton, Connecticut): evidence for punctuated rise in relative sealevel during the latest Holocene. Journal of Coastal Research, Special Issue 11, 125-158.

Trenberth, K.E., Jones, P.D., Ambenje, P., Bojariu, R., Easterling, D., Klein Tank, A., Parker, D., Rahimzadeh, F., Renwick, J.A., Rusticucci, M., Soden, B., Zhai, P., 2007. Observations: Surface and Atmospheric Climate Change. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L., (Eds.), Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press: Cambridge, United Kingdom and New York, NY, USA, pp. 235-335.

Tsimplis, M.N., Baker, T.F., 2000. Sea level drop in the Mediterranean Sea: an indicator of deep water salinity and temperature changes? Geophysical Research Letters 27, 1731-1734, doi:10.1029/1999GL007004.

Van de Plassche, O., Bohncke, S.J.P., Makaske, B., van der Plicht, J., 2005. Water-level changes in the Flevo area, central Netherlands (5300-1500 BC): implications for relative mean sea-level rise in the western Netherlands. Quaternary International 133-134, 77-93, doi:10.1016/j.quaint.2004.10.009.

van Veen, J. 1945. Bestaat er een geologische bodemdaling te Amsterdam sedert 1700? Tijdschrift Koninklijk Nederlandsch Aardrijkskundig Genootschap 62, 2-37.

Wake, L.M., Huybrechts, P., Box, J.E., Hanna, E., Janssens, I., Milne, G.A., 2009. Surface mass-balance changes of the Greenland ice sheet since 1866. Annals of Glaciology 50, 178-184, doi: 10.3189/172756409787769636.

Watson, P.J., 2011. Is there evidence yet of acceleration in mean sea level rise around mainland Australia? Journal of Coastal Research 27, 368-377, doi:10.2112/JCOASTRES-D-10-00141.1.

Wenzel, M., Schröter, J. 2010. Reconstruction of regional mean sea level anomalies from tide gauges using neural networks. Journal of Geophysical Research 115, C08013, doi:10.1029/2009JC005630.

Woodroffe, S.A., 2009. Testing models of mid- to late Holocene sea-level change, north Queensland, Australia. Quaternary Science Reviews 28, 2474-2488, doi:10.1016/j.quascirev.2009.05.004.

Woodroffe, S.A., Horton, B.P. 2005. Holocene sea-level changes in the Indo-Pacific. Journal of Asian Earth Science 24, 29-43, doi:10.1016/j.jseaes.2004.01.009.

Woodworth, P.L. 1990. A search for accelerations in records of European mean sea level. International Journal of Climatology, 10, 129-143, doi: 10.1002/joc.3370100203.

Woodworth, P.L. 1999. High waters at Liverpool since 1768: the UK's longest sea level record. Geophysical Research Letters 26, 1589-1592, doi:10.1029/1999GL900323.

Woodworth, P.L., Player, R., 2003. The Permanent Service for Mean Sea Level: an update to the 21st century. Journal of Coastal Research 19, 287-295.

Woodworth, P.L., White, N.J., Jevrejeva, S., Holgate, S.J., Church, J.A., Gehrels, W.R., 2009. Evidence for the accelerations of sea level on multi-decade and century timescales. International Journal of Climatology 29: 777-789, doi:10.1002/joc.1771.

Woodworth, P.L., Menéndez, M., Gehrels, W.R., 2011a. Evidence for century-timescale acceleration in mean sea levels and for recent changes in extreme sea levels. Surveys in Geophysics 32, 603-618 (erratum page 619), doi:10.1007/s10712-011-9112-8.

Woodworth, P.L., Gehrels, W.R, Nerem, R.S., 2011b. Nineteenth and twentieth century changes in sea level. Oceanography 24, 80-93, doi:10.5670/oceanog.2011.29.

Woodworth, P.L. 2012. Tidal measurements. To be published in the History of Cartography. Volume 6. University of Chicago.

Wöppelmann, G., Pouvreau, N., Simon, B., 2006. Brest sea level record: a time series construction back to the early eighteenth century. Ocean Dynamics 56, 487-497, doi:10.1007/s10236-005-0044-z.

