Coastal resilience and late Holocene tidal inlet history: the evolution of Dungeness Foreland and the Romney Marsh depositional complex (U.K.)

A.J. Long[a,]*, M.P. Waller[b], A.J. Plater[c]

[a]Department of Geography, Durham University, Science Site, South Road, Durham
DH1 3LE, U.K.

[b]Centre for Earth and Environmental Science Research, School of Earth Sciences and
Geography, Kingston, University, Penrhyn Road, Kingston upon Thames, Surrey
KT1 2EE, U.K.

[c.]Department of Geography, Roxby Building, University of Liverpool, PO Box 147,
Liverpool L69 3BX, U.K.

Abstract

Dungeness Foreland is a large sand and gravel barrier located in the eastern English Channel that during the last 5000 years has demonstrated remarkable geomorphological resilience in accommodating changes in relative sea-level, storm magnitude and frequency, variations in sediment supply as well as significant changes in back-barrier sedimentation. In this paper we develop a new palaeogeographic model for this depositional complex using a large dataset of recently acquired litho-, bio- and chrono-stratigraphic data. Our analysis shows how, over the last 2000 years, three large tidal inlets have influenced the pattern of back-barrier inundation and sedimentation, and controlled the stability and evolution of the barrier by determining the location of cross-shore sediment and water exchange, thereby moderating sediment supply and its distribution. The sheer size of the foreland has contributed in part to its resilience, with an abundant supply of sediment always available for ready redistribution. A second reason for the landform's resilience is the repeated ability of the tidal inlets to narrow and then close, effectively healing successive breaches by back-barrier sedimentation and ebb- and/or flood-tidal delta development. Humans emerge as key agents of change, especially through the process of reclamation which from the Saxon period onwards has modified the back-barrier tidal prism and promoted repeated episodes of fine-grained sedimentation and channel/inlet infill and closure. Our palaeogeographic reconstructions show that large barriers such as Dungeness Foreland can survive repeated “catastrophic” breaches, especially where tidal inlets are able to assist the recovery process by raising the elevation of the back-barrier area by intertidal sedimentation. This research leads us to reflect on the concept of “coastal resilience” which, we conclude, means little without a clearly defined spatial and temporal framework. At a macro-scale, the structure as a whole entered a phase of recycling and rapid progradation in response to changing sediment budget and coastal dynamics about 2000 years ago. However, at smaller spatial and temporal scales, barrier inlet dynamics have been associated with the initiation, stabilisation and breakdown of individual beaches and complexes of beaches. We therefore envisage multiple scales of “resilience” operating simultaneously across the complex, responding to different forcing agents with particular magnitudes and frequencies.

Keywords: Barrier; Gravel; Inlet system; Saltmarsh; Reclamation; Sea-level change

*Corresponding author (email: A.J.Long@Durham.ac.uk)

1. Introduction

“Coastal resilience” describes the self-organising ability of a coast to respond in a sustainable manner to morphological, biological and/or socio-economic pressures (Klein et al., 1998). From a morphological perspective, it is a concept that is helpful in understanding the ability of a coastal landform to respond to external drivers that include relative sea-level (RSL) rise, an increase in storm magnitude/frequency, or a fall in sediment supply. A morphologically resilient coast is one that can maintain its long-term form despite experiencing short-term variations in the forcing processes, including sediment supply, on which it depends. Sand and gravel barriers can demonstrate morphological resilience over a range of temporal and spatial scales.

Over short-timescales (days to years), barriers may demonstrate resilience by dissipating or reflecting the energy of high waves or large storms and, though experiencing a degree of morphological change, are nevertheless able to maintain their structural integrity. However, barrier resilience over longer time-scales can involve larger-scale dynamic adjustment due to changes in RSL or variations in sediment supply with, for example, changes in inlet dynamics or phases of barrier progradation or erosion. Observational and geological records also show that the morphological and dynamic resilience of barriers can be exceeded when a geomorphological threshold is crossed and a barrier breaks down (Forbes et al., 1995; Jennings et al., 1998).

Conceptual models for the development of coarse clastic barriers suggest that barrier evolution, and by extension barrier resilience, is controlled by the interplay between the rate of RSL rise, sediment supply, longshore drift, basement topography, as well as storm incidence and magnitude (Orford et al., 1991, 2002). In general, low-gradient swash-aligned coasts that experience RSL rise are characterised by transgressive barriers whose resilience is strongly determined by the rate at which sediment is eroded from the shoreface and transported into the back-barrier (Roy et al., 1994). In the case of headland spits, such as Dungeness Foreland, these are progradational barrier complexes that extend down-drift in response to a dominant littoral drift pattern. Their morphological resilience depends on sufficient sediment supply along the barrier length. Progressive down-drift extension of these drift-aligned structures is characterised by the deposition of successive beach ridges. New ridge progradation occurs episodically during storms, although tidal currents are involved in building the sub-tidal platform on which the beaches accumulate.

Maintaining coastal resilience is increasingly viewed as a desirable outcome for coastal management since a resilient coast is better able to accommodate perturbations driven by natural and anthropogenic processes than one that has limited capacity for internal change (Nicholls and Bransen, 1998). However, we still do not thoroughly understand the roles that morphology, palaeogeography and sediment supply play in the resilience of sand and gravel barriers over century to millennial time-scales. This information gap has meant that many current coastal defence and erosion protection measures may not be attuned to the response mechanisms through which coastal resilience is enhanced.

In this paper we explore the evolution and resilience of Dungeness Foreland, a large sand and gravel barrier located in the eastern English Channel (Fig. 1) that comprises several hundred storm beaches that date from at least 4000 calibrated years before present (cal. yrs BP, AD 1950). Dungeness Foreland encloses an extensive area of fine-grained marsh sediments that have accumulated in the lee of the barrier (now Romney, Walland and Denge marshes). A recent programme of research, involving coring and analyses of back-barrier and barrier sediments, enables us to develop a new palaeogeographic model for the study area based on a review of a large database of palaeoenvironmental information. This model is based on several thousand boreholes, over 40 radiocarbon-dated microfossil diagrams, 39 new Optically Stimulated Luminescence (OSL) age determinations, as well as a rich archival and archaeological record summarised in a set of recent research monographs (Eddison and Green, 1988; Eddison, 1995; Eddison et al., 1998; Long et al., 2002) and web-based material (www.romneymarsh.net). Our focus is on the role of coastal morphology and palaeogeography in controlling coastal resilience over different spatial and temporal scales.

Central to our work is the history of three large tidal inlets which we consider have played a strong role in determining the morphological resilience of the Dungeness Foreland. When operating, these inlets were kept open by tidal flux which is a function of tidal range, wave climate and tidal prism (Bruun and Gerritsen, 1959). Studies elsewhere show that tidal inlets enable the deposition of large volumes of sediment in back-barrier areas that can provide a platform against which a barrier can stabilise and over which it may then migrate (FitzGerald et al., 2002; Cleary and FitzGerald, 2003; Davis and Barnard, 2003). Moreover, the shoreface sediment platform is also an important precursor to the deposition of new beaches on prograding barriers such as Dungeness Foreland. The location of tidal inlets relative to the barrier coastline is significant since it influences the input of sediment to the coastal cell and, hence, the pattern of sediment processing. This is particularly so for drift-aligned barrier complexes.

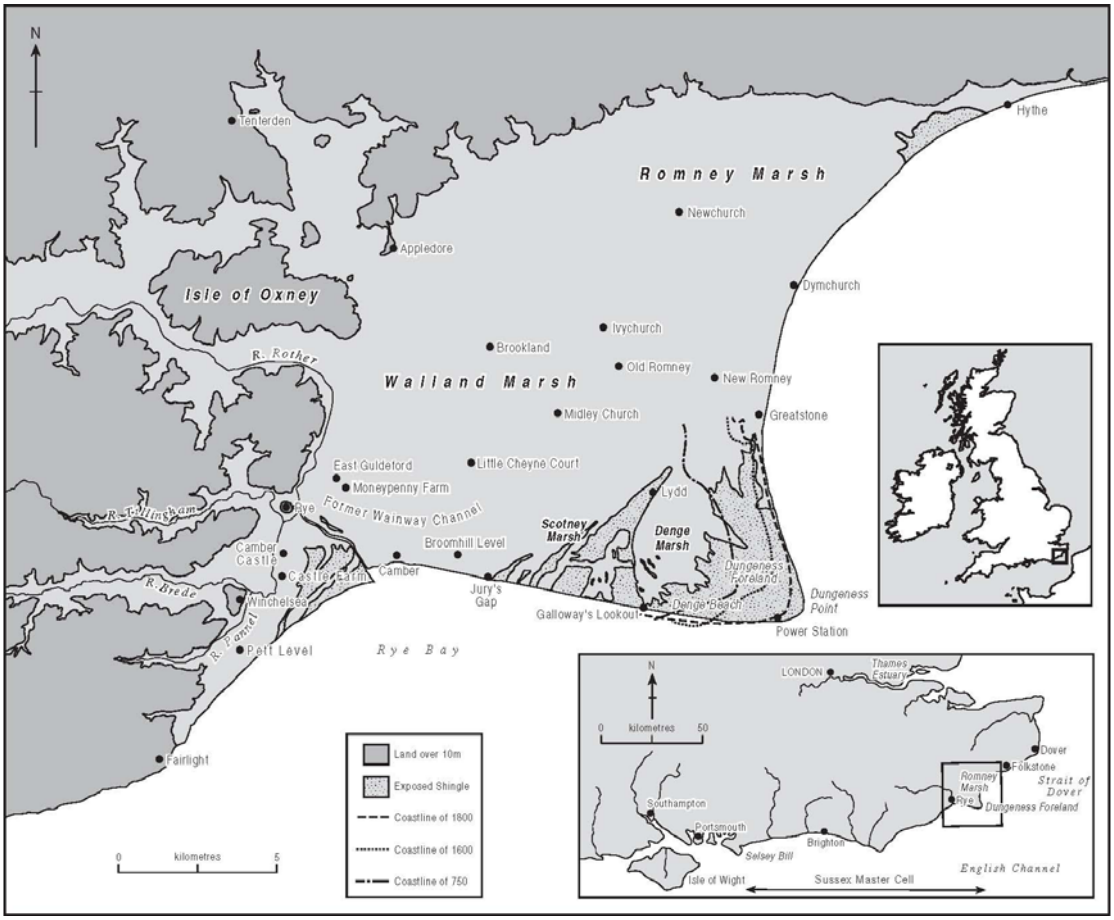

Fig. 1. Location map showing Dungeness Foreland and Romney/Walland Marshes in the eastern English Channel. The hypothetical shorelines shown are from Lewis and Balchin (1940).

The four principal aims of the paper are as follows:

1. To provide a broad-scale stratigraphic framework for the region by linking pre-existing data from the back-barrier area with newly acquired data from the Rye and Romney inlets and the beach complex of Dungeness Foreland;
2. Using this framework, to reconstruct the depositional history of Dungeness Foreland, with particular attention paid to the role of the tidal inlets that formerly existed at Hythe, Romney and Rye as a controlling influence on barrier resilience;
3. To explore the interdependence of the barrier, tidal inlet, and back-barrier sediments, and;
4. To examine the natural and human influences that contribute to the morphological resilience of Dungeness Foreland, and to explore the implications of this research for existing models of barrier evolution.

2. Physical setting and previous research

2.1 The location of Dungeness Foreland and relative sea-level change

Dungeness Foreland is located on the south coast of England, towards the eastern end of the English Channel (Fig. 1), at the down-drift limit of a sediment cell that extends from Selsey Bill to Dungeness (Nicholls, 1991). The main sources of gravel derive from offshore and the longshore movement of sediment released from the erosion of the chalk cliffs and associated Pleistocene deposits (Long et al., 1996) (Fig. 1). A platform of subtidal and intertidal deposits underlies much of Rye Bay, Dungeness Foreland and the back-barrier marshes (Greensmith and Gutmanis, 1990; Long and Innes, 1995; Dix et al., 1998). The deposition of the Rye Bay shelf sand body (SSB) is explained by several factors (Dix et al., 1998). First, the area is located at the downwind end of the English Channel, one of the stormiest seas in the U.K, and it therefore experiences the typically high energy wind/wave conditions necessary for SSB deposition. Second, SSB development is favoured by a steep (>1°) shoreface (Roy et al., 1994). Outside the limits of Rye Bay, bedrock outcrops as an eroded planar surface with a gradient typically <0.6°, whereas across Rye Bay bedrock gradients steepen from the exposed bedrock platform that outcrops in the intertidal zone beneath Fairlight Cliffs (Fig. 1). Finally, following the early Holocene opening of the Strait of Dover, stratigraphic evidence and palaeotidal modelling (Austin, 1991) suggests that the tidal range increased significantly and a strong nearshore easterly movement of sand began towards the Strait of Dover (Long et al., 1996).

The beach gravel of Dungeness Foreland mostly comprises cherty sandstones derived from the Upper Greensand, fine-grained sandstones from the Upper Greensand and various quartzites. The gravel on the beach ridge crests are generally finer than that in the lows, and grain size varies between c. 8 mm and 150 mm (Green, 1968). Beach ridge amplitude varies between c. 0.5 m and 2 m (Plater and Long, 1995).

