Coastal Geomorphology of Great Britain

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JOINT NATURE CONSERVATION COMMITTEE

Chapter 6

Gravel and 'shingle' beaches – GCR site reports

INTRODUCTION

V.J. May

The gravel structures and beaches of the British coast are among its best-known and longest-studied geomorphological features. In England, deposits of well-rounded beach gravel are known as 'shingle', and this less geomorphologically precise word is retained for the English and Welsh sites described in the present chapter, where the usage is more common in the literature.

'Shingle' is characterized by grain sizes between 4 and 64 mm (–2 to –6 phi). Many shingle structures are formed predominantly of clasts within this size range, but even the most distinctively sorted such as Dungeness and Chesil Beach contain clasts of many different sizes. Some beaches are characterized by clasts whose long axis exceeds –6 phi and are described as 'cobbles'. The eastern part of Chesil Beach, and the majority of the materials at Budleigh Salterton and Westward Ho! exceed shingle size. Many shingle beaches are, in reality, of mixed clast sizes, with varying quantities of both finer- and coarser-grained materials.

Although most shingle beaches in England and Wales are formed of flint or chert, many include clasts comprising relatively weak materials such as sandstone or chalk or other harder materials; a wide range of clasts formed from harder materials are characteristic of western and northern beaches and in Scotland.

Scolt Head Island, Chesil Beach and Dungeness are three of the most scientifically well-known coastal shingle structures of international renown, but there are many other geomorphologically important types of shingle and gravel features in Britain including small cliff-foot beaches, bay-bars, small recurved spits and beach plains. Some coastal gravel/shingle structures in Britain are at least 5500 years old. Others are relatively young, such as recently formed cliff-foot beaches.

About 1040 km of the British coast is formed by gravel structures (excluding any cliff-foot beaches). But gravel structures also form the base of many sand spit and dune features. If the sand structures with a gravel *base* are added, then the British coast is fringed by about 2900 km of gravel-dominated beaches.

The extent to which gravel beaches are well-sorted affects permeability and the extent to which they behave as reflective or dissipative structures.

Gravel/shingle features have been classified according to their plan-form (e.g. Pethick, 1984; see Figure 6.1), and their profile (e.g. Wright and Short, 1983). Pethick (1984), following many earlier writers, summarized the plan-form of beaches at different scales, ranging from the small rhythmic features such as *cusps* that occur on many beaches, to the very large detached beaches such as spits. Wright and Short (1983), in contrast, focused on the relationship between beach profiles and wave conditions, with a specific emphasis on the differences between a dissipative domain in which beaches display a flat shoaling slope and wide surf zone, with multiple parallel nearshore bars and a reflective domain, characterized by steep beaches (> 6°), with no nearshore bars. Between these extremes there are several intermediate domains that are dominated by longshore bar-troughs, rhythmic bars, transverse (welded) bars or low-tide terraces.

The GCR sites descibed in the present chapter (Figure 6.2 and Table 6.1) contain many short-timescale features, but as major landforms owe their origins to processes acting over considerable periods of time. They include cliff-foot fringing beaches, pocket beaches, bay-bars, cheniers, spits (some with complex recurves), barrier beaches and cuspate forelands.

The simplest gravel and shingle structures are cliff-foot fringing beaches. They fall broadly into two groups:

1. beaches at the foot of cliffs that are the present-day source of most clasts in the beach.
2. cliff-foot beaches where the gravel/shingle is derived from longshore transport. In southern Britain, many shingle beaches are formed from flint that was eroded from the chalk under periglacial conditions and, in northern Britain, under glacial conditions from a range of lithologies, and then transported landwards by the postglacial transgressing seas.

