# Relative sea-level rise and climate change over the last 1500 years

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## ABSTRACT

We constructed a detailed relative sea-level rise curve for the last 1500 years using a novel approach, i.e. charting the rate of relative sea-level rise using microfaunal and geochemical data from a coastal salt marsh sequence (Clinton, CT, USA). The composition of benthic foraminiferal assemblages and the iron abundance in peats were used to describe shifts in marsh environment through time quantitatively. The resulting sea-level rise curve, with age control from $^{14}$C dating and the onset of anthropogenic metal pollution, shows strong increases in the rate of relative sea-level rise during modern global warming (since the late nineteenth century), but not during the Little Climate Optimum (AD 1000–1300). There was virtually no rise in sea-level during the Little Ice Age (AD 1400–1700). Most of the relative sea-level rise over the last 1200 years in Clinton appears to have occurred during two warm episodes that jointly lasted less than 600 years. Changes from slow to fast rates of relative sea-level rise apparently occurred over periods of only a few decades. We suggest that changes in ocean circulation could contribute to the sudden increases in the rate of relative sea-level rise along the northeastern USA seaboard. Relative sea-level rise in that area is currently faster than the worldwide average, which may result partially from an ocean surface effect caused by hydrodynamics. Our data show no unequivocal correlation between warm periods (on a decaal to centennial time-scale) and accelerated sea-level rise. One period of acclerated sea-level rise may have occurred between about AD 1200 and 1450, which was the transition for the Little Climate Optimum to the Little Ice Age, i.e. a period of cooling (at least in northwestern Europe). Local changes in tidal range might also have contributed to this apparent increase in the rate of relative sea-level, however. The second period of accelerated sea-level rise occurred during the period of modern global warming that started at the end of the last century.

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## INTRODUCTION

Sea-level rise (SLR) is thought to be one of the consequences of global warming, and different scenarios exist to predict future SLR in response to antropogenic greenhouse warming (IPCC report, 1990; Lambeck, 1990; Gornitz, 1991). Unequivocal evidence of the response of SLR to short-term (< $10^3$ yr) climate oscillations is rare, (Fairbridge and Hillaire-Marcel, 1977; Shennan, 1986; Van de Plassche and Roep, 1989). Tide gauge records provide detailed relative SLR (RSLR) records for several locations, mainly on the northern hemisphere, over the last 200 years (Gornitz *et al.*, 1982; Barnett, 1988; Meier, 1984; Wigley and Raper, 1987; Pirazzoli, 1989). These records, however, are too short to derive meaningful correlations between climate and RSLR. We developed new methods to chart the rate of RSLR using tidal salt marsh sequence (Thomas and Varekamp, 1991; Van de Plassche, 1991). In this paper we provide a detailed record of RSLR over the last 1500 years, based on a study of salt marsh sequences in Clinton (Connecticut, USA). We cannot correlate fluctuations in the rate of RSLR unequivocally with periods of global warming and cooling, and speculate on the nature of the linkage between climate change and RSLR on a decadal to centennial time scale.

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## FLUCTUATIONS IN THE RATE OF RELATIVE SEA-LEVEL RISE: EVIDENCE FROM SALT MARSH SEQUENCES

Many sea-level rise studies have as their conceptual basis (or tacitly assume) that coastal salt marsh accretion keeps up with RSLR. Depth–age data from peat sequences then provide an envelope for a RSLR curve, which is commonly fitted with line segments or with a polynomial function (Pinter and Gardner, 1989) on scattered observations (Field *et al.*, 1979; Kraft *et al.*, 1987). Other studies use basal peats only, in an array landinwards. These RSLR curves are more precise (Van de Plassche *et al.*, 1989; Scott and Medioli, 1986; Scott *et al.*, 1987), but lack detail in the documentation of fluctuations in the rate of RSLR, as a result of inherent uncertainties in $^{14}$C dating. Our method of charting RSLR from salt marsh sequences derives information from the

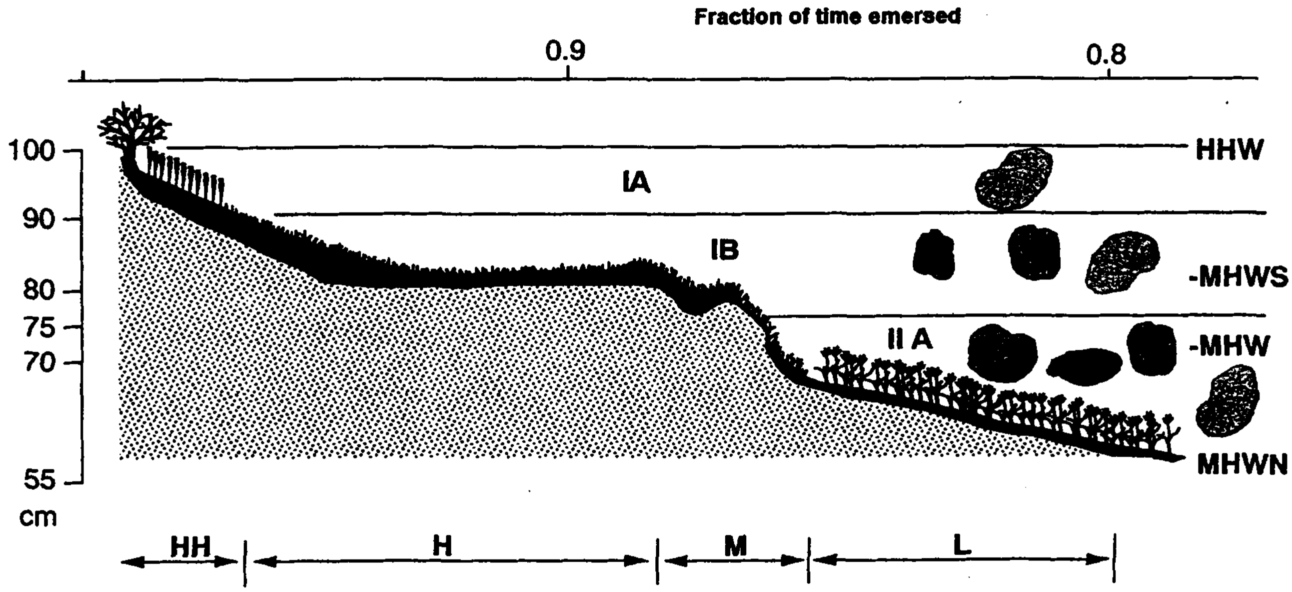



**Fig. 1.** *Cross section of a coastal salt marsh in New England. Indicated are the floral zones (L = low marsh with predominantly* Spartina alterniflora, *M = middle marsh, H = high marsh with predominantly* Spartina patens *and* Distichlis spicata, *and HH = highest high marsh with abundant* Phragmites*) and microfaunal zones with foraminiferal assemblages (IA =* T. macrescens; *IB:* T. macrescens, T. comprimata, H. manilaensis; *IIA:* T. macrescens, T. inflata, M. fusca, T. comprimata, A. mexicana*) and the approximate flooding frequency, shown as fraction of time of emersion (adapted from Gordon, 1980).*

disequilibrium between marsh growth and RSLR: the marsh drowns when the rate of RSLR exceeds the marsh accretion rate, whereas it becomes dominated by high and highest marsh environments when RSLR is slow.

The salt marsh environment is subdivided into lower, middle, high and highest marsh zones which may stretch over a vertical distance of several metres, depending on the local tidal range (e.g. Nixon, 1982; Long and Mason, 1983). These sub-environments differ in lithological, macrofloral, microfaunal and geochemical characteristics (Fig. 1; Scott and Medioli, 1980). Lithostratigraphic and floral studies of marsh sequences document regressive and transgressive tendencies (Van de Plassche, 1991), but the vertical magnitude of the shifts in depositional environment cannot be quantified precisely. We concluded that flooding frequency (Fig. 1) is a controlling parameter of the macrofaunal and geochemical signatures that characterize the subenvironments (Thomas and Varekamp, 1991).

The microfaunal zonation of salt marshes is based on assemblages of benthic, agglutinated foraminifera, which can be used to locate former sea-levels in subsurface sediments (Scott and Medioli, 1978; 1980; 1986; Scott and Leckie, 1990; Scott *et al.*, 1977; 1980; 1987; Fig. 1). The highest high water datum level can be most accurately located: the highest faunal subzone (IA, Fig. 1) has a vertical range of only about 10 cm. The species *Trochammina macrescens* is a key indicator for this subzone, and can survive the most prolonged periods of atmospheric exposure between flooding events. Closer to mean high water (MHW) the abundance of *T. macrescens* decreases rapidly, and more diverse assemblages are found, with *Milliammina fusca* as an indicator species for the middle to low marsh subenvironments. The relative abundance of *T. macrescens* is thus inversely related to the flooding frequency and the relative abundance of species other than *T. macrescens* is the parameter 'other species', used as a proxy for flooding frequency. Depositional environments in faunal zone II, especially subzone IIB (not recognized in our samples) can not be located as accurately relative to MHW, because of their larger vertical range. With increased tidal range the vertical extent of the uppermost three subzones (IIA, IB, and IA) changes little, because most of the difference is absorbed with faunal subzone IIB (Scott and Medioli, 1980).