The present tidal range at Dungeness is 6.7 m and the height of mean high water spring tides at Dungeness Point is +4.03 m OD. Relative sea-level (RSL) in this region has generally risen during the Holocene (Fig. 2) (Waller and Long, 2003;

Long et al., 2006). The rate of rise was greatest before c. 5000 cal. yrs BP, typically >3 mm yr^{-1}, after which there was a pronounced slow-down to <2 mm yr^{-1} as the global production of meltwater fell. Trends from the late Holocene are difficult to establish because most of the data from the last 4000 years have been lowered from their original elevation by sediment (including peat) compaction. The exposed gravel at Dungeness Foreland covers c. 2160 ha, with a further 1150 ha of gravel lying buried beneath marsh sediments.

2.2 Previous work

2.2.1 Dungeness Foreland

A continuous barrier running from Fairlight towards Hythe has been an element of most evolutionary models of Dungeness Foreland, including those dating from the nineteenth century (Elliott, 1847; Lewin, 1862; Drew, 1864; Burrows, 1884; Gulliver, 1897). However, these early models were largely speculative being based on scant stratigraphic evidence and no absolute dating. Following a systematic mapping of the morphology of the Dungeness beaches, Lewis (1932) and Lewis and Balchin (1940) assigned ages to individual shorelines based on cartographic and documentary evidence. They suggested that the beach ridges near Broomhill Level (Fig. 1), which are the lowest that outcrop at the surface in the study area, accumulated in the pre-Roman period. This suggestion has subsequently been confirmed by radiocarbon dates of c. 3500 cal. yrs BP from organic deposits that overlie the beaches (Tooley and Switsur, 1988; Plater et al., 2002).

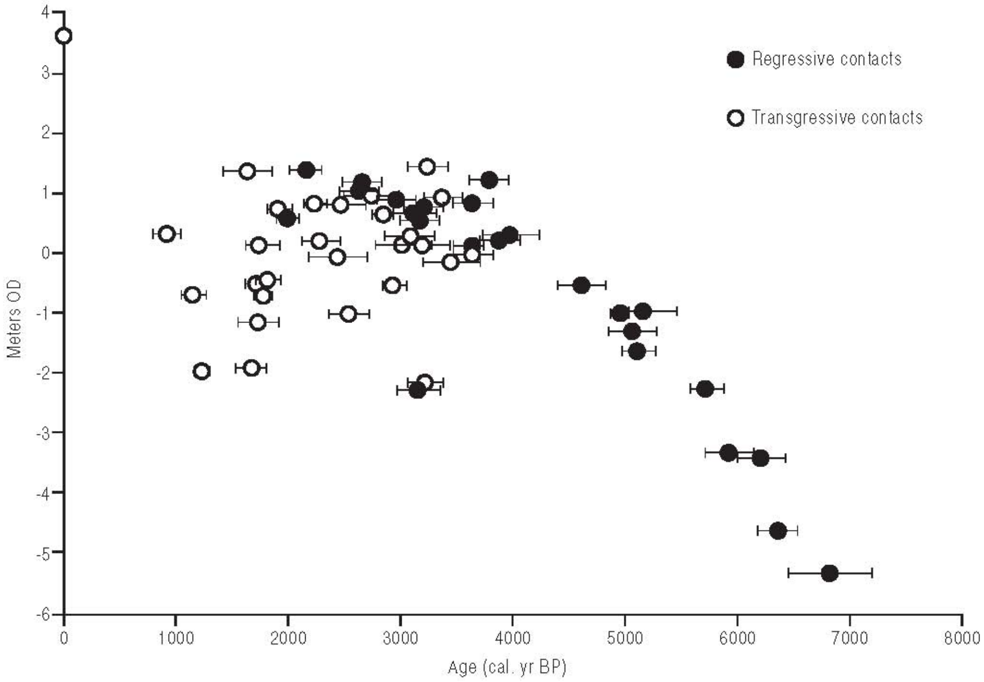

Fig. 2. Age/altitude graph depicting the age and elevation of transgressive and regressive contacts from the Dungeness Foreland/Romney Marsh depositional complex (from Long et al., 2006). All of the data have been lowered from their original elevation by sediment compaction, a process that particularly affects points from the transgressive contact to the thick main marsh peat.

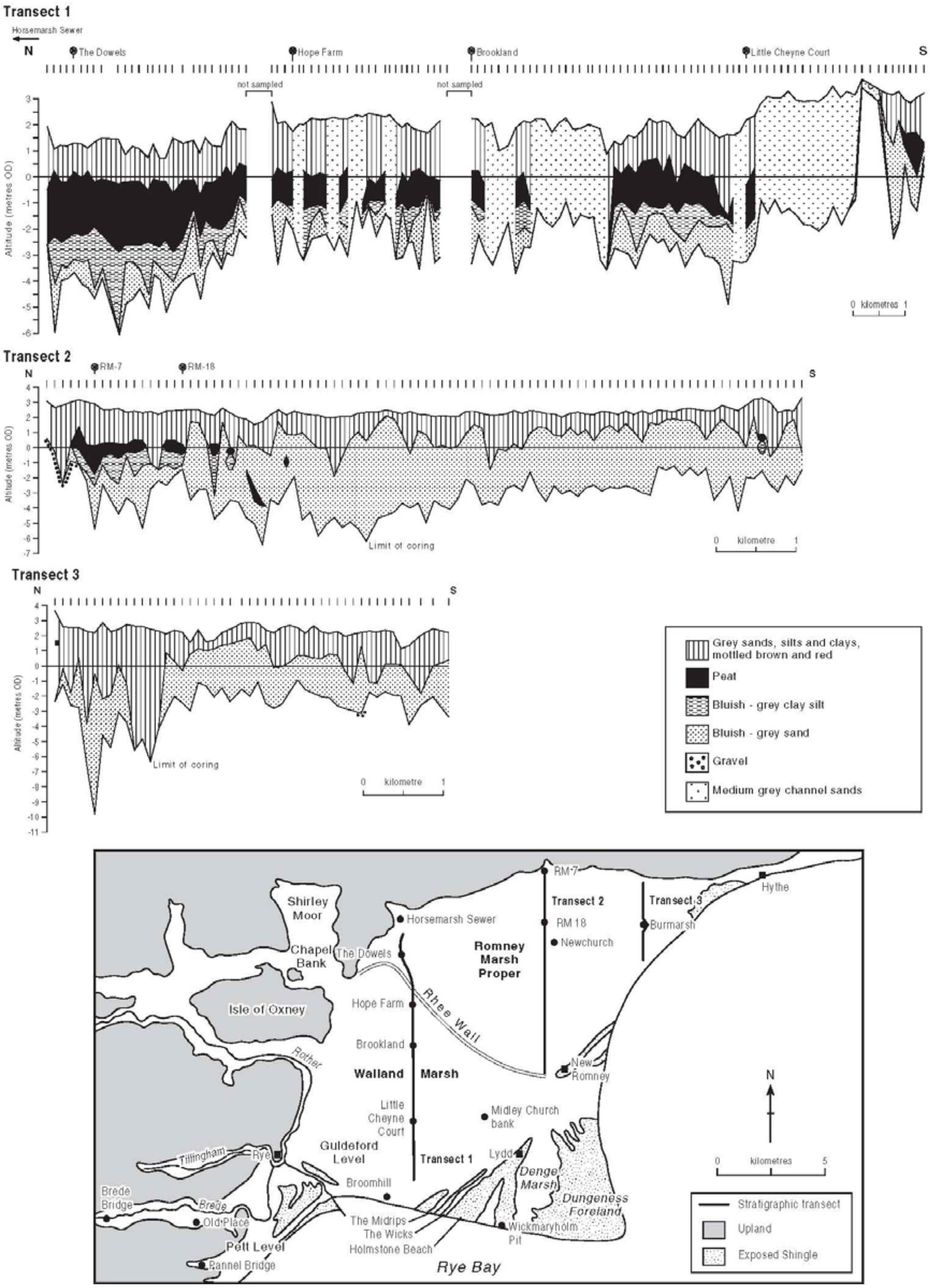

Fig. 3. Simplified lithostratigraphy of Romney and Walland Marshes; Transect 1 (Long and Innes, 1995), Transects 2 and 3 (Long et al., 1998). The location map shows the transect locations.

A series of deep boreholes in the region of the Dungeness nuclear power station (Fig. 1) penetrate to bedrock (at -30 m to -35 m OD) and reveal a stratigraphic sequence indicating coastal emergence, with offshore and lower shoreface sands that are overlain by upper shoreface sands, surface gravel and storm beach gravels (Hey, 1967; Greensmith and Gutmanis, 1990). This sequence is partly repeated in a borehole from Holmstone Beach between Jury's Gap and Galloway's Lookout (Fig. 1) (Plater et al., 2002). Further evidence of a progradational shoreface that pre-dates beach ridge deposition comes from an offshore seismic investigation which identified a series of convex-upward reflectors indicative of a seaward-prograding shelf sand body that provided the platform on which the foreland beaches accumulated (Dix et al., 1998, see above). At the nuclear power station site, Greensmith and Gutmanis (1990) dated detrital shell and other organic material to establish a minimum age of c. 3100 cal. yrs BP for a sandy facies (beneath the 5 to 6 m thick surface gravels). However, radiocarbon and OSL dates, detailed below in Section 5, indicate that the main body of the extant surface gravel beaches here accumulated more recently, probably in the last two thousand years.

2.2.2 Tidal inlet history

Previous research has described late Holocene tidal inlets at Hythe, New Romney and Rye (Fig. 1) (e.g. Green, 1968; Cunliffe, 1980, 1988; Green, 1988; Eddison, 2000; Rippon, 2002). The Hythe inlet, which is likely to have been an early conduit for the river Rother (or "Limen"), existed during the Roman period (Cunliffe, 1980). By the late Saxon period this inlet was much reduced in size and a breach had developed at Romney, where a Medieval port prospered before sediment infilled the harbour in the 13th century AD. The third inlet at Rye is widely believed to have developed following a catastrophic breach of the barrier during storms in the 13th century AD. This inlet and its associated harbours persisted until the 17th century AD (Eddison, 1998, 2000).

2.2.3 Romney and Walland Marshes

Lithostratigraphic investigations across Romney and Walland Marshes (Waller et al., 1988; Long and Innes, 1995; Long et al., 1996, 1998; Spencer et al., 1998a, b) corroborate a basic stratigraphic model proposed by Green (1968) who identified four stratigraphic units above bedrock: a lower sand, blue clay, main marsh peat and younger alluvium (Fig. 3). The lower sand accumulated from c. 7800 cal. yrs BP, when RSL rise was rapid and when the tidal range increased following the opening of the Strait of Dover (Long et al., 1996). Diatoms and foraminifera demonstrate that the blue clay immediately beneath the peat accumulated under intertidal mudflat and saltmarsh environments. The peat is up to 6 m thick in the valleys on the western side of Walland Marsh and thins eastward - near Midley Church the peat is only c. 0.5 m thick (Long and Innes, 1993). The basal and upper contacts of this unit rise in altitude in an easterly direction, suggesting that their deposition occurred as a back-barrier inlet infilled (Allen, 1996; Spencer et al., 1998a). Radiocarbon dates from the base of the peat show that it spread from the valleys in the west after c. 6000 cal. yrs BP and by c. 3000 cal. yrs BP was forming across Walland Marsh and abutting the western edge of Dungeness Foreland. Peat is restricted to the northern edge of Romney Marsh proper with thick deposits of laminated sands and silts present across much of the central and southern part of the marsh (Long et al., 1998; Fig. 3).

Dates of between c. 3000 and c. 1700 cal. yrs BP have been obtained from the top of the peat (Long et al., 1998; Waller et al., 1999). The upper contact of the peat is almost always abrupt and locally shows signs of erosion. In places, tidal creeks have cut through the peat and removed it entirely. The largest channel (historically referred to as the "Wainway Channel") is recorded to the south of Moneypenny Farm and Little Cheyne Court (Figs. 1 and 3 (Transect 1)). Other smaller channels (typically several tens to hundreds of meters across) are common in central and southern parts of Walland Marsh, although they are much less frequent across the northern marshlands.

3. Methodology

The lithological data presented in this paper for the marshland sediments were mainly collected using a gouge corer, with sample cores for laboratory analysis retrieved using a percussion or "Russian" corer. The deep boreholes that extend through the gravel of Dungeness Foreland, and which were used for the collection of samples for OSL dating, were drilled using

Table 1. Radiocarbon dates used in Fig. 4. Dates are calibrated using Calib 5.0.1 (Reimer et al., 2004).