Gravel and shingle beach ridges

The process of migration of a ridge across the sea floor during a marine transgression has often been used to explain the establishment of large linear features such as Slapton bar, Chesil Beach, and Blakeney Point. They may be parts of barrier beaches that have assumed their present form

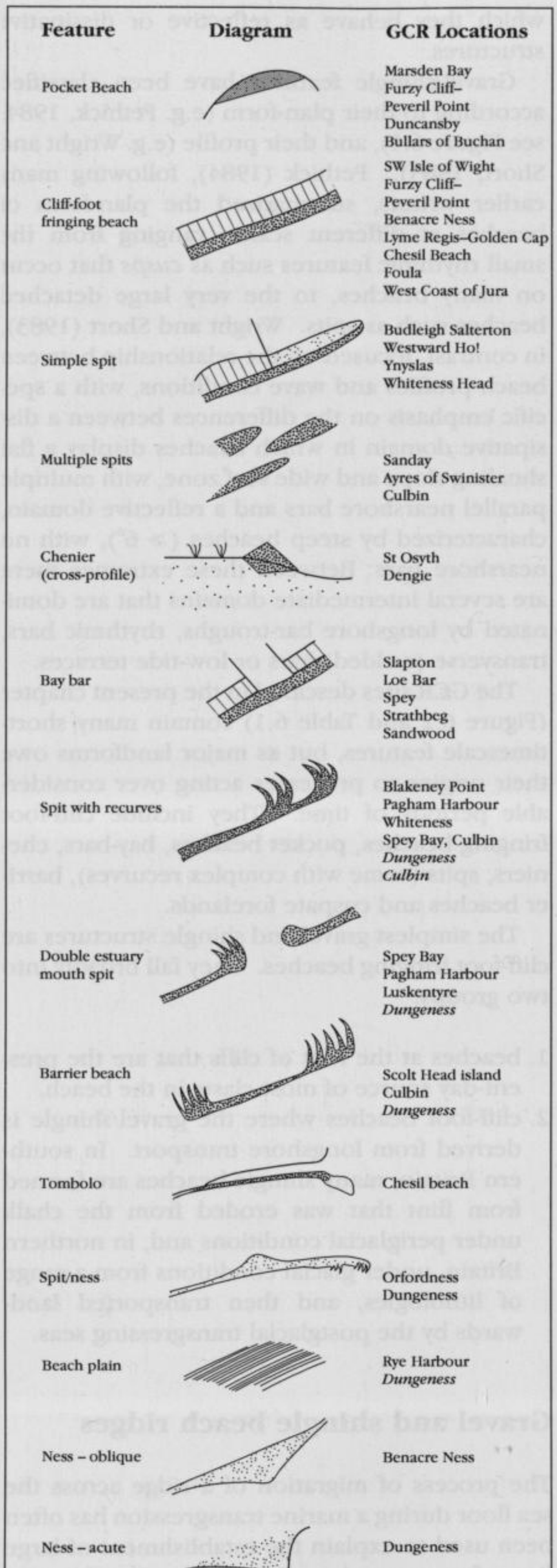


Figure 6.1 Forms and typology of gravel and shingle structures and the GCR sites that represent them. The schematic diagrams show the plan form of the structure concerned. Italic type indicates presence of relict features at a site. In some cases gravel forms the core of the feature, and is now covered in sand.

because they have been built up against the coast. As they transgress the pre-existing landscape, these fringing beaches can become compartmentalized by headlands into smaller embayment beaches, as for example between Osmington and Kimmeridge and in Spey Bay. Chesil Beach is undergoing a similar change between West Bay and Abbotsbury. Slapton bar is attached to headlands at both ends and is in continuity with a cliff-foot fringing beach, whereas Chesil Beach fringes cliffs for much of its western half and then stands separately as a tombolo joining the Isle of Portland to the mainland. In contrast, Blakeney Point is only attached at its eastern end where it continues the line of fringing beach below Sheringham cliffs. It then becomes aligned towards the dominant waves from the north-east. However at its distal end it lies in deeper water, and waves from the north and north-west have been instrumental in developing a series of modern and relict recurves. Gradual elongation of a barrier partially blocking an embayment or the extension of a spit may give rise to a bay-bar.

Larger shingle structures are characteristically built up of series of beach ridges, several hundred in the case of Dungeness, which preserve earlier episodes of beach-construction. Beaches that are initially oriented alongshore as drift-aligned features show a tendency to swash-alignment through time (Davies, 1972). This process involves erosion updrift and the truncation of former ridges, often at a significant angle to the present-day ridge orientation. As spits extend into deeper water, they develop recurves under the influence of different wave directions. Recurves are often grouped, as for example at Scolt Head Island and Blakeney Point, and may be related to pulses of greater sediment flux. Such pulses have been identified at Spey Bay by Gemmell *et al.* (2001a,b) along the main gravel structures but the extent to which these might contribute to recurving at the western end of Spey Bay is less clear.