The geochemistry of marsh sediments is determined by the abundance of clastic materials, the accretion rate of the organic matrix, diffusional and diagenetic processes, and atmospheric deposition (Given, 1975; Varekamp, 1991). Most clastic matter is deposited on mudflats, although fine sand is common in the matrix between plant rootlets in lower salt marsh peats (Stevenson *et al.*, 1988). The abundance of organic matter may therefore serve as a crude parameter for the flooding frequency at a given site, but depends on marsh accretion rate and hydrodynamic regime. Fine-grained, suspended sediments may be trapped efficiently by the marsh flora during flooding. The abundance of very fine particulates in peat sequences can be used as an indicator for flooding frequency or for the volume of sea water that washes over a given marsh zone per time unit. Very fine matter is the main host for metals in most clastic sediments (Foerstner and Wittmann, 1983); it consists partially of hydrous Fe-oxides with adsorbed heavy metals, precipitated in the estuarine mixing zone (e.g. Lee, 1975; Hem, 1975; Turekian, 1977). The Fe and Zn concentrations in marsh sediments are excellent proxies for the abundance of fine-grained matter in the marsh sediments and therefore of flooding frequency.

Analysis of peat cores for benthic foraminifera and metal abundances provides information on shifts in depositional environment during development of a marsh. These shifts may result from changes in the morphology of the marsh (e.g. spit growth), changes in tidal range or marsh accretion rate, or vertical movement of the land (e.g. seismic events; Nelson and Jennings, 1988). In many cases one can distinguish between these different processes, and we interpret the environmental changes as observed in the Clinton marshes to be a result of fluctuations in the rate of RSL, because of the absence of evidence for changes in marsh morphology or sudden tectonic movements.

We charted changes in depositional environment at coring sites over time as fluctuations in vertical distance from MHW, which is sensitive to RSLR as well as to changes in tidal range. The vertical distances from MHW for all core samples were derived from foraminiferal data and Fe abundances. We constructed a general sequence of drowning and emergence, which was put into a $^{14}C$ time frame. We then derived a RSLR curve with a high resolution for the sequence of events. The relatively poor calibration of $^{14}C$ ages for the last 2 millennia (Stuiver and Pearson, 1986) makes the absolute time control on the drowning and emergence events in the marshes not precise. In the upper part of the sequence, the anthropogenic input of metals such as Cu and Zn provides a useful marker of the AD 1860 level (McCaffrey and Thomson, 1980; Varekamp, 1991; Allen and Rae, 1987, 1988), but Zn abundances in the peats have no palaeo-environmental significance above that level because of these increased atmospheric and aqueous pollution inputs.

## STUDY LOCATION, METHODS AND RESULTS

We studied peat cores from the Hammock river marshes near Clinton (CT, USA, Fig. 2; Thomas and Varekamp, 1991; Van de Plassche, 1991; Varekamp, 1991). Cores for faunal and geochemical analysis were taken in two lobes of the marsh (Fig. 2), and sliced into a continuous set of samples. Up to five core sections of intervals with significant changes in environment were studied from other locations in the marsh (Thomas and Varekamp, 1991). Two cores, taken about 40 cm apart at the F locality, were sampled at the same depths for faunal and chemical studies, with a depth resolution for correlation of cores of about 5 cm. Two other cores were taken within 50 cm (to the north-

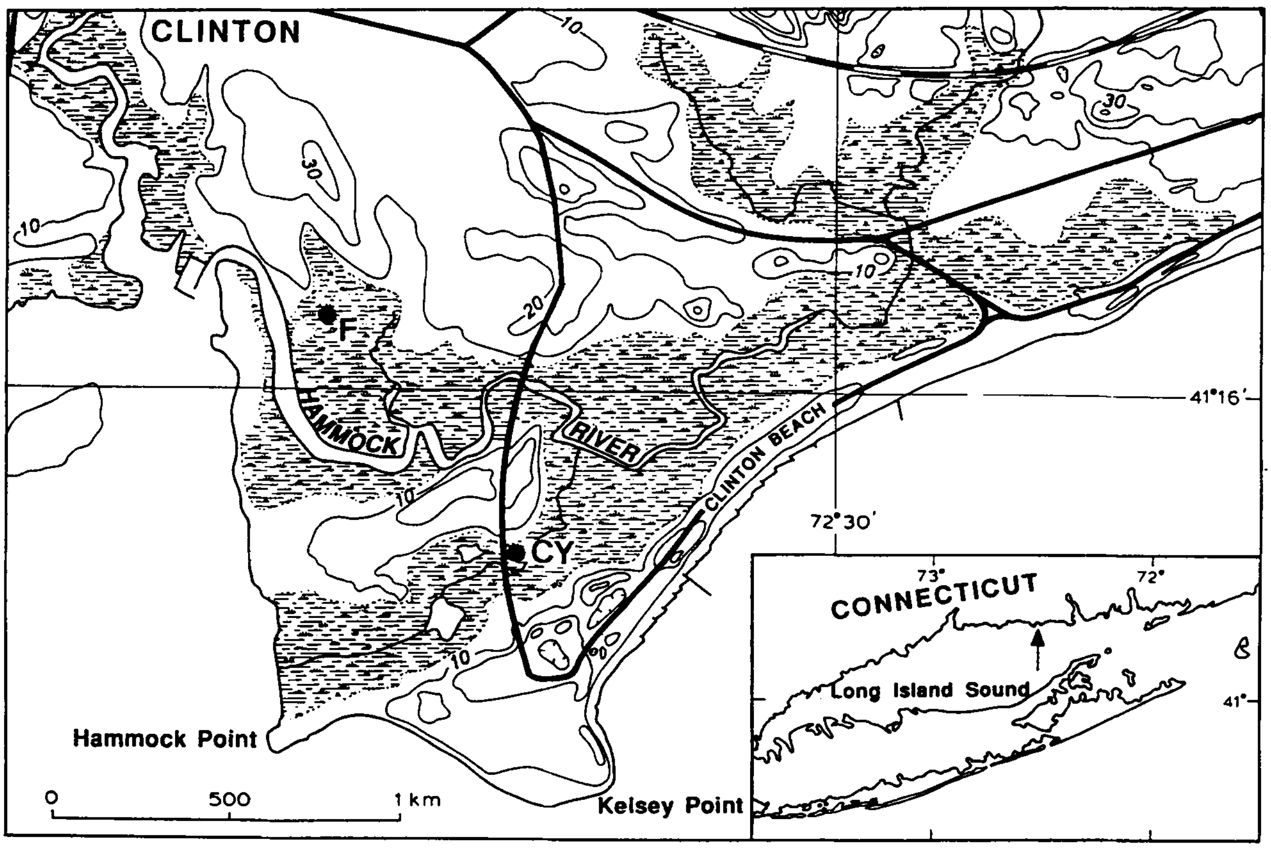



**Fig. 2.** *The Hammock River marshes near Clinton, Connecticut, USA. Indicated are the two marsh lobes with the coring sites F and CY.*

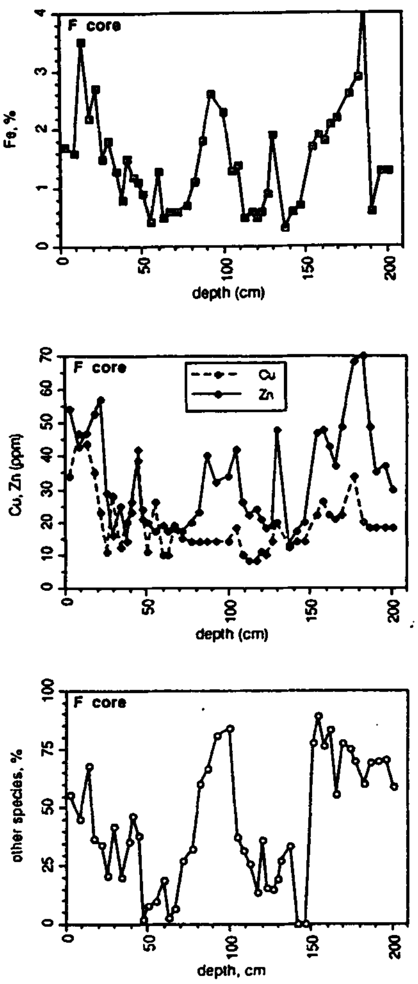



**Fig. 3.** *Relative abundance of species other than* T. macrescens *('other species' parameter), and Fe and Zn contents of core F.*

east and to the east). Six samples from the core taken to the northeast, and one sample (Table 1, UtC-2002) from the core taken to the east, were used for radiocarbon analyses by AMS. For radiocarbon analysis cores were split and each half was searched for substrate stem remains of *Spartina alterniflora, Spartina patens, Distichlis spicata, Scirpus robustus,* and *Phragmites australis.* Rhizomes were excluded carefully from analysis because of their unknown depth of penetration to deeper levels in the soil. Subsurface stem remains indicate a position about 1–5 cm above the contemporaneous marsh surface; we did not correct for this small offset in vertical distance. Samples were cleaned ultrasonically to remove clay and amorphous organic material. Younger rootlets were removed from the sample using a binocular microscope. All samples were analysed by AMS at the Robert van der Graaff laboratory (University of Utrecht, Netherlands).