Location	**Material dated**	**Laboratory code**	**Radiocarbon age ±1 SD**	**Calibrated radiocarbon age, yrs BP (± 2 SD)**	**Source**
Pewis Marsh	Peat (humin) Peat (humic acid)	GrN-27876 GrN-27913 Pooled mean	3500±30 3380±80 3485±28	3650-3838	Waller et al. (2006)
West Winchelsea	Fine rootlets Fine rootlets	GrA-25302 OxA-13460 Pooled mean	1170±35 1297±28 1248±27	1088-1269	Long et al. (2006)
West Winchelsea	Peat (humin) Peat (humic acid)	GrN-28734 GrN-28735 Pooled mean	1360±30 1300±60 1348±27	1185-1309	Long et al. (2006)
Rye 11	Peat	Beta-75451	5590±70	6222-6533	Long et al. (1996)
Little Cheyne Court	Peat	SRR-5611	1050±45	804-1062	Waller et al. (1999)
Little Cheyne Court	Peat	SRR-5614	4410±45	4860-5276	Waller et al. (1999)
Wainway Channel	*Cerastoderma edule*	Beta-127959	1210±50	655-925	Evans et al. (2001)
Tishy's Sewer, Broomhill Level	Peat	Q2651	3410±60	3484-3832	Tooley and Switsur (1988)
Tishy's Sewer, Bromhill Level	Peat	Q2652	3160±60	3219-3554	Tooley and Switsur (1988)
Wickmaryholm Pit	Plant macrofossil	OxA-12685	1652±25	1422-1686	This paper
Muddymore Pit	Plant macrofossil	GrA-22408	930±30	782-925	Schofield and Waller (2005)
Manor Farm	*Cerastoderma edule* *Cerastoderma edule*	Beta-160061 Beta-160060 Pooled mean	1620±40 1590±40 1605±28	1048-1306	Plater et al. (2002)
Cockles Bridge	Whale skull Whale bone	UB-4175 UB-4176 Pooled mean	1448±24 1468±24 1458±17	914-1175	Gardiner et al. (1999)

Marine shells and whale bones are calibrated using the Marine04 dataset, corrected for the local marine reservoir effect (ΔR) in the Eastern English Channel of -32 ± 56 yrs (Harkness, 1983). Dates from peat at Pewis Marsh and West Winchelsea, on the humic acid and humin fractions of the same sample, are also presented as pooled means. The West Winchelsea AMS dates on fine rootlets fine rootlets are AMS dates on adjacent samples from the same depositional unit (Long et al., 2006).

casing with a steel liner to prevent back-filling in conjunction with a 38 mm diameter plastic-lined steel sampling chamber that was percussion driven into the sands. The OSL dates were obtained from coarse (sand-sized) quartz grains using the Single Aliquot Regenerative (SAR) dose measurement protocol (see Roberts and Plater, 2005). All site elevations are reported in meters with respect to the U.K. Ordnance Datum (OD) which approximates mean sea-level. The detailed methods and results of the newly collected litho-, bio and chronostratigraphic data summarised in this paper can be found in Roberts and Plater (2005), Schofield and Waller (2005), Long et al. (2006), Waller et al. (2006), Waller and Schofield (2006) and Stupples and Plater (submitted).

This paper uses a chronology based on a variety of data sources. The radiocarbon dates (Table 1) are cited in calibrated years before present (cal. yrs BP), where BP is AD 1950. Freshwater (terrestrial) samples are calibrated using the CALIB 5.0.1 programme (Reimer et al., 2004). Marine samples are calibrated using the Marine04 dataset with a local marine reservoir effect (ΔR) for the English Channel of -32 ± 56 yrs (Harkness, 1983). The OSL dates were originally calculated in years before 2000 AD (Roberts and Plater, 2005), but here we adjust them to years before AD 1950 to allow direct comparison to the calibrated radiocarbon dates (Table 2). Finally, historical dates are referred to in years AD.

4. A new stratigraphic model for the Romney Marsh/Dungeness Foreland depositional complex

In this section we present a new stratigraphic model for the south coast of the depositional complex that links the barrier and back-barrier sediments, including the tidal inlets that once existed at Romney and Rye (Fig. 4). The transect is divided into seven sections.

4.1. West of the River Brede (Fig. 4, Section 1)

Peat accumulation slowed from c. 4000 cal. yrs BP at the upland edge on the western side of the study area. At Pewis Marsh, for example, highly humified peat from the upper contact has been dated to c. 3700 cal. yrs BP (Waller et al., 2006). A thin slope-wash deposit that mantles the peat is dated to between c. 1700-2200 cal. yrs BP (Waller et al., 2006). This provides a minimum age for the return

Table 2. Equivalent dose (D_e), dose-rates, and optically stimulated luminescence (OSL) ages used in Fig. 4 (Roberts and Plater, 2005).

[†]. Water content expressed as a percentage of the mass of dry sediment, calculated using field values in conjunction with a model of water history for the area.
[*]. The error shown on D_e is the standard error on the mean.
[‡]. 'n' is the number of D_e determinations.
[||]. Dose-rate values (Gy/10^3 yr) were calculated using the conversion factors of Adamiec and Aitken (1998) and are shown rounded to 3 decimal places, although the total dose-rates and ages were calculated using values prior to rounding. Central values are given for dose-rates – errors are incorporated into that given for the total dose-rate.

Sample No.[*]	Depth (m)	Water content (%)[†]	Grain size analysed (μm)	D_e (Gy)[*]	'n'[‡]	Total dose-rate[||]	Age (10^3 yr)[¶]
73BH-1/1	3.75	21 ± 5	150-180	3.70 ± 0.06	31	0.79 ± 0.03	4650 ± 200
73BH-1/2	5.95	24 ± 5	150-180	3.11 ± 0.05	33	0.77 ± 0.03	4020 ± 170
73BH-1/3	8.05	25 ± 5	150-180	3.44 ± 0.06	18	0.67 ± 0.03	5070 ± 220
73BH-2/1	11.25	25 ± 5	150-180	4.31 ± 0.09	18	0.92 ± 0.04	4660 ± 220
73BH-2/2	12.25	25 ± 5	150-180	3.81 ± 0.08	17	0.92 ± 0.04	4120 ± 190
73BH-2/3	13.60	25 ± 5	150-180	3.55 ± 0.09	12	0.84 ± 0.03	4190 ± 190
73BH-3/1	12.90	25 ± 5	150-180	2.37 ± 0.06	17	0.65 ± 0.03	3580 ± 160
73BH-3/2	13.90	25 ± 5	150-180	2.42 ± 0.06	18	0.63 ± 0.03	3800 ± 180
73BH-3/3	15.75	25 ± 5	150-180	3.73 ± 0.09	17	0.92 ± 0.04	4020 ± 190
73BH-4/1	9.95	25 ± 5	150-180	1.34 ± 0.03	17	0.74 ± 0.03	1780 ± 90
73BH-4/2	11.55	25 ± 5	150-180	1.75 ± 0.04	18	0.78 ± 0.03	2210 ± 110
73BH-4/3	14.55	25 ± 5	125-150	2.30 ± 0.05	13	0.92 ± 0.04	2460 ± 110
73BH-5/1	9.90	25 ± 5	150-180	1.41 ± 0.04	19	0.71 ± 0.03	1940 ± 100
73BH-5/2	10.85	25 ± 5	150-180	1.39 ± 0.04	18	0.78 ± 0.03	1740 ± 80
73BH-6/1	10.65	25 ± 5	150-180	1.10 ± 0.02	14	0.57 ± 0.02	1890 ± 80

*73*BH-6/2	13.55	25 ± 5	150-180	1.40 ± 0.05	15	0.70 ± 0.03	1940 ± 100
*73*BH-6/3	14.70	25 ± 5	150-180	1.55 ± 0.03	26	0.80 ± 0.04	1890 ± 100
*73*BH-7/1	10.30	25 ± 5	125-150	1.18 ± 0.02	12	1.20 ± 0.05	930 ± 40
*73*BH-7/2	12.55	25 ± 5	125-150	1.17 ± 0.04	14	0.85 ± 0.03	1340 ± 70
*73*BH-7/3	14.70	25 ± 5	125-150	1.20 ± 0.03	29	0.81 ± 0.03	1450 ± 70
*73*BH-8/1	17.25	25 ± 5	150-180	0.78 ± 0.03	15	0.57 ± 0.02	1310 ± 70
*73*BH-8/2	14.00	25 ± 5	150-180	0.65 ± 0.02	13	0.50 ± 0.02	1250 ± 70
*73*BH-8/3	21.00	25 ± 5	150-180	0.91 ± 0.03	14	1.24 ± 0.05	690 ± 40
*73*BH-9/1	4.85	25 ± 5	180-212	0.60 ± 0.03	15	0.57 ± 0.02	1000 ± 70
*73*BH-9/2	7.35	25 ± 5	150-180	1.04 ± 0.03	17	1.01 ± 0.04	990 ± 50
*73*BH-9/3	9.15	25 ± 5	150-180	1.03 ± 0.03	14	1.04 ± 0.04	940 ± 50
*73*BH-10/1	5.85	25 ± 5	150-180	0.56 ± 0.01	44	0.57 ± 0.02	930 ± 40
*73*BH-10/2	7.85	25 ± 5	150-180	0.71 ± 0.01	27	0.74 ± 0.03	910 ± 40
*73*BH-10/3	9.75	25 ± 5	150-180	1.24 ± 0.03	20	1.21 ± 0.05	980 ± 50
*73*BH-11/1	1.80	23 ± 5	150-180	0.35 ± 0.01	35	0.82 ± 0.03	380 ± 20
*73*BH-11/2	4.50	25 ± 5	150-180	0.46 ± 0.02	12	0.91 ± 0.04	460 ± 30
*73*BH-11/3	6.75	25 ± 5	150-180	0.81 ± 0.03	14	0.93 ± 0.04	820 ± 50
*73*BH-12/1	15.05	25 ± 5	150-180	0.51 ± 0.01	32	0.75 ± 0.03	630 ± 30
*73*BH-12/2	15.50	25 ± 5	150-180	0.46 ± 0.01	14	0.75 ± 0.03	560 ± 30
*73*BH-12/3	15.95	25 ± 5	150-180	0.47 ± 0.02	15	0.67 ± 0.03	650 ± 40
*73*BH-USS	0.01	25 ± 5	150-180	0.03 ± 0.03	12	0.76 ± 0.03	-10 ± 40
*73*BH-SSR	0.01	25 ± 5	150-180	0.02 ± 0.02	11	1.15 ± 0.05	-35 ± 15

¶Luminescence ages are expressed as years before 1950 AD, and rounded to the nearest 10 years.

of marine conditions to this area. Peat formation was sustained for a much longer period in the central part of the Brede valley near West Winchelsea, where saltmarsh developed on top of the peat only after c. 1300 cal. yrs BP (Long et al., 2006). Several centuries after this initial inundation there was a rapid increase in water depth and tidal energy, probably associated with historically documented flooding in the 13th century AD which resulted in the deposition of c. 4 m of tidally-laminated sediments before site reclamation by c. 1460 AD.

4.2. The Rye breach; River Brede to East Guldeford (Fig. 4, Section 2)

The peat and overlying clastic sediments referred to in Section 4.1 are extensive to the west of the River Brede, whereas to the east is a wide expanse of late Holocene barrier beaches. We suspect that the peat once extended uninterrupted across this area but has been eroded by the large tidal channels that developed here in the Medieval period (the Rother, Wainway Channel and Brede channels). A deep borehole that penetrated the late Holocene beaches at Castle Farm (Fig. 4) yielded three OSL dates in age sequence and record sand accumulation from 890 ± 70 yrs BP through to 560 ± 30 yrs BP after which the surface gravel was deposited (Roberts and Plater, 2005). The oldest of these dates corroborates the radiocarbon evidence from the lower Brede (West Winchelsea) suggesting a tidal inlet in this area from at least 1000 yrs BP, with sand accumulating on an intertidal or subtidal shoreface in front of the contemporaneous shoreline.

The transect extends across the former Wainway Channel, now infilled by a deep sequence (from c. +2.5 m OD to at least to -3 m OD) of tidally laminated sands and silts with occasional pockets of eroded peat and broken shells. The northern edge of this channel marks the limit of the erosion associated with the Rye inlet and is notable for a surface outcrop of gravel which forms a series of small cuspate beaches in the Moneypenny Farm area (Fig. 4). The conventional marshland sequence (see section 2.2.3) occurs to the north of this gravel outcrop where the upper levels of the peat contain a layer rich in *Sphagnum* macrofossils indicating the development of a raised bog during the later stages of peat growth; dating evidence at Little Cheyne Court suggests raised bog development after c. 2700 cal. yrs BP (Waller et al., 1999). The bog was inundated soon after c. 900 cal. yrs BP at East Guldeford and Little Cheyne Court, and buried by marine/brackish sediments that extend to the modern surface. In contrast, at Moneypenny Farm these post-peat sediments are overlain by sand and the surface outcrop of gravel. A sample of this sand yielded an OSL age of 460 ± 30 yrs BP (Roberts and Plater, 2005).

4.3. Little Cheyne Court to Jury’s Gap (Fig. 4, Section 3)

Inundation of the raised bog at Little Cheyne Court occurred as a result of flooding from the adjacent Wainway Channel. This channel is c. 1000 m wide between Little Cheyne Court and Sandylands (Broomhill Level) (Green, 1968; Long and Innes, 1995). Near to Little Cheyne Court, there is evidence for a marked up-core change in channel stratigraphy, from coarse, saturated sands to finer-grained, well-laminated sands and silts above c. -1 m OD (Fig. 4). A single valve of the mollusc *Cerastoderma edule* from the base of the laminated sand yielded a date of 655-925 cal. yrs BP (Evans et al., 2001).

Buried gravel of mid to late Holocene age occurs across much of Broomhill Level (Fig. 4). OSL dates indicate the shoreface sands below this gravel were in place by c. 4700 yrs BP (Roberts and Plater, 2005). Locally the gravel subcrop is overlain by a 0.5 m-thick organic deposit which, dating at Tishy's Sewer from c. 3200-3800 cal. yrs BP, provides a maximum age for gravel deposition (Tooley and Switsur, 1988). To the north-east, in the Scotney Marsh area (Fig. 1), peat abuts the western margin of the gravel outcrop of Dungeness Foreland (Spencer, 1997; Spencer et al., 1998a, b). Radiocarbon dates indicate several phases of peat development here between c. 3800 and c. 1600 cal. yrs BP (Spencer and Woodland, 2002). The clastic sediments overlying and intercalated with these peats suggest that tidal waters penetrated behind the gravel barrier from early as c. 3200 cal. yrs BP, with repeated tidal flooding until at least the late Roman period. These tidal waters may have penetrated the barrier from the south coast of the foreland, or from a tidal inlet at Hythe or Romney (see below). A marine/brackish influence in the Scotney Marsh area during the Roman period is supported by archaeological evidence for saltmaking (Barber, 1998).

