Coastal peat-beds and peatlands of the southern North Sea: their past, present and future

Martyn Waller[1,*] and Jason Kirby[2]

[1]*Department of Geography and Geology, Kingston University London, Penrhyn Road, Kingston upon Thames, Surrey, KT1 2EE, UK*

[2]*School of Biological and Environmental Sciences, Liverpool John Moores University, Byrom Street, Liverpool, L3 3AF, UK*

*Author for correspondence (E-mail: martynwaller58@gmail.com; Tel.: 01444 819083).

ABSTRACT

Peat layers are well represented in the Holocene coastal deposits of the southern North Sea and provide evidence as to the extent and nature of the fens and bogs that occupied the region in the mid and late Holocene. While natural processes contributed to their demise, without human interference extensive areas of peatland would remain. We review the characteristics of the vegetation of these peatlands along with the processes that influenced their development. Spatial and temporal trends are explored through the use of palaeogeographic maps from three areas: the East Anglian Fenland, the Romney Marsh area and the Netherlands. The palaeoecological evidence indicates that eutrophic vegetation promoted by rising relative sea level (RSL) dominated in the mid Holocene, with a trend towards the development of oligotrophic and ombrotrophic vegetation in the late Holocene as the rate of RSL rise declined. Nevertheless, areas of eutrophic vegetation appear capable of long-term stability with areas of fen woodland and herbaceous fen persisting at some locations for several thousand years in the mid and late Holocene. Areas of active peat growth in the region are now largely confined to small remnants within agricultural settings. To retain their characteristic biodiversity these remnants have been managed using traditional practices, although their small size and fragmented distribution limits their biodiversity value. Biodiversity concerns and the ecosystem services peatlands provide, notably carbon sequestration and flood attenuation, underlie recent restoration projects. These efforts are likely to receive additional impetus as a consequence of rising water levels, given projected rates of RSL rise. Future large-scale restoration can be informed by a greater understanding of the processes that formed and sustained coastal peatlands in the past. We identify advances in palaeoenvironmental research that could enhance restoration efforts and help maximise the ecosystem services delivered through such projects.

Key words: bogs, peatlands, coastal lowlands, ecosystem services, fens, Holocene, palaeoecology, palaeogeography, relative sea-level change, restoration.

CONTENTS

I. INTRODUCTION

(1) General introduction

Peat-beds, buried within Holocene sedimentary fills, submerged on the sea floor and outcropping on the modern foreshore (as 'submerged forests'), are common features of the estuaries and former wetland areas of the southern North Sea and adjacent regions. Such layers have long been used as evidence for relative sea-level (RSL) change (Reid, 1913; Swinnerton, 1931; Godwin, 1940) and since the development of pollen analysis in the early 20th century AD, have been used for palaeoenvironmental reconstruction and the correlation and dating of coastal deposits (Godwin, 1943, 1945, 1978; Godwin, Suggate & Willis, 1958). Building on such foundations, research during the last 40 years has considerably enhanced our knowledge of the spatial and temporal distribution of these deposits, facilitating a greater understanding of the interacting processes that produced them.

In contrast to their representation in the Holocene, the region contains few extant peatlands. Most that survive have been heavily modified by human usage, although they have a high conservation status where they retain elements of their characteristic biodiversity. With additional impetus provided by the potential of coastal wetlands to mitigate against projected sea-level rise (Schuerch *et al*., 2018) and the ability of active peatlands to act as carbon sinks and provide other ecosystem services (Charman, 2002; Rydin & Jeglum, 2006; Bonn *et al*., 2016), there is increasing interest in their restoration and recreation, undertakings which, given the paucity of modern examples, may be informed by a better understanding of their Holocene predecessors.

This review examines the nature of the peat-forming vegetation, then reviews current knowledge of the processes involved in peat formation and changes in peatland extent. Recent conservation and restoration schemes are outlined and we discuss how palaeoenvironmental research can be used to inform large-scale restoration and enhance the ecosystem services these peatlands deliver. The southern North Sea and adjacent areas (Fig. 1) have been selected as our focus as the coastal deposits here are relatively well documented, with peat layers common. Also, the region experienced a consistent general trend of (albeit differential) land subsidence and rising RSL (Vink *et al*., 2007; Shennan, Milne & Bradley, 2012; Shennan, Bradley & Edwards, 2018) during the Holocene, which resulted in the accumulation of thick sequences of wetland deposits preserving a rich archive of palaeoenvironmental information.

(2) Definitions for Holocene coastal peats

Peat, comprising partially decomposed plant matter, is a variable substance and as Godwin (1943, p. 203) recognised, the term has been rather loosely applied in coastal settings. Definition can be based on the percentage organic content (dry mass). Allen (1990), for example, used >75% to define peat, although a figure of >65% is more typically applied (e.g. Charman, 2002). Problems arise when describing the often extensive bodies of sediment in coastal settings as peat, as the proportion of organic material in such layers varies considerably both vertically and horizontally, even when minor minerogenic layers and transitional sediments are discounted. The precise organic content has generally not been considered when describing units as ‘peats’ (semi-quantitative estimates are common in field descriptions, e.g. Tröels-Smith, 1955) although organic sediments with a high inorganic content are often referred to using terms such as organic clays or peaty clays.

Along the east coast of North America predominantly organogenic sediments described as peat derive from salt-marsh environments (Johnson, 1913; Allen, 2000). However, this is not the case in the southern North Sea region where equivalent coastal and estuarine marshes give rise to sediments that are predominantly minerogenic (Pratolongo *et al*., 2018). In this region, where deposits have high organic content, the *in situ* organic matter derives from plants that grew in transitional (e.g. reedswamp) and terrestrial freshwater environments. Coastal peat can therefore be regarded as sediment from which the organic content derives predominantly from freshwater plants growing *in situ* in telmatic (between high and low water levels) or terrestrial environments (collectively here termed peatlands), rather than from halophytes in salt-marshes. A freshwater origin raises the issue as to in what sense peat/peatlands can be considered 'coastal'. Peat formation is associated with areas where the water level is close to the sediment surface and decay is inhibited by anaerobic conditions induced by waterlogging. In estuaries and coastal wetlands there is an expectation that the water table will be influenced by sea level and this could be considered a useful distinguishing feature (Tooley, 1986). For example, Hageman (1969, p. 377) defined a 'perimarine' area 'where sedimentation or seditation took place under the direct influence of the relative sea-level movements but where marine or brackish sediments themselves are absent'. However, the utility of such terms is open to question as it can be very difficult to determine whether deposition was under the influence of sea level or not; in a number of circumstances (see Section III.2) water levels in coastal wetlands can become decoupled from sea level. While many of the peat layers of the southern North Sea region are likely to have formed in what can be termed 'tidal freshwater wetlands' (between the upstream limits of the tide and downstream saline estuarine environments; Barendregt & Swarth, 2013), this is not universally the case. A 'coastal' peat is used here to refer to a deposit that formed at the interface between the marine and fluvial domains, irrespective of whether this occurred under the direct influence of sea level or tides.

Gyttja – organic sediment formed below water level (in a limnic environment) – also occurs within the coastal deposits of the region and can form extensive layers (e.g. the calcareous gyttja/shell marl of the Fenland basin; Waller, 1994*b*). Gyttja rich in terrestrial plant matter can also accumulate within certain terrestrial environments (Bos, Busschers & Hoek, 2012; and Section II.2) and, resembling humified peat, can be difficult to distinguish without laboratory-based examination. As a consequence, organic material deposited below water level is likely to have been described as peat in many studies of Holocene coastal sediments. Recognising this, the term peat is used here *sensu lato*, following published descriptions.

(3) Stratigraphic architecture

The contribution of peat to the Holocene sediments of the region varies considerably through space and time, with the complex three-dimensional geometry of the coastal deposits reflecting the variety and size of coastal settings (open coasts, back-barrier locations, estuaries, and tidal basins notably including the Rhine–Meuse delta) and the number of processes that influence both sediment type and sediment preservation. Variation in peat presence and thickness within the depositional complexes of the region also reflects the fundamental sedimentary architecture of back-barrier and estuarine environments (Fig. 2). Where fluvial systems enter the coastal zone, three sedimentary complexes can be distinguished. Firstly, a landward sediment body comprising predominantly minerogenic sediment, although isolated peat layers may be encountered, deposited within fluvial systems (e.g. the Echteld Formation in the Netherlands; De Mulder *et al*., 2003). Secondly, a peat complex (the Nieuwkoop Formation of De Mulder *et al*., 2003) that landward can either interdigitate with, or grade into, the fluvial complex. Peat can fill the vertical sedimentary sequence, the stratigraphic column, particularly adjacent to upland areas where major fluvial systems are absent and basal layers often extend both landward and seaward of the main bodies of peat. Seaward, peat layers interdigitate with sediment deposited under marine/brackish conditions. Such layers are referred to here as ‘terrestrialization’ peats. Minerogenic sediments deposited under marine/brackish conditions comprise the third sedimentary complex (the Naaldwijk Formation of De Mulder *et al*., 2003). The growth of peat over pre-Holocene surfaces (basal peat) is influenced by groundwater level rise and topography. Such peat often occurs extensively in valley bottoms, on low-gradient slopes and where tributaries join (Vis *et al*., 2015) and, by contrast, is frequently absent from steeply sloping surfaces.

Not all organic deposits situated in what might be considered as the coastal zone owe their origin to high regional groundwater levels. Peats, the formation of which commenced in the Late Glacial or early Holocene at altitudes well above contemporaneous sea level, occur in valley bottoms or small closed depressions and relate to the drainage topography. Such deposits feature prominently in the research of earlier workers, particularly Godwin in the East Anglian Fenland, as they enabled long pollen sequences to be constructed (providing a chronostratigraphic framework), and some were of notable archaeological interest (Godwin, 1940; Clark & Godwin, 1962). Additionally, in the late Holocene areas of coastal peat merged laterally with ombrotrophic (raised) bogs, which originated in depressions in the uplands of the northern Netherlands (Vos, 2015). Such stratigraphic continuity between deposits of differing genesis presents difficulties for the understanding Holocene coastal evolution, with information on the age, altitude and relationship to contemporaneous sea level required to determine whether a particular deposit can be considered 'coastal'.

II. THE VEGETATION OF COASTAL PEATLANDS

(1) Peatland terminology

Peatlands have been classified in a variety of ways (Wheeler & Proctor, 2000; Charman, 2002; Rydin & Jeglum, 2006). The vegetation is influenced by water chemistry and nutrient status (Fig. 3), which in turn is dependent upon hydro-topographic relationships. Peatlands that are rain-fed are termed ombrotrophic while those that receive groundwater are termed minerotrophic. The former are inevitably nutrient poor (oligotrophic) and acidic and (following Wheeler & Proctor, 2000) termed bogs, while the latter tend to be more nutrient rich (meso- or eutrophic). Those with a high pH are termed fens. However, minerotrophic peatlands are by no means always base-rich and some give rise to vegetation communities similar to those associated with ombrotrophic bog. Such communities can be termed poor fen or, as preferred here, mesotrophic bog. Vegetation communities where the water level is above the ground surface for most of the year are referred to as swamps (Wheeler & Proctor, 2000).

(2) Botanical and physical characteristics of coastal peatland habitats

The modern vegetation types likely to have contributed to the formation of Holocene peat deposits of the region are described in a number of sources (notably Zonneveld, 1960; Wheeler, 1980*a*,*b*; Rodwell, 1991*a*,*b*; 1995; Den Held, Schmitz & Van Wirdum, 1992; Wiegers, 1992). The variable nature of the depositional environments present in coastal wetlands is reflected in the complexity of the vegetation. Den Held *et al*. (1992), for example, defined 35 types of open terrestrialising vegetation from the Netherlands alone. Palaeoecological studies often suggest that Holocene peat deposits contain communities that are closely analogous to extant types, although non-analogue communities, such as the occurrence of *Taxus* in fen woodland (Godwin, 1968; Deforce & Bastiaens, 2007; Branch *et al*., 2012; Waller & Early, 2015), have been reported and additional communities (e.g. pools created by beaver dams) are likely to have occurred. The communities represented in coastal peat deposits of the southern North Sea region can be divided into five major types. First, swamps are species-poor communities dominated by tall monocotyledons, which occur at transitions to open water, where the sediment surface is seasonally or permanently submerged. The species commonly associated with such areas is the reed *Phragmites australis*, although other grasses with similar growth forms can also dominate (e.g. *Phalaris arundinacea*, *Glyceria maxima*). *Scirpus* spp. typically occur in brackish swamps (Den Held *et al*., 1992; Rodwell, 1995). *Phragmites*-dominated swamps (reedswamp) generally give rise to sediment in which fragments (up to 2 cm) of the leaves, stems and rhizomes are embedded within a fine-textured clay-rich matrix (Bos *et al*., 2012).

Second, herbaceous fens occur in seasonally or periodically flooded areas. These communities are both floristically and ecologically varied. Sedges are often, although not always dominant, with other monocotyledons including rushes (e.g. *Juncus subnodulosus*) and grasses (e.g. *Calamagrostis canescens*) frequently abundant. The sedges present include both tussock-forming species, individuals of which in the case of *Carex paniculata* can attain heights of *c*. 0.6 m and be separated by water or bare peat, and species with stout shoots that form dense clumps (e.g. *Cladium mariscus*). Sedge-dominated communities occupy a continuum between eutrophic and oligotrophic conditions (Fig. 3C) with species such as *Carex rostrata* characteristic in nutrient-poor environments (mesotrophic bog/oligotrophic fen). The sediment commonly produced comprises flat dark-surfaced rhizomes <0.5 cm wide, some of which are distinct (e.g. the salmon-pink axes of *Cladium* have long been distinguished in coastal peats; Godwin, 1975).

Third, fen carr is a term used to describe a number of woodland and shrub communities associated with freshwater wetlands, which are generally regarded as representing a comparatively dry environment where the sediment surface is close to the average water level (Wiegers, 1992). On nutrient-rich substrata, *Alnus glutinosa* frequently forms the canopy layer in what is often a species-diverse community. Typically, the vegetation additionally comprises an under-storey shrub layer, often with *Salix cinerea*, but many other species are possible. At ground level, conditions vary from the presence of tall herbs (those associated with herbaceous fens) and climbing/trailing plants, to a carpet of small herbs (e.g. *Chrysosplenium oppositifolium*) and grasses, to ‘hummock’ areas (moss-covered tree bases and sedge tussocks) interspersed between pools of water. Carr dominated by *Salix* can form communities or transitional zones between herbaceous fen and *Alnus* carr, while *Betula*-dominated communities (often again with *Salix*) occur in meso- and oligotrophic situations (Fig. 3B). The sediment produced comprises a mixture of wood (from fallen trunks to the remains of twigs), coarse organic detritus (e.g. fruits and cones of *Alnus*), and fine gyttja deposited in the pools. In riparian areas, organic content can be low (<50% dry mass) due to the inwash of alluvial clay sediment.

Fourth, mesotrophic–oligotrophic bog (Fig. 3A) refers here to a range of vegetation types that occur in environments with a low pH, but remain influenced by groundwater (which occurs close to the surface). The species present include those found at the mesotrophic end of the herbaceous fen continuum to those present in ombrotrophic bog (members of the Ericaceae and *Sphagna*). The sediments produced vary accordingly, potentially containing fibrous sedge remains, twigs and roots of the Ericaceae and the spongy remains of *Sphagnum* spp.

Fifth, ombrotrophic bog is distinguished from the previous community by the vegetation not being influenced by the groundwater, with rainfall being sufficient to maintain surface saturation. The Ericaceae and *Sphagna* dominate along with sedges, notably *Eriophorum*. Such an environment does not produce vegetation, or consequently sediment, that is fundamentally different from other bog types. Some species (e.g. *Sphagnum austinii*, *Empetrum nigrum*, *Andromeda polifolia*) can be considered indicative of ombrotrophy, however, recognition from peat deposits may depend more on the balance between species that thrive in ombrotrophic conditions and those that, as a result of competition, rarely dominate in more nutrient-poor environments (Tuittila *et al*., 2013).

Spatial (zonal) and temporal (successional) relationships between these communities are often seen as expressions of their tolerance of a different range of water depths (between the sediment surface elevation and the water level). In practice, the range of water levels that these communities can tolerate and their relationship (if any) to tidal constants is poorly understood (see Table 1 for various estimates) due to the fragmented and much altered state of the extant coastal wetlands of NW Europe. With tussock-forming plants and deep pools common features in a number of these communities, a sediment surface level may be impossible to define and measurement is therefore limited to the range of fluctuations (Wiegers, 1992). Wheeler & Proctor (2000) caution that in addition to average water depth and the frequency and duration of fluctuations in water surface level, the distribution of species and communities can be influenced by factors such as extreme minima and maxima and the timing of these events. In addition to such hydrological stresses, vegetation patterns in tidal freshwater wetlands are strongly influenced by salinity and tolerance to burial by sedimentation (Leck *et al*., 2009).

III. PROCESSES INFLUENCING THE FORMATION OF COASTAL PEATS

(1) Introduction

Peat formation requires the elevation of the sediment surface in relation to the water table to be suitable for the growth of wetland plants and the preservation of organic material. Factors influencing accumulation include processes that contribute sediment, either autochthonously (deposited *in situ* by the peat-forming vegetation) or allochthonously (transported to the site of deposition), and processes that result in a reduction in the elevation of the surface: decomposition and compaction. Other processes, notably RSL movements in coastal situations, can change the absolute height of the water table. Allen (1990) introduced a model of peat growth in coastal marshes under rising RSL that incorporates these processes (Equation 1)

$$\frac{dE}{dt} = \frac{dSorg}{dt} + \frac{dSmin}{dt} - \frac{dM}{dt} - \frac{dP}{dt} \qquad (1)$$

The elevation (E) of sediment surface over time ($\frac{dE}{dt}$) is determined by the supply of sediment, the net accumulation of organic matter $\frac{dSorg}{dt}$ and any minerogenic sediment ($\frac{dSmin}{dt}$), RSL movement ($\frac{dM}{dt}$), a rise in which was treated as positive, and the shortening of the sediment column due to compaction through loading ($\frac{dP}{dt}$).

The influences of these factors on peat formation are discussed below, although many are interrelated, with interactions among processes underlying both peatland stability and change.