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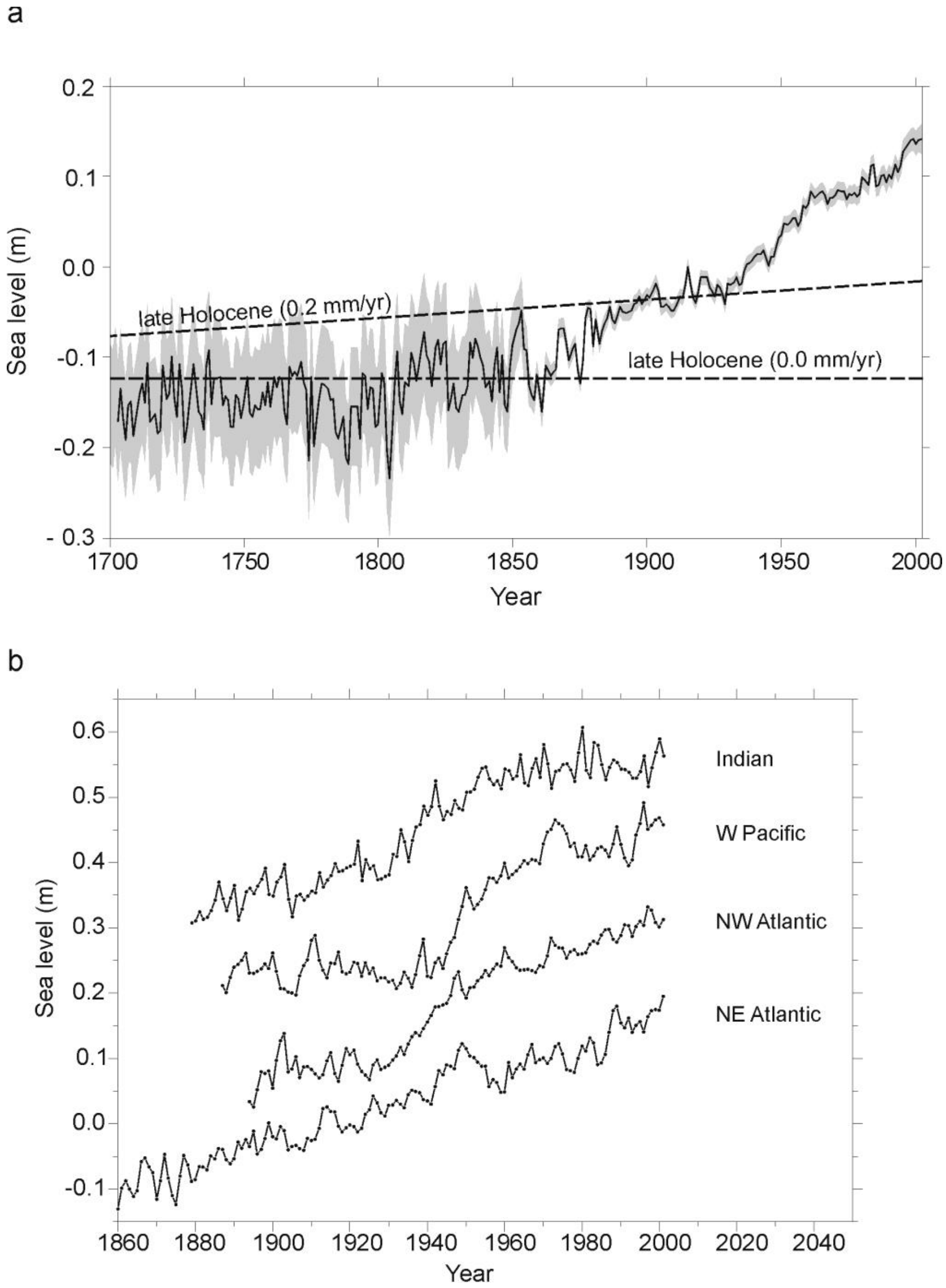

Figure 1. a. Global tide-gauge compilations from Jevrejeva et al. (2008) and Church and White (2011). Increased errors bands before 1850-1900 reflect the low number of tide-gauge records that cover the 18th and 19th centuries. The range of global late Holocene sea-level trends (0-0.2 mm/yr), as proposed by Jansen et al. (2007), is also shown. b. Compilations of tide-gauge records for four oceanic regions relevant to this study. From Milne et al. (2009), based on data from Jevrejeva et al. (2006).