4.4. Dungeness Foreland – Jury's Gap to Galloway's Lookout (Fig. 4, Section 4)

The gravel outcrop has a surface relief of +3 m to +4 m OD at Jury's Gap and rises in an easterly direction to c. +5 m OD on Holmstone Beach (Fig. 4). OSL dates from the Midrips indicate that the shoreface sands below the SW-NE trending gravel ridges here were in place soon after the equivalent deposits at Broomhill Level, c. 4700 yrs BP (Roberts and Plater, 2005). These older inner beach ridges extend NW beyond Lydd and indicate an early period of drift-aligned sediment movement and barrier extension towards Hythe.

The rate of eastward barrier development appears to rise after c. 2000 yrs BP (Fig. 4), though this may equally reflect a shift in the axis of foreland progadation (Roberts and Plater, 2005). Thus, OSL dates from Holmstone Beach, South Brooks and Lydd Beach all provide a similar minimum age for gravel deposition of c. 1900 BP. Independent support for this chronology comes from a radiocarbon-dated plant macrofossil at the base of Wickmaryholm Pit; a natural waterlogged depression on the foreland surface (Long and Hughes, 1995) (Fig. 4).

Several areas of marshland sediment interrupt the continuity of the gravel beaches along the south coast of the foreland (e.g. the Midrips, Wicks and South Brooks). Each is infilled with up to 4 m of sediment, typically comprising a lower laminated sandy silt and an upper, mottled silt clay. Diatoms from a core collected from South Brooks (Long and Hughes, 1995) demonstrate deposition occurred under tidal channel conditions, whilst the tidally-laminated sediments within these sites are indicative of sediment accretion rates of the order of 0.3 m yr^{-1} (Stupples and Plater, submitted). Air photographs, as well as Green's (1968) soil map, show that some of these tidal channels cross-cut older beaches and, therefore, post-date gravel deposition. The origin and, in particular, the age of these marsh sediments are difficult to establish due to the lack of *in situ* carbonaceous material for dating. Eddison (1983) has observed ridge and furrow patterns in the Wicks, which are interpreted as evidence for an "early phase" of agriculture. A "later phase" of more intense activity, perhaps associated with saltmaking, is suggested by a set of closely spaced rectangular ditches, on top of which is a sea wall that dates from the mid-13th century AD and which was part of an extensive set of walls across Walland Marsh built in response to the threat of flooding at this time (Eddison, 1983; Eddison and Draper, 1997).

4.5. Galloway's Lookout to Denge Marsh (Fig. 4, Section 5)

The gravel beaches rise further in elevation between the eastern edge of Lydd Beach at Galloway's Lookout and Muddymore Pit, with one set of beaches reaching a maximum altitude of c. +6.5 m OD (Lewis and Balchin, 1940; Plater and Long, 1995). These high beaches lie to the west of Denge Marsh and several of them curve back on themselves to form a prominent beach that runs in an arc back along the line of the Dengemarsh Road and defines the eastern edge of the town of Lydd. This beach can be traced northwards as a thin finger of gravel that extends towards New Romney. To the east and north of this beach are the fine-grained marsh sediments of Denge Marsh.

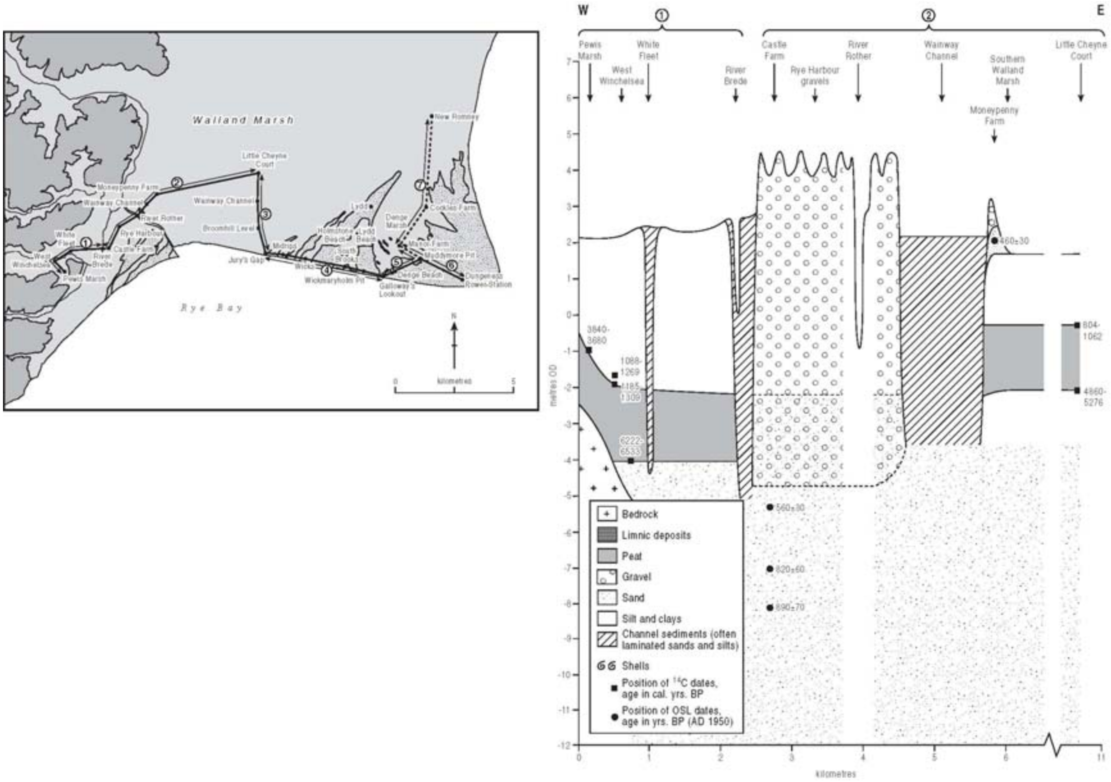

Fig. 4. Stratigraphic transect along the south coast of the Romney Marsh/Dungeness Foreland depositional complex. Radiocarbon dates are cited in calibrated years before present (AD 1950) (cal. Yr BP) with a two sigma age range. Marine samples are calibrated using the Calib Marine04 dataset with a local marine reservoir effect (ΔR) for the English Channel of −32±56 yr (Harkness, 1983). OSL ages are presented in full in Roberts and Plater (2005) and Roberts and Plater (in press), and are cited in years before AD 1950 to enable comparison with the radiocarbon dates.

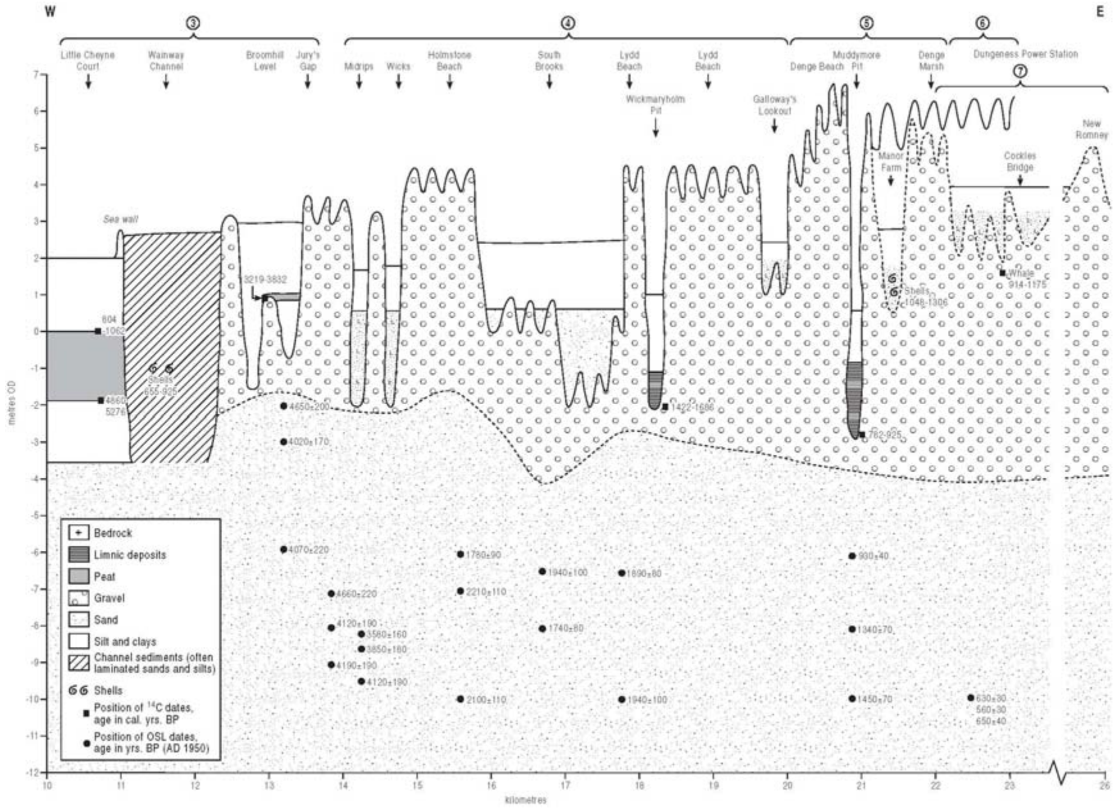

Fig. 4 (*continued*)

An age of c. 1000 yrs BP for the prominent Dengemarsh Road beach is derived from several sources. OSL dates from shoreface sands at Manor Farm suggest a minimum age for the overlying gravel of 930 ± 40 yrs BP (Fig. 4) (Roberts and Plater, 2005) and a radiocarbon date from the base of Muddymore Pit, provides a maximum age for gravel deposition of 782-925 cal. yrs BP (Schofield and Waller, 2005). The age difference between Lydd Beach (c. 1900 yrs BP) and the Dengemarsh Road beach implies a decline in the rate of foreland progradation. However, this simply reflects a change in the axis of progradation, as the intervening ridges were clearly deposited at the distal portion of the prograding foreland. Together, these data suggest that the foreland migrated rapidly in a north-easterly direction between c. 1900 and 1000 yrs BP.

Shells of *Cerastoderma edule* from laminated sands and silts that overlie gravel at Manor Farm (Fig. 4) are radiocarbon dated to 1048-1306 cal. yrs BP (Plater et al., 2002), whilst two dates on a whale skeleton found during gravel extraction to the north at Cockles Bridge indicate marine sedimentation here at 914-1175 cal. yrs

BP (Gardiner, 1998; Gardiner et al., 1999). Historical sources indicate that much of the area was used for salt manufacture from c. 1000 AD onwards (Vollans, 1995).

4.6. Denge Marsh to Dungeness Point (Fig. 4, Section 6)

East of Muddymore Pit, the main transect continues across the gravel beaches to the nuclear power station and Dungeness Point itself, which marks the easternmost limit of the complex. Beach ridge elevations continue to rise to the present ness, where altitudes of between +6 m and +7 m OD are attained. OSL ages on the sub-gravel shoreface are indicative of rapid coastal progradation between 930 ± 40 and 630 ± 30 yrs BP. Historical data, summarised in Lewis and Balchin (1940), indicate eastward progradation rates between 1617 AD and 1844 AD of at least 5.5 m yr^{-1}, and up to 3.6 m yr^{-1} between 1844 AD and 1939 AD (Fig. 1).

4.7. Denge Marsh to Romney (Fig. 4, Section 7)

Northeastward from Denge Marsh, OSL age data suggest rapid shoreface extension between 1310 ± 70 and 400 ± 20 yrs BP, with extension beyond Dengemarsh Road probably commencing as recently as 930 ± 40 yrs BP (Roberts and Plater, 2005). These observations confirm the suggestion of Lewis (1932) for easterly growth of the foreland during the post-Roman era, switching to more northerly accretion from late Saxon times. Indeed, the tightly recurved, short gravel beaches east of Denge Beach do not extend across Denge Marsh. This indicates that at this time the shoreface sand platform was not sufficiently developed to the north and north-east to enable continuous gravel barrier extension across the entire shore. This phase of barrier progradation clearly occurred under a very different set of dynamic controls to the previous 4000 years (i.e. 5000-1000 cal. yrs BP).

5. Palaeogeography of the Romney Marsh/Dungeness Foreland depositional complex

In the following sections we reconstruct the history of the Romney Marsh/Dungeness Foreland depositional complex. Particular emphasis is placed on the evolution of the three tidal inlets (first at Hythe, then Romney and Rye) and their influence on barrier development. The reconstructions allow us to explore the driving mechanisms responsible for the long-term resilience of Dungeness Foreland, as well as models of barrier evolution more generally.