(2) Water-table relationships and RSL

The Holocene RSL history for the region is summarised in Fig. 4. Spatial variability in several Holocene geological processes, mainly glacio and hydro-isostatic movements, resulted in differential subsidence across the southern North Sea area, estimated at between 0.1 and 1 mm yr^{-1}, which has operated non-linearly over time (Kiden, Denys & Johnston, 2002; Vink *et al*., 2007). The northern areas experienced the highest rates of RSL rise in the early Holocene (with the greatest subsidence), but the RSL curves presented in Fig. 4 converge so that rates of RSL rise in the late Holocene are broadly similar across the region. During the early Holocene, defined as *c*. 11,700–8200 cal. yrs BP [following Walker *et al*. (2012), with all radiocarbon dates expressed here in calibrated years before present, i.e. 1950 AD], the southern North Sea region experienced average rates of RSL rise >7 mm yr^{-1} with short-lived jumps likely relating to sudden meltwater release events associated with the disintegration of major ice sheets (Hijma & Cohen, 2019). The latter have been implicated in the final inundation of Doggerland and other now-submerged early Holocene landscapes in the southern North Sea (e.g. Brown *et al*., 2018). The rate of RSL rise gradually decreased during the mid-Holocene (*c*. 8200–4200 cal. yrs BP) and was further reduced in the late Holocene (*c*. 4200 cal. yrs BP to present) to <1 mm yr^{-1} after 3000 cal. yrs BP (van de Plassche, 1982; Denys & Baeteman, 1995; Long, Waller & Plater, 2006*a*; Hijma & Cohen, 2019).

There has been considerable debate as to the influence of RSL movements on the formation of coastal peats. Basal peats form under the influence of rising water levels and can preserve a record of increasing *E* with eutrophic vegetation communities indicative of increasing water depth (a retrogressive series; Behre, 1986) occurring successively over time (Fig. 5A). In such circumstances rising RSL is providing ‘accommodation space’ into which the peat can accumulate. In addition to the rate of RSL rise and topography, other processes that add to or cause a reduction in the sediment column will influence whether peat formation continues over an extended time period. However, that rising RSL is conducive to peat formation, if other conditions are met, is suggested by the accumulation of thick deposits of fen origin in back-barrier and deltaic settings in a region that experienced a general trend of rising RSL during the Holocene (see Section IV).

The formation of peat above intertidal sediments (terrestrialisation) has long been regarded as indicative of falling RSL, an interpretation that persists in the southern North Sea region (e.g. Behre, 2007). Issues such as the reliability of the evidence for such falls (e.g. the use of sea-level index points potentially subject to compaction, see Section III.5) and the lack of known mechanisms for large-amplitude RSL fluctuations, particularly during the mid- to late Holocene, has led others to questions this (e.g. Baeteman, Waller & Kiden, 2011). The seaward extension of freshwater conditions clearly requires an increase in *E*, although in addition to RSL change this can also be produced by an excess supply of minerogenic sediment that fills the sediment column (Gerrard, Adam & Morris, 1984). Following a fall in RSL peat could fill any residual accommodation space, with communities indicative of decreasing water depth (a progressive series; Behre, 1986; Fig. 5B) occurring successively over time. However, the continued long-term accumulation of eutrophic peat is doubtful under falling or stable RSL. Freshwater input would need to exceed water loss through seepage, the latter being influenced by factors such as the depth, composition and degree of humification of the peat. Confined estuarine settings would be more suitable environments for this to occur than large back-barrier/deltaic environments. Relatively low rates of peat accumulation would be expected as would, with movement of water through the profile promoting acidification, the development of bog communities and eventually ombrotrophy.

More northerly regions, which have experienced isostatic uplift and provide evidence for the onward growth of peat in these circumstances, show such successional trends [e.g. the work of Tuittila *et al*. (2013) on eastern coast of the Gulf of Bothnia]. However, such areas will have experienced precipitation and evaporation rates that are more favourable for bog development than the southern North Sea region. Given the lack of independent evidence for large-amplitude falls in RSL we interpret the presence of progressive series in the terrestrialisation peats of the southern Northern Sea region as indicating a state where peat accumulation exceeded the rate of RSL rise, a situation (along with ombrotrophy) that would be expected to become increasingly common as the rate of RSL rise declined in the late Holocene (see Section IV.4).

In addition, it cannot simply be presumed that the water table in freshwater areas reflects coastal sea level. Higher or lower water levels can be created by a number of processes (van de Plassche, 1982; Vink *et al*., 2007; Kiden, 1995; Kiden, Makaske & van de Plassche, 2008; Vis *et al*., 2015). These include the effect of estuary morphology on tidal amplitude, where high water level increases up-estuary as the tidal system narrows. There is also the ‘flood basin effect’, which dampens the tide range as a result of water storage in tidal basins, and the ‘river gradient effect’ whereby, as a result of a sloping groundwater table, local water levels increase upstream along a tidal river. Channel migration (avulsion) in river deltas can also influence water levels. For example, new channels may lack well-developed levees, with consequent effects on floodplain groundwater levels.

(3) The organic component

In addition to external (allogenic) influences on the water table, the development of peatlands is influenced by internal (autogenic) processes, most notably by the change in *E* brought about by the accumulation of peat (hydroseral succession). The quantity of autochthonous organic sediment added through time ($\frac{dSorg}{dt}$) is determined by the difference between net primary production and losses through decay, herbivory and, in areas influenced by tides, the export of litter. Vegetation type and water level are likely to be important in determining rates of production and decomposition, although both processes are influenced by a much wider set of factors.

For production, information from analogous communities and situations to coastal peatlands is lacking, as are data on below-ground productivity in general. However, productivity is influenced by nutrient availability and therefore in general terms should increase along a gradient from bog to fen, with some authors indicating that the highest productivity occurs in wooded fens (Szumigalski & Bayley, 1996). With nitrogen potentially a limiting factor, nitrogen fixation by fungi (*Frankia* spp) in the root nodules of *Alnus glutinosa* is likely to influence the relative productivity of carr communities. In terms of water level, productivity increases with water depth in certain herbaceous peatland types but decreases in shrub communities (e.g. Thormann & Bayley, 1997), with evidence that the growth rate of wood in *Alnus* carr decreases from wet to very wet conditions (Schäfer & Joosten, 2005).

However, rather than high primary productivity, it is low decomposition rates that are largely responsible for the accumulation of organic matter in fens and bogs (Clymo, 1984). As well as environmental factors, the chemical composition of the plant material has a strong influence on the processes that cause decay. Decomposition can occur syn- and post-depositionally and ultimately transforms fibrous peat into a black amorphous homogeneous deposit with a reduced volume. With aerobic decay faster than anaerobic, surface aeration caused by fluctuating water levels and drainage will significantly increase rates of decomposition (see Section V.1). In general terms, decomposition is slower in bogs than fens (Aerts, Verhoeven & Whigham, 1999), with rates for *Sphagnum* notably lower than for other growth forms. Wood and roots decay at slower rates than leaf litter (Clymo, 1983), while rates for the latter are higher in deciduous wet woodland than herbaceous fen (Aerts *et al*., 1999). Barthelmes (2009) suggests that rates are relatively high for *Alnus glutinosa*-dominated communities as a consequence of fluctuating water levels, the transportation into and release of O_2 by deep roots, and high nutrient availability.

The accumulation of partially decomposed vegetation results in large amounts of carbon sequestration in peatlands (see Section V.3). As a result of the balance between production and decomposition, in general terms fens show similar or lower carbon sequestration rates compared to bogs (see Table 1 and Limpens *et al*., 2008). Peat derived from 'lowland fens' (mostly coastal lowland deposits and therefore with a complex genesis) is thought to account for between a quarter and a third of the total soil carbon stored within UK peatlands (Natural England, 2010), with carbon storage within the Fenland basin estimated at around 53 Tg (Holman & Kechavarzi, 2011).

(4) The minerogenic component

The minerogenic component of coastal peats may be fluvial, marine or aeolian in origin. However, with peat formation associated with freshwater environments, in estuarine/deltaic settings the non-organic component is most likely to be fluvially derived and, with peat grading into the fluvial minerogenic domain, forms an increasing proportion of the sediment matrix landward (Fig. 6). In such settings, variations in sediment supply, and therefore catchment processes (climate and land-cover changes), are a potentially important influence on the vertical distribution of the minerogenic component (Burrin & Scaife, 1984; Buckland & Sadler, 1985). Other factors that control the proportion of minerogenic sediment in coastal peats include the elevation of sediment surface, proximity to channels and the net productivity of the plant community. Reedswamp, sedge fen and fen carr will all occur over mineral substrata and unsurprisingly there is considerable variation in the minerogenic content of sediment derived from these communities. For example, organic content of the peat units present in the valleys feeding into Romney Marsh is highly variable (Fig. 6) and can be as low 30% (dry mass), with similar figures reported from sediment bodies described as peat elsewhere (Bos *et al*, 2012). Oligotrophic, and particularly ombrotrophic communities, isolated from sediment-rich water sources produce peats with a low minerogenic content (e.g. Little Cheyne Court; Fig. 6). Irrespective of the community, the minerogenic content of basal peats is often high immediately above the bedrock surface, which can be attributed to bioturbation in the early stages of peat formation over a terrestrial soil.

Minerogenic (and detrital organic) sediment not only provides nutrients but also fills accommodation space. By increasing *E*, the deposition of such sediment within marine–brackish environments is required to create suitable conditions for peat to form above sub-tidal sediments. In addition, during peat formation it has been suggested that, under conditions of rising RSL, the addition of allochthonous material can be important in preventing ‘drowning’, with the shutting off of sediment supply providing an explanation for the cessation of peat formation (Koshers, Chmura & Bailey, 1987). Neubauer (2008), however, notes that per unit mass, organic material contributes four times more to sediment volume as a result of water and air held in interstitial spaces.

(5) Compaction

Compaction refers to the reduction in volume of sediment that occurs as a result of sediment burial, self-weight and any subsequent loading and is caused by biological and chemical (decomposition), as well as physical processes (van Asselen, Stouthamer & van Asch, 2009; Brain, 2015). Peats are particularly susceptible to compaction and deposits with a high organic content can experience a reduction in volume of up to 90% (Shennan & Horton, 2002). The lowering of *E* that may result has important implications, both for the evolution of the coastal wetlands and for palaeoenvironmental and RSL reconstructions that utilize coastal peat deposits.

Peat compaction can, by reducing sediment volume, result in continued, renewed or increased rates of sedimentation (Haslett *et al*., 1998; Allen, 1999). If sediment accumulation keeps pace with the creation of accommodation space, compaction need not result in land subsidence (van Asselen *et al*., 2009). However, water-table lowering can result in the differential compaction of lithologically varied sediments and produce subsidence, which affects both wetland topography and evolution. Preferential compaction of peat-dominated sequences can magnify the elevation of channel and levee deposits composed of sand and silt. Such features, known locally in the East Anglian Fenlands as 'roddons', have been widely reported across the region (Godwin, 1938; Green, 1968; Vos & Van Heeringen, 1997; Berendsen, 2007; Smith *et al*., 2010), and are often co-incident with patterns of early settlement. Differential compaction can also produce channel avulsion, bringing river systems to new areas (van Asselen *et al*., 2009) and raise groundwater levels and thereby change the distribution of vegetation. For example van Asselen, Cohen & Stouthamer (2017) report the replacement of a highly organic wood peat by a low-organic reed peat in the vicinity of an avulsed channel.

Peat compaction distorts sediment accumulation rates (see Section IV.5) and in sea-level research lowers the elevation of sea-level index points, that is sediment or fossil remains, the deposition of which was controlled by palaeo sea level (Shennan, Long & Horton, 2015). Mitigation is therefore important and there is increasing interest in using geotechnical modelling to make corrections, although these require local calibration (van Asselen *et al*. 2009; Brain, 2015) and will increase altitudinal (vertical) error ranges. Sea-level researchers have long made use of basal peats (Jelgersma, 1961; van de Plassche, 1982; Gehrels, 1999; Meijles *et al*., 2018) as such layers usually directly overlie incompressible substrates. Peat initiation in such circumstances can however be influenced by local groundwater conditions and therefore factors such as topography and substrate permeability. For sea-level reconstructions, Vis *et al*. (2015) advocate using the uppermost part of such layers, combined with a good understanding of palaeogeography, to eliminate points with anomalous elevations unrelated to RSL.

IV. TEMPORAL AND SPATIAL TRENDS

(1) Introduction

Many authors have attempted to identify trends in peat formation across the southern North Sea region by correlating deposits using stratigraphic position and chronostratigraphy, both as individual units and schemes of stratigraphic subdivision, notably the Calais–Dunkirk system (De Jong, 1971; Roeleveld, 1974; Behre, 2003, 2007). Such schemes were underpinned by the belief that regional reductions in RSL determined peat formation and consequently they have now largely been abandoned (Streif & Zimmermann, 1973; Baeteman, 1981; van Loon, 1981; Denys, 1999; Wheeler & Waller, 1995; Ebbing, Weerts & Westerhoff, 2003; Weerts *et al*., 2005). Peat has been recorded from a large number of coastal locations in the region with considerable variation in the quantity of the litho-, bio-, and chronostratigraphic data available. Therefore rather than attempting to correlate sequences from many different areas, in the following discussion emphasis is placed upon regions where palaeogeographic maps, showing broad changes in the extent of peat formation, are available.

The maps utilized are from three contrasting coastal settings. These are: the Fenland basin (Waller, 1994*b*), where the coast is believed to have been open throughout the Holocene (Shennan, 1986*b*), the Romney Marsh area in the eastern English channel (Long *et al*., 2006*a*; Long, Waller & Plater, 2007), where peat formation occurred in back-barrier environments and the adjacent valleys, and the Netherlands (Vos, 2015). In the Netherlands, in addition to barrier islands, complexity is added by the presence of the Rhine–Meuse delta and, in the north, by the coalescence of coastal and inland peatlands and differences in tidal range along the coast.

One of the difficulties in comparing peat sequences among different depositional complexes is highlighted by this selection: that of scale. The Romney Marsh area comprises *c*. 270 km^2 (Eddison, 2000), the Fenland basin *c*. 4,000 km^2 (Waller, 1994*b*) and the coastal plain of the Netherlands *c*. 15,000 km^2 (Pons, 1992). In general terms, small systems are more prone to local influences with such differences likely to influence peat initiation (e.g. the amount of sediment required to fill the accommodation space to raise *E*), onward growth (influencing for example the freshwater input relative to the size of the peatland) and cessation (with peripheral areas in large wetlands less likely to be susceptible to marine inundation). The preservation of sequences is also likely to be influenced by scale: for example Waller & Long (2010) suggest that the absence of peat from some of the narrow Sussex valleys may reflect the potential for channel migration to rework sediment in such confined settings.

Limitations in the construction of the palaeogeographic maps include the quantity of data available for the early Holocene and the upper part of the stratigraphic column being preferentially impacted by various destructive processes, as outlined in Section V. In addition, with the timing of events based on radiocarbon dates that cannot be resolved to less than a few hundred years, correlating events within individual basins is inevitably subject to chronological uncertainties. The presence of bog vegetation is shown using the palaeoenvironmental data summarised in Waller (1994*b*), Long *et al*. (2007) and Pons (1992). Such evidence is however particularly vulnerable to destruction, inevitably less abundant than lithostratigraphic data, and the boundaries of such communities are inherently difficult to define.

(2) Trends in peat formation

In the early Holocene, prior to the connection of the North Sea with the English Channel (*c*. 9000 cal. yrs BP; e.g. Shennan *et al*., 2000), the Fenland rivers fed into a southward expanding North Sea coastline while the eastern English Channel and the Rhine–Meuse palaeovalley were inundated from the south-east (Fig. 1). Thin basal peats have been recorded from both basins (e.g. Jelgersma, 1961; Shennan *et al*., 2000; Brown *et al*., 2018). Hijma & Cohen (2011, 2019) provide detailed palaeogeographic maps showing that basal peats/organic palaeosol layers are commonly present in the Rhine–Meuse palaeovalley. However, they caution that only those occurring over permeable surfaces (e.g. on the flanks of aeolian dune fields in the Rhine–Meuse palaeovalley) are likely directly to reflect rising water levels in response to RSL (see also

Hepp *et al*., 2019). More surprisingly, terrestrializing peats have also been recorded from both the Rhine–Meuse delta and the south coast of England. The latter are also situated in palaeovalleys, the Arun (Gupta *et al*., 2004) and at Eastbourne (Jennings & Smyth, 1987), and range in age from *c.* 11,200 to 9650 cal. yrs BP. They imply that sufficient minerogenic sediment was locally available to fill the accommodation space created by RSL rise and bring the elevation of the sediment surface within the range in which organic sedimentation could occur, allowing brief periods of peat growth before continued RSL rise resulted in further flooding. Jennings & Smyth (1987) considered that a protected setting, coupled with early temporary barriers were required at Eastbourne, however, the Rhine–Meuse and Arun peats are seen as being distal to active tidal channels (Gupta *et al*., 2004; Hijma & Cohen, 2019).

The connection of the English Channel to the North Sea resulted in what are likely to have been rapid changes in tidal regime and patterns of sediment movement in the region at the beginning of the mid-Holocene (*c.* 8200–4200 cal. yrs BP). Initially rates of RSL rise remained high and peat development relatively limited. Sufficient data are available to map the distribution of basal peat across the Netherlands *c.* 7450 cal. yrs BP (Fig. 7), which was particularly extensive in the vicinity of the Rhine–Meuse delta (Vos, 2015). Elsewhere terrestrializing peats appear still to be confined to valley situations, with the earliest recorded in Fenland occurring at Adventurers' Land *c.* 7200 cal. yrs BP and Welney Washes and Spalding *c.* 6700 cal. yrs BP (Shennan, 1986*a*; Waller, 1994*b*), in the Thames estuary *c.* 7900 cal. yrs BP (Devoy, 1979) and in the Romney Marsh area *c.* 7800 cal. yrs BP (Waller & Kirby, 2002). In general terms, however, the scarcity of terrestrializing peats in the stratigraphic record during the early part of this period is an indication that away from valley situations and the availability of local source material (as indicated for Fenland by Brew *et al*., 2000), the creation of accommodation space as consequence of RSL rise continued to outstrip sediment supply.

River valley locations remain the focus of peat formation until after *c*. 6000 cal. yrs BP with the onset of extensive peat formation in the Romney Marsh area beginning in the Brede and Rother valleys from *c.* 6600 cal. yrs BP (Waller, Burrin & Marlow, 1988) and the sea and landward expansion of peat in the Rhine–Meuse delta by 5800 cal. yrs BP (Vos, 2015). The essential prerequisite to provide a platform upon which peat could expand seaward, i.e. the build-up of sediment onshore, combined with further reductions in the rate of RSL rise, seem not to have been more generally met in the region until after *c.* 5500 cal. yrs BP.