On some coasts, large cuspate forms develop.

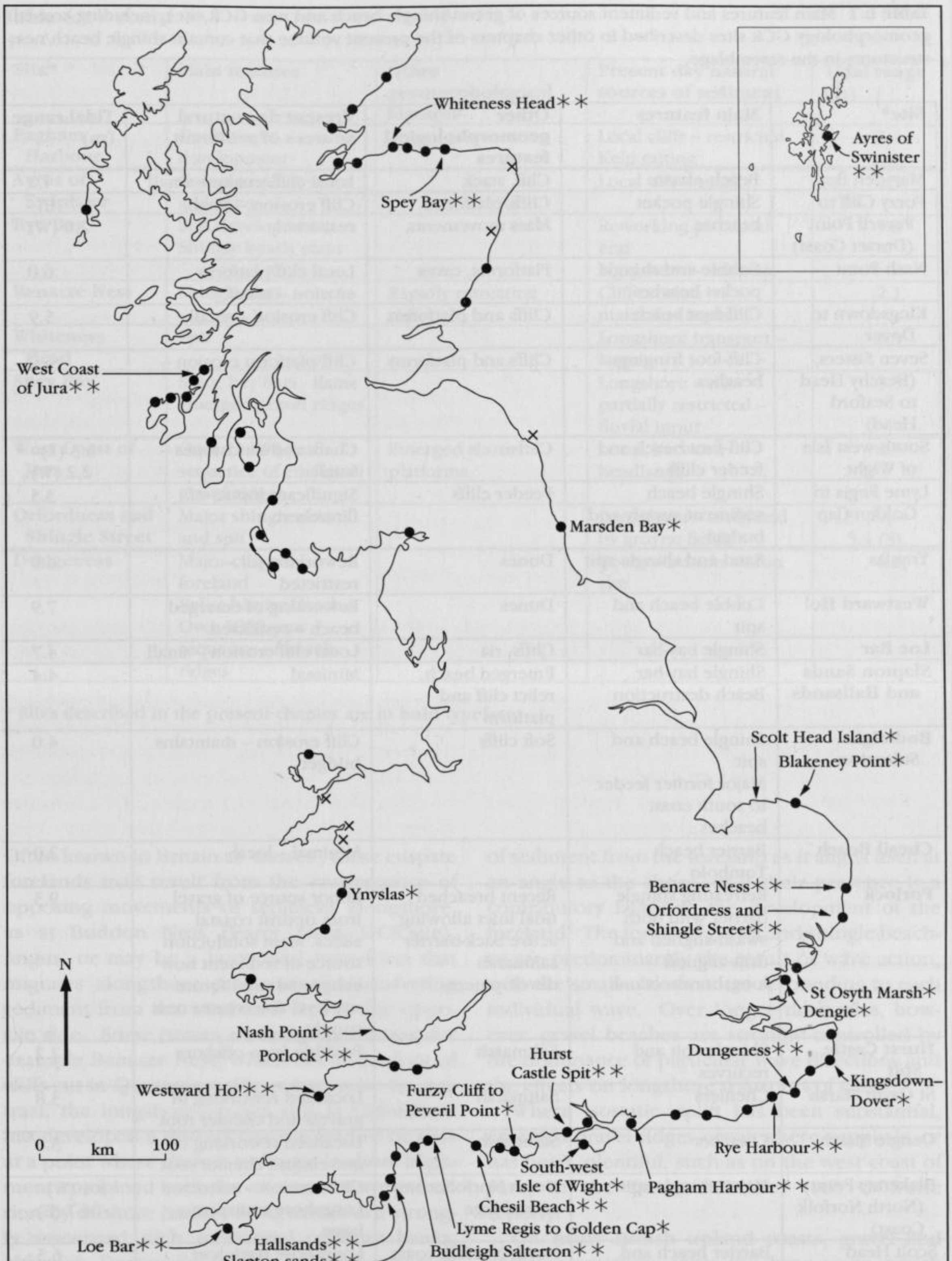


Figure 6.2 Coastal shingle and gravel structures around Britain, showing the location of the sites selected for the GCR specifically for gravel/shingle coast features, and some of the other larger gravel structures.