5.1. The Hythe inlet (Fig. 5)

Geomorphic and archaeological evidence demonstrate a tidal inlet persisted in the north-eastern part of Romney Marsh proper during the late Holocene. Green (1968) attributed a large area of calcified soils in the area to this inlet. Indeed, the position of these deposits, stratigraphically above the peat along the northern edge of the inlet, shows that an expansion in tidal inundation across former freshwater wetlands occurred during the late Holocene. Radiocarbon dates of c. 3200 cal. yrs BP (RM18) and c. 2300 cal. yrs BP (RM7) from Romney Marsh proper (Fig. 1, Long et al., 1998) provide minimum ages for this inundation. However, the upper peat contact in these cores and elsewhere across Romney Marsh is abrupt and we suspect the inundation occurred later, possibly in the Roman period given the charcoal and burnt silt in the overlying sediment of RM18. Certainly archaeological evidence for saltmaking is abundant across large areas of Romney Marsh during the 1st and 2nd centuries AD (Cunliffe, 1988; Reeves, 1995), reflecting a pattern that was widespread in other coastal lowland areas in the UK and elsewhere on North Sea coasts during the Roman period (e.g. Hall and Coles, 1994; Rippon, 1996; Bonnot-Courtois et al., 2002; Behre, 2004). A third century AD Roman port known as *Limana* (below the fort now known as Stutfall Castle, Fig. 5) existed close to Hythe, near to the mouth of a tidal inlet (Cunliffe, 1980), and Gardiner et al. (2001) believes that there may well have been an earlier fort here too, perhaps dating from as early as 130 AD.

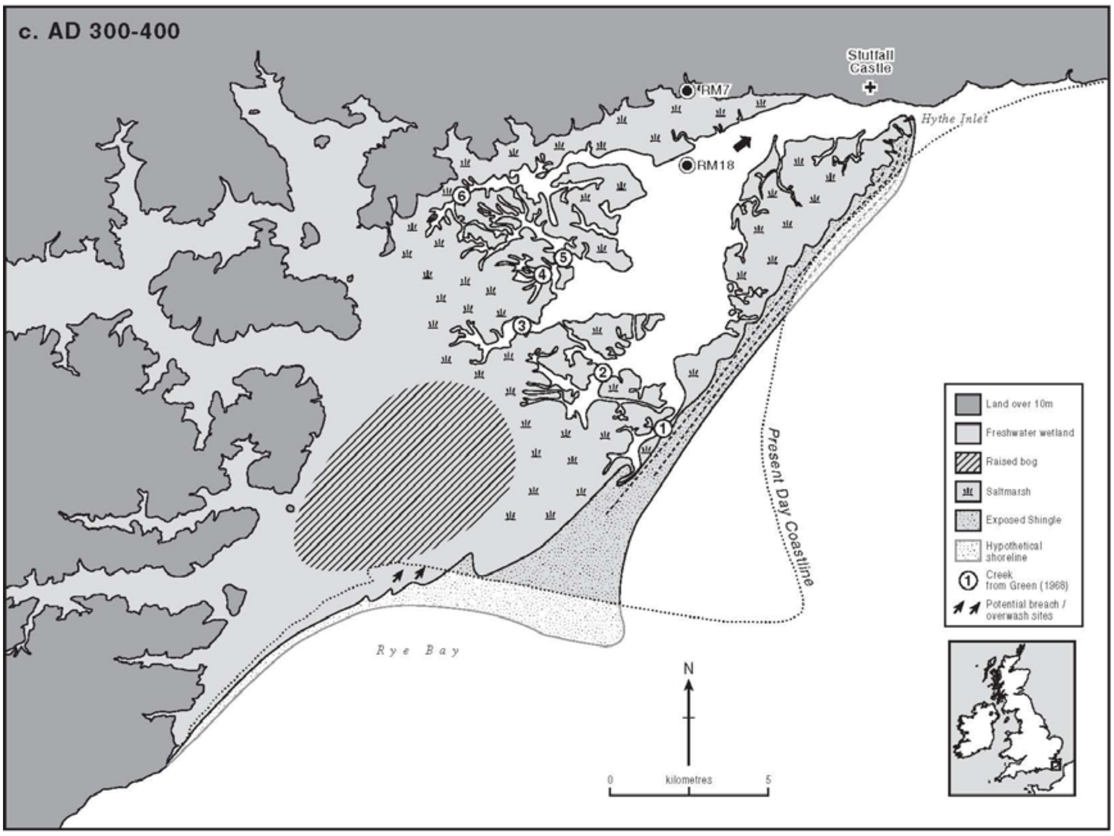

Fig. 5. Palaeogeographical reconstruction of the Romney Marsh/Dungeness Foreland depositional complex when the Hythe inlet was dominant (c. AD 300–400).

There is little archaeological evidence for the occupation of Romney Marsh proper between the 2nd century AD and the Saxon period. One hypothesis is that the area was inhospitable due to the marine flooding noted above increasing in intensity for at least several centuries after 200 AD (Cunliffe, 1988; Reeves, 1995). This could be related to a period of increased storm magnitude and frequency and/or a rise in RSL, with inundation accompanied by the localised erosion and subsequent compaction of the peatlands. Green (1968) suggests that the western limit of thick, near-surface peat, is close to a line that connects Lydd, Old Romney, Ivychurch and Newchurch (Fig. 1). However, tidal inundation probably reached further west, across onto what is today eastern Walland Marsh via several major west-east aligned creeks that are mapped by Green (1968) close to Brookland and Wheelsgate (creek numbers 2 and 3, Fig. 5). There is no evidence for a tidal inlet at either Romney or Rye prior to the 7th century AD.

Increased occupation of Romney Marsh from the mid-Anglo-Saxon period onward suggests that tidal waters were retreating by this time as the Hythe inlet infilled. Thus, Reeves (1995) records pottery dating from the 8th to the 10th century AD across much of Romney Marsh proper, indicating that reclamation and land settlement were well advanced by this time. Gardiner et al. (2001) track the infilling and closure of the Hythe inlet during this interval caused by the northward extension of sand and gravel beaches close to Sandtun; a port located on a sand spit close to Hythe. Two occupation layers here are dated to 690-775 AD and up to c. 840 AD, after which the site was sealed by blown sand. The closure of the Hythe inlet probably reflected a combination of a reduction in tidal prism (itself possibly associated a change in the outfall of the River Rother) and the northward drift of sand and gravel beaches from the Lydd-Dymchurch area. Aeolian deposition may also have been important and suggests an abundance of shoreface sand at this time.

5.2. The Romney inlet (Fig. 6)

Historical documents refer to the existence of Romney (or "Rumensea"), a port that stood on the shores of a tidal inlet, from at least AD 741 (Brooks, 1988; Eddison, 2000). This indicates that the east coast barrier was breached at broadly the same time (if not before) as the Hythe inlet was closing. We suspect that the two events may have been linked, and that the Romney breach resulted from cannibalisation of the proximal part of the spit complex. Such a process is an inherent tendency of drift-aligned barriers and would have been encouraged by increased northward drift of sediment towards Hythe. A second possibility is that a tidal channel that flowed along the inside of the Lydd to Hythe beaches, possibly Green's (1968) creek 1 (Fig. 5), weakened the barrier in the Romney area and promoted breaching.

It is not easy to reconstruct the former geometry of the newly established Romney inlet and the size of the associated back-barrier tidal prism (Fig. 6). Today, much of the inner part of the inlet is infilled with mounds of sand and silt, some thought to be associated with Medieval salt making (Vollans, 1995). However, the absence of the tidal flooding of Romney Marsh proper at this time suggests that a "probable" sea wall (Green, 1968) that can be traced in a westerly direction from New Romney, acted as a northerly boundary to this inlet (Fig. 6). This wall can be tracked further inland as the Rumensea Wall, which, according to Allen (1996, 1999), was a major sea defence that separated Walland from Romney Marsh proper from as early as perhaps 700 AD. The Rumensea Wall extends nearly all the way to the upland near Appledore, suggesting that it was built after the development of the Romney inlet and was designed to protect the economically important regions of Romney Marsh proper from flooding along its entire length.

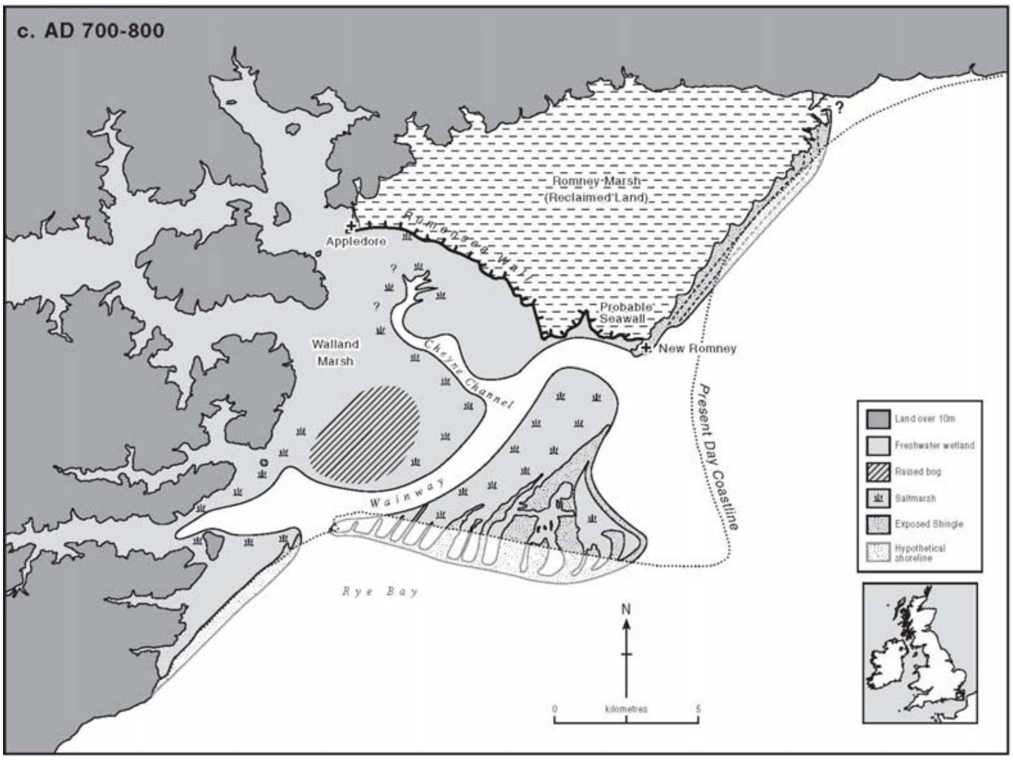

Fig. 6. Palaeogeographical reconstruction of the Romney Marsh/Dungeness Foreland depositional complex when the Romney inlet was dominant (c. AD 700–800).

The Rumensea Wall was constructed on the east bank of a small creek (the Rumensea). A second more substantial creek that appears to be associated with the Romney inlet can be identified to the south of Brookland in the stratigraphic transect of Long and Innes (1995) (Fig. 3, Transect 1). This channel is infilled with thick deposits of laminated sands and silts. It cuts across the main marsh peat which appears to been totally removed by erosion. Spencer et al. (2002) record similar tidal channel deposits, also incised through the peat, to the northwest of Midley Church. The distribution of the Brookland and Midley channel deposits closely matches a swath of Snargate-Finn soils mapped by Green (1968) that extend in a loop from close to Snargate, around Brookland and then south towards Cheyne Court. Compared to the Rumensea, this channel, termed here the “Cheyne Channel”, appears to be a much larger and wider (c. 1.2 km) feature and, therefore, one of considerably more significance to the Medieval landscape of Walland Marsh. Whether this provided a route for the river Rother south across Walland Marsh to Romney and thus the English Channel is not yet certain, although it must be considered likely.

The extent of flooding of Walland Marsh appears initially to have been limited by the presence of the raised bog. Indeed, the Cheyne Court area (immediately west of the Cheyne Channel) was not flooded until at least c. 900 AD, suggesting that it must have stood above the tidal waters for at least 200 years after the Romney breach. The soil survey map (Green, 1968) shows that the Cheyne Channel was confined between two major sets of embankments which were evidently constructed to limit flooding, both eastwards towards the Rumensea Wall and also to the southwest across Walland Marsh.

After about 1000 AD, the back-barrier area of the Romney inlet, including the Cheyne Channel, began to infill and the tidal prism fell. This process was greatly enhanced by reclamation from at least the 11th century AD onwards, if not before (Eddison and Draper, 1997). Much of Romney Marsh proper was densely populated at this time, the pressure to acquire new land was high, and the intertidal expanses associated with the Romney inlet must have been a prime target for reclamation. A reduction in size of the Romney tidal prism would have had several consequences. The first, as observed on the German and Belgium coasts during the Medieval period (Behre, 2004; Meier, 2004), is to cause a reduction in the area available for accommodating storm waters. This would have increased the likelihood of catastrophic coastal inundation during storms, especially when combined with compaction and lowering of the reclaimed and drained land behind the sea walls. A second consequence would have been to reduce the cross-sectional area of the tidal inlet and promote sediment infilling of the breach. Tidal inlet reduction can cause a decrease in the size of the ebb-tidal delta and the onshore migration of large swash bars, as noted by FitzGerald et al. (2002) at Chatham Harbor Inlet (Cape Cod, U.S.A). This appears to have occurred as the Romney inlet infilled, with the development of a wide intertidal platform of sands and silts providing the foundation on top of which Dungeness Foreland could prograde.

There is abundant historical and geomorphological evidence that these processes were accelerating from the 11th century AD onwards. The progressive infilling of the tidal inlet was aided by silting of the back-barrier area and tidal inlets across Walland Marsh, including parts of Denge Marsh and Belgar (Fig. 6). This process continued despite the construction of the Rhee Wall; an artificial watercourse that parallels the Rumensea Wall that was built in a failed attempt to flush out the harbour during the 13th century AD with tidal water which by this time extended northwards from the Rye area along the western edge of Walland Marsh to Appledore (Vollans, 1988; Green, 1988; Eddison, 2002).