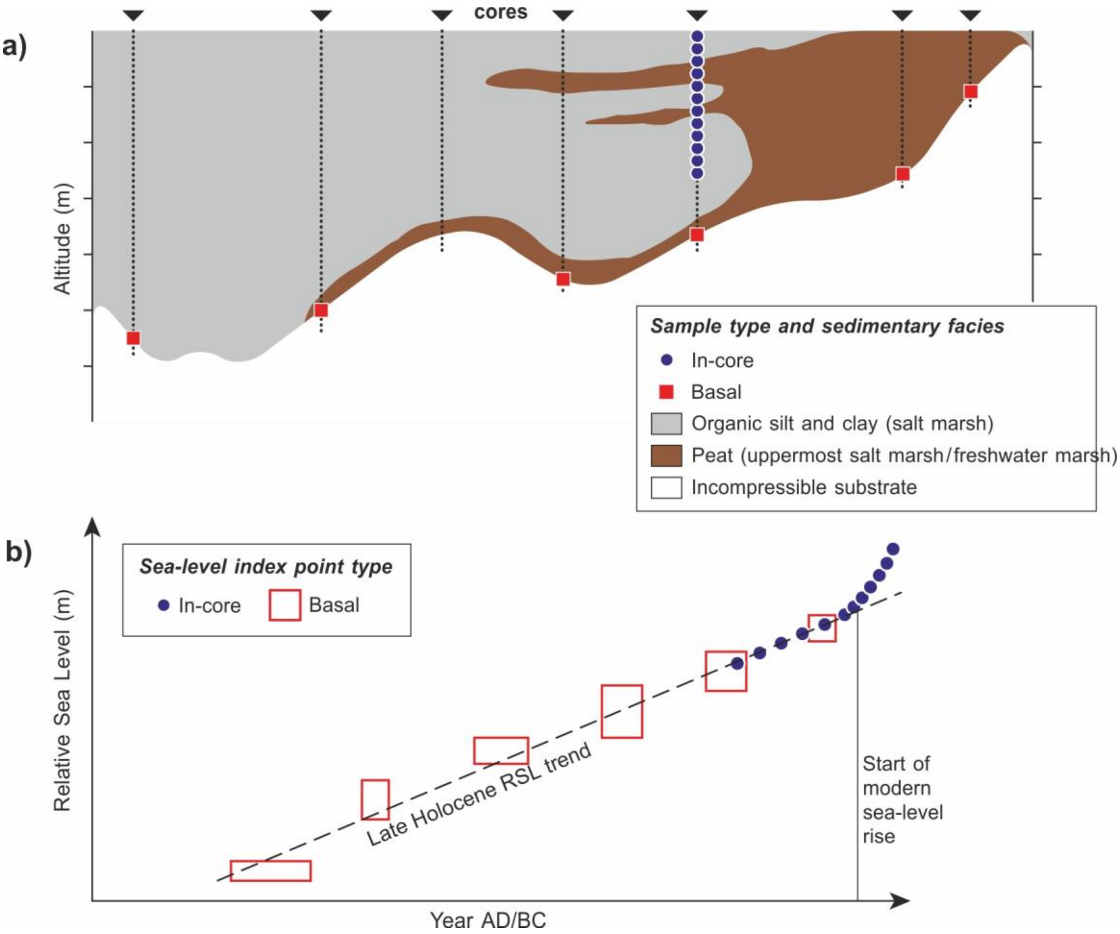

Figure 2. Salt-marsh stratigraphy and sea-level index points. a. Dates on basal sediments (squares) provide a long-term (millennial-scale) trend of sea-level change. The most recent record of sea-level change is captured by the upper part of the salt-marsh stratigraphy (dots). In the coastal lowlands of northwest Europe the salt-marsh units are typically very thin and largely replaced by tidal flat deposits. In these settings only millennium-scale sea-level reconstructions are possible from basal sea-level index points. b. Age-altitude graph of sea-level index points. The squares and dots correspond with the basal points and upper salt-marsh points in a, respectively. The sizes of the boxes reflect age and altitudinal errors. For the recent record (dots) errors are typically much smaller. In this paper we define the onset of modern rates of sea-level rise as the timing of the divergence of the recent sea-level record from the long-term trend. Figure is adapted from Engelhart et al. (2011b).

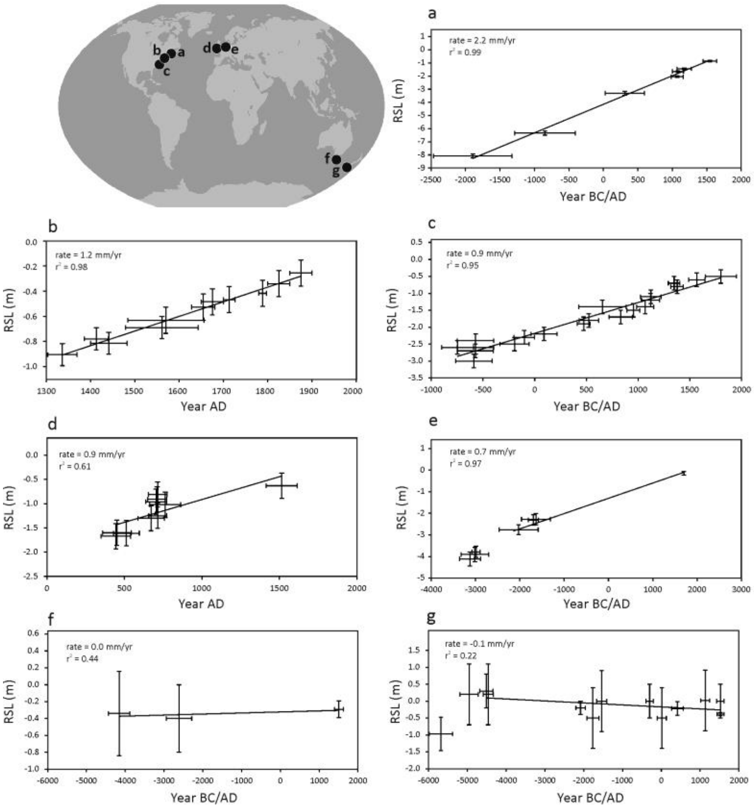

Figure 3. Late Holocene rates of sea-level rise for the seven coastal sites considered in this study as determined by linear regression through sea-level index points. Crosses reflect altitudinal and age uncertainties. a. Chezzetcook, Nova Scotia (Scott et al., 1995; Gehrels et al., 2004, 2005). b. Barn Island, Connecticut, USA (Donnelly et al., 2004). c. Sand Point, North Carolina, USA (Kemp et al., 2009, 2011; Engelhart et al. 2011a). d. Thurlestone, Devon, United Kingdom (Gehrels et al., 2011). e. Schokland, the Netherlands (van de Plassche et al., 2005). f. Little Swanport, Tasmania, Australia (Gehrels et al., 2012). g. Pounawea and Blueskin Bay, southeastern New Zealand (Gibb 1986; Gehrels et al., 2008). All sea-level index points from North Atlantic sites are from basal organic sediments (see Figure 2). Those from Tasmania and New Zealand are from intertidal shells found in tidal flat deposits and from salt-marsh sediments (the three dates with the smallest vertical errors). Only a few of these are basal index points.