Extensive peat formation began over the western side of the Romney Marsh basin at this time, probably promoted by the growth of a gravel barrier (Fig. 8; Long *et al*., 2006*a*). In the Netherlands, peat development over the central tidal basin of the Oer-Vecht had taken place by 4700 cal. yrs BP (Vos, 2015), while the first major seaward expansion of a thin peat in western Fenland reached its maximum extent *c.* 4600 cal. yrs BP (Fig. 9; Waller, 1994*b*).

At the beginning of the late Holocene (*c.* 4200 cal. yrs BP to present), in the Romney Marsh area and Zeeland and North Holland, terrestrializing peats continued to expand seaward reaching the coastal barrier systems *c.* 3300 and *c.* 3450 cal. yrs BP respectively (Long *et al*., 2006*a*; Vos, 2015). By contrast, possibly a consequence of vulnerability to flooding events as a result of a more open coast, the Fenland basin experienced fluctuations between freshwater and marine brackish conditions. The maps (Fig. 9) suggest differences between the south-eastern and south-western parts of the Fenland basin *c.* 4600 cal. yrs BP and *c.* 3700 cal. yrs BP, with marine conditions reaching their maximum inland extent in one area while peat expanded seaward in the other (Waller, 1994*b*). Spatial variability is also recorded in the Netherlands, where peat expansion over the tidal basins in the north began in the late Holocene 1000–2000 years later than in the western Netherlands (Vos, 2015). In the Netherlands, differences in tidal range along the coast (Van der Molen & De Swart, 2001) may provide a partial explanation, while in the smaller systems such disparities are most likely to reflect variation in the supply of sediment to the shoreline although this requires further research. More certainly the relatively rapid development of peat over large parts of the Fenland basin (e.g. *c.* 3400 and 2900 cal. yrs BP) suggests a pre-peat surface of tidal flats, while the markedly diachronous peat growth recorded in the Romney Marsh area (from *c.* 6600 cal. yrs BP in the valleys to its maximum extent *c.* 3000 cal. yrs BP) is indicative of the gradual and progressive infilling of a basin. In both Fenland and the Netherlands, the palaeogeographic maps (Figs 9 and 7) show that the margin of peat growth continued to encroach landward during the mid and well into the late-Holocene (this is not evident on the Romney Marsh maps due to the much steeper topography). The rising water levels implied are consistent with the trend of rising RSL (albeit rising at a declining rate). In the Netherlands this resulted in the coastal peatlands coalescing with inland bogs (e.g. Bourtanger Moor at *c.* 3500 cal. yrs BP). Pons (1992) indicates their maximum extent was attained *c.* 3200 cal. yrs BP when almost all of the coastal plain from Calais to southern Denmark was peatland.

The subsequent history of the peatlands of the North Sea region is one of diminution as a result of freshwater and marine flooding and human activity. For example, in the delta of the Rhine–Meuse, peat was increasingly replaced by the deposition of fluvial clays from *c.* 3500 cal. yrs BP onwards (Vos, 2015). It is likely that the progressive removal of catchment woodland from the Bronze Age onwards resulted in the erosion of soils and increased alluviation (Burrin & Scaife, 1984; Buckland & Sadler, 1985; Smyth & Jennings, 1988; Erkens, 2009). Such processes are also likely to have been an important sediment source for material deposited under marine–brackish conditions.

The late Holocene saw major marine incursions in all three regions. Whilst locally variable in composition, silty clays typically 1–3 m thick form the uppermost unit in areas fringing the coast. Given the decline in the rate of RSL rise, compaction and the erosion of the underlying peat is likely to have provided the accommodation space required for deposition of these sediments (Long, Waller & Stupples, 2006*b*). Deposition is frequently associated with tidal channels that are incised into the underlying organic deposits (Baeteman, 2005). Consistent dates can therefore be difficult to obtain from the upper contact of these peats (Waller, Long & Schofield, 2006), although inundation appears to have occurred in Fenland *c.* 1900 cal. yrs BP and on the eastern side of the Romney Marsh area after *c.* 1800 cal. yrs BP (Waller, 1994*b*; Long *et al.*, 2006*a*,*b*). In the Netherlands, the area covered with peat also contracted from *c.* 1850 cal. yrs BP with marine inlets developing in Zeeland and at the Rhine–Meuse mouth (Vos, 2015). By *c.* 1150 cal. yrs BP Zeeland had been extensively inundated.

A number of potentially interacting natural and anthropogenic factors have been suggested as drivers of these inundations. The natural factors include the depletion of near-shore sediment sources and resultant erosion of barrier systems, an increase in storm conditions and a lack of vertical peat accumulation due to the slow RSL rise lowering the peat surface relative to the tidal frame, making the peatlands vulnerable to inundation (Long *et al.*, 2006*a*; Hamilton *et al.*, 2019). Baeteman (2005) suggests excessive run-off during the late Holocene (either relating to climatic deterioration or a change in hydrological conditions connected to anthropogenic deforestation) resulted in the flushing of sediment from channels draining the peat, enabling tidal conditions to re-enter these areas. Others have implicated human activity more directly, with inundation initiated by peat cutting (Vos & van Heeringen, 1997) or subsidence occurring as a consequence of drainage associated with early reclamation (Pierik *et al.*, 2017).

Human activity has certainly played an important role in the history of these areas over the last 1000 years (Section V). Organic sedimentation ceased over large parts of central Holland from c. 1100 cal. yrs BP as a result of reclamation, and the subsequent reconstructions of Vos (2015), from *c.* 450 cal. yrs BP through to the modern era (Fig. 7), show the further diminution of coastal peatlands due to human intervention. Peat formation effectively ended in the Romney Marsh region when the unreclaimed western side was flooded *c.* 1200 cal. yrs BP as a consequence of barrier breaching (Fig. 8; Long *et al*., 2006*a*). However, in Fenland human intervention appears not to have prevented a renewed phase of extensive peat growth from *c.* 1650 cal. yrs BP (Fig. 9; Waller, 1994*b*), and although this uppermost deposit is now poorly preserved/absent, over large areas of southern Fenland active peat formation appears to have continued until drainage from *c.* 400 cal. yrs BP onwards (Darby, 1983).

(3) Vegetation trends: basal peats

While predicated by rising water levels, the nature of basal peats is controlled by local groundwater conditions and influenced by factors such as topography, substrate permeability and water source. In the Rhine–Meuse palaeovalley, from *c.* 10,000 to 7000 cal. yrs BP, Bos *et al*. (2012) record the presence of gyttja and a range of peat types reflecting the variety of water sources. Oligotrophic peats are rare but occur over cover-sand ridges and are fed by precipitation. In the mid-Holocene slower rates of RSL rise and areas with shallow slopes provided the opportunity for the development of thicker peats accumulating at the landward margins of sedimentation. In the northern Netherlands and eastern Belgian coastal plain the nutrient-poor cover-sands enabled the development of meso-oligotrophic bog, with communities dominated by *Eriophorum* and *Sphagnum* occurring above *Phragmites* or *Betula* carr (Pons, 1992; Allemeersch, 1991). A rare example of such vegetation in Fenland occurred in Holme Fen basin where oligotrophic bog developed from *c.* 5700 cal. yrs BP (Godwin & Vishnu-Mittre, 1975). Comparable ages have been obtained for basal peat development in eastern Belgium (Allemeersch, 1991) and the onset of acidification in the northern Netherlands (from *c.* 5300 cal. yrs BP; Pons, 1992).

By contrast, over most of Fenland and in the Romney Marsh area, basal peats and the peats that remained beyond the landward marine limit (the latter are poorly preserved) largely appear to have been eutrophic. Marine incursions into the basins of southern Fenland over the period *c.* 5400–3400 cal. yrs BP gave rise to basal peats *c.* 30–40 cm thick, with eutrophic communities indicative of increasing water depth occurring successively over a few hundred years (Fig. 5A) as rising water levels outstripped sediment supply. Southern Fenland is also noted for an abundance of ‘bog oaks’, trees that have emerged at the surface, beyond the landward limits of marine sedimentation, since drainage. The prevalence of oak in such situations is likely to reflect relative tolerance to waterlogging, although may also be influenced by preservation bias with early workers (Godwin, Godwin & Clifford, 1935; Godwin & Clifford, 1938) indicating the presence of additional taxa.

(4) Vegetation trends: territorialisation peats

The terrestrializing peats of the region record progressive series in their initial stages of growth. Saltmarsh communities are replaced by reedswamp (with visible *Phragmites* remains common), and in turn these are replaced by sedge-dominated communities. Where macrofossil data are available, *Cladium mariscus* and *Carex* spp., notably *Carex elata* (a eutrophic species associated with the margins of open water) are frequently recorded (Allemeersch, 1991).

Above the initial stages, a distinct eutrophic pathway can be identified with *Alnus*- and *Salix*-dominated fen carr communities becoming established and being sustained for long periods (Waller *et al*., 1999). Such communities developed (and bog communities were absent) where fluvially derived water is likely to have retained a dominant influence, in confined valley settings such as the Sussex Ouse valley (Waller & Hamilton, 2000; Waller & Early, 2015), the western valleys of Romney Marsh (Waller, 1994*a*) and the Thames estuary (Devoy, 1979; Branch *et al*., 2012; Waller & Grant, 2012). On the Rhine–Meuse deltaic plain Van der Woude (1984) describes a fluviolagoonal environment, consisting of a mosaic of fen carr, *Phragmites* swamp and lakes, that was present for much of the mid-Holocene. In south-eastern Fenland open eutrophic communities also persisted in the thicker terrestrializing peats of the late Holocene (Waller, 1994*b*), occurring both above fen carr deposits (e.g. Redmere) and throughout peat formation (Welney Washes).

A wide variety of vegetation types that fall within the category of bog have been recorded within terrestrializing peats. *Betula*-dominated fen carr commonly occurs and can be preceded by both open and woody fen (e.g. Allemeersch, 1991; Waller *et al*., 1999) and open bog communities (e.g. Pons, 1992). On both the coastal plains of eastern Belgium and the Netherlands, communities with *Carex*/*Eriophorum* and mosses indicative of meso-oligotrophic conditions commonly occur above the initial terrestrializing stages (Allemeersch, 1991; Pons, 1992). Such communities often merge into vegetation in which Sphagna are abundant. The presence of *Sphagnum* Section Acutifolia and Section Cuspidata, which are frequently recorded alongside mesotrophic taxa, indicates these communities were not all initially ombrotrophic. *Sphagnum* subsequently dominates, occupying the upper, and often greater, parts of the terrestrializing deposits of Belgium (>60%; Allemeersch, 1991) and the Netherlands (Pons, 1992). Here, raised bogs occupied most of the coastal plain in the late Holocene. Ombrotrophic conditions are confirmed by the presence of taxa such as *Sphagnum imbricatum* which also demonstrates the presence of ombrotrophic bog at Little Cheyne Court on Walland Marsh (Fig. 8; Waller *et al*., 1999). Pons (1992) provides descriptions of the bogs of the Netherlands envisaging a complex of meso-scale bogs intersected by mesotrophic and eutrophic communities along drainage routes. Their size reflects the much more extensive coastal plain, that over Schouwen, Zeeland is envisaged as having a diameter of 12 km, occupying 12,000 ha, with the dome rising to *c.* 2.5 m above modern sea level.

In Fenland, aside from the Holme Fen sequence, there is no evidence that sequences progressed beyond meso-oligotrophic bog, although the highly humified nature of the late Holocene deposits generally prevents the identification of macrofossils. Pollen assemblages with *Sphagnum* also contain an abundance of Ericaceae, Cyperaceae and *Pinus* pollen (Godwin *et al*., 1935; Waller, 1994*b*). The extent of such bogs in the late Holocene in Fenland is largely a matter of conjecture due to their subsequent destruction (Section V), although is likely to have been greater than indicated in Fig. 9 and they may have been widespread (Godwin, 1978). On the northern and western fringes of Walland Marsh mesotrophic vegetation developed in the late Holocene with *Myrica gale* an important vegetation component, occurring above fen carr (both *Alnus*- and *Betula*-dominated). With its establishment requiring either dry elevated surfaces or little lateral movement of water (Wheeler, 1980*b*), it appears to form an alternative minerotrophic pathway (Long *et al*., 2007; Kirby *et al*., 2010).

As Godwin & Clifford (1938) recognised, isolation from eutrophic water sources appears to be the critical factor in enabling the widespread development of bog vegetation. Initial development occurs where there is spatial separation from such sources, either basins remote from the major rivers (e.g. the Holme–Woodwalton and Wood Fen basins in Fenland) or during periods when peat formation would have been particularly extensive in a region (e.g. Walland Marsh). Bogs did not become widespread in the region until the late Holocene: *c.* 2800 cal. yrs BP in Fenland (Waller, 1994*b*), *c.* 3000 cal. yrs BP on Walland Marsh (Waller *et al*., 1999), and from *c.* 4600 cal. yrs BP in North Holland and Zeeland (Pons, 1992). That their most extensive development occurs in the upper part of the Holocene sequence suggests that vertical separation – the ability of peat to keep pace or outgrow rising water levels (through autogenic succession) – is also likely to have been an influential factor.

Not only was the rate of RSL rise declining but precipitation was increasing. Climate change, the shift from relatively warm continental to wetter oceanic conditions in the period *c.* 2800–2700 cal. yrs BP (Kilian, van der Plicht & van Geel, 1995), appears to have been a significant factor promoting acidification. Witte & van Geel (1985) record a shift from *Phragmites* to mesotrophic *Molinia* at Assendelver Polder at this time, and it is also likely to have resulted in the switch from minerotrophic to ombrotrophic conditions in many pre-existing bogs (e.g. Waller *et al*., 1999). In modern mire systems the shift from fen to bog can occur within a few decades if vertically mobile surfaces (floating mats of herbaceous vegetation) develop and prevent flooding with eutrophic water (Giller & Wheeler, 1988; Van Digglen, Molenaar & Kooijman, 1996).

(5) Stability

Stratigraphic and palaeoecological data suggest that stability, both in terms of continuous peat growth and individual communities, can be measured over not just a few hundreds of years but several thousands of years in the southern North Sea region. In the case of the extensive accumulations of peat within ombrotrophic communities, in the upper part of sequences such stability is unsurprising. This would be expected as a result of the independence of such bogs from minor fluctuations in the groundwater table and, under at least the climatic conditions of the late Holocene (with active peat accumulation continuing in parts of the Netherlands), ombrotrophic peat growth being self-sustaining.

More interestingly the data also suggest the long-term (>1000 cal. yrs) stability of fen vegetation, both carr and herbaceous communities. Waller (1993, 1994*a*), Deforce (2011), Branch *et al*. (2012), Deforce *et al*. (2014) and Waller & Early (2015) all record the extended presence of fen carr during the mid-Holocene (with sustained sediment accumulation rates >2 mm yr^{-1} recorded), while the persistence of areas of swamp and herbaceous communities is indicated by Waller (1994*b*), Long & Innes (1995), Waller *et al*. (1999) and

Jennings *et al*. (2003) in the late Holocene. This raises the question as to what factors enabled these plant communities to sustain a sediment surface elevation within their range of tolerance over an extended period of time. In the mid-Holocene the rate of RSL rise had evidently declined to the point where in protected settings the accumulation of organic material was capable of filling the accommodation space created. However, as the rate of rise is unlikely to have remained constant over long periods of time, other factors are likely to have been involved. These include the wide ecological amplitude in respect of the height of the water table of many of the dominant plants including *Alnus glutinosa* (see Table 1). In addition, higher rates of decomposition would be expected if the supply of organic material outstripped the rate of water table rise, while increased flooding, and therefore potentially the supply of additional allochthonous sediment, would be expected if RSL rise outstripped organic production. Data from the valleys leading into Romney Marsh (Fig. 6) show that while the minerogenic component is highly variable, temporal trends are difficult to discern. Unfortunately, the distortion of sediment accumulation rates by compaction and a lack of detailed chronologies hinder our ability to analyse the sedimentary responses to water-level changes. In addition, although methods are available to determine the degree of peat decomposition or 'humification' (as a proxy for hydrological change), these have been considered not to be worthwhile in coastal areas that have been subject to modern drainage. Factors that may limit the successional trend towards fen carr in open eutrophic communities, with herbaceous peats tending to persist in seaward locations, include exposure and high salinity. The inability of thick peat sequences to support the weight of large trees may also promote and sustain a trend towards open vegetation through time (Waller & Early, 2015) with subsidence an additional factor in creating the accommodation space needed to maintain eutrophic peats into the late Holocene.

V. EXPLOITATION, CONSERVATION AND POTENTIAL RESURRECTION

(1) Exploitation

During the late Holocene the much-reduced areas of peatland in the Netherlands and Fenland were subject to extensive human exploitation. Peat cutting to provide fuel for domestic and industrial purposes (e.g. salt production) was practiced across the region from the Roman period onwards (Hall & Coles, 1994; Vos & van Heeringen, 1997; Baeteman, 2007). Becoming increasingly extensive in medieval times, the custom continued into 20th century AD (Borger, 1992). Flooding of the resultant turbaries produced areas of open water, notably the Norfolk Broads (Lambert *et al*., 1961), with exploitation continuing below water level in the central Netherlands (Borger, 1992). Other widespread practices included mowing, the cutting of reeds, rushes and sedges for construction materials (particularly thatching), and the use of peatland for summer pasture, fishing and wildfowling (Rippon, 2000). In the Netherlands, areas of peat remained sufficiently above the groundwater table to enable, if fertilised, grain cultivation into the 14th century AD (Borger, 1992).

Reclamation through the use of dykes, ditches and sluices extends back in the region to the late Iron Age (Lascaris & De Kraker, 2013), although settlement in tidal environments continued along the Belgian coastal plain into the 13th century AD (Tys, 2013). The use and spread of more-advanced water-management techniques from the 17th century AD onwards, including networks of pumping stations and areas set aside to retain flood water (e.g. the Fenland Washes), facilitated agricultural use of the remaining peatland areas. The cycle of events initiated by such drainage is well documented (e.g. Hutchinson, 1980). The loss of water results in contraction of the sediment volume and subsidence. Oxidised peat begins to decompose and those constituents not converted to volatile products are washed or blown away: a process that is collectively referred to as ‘peat wastage’. This lowering of the land surface then necessitates further action to lower the water levels. In coastal areas, the lack of sedimentation in the enclosed areas in itself contributes to land lowering relative to the coastal zone, increasing vulnerability to flooding from storm surges (Section IV.2).