5.3. The Rye inlet (Fig. 6)

Historical evidence demonstrates the presence of a major tidal inlet at Rye in the 13th century AD (Green, 1968; Eddison, 1998). However, it is now clear that a breach in the barrier occurred here from c. 700 AD, at which time the Romney and Rye inlets may have been joined (Fig. 6). Evidence for this early breach is provided by the radiocarbon dates (c. 1300 cal. yrs BP) for the end of peat formation at West Winchelsea and the development of saltmarsh environments here and at East Guldeford shortly after. A breach near Rye would have caused a substantial reduction in the longshore drift of sediment across Rye Bay and the reworking of down-drift portions of the foreland along the south coast (i.e. to the east of the breach). The marked change in beach ridge orientation and distal extent in the beach ridges east of Galloway’s Lookout suggest rapid north-eastward progradation of the ness at this time. With both tidal inlets open, tidal waters initially extended across the upstanding peatlands of southern Walland Marsh, including the margins of the raised bog. However, as noted above, reclamation of northeastern Walland Marsh, notably the area between the Cheyne Channel and the Rumensea Wall, caused the size of the Romney inlet to fall. It appears that simultaneously the Rye inlet expanded until it came to dominate the back-barrier area of Walland Marsh, west of the large walls (or “Great Cordon” (Eddison and Draper, 1997) that delimit the western margin of the Cheyne Channel. The contraction of the Romney inlet was greatly facilitated by the construction of a set of embankments from the Cheyne Channel to Midley and thence onto the gravel beaches at Lydd. A pronounced west to east fall in ground surface elevation across these embankments, of between c. 1 m to 2 m (Long and Innes, 1995; Spencer, 1997), confirms the relative timing of these changes in inlet dimensions.

Reclamation of the remaining intertidal areas of Walland Marsh continued until at least 1234 AD (Eddison, 1998), but from the middle of the 13th century AD onwards there is growing reference in historical documents to the construction (or strengthening) of sea defences across Walland Marsh to protect land from flooding from the Rye inlet, as opposed to new land claim (Eddison, 1998). This switch to a more defensive mode of land management records the start of a period of renewed flooding of the back-barrier area which was aggravated by the major storms of the middle and later 13th century AD.

Our reconstruction indicates that the 13th century AD storms enlarged a pre-existing early tidal inlet at Rye. An indication of the dimensions of the breach can be estimated from the distribution of peat, which has been eroded from a 4 km wide corridor in the Rye area (Long et al., 2006). A gravel outcrop along the northern shore of the Wainway Channel at Moneypenny Farm (Section 5.2) is derived from material pushed northward during this breach and subsequently redistributed by littoral drift down the Rother estuary and into the Wainway Channel, with the recurves which mark its eastward limit forming as late as c. 500 yrs BP (Section 4.2). However, there is no evidence in the stratigraphy around Rye for further substantial deposits of gravel inland of this inlet (Long et al., 1996). This suggests that the gravel component of the Rye barrier was already of limited volume by the 13th century AD. Elsewhere, the widening of the Rye breach allowed tidal waters to inundate much of Walland Marsh, extending as far east as the “Great Cordon” (Fig. 7). This flooding resulted in a final inundation and rapid burial by collapse of the remaining areas of raised bog on Walland Marsh, as well as extensive areas of peat in the valleys to the west of the study area (Long et al., 2006).

The widening of the Rye inlet allowed the town to develop as one of the most important ports in southern England. However, its prosperity was short-lived since renewed land claim after the thirteenth century storms once again rapidly diminished the back-barrier tidal prism (Gardiner, 2002). Gravel began to accumulate again both sides of the breach from about 1400-1600 AD. By this time the inlet was sufficiently full of sediment (mostly sands and silts) to provide a platform on which the gravel beaches of Rye Harbour could develop (Fig. 7). This marked the final stages in the history of the Rye inlet, which from this point onwards became a narrow tidal channel with only limited anchorage inland.

6. Discussion

The Romney Marsh/Dungeness Foreland depositional complex has displayed a remarkable morphological resilience over the last 4000 years, responding to changes in RSL, sediment supply, storms, and two major breaches associated with widespread back-barrier flooding. What, then, are the factors that contributed to this resilience and what are the implications of our work for a wider understanding of mixed sand and gravel barrier dynamics?

6.1 Sediment supply and transport

An abundant sediment supply has clearly been critical, both from external sources and also from internal reworking. Smaller mixed sand and gravel barriers on paraglacial coasts generally complete a life-cycle in a matter of a few centuries or millennia, their geologically ephemeral existence dictated by a finite supply of sediment released by localised erosion of cliff material (e.g. Orford et al., 1991; Forbes et al., 1995). Once the source of material has been exhausted, barriers switch into a breakdown phase associated with up-drift cannibalisation followed by breaching and structural re-organisation. In contrast to these smaller-scale structures, the large barrier system of Dungeness Foreland has a geomorphological persistence that is quantified in thousands of years.

External sediment supplies to Dungeness Foreland include episodic longshore delivery of sediment from the regional Selsey Bill – Dungeness coastal cell, as well as offshore sources from the floor of the English Channel. The former will have been prevalent during periods of high storm incidence when headland by-passing was possible, and/or when the tidal inlets along the Sussex coast were closed, thereby facilitating the easterly movement of sediment to the Dungeness depocentre (Jennings and Smyth, 1990; Nichols, 1991). The latter is not thought to be a significant contributor during the late Holocene and, as demonstrated by a seismic survey of Rye

Bay, the majority of offshore sediment at or below wave-base is sand and not gravel (Dix et al., 1998).

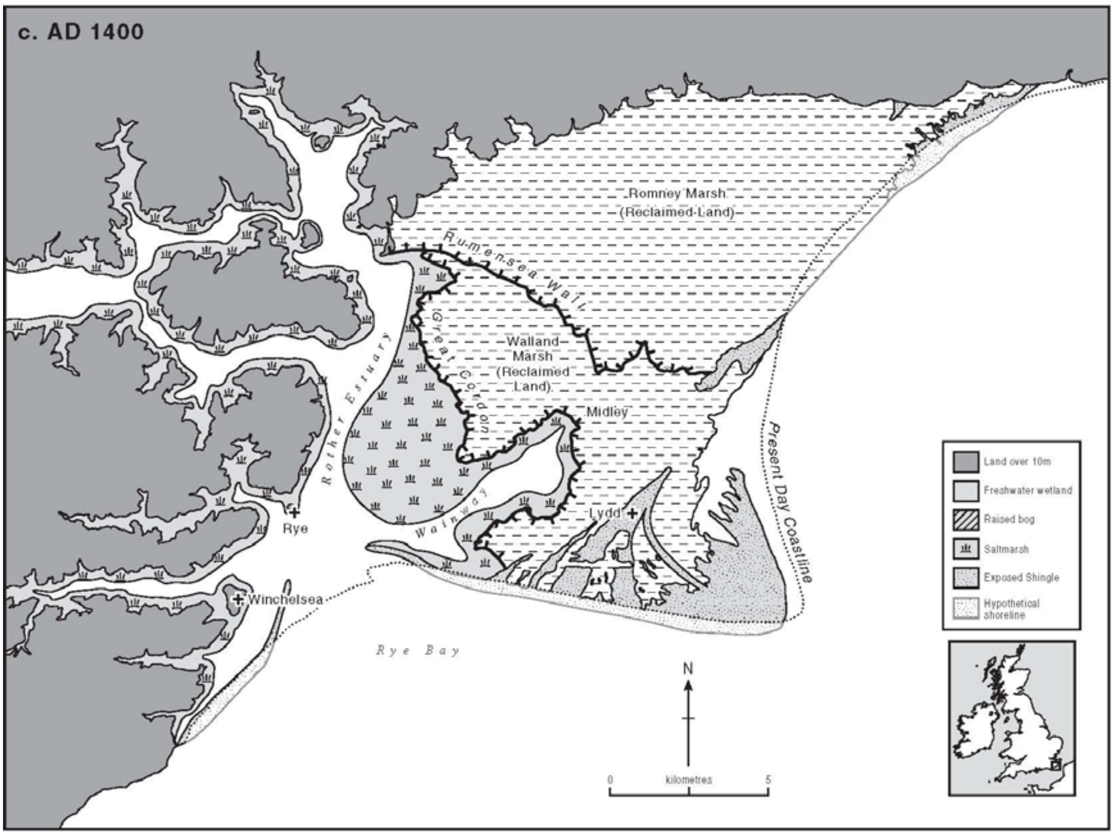

Fig. 7. Palaeogeographical reconstruction of the Romney Marsh/Dungeness Foreland depositional complex when the Rye inlet was dominant (c. AD 1400).

Dungeness Foreland has been periodically nourished by locally available sediment released through cannibalisation of the updrift portions of the barrier, which were presumably once more extensive in Rye Bay than present based on the orientation of extant beaches along the foreland's south coast. This process is similar to that described in other case studies from North America (e.g. Forbes et al., 1995) and Argentina (Isla and Bujalesky, 2000). Drift-aligned barriers are especially sensitive to reductions in sediment supply with up-drift reworking leading to the creation of distinct erosion/accretion cells. Cell fragmentation may eventually lead to breaching and tidal inlet formation, such as occurred at Romney and at Rye. Carter and Orford (1991) note that localised protuberances of sediment may develop at downdrift locations in these cells, and such a process appears to explain some of the abrupt changes in beach ridge orientation and extension observed on the south coast of Dungeness Foreland, east of Galloway's Lookout (Fig. 1). These changes in beach ridge morphology date from around the time of the establishment of the Rye inlet and probably record accumulation of sediment at the downdrift end of a newly formed Rye – Dungeness sediment cell.

The history of Dungeness Foreland is essentially one of fragmentation; an initial single drift-aligned structure that extended across much of the study area developed into a succession of smaller cells characterised by accretion and erosion associated with the opening and closing of different tidal inlets. In this sense, the term "coastal resilience" means little without a clearly defined spatial and temporal framework. At a landform scale, Dungeness Foreland has demonstrated remarkable morphological resilience. But at small spatial scales and shorter temporal intervals, barrier inlet dynamics have been associated with the repeated creation and destruction of individual beaches and complexes of beaches. The landform demonstrates multiple scales of "resilience" that operated simultaneously across the complex in response to different forcing agents of different magnitudes and frequencies. These issues of scale are relevant when considering the application of Orford et al.'s (1991) three phase model of barrier initiation, stability and breakdown to large sand and gravel barriers such as Dungeness. In particular, it is helpful to reflect further on the interactions between sediment supply, sediment transport and nearshore bathymetry and accommodation space.

For drift-aligned structures, multiple modes of barrier behaviour arise from the interplay between sediment supply, nearshore bathymetry, accommodation space and wave climate (e.g. Cooper and Navas, 2004). During the early and mid Holocene, the Rye Bay bathymetry was relatively deep and the area would have been a sediment sink (Dix et al., 1998). However, by the post-Roman period the bay had shallowed and infilled. From this time onwards, the down-drift export of upper-shoreface gravel increased and Rye Bay developed a sediment deficit. This process caused the Rye Bay barrier to erode and made it increasingly vulnerable to overtopping, overwashing and eventual breaching. Once breached, new deep water conditions developed in the breach vicinity and sediment accumulation and the process of barrier healing began. Thus, as the Rye Bay completed a cycle of initiation, stability, breakdown and reformation, so the down-drift part of Dungeness Foreland grew and contracted in harmony.

6.2 Relative sea-level change

The Romney Marsh/Dungeness Foreland depositional complex is located in a region of the northwest Europe that is experiencing gradual, long-term crustal subsidence as a result of on-going glacio-isostatic adjustment following the deglaciation of the British and Fennoscandanavian ice sheets. As a result, the trend in RSL throughout the Holocene has been upwards although, as is apparent from Fig. 2., the rate of RSL has changed quite significantly. Regrettably, the quality of the RSL observations from the study area is not high, due to the contaminating effects of sediment compaction. Nevertheless, the abrupt slow-down in the rate of RSL at c. 4000-5000 cal. yrs BP is a regional feature observed elsewhere in southern England (Waller and Long, 2003) and it is noteworthy that this period of time coincides with the earliest dates that are presented in this paper for the deposition of the gravel beaches across Broomhill Level (Section 4.3 above). Indeed, Jennings and Smyth (1990) have previously argued that this pronounced slow-down in mid Holocene RSL facilitated the onshore transfer of sediment to form gravel barriers along the Sussex coast under a strongly dissipative nearshore regime. The altitude of the Dungeness beach ridges rise through time, from c. +1 m OD on Broomhill Level to c. +6 to +7 m OD on Denge Beach and the current eastern shore of the complex (Fig. 4). Some of this increase may be explained by changes in beach ridge orientation, sediment supply, as well as storminess (Plater and Long, 1995), but based on the trends in Fig. 2, at least 3 m or so of this increase must record the millennial-scale rise in RSL from the mid Holocene to present. These observations indicate that conditions for the development of the foreland probably originated in response to the mid Holocene slow-down in RSL, but that the foreland has continued to grow over millennial timescales, regardless of a continuing upwards trend in long-term RSL.

A more challenging question is to determine whether the reslience of the foreland has been affected over shorter timescales by high frequency variations in RSL and, in particular, variations in storminess. During the late Holocene, our reconstruction identifies two time periods of particular importance to the history of the Romney Marsh/Dungeness Foreland depositional complex; the first around c. 2000 cal. yrs BP and the second at c. 700 AD. Changes in barrier behaviour at these times had far-reaching and inter-connected consequences for the foreland and the back-barrier areas. Although there is little evidence for a period of enhanced storms during the Roman period, studies elsewhere suggest an increase in dune formation (e.g. Tooley, 1990; Orford et al., 2000) during the Medieval Warm Period, when the Rye and Romney inlets developed. However, the latter provide, at best, loose chronological correlatives to the c. 1300 cal. yrs BP breaches at Romney and Rye, and the dating evidence is currently too weak to invoke enhanced storminess as a common cause for their development.