Peat samples for geochemical analysis were dried, gently pulverized, and the fraction < 180 μm was leached with acids. The leachates were analysed for Fe, Zn, Cu, S and the solid sample (< 180 μm) for approximate organic carbon contents by resp. Atomic Absorption Spectroscopy, Ion Chromatography and weight-loss-upon-combustion.

Quantitative faunal analysis (> 63 μm fraction) showed intervals dominated by *T. macrescens* alternating with intervals with a more abundant and diverse foraminiferal assemblages (Fig. 3). Fluctuations in abundance of Fe, Zn and S are strongly positively cross-correlated. Variations in Cu and 'organic carbon' values are less systematic (Thomas and Varekamp, 1991; Varekamp, 1991). The parameter 'other species' is positively correlated with the Fe and Zn abundances in both cores ($R_{Fe} = 0.63$; $R_{Zn} = 0.68$), and between the two cores indicating fluctations in depositional regime with depth-in-core (Fig. 3). The coherence between the chemical and faunal records suggests that the fluctuations in Fe and Zn abundances are largely depositional. The correlation of the faunal parameter with the sulphur abundance is largely an indirect effect of the depositional environment: the extent of diagenetic S-fixation in this environment rich in organic matter is mostly limited by the Fe abundance (Varekamp, 1991; Thomas and Varekamp, 1991).

## A RSLR CURVE FOR THE CLINTON AREA (CT, USA)

Faunal and chemical records indicate sudden transgressions at about 140, 110 and 50 cm depth (Fig. 3), followed by more gradual recoveries of the marsh. The changes in depositional environment can be quantified by application of the vertical spacing of the foraminiferal assemblages in the marshes with respect to HHW (Scott and Medioli, 1981; Thomas and Varekamp, 1991) and through the use of a transfer function for the Fe abundances. We assumed a constant tidal range equal to the modern mean tidal range (1.45 m; Harrison and Bloom, 1977), and derived marsh palaeoenvironment (MPE) curves from the microfaunal data. We then calibrated the Fe contents of the peats with distance from MHW (Fig 4): the lowest measured Fe content represents the zone near the HHW line, whereas the peak of the transgression at about 90 cm depth in the F core was taken as representative for the environment near the MHWN level, as derived from the foraminiferal data. These two data points were then fitted to a curve that represents the fraction of emersion time for the marsh environments (Fig. 1), and the curve was subsequently slightly adjusted to samples that plot on the transition from the IB-IIA faunal subzones. The distances to MHW for each sample were then graphically derived from this calibration curve.

The MPE curves (fauna-based MPE-1, and Fe-based, MPE-2; Fig. 5) delineate the shifting environment of deposition at the coring sites through time. The two curves derived for the F-core are very similar in shape and only differ slightly in amplitude in the uppermost part. This difference in amplitude may result from the more complex dependence of sediment chemistry on bulk sediment make-up or be influenced by an enhanced Fe flux in modern times. These differences may be resolved through the use of additional element indicators of flooding frequency (e.g. Mo; Thomas and Varekamp, 1991). The MPE curves for the CY and F sites show great similarity in the sequence of events, although there are slight differences in depth-in-core for some events (Fig. 5). This similarity suggests that the changes in subenvironments are not the result of extremely local effects, such as the changing course of a channel. Correlation of our MPE curves with stratigraphic analyses of hundreds of cores in the whole marsh confirms that the environmental changes occurred marshwide (Van de Plassche, 1991).

A transgression occurred at 190 cm (F-core), followed by a gentle shoaling in MPE-2. A sudden change occurs near 150 cm, with emergence of the marsh followed by a sudden drowning: the relative position of the coring site with

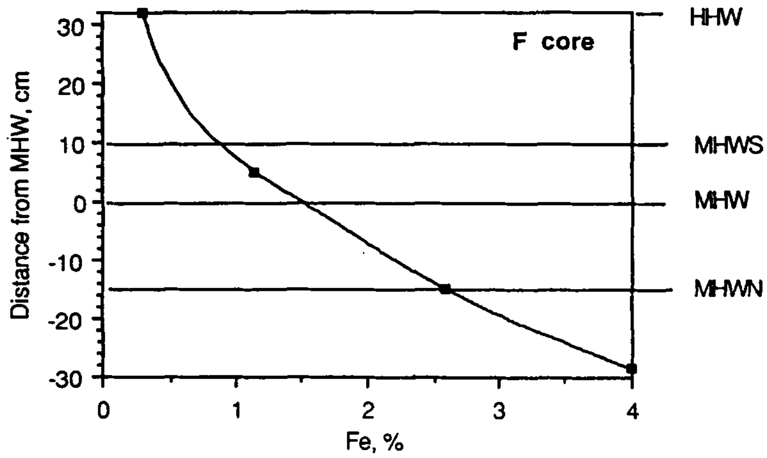



**Fig. 4.** *Calibration of Fe contents with vertical location with respect to MHW, as derived from correlation with foraminifera data from three samples (squares) and the shape of the flooding frequency curve.*

respect to MHW changed by about 40 cm vertically over less than 5 cm core in the foraminifer-based curve. The subsequent transgression created a depositional environment close to that existing before the regression. After this transgression, shoaling occurred, followed by a major, more gradual transgressive episode at about 110 cm depth. An initial drowning of 25–35 cm took place, followed by a long period of marsh recovery until the marsh was again close to emergence (highest high marsh) at about 50 cm depth in the core. At this level, a another transgression took place, followed by a short regressive period in the MPE-1 curve (less well resolved in MPE-2, but present in the Fe-based MPE curve from the CY site), followed by a transgressive period that is ongoing presently.

We then converted the two MPE curves for the F site to MHW rise curves (Appendix). The two MHW rise curves for location F are similar, with a consistently 'deeper MHW level' derived from the Fe data set (Fig. 6). The short period of regression around AD 1860 (25 cm) is not resolved in the Fe data set, and the shape and timing of the early regression transgression couplet (T1) is slightly different for the two curves. We obtained a long-term average of 1.1 mm $yr^{-1}$ in MHW rise over the last 1500 yr (confirming Bloom and Stuiver, 1963), and a mean rate of about 2.2 mm $yr^{-1}$ over the last 100 years, in agreement with tide gauge data for this region (Lyles *et al.*, 1988).

The curves show rapid changes in the rate of RSLR around AD 650 to 750 (drowning followed by shallowing), AD 1200 (increase in RSLR), AD 1450 (decrease in RSLR), AD 1700 (increase in RSLR) and AD 1880 (increase in RSLR). Very low rates prevailed between AD 750 and 1200, and between AD 1750 and 1880. Given the errors in sampling depth, in $^{14}C$ dating and in age calibration, these ages have an uncertainty of up to ±150 yr for some intervals. We have no age data between 105 cm and 135 cm (AD 700–1400), and changes in sedimentation rate within this interval are thus not resolved. The ages for the