Peat wastage rates are highly variable, and depend on site-related factors such as peat thickness and land use. In Fenland, the Holme Fen post indicates a fall in the ground surface of 3.74 m between 1848 and 1978. However, the wider applicability of this often-quoted figure is doubtful, since the post provides a guide to subsidence rates over thick peat during the early dewatering phases (2.34 m between 1848 and 1871), but such deposits are rare in Fenland and subsequent land use around the post (pasture from as early as 1880s and woodland from the 1920s) is atypical (Hutchinson, 1980). In the Netherlands, Erkens, van der Meulen & Middelkoop (2016) have estimated peat-surface lowering as a result of wastage and excavation at an average of 1.9 m over the past 1000 years. Current estimated wastage rates in Fenland vary from 2.1 cm yr^{-1} for thick peat under arable land use to 0.1 cm yr^{-1} for thin peats under semi-natural vegetation (Holman &

Kechavarzi, 2011), while in the Netherlands, Acreman & Miller (2007) provide an estimate of 1 cm yr^{-1} for peatland under normal agricultural use.

Drainage and conversion to agriculture alters the carbon balance in peatlands from a sink to a source, resulting in the oxidative release of greenhouse gases (CO_2 and N_2O). The figures Holman & Kechavarzi (2011) provide for the current rate of release in Fenland at around 0.4 Tg C yr^{-1} are equivalent to about 0.4% of the UK's current industrial emissions of CO_2 (DBEIS, 2018).

(2) Conservation of the remnants

At the start of the 20th century the active coastal peatlands remaining in the region survived as remnant semi-natural ecosystems within agricultural settings. These consisted of fen vegetation, with only small areas of bog (<10,000 ha in 1985) persisting in the Netherlands (Vermeer & Joosten, 1992). Emphasis was placed first upon the acquisition and then the management of these remnants for biodiversity conservation. Management for conservation has frequently been aimed at preventing succession, maintaining minerotrophic vegetation by biomass removal. This has generally involved continuing or reinstating historic practices, particularly mowing, but also grazing and peat-cutting. The rationale for such practices is that vegetation such as fen meadow is rare; fen forms less than 1% of UK peatland (JNCC, 2011), and has a high biodiversity value (Van Wirdum, Den Held & Schmitz, 1992; Wheeler, 1988). In the absence of such management, nutrients build up and competitive fast-growing species soon dominate and begin to outgrow the groundwater level.

The natural successional trend towards a bog system is not the only challenge to maintaining the nutrient balance in peatland remnants. Draining of the surrounding agricultural land increases flow rates and, by lowering the hydrologic head, shifts the balance of water input from groundwater to precipitation. Acidification can also be accelerated by air pollution, particularly the influx of HNO_3. Consequent vegetation changes include the loss of minerotrophic vegetation with the invasion of *Sphagnum* and the development of embryonic bog (Van Digglen *et al*., 1996). Efforts to mitigate acidification include the removal of surface peat, liming and the use of manure (Vermeer & Joosten, 1992; Beltman *et al*., 1996; Leon *et al*., 2002; Van Digglen *et al*., 2015). Increased levels of nitrogen or phosphorus (i.e. eutrophication) are also a major threat, particularly to the surface waters associated with such remnants. External inputs arise through increased atmospheric deposition and run-off or drainage water derived from agricultural lands. In addition, the release of internal nutrient sources can occur as a result of low water levels through the mineralisation of surface peat (Koerselman & Verhoeven, 1995).

While management can be directed at maintaining past diversity and rare species, these aims are intrinsically threatened by the small size and fragmented distribution of the remnants. Size, with the individual communities within the remnants inevitably small, can prevent the long-term maintenance of viable populations and/or result in the genetic deterioration of small populations. The challenges posed can be illustrated through the fate of two species, the fen violet (*Viola stagnina*) and the large copper butterfly (*Lycaena dispar*) in the UK. In the early 1990s the violet was confined to two Fenland peatland remnants (Woodwalton and Wicken Fen), although only one plant was recorded in 2005 from Woodwalton and none were found for a 15-year period at Wicken Fen. The plant has a persistent seed bank and requires periodic disturbance (e.g. peat cutting), although in addition to competition from surrounding vegetation, the maintenance of pure populations is threatened by its readiness to hybridise with another *Viola* species (Porter & Foley, 2017). After UK extinction in the 19th century the large copper butterfly was reintroduced into Woodwalton Fen in 1927. The population was subsequently maintained, through the management of open fen communities, but also in part by protecting larvae and the supplementary release of adults. However, periodic prolonged flooding events killed the larvae and by the 1990s it was concluded that the fen was of insufficient size to maintain a self-sustaining population and attempts at reintroduction were abandoned (Barnett & Warren, 1995).

(3) Large-scale restoration and the future

These problems highlight the need for management at the landscape scale and recent years have seen the development of large schemes aimed at peatland restoration [see Bonn *et al*. (2016) for detailed discussion of this topic and further examples]. In coastal lowland regions these typically involve areas from which peat was formerly excavated (e.g. Thorne and Hatfield Moors comprising 3,300 ha in the Humberhead Levels and the Weerribben-Wieden, a 10,500 ha reserve in the central Netherlands) or the retransformation of agricultural land back into peatlands (e.g. The Great Fen and Wicken Fen Projects in the UK Fenland). Restoration of the former has generally involved raising the water table and revegetating areas of bare peat.

It is intended that the Fenland projects will also raise water levels and, by utilising differences in the underlying substrate and topography, create gradients within the landscape, and re-establish a variety of wetland communities. Greater connectivity is also an important element of these schemes. The Great Fen Project (initiated in 2001) will reconnect two existing remnants in Fenland (Holme Fen and Woodwalton Fen), creating, in addition to other habitats, 2250 ha of reed bed, tall herb fen and fen meadow (Mountford *et al*., 2002). It is nevertheless recognised that establishment of fen from agricultural land through natural recolonisation is likely to be a slow process and species reintroduction, through the use of seed, hay or animal vectors, will be required. The Wicken Fen Project (initiated in 1999), which will add an additional 5,300 ha of wetland to the current 255 ha area, is non-proscriptive about both the habitats to be created and the target species. Here, in what is intended to be a more financially sustainable form of management, free-roaming self-reliant animals ('naturalistic grazing') are being used to influence vegetation development (National Trust, 2009).

As well as benefits for biodiversity, such schemes have the potential to provide additional regulating ecosystem services, including turning areas from carbon sources to sinks. Unfortunately peatlands can exhibit initial increases in the emission of greenhouse gases after rewetting, with carbon loss occurring as a result of increasing pH and high rates of organic breakdown, and the higher water levels produced increasing emissions of CH_4 (Lamers *et al*., 2015). Nevertheless, it is has been estimated that the Great Fen Project will prevent an annual loss of 325,000 tonnes of CO_2 equivalents and over a 100-year period the area should act as a carbon sink, with the presence of sulphate-rich clays in the subsoil potentially mitigating against the production of CH_4 (Gauci, 2008). Other studies have also shown the potential for rewetted former agricultural pasture to reduce carbon losses over shorter timescales, providing the water table can be carefully managed and kept close to the peat surface throughout the year (Huth *et al*., 2018; Peacock *et al*., 2019).

Other ecosystem services result from the catchment-scale hydrological management that such schemes require. Areas of open water at the Weerribben-Wieden and envisaged in the Great Fen and Wicken Fen Projects are designed to enable excess upland run-off to be stored and used to control the degree of water level fluctuation. Such stores can additionally help with groundwater regulation and flood attenuation downstream. In addition, such wetland complexes act as nutrient sinks and purify water. Nitrogen and phosphorus pollution has been a particular problem at Weerribben-Wieden, which receives water derived from adjacent fertilised agricultural polders. Here through effective water regulation, the wetlands now act as a filter, so that the more isolated parts can sustain the water quality required for the persistence of biodiverse rich fen communities (Cusell *et al*., 2014).

Functioning peatlands also provide cultural and natural heritage services that include the educational value of continuing with historic management practices and preserving what can be extensive waterlogged archaeological remains (e.g. Hall & Coles, 1994) that occur both within and beneath the peat along with palaeoenvironmental records. They also afford outdoor recreational opportunities otherwise lacking in what are now heavily populated regions (National Trust, 2009).

The benefits of restoration schemes are challenging both to measure and monitor (Hughes *et al.*, 2016) and they inevitably involve trade-offs, both in terms of potentially conflicting ecosystem services (Acreman *et al*., 2011; Lamers *et al*., 2015) and their value when set against productive agricultural land and issues relating to food security. Nevertheless, Peh *et al*. (2014) suggest that in 2011 the Wicken Fen Project had achieved a net gain to society. With the cycle of wastage necessitating further drainage of agricultural land likely to be exasperated by future climate change and sea-level rise, the balance between costs and benefits is likely to increasingly favour peatland restoration (Alonso *et al*., 2012).

The current rate of global mean sea-level rise, estimated at 3.1 ± 0.3 mm yr^{-1} for the last 25 years (Cazenave, Palanisamy & Ablain, 2018) has not been experienced since the mid-Holocene, when the palaeoecological evidence shows not only the expansion, but also the long-term persistence of peatland including eutrophic vegetation across the southern North Sea region. Rising groundwater levels will provide the accommodation space for peat accumulation and the opportunity to recreate the eutrophic communities abundant in the past including fen carr. The presence of *Alnus glutinosa* in the early stages of rewetting can reduce CH_4 emissions (Huth *et al*., 2018). If unmanaged and therefore with biomass retained, fen carr communities have the ability to store large quantities of carbon, particularly when the water level is maintained close to the surface (Barthelmes, 2009; Table 1). *A. glutinosa* is also reported as showing long-term resilience (the ability of a system to tolerate change while retaining essentially the same function and structure) to climate change with stands regenerating after diebacks if environmental conditions improve

(Latałowa *et al*., 2019). As peatland areas are expanded, greater consideration should therefore be given to converting areas into *A. glutinosa* woodland either by planting or allowing succession to occur.

The ability of peatlands to response to water-table rises and provide ecosystem services adds considerable value to such projects and brings into question the use of heavy grazing regimes in coastal rewilding projects. For example, at Oostvaardersplassen, a 5,600 ha reclaimed polder in Flevoland, the desire to maintain the early successional stages for biodiversity conservation has been achieved through naturalistic grazing with herd size determined by availability of natural food. Such an approach limits the build-up of biomass, potentially negating the benefits that might arise through uninhibited peatland development.

Research on biodiversity and ecosystem functioning has demonstrated that plant traits interact with abiotic variables to modulate ecosystem properties (e.g. soil carbon, above-ground biomass) in peatlands (Lavorel & Garnier, 2002; Moor *et al*., 2017). Restoration that focuses on ecosystem services should therefore aim at restoring vegetation with traits that maximise specific functions. For example, carbon sequestration in fens may be enhanced by the presence of species with specific leaf traits (Carvalho *et al*., 2019*a*). Here, palaeoecological evidence has the potential to provide information about how different plant traits affect the rates of delivery of ecosystem processes through time. Numerical techniques are being developed that link plant functional information to pollen data (Reitalu *et al*., 2015) so that inferences about ecosystem properties such as biomass accumulation and decomposition rates can be made using Holocene pollen-based trait reconstructions from fen deposits (Carvalho *et al*., 2019*b*).

It is currently difficult to assess the long-term role of minerogenic sediment supply in maintaining coastal peatland. This is an important area for future palaeoenvironmental research as the re-establishment of a minerogenic component through natural flood regimes would require the considerable obstacles (e.g. costs) to restoration of free-flowing rivers and estuaries to be overcome.

VI. CONCLUSIONS

(1) The combination of rising RSL and the onshore movement of sediment in the early and mid-Holocene created conditions suitable for peat formation in coastal lowland areas adjacent to the North Sea. From *c.* 5500 cal. yrs BP onwards peat-forming vegetation, initially largely eutrophic communities, became

widespread. A decline in the rate of RSL rise and autogenic succession then promoted bog vegetation as these peatlands expanded to reach their maximum extent in the region *c.* 3000 cal. yrs BP.

(2) Natural processes, such as declining sediment supply at the coast and a slower rate of rise in RSL, contributed to the loss of peatland in the late Holocene. However, human interference has been both a major and long-term influence, without which extensive areas of peatland would exist today.

(3) Biodiversity conservation and a growing awareness of the ecosystem services provided by peatland communities (notably carbon storage in active peatlands) have led to increased efforts towards their restoration. This development is likely to receive further impetus as a result of future RSL rise. In addition to adding to the costs of maintaining current agricultural activities, rising RSL could provide the accommodation space needed for future peat accumulation and the opportunity for the restoration of a variety of wetland communities.

(4) Restoration would be aided by a better understanding of the factors that promote both peat formation and onward growth in coastal lowland situations. These include further investigations into the range of water levels that extant communities can tolerate and their relationship to tidal constants, the decompaction of sequences and more detailed chronologies. Such research would enable a better understanding of the sedimentary response to water-level changes, including the role played by minerogenic sediment supply and flood regime in maintaining conditions suitable for peat accumulation.

(5) Linking palaeoecological data and plant traits has the potential to provide information as to how ecosystem services can be maximised in peatland-restoration projects.

VII. ACKNOWLEDGEMENTS

We would like to thank Dr Kerry Brown at Kingston University and Dr Cecile Baeteman of the Royal Belgian Institute of Natural Sciences for their helpful comments on an earlier draft of this review, together with Patrick Kiden and an anonymous referee for their insightful suggestions. This paper is dedicated to Jill Eddison who did so much facilitate a greater understanding of the Romney Marsh area.

VIII. REFERENCES

ACREMAN, M. C., HARDING, R. J., LLOYD, C., MCNAMARA, N. P., MOUNTFORD, J. O., MOULD, D. J., PURSE, B. V., HEARD, M. S., STRATFORD, C. J. & DURY, S. J. (2011). Trade-off in ecosystem services of the Somerset Levels and Moors wetlands. *Hydrological Sciences Journal* **56**, 1543–1565.

ACREMAN, M. C. & MILLER, F. (2007). Hydrological impact assessment of wetlands. In *Proceedings of the ISGWAS Conference on Groundwater Sustainability, Spain, January 2006*. pp 225–255. http://aguas.igme.es/igme/isgwas/Ponencias%20ISGWAS/16-Acreman.pdf

AERTS, R., VERHOEVEN, J. T. A. & WHIGHAM, D. E. (1999). Plant-mediated controls on nutrient cycling in temperate fen and bogs. *Ecology* **80**, 2170–2181.

ALLEMEERSCH, L. (1991). Peat in the Belgian eastern coastal plain. *Aardkundige Mededelingen* **6**, 1–54.

ALLEN, J. R. L. (1990). The formation of coastal peat marshes under an upward tendency of relative sea level. *Journal of the Geological Society of London* **147**, 743–745.

ALLEN, J. R. L. (1999). Geological impacts on coastal wetland landscapes: some general effects of sediment autocompaction in the Holocene of northwest Europe. *The Holocene* **9**, 1–12.

ALLEN, J. R. L. (2000). Morphodynamics of Holocene salt marshes: a review sketch from the Atlantic and Southern North Sea coasts of Europe. *Quaternary Science Reviews* **19**, 1155–1231.

ALONSO, I., WESTON, K., GREGG, R. & MORECROFT, M. (2012). *Carbon storage by habitat: review of the evidence of the impacts of management decisions and condition on carbon stores and sources*. Natural England Research Reports NERR043. Natural England, Peterborough, UK.

BAETEMAN, C. (1981). An alternative classification and profile type map applied to the Holocene deposits of the Belgian coastal plain. *Bulletin Belgische Verenigingvoor Geologie* **90**, 257–280.

BAETEMAN, C. (2005). How subsoil morphology and erodibility influence the origin and pattern of late Holocene tidal channels: case studies from the Belgian coastal lowlands. *Quaternary Science Reviews* **24**, 2146–2162.

BAETEMAN, C. (2007). Roman peat-extraction pits as possible evidence for the timing of coastal changes: an example from the Belgian coastal plain. In *Landschap in ruimte en tijd* (eds J. J. BEENAKKER, F. H. HORSTEN, A. M. J. DE KRAKER & H. RENES), pp. 16–25. Aksant, Amsterdam.

BAETEMAN, C., WALLER, M. & KIDEN, P. (2011). Reconstructing middle to late Holocene sea-level change: A methodological review with particular reference to ‘A new Holocene sea-level curve for the southern North Sea’ presented by K.-E. Behre. *Boreas* **40**, 557–572.

BARENDREGT, A. & SWARTH, C. W. (2013). Tidal freshwater wetlands: variation and changes. *Estuaries and Coasts* **36**, 445–456.

BARNETT, L. K. & WARREN, M. S. (1995). *Species Action Plan: Large Copper Lycaena dispar*. Butterfly Conservation, Wareham.

BARTHELMES, A. (2009). *Vegetation dynamics and carbon sequestration of Holocene alder (*Alnus glutinosa*) carrs in NE Germany*. PhD Thesis. Ernst-Moritz-Arndt-Universität Greifswald.

BEHRE, K.-E. (1986). Analysis of botanical macro-remains. In *Sea-level research: a manual for the collection and evaluation of data* (ed. O. VAN DE PLASSCHE), pp. 413–433. Geo Books, Norwich.

BEHRE, K.-E. (2003). Eineneue Meeresspiegelkurvefür die südliche Nordsee. Transgressionen und Regressionen in den letzten 10.000 Jahren. *Probleme der Küstenforschungimsüdlichen Nordseegebiet* **28**, 9–63.

BEHRE, K.-E. (2007). A new Holocene sea-level curve for the southern North Sea. *Boreas* **36**, 82–102.

BELTMANN, B., VAN DEN BROEK, T., VAN MAANAN, K. & VANEVELD, K. (1996). Measures to develop a rich-fen wetland landscape with a full range of successional stages. *Ecological Engineering* **7**, 299–313.

BERENDSEN, H. J. A. (2007). History of geological mapping of the Holocene Rhine-Meuse delta, the Netherlands. *Netherlands Journal of Geosciences* **86**, 165–177.

BONN, A., ALLOTT, T., EVANS, M., JOOSTEN, H. & STONEMAN, R. (2016). *Peatland Restoration and Ecosystem Services: Science, Policy and Practice*. Cambridge University Press, Cambridge.

BORGER, G. J. (1992). Draining-digging-dredging; the creation of a new landscape in the peat areas of the low countries. In *Fens and Bogs in the Netherlands: Vegetation, History, Nutrient Dynamics and Conservation* (ed. J. T. A. VERHOEVEN), pp. 131–172. Kluwer Academic Publishers, Dordrecht.