The evidence for tidal inundation of Romney and Walland Marshes during the Roman period is matched elsewhere in the UK and the southern North Sea basin. In the Severn Estuary, for example, coastal flooding is well documented on the Somerset Levels, the Avon Levels, as well as the Gwent Levels (Godwin, 1943; Rippon, 1997). In North Germany, Behre (2004) describes how in the 1st century AD coastal dwellings were abandoned and the remaining inhabitants protected themselves by constructing their houses on raised mounds known as Wurten. Many of the coastal dwelling sites in Lower Saxony were abandoned by the 3rd century AD, probably due to coastal flooding (Mier, 2004). Such a regional trend suggests RSL rise and associated erosion at this time was a likely contributory factor to inundation and abandonment. It is probable that there was a significant anthropogenic component to these inundations, with land claim likely to have significantly increased flood levels.

6.3 Barrier and back-barrier interactions

In addition to abundant sediment supply, barrier resilience is promoted by the close interaction between the barrier and the back-barrier. This is well-illustrated by the repeated ability of the tidal inlets to narrow and then close, effectively healing the original breach via back-barrier sedimentation and ebb- and/or flood-tidal delta development. For example, during the Medieval period, when the Romney and Rye inlets were simultaneously open, the foreland became an island. Back-barrier infilling and tidal inlet closure ensured that the landform reformed into a single structure relatively quickly, probably within a few hundred years. However, only with the Hythe inlet does the process of inlet closure appear to have been an entirely natural process (although Reeves (1995) suggests even here there was a human-dimension);

inlet closure/contraction at Romney and Rye were strongly influenced by reductions in tidal prism due to aggressive reclamation and by artificial maintenance of the tidal Rother as a navigable channel.

6.4 Implications for the management of sand and gravel barriers

These observations on self-organisation have implications for coastal management strategies. Firstly, it is clear that drift-aligned sand and gravel barriers are dynamic landforms at a wide range of spatial and temporal scales. Any attempt to restrict this dynamism ignores the fact that this is an inherent element of the long-term resilience of these landforms. Secondly, the Rye-Dungeness barrier system has a particular response to reduced sediment supply – it breaches. These breaches, at Romney and Rye, interrupted the efficient drift-cell export of sediment which would clearly have been unsustainable in the long-term. Each was followed by re-sealing as cell fragmentation and deeper nearshore waters prompted renewed sediment accumulation. This tendency to breach under periods of sediment stress could potentially be used to the long-term future benefit of the landform.

7. Conclusions

We have proposed a new stratigraphic model for the Dungeness Foreland/Romney Marsh depositional complex based on a detailed review of a wide range of data. An early drift-aligned structure that extended uninterrupted from Fairlight towards Hythe, is envisaged in line with previous reconstructions. This barrier was in place by c. 5000 to 4000 cal. yrs BP. For the next 2000 to 3000 years, the barrier remained a largely stable form, building relatively slowly to the east, increasingly nourished with sediment cannibalised from up-drift sources in Rye Bay.

Marine conditions returned across Romney Marsh proper after c. 2000 cal. yrs BP, penetrating across parts of Walland Marsh. The phases of tidal flooding have not been conclusively dated, but one may correlate with the marine inundation that is recorded in many coastal lowlands in the UK and the southern North Sea basin during the Roman period. Between c. 700-1700 AD, the evolution of Dungeness Foreland was closely linked to the opening and subsequent contraction/closure of tidal inlets at

Romney and Rye. For a period of time, perhaps lasting a few hundred years, we believe that the two inlets operated simultaneously and that Dungeness Foreland became an island. Closure of the Romney inlet was aided by extensive reclamation that reduced the back-barrier tidal prism and probably also diminished the size of the Romney (and Rye) ebb-tidal delta. An initially small inlet at Rye was significantly widened in the 13th century AD by a period of intense storms. There followed a relatively brief interval of renewed flooding across Walland Marsh, and then a more protracted infilling of the main tidal channels – the Wainway, Rother and Brede. Renewed reclamation accelerated tidal prism reduction and inlet closure, with significant infilling occurring after c. 1500 AD by sand and gravel derived from up-drift sources.

Our work highlights that resilient coasts are not necessarily stable coasts. Certainly, with respect to our study area, much of the Dungeness resilience can be attributed to the sheer size of the depositional complex that includes the foreland and the back-barrier marshland. Their co-dependence demonstrates that current management schemes that preclude the possibility of significant cross-barrier sediment and water exchange are at odds with the longer-term dynamic resilience of this landform. Indeed, the large height difference between the barrier and back-barrier areas, and the extensive use of hard defences along Pett Level, at Broomhill, and also Dymchurch (Robinson, 1988), combine to create a coastal landform that has little capacity for internal readjustment in response to future changes in RSL, storms and sediment supply without radical readjustment of its boundary conditions. Thus, the Holocene record demonstrates that the Romney Marsh/Dungeness Foreland depositional complex retains significant potential for coastal resilience but that, as with many managed coastal lowlands in the developed world, this will only be realised if the constraints of ‘hard’ engineered coastal protection measures are loosened and ‘soft’ engineering solutions are considered on spatial and temporal scales more attuned to inherent (and previously demonstrated) resilience characteristics.

Large mixed sand and gravel barriers like Dungeness Foreland are inherently resilient landforms capable of internal recycling of sediment to maintain overall landform integrity. We identify multiple spatial and temporal scales of morphological resilience, often with different elements of the same landform experiencing synchronous phases of erosion, stability or accretion. The repeated development of tidal inlets facilitates cross-barrier exchange of water and sediment, but these inlets also disrupt any tendency toward landform self-destruction that is inevitable in an efficient drift-dominated system that experiences sediment depletion.

Acknowledgements

This work was completed with the support of two grants from English Heritage under the Aggregate Levy Sustainability Fund as well as support from the Romney Marsh Research Trust and the Colyer-Fergusson Charitable Trust. We thank our collaborators in this project for their assistance in the field, the laboratory and in the development of the ideas included in this paper, particularly Helen Roberts, Paul Stupples, Damien Laidler, Ed Schofield, Jonathan Lageard, Peter Wilson and Kate Elmore. We thank Alex Bayliss and John Meadows of the English Heritage Scientific Dating Section for their contribution to the dating programme and discussions of the chronology. This paper benefited from stimulating discussions with Dr Cecile Baeteman, Mrs Jill Eddison and Dr Gillian Draper, although the views expressed here are ours. Two referees made helpful observations that strengthened the paper. Frank Davies, Eddie Million and Neil Tunstall assisted in the laboratory and the diagrams were drawn by David Hume. This paper is a contribution to IGCP Project 495 “Quaternary Land-Ocean Interactions: Driving Mechanisms and Coastal Responses” and to the INQUA Sub-Commission on Coastal Processes and Sea-level Changes.

References

Adamiec, G., Aitken, M., 1998. Dose-rate conversion factors: update. Ancient TL 16, 37-49.

Allen, J.R.L., 1996. The sequence of early land claims on the Walland and Romney Marshes, southern Britain: A preliminary hypothesis and some implications. Proc. Geol. Assoc. 107, 271-280.

Allen, J.R.L., 1999. The Rumensea Wall and the early settled landscape of Romney Marsh (Kent). Landscape History 21, 5-18.

Austin, R.M., 1991. Modelling Holocene tides on the NW European continental shelf. Terra Nova , 276–288.

Barber, L., 1998. An early Romano-British salt-working site at Scotney Court. Arch. Cant. 118, 327-353.

Behre, K-E., 2004. Coastal development, sea-level change and settlement history during the later Holocene in the Clay District of Lower Saxony (Niedersachsen), northern Germany. Quat. Int. 112, 37-53.

Bonnot-Courtois, C., Caline, B., L'Homer, A., Le Vot, M. (Eds.), 2002. La baie du Mont-Saint-Michel et l'estuarire de la Rance. Environments sédimentaires et evolution récente (The Bay of Mon-Saint-Michel and the Rance Estuary: Recent Development and Evolution of the Depositional Environments). Bull. Centre Rech. Elf Exploration. Prod. Mém. 26, 256 pp.

Brooks, N., 1988. Romney Marsh in the Early Middle Ages. In: Eddison, J., Green, C. (Eds.), Romney Marsh: Evolution, Occupation, Reclamation. Oxford University Committee for Archaeology, Oxford, No. 24, pp. 90-104.

Bruun, P., Gerritsen, F., 1959. Natural bypassing of sand at coastal inlets. Waterw. Harb. Div., Proc. ASCE 85, 75-107.

Burrows, A.J., 1884. Romney Marsh, Past and Present. Surv. Inst. Trans. 27, 338.

Carter, R.W.G., Orford, J.D. 1991. The sedimentary organisation and behaviour of drift-aligned barriers. In: Coastal Sediments '91. Amer. Soc. Civ. Eng., New York, pp. 934-948.

Cleary, W.J., FitzGerald, D.M., 2003. Tidal inlet response to natural sedimentation processes and dredging-induced tidal prism changes: Mason Inlet, North Carolina. J. Coast. Res. 38, 1187-1201.

Cooper, J.A.G., Navas, F., 2004. Natural bathymetric change as a control on century-scale shoreline behaviour. Geology 32, 513-316.

Cunliffe, B.W., 1980. The evolution of Romney marsh: a preliminary statement. In: Thompson (Ed.), Archaeology and Coastal Change. Soc. Antiquaries, Occas. Pap., NSL, London, pp. 37-55.

Cunliffe, B.W., 1988. Romney Marsh in the Roman Period. In: Eddison, J., Green, C. (Eds.), Romney Marsh: Evolution, Occupation, Reclamation. Oxford University Committee for Archaeology, Oxford, No. 24, pp. 83-87.

Davis, R.A., Barnard, P., 2003. Morphodynamics of the barrier-inlet system, west-central Florida. Mar. Geol. 200, 77-101.

Dix, J., Long, A.J., Cooke, R., 1998. The evolution of Rye Bay and Dungeness Foreland, the offshore seismic record. In: Eddison, J., Gardiner, M., Long, A.J. (Eds.), Romney Marsh: Environmental Change and Human Occupation in a Coastal Lowland. Oxford University Committee for Archaeology, Oxford, No. 46, pp. 3-12.

Drew, F., 1864. Geology of the Country between Folkstone and Rye Including Romney Marsh. Mem. Geol. Surv. UK, Harpenden.

Eddison, J., 1983. An intensive ditching system in the Wicks south-west of Lydd. Arch. Cant. 139, 273-276.

Eddison, J., 1995. Romney Marsh: The Debatable Ground. Oxford University Committee for Archaeology, Oxford, No. 41.

Eddison, J., 1998. Catastrophic changes: the evolution of the barrier beaches of Rye Bay. In: Eddison, J., Gardiner, M., Long, A.J. (Eds.), Romney Marsh: Environmental Change and Human Occupation in a Coastal Lowland. Oxford University Committee for Archaeology, Oxford No. 46, pp. 65-88.

Eddison, J., 2000. Romney Marsh: Survival on a Frontier. Tempus, Stroud.

Eddison, J., 2002. The purpose, construction and operation of a 13th-century watercourse. In: Long, A.J., Hipkin, S., Clarke, H. (Eds.), Romney Marsh: Coastal and Landscape Change through the Ages. Oxford University Committee for Archaeology, Oxford No. 56, pp. 127-139.

Eddison, J., Draper, G., 1997. A landscape of Medieval reclamation: Walland Marsh, Kent. Landscape History 19, 75-88.

Eddison, J., Gardiner, M., Long, A.J., 1998. Romney Marsh: Environmental Change and Human Occupation in a Coastal Lowland. Oxford University Committee for Archaeology, Oxford No. 46

Eddison, J., Green, C., 1988. Romney Marsh: Evolution, Occupation, Reclamation. Oxford University Committee for Archaeology, Oxford, No. 24.

Elliott, J., 1847. Account of the Dymchurch Wall, which forms the sea defences of Romney Marsh. Minut. Proc. Instn. Civ. Engrs 6, 466-484.

Evans, J.R., Kirby, J.R., Long, A.J., 2001. The litho- and biostratigraphy of a late Holocene tidal channel Romney Marsh, Southern England. Proc. Geol. Ass. 112, 111-130.

FitzGerald, D.M., Buynevich, I.V., Davis, R.A., Fenster, M.S., 2002. New England tidal inlets with special reference to riverine-associated inlet systems. Geomorphology 48, 179-208.

Forbes, D.L., Orford, J.D., Carter, R.W.G., Shaw, J., Jennings, S., 1995. Morphodynamic evolution, self organisation, and instability of coarse clastic barriers on paraglacial coasts. Mar. Geol. 126, 63-86.

Gardiner, M.F., 1998. The exploitation of sea-mammals in Medieval England: bones and their social context. Arch. J. 154,173-95.

Gardiner, M., 2002. The late Medieval “Antediluvian” landscape of Walland Marsh. In: Long, A.J., Hipkin, S., Clarke, H. (Eds.), Romney Marsh: Coastal and Landscape Change Through the Ages. Oxford University Committee for Archaeology, Oxford No. 56, pp. 101-120.