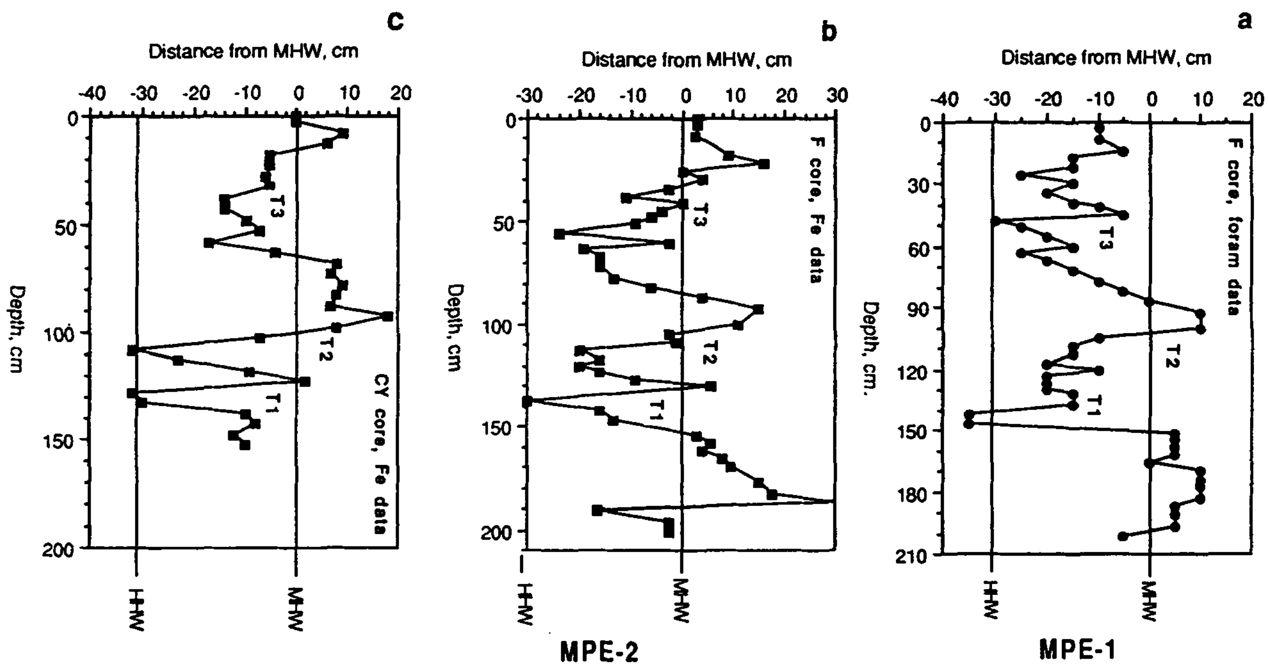



**Fig. 5.** *Marsh-Palaeo-Environmental graphs (MPE curves) based on fauna (MPE-1; (**a**) and based on Fe contents of sediments (MPE-2, (**b**) for the F core and an Fe-based MPE curve (MPE-2) for the CY site. (**c**) The three transgressive events are labelled T1, T2 and T3.*

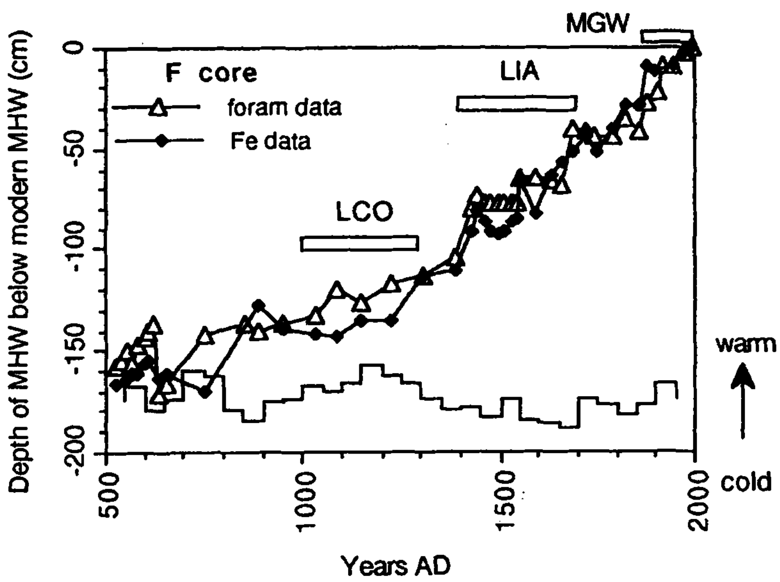



**Fig. 6.** *Graph of 'MHW below modern MHW' vs age derived from Fig. 5, using age data in Fig. 8, Table 1. Climate periods are indicated by rectangular boxes (LCO = Little Climate Optimum, LIA = Little Ice Age, MGW = Modern Global Warming) as well as a temperature record after Hammer* et al. *(1980). The age of the part of the curve between* AD *700 and 1450 is derived from samples of* S. alterniflora, *which on average grows below* S. patens. *(Van de Plassche, 1991). Ages for this part of the curve are thus minimum ages, and might be higher by up to 100–150 yr.*

samples between 90 and 105 cm (Table 1) are derived from *S. alterniflora* material. This plant grows on average lower than *S. patens*, and the age for the apparent increase in the rate of RSLR (AD 1450) is thus a minimum age.

The T3 transgression is relatively well constrained in age because it occurs at the onset of anthropogenic metal pollution. We consistently derived the same sequence of events over time with our two independent methods at several locations in the marsh and in another Connecticut marsh, Pataguanset (Varekamp and Thomas, unpubl. data), but the 'absolute' ages of the events are by necessity approximate. Our data thus agree with earlier interpretations that relative sea-level rise was not gradual, but punctuated (Fairbridge, 1961, 1981; Meyerson, 1972; Moerner, 1976, 1980, 1987; Van de Plassche, 1982).

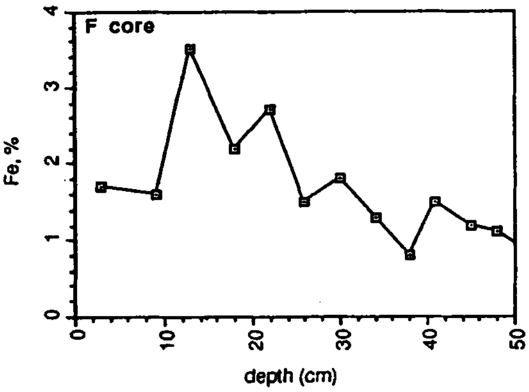
## DISCUSSION

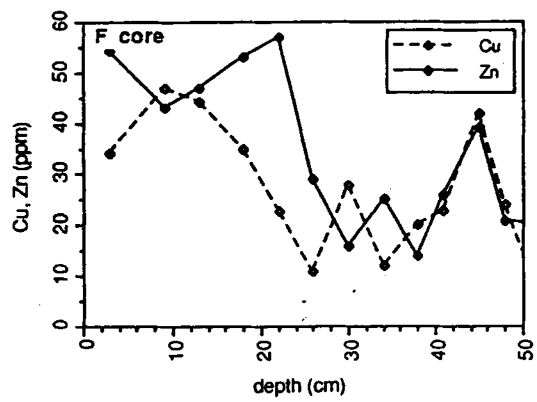
The changes in depositional environment over time may represent changes in RSL, which could relate to vertical changes in the position of land (seismic movements, rate changes in marsh accretion), changes in tidal range, or changes in the rate of true SLR. Changes in local geomorphology (e.g. changes in the course of major tidal channels, build-up of a coastal spit; Shennan, 1986; Pearson *et al.*, 1989) can dramatically change the marsh environment.

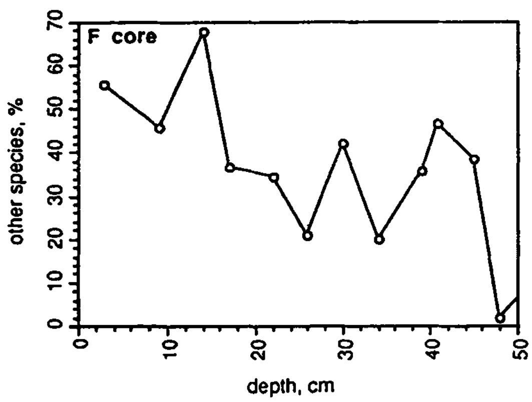
The time-MHW curves (Fig. 6) show three intervals of relatively sudden transgression, punctuated by smaller events. The T1 event (AD 600) is characterized by the presence of the sudden, preceding regression (AD 700). The rapid transgression brings sea level roughly back to where it was before the regression (MPE-1) or to a higher level with a following regression (MPE-2; Fig. 5). The marsh emerged between these events and there is no time control on the duration of the emergence. We used the marsh accretion rate derived from the deeper core section between the regression and transgression (Fig. 6). The suddenness of the regression and T1 transgression are probably related to local (but marsh-wide) factors. Growth of a spit may have led to complete closure of the marsh, preventing sea water from entering the marsh. The marsh would then either dry or turn into a fresh water marsh. We cannot distinguish between these two possibilities using foraminiferal evidence, because marsh foraminifera do not occur in fresh water marshes. Storms could have removed the barrier, leading to the rapid T1 transgression. We cannot exclude, however, that a true regression (decrease in RSL) occurred around AD 700. The shallowing–deepening sequence occurred at the end of a period of relatively rapid RSLR (up to about AD 650 in MPE-1), and at the beginning of a long episode of more sluggish RSLR (about AD 750–1200; Fig. 6), coinciding with a period of warming into the Little Climate Optimum.