Table 6.1 Main features and sediment sources of gravel/shingle beach and ness GCR sites, including coastal geomorphology GCR sites described in other chapters of the present volume that contain shingle beach/ness structures in the assemblage.

Site*	Main features	Other geomorphological features	Present day natural sources of sediment	Tidal range (m)
Marsden Bay	Beach phases	Cliff, stack	Local cliff erosion– small	4.2
Furzy Cliff to Peveril Point (Dorset Coast)	Shingle pocket beaches	Cliffs/platforms Mass movements	Cliff erosion – small, restricted	1.7 (E)–2.0 (W)
Nash Point	Cobble and shingle pocket beaches	Platforms, caves	Local cliff/platform erosion – small	6.0
Kingsdown to Dover	Cliff-foot beach	Cliffs and platforms	Cliff erosion – small	5.9
Seven Sisters, (Beachy Head to Seaford Head)	Cliff-foot fringing beaches	Cliffs and platforms	Cliff/platform erosion – small	6.0
South-west Isle of Wight	Cliff-foot beach and feeder cliffs	Cliffs	Chalk and sandstones – small	3.3 (E)–2.2 (W)
Lyme Regis to Golden Cap	Shingle beach sediment supply and budget	Feeder cliffs	Significant inputs of flint/chert	3.5
Ynyslas	Sand and shingle spit	Dunes	Reworking till – restricted	4.0
Westward Ho!	Cobble beach and spit	Dunes	Reworking of emerged beach – restricted	7.9
Loe Bar	Shingle bay-bar	Cliffs, ria	Local cliff erosion – small	4.7
Slapton Sands and Hallsands	Shingle bay-bar Beach destruction	Emerged beach, relict cliff and platform	Minimal	4.4
Budleigh Salterton	Shingle beach and spit Major former feeder to south coast beaches	Soft cliffs	Cliff erosion – maintains budget	4.0
Chesil Beach	Barrier beach Tombolo		Minimal – local	2.0
Porlock	Retreating shingle barrier with both swash-aligned and drift-aligned longshore sections	Recent breached tidal inlet allowing active back-barrier saltmarsh development	Minor source of gravel from updrift coastal slides. Main solifluction source of sediment now exhausted until future sea-level rise creates new supply	9.3
Hurst Castle Spit	Shingle spit and recurves	Saltmarsh	Possible from offshore	2.2
St Osyth Marsh	Cheniers	Saltmarsh	Localized reworking of gravels and chenier root	3.8
Dengie Marsh	Cheniers	Saltmarsh	Localized reworking of gravels and chenier root	3.8
Blakeney Point (North Norfolk Coast)	Major shingle spit	North Norfolk coast assemblage	Cliff erosion – restricted Longshore transport – large	6.4 (W)–4.7 (E)
Scolt Head Island (North Norfolk Coast)	Barrier beach and spits	North Norfolk coast assemblage	Longshore transport – large	6.5

* Sites described in the present chapter are in **bold** typeface

Table 6.1 – *contd*

Site*	Main features	Other geomorphological features	Present day natural sources of sediment	Tidal range (m)
Pagham Harbour	Double spit development		Local cliffs – restricted Kelp rafting	3.4
Ayres of Swinister	Complex of bay bars and spits		Local tills – small	1.5
Rye Bay	Spit developments Shingle beach plain		Reworking proximal end Longshore – minimal	5.8
Benacre Ness	Shingle ness	Rapidly retreating cliffs	Cliff erosion – maintains input	2.1
Whiteness Head	Spit		Longshore transport – large	3.5
Spey Bay	Spits, bay bars, emerged gravel ridges		Longshore – now partially restricted – fluvial input	3.5
West Coast of Jura	Over 11 000 year sequence of emerged gravel ridges	Emerged shore platforms	Local, between headlands	2.5
Orfordness and Shingle Street	Major shingle ness and spit		Longshore – restricted by