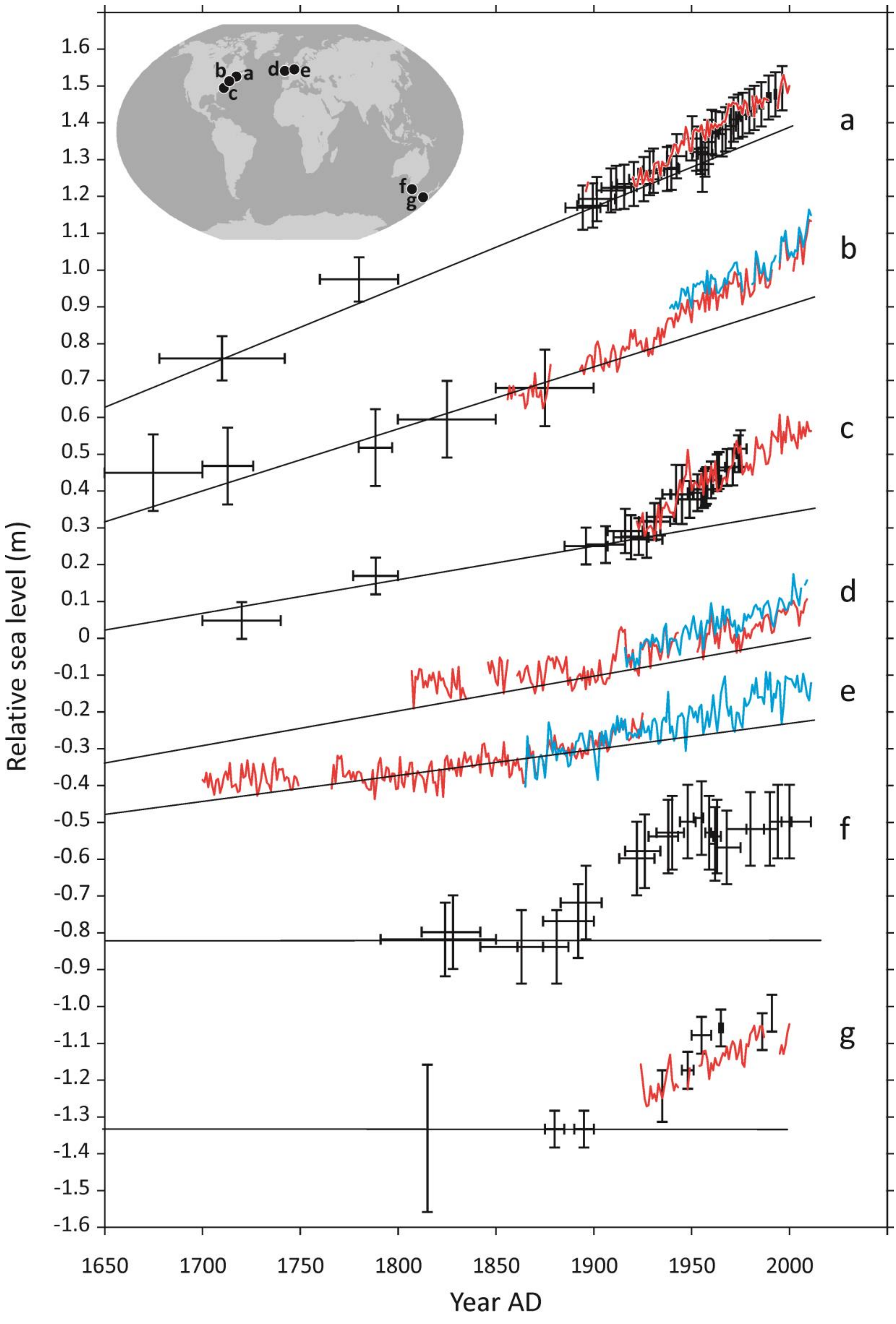

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Figure 4. Recent sea-level changes (last 200-350 years) compared with late Holocene background trend of sea-level change (from Figure 3). Red and blue lines are tide-gauge records. Sea-level index points from proxy data are shown as crosses reflecting age and altitudinal uncertainties. a. Chezzetcook, Nova Scotia (Gehrels et al., 2005), with tide-gauge record from Halifax. b. Barn Island, Connecticut, USA (Donnelly et al., 2004), with tide-gauge records from New York City (red line) and New London (green line). c. Sand Point, North Carolina, USA (Kemp et al., 2009, 2011), with tide-gauge record from Charleston, South Carolina. d. Tide-gauge records from Brest (red line) and Newlyn (green line) compared with late Holocene trend of relative sea-level change at Thurlestone, Devon, United Kingdom (Gehrels et al., 2011). e. Instrumental sea-level record from Amsterdam (red line) and tide-gauge record from Den Helder (blue line), compared with late Holocene trend of relative sea-level change at Schokland, the Netherlands (van de Plassche et al., 2005). f. Little Swanport, Tasmania, Australia (Gehrels et al., 2012). g. Pounawea southeastern New Zealand (Gehrels et al., 2008), with tide-gauge record from Lyttelton.