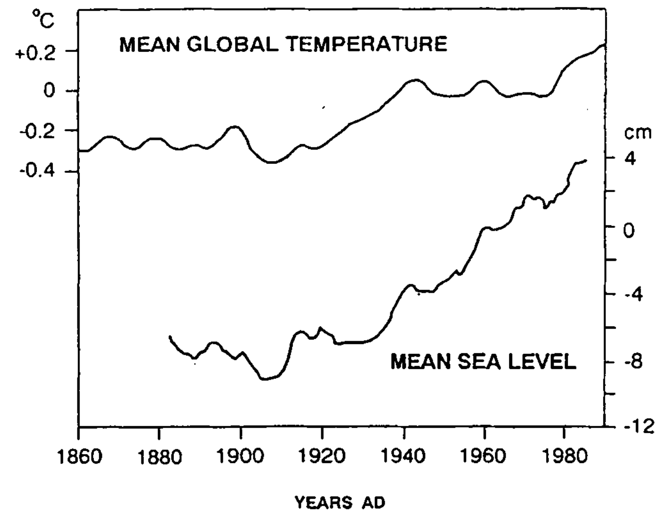
Transgressive periods T2 and T3 are fundamentally different from T1, and are interpreted as periods of accelerated RSLR. A markedly higher rate of RSL was reached at the end of each of these periods. Our age model puts T2 from about AD 1200–1450 (but the event might have occurred somewhat earlier; see Fig. 6, Appendix). The T3 episode started around AD 1700 in MPE-1 and slightly earlier for MPE-2. The period of slow rates of RSLR from AD 1800 to 1860 (MPE-1) is not resolved in the MPE-2 curve. The rate of RSLR increased again (MPE-1) around AD 1876–1900, and this transgression continues today.

Studies of other marshes in Long Island Sound show that the youngest depositional history (last 100–200 yr) was characterized by transgression of high marsh peat over highest high marsh peat (Rampino and Sanders, 1981). The main expansion started at 0.3–0.5 m below the marsh surface, suggesting that the modern transgressive phase occurred in a wide region. Further evidence on the occurrence of the T2 and T3 episodes in other areas is scarce, mainly because of a lack of high resolution data, although our preliminary chemical and faunal data from the Pattagansett marsh near New London (CT) also show emergence near 130 cm depth and transgressive regimes around 0.9 and 0.6 m depth (Varekamp and Thomas, unpubl. data).



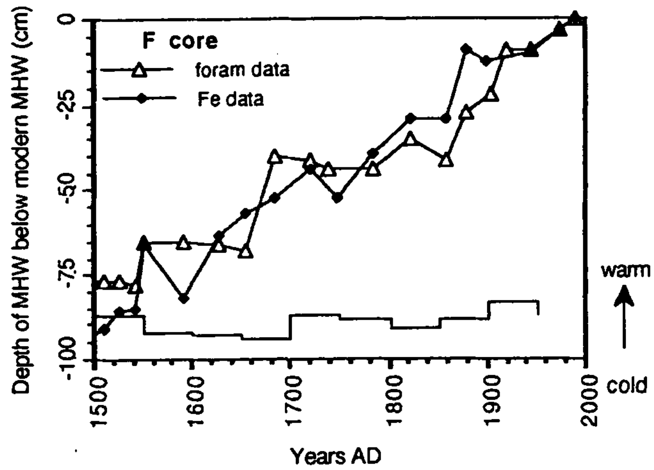




Fig. 7. (a) *Records for Fe, Cu and Zn and 'other species' as well as MHW below modern MHW versus depth for the last 150 years in the F and CY cores.* (b) *Record of a global tide gauge summary (Barnett, 1988), a global temperature summary (IPCC, 1990) and a detailed MHW rise curve derived from the foraminiferal data of core F.*

## CORRELATION OF RELATIVE SEA-LEVEL RISE WITH PALAEOCLIMATE RECORDS

The palaeoclimate record over the last 2000 yr is known from historic records as well as from climate proxies such as stable isotope abundances in ice cores, and pollen studies (Butler, 1959; Clark, 1986; Clark and Patterson, 1985; Clark *et al.*, 1978; Lamb, 1989). Most climate data are regional and obtained in Europe or Northeastern America (Wigley *et al.*, 1979; Gornitz *et al.*, 1982). A generalized summary of available data (IPCC, 1990) shows a warm period beginning near AD 0, a cold period in the last part of the first millenium, followed by the warm Little Climate Optimum (AD 1000–1300), and the Little Ice Age (AD 1400–1700). Modern global warming is thought to have started at the end of last century.

Ealier studies suggested the possibility of a correlation between the extent of glaciers and a RSLR curve from Clinton based on lithostratigraphic analyses of the marshes (Van de Plassche, 1991). Here we compare the more detailed RSLR curves derived by the faunal/geochemical techniques with a mean temperature curve for the northern hemisphere over the last 1500 years (Fig. 6). The time resolution over the interval before AD 100 is poor, and we do not want to speculate here on possible correlation between badly documented and dates climate change and unprecisely dated transgressional and regressional events. The T2 transgressive period (AD 1200–1450) did not coincide with the Little Climate Optimum (AD 1000–1300), but with the ending of this warm period and at the beginning of the Little Ice Age (Fig. 6). The ages for T2 might be slightly older than estimated here, but the beginning of the more rapid rate of RSLR postdates the period of increasing temperatures in the beginning of the Little Climate Optimum, even if we use the oldest acceptable estimate. The flat section in the curve after T2 coincides with the Little Ice Age, whereas the steepest inclination of the RSLR curve is found in the post-AD 1700 segment of the curve (Modern Global Warming). The initial part of the T3 transgression correlates well with the sudden warming in the early 1700s (Lamb, 1982, 1984), and the following period of slow RSLR coincides with the cold 'tail' of the Little Ice Age (Lamb, 1984). The last steep part of the T3 transgression coincides with modern global warming.

Within the limits of time resolution and the poor time constraints of palaeoclimate records, there is no clear correlation between intervals with high rates of RSLR and warm periods of longer than 100 years. RSL rose only slowly during the warm Little Climate Optimum, and there was just as slow a rate of RSLR during the cold Little Ice Age.

Absolute rates of RSLR can be approximately estimated, although marsh accretion rates probably varied somewhat with the pace of RSLR. During the T2 episode about 40 cm of RSLR occurred in about 250 yr (1.6 mm $yr^{-1}$). We found a rate of 2.2 mm $yr^{-1}$ RSLR over the last 100 yr, using averages of the onset of Cu pollution and increased rates of RSLR in cores F and CY, close to the rate of RSLR measured in New London (CT) over the last 50 yr (2.1 mm $yr^{-1}$, NOAA, 1988). Virtually all the RSLR over the last 1200 yr took place during the drowning episodes of AD 1200–1450 and during the period from AD 1700 to modern times.

We investigated in detail the record of the modern episode (since AD 1700). We plotted the Fe and Cu abundances, 'other species' parameter and the MHW record for the CY and F cores against depth (Fig. 7). We used the Cu abundances as an indicator for anthropogenic metal pollution (Varekamp, 1991), and the Fe abundances and 'other species' parameter as environmental indicators from which the MHW curve was derived. We also plotted details of the foraminifer-based MHW curve 'other species' parameter, Fe contents, an average tide gauge record (Barnett, 1988), and the global mean temperature against time (Fig. 7).

There is a positive correlation between tide gauge records, our data and global temperature data. The historical temperature and global tide gauge records show little rise before AD 1900, which coincides with the flat region in our MHW rise curve. From the late 1800s on, RSL started rising fast, as did the temperature, with a fall back after the 1940s. Even small variations in the temperature record, e.g. the warm period of the 1940s, may be present in our data set as peaks in Fe abundance and the 'other species' parameter, suggesting increased flooding frequencies over a period of a single decade. This increase may be related to increased frequency or vigour of storms, or to increased rates of RSLR. The onset of accelerated RSLR slightly lags behind the start of anthropogenic metal pollution in core CY and occurs at similar level in core F. The foram-based MHW-curve (Fig. 6) is not systematically influenced by heavy metal pollution and the observed broad correlation between the increased rate of RSLR and anthropogenic pollution is thus probably not inherent to our methods.