BOS, I. J., BUSSCHERS, F. S. & HOEK, W. Z. (2012).Organic-facies determination: a key for understanding facies distribution in the basal peat layer of the Holocene Rhine-Meuse delta, The Netherlands. *Sedimentology* **59**, 679–703.

BRAIN, M. (2015). Compaction. In *Handbook of Sea-level Research* (eds I. Shennan, A. J. Long & B. P. Horton), pp. 452–469. Wiley, Chichester.

BRANCH, N. P., BATCHELOR, C. R., CAMERON, N. G., COOPE, G. R., DENSEM, R., GALE R., GREEN, C. & WILLIAMS, A. N. (2012). Holocene environmental changes in the Lower Thames Valley, London, UK: Implications for understanding the history of *Taxus* woodland. *The Holocene* **22**, 1143–1158.

BREW, D. S., HOLT, T., PYE, K. & NEWSHAM, R. (2000). Holocene sedimentary evolution and palaeocoastlines of the Fenland embayment. *Geological Society of London Special Publication* **166**, 253–273.

BROWN, A., RUSSELL, J., SCAIFE, R., TIZZARD, L., WHITTAKER, J. & WYLES, S. F. (2018). Late glacial/early Holocene palaeoenvironments in the southern North Sea Basin: new data from the Dudgeon offshore wind farm. *Journal of Quaternary Science* **33**, 597–610.

BUCKLAND, P. C. & SADLER, J. P. (1985). The nature of late Flandrian alleviation in the Humberhead levels. *East Midland Geographer* **8**, 239–251.

BURRIN, P. J. & SCAIFE, R. G. (1984). Aspects of Holocene valley sedimentation and floodplain development in southern England. *Proceedings of the Geological Association* **95**, 81–96.

CARVALHO, F., BROWN, K. A., WALLER, M. P. & BOOM, A. (2019*a*). Leaf traits interact with management and water table height to modulate ecosystem properties in fen peatlands. *Plant and Soil* **441**, 331–347.

CARVALHO, F., BROWN, K., WALLER, M. P., BUNTING, M. J., BOOM, A. & LANG, M. (2019*b*). A method for reconstructing temporal changes in vegetation functional trait composition using Holocene pollen assemblages. *PLoS ONE* **14**, e0216698

CAZENAVE, A., PALANISAMY, H. & ABLAIN, M. (2018). Contemporary sea level changes from satellite altimetry: what have we learned? What are the new challenges? *Advanced Space Research* **62**, 1639–1653.

CHARMAN, D. (2002). *Peatlands and environmental change*. Wiley, Chichester.

CLARK, J. G. D. & GODWIN, H. (1962). The Neolithic in the Cambridgeshire Fens. *Antiquity* **36**, 10–23.

CLYMO, R. S. (1983). Peat. In *Mires: Swamp, Bog Fen and Moor. Ecosystems of the World 4A* (ed. by A.T.P. Gore), pp. 159–224. Elsevier, Amsterdam.

CLYMO, R.S. (1984). The limits to peat bog growth. *Philosophical Transactions of the Royal Society of London* **B303**, 605–654.

CUSELL, C., KOOIJMAN, A., FERNANDEZ, F., VAN WIRDUM, G., GEURTS, J. J. M., VAN LOON, E. E., KALBITZ, K. & LAMERS, L. P. M. (2014). Filtering fens: mechanisms explaining phosphorus-limited hotspots of biodiversity in wetlands adjacent to heavily fertilized areas. *Science of the Total Environment* **481**, 129–141.

DARBY, H.C. (1983). *The Changing Fenland*. Cambridge University Press, Cambridge.

DBEIS (Department for Business, Energy and Industrial Strategy) (2018). *2017 UK greenhouse gas emissions, provisional figures*. National Statistics, Department for Business, Energy and Industrial Strategy.

DEFORCE, K. (2011). Middle and late Holocene vegetation and landscape evolution of the Scheldt Estuary. A palynological study of a peat deposit from Doel (N-Belgium). *Geologica Belgica* **14**, 277–288.

DEFORCE K. & BASTIAENS J. (2007). The Holocene history of *Taxus baccata* (yew) in Belgium and neighbouring regions. *Belgian Journal of Botany* **140**, 222–237.

DEFORCE, K., STORME, A., BASTIAENS, J. DEBRUYNE, S., DENYS, L., ERVYNCK, A., MEYLEMANS, E., STIEPERAERE, H., VAN NEER, W. & CROMBÉ, P. (2014). Middle-Holocene alluvial forests and associated fluvial environments: a multi-proxy reconstruction. *The Holocene* 24, 1550–1564.

DE JONG, J. D. (1971). The scenery of the Netherlands against the background of Holocene geology: a review of the recent literature. *Revue de Géographie Physique et de Géologie Dynamique* **13**, 143–162.

DE MULDER, E. F. J., GELUK, M. C., RITSEMA, I., WESTERHOFF, W. E. & WONG, TH. E. (2003). *De ondergrond van Nederland. Geologie van Nederland, deel 7*. Utrecht: Nederlands Instituutvoor Toegepaste Geowetenschappen TNO.

DEN HELD, A. J., SCHMITZ, M. & VAN WIRDUM, G. (1992). Types of terrestrializing fen vegetation in the Netherlands. In *Fens and Bogs in the Netherlands: Vegetation, History, Nutrient Dynamics and Conservation Geobotany* (ed. J. T. A. VERHOEVEN), pp. 237–321. Kluwer Academics Publishers, Dordrecht.

DENYS, L. (1999). A diatom and radiocarbon perspective of the palaeoenvironmental history and stratigraphy of Holocene deposits between Oostende and Nieuwpoort (western coastal plain, Belgium). *Geologica Belgica* **2**, 111–140.

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.

DEVOY, R. J. N. (1979). Flandrian sea level changes and vegetational history of the Lower Thames Estuary. *Philosophical Transactions of the Royal Society of London* **B285**, 355–407.

EBBING, J. H. J., WEERTS, H. J. T. & WESTERHOFF, W. E. (2003). Towards an integrated land-sea stratigraphy of The Netherlands. *Quaternary Science Reviews* **22**, 1579–1587.

EDDISON, J. (2000). *Romney Marsh: survival on a frontier*. Tempus, Stroud.

ERKENS, G. (2009). *Sediment Dynamics in the Rhine Catchment: Quantification of Fluvial Response to Climate Change and Human Impact*. The Royal Dutch Geographical Society/ Faculty of Geographical Sciences, Nederlandse Geografische Studies 388, Utrecht University, Utrecht, The Netherlands.

ERKENS, G., VAN DER MEULEN, M. J. & MIDDELKOOP, H. (2016). Double trouble: subsidence and CO2 respiration due to 1,000 years of Dutch coastal peatlands cultivation. *Hydrogeological Journal* **24**, 551–568.

GAUCI, V. (2008). *Carbon balance and offset potential of the Great Fen Project*. The Open University and GLCC.

GEHRELS, W. R. (1999). Middle and late Holocene sea-level changes in eastern Maine reconstructed from foraminiferal saltmarsh stratigraphy and AMS 14C dates on basal peat. *Quaternary Research* **52**, 350–359.

GERRARD, A. J., ADAM, B. H. & MORRIS, L. (1984). Holocene coastal changes: methodological problems. *Quaternary Newsletter* **44**, 7–14.

GILLER, K. E. & WHEELER, B. D. (1988). Acidifiation and succession in a floodplain mire in the Norfolk Broadland, UK. *Journal of Ecology* **76**, 849–866.

GODWIN, H. (1938). The origin of roddons. *Geographical Journal* **91**, 241-250.

GODWIN, H. (1940). Studies of the post-glacial history of British vegetation. III. Fenland pollen diagrams. IV. Post-glacial changes of relative land and sea-level in the English Fenland. *Philosophical Transactions of the Royal Society of London* **B230**, 239–303.

GODWIN, H. (1943). Coastal peat beds of the British Isles and North Sea: presidential address to the British Ecological Society 1943. *Journal of Ecology* **31**, 199–247.

GODWIN, H. (1945). Coastal peat-beds of the North Sea region, as indices of land- and sea-level changes. *New Phytologist* **44**, 29–69.

GODWIN, H. (1968). Terneuzen and buried forests of the East Anglian Fenland. *New Phytologist* **67**, 733–738.

GODWIN, H. (1975). *History of the British Flora*. 2nd edn. Cambridge University Press, Cambridge.

GODWIN, H. (1978). *Fenland: its Ancient Past and Uncertain Future*. Cambridge University Press, Cambridge.

GODWIN, H. & CLIFFORD, M. H. (1938). Studies of the post-glacial history of British vegetation. I. Origin and stratigraphy of Fenland deposits near Woodwalton, Hunts. II. Origin and stratigraphy of deposits in southern Fenland. *Philosophical Transactions of the Royal Society of London* **B229**, 323–406.

GODWIN, H., GODWIN, M. E. & CLIFFORD, M. H. (1935). Controlling factors in the formation of fen deposits, as shown by peat investigations at Wood Fen, near Ely. *Journal of Ecology* **23**, 509–535.

GODWIN, H., SUGGATE, R. P. & WILLIS, E. H. (1958). Radiocarbon dating of the Eustatic rise in ocean-level. *Nature* **181**, 1518–1519.

GODWIN, H. & VISHNU-MITTRE. (1975). Studies of post-glacial history of British vegetation. XVI. Flandrian deposits of the Fenland margin at Holme Fen and Whittlesey Mere, Hunts. *Philosophical Transactions of the Royal Society of London* **B270**, 561–608.

GREEN, R. D. (1968). *Soils of Romney Marsh*. Bulletin of the Soil Survey of England and Wales, Harpenden.

GUPTA, S., COLLIER, J., PALMER-FELGATE, A., DICKINSON, J., BUSHE, K. & HUMBER, S. (2004). *Submerged Palaeo-Arun River: Reconstruction of Prehistoric Landscapes and Evaluation of Archaeological Resource Potential* (Integrated Projects 1 and 2). Final Project Report for English Heritage.

HAGEMAN, B. P. (1969). Development of the western part of the Netherlands during the Holocene. *Geologieen Mijnbouw* **48**, 373–388.

HALL, D. & COLES, J. (1994). *Fenland Survey: an Essay in Landscape and Persistence*. English Heritage, London.

HAMILTON, C. A., KIRBY, J. R., LANE, T. P., PLATER, A. J. & WALLER, M. P. (2019). Sediment supply and barrier dynamics as driving mechanisms of Holocene coastal change for the southern North Sea basin. *Quaternary International* **500**, 147–158.

HASLETT, S. K., DAVIES, P., CURR, R. H. F., DAVIES, C. F. C., KENNINGTON, K., KING, C. P. & MARGETTS, A. J. (1998). Evaluating late-Holocene relative sea-level change in the Somerset Levels, southwest England. *The Holocene* **8**, 197–207.

HEPP, D. A., ROMERO, O. E., MORZ, T. DE POL-HOLZ, R. & HEBBELN, D. (2019). How a river submerges into the sea: a geological record of changing a fluvial to a marine paleoenvironment during early Holocene sea level rise. *Journal of Quaternary Science* **34**, 581–592.

HIJMA, M. P. & COHEN, K. M. (2011). Comment on: mid-Holocene water-level changes in the lower Rhine-Meuse delta (western Netherlands): implications for the reconstruction of relative mean sea-level rise, palaeo river-gradients and coastal evolution by Van de Plassche et al. (2010). *Netherlands Journal of Geosciences* **90**, 51–54.

HIJMA, M. P. & COHEN, K. M. (2019). Holocene sea-level database for the Rhine-Meuse Delta, The Netherlands: implications for the pre-8.2 ka sea-level jump. *Quaternary Science Reviews* **214**, 68–86.

HOLMAN, I. P. & KECHAVARZI, C. (2011). *A revised estimate of peat reserves and loss in the East Anglian Fens*. Commissioned by the RSPB. Cranfield University, Cranfield.

HUGHES, F. M. R., ADAMS, W. M., BUTCHART, S. H. M., FIELD, R. H., PEH, K. S.-H. & WARRINGTON, S. (2016). The challenges of integrating biodiversity and ecosystem services monitoring and evaluation at a landscape-scale wetland restoration project in the UK. *Ecology and Society* **21**, http://dx.doi.org/10.5751/ES-08616-210310

HUTCHINSON, J. N. (1980). The record of peat wastage in the East Anglian Fenlands at Holme Post, 1848-1978 A.D. *Journal of Ecology* **68**, 229–249.

HUTH, V., HOFFMANN, M., BERESWILL, S., POPOVA, Y., ZAK, D. & AUGUSTIN, J. (2018). The climate warming effect of a fen peat meadow with fluctuating water table is reduced by young alder trees. *Mires and Peat* **21**, 1–18.

JELGERSMA, S. (1961). Holocene sea-level changes in the Netherlands. *Mededelingen Geologische Stichting* **7**, 1–101.

JENNINGS, S. C., GREATOREX, C., SMYTH, C. & SPURR, G. (2003). The environmental archaeology of the Late Bronze Age occupation platform at Shinewater, near Eastbourne. In *Alluvial Archaeology in Europe* (eds A. J. HOWARD, M. G. MACKLIN & D. G. PASSMORE) pp 93–110, Swets, Rotterdam.

JENNINGS, S. C. & SMYTH, C. (1987). Coastal sedimentation in East Sussex during the Holocene. *Progress in Oceanography* **18**, 205–241.

JNCC (2011). *Towards an assessment of the state of UK Peatlands*. Joint Nature Conservation Committee Report No. 445. JNCC, Peterborough.

JOHNSON, D. W. (1913). Botanical phenomena and the problem of recent coastal subsidence. *Botanical Gazette* **56**, 449–468.

KHAN, N. S., VANE, C. H., HORTON, B. P., HILLIER, C., RIDING, J. B. & KENDRICK, C. P. (2015). The application of d13C, TOC and C/N geochemistry to reconstruct Holocene relative sea levels and paleoenvironments in the Thames Estuary, UK. *Journal of Quaternary Science* **30**, 417–433.

KIDEN, P. (1995) Holocene sea-level change and crustal movement in the southwestern Netherlands. *Marine Geology* **124**, 21–41.

KIDEN, P., DENYS, L. & JOHNSTON, P. (2002). Late Quaternary sea-level change and isostatic and tectonic land movements along the Belgian–Dutch North Sea coast: geological data and model results. *Journal of Quaternary Science* **17**, 535–546.

KIDEN, P., MAKASKE, B. & VAN DE PLASSCHE, O. (2008). Waarom verschillen de zeespiegelreconstructies voor Nederland? *Grondboor en Hamer* **62**, 54–61.

KILIAN, M. R., VAN DER PLICHT, J. & VAN GEEL, B. (1995). Dating raised bogs: new aspects of AMS 14C wiggle matching, a reservoir effect and climate change. *Quaternary Science Reviews* **14**, 959–966.

KIRBY, J. R., CLARKE, D., SHAW, T. & TOOLE, E. (2010). The Mid–Late Holocene Evolution of Southern Walland Marsh and the Origin of the 'Midley Sand'. In *Romney Marsh: Persistence and change in a coastal lowland* (eds M. P. WALLER, E. EDWARDS & L. BARBER), pp 23–51. Romney Marsh Research Trust, Sevenoaks.

KOERSELMAN, W. & VERHOEVEN, J. T. A. (1995). Eutrophication of fen ecosystems: external and internal nutrient sources and restoration strategies. In *Restoration of Temperate Wetlands* (eds B.D. WHEELER, S. C. SHAW, W. J. FOJT & R. A. ROBERTSON), pp 91–112. Wiley, Chichester.

KOSTERS, E. C., CHMURA, G. L. & BAILEY, A. (1987). Sedimentary and botanical factors influencing peat accumulation in the Mississippi delta. *Journal of the Geological Society of London* **144**, 423–434.

LAMBERT, J. M., JENNINGS, M. A., SMITH, C. T., GREEN, C. & HUTCHINSON, J. N. (1961). *The making of the Broads: A reconsideration of their origin in the light of new evidence*. The Royal Geographical Society, London.

LAMERS, A. P. M., VILE, M. A., GROOTJANS, A. P., ACREMAN, M. C., VAN DIGGELEN, R., EVANS, M.G., RICHARDSON, C. J., ROCHEFORT, L., KOOIJMAN, A. M., ROELOFS, J. G. M. & SMOLDERS, A. J. P. (2015). Ecological restoration of rich fens in Europe and North America: from trial and error to an evidence-based approach. *Biological Reviews of the Cambridge Philosophical Society* **90**, 182–203.

LASCARIS, M. A. & DE KRAKER, A. M. J. (2013). Dikes and other hydraulic engineering works from the late Iron Age and Roman Period on the coastal area between Dunkirk and the Danish Bight. In *Landscapes or seascapes? The history of the coastal area in the North Sea area reconsidered* (eds E. THOEN, G. J. BORGER, A. M. J. DE KRAKER, T. SOENS, D. TYS, L. VERVAET & H. J. T. WEERTS), pp. 177–198. CORN Publication Series 13, Turnhout.

LATAŁOWA, M., ŚWIĘTA-MUSZNICKA, F., SŁOWIŃSKI, M., PĘDZISZEWSKA, A., NORYŚKIEWICZ, A. M., ZIMNY, M., OBREMSKA, M., OTT, F., STIVRINS, N., PASANEN, L., ILVONEN, L., HOLMSTRÖM, L. & SEPPÄ, H. (2019). Abrupt *Alnus* population decline at the end of the first millennium CE in Europe - The event ecology, possible causes and implications. *The Holocene* **29**, 1335–1349.

LAVOREL, S. & GARNIER, E. (2002). Predicting changes in community composition and ecosystem functioning from plant traits: revisiting the holy grail. *Functional Ecology* **16**, 545–556.

LECK, M. A., BALDWIN, A. H., PARKER, V. T., SCHILE, L. & WHIGHAM, D. F. (2009). Plant communities of tidal freshwater wetlands of the continental USA and Canada. In *Tidal freshwater wetlands* (eds A. BARENDREGT, D. F. WHIGHAM, A. H. BALDWIN), pp. 41–51. Backhuys, Leiden.

LEON, P. M., LAMERS, A. J. P., SMOLDER, S. & ROELOFS, J. G. M. (2002). The restoration of fens in the Netherlands. *Hydrobiologia* **478**, 107–130.

LIMPENS, J., BERENDSE, F., BLODAU, C., CANADELL, J. G., FREEMAN, C., HOLDEN, J., ROULET, N., RYDIN, H. & SCHAEPMAN-STRUB, G. (2008). Peatlands and the carbon cycle: from local processes to global implications - a synthesis. *Biogeosciences* **5**, 1475–1491.