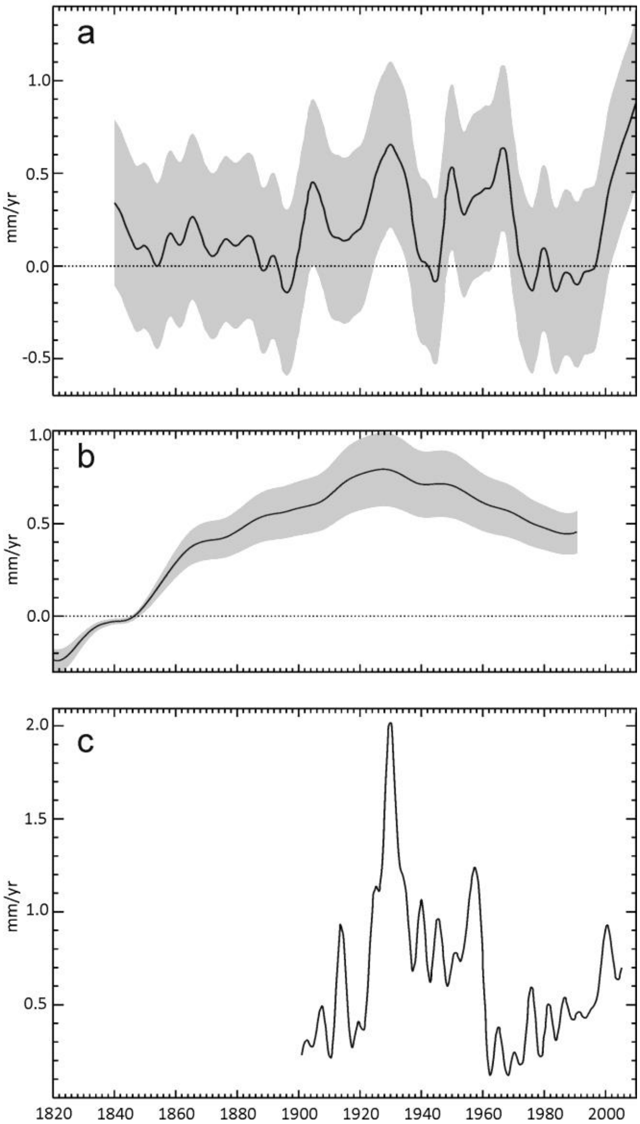

Figure 5. a. Contributions to global sea-level rise from the Greenland Ice Sheet reconstructed from modelling of surface mass balance and ice discharge (13 yr running average). The decade 1922-1932 produced 5.4 mm sea-level rise (Box et al., in review). b. Contributions to global sea-level rise from glaciers and ice caps reconstructed from glacier length records (Leclercq et al., 2011; Gregory et al., in revision). c. Contributions to global sea-level rise (5 yr running average) from glaciers in Alaska, Canada, western USA, Greenland, Iceland, Svalbard, Scandinavia and Russian Arctic (Marzeion et al., 2012). Uncertainty estimates in a and b are shown in grey shading. Uncertainties in c vary by region (see Marazeion et al., 2012).

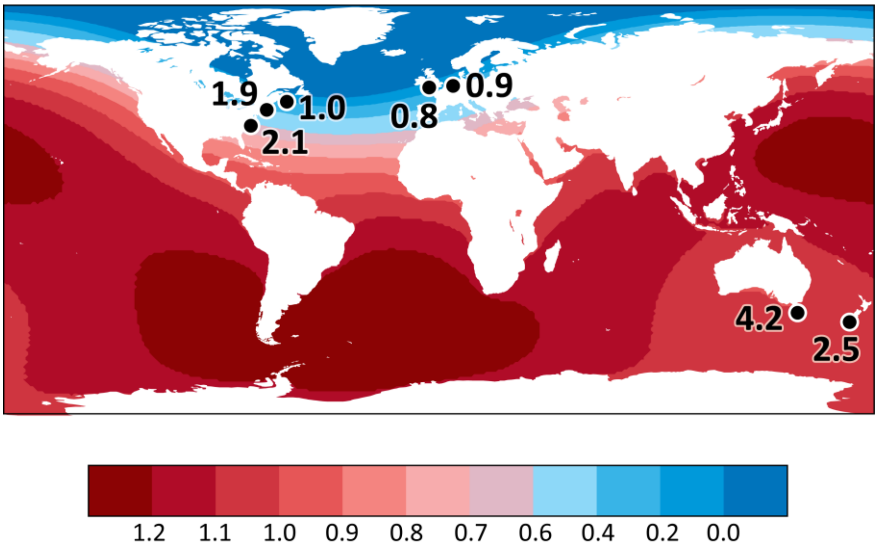

Figure 6. Model prediction (in mm) of how 1 mm of sea-level equivalent ice melt would be redistributed across the world's oceans were that water to come from the melting of the Greenland Ice Sheet (Mitrovica et al., 2001). Superimposed are differences between late Holocene and post-1930 rates of sea-level rise (mm/yr) from Table 3. The rate shown for Tasmania (4.2 mm/yr) is for the period 1900-1950 and is based on proxy data (Gehrels et al., 2012). Fastest increases of 20th century sea-level rise have occurred in the two southern hemisphere sites.