The correlation between climate and rates of RSLR and the correlation in Fig. 7 suggests that climate change and rates of SLR are in phase (within our time resolution) during the Modern Global Warming. For this period the global tide gauge and temperature records show a direct correlation, with little or no lag time between the onset of sea-level rise and global warming (Fig. 7). We can not assume without further data that a direct causal relationship exists between atmospheric warming and the inferred increase in the rate of RSLR, especially because such a correlation does not appear in our records for the Little Climate Optimum. Our records, however, clearly suggest the existence of rapid transitions from slow to fast rates of RSLR, with changes taking place over periods on the order of decades.

If climate change indeed drives the change in RSLR, then what could cause such a rapid response of RSLR? We argued elsewhere (Thomas and Varekamp, 1991; Van de Plassche, 1991) for a correlation between glacier extent and the Clinton RSLR record. We did not imply that all RSLR is caused by global sea-water volume increase as a result of glacier melting: both features are likely to be a result of climate change. It seems unlikely that continental glacier melting could cause a rise in global sea-level of 20 cm in 50 years (as observed in our data). Recent data and climate models suggest that rising temperatures at high latitudes result, in fact, in increased volume of the Antarctic ice sheet as a result of increased precipitation for at least decades to one of two centennia (e.g. Oerlemans, 1982, 1989; Drewry, 1991). Surging of the West Antarctic ice sheet has been suggested as a causal mechanism for rapid changes in RSL, but we do not have evidence that surging of that ice sheet actually occurred (Kuhn, 1990).

Thermal expansion of the oceanic mixed layer may give rise to global SLR on the order of 2–3 cm for a 1°C rise in temperature (Wigley and Raper, 1989), with a response time on the order of 1–2 decades. Thermal expansion may have contributed to the accelerated rates of RSLR during the T3 episode, but is unlikely to be the only cause.

The T2 episode occurred during a period of overall declining temperatures (at least in the northern hemisphere), and lagged the temperature increase at the beginning of the Little Climate Optimum by several hundreds of years. During the Little Climate Optimum temperatures were higher at intermediate and high latitudes, but only changes in climate characteristics, not necessarily higher temperatures, have been documented at lower latitudes (Lamb, 1977). The Little Climate Optimum might thus have been a period of more efficient distribution of heat to higher latitudes, possibly by ocean currents, instead of a period with a higher global heat budget. This heat transport could have led to increased precipitation at higher latitudes, thus increasing polar precipitation, ice volumes and decreased rates of RSLR (e.g. Oerlemans, 1982, 1989; Drewry, 1991). On the other hand, the Little Climate Optimum might have been too local a phenomenon to be perceptible as a rise in RSLR. Further studies on coastal marshes will enable us to judge if the drowning episode that occurred at the end of the Little Climate Optimum was a worldwide phenomenon, or occurred only or more intensely along the northeastern US seaboards. Further studies are also needed to assess why and in which aspects the period of Modern Global Warming (temperature rise and rapid rise in RSL) differs from the Little Climate Optimum (temperature rise, slow rise in RSL).

Part of the signal of accelerated RSLR in modern times in the Clinton marsh sediments could be related to an Atlantic ocean water surface effect (e.g. Kidson, 1982). Changes in ocean circulation patterns may lead to redistribution of the ocean water mass leading to a regional rise in sea-level. Mean sea-level at the northeastern US seaboards is slightly higher in the summer than in the winter (Gordon, 1980), as a result of slightly different ocean currents and wind fields during the summer. A change in position of the Gulf stream could 'push' the ocean waters higher up the northeastern US coast (Fairbridge, 1987). The modern rate of RSLR in the northeastern USA (about 2–3 mm $yr^{-1}$; Fairbridge, 1987) is higher than the estimated global average of about 1–1.5 mm $yr^{-1}$ (Pirazzoli, 1989; Lambeck, 1990), possibly indicative of such an ocean surface effect. Studies from the southern hemisphere provide more evidence for falling sea levels than SLR over the last few millenia (Isla, 1989), again suggesting the importance of watermass redistribution and dynamic, ocean water surface effects.

Pulses in RSLR could possibly be related to changes in ocean circulation, because ocean water is an important heat conveyor (Broecker, 1989; Broecker *et al.*, 1990). Thermal expansion of the ocean mixed layer and glacier melting are necessary elements of true, global sea-level rise, and Modern Global Warming might possibly involve global sea-level rise, in contrast with the Little Climate Optimum. The cool period of the Little Ice Age was presumably a global phenomenon (Grove, 1987), and may have led to changes in ocean circulation, which with the documented growth in glaciers (Roethlisberger, 1986) may have led to the decrease in the rate of RSLR over the period from AD 1200–1450 (or slightly earlier).

## CONCLUSIONS

Application of our new study methods provides detailed information on evolution and growth of coastal salt marsh sequences. The correlation between a local tide-gauge record and our marsh palaeoenvironment database for the last 100 years suggests that our technique may be used as an extension of tide gauge data over the last few thousand years. We documented at least three periods of sudden drowning in the Clinton marshes. Two periods of sudden marsh drowning are interpreted to result from increases in the rate of RSLR. One of these two episodes (AD 1200–1450) is broadly correlated with a period of cooling, the end of the Little Climate Optimum. Uncertainties in the dating include the possibility that this episode of increased rate of RSLR overlapped with the period of peak warmth (AD 1150–1200), but it postdated the period of increasing temperatures during the first half of the Little Climate Optimum. The second period of marsh drowning is correlated with a period of increasing temperatures, from AD 1700 through 1850, and then from AD 1860 to the present time (Modern Global Warming). There was little or no RSLR in the intervening cold period (Little Ice Age).

The signal in RSLR during the Modern Global Warming might have been caused partially by changes in ocean water surface circulation and thus morphology, in addition to thermal expansion and glacier melting. We stress that the RSLR response to climate change may be different in different regions as a result of such a redistribution of ocean water masses. The northeastern USA seaboard may be sensitive to this effect and future RSL changes in response to climate change may be more significant there than elsewhere. If effects of ocean surface topography are important in RSLR in some regions, we can expect different effects of RSLR from different types of climate change, e.g. greenhouse warming with an increase in total global heat budget versus climate change related to changes in heat distribution. During some periods of warming on a centennial time-scale increased humidity at high latitudes might possibly lead to increased precipitation at high latitudes, and thus increased ice volume and a lowering of sea-level (Little Climate Optimum). The modern global warming, however, appears to have co-occurred with rising sea-level over the last hundred years, although several authors suggest that the Antarctic ice sheet is expanding presently (e.g. Drewry, 1991).

We found a broad correlation between the onset of modern global warming, anthropogenic pollution and elevated rates of RSLR. The exact linkage between these observations needs further study, but our data suggest that if a causal linkage exists, the response times are extremely short, on the order of decades or less. Future anthropogenically induced global warming might manifest itself rapidly in raised sea-level, possibly with a different magnitude in different areas.

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## REFERENCES

Allen J.R.L. and Rae J.E. (1987) Late Flandrian shoreline oscillations in the Severn Estuary: a geomorphological and stratigraphical reconnaissance, *Phil. Trans. R. Soc. London*, **B315**, 185–230.

Allen J.R.L. and Rae J.E. (1988) Vertical salt marsh accretion since the Roman period in the Severn Estuary, Southwest Britain, *Mar. Geol.*, **85**, 225–235.

Barnett T.P. (1988) Global Sea level change. In: *NCPO Climate variations over the last century and the greenhouse effect.* Report from the first climate trends workshop, September 1988, Washington DC, USA, National Climate Program Office, NOAA, Rockville, Maryland.

Bloom A.L. and Stuiver M. (1963) Submergence of the Connecticut coast, *Sci.*, **139**, 332–334.

Bloom A.L. (1964) Peat accumulation and compaction in a Connecticut coastal marsh, *J. Sedim. Petrol.*, **34**, 599–603.

Broecker W.S. (1989) The salinity contrast between the Atlantic and Pacific oceans, *Paleoceanography*, **5**, 207–212.

Broecker W.S., Bond G. and Klas M. (1990) A salt oscillator in the glacial Atlantic? 1. The concept, *Paleoceanography*, **5**, 469–477.

Butler P. (1959) Palynological studies of the Barnstable Marsh, Cape Cod, Massachusetts, *Ecology*, **40**, 735–737.

Clark J.A., Farrell W.E. and Peltier W.R. (1978) Global changes in postglacial sea level: a numerical calculation, *Quat. Res.*, **9**, 265–287.

Clark J.S. (1986) Dynamism in the barrier-beach vegetation of Great South Beach, New York. *Ecol. Monogr.*, **56**, 97–126.

Clark J.S. and Patterson W.A. (1985) The development of a tidal marsh: upland and oceanic influences, *Ecol. Monogr.*, **55**, 189–217.