LONG, A. J. & INNES, J. B. (1995). The back-barrier and barrier depositional history of Romney Marsh, Walland Marsh and Dungeness, Kent, England. *Journal of Quaternary Science* **10**, 267–283.

LONG, A. J., WALLER, M. P. & PLATER, A. J. (2006*a*). Coastal resilience and late Holocene tidal inlet history: the evolution of Dungeness Foreland and the Romney Marsh (UK) depositional complex. *Geomorphology* **82**, 309–330.

LONG, A. J., WALLER, M. P. & PLATER, A. J. (2007). *Dungeness and the Romney Marsh: Barrier dynamics and marshland evolution.* Oxbow Books, Oxford.

LONG, A.J., WALLER, M. P. & STUPPLES, P. (2006*b*). Driving mechanisms of coastal change: sediment autocompaction and the destruction of late Holocene coastal wetlands. *Marine Geology* **225**, 63–84.

MEIJLES, E. W., KIDEN, P., STREURMAN, H-J., VAN DER PLICHT, J., VOS, P. C., GEHRELS, R. & KOPP, R. E. (2018). Holocene relative mean sea-level changes in the Wadden Sea area, northern Netherlands. *Journal of Quaternary Science* **33**, 905–923.

MILNE, G. A., LONG, A. J. & BASSETT, S. E. (2005). Modelling Holocene relative sea-level observations from the Caribbean and South America. *Quaternary Science Reviews* **24**, 1183–1202.

MOOR, H., RYDIN, H., HYLANDER, K., NILSSON, M. B., LINDBORG, R. & NORBERG, J. (2017). Towards a trait-based ecology of wetland vegetation. *Journal of Ecology* **105**, 1623–1635.

MOUNTFORD, J. O., MCCARTNEY, M. P., MANCHESTER, S. J. & WADSWORTH, R. A. (2002). *Wildlife Habitats and their Requirements within the Great Fen Project*. *Final report.* NERC/Centre for Ecology & Hydrology.

NATIONAL TRUST (2009). *Wicken Fen Vision: a National Trust strategy to create a 53 square kilometre nature reserve for wildlife and people in Cambridgeshire*. National Trust, Wicken, Cambridgeshire.

NATURAL ENGLAND (2010). *England's peatlands: carbon storage and greenhouse gases* (NE257). Report by Natural England. http://publications.naturalengland.org.uk/publication/

30021

NEUBAUER, S.C. (2008). Contributions of mineral and organic components to tidal freshwater marsh accretion. *Estuarine, Coastal and Shelf Science* **78**, 78–88.

PEACOCK, M., GAUCI, V., BAIRD, A. J., BURDEN, A., CHAPMAN, P. J., CUMMING, A., EVANS, J. G., GRAYSON, R .P., HOLDEN, J., KADUK, J., MORRISON, R., PAGE, S., PAN, G., RIDLEY, L. M., WILLIAMSON, J., *et al*. (2019). The full carbon balance of a rewetted cropland fen and a conservation managed fen. *Agriculture, Ecosystems and Environment* **269**, 1–12.

PEH, K., S-H., BALMFORD, A., FIELD, R. H., LAMB, A., BIRCH, J. C., BRADBURY, R. B., BROWN, C., BUTCHART, S. H. M., LESTER, M., MORRISON, R., SEDGWICK, I., SOANS, C., STATTERSFIELD, A. J., STROH, P. A., SWETNAM, R. D., *et al*. (2014). Benefits and costs of ecological restoration: rapid assessment of changing ecosystem service values at a U.K. wetland. *Ecology and Evolution* **4**, 3875–3886.

PIERIK, H. J., COHEN, K. M., VOS, P. C., VAN DER SPEK, A. J. F. & STOUTHAMER, E. (2017). Late Holocene coastal-plain evolution of the Netherlands: the role of natural preconditions in human-induced sea ingressions. *Proceedings of the Geologists' Association* **128**, 180–197.

PONS, L. (1992). Holocene peat formation in the lower parts of the Netherlands. In *Fens and Bogs in the Netherlands: Vegetation, History, Nutrient Dynamics and Conservation Geobotany* (ed. J. T. A. VERHOEVEN), pp. 7-79. Kluwer Academics Publishers, Dordrecht.

PORTER, M. & FOLEY, M. (2017). *Violas of Britain and Ireland*. BSBI Handbook No. 17. Botanical Society of the Britain and Ireland, Bristol.

PRATOLONGO, P., LEONARDI, N., KIRBY, J. R. & PLATER, A. J. (2018). Temperate coastal wetlands: morphology, sediment processes and plant communities. In *Coastal Wetlands, an integrated ecosystem approach* (eds G. M. E. PERILLO, E. WOLANSKI, & D. CAHOON), pp. 105–152. Elsevier, Amsterdam.

REID, C. (1913). *Submerged Forests.* Cambridge University Press, Cambridge.

REITALU T., GERHOLD P., POSKA A., PÄRTEL M., VÄLI V. & VESKI S. (2015). Novel insights into post-glacial vegetation change: functional and phylogenetic diversity in pollen records. *Journal of Vegetation Science* **26**, 911–922.

RIPPON, S. (2000). *The transformation of coastal wetlands*. Oxford University Press, Oxford.

RODWELL, J. S. (1991*a*). *British plant communities: Volume 1, Woodlands and scrub*. Cambridge University Press, Cambridge.

RODWELL, J. S. (1991*b*). *British Plant Communities: Volume 2, Mires and Heath*. Cambridge University Press, Cambridge.

RODWELL, J. S. (1995). *British Plant Communities: Volume 5, Aquatic Communities, Swamps and Tall-herb Fens*. Cambridge University Press, Cambridge

ROELEVELD, W. (1974). The Holocene evolution of the Groningen marine-clay district - Berichten van de Rijksdienstvoor het Oudheidk. *Bodemonderzoek* **24**, 1–132.

RYDIN, H. & JEGLUM, J. (2006). *The Biology of Peatlands*. Oxford University Press, Oxford.

SCHÄFER, A. & JOOSTEN, H. (2005). Erlenaufforstung auf wiedervernässten Niedermooren. Institut für dauerhaft umwelt-gerechte Entwicklung von Naturräumen der Erde, Greifswald.

SCHUERCH, M., SPENCER, T., TEMMERMAN, S., KIRWAN, M. L., WOLFF, C. LINCKE, D., MCOWEN, C. J., PICKERING, M. D., REEF, R., VAFEIDIS, A. T., HINKEL, J., NICHOLLS, R. J. & BROWN, S. (2018). Future response of global coastal wetlands to sea-level rise. *Nature* **561**, 231–234.

SHENNAN, I. (1986*a*). Flandrian sea-level changes in the Fenland. I: the geographical setting and evidence of relative sea-level changes. *Journal of Quaternary Science* **1**, 119–154.

SHENNAN, I. (1986*b*). 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.

SHENNAN, I., BRADLEY, S. L. & EDWARDS, R. (2018). Relative sea-level changes and crustal movements in Britain and Ireland since the Last Glacial Maximum. *Quaternary Science Reviews* **188**, 143–159.

SHENNAN, I. & HORTON, B. (2002). Holocene land- and sea-level changes in Great Britain. *Journal of Quaternary Science* **17**, 511–526.

SHENNAN, I., LAMBECK, K., FLATHER, R., HORTON, B., MCARTHUR, J., INNES, J., LLOYD, J., RUTHERFORD, M. & WINGFIELD, R. (2000). Modelling western North Sea palaeogeographies and tidal changes during the Holocene. *Geological Society of London Special Publication* **166**, 298–319.

SHENNAN, I., LONG, A. J. & HORTON, B. P. (2015). Handbook of sea-level research: framing research questions. In *Handbook of Sea-level Research* (eds I. SHENNAN, A. J. LONG & B. P. HORTON), pp. 1–25. Wiley, Chichester.

SHENNAN, I., MILNE, G. & BRADLEY, S. (2012). Late Holocene vertical land motion and relative sea-level changes: lessons from the British Isles. *Journal of Quaternary Science* **27**, 64–70.

SMITH, D. M., ZALASIEWICZ, J. A., WILLIAMS, M., REDDING, M. & BEGG, C. (2010). Holocene drainage systems of the English Fenland: roddons and their environmental significance. *Proceedings of the Geologists Association* **121**, 256–269.

SMYTH, C. & JENNINGS, S. (1988). Mid- to late-Holocene forest composition and the effects of clearances in the Combe Haven valley, East Sussex. *Sussex Archaeological Collections* **126**, 1–20.

STREIF, H. & ZIMMERMANN, B. (1973). Das Küstenholozän von Rysum/Knock imGebied der Emsmündung (Nordsee). G*eologischesJahrbuch* **A9**, 3–20.

SWINNERTON, H. H. (1931). The post-glacial deposits of the Lincolnshire coast. *Quarterly Journal of the Geological Society of London* **87**, 360–375.

SZUMIGALSKI, A. R. & BAYLEY, S. E. (1996). Net above-ground production along a bog-rich fen gradient in central Alberta, Canada. *Wetlands* **16**, 467–476.

THORMANN, M. N. & BAYLEY, S. E. (1997). Above-ground plant production and nutrient content of the vegetation in six peatlands in Alberta, Canada. *Plant Ecology* **131**, 1–16.

TOOLEY, M. J. (1986). Sea Levels. *Progress in Physical Geography* **10**, 120–129.

TRÖELS-SMITH, J. (1955). Characterisation of unconsolidated sediments. *Danmarks Geologiske Undersogelse* Series IV, 38–73.

TUITTILA, E-S., JUUTINEN, S., FROLKING, S., VÄLIRANTA, M., LAINE, A. M., MIETTINEN, A., SEVÄKIVI, M-L., QUILLET, A. & MERILÄ, P. (2013). Wetland chronosequence as a model of peatland development: Vegetation succession, peat and carbon accumulation. *The Holocene* **23**, 25–35.

TYS, D. (2013). The medieval embankment of coastal Flanders. In *Landscapes or seascapes? The history of the coastal area in the North Sea area reconsidered* (eds E. THOEN, G. J. BORGER, A. M. J. DE KRAKER, T. SOENS, D. TYS, L. VERVAET & H. J. T. WEERTS), pp. 199–239. CORN Publication Series 13, Turnhout.

VAN ASSELEN, S., COHEN, E. & STOUTHAMER, E. (2017). The impact of avulsion on groundwater level and peat formation in delta floodbasins during the middle-Holocene transgression in the Rhine-Meuse delta, The Netherlands. *The Holocene* **27**, 1694–1706.

VAN ASSELEN, S., STOUTHAMER, E. & VAN ASCH, W. J. (2009). Effects of peat compaction on delta evolution: a review on processes, responses, measuring and modelling. *Earth-Science Reviews* **92**, 35–51.

VAN DE PLASSCHE, O. (1982). Sea-level change and water-level movements in the Netherlands during the Holocene. *Mededelingen Rijks Geologische Dienst* **36**, 1–93.

VAN DER MOLEN, J. & DE SWART H. E. (2001). Holocene tidal conditions and tide-induced sand transport in the southern North Sea. *Journal of Geophysical Research* **106**, 9339–9362.

VAN DER WOUDE, J. D. (1984). The fluviolagoonal palaeoenvironment in the Rhine/Meuse deltaic plain. *Sedimentology* **31**, 395–400.

VAN DIGGLEN, R., BENSE, I. H. M., BROUWER, E., LIMPENS, J., VAN SCHIE, J. M., SMOLDERS, A. J. P., LEON P. M. & LAMERS, A. P. M. (2015). Restoration of acidified and eutrophied rich fens: Long-term effects of traditional management and experimental liming. *Ecological Engineering* **75**, 208–216.

VAN DIGGLEN, R., MOLENAAR, W. J. & KOOIJMAN, R. H. (1996). Vegetation succession in a floating mire in relation to management and hydrology. *Journal of Vegetation Science* **7**, 809–820.

VAN LOON, A. J. (1981). Problems of Holocene lithostratigraphy. *Geologie en Mijnbouw* **60**, 353–361.

VAN WIRDUM, G., DEN HELD, A. J. & SCHMITZ, M. (1992). Terrestrializing fen vegetation in former turbaries in the Netherlands. In *Fens and Bogs in the Netherlands: Vegetation, History, Nutrient Dynamics and Conservation* (ed. J. T. A. VERHOEVEN), pp. 323–360. Kluwer Academic Publishers, Dordrecht.

VERMEER, J. G. & JOOSTEN, J. H. J. (1992). Conservation and management of bog and fen reserves in the Netherlands. In *Fens and Bogs in the Netherlands: Vegetation, History, Nutrient Dynamics and Conservation* (ed. J. T. A. VERHOEVEN), pp. 433–478. Kluwer Academic Publishers, Dordrecht.

VINK, A., STEFFEN, H., REINHARDT, L. & KAUFMANN, G. (2007). Holocene relative sea level change, isostatic subsidence and the radial viscosity structure of the mantle of northwest Europe (Belgium, the Netherlands, Germany, southern North Sea). *Quaternary Science Reviews* **26**, 3249–3275.

VIS, G-J., COHEN, K. M., WESTERHOFF, W. E., TEN VEEN, J. H., HIJMA, M. P., VAN DER SPEK, A. J. F. & VOS, P. C. (2015). Paleogeography. In *Handbook of Sea-level Research* (eds I. SHENNAN, A.J. LONG & B. P. HORTON), pp. 514–535. Wiley, Chichester.

VOS, P.C. (2015). *Origin of the Dutch Coastal Landscape– Long-Term Landscape Evolution of the Netherlands during the Holocene, Described and Visualized in National, Regional and Local Map Series*. Barkhuis: Groningen.

VOS, P. C. & VAN HEERINGEN, R. M. (1997). Holocene geology and occupation history of the Province of Zeeland. *Mededelingen Rijks Geologische Dienst* **59**, 5–110.

WALKER, M. J. C., BERKELHAMMER, M., BJÖRCK, S., CWYNAR, L. C., FISHER, D. A., LONG, A. J., LOWE, J. J., NEWNHAM, R. M., RASMUSSEN, S. O. & WEISS, H. (2012). Formal subdivision of the Holocene Series/Epoch: a discussion paper by a working group of INTIMATE (Integration of ice-core, marine and terrestrial records) and the Subcommission on Quaternary Stratigraphy (International Commission on Stratigraphy). *Journal of Quaternary Science* **27**, 649–659.

WALLER, M. P. (1993). Flandrian vegetational history of south-eastern England. Pollen data from Pannel Bridge, East Sussex. *New Phytologist* **124**, 345–369.

WALLER, M. P. (1994*a*). Flandrian vegetational history of south-eastern England. Stratigraphy of the Brede valley and pollen data from Brede Bridge. *New Phytologist* **126**, 369–392.

WALLER, M. P. (1994*b*) *The Fenland Project, Number 9: Flandrian environmental change in Fenland*. East Anglian Archaeology Monograph No. 70.

WALLER, M. P., BURRIN, P. J. & MARLOW, A. D. (1988). Flandrian sedimentation and palaeoenvironments in Pett Level, the Brede and Lower Rother valleys and Walland Marsh. In *Romney Marsh Evolution, Occupation, Reclamation* (eds J. EDDISON & C. P. GREEN), pp. 3-29. Oxford University Committee for Archaeology, Monograph No. 24.

WALLER, M. P. & EARLY R. (2015). Vegetation dynamics from a coastal peatland: insights from combined plant macrofossil and pollen data. *Journal of Quaternary Science* **30**, 779–789.

WALLER, M. P. & GRANT, M. J. (2012). Holocene pollen assemblages from coastal wetlands: differentiating natural and anthropogenic causes of change in the Thames estuary, UK. *Journal of Quaternary Science* **27**, 461–474.

WALLER, M. P. & HAMILTON, S. (2000). The vegetation history of the English chalkland: a mid-Holocene pollen sequence from the Caburn, East Sussex. *Journal of Quaternary Science* **15**, 253–272.

WALLER, M. P. & KIRBY, J. (2002). Late Pleistocene/Early Holocene Environmental Change in the Romney Marsh region: New evidence from Tilling Green, Rye. In *Romney Marsh: Coastal and Landscape Change through the Ages*, (eds A. J. LONG, S. HIPKIN & H. CLARKE), pp 22–39. Oxford University School of Archaeology Monograph 56.

WALLER, M. P. & LONG, A. J. (2010). The Holocene coastal deposits of Sussex: a re-evaluation. In *Romney Marsh: Persistence and change in a Coastal Lowland* (eds M. P. WALLER, E. EDWARDS & L. BARBER), pp. 1–21. Romney Marsh Research Trust, Sevenoaks.

WALLER, M. P., LONG, A. J., LONG, D. & INNES, J. B. (1999). Patterns and processes in the development of coastal mire vegetation: multi-site investigations from Walland Marsh, Southeast England. *Quaternary Science Reviews* **18**, 1419–1444.

WALLER, M. P., LONG, A. J. & SCHOFIELD, J. E. (2006). The interpretation of radiocarbon dates from the upper surface of late Holocene peat layers in coastal lowlands. *The Holocene* **16**, 51–61.

WEERTS, H. J. T., WESTERHOFF, W. E., CLEVERINGA, P., BIERKENS, M. F.P., VELDKAMP, J. G. & RIJSDIJK, K. F. (2005). Quaternary geological mapping of the lowlands of The Netherlands, a 21st century perspective. *Quaternary International* **133/134**, 159–178.

WHEELER, A. J. & WALLER, M. P. (1995). The Holocene lithostratigraphy of Fenland, eastern England: a review and suggestions for redefinition. *Geological Magazine* **132**, 223–233.

WHEELER, B. D. (1980*a*). Plant communities of rich-fen systems in England and Wales: I. Introduction. Tall sedge and reed communities. *Journal of Ecology* **68**, 365–395.

WHEELER, B. D. (1980*b*). Plant communities of rich-fen systems in England and Wales: III. Fen meadow, fen grassland and fen woodland communities, and contact communities. *Journal of Ecology* **68**, 761–788.

WHEELER, B. D. (1988). Species richness, species rarity and conservation evaluation of rich-fen vegetation in lowland England and Wales. *Journal of Applied Ecology* **25**, 331–352.

WHEELER, B. D. & PROCTOR, M. C. F. (2000). Ecological gradients, subdivisions and terminology of north-west European mires. *Journal of Ecology* **88**, 187–203.

WIEGERS, J. (1992). Carr vegetation: communities and succession of dominant tree species. In *Fens and Bogs in the Netherlands: Vegetation, History, Nutrient Dynamics and Conservation Geobotany* (ed. J.T.A. VERHOEVEN), pp. 361–395. Kluwer Academics Publishers, Dordrecht.