Gardiner, M.F., Cross, R., Macpherson-Grant, N., Riddler, I., 2001. Continental trade and non-urban ports in Mid-Anglo-Saxon England: excavations at Sandtun, West Hythe, Kent. Arch. J. 158, 161-290.

Gardiner, M.F., Stewart, J., Priestley-Bell, G., 1999. Exploitation of whales in Anglo-Saxon England: some evidence from Denge Marsh, Lydd, Kent. Med. Arch. 42, 96-101.

Godwin, H., 1943. Coastal peat beds of the British Isles and North Sea. J. Ecol. 31, 199-247.

Green, C., 1988. Palaeogeography of marine inlets of the Romney Marsh area. In: Eddison, J., Green, C. (Eds.), Romney Marsh: Coastal and Landscape Change Through the Ages. Oxford University Committee for Archaeology, Oxford, No. 24, pp. 167-174.

Green, R.D., 1968. Soils of Romney Marsh. Soil Survey of Great Britain Bulletin 4, Harpenden.

Greensmith, J.T., Gutmanis, J.C., 1990. Aspects of the late Holocene history of the Dungeness area, Kent. Proc. Geol. Assoc. 101, 225-237.

Gulliver, F.P., 1897. Dungeness Foreland. Geogr. J. 9, 536-546.

Hall, D., Coles, J., 1994. Fenland Survey: an essay in landscape and persistence. English Heritage, London.

Harkness, D.D., 1983. The extent of the natural ^{14}C deficiency in the coastal environment of the United Kingdom. J. Europ. Study Grp. Phys., Chem., Math. Tech. Appl. to Arch. PACT 8 (IV.9), 351-364.

Hey, R.W., 1967. Sections in the beach plain of Dungeness. Geol. Mag. 104, 361-370.

Isla, F.I., Bujalesky, G.G., 2000. Cannibalisation of Holocene gravel beach plains, northern Tierra del Fuego, Argentina. Mar. Geol. 170, 105-122.

Jennings, S., Smyth, C., 1990. Holocene evolution of the gravel coastline of East Sussex. *Proc. Geol. Ass.* 101, 213-224.

Jennings, S.C., Orford, J.D., Canti, M., Devoy, R.J.N., Straker, V., 1998. The role of sea-level rise and changing sediment supply on Holocene gravel barrier development; the example of Porlock, Somerset, UK. Holocene 8, 165-181.

Klein, R.J.T., Smit, M.J., Goosen, H., Hulsbergen, C.H., 1998. Resilience and vulnerability: coastal dynamics or Dutch dikes? Geogr. J. 164, 259-268.

Lewin, T., 1862. The invasion of Britain by Julius Caesar with replies to the Astronomer-Royal and of the late Camden professor of Ancient history at Oxford. London.

Lewis, W.V., 1932. The formation of Dungeness Foreland. Geogr. J. 80, 309-324.

Lewis, W.V., Balchin, W.G.V., 1940. Past sea-levels at Dungeness. Geogr. J. 96, 258-285.

Long, A.J., Hughes, P., 1995. Evolution of the Dungeness foreland during the last 4000 years. Mar. Geol. 124, 253-271.

Long, A.J., Innes, J.B., 1993. Holocene sea-level changes and coastal sedimentation in Romney Marsh, southeast England, UK. Proc. Geol. Assoc. 104, 223-237.

Long, A.J., Innes, J.B., 1995. The back-barrier and barrier depositional history of Romney Marsh, Kent, UK. J. Quat. Sci. 10, 267-283.

Long, A. J., Plater, A.J., Waller, M.P., Innes, J.B., 1996. Holocene coastal sedimentation in the Eastern English Channel: new data from the Romney Marsh region, United Kingdom. Mar. Geol. 136, 97-120.

Long, A.J., Hipkin, S., Clarke, H., 2002. Romney Marsh: Coastal and Landscape Change Through the Ages. Oxford University Committee for Archaeology, Oxford, No. 56.

Long, A. J., Waller, M.P., Hughes, P., Spencer, C., 1998. The Holocene depositional history of Romney Marsh proper. In: Eddison, J., Gardiner, M., Long, A.J., (Eds.), Romney Marsh: Environmental Change and Human Occupation in a Coastal Lowland. Oxford University Committee for Archaeology, Oxford, No. 46, pp. 45-63.

Long, A. J., Waller, M.P., Stupples, P., 2006. Driving mechanisms of coastal change: sediment autocompaction and the destruction of late Holocene coastal wetlands. Mar. Geol. 225, 63-84.

Meier, D., 2004. Man and environment in the marsh area of Schleswig-Holstein from Roman until late Medieval times. Quat. Int. 112, 55-69.

Nicholls, R.J., 1991. Holocene evolution of the gravel coastline of East Sussex: discussion. Proc. Geol. Assoc. 102, 301-306.

Nicholls, R.J., Branson, J., 1998. Coastal resilience and planning for an uncertain future: an introduction. Geogr. J. 164, 255-258.

Orford, J.D., Carter R.W.G., Jennings, S., 1991. Coarse clastic barrier environments: evolution and implications for Quaternary sea level interpretation. Quat. Int. 9, 87-104.

Orford, J.D., Wilson, P., Wintle, A.G., Knight, J., Braley, S., 2000. Holocene coastal dune initiation in Northumberland and Norfolk, eastern UK: climate and sea-level changes as possible forcing agents for dune initiation. In: Shennan, I., Andrews, J. (Eds.), Holocene Land-Ocean Interaction and Environmental Change around the North Sea. Geological Society, London, Special Publications, 166, 197-217.

Orford, J.D., Forbes, D.J., Jennings, S., 2002. Organisational controls, typologies and timescales of paraglacial gravel-dominated coastal systems. Geomorphology 48, 51-85.

Plater, A.J., Long, A.J., 1995. The morphology and evolution of Denge Beach and Denge Marsh. In: Eddison, J. (Eds.), Romney Marsh: the Debatable Ground. Oxford University Committee for Archaeology, Oxford, No. 41, pp. 8-36.

Plater, A.J., Stupples, P., Roberts, H., Owen, C., 2002. The evidence for late Holocene foreland progradation and rapid tidal sedimentation from the barrier and marsh sediments of Romney Marsh and Dungeness: a geomorphological perspective. In: Long, A.J., Hipkin, S., Clarke, H. (Eds.), Romney Marsh: Coastal and Landscape Change Through the Ages. Oxford University Committee for Archaeology, Oxford, No. 56, pp. 40-57.

Reeves, A., 1995. Romney Marsh: the field walking evidence. In: Eddison, J. (Eds.), Romney Marsh: the Debatable Ground. Oxford University Committee for Archaeology, Oxford, No. 41, pp. 78-91.

Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C., Blackwell, .G., Buck, C.E., Burr, G., Cutler, K.B., Damon, P.E., Edwards, R.L., Fairbanks, R.G., Friedrich, M., Guilderson, T.P., Hughen, K.A., Kromer, B., McCormac, F.G., Manning, S., Bronk Ramsey, C., Reimer, R.W., Remmele, S., Southon, J.R., Stuiver, M., Talamo, S., Taylor, T.W., van der Plicht, J., Weyhenmeyer, C.E., 2004. IntCal04 Terrestrial radiocarbon age calibration, 26 - 0 ka BP. Radiocarbon 46, 1029-1058.

Rippon, S., 1996. Gwent Levels: the evolution of a wetland landscape. Coun. Brit. Arch. Res. Rep.105, York.

Rippon, S., 1997. The Severn Estuary: landscape evolution and wetland reclamation. Leicester University Press, London.

Rippon, S., 2002. Romney Marsh: evolution of the historic landscape and its wider setting. In: Long, A.J., Hipkin, S., Clarke, H. (Eds.), Romney Marsh: Coastal and Landscape Change Through the Ages. Oxford University Committee for Archaeology, Oxford No. 56, 84-100.

Roberts, H., Plater, A.J., 2005. Optically Stimulated Luminescence (OSL) Dating of Sands Underlying the Gravel beach Ridges of Dungeness and Camber, Southeast England, UK. English Heritage, Centre Archaeol. Rep. 27/2005.

Robinson, G., 1988. Sea defence and land drainage of Romney Marsh. In: Eddison, J., Green, C. (Eds.), Romney Marsh: Evolution, Occupation, Reclamation. Oxford University Committee for Archaeology, Oxford, No. 24, pp. 163-166.

Roy, P.S., Cowell, P.J., Ferland, M.A., Thom, B.G., 1994. Wave-dominated coasts. In: Carter, R.W.G., Woodroffe, C.D. (Eds.), Coastal Evolution, Cambridge University Press, Cambridge.

Schofield, J.E., Waller, M.P., 2005. A pollen analytical record for hemp retting from Dungeness Foreland, UK. J. Arch. Sci. 32, 715-726.

Spencer, C.D., 1997. The Holocene evolution of Romney Marsh: a record of sea-level change. Unpublished Ph.D. thesis, University of Liverpool.

Spencer, C.D., Woodland, W., 2002. Palaeoenvironmental changes during the last 4000 years at Scotney Marsh, Romney Marsh, Kent: a multiproxy approach. In: Long, A.J., Hipkin, S., Clarke, H. (Eds.), Romney Marsh: Coastal and Landscape Change Through the Ages. Oxford University Committee for Archaeology, Oxford, No. 56, pp. 58-74.

Spencer, C.D., Plater, A.J., Long, A.J., 1998a. Holocene barrier estuary evolution: the sedimentary record of the Walland Marsh region. In: Eddison, J., Gardiner, M., Long, A.J. (Eds.), Romney Marsh: Environmental Change and Human Occupation in a Coastal Lowland. Oxford University Committee for Archaeology, Oxford No. 46, pp. 13-29.

Spencer, C.D., Plater, A.J., Long, A.J., 1998b. Rapid coastal change during the mid- to late-Holocene: the record of barrier estuary sedimentation in the Romney Marsh region, south-east England. Holocene 8, 143-163.

Stupples, P., Plater, A.J., Submitted. Controls on the temporal and spatial resolution of tidal signal preservation in late-Holocene tidal rhythmites, Romney Marsh, Southeast England. Int. J. Earth Sci.

Tooley, M.J. 1990. The chronology of coastal dune development in the United Kingdom. Catena Supplement 18, 81-88.

Tooley, M.J., Switsur, V.R., 1988. Water level changes and sedimentation during the Flandrian Age in the Romney Marsh area. In: Eddison, J., Green, C. (Eds.), Romney Marsh: Evolution, Occupation, Reclamation. Oxford University Committee for Archaeology, Oxford, No. 24, pp. 53-71.

Vollans, E., 1988. New Romney and the ‘river of Newenden’ in the later Middle Ages. In: Eddison, J., Green, C. (Eds.), Romney Marsh: Evolution, Occupation, Reclamation. Oxford University Committee for Archaeology, Oxford, No. 24, pp. 128-141.

Vollans, E., 1995. Medieval salt-making and the inning of tidal marshes at Belgar, Lydd. In: Eddison, J. (Eds.), Romney Marsh: the Debatable Ground. Oxford University Committee for Archaeology, Oxford, No. 41, pp. 118-126.

Waller, M.P., Long, A.J., 2003. Holocene coastal evolution and sea-level change on the southern coast of England: a review. J. Quat. Sci. 18, 351-359.

Waller, M.P., Schofield, J.E. 2005. Mid to late Holocene vegetation change in the Weald of south-eastern England: multiple pollen profiles from the Rye Area. Veget. Hist. Archaeobot.

Waller, M.P., Burrin, P.J., Marlow, A., 1988. Flandrian sedimentation and palaeoenvironments in Pett Level, the Brede and Lower Rother valleys and Walland Marsh. In: Eddison, J., Green, C. (Eds.), Romney Marsh: Evolution, Occupation, Reclamation. Oxford University Committee for Archaeology, Oxford, No. 24, pp. 3-30.

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, South-east England. Quat. Sci. Rev. 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. Holocene 16, 51-61.

List of Figures

Fig. 1	Location map showing Dungeness Foreland and Romney/Walland Marshes in the eastern English Channel. The hypothetical shorelines shown are from Lewis and Balchin (1940).
Fig. 2	Age/altitude graph depicting the age and elevation of transgressive and regressive contacts from the Dungeness Foreland/Romney Marsh depositional complex (from Long et al., 2006). All of the data have been lowered from their original elevation by sediment compaction, a process that particularly affects points from the transgressive contact to the thick main marsh peat.
Fig. 3	Simplified lithostratigraphy of Romney and Walland Marshes; Transect 1 (Long and Innes, 1995), Transects 2 and 3 (Long et al., 1998). The location map shows the transect locations.
Fig. 4	Stratigraphic transect along the south coast of the Romney Marsh/Dungeness Foreland depositional complex. Radiocarbon dates are cited in calibrated years before present (AD 1950) (cal. yrs BP) with a two sigma age range. Marine samples are calibrated using the Calib Marine04 dataset with a local marine reservoir effect (ΔR) for the English Channel of -32 ± 56 yrs (Harkness, 1983). OSL dates are cited in years before AD 1950 to enable comparison with the radiocarbon dates.
Fig. 5	Palaeogeographical reconstruction of the Romney Marsh/Dungeness Foreland depositional complex when the Hythe inlet was dominant (c. AD 300-400).
Fig. 6	Palaeogeographical reconstruction of the Romney Marsh/Dungeness Foreland depositional complex when the Romney inlet was dominant (c. AD 700-800).

Fig. 7 Palaeogeographical reconstruction of the Romney Marsh/Dungeness Foreland depositional complex when the Rye inlet was dominant (c. AD 1400).