^{14}C lab code / marker	^{14}C age	Cal yr age BP (2σ range)	RSL	RSL error	Reference
Chezzetcook, Nova Scotia, Canada (44°44'N, 063°16'W)					
GX-18458	1735±120	1638.5 (1925-1353)	-3.3	0.15	Scott et al. (1995)
GX-18454	2710±155	2797.5 (3239-2356)	-6.34	0.15	Scott et al. (1995)
GX-5708	2495±115	3844 (4414-3274)	-8.08	0.15	Scott et al. (1995)
AA-47216	350±34	405.5 (308-503)	-0.86	0.06	Gehrels et al. (2004)
AA-47218	841±35	778 (674-882)	-1.45	0.06	Gehrels et al. (2004)
AA-47219	967±36	870 (789-951)	-1.65	0.06	Gehrels et al. (2004)
AA-47220	996±36	880.5 (794-967)	-2.05	0.06	Gehrels et al. (2004)
Barn Island, Connecticut, USA (41°20'N, 071°52'W)					

OS-26454	265±30	161.5 (153-170)	-0.42	0.10	Donnelly et al. (2004)
OS-26654	15±40	237 (224-250))	-0.47	0.10	Donnelly et al. (2004)
OS-27765	240±35	295.5 (269-322)	-0.53	0.10	Donnelly et al. (2004)
OS-26452	305±40	380 (294-466)	-0.63	0.10	Donnelly et al. (2004)
OS-29653	330±35	389 (307-471)	-0.69	0.09	Donnelly et al. (2004)
OS-27764	540±40	537 (510-564)	-0.78	0.09	Donnelly et al. (2004)
OS-33644	475±40	509 (468-550)	-0.82	0.09	Donnelly et al. (2004)
OS-29652	570±35	614 (581-647)	-0.91	0.09	Donnelly et al. (2004)
pollen (*Rumex*)	n/a	275 (250-300)	-0.49	0.10	Donnelly et al. (2004)
pollution (Cu, Pb)	n/a	75 (50-100)	-0.26	0.10	Donnelly et al. (2004)
pollen (*Plantago*)	n/a	125 (100-125)	-0.34	0.10	Donnelly et al. (2004)
Sand Point, North Carolina, USA (35° 53'N, 075° 41'W)					
OS-43066	185±30	150 (0–300)	-0.50	0.20	Engelhart et al. (2011a)
OS-43067	900±50	827 (727–927)	-1.10	0.20	Engelhart et al. (2011a)
OS-43068	1520±40	1427 (1333–1521)	-1.80	0.20	Engelhart et al. (2011a)
OS-43069	1920±45	1860 (1734–1986)	-2.20	0.20	Engelhart et al. (2011a)
OS-43070	2090±35	2051 (1951–2151)	-2.30	0.20	Engelhart et al. (2011a)
OS-43071	2420±35	2524 (2349–2699)	-2.70	0.20	Engelhart et al. (2011a)
OS-43266	2470±45	2538.5 (2363–2715)	-3.00	0.20	Engelhart et al. (2011a)
OS-58902	315±25	383 (305–461)	-0.60	0.20	Engelhart et al. (2011a)
OS-58897	535±30	572 (512–632)	-0.80	0.20	Engelhart et al. (2011a)
OS-58901	910±30	830 (743–917)	-1.20	0.20	Engelhart et al. (2011a)
OS-58896	1000±25	882 (800–964)	-1.40	0.20	Engelhart et al. (2011a)
OS-58713	1080±30	995 (933–1057)	-1.50	0.20	Engelhart et al. (2011a)
OS-58712	1190±30	1118 (1006–1230)	-1.70	0.20	Engelhart et al. (2011a)
OS-58711	1600±25	1475.5 (1413–1539)	-1.90	0.20	Engelhart et al. (2011a)
OS-58710	2120±25	2145 (2003–2287)	-2.50	0.20	Engelhart et al. (2011a)
OS-62716	2620±45	2696 (2543–2849)	-2.60	0.20	Engelhart et al. (2011a)
OS-64687	615±35	602 (546–658)	-0.70	0.20	Engelhart et al. (2011a)
OS-64688	2410±35	2522 (2346–2698)	-2.40	0.20	Engelhart et al. (2011a)
OS-64813	1390±110	1295 (1067–1523)	-1.40	0.20	Engelhart et al. (2011a)
OS-64689	2410±40	2522 (2345–2699)	-2.60	0.20	Engelhart et al. (2011a)
Thurlestone, SW England (50° 18'N, 003° 51'W)					
SUERC-20170	1321±35	1239 (1178-1300)	-0.91	0.26	Gehrels et al. (2011)
SUERC-20041	1310±35	1236 (1178-1294)	-1.25	0.26	Gehrels et al. (2011)
SUERC-20171	1385±37	1278 (1193-1363)	-1.3	0.26	Gehrels et al. (2011)
SUERC-20172	1306±37	1235.5 (1175-1296)	-0.81	0.26	Gehrels et al. (2011)
SUERC-20173	1342±37	1245.5 (1178-1313)	-0.96	0.26	Gehrels et al. (2011)
SUERC-20174	1270±37	1187 (1087-1287)	-1.02	0.26	Gehrels et al. (2011)
SUERC-20175	1539±35	1437.5 (1354-1521)	-1.61	0.26	Gehrels et al. (2011)
SUERC-23074	1610±35	1498.5 (1406-1591)	-1.6	0.26	Gehrels et al. (2011)
SUERC-23075	1619±35	1505 (1410-1600)	-1.67	0.26	Gehrels et al. (2011)
SUERC-23081	439±35	435.5 (335-536)	-0.63	0.26	Gehrels et al. (2011)