Drewry D.J. (1991) The response of the Antarctic Ice Sheet to climatic change. In: *Antarctica and Global Change* (ed. by C.M. Harris and B. Stonehouse), pp.90–106. Belhaven Press, London, UK.

Fairbridge R.W. (1961) Eustatic changes in sealevel. In: *Physics and Chemistry of the Earth* (ed. by L.H. Arhens, K. Press, K. Rankama and S.K. Runcorn), pp. 99–185.

Fairbridge R.W. (1981) Holocene wiggles, *Nature*, **292**, 670.

Fairbridge R.W. (1987) The spectra of sea level in a Holocene. In: *Climate: History, Periodicity and Predictability* (ed. by M.R. Rampino, J.E. Sanders, W.E. Newman and L.K. Koenigson), pp. 127–142.

Fairbridge R.W. and Hillaire-Marcel C. (1977) An 8,000 yr paleoclimate record of the Double Hale 45 yr solar cycle, *Nature*, **268**, 413–416.

Field M.E., Meisburger E.P., Stanley E.A. and Williams S.J. (1979) Upper Quaternary peat deposits on the Atlantic inner shelf of the United States, *Bull. geol. Soc. Am.*, **90**, 618–628.

Foerstner U. and Wittmann G.T.W. (1983) *Metal Pollution in the Aquatic Environment.* Springer-Verlag, New York, 485 pp.

Given P. (1975) Environmental organic chemistry of bogs, marshes and swamps, *Environ. Chem.*, **1**, 55–80.

Gorden R.B. (1980) The sedimentary system of Long Island Sound. In: *Estuarine Physics and Chemistry: Studies in Long Island Sound* (ed. by B. Saltzman). *Adv. Geophys.*, **22**, 1–39.

Gornitz V. (1990) Mean Sea level changes in the recent past. In: *Climate and Sea Level Change: Observations, Implications and Projections* (ed. by R.A. Warrick and T.M.L. Wigley). Cambridge University Press, Cambridge, UK, in press.

Gornitz V. (1991) Global coastal hazards from future sea level rise, *Palaeogeogr. Palaeoclim. Palaeoecol. (Global Planet. Change sect.)*, **89**, 379–398.

Gornitz V., Lebedeff S. and Hansen J. (1982) Global Sea Level Trend in the Past Century, *Science*, **215**, 1611–1614.

Grove J.M. (1988) *The Little Ice Age.* Methuen, London, 498 pp.

Hammer C.U., Clausen H.B. and Dansgaard W. (1980) Greenland ice sheet evidence of post-glacial volcanism and its climatic impact, *Nature*, **288**, 230–235.

Harrison E.Z. and Bloom A.L. (1977) Sedimentation rates on tidal salt marshes in Connecticut, *J. Sedim. Petrol.*, **47**, 1484–1490.

Hem J.D. (1975) Role of hydrous metal oxides in the transport of heavy metals in the environment (discussion of the paper of G.F. Lee). In: *Heavy Metals in the Aquatic Environment.* (ed. by P.A. Krenkel), pp. 149–153. Pergamon Press, Oxford.

Intergovernmental Panel on Climate Change (IPCC) (1990) *Climate Change.* Cambridge University Press, Cambridge, 364 pp.

Isla F.I. (1989) Holocene sea level fluctuation in the Southern Hemisphere, *Quat. Sci. Revs*, **8**, 359–368.

Kidson C. (1982) Sea level changes in the Holocene, *Quat. Sci. Revs*, **1**, 121–161.

Kraft J.C., Belknap D.F. and Demarest J.M. (1987) Prediction of Effects of Sea level change from paralic and inner shelf stratigraphic sequences. In: *Climate: History, Periodicity and Predictability* (ed. by M.R. Rampino, J.E. Sanders, W.E. Newman and L.K. Koenigson), pp. 166–192.

Kuhn M. (1990) Contributions of glaciers to sea level rise. In: *Climate and Sea Level Change: Observations, Implications, and Projections* (by R.A. Warrick and T.M.L. Wigley). Cambridge University Press, Cambridge, UK, in press.

Lamb H.H. (1977) *Climate: present, past and future, Vol 2: Climatic history and the future.* Methuen, London, 835 pp.

Lamb H.H. (1982) *Climate, History and the Modern World.* Methuen, London, 387 pp.

Lamb H.H. (1984) Climate in the last 1000 years: natural climatic fluctuations and change. In: *The Climate of Europe: Past, present and future* (ed. by H. Flohn and R. Fantechi), pp. 25–64. Riedel, Dordrecht.

Lambeck K. (1990) Late Pleistocene, Holocene and present sea-levels: constraints on future change, *Palaeogeogr. Palaeoclim. Palaeoecol. (Global Planet. Change sect.)*, **89**, 205–217.

Lee G.F. (1975) Role of hydrous metal oxides in the transport of heavy metals in the environment. In: *Heavy Metals in the Aquatic Environment* (ed. by P.A. Krenkel), pp. 137–147. Pergamon Press, Oxford.

Long S.P. and Mason C.F. (1983) *Saltmarsh Ecology.* Chapman and Hall, New York, 160 pp.

Lyles L.D., Hickman L.E. Jr. and Debaugh H.A. Jr. (1988) *Sea Level Variations for the United States 1855–1986.* NOAA, U.S. Department of Commerce, Rockville, Md.

McCaffrey R.J. and Thomson J. (1980) A record of the accumulation of sediment and trace metals in a Connecticut salt marsh. In: *Estuarine Physics and Chemistry: Studies in Long Island Scound* (ed. by B. Saltzman). *Adv. Geophys.*, **22**, 165–236.

Meier, M.F. (1984) Contribution of small glaciers to global sea level, *Sci.*, **226**, 1418–1421.

Meyerson L.A. (1972) Pollen and paleosalinity analyses from a Holocene tidal marsh sequence, Cape May County, New Jersey, *Mar. Geol.*, **12**, 335–357.

Moerner N.-A. (1976) Eustatic changes during the last 8,000 years in view of radiocarbon calibration and new information from the Kattegatt region and other northwestern European coastal