WITTE, H. J. L. & VAN GEEL, B. (1985). Vegetational and environmental succession and net organic production between 4500 and 800 B.P. reconstructed from a peat deposit in the western Dutch coastal area (Assendelver Polder). *Review of Palaeobotany and Palynology* **45**, 239–300.

ZONNEVELD, I. S. (1960). *De Brabantse Biesbosch: een studie van bodem en vegetatie van een zoetwatergetijdendelta bestaande uit de drie gedeelten A, B en C.* Stichting voor Bodemkartering, Wageningen.

Figures

Fig. 1. Map of the southern North Sea and eastern English Channel. The dotted line shows the approximate position of the coastline *c*. 9000 cal. yrs BP from Shennan *et al*. (2000).

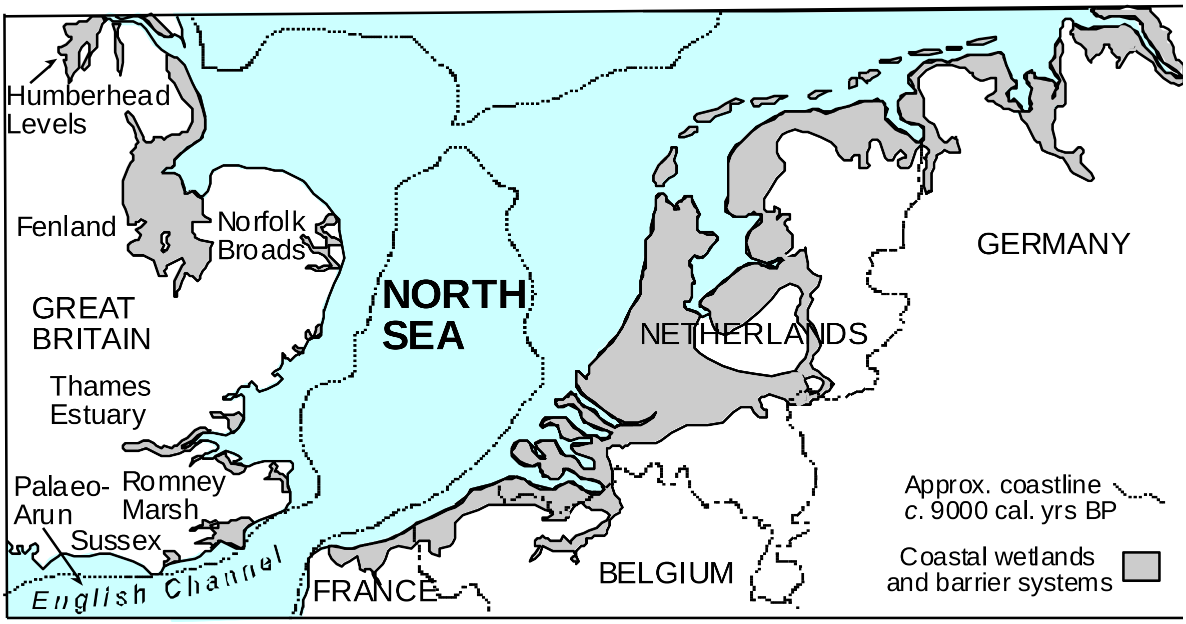


Fig. 2. Schematic representation of the stratigraphic architecture of the coastal peat deposits of the southern North Sea region.

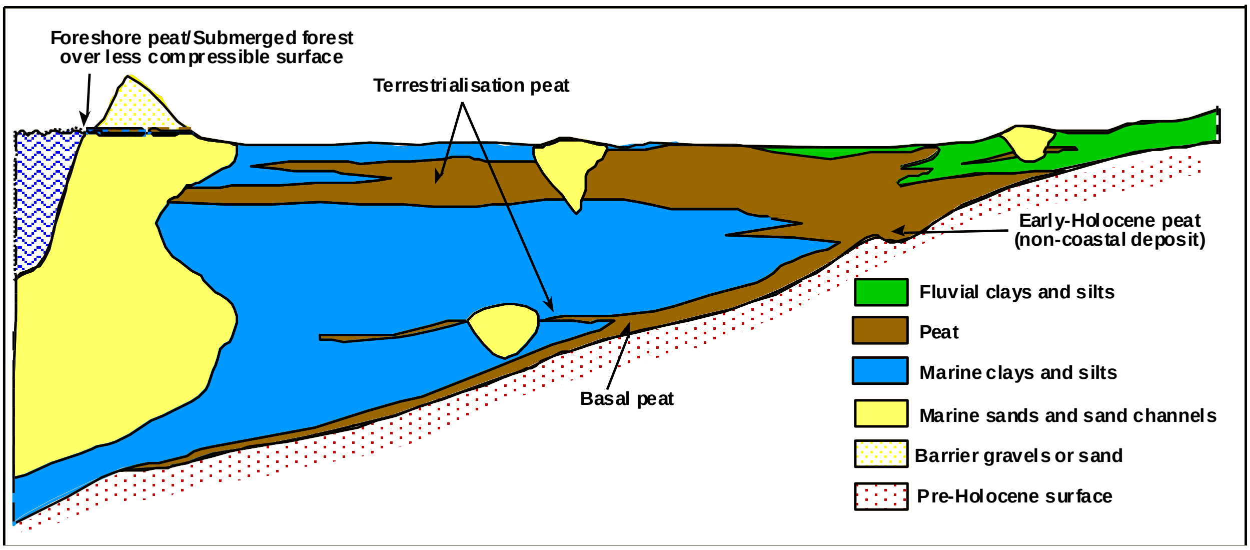


Fig. 3. The position of modern vegetation analogous to Holocene coastal communities in relation to variations in pH and substratum fertility (modified from Wheeler & Proctor, 2000). See Wheeler & Proctor (2000) for an explanation of the measure of fertility. (A) Main phytosociological alliances of mires. (B) Position of the main categories of mire vegetation assumed to be self-sustaining. (C) Position of herbaceous vegetation types and some important herbaceous taxa.

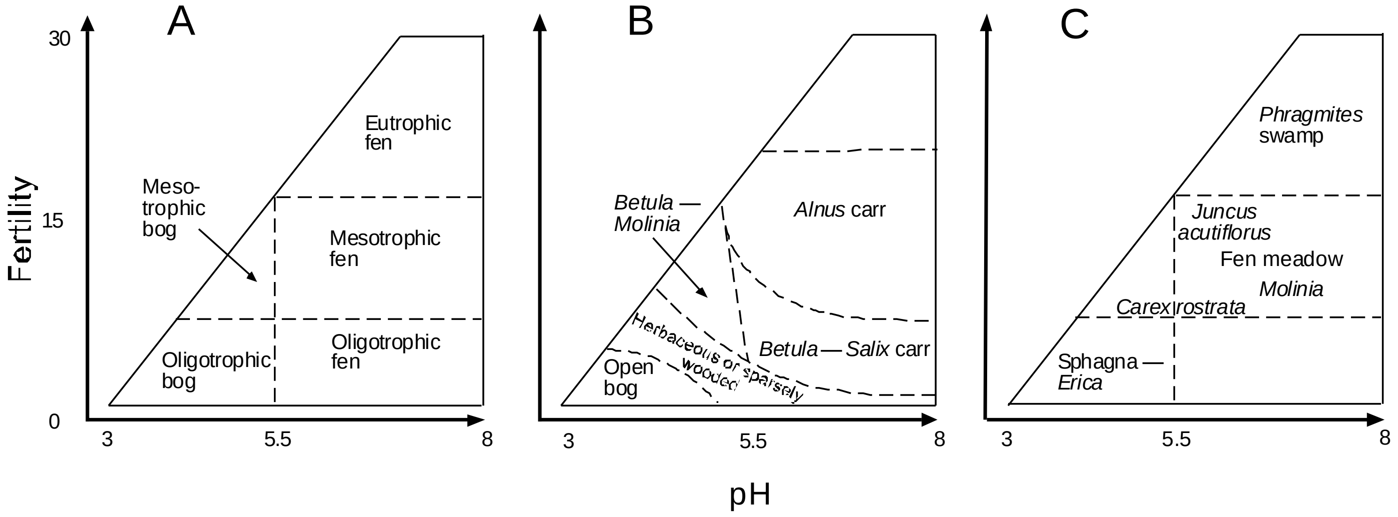


Fig. 4. Relative sea-level curves from the eastern North Sea, showing differential north–south subsidence across the southern North Sea area, with the eustatic element distinguished. NN refers to the German ordnance datum (Normalnull). Modified from Vink *et al*. (2007).

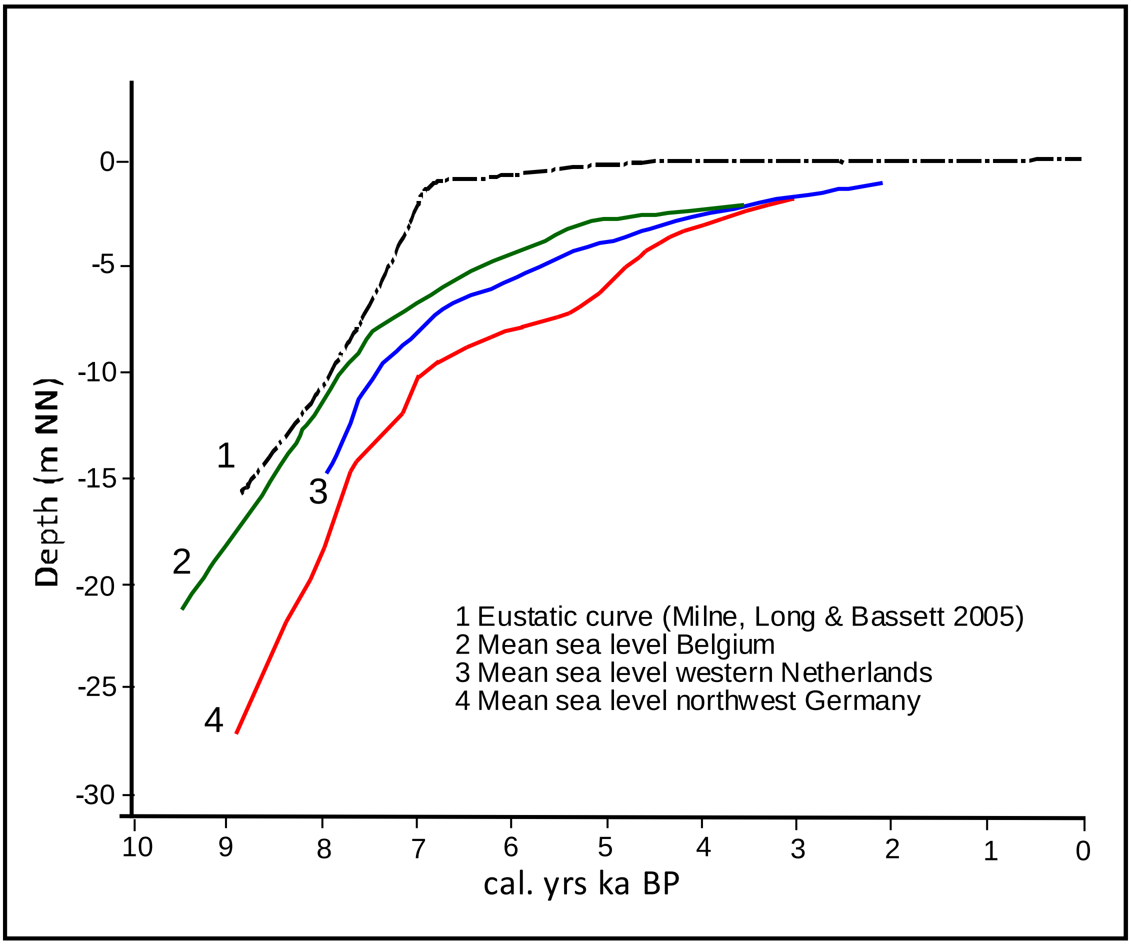


Fig. 5. Examples of retrogressive (communities indicative of successively higher water levels relative to the sediment surface) and progressive (communities indicative of successively lower water levels relative to the sediment surface) series from Fenland. (A) Lithology, organic content (determined by loss-on-ignition) and select pollen data showing a retrogressive series from a basal peat at Lade Bank. (B) Lithology, organic content (determined by loss-on-ignition) and select pollen data showing a progressive series from a terrestrialisation peat at Wiggenhall. Data from Waller (1994*b*) and M. Waller (unpublished data). OD refers to the UK ordnance datum. See Fig. 9 for site locations.

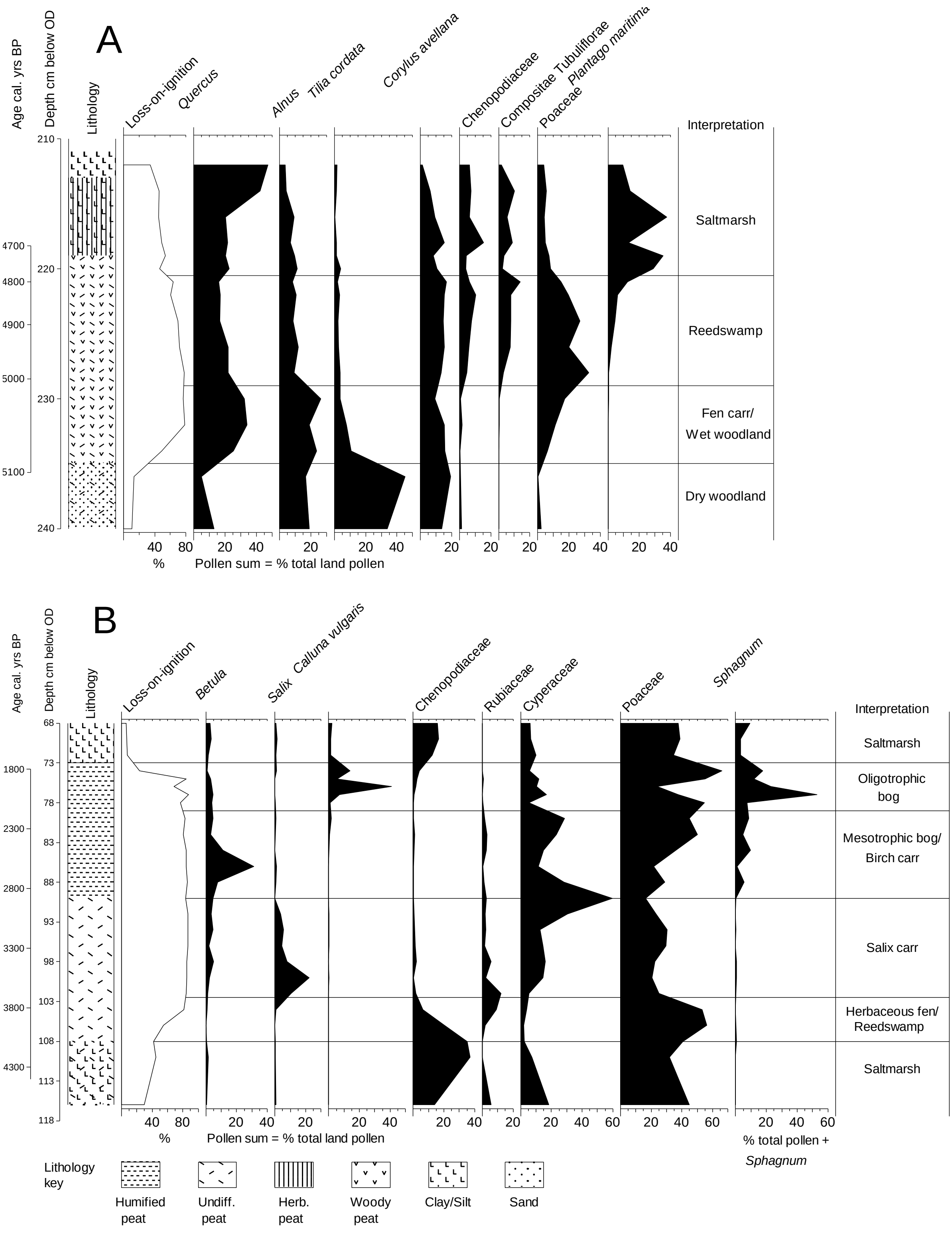


Fig. 6. Organic content (determined by loss-on-ignition) from the main peat bed of the Romney Marsh region (excluding transitional layers and thin clay lenses). Valley sites are represented by open symbols, sites from Walland Marsh by filled symbols, see Fig. 8 for site locations. Data from Waller (1994*a*) and M. Waller (unpublished data).