Schokland, Netherlands (52° 39’N, 005° 47’E)					
Amsterdam tide gauge	n/a	250	-0.15	0.05	Van Veen (1945)
GrA-16219/16225	3365±40	3588 (3481-3695)	-2.28	0.25	van de Plassche et al. (2005)
GrN-16381	3350±140	3647 (3321-3973)	-2.31	0.26	van de Plassche et al. (2005)
GrA-16216/16217	3655±40	3979 (3867-4091)	-2.76	0.24	van de Plassche et al. (2005)
GrN-16382	3740±160	4086.5 (3644-4529)	-3.14	0.23	van de Plassche et al. (2005)
GrA-12714	4340±50	4936.5 (4846-5027)	-3.79	0.26	van de Plassche et al. (2005)
GrN-15128	4330±70	4966 (4655-5277)	-3.91	0.33	van de Plassche et al. (2005)
GrN-15129	4420±100	5075.5 (4838-5313)	-4.12	0.33	van de Plassche et al. (2005)
Little Swanport, Tasmania, Australia (42° 21’S, 147° 56’E)					
SUERC-29114/29115	664±11	442 (329-555)	-0.29	0.1	Gehrels et al. (2012)
SUERC-28447	4414±37	4563 (4236-4890)	-0.4	0.4	Gehrels et al. (2012)
SUERC-28448	5677±38	6106 (5830-6382)	-0.34	0.5	Gehrels et al. (2012)
Pounawea, SE New Zealand (46° 29’S, 169° 41’E)					
Wk20397	1665±48	1533 (1382-1684)	-0.22	0.2	Gehrels et al. (2008)
Wk15813	3756±36	4032 (3911-4135)	-0.2	0.2	Gehrels et al. (2008)
Beta 20652	413±23	425 (335-515)	-0.385	0.05	Gehrels et al. (2008)
NZ5270	413±30	425.5 (331-520)	0	0.5	Gibb (1986)
NZ6485	907±62	814 (698-930)	0.02	0.9	Gibb (1986)
NZ1973	1970±50	1935 (1818-2052)	-0.5	0.9	Gibb (1986)
NZ5269	2250±50	2249 (2151-2347)	0	0.5	Gibb (1986)
NZ1975	3240±60	3487 (3359-3615)	0	0.9	Gibb (1986)
NZ1974	3440±60	3714 (3560-3868)	-0.5	0.9	Gibb (1986)
NZ1978	5600±70	6413.5 (6281-6546)	0.2	0.9	Gibb (1986)
NZ1976	5640±70	6461 (6295-6627)	0.3	0.5	Gibb (1986)
NZ1977	6000±70	6905.5 (6668-7143)	0.2	0.9	Gibb (1986)
NZ6484	6750±150	7633.5 (7334-7933)	-0.97	0.5	Gibb (1986)

Table 1. Proxy data used to calculate late Holocene trends of sea-level change in our study sites (Figure 3).

Year AD	Age error (yr)	RSL (m)	RSL error (m)	dating method/marker
Chezzetcook, Nova Scotia, Canada (Gehrels et al. 2005)				
1996	0.2	-0.01	0.06	^{210}Pb
1993	0.4	-0.02	0.06	^{210}Pb
1989	1.0	-0.03	0.06	^{210}Pb
1986	1.1	-0.05	0.06	^{210}Pb
1982	1.0	-0.06	0.06	^{210}Pb
1979	1.3	-0.07	0.06	^{210}Pb
1976	1.8	-0.08	0.06	^{210}Pb
1974	1.2	-0.08	0.06	^{210}Pb
1971	1.9	-0.09	0.06	^{210}Pb
1970	2.6	-0.11	0.06	^{210}Pb
1967	2.8	-0.12	0.06	^{210}Pb
1966	1.8	-0.13	0.06	^{210}Pb
1962	1.6	-0.13	0.06	^{210}Pb
1960	2.5	-0.12	0.06	^{210}Pb
1959	3.6	-0.15	0.06	^{210}Pb
1957	3.7	-0.19	0.06	^{210}Pb
1956	3.2	-0.23	0.06	^{210}Pb
1955	3.9	-0.18	0.06	^{210}Pb
1953	3.1	-0.17	0.06	^{210}Pb
1950	4.5	-0.14	0.06	^{210}Pb
1942	4.9	-0.19	0.06	^{210}Pb
1940	3.4	-0.22	0.06	^{210}Pb
1938	6.1	-0.23	0.06	^{210}Pb
1931	5.5	-0.23	0.06	^{210}Pb
1929	5.9	-0.24	0.06	^{210}Pb
1926	7.5	-0.25	0.06	^{210}Pb
1920	8.5	-0.27	0.06	^{210}Pb
1916	6.4	-0.27	0.06	^{210}Pb
1912	7.6	-0.28	0.06	^{210}Pb
1909	7.3	-0.28	0.06	^{210}Pb
1902	9.3	-0.31	0.06	^{210}Pb
1899	7.9	-0.32	0.06	^{210}Pb
1894	8.9	-0.33	0.06	^{210}Pb
1780	20.0	-0.53	0.06	pollen
1710	32.0	-0.74	0.06	^{14}C
Sand Point, North Carolina, USA (Kemp et al. 2009, 2011)				
1975	3.0	-0.06	0.05	^{210}Pb
1974	1.0	-0.07	0.05	^{14}C
1971	4.0	-0.11	0.05	^{210}Pb
1968	4.0	-0.11	0.05	^{210}Pb
1964	4.0	-0.12	0.05	^{210}Pb
1963	0.0	-0.12	0.05	^{137}Cs
1960	5.0	-0.14	0.05	^{210}Pb