areas, *Palaeogeogr. Palaeoclim. Palaeoecol.*, **19**, 63–85.
Moerner N.-A. (1980) The northwest European "sea level laboratory" and regional Holocene eustasy, *Palaeogeogr. Palaeoclim. Palaeoecol.*, **29**, 281–300.
Moerner N.-A. (1987) Models of global sea level change. In: *Sea Level Changes* (ed. by M.J. Tooley and I. Shennan), pp. 332–355.
Nelson A.R. and Jennings A.E. (1988) Use of salt marsh foraminifera to identify coseismic changes in sea level in Oregon, *Am. Quat. Ass., Progr. and Abstr., 10th Bienn. Meeting, Amherst, Ma*, 142 pp.
Nixon S.W. (1982) *The Ecology of New England High Salt Marshes: a community profile*. U.S. Fish and Wildlife Service, Office of Biological Services, Washington, D.C., FWS/OBS-81-55, 70 pp.
Oerlemans J. (1982) Response of the Antarctic ice sheet to a climatic warming: a model study, *J. Climatol.*, **2**, 1–11.
Oerlemans J. (1989) A projection of future sea-level, *Climatic Change*, **15**, 151–174.
Pearson I., Funnell B.M. and McCave I.N. (1990) Sedimentary environments of the sandy barrier/tidal marsh coastline of North Norfolk, *Bull. geol. Soc. Norfolk*, **39**, 3–44.
Pinter N. and Gardner T.W. (1989) Construction of a polynomial model of glacio-eustatic fluctuation: estimating paleosea levels continuously through time, *Geol.*, **17**, 295–298.
Pirazzoli P.A. (1989) Present and near future global sea level changes, *Palaeogeogr. Palaeoclim. Palaeoecol. (Global Planet. Change sect.)*, **75**, 241–258.
Rampino M.R. and Sanders J.E. (1981) Episodic growth of Holocene tidal marshes in the northeastern United States: a possible indicator of eustatic sealevel fluctuations, *Geology*, **9**, 63–67.
Roethlisberger E. (1986) *10.000 Jahre Gletschergeschichte der Erde*. Verlag Sauerlaender, Aarau.
Scott D.B. and Medioli F.S. (1978) Vertical zonations of marsh foraminifera as accurate indicators of former sea-levels, *Nature*, **272**, 528–531.
Scott D.B. and Medioli F.S. (1980) Quantitative studies of marsh foraminiferal distributions in Nova Scotia: implications for sea level studies, *Spec. Publ. Cushman Found. Foram. Res.*, **17**, 58 pp.
Scott D.B. and Medioli F.S. (1986) Foraminifera as sea level indicators. In: *Sea level Research: a manual for the collection and evaluation of data* (ed. by O. Van de Plassche), pp. 435–455. Geo Books, Norwich.
Scott D.B., Medioli F.S. and Miller A.A.L. (1987) Holocene sea levels, paleoceanography, and the *late glacial ice configuration near the Northumberland Strait, Maritime Provinces, Can. J. Earth Sci.*, **24**, 668–675.
Scott D.B., Medioli F.S. and Schafer C.T. (1977) Temporal changes in foraminiferal distributions in Miramichi River estuary, New Brunswick, *Can. J. Earth Sci.*, **14**, 1566–1587.
Scott D.B., Schafer C.T. and Medioli F.S. (1980) Eastern Canadian estuarine foraminifera: a framework for comparison, *J. Foram. Res.*, **10**, 205–234.
Scott D.K. and Leckie R.M. (1990) Foraminiferal zonation of Great Sippewissett Salt Marsh (Falmouth, Massachusetts), *J. Foram. Res.*, **20**, 248–266.
Shennan I. (1986) Flandrian sea level changes in the Fenland II: tendencies of sea level movement, altitudinal changes and local and regional factors, *J. Quat. Sci.*, **1**, 155–179.
Stevenson J.C., Ward L.G. and Kearney M.S. (1988) Sediment transport and trapping in marsh systems: implications of tidal flux studies, *Mar. Geol.*, **80**, 37–59.
Stuiver M. and Pearson G.W. (1986) High precision calibration of the radiocarbon time scale, AD 1950–500 BC, *Radiocarbon*, **28**, 805–838.
Thomas E. and Varekamp J.C. (1991) Paleo-environmental analyses of marsh sequences (Clinton, CT): evidence for punctuated rise in relative sea level during the latest Holocene, *J. Coastal Res., Spec. Iss.*, **11**, 125–158.
Turekian K.K. (1977) The fate of metals in the oceans, *Geochim. Gosmochim. Acta*, **41**, 1139–1144.
Van de Plassche O. (1982) Sea level change and water-level movements in the Netherlands during the Holocene, *Med. Rijks Geol. Dienst*, **36–1**, 1–93.
Van de Plassche O. (1991) Late Holocene sea-level fluctuations on the shore of Connecticut inferred from transgressive and regressive overlap boundaries in salt marsh deposits, *J. Coastal Res., Spec. Iss.*, **11**, 159–180.
Van de Plassche O., Mook W.G. and Bloom A.L. (1989) Submergence of coastal Connecticut 6000–3000 $^{14}$C-years BP, *Mar. Geol.*, **86**, 349–354.
Van de Plassche O. and Roep Th. B. (1989) Sea level changes in The Netherlands since 6500 yr B.P.: basal peat vs coastal barrier data. In: *Late Quaternary Sea Level Correlation and Applications* (ed. by D.B. Scott, P.A. Pirazzoli and D.H. Honig), NATO ASI series C, pp. 41–55. Kluwer, Dordrecht.
Varekamp J.C. (1991) Trace element geochemistry and pollution history of mudflat and marsh sediments from the Connecticut River estury, *J. Coastal Res., Spec. Iss.*, **11**, 105–124.
Walcott R.I. (1972) Late Quaternary vertical movements in eastern North America: quantitative evidence of glacio-eustatic rebound, *Rev. Geophys. Space Phys.*, **10**, 849–884.
Wigley T.M.L. and Raper S.C.B. (1987) Thermal expansion of sea water associated with global warming, *Nature*, **330**, 127–132.

## APPENDIX 1

### Derivation of MHW curves

We converted the MPE curves to MHW rise curves in sereval steps, starting with plotting depth-in-core versus 'depth of MHW below datum'. The palaeo-depth of MHW at a given core level was obtained by adding or subtracting the vertical distance from MHW for that level (value derived from the MPE curves) to its depth-in-core. This results in a MHW rise curve along an undefined time-coordinate axis (the depth-in-core axis). Seven peat samples from core F were dated with $^{14}$C (Table 1), and calibrated according to Stuiver and Pearson (1986). We used the onset of anthropogenic pollution (as heralded by an increased Cu abundance in the records) as the AD 1860 level (Varekamp, 1991). We did not use the ages obtained for the samples at 75–78 cm and 135 cm in the curve in Fig. 8, because these samples show age inversion, combined with unacceptable values for $\delta^{13}$C. We derived an age–depth curve from the remaining data points by using the age markers and their uncertainties in depth, and error boxes for age calibration based on 1$\sigma$ error in the $^{14}$C measurement (Fig. 8). To interpolate between age markers, we used a constant accretion rate for segments of the MPE curve with the same microfaunal trend, e.g. the long, gradual shoaling trend from 100 to 50 cm depth (Fig. 8). We subsequently 'stretched' the MHW rise curve versus depth-in-core into time by plotting depth of MHW below datum versus age using the age–depth information. We then converted the 'MHW below datum' axis to 'depth of MHW below modern MHW', to obtain an age–MHW rise curve, using a modern depositional environment at the F coring site of about 13 cm above MHW for MPE-1 and 6 cm below MHW for MPE-2. The estimated (true) modern position of coring site F is between 5 and 10 cm above modern

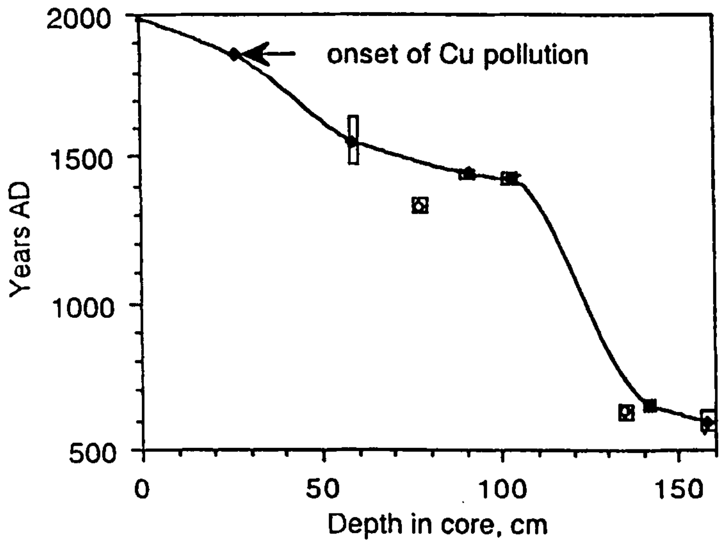


**Fig. 8.** *Depth–age graph with calibrated* $^{14}C$ *age index points; the* AD *1860 date level represents the onset of anthropogenic Cu pollution (Varekamp, 1991). The samples with ages between* AD *1300 and 1400 consisted of material from* S. alterniflora *(Table 1) which grows at lower average levels (5–35 cm) than* S. patens. *Therefore, the slope of the curve in its steepest part (*AD *700–1450) is a maximum slope, and ages for this part of the core are minimum ages.*

MHW (Van de Plassche, 1991). The differences in position of modern depositional environments with respect to MHW reflect the relatively poor agreement between the two MPE methods in the upper 40 cm of core F, but internally they are consistent and correlate well. We did not take the compaction of the lower peat beds into account in this derivation. Some compaction may have taken place in the lower core sections (Bloom, 1964), although the studied salt marsh deposits are rich in terrigenous material, and are thus unlikely to have been compacted as much as the sedge peat beds studied by Bloom. McCaffrey and Thomson (1980) argued that compaction of salt marsh sequences is minimal in the upper few, water-saturated metres because of the neutral or slightly positive buoyancy of the marsh sediments. If compaction was more than just a few cm, our rate estimates for RSLR in the lower core section are minimum values. The RSLR rate changes (drowning and shoaling episodes) are based on changes in depositional environment and are not influenced by compaction.

**Table 1.** Age Data for the F Core, calibrated according to Stuvier and Pearson, 1986.

| Lab. nr. | Plant material | Depth | Age $^{14}C$ BP | Years AD | $\delta^{13}C$ |
|---|---|---|---|---|---|
| UtC-2005 | *Sp. patens* | 58–60 | 340±40 | 1474–1636 | −13.8 |
| UtC-2004 | *Sp. alterniflora* | 75–78 | 540±70 | 1311–1353 | −4.0 |
| UtC-2003 | *Sp. alterniflora* | 90–93 | 440±30 | 1433–1454 | −13.6 |
| UtC-2002 | *Sp. alterniflora* | 101–105 | 500±30 | 1409–1434 | −13.4 |
| UtC-2001 | *D. spicata* | 133–135 | 1410±40 | 608–658 | −9.3 |
| UtC-2000 | *Sc. robustus* | 140–143 | 1370±50 | 636–673 | −22.5 |
| UtC-1999 | *Sp. patens* | 157–159 | 1460±40 | 559–638 | −12.7 |