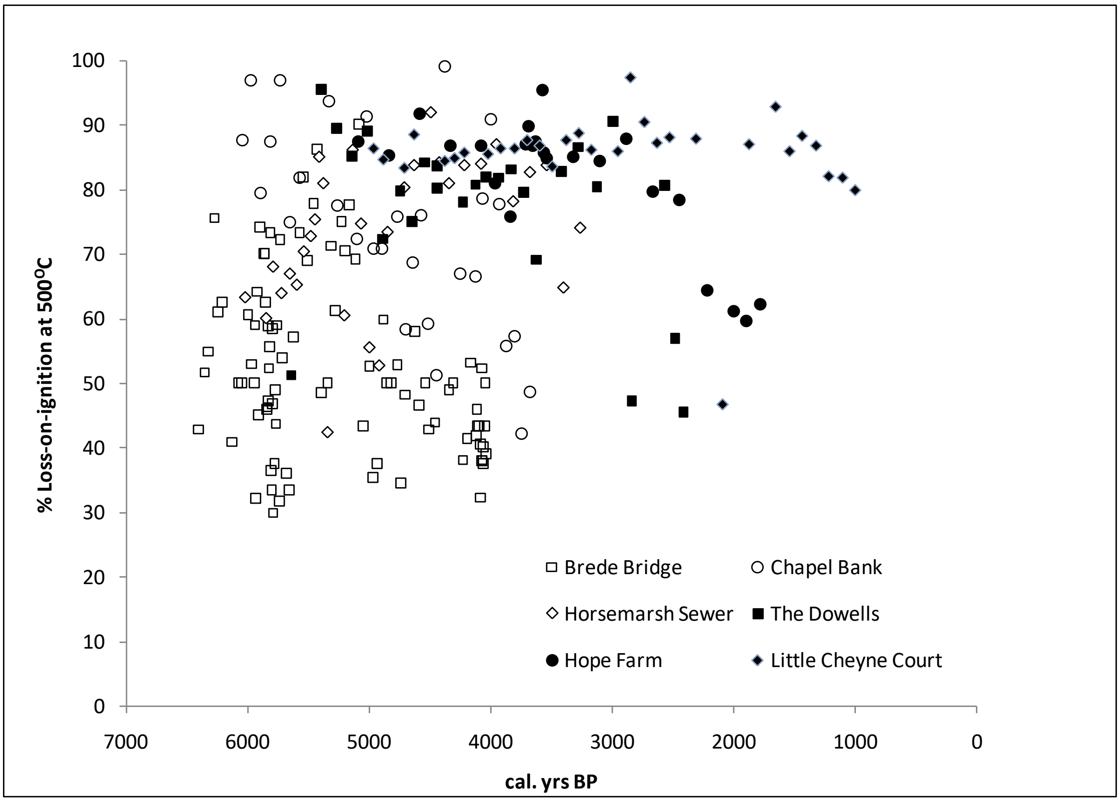
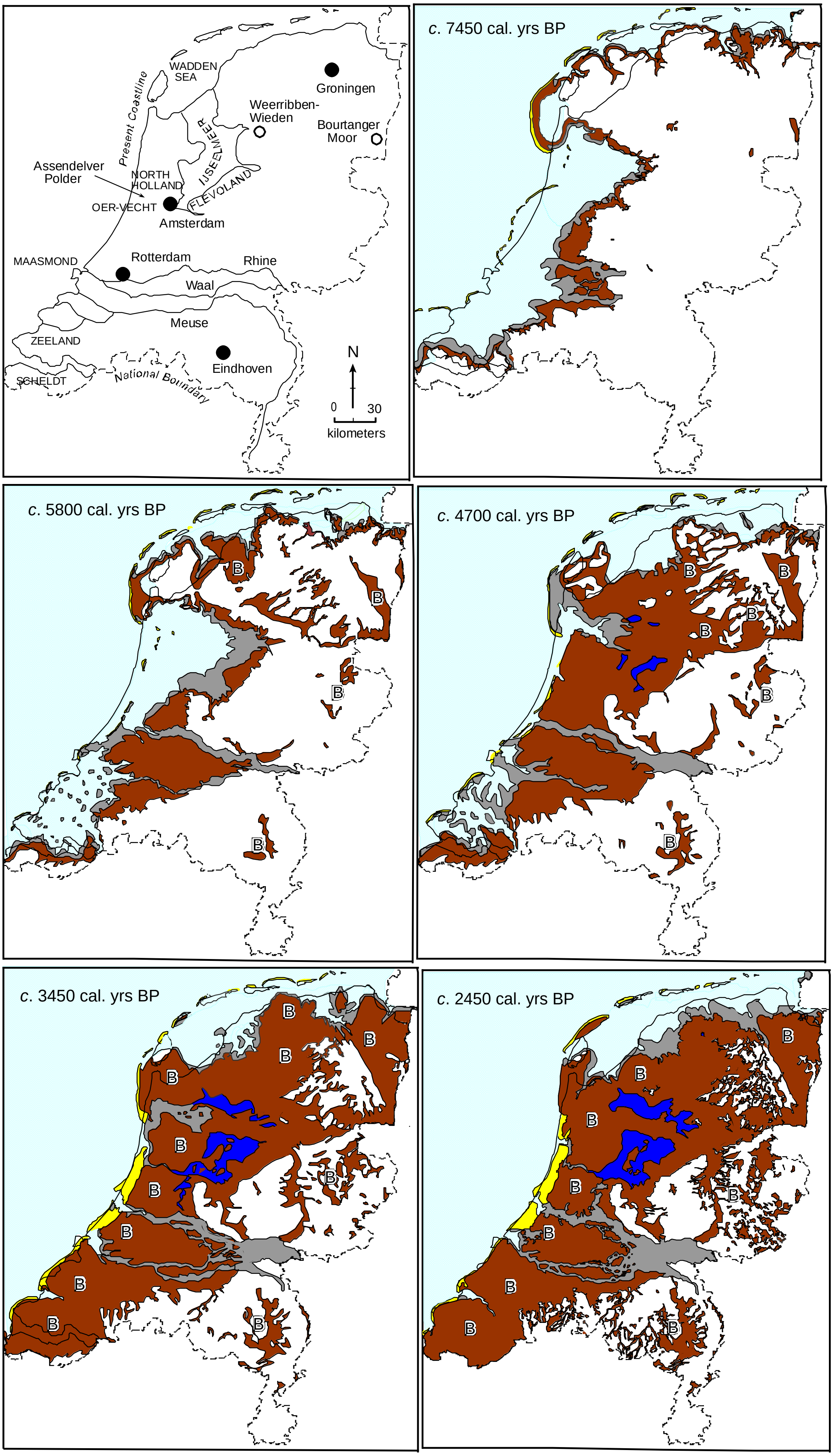


Fig. 7. Location of sites and palaeographic maps from the Netherlands (after Vos, 2015). See Fig. 9 for key.



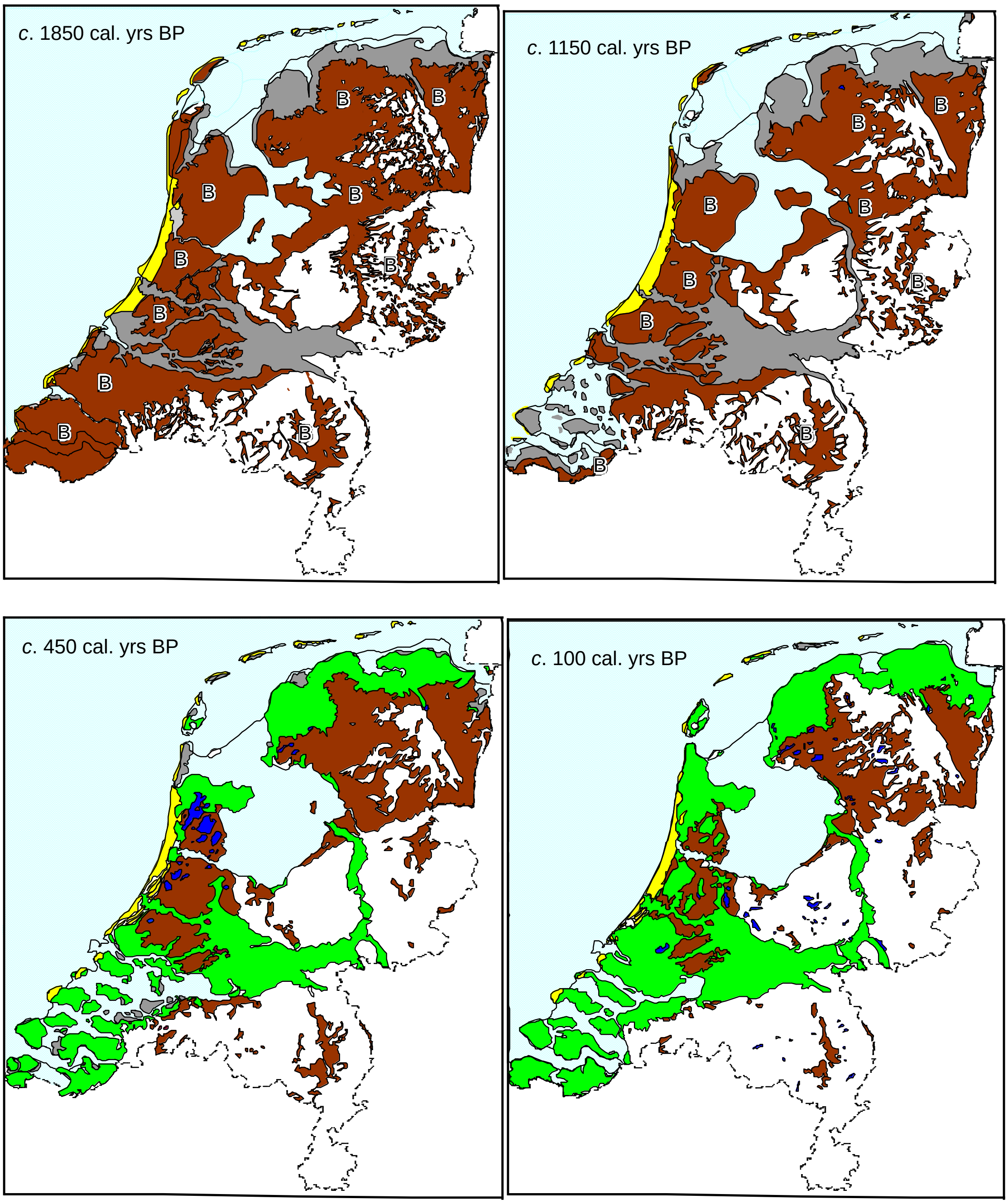

c. 1850 cal. yrs BP
c. 1150 cal. yrs BP
c. 450 cal. yrs BP
c. 100 cal. yrs BP

Fig. 8. Location of sites and palaeogeographic maps from the Romney Marsh area (after Long *et al*., 2007). See Fig. 9 for key.

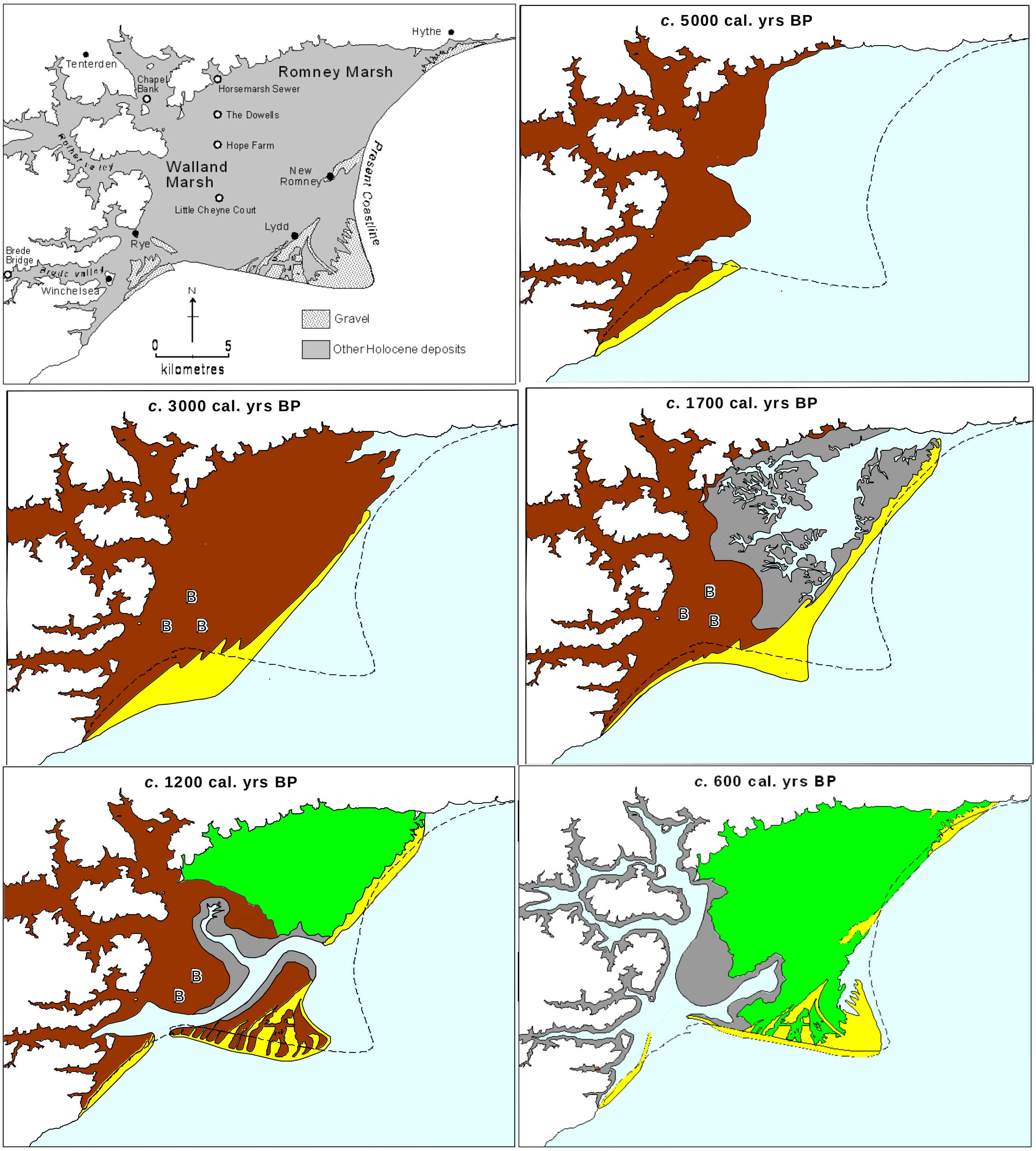


Fig. 9. Location of sites and palaeogeographic maps for the Fenland area (after Waller, 1994*b*).

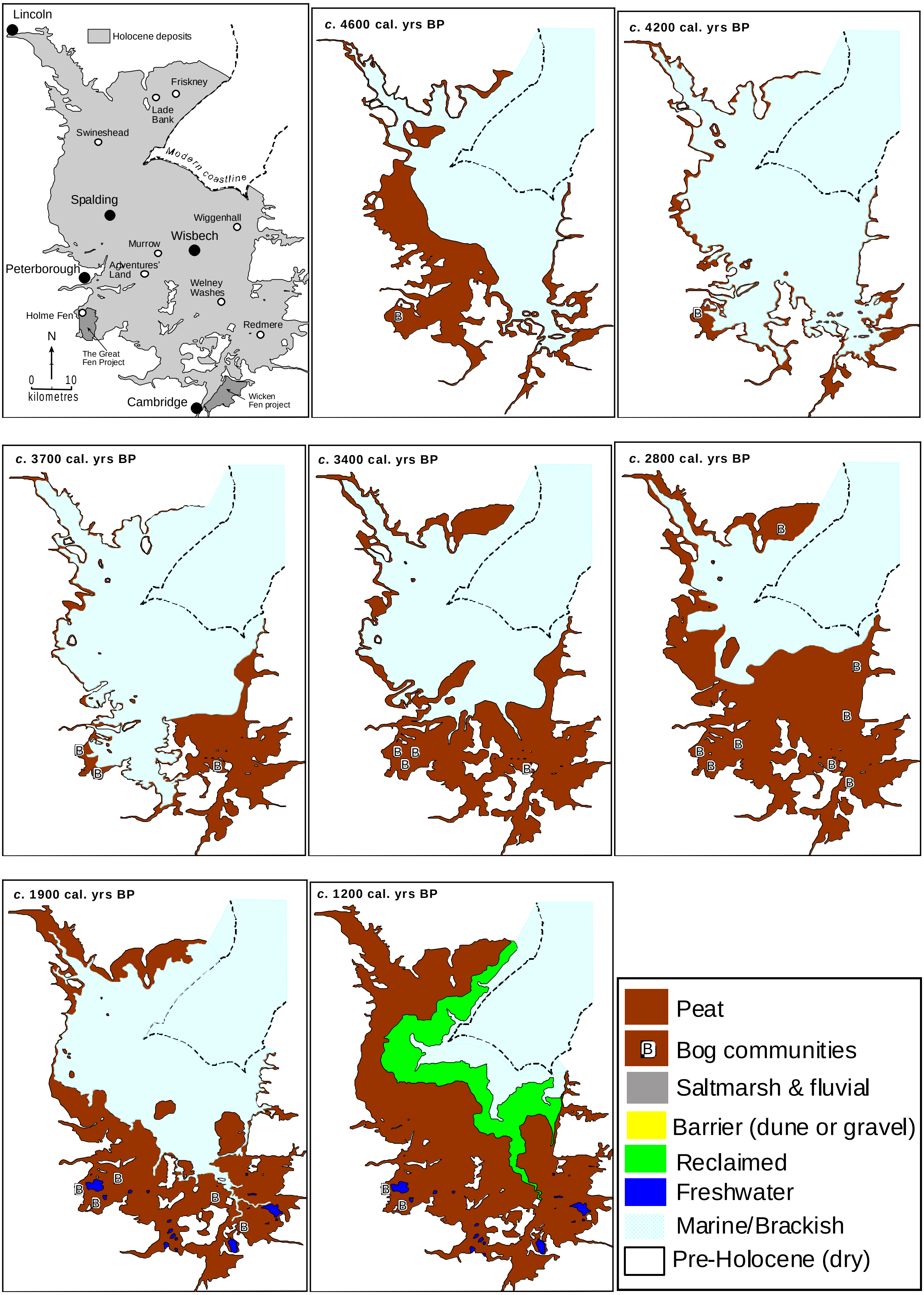


Table 1. The height relationships and properties of coastal peat types. Reference water level refers to the modern equivalent elevation at which the vegetation type or transition occurs, defined with reference to a tide level; mean high water of spring tides (MHWST), mean tide level (MTL), highest astronomical tide (HAT), mean high water (MHW). TOC, total organic carbon based on loss-on-ignition; NA, not applicable.

Vegetation/ peat type	**Reference water-level (m) MHWST Shennan *et al*. (2018)**	**Reference water level (m) MHW [1]Zonneveld (1960) [2]Behre (2007) [3]Hijma & Cohen (2019)**	**Sediment surface elevation (m) below & above water-level De Held *et al*. (1992)**	**Salinity (mg Cl/l) De Held *et al*. (1992)**	**TOC (%) Bos *et al*. (2012) Basal peat**	**TOC (%) Khan *et al*. (2015) Modern sites**	**C/N Khan *et al*. (2015) Modern sites**	**t C ha^{-1} yr^{-1} Barthelmes (2009)**
Swamp with *Phragmites*	*Phragmites* peat directly above fen wood peat –0.1 ± 0.2 *Phragmites* peat directly below tidal-marsh deposit –0.2 ± 0.2	[1]–0.6 to 0 [2]–0.5 to *c*. 0	–1.5 to 0	<2000	35–75	37.9 ± 9.8	13.9 ± 1.2	0.75 Netherlands
Sedge-dominated fen		[1]–0.8 to 0.20 (includes emergents) [2]*c*. 0	–0.3 to 0	<1000	72–96			0.24–0.38 NE Germany
Fen carr	Woody peat that does not show a direct relationship to a contemporaneous tide MTL+HAT/2 (range MTL to HAT)	[1]–0.35 to 0.25 [3]0 ± 0.1			30–83	46.3 ± 1	18.2 ± 3.8	0.34–1.64 NE Germany
Mesotrophic bog	NA	NA	0 to +0.2	<800				
Ombrotrophic bog	NA	NA	+0.05 to +0.3	<100	69–91			0.14–0.72 S. Sweden