1958	0.4	-0.16	0.05	^{14}C
1957	0.7	-0.16	0.05	^{14}C
1956	5.0	-0.17	0.05	^{210}Pb
1953	5.0	-0.18	0.05	^{210}Pb
1949	6.0	-0.19	0.05	^{210}Pb
1945	5.6	-0.18	0.08	^{14}C
1945	6.0	-0.18	0.08	^{210}Pb
1942	7.0	-0.18	0.08	^{210}Pb
1934	7.0	-0.24	0.05	^{210}Pb
1931	8.0	-0.26	0.05	^{210}Pb
1927	8.0	-0.30	0.05	^{210}Pb
1923	9.0	-0.30	0.05	^{210}Pb
1919	9.0	-0.30	0.06	^{210}Pb
1916	9.0	-0.28	0.06	^{210}Pb
1906	10.0	-0.32	0.05	^{210}Pb
1896	11.0	-0.32	0.05	^{210}Pb
1789	11.5	-0.40	0.05	^{14}C
1720	20.0	-0.52	0.05	pollen

Little Swanport, Tasmania, Australia (Gehrels et al. 2012)

2000	7.5	0.03	0.10	^{210}Pb
1994	5.5	0.03	0.10	^{14}C
1994	5.5	0.03	0.10	^{210}Pb
1990	8.0	0.01	0.10	^{14}C
1990	8.0	0.01	0.10	^{210}Pb
1990	8.0	0.01	0.10	geochemistry
1980	9.5	0.01	0.10	^{210}Pb
1968	6.0	-0.04	0.10	^{210}Pb
1963	2.0	-0.01	0.10	^{14}C
1963	2.0	-0.01	0.10	^{210}Pb
1963	2.0	-0.01	0.10	^{137}Cs
1962	2.0	-0.03	0.10	^{210}Pb
1962	2.0	-0.03	0.10	^{14}C
1959	1.5	0.00	0.10	^{14}C
1959	1.5	0.00	0.10	^{210}Pb
1955	2.5	0.04	0.10	^{14}C
1948	4.0	0.03	0.10	geochemistry
1940	7.0	0.00	0.10	^{210}Pb
1938	7.5	-0.01	0.10	^{210}Pb
1938	7.5	-0.01	0.10	^{14}C
1926	9.0	-0.05	0.10	geochemistry
1922	9.0	-0.07	0.10	^{14}C
1896	10.5	-0.19	0.10	^{14}C
1892	13.0	-0.24	0.10	geochemistry
1881	13.0	-0.31	0.10	^{14}C
1863	16.0	-0.31	0.10	pollen, geochemistry

1828	15.0	-0.27	0.10	pollen, charcoal
1824	29.5	-0.29	0.10	^{14}C
Pounawea, New Zealand (Gehrels et al. 2008)				
1991	0	0.03	0.05	geochemistry
1986	0	-0.02	0.05	geochemistry
1965	1	-0.01	0.05	^{137}Cs
1955	5	-0.03	0.05	pollen
1948	3	-0.125	0.05	geochemistry
1935	0	-0.195	0.05	charcoal
1895	5	-0.285	0.05	geochemistry
1880	5	-0.285	0.07	pollen
1815	0	-0.31	0.05	geochemistry

Table 2. Proxy data used to reconstruct recent sea-level change in four of our sites (Figure 4a,c,f,g).

	Inflexion	Late Holocene rate (mm/yr)	Post-1930 rate (mm/yr)	Tide gauge	Reference(s)
NW Atlantic					
Nova Scotia*	1930-1940	2.2	3.2	Halifax	Scott et al., (1995), Gehrels et al., (2005)
Connecticut*	1925-1935	1.2	3.1	New York City	Donnelly et al. (2004)
North Carolina*	1925-1935	0.9	3.0	Charleston	Kemp et al., (2009), (2011)
Tide gauges□	1925-1935	n/a	2.7	many	Jevrejeva et al., (2006), Woodworth et al., (2009); Milne et al., (2009)
NE Atlantic					
SW England*	1920-1940	0.9	1.7	Newlyn	Gehrels et al., (2011), Wōppelmann et al., (2006)
Netherlands*	1905-1915	0.7	1.6	Den Helder	Van de Plassche (2005), Woodworth et al., (2011a)
Tide gauges□	none	n/a	1.8	many	Jevrejeva et al., (2006), Woodworth et al., (2009), Milne et al., (2009)
SW Pacific and Tasman Sea					
SE New Zealand*	1895-1925	-0.1	2.4	Lyttelton	Gehrels et al., (2008)
Tasmania*	1895-1920	0.0	n/a	none	Gehrels et al., (2011)
W Pacific tide gauges□	1930-1945	n/a	3.6	many	Jevrejeva et al., (2006), Woodworth et al., (2009); Milne et al., (2009)
Indian Ocean tide gauges□	1925-1935	n/a	1.8	many	Jevrejeva et al., (2006), Woodworth et al., (2009); Milne et al., (2009)
Global					
Tide gauges	1935	0.0-0.2	1.8	many	Jansen et al., (2007), Church and White (2011)

Table 3. Onset of recent rapid sea-level rise ('inflexion') determined from proxy sea-level records, tide gauges near proxy sites, and compilations of regional tide-gauge records. The post-1930 rates are calculated from tide-gauge records that are listed in the next column (www.psmsl.org). The tide-gauge compilation for the northeast Atlantic does not have a clear inflexion, but the longest individual records do (see text for discussion). Sites with proxy data are marked by an asterisk (*) and regional tide gauge compilations by a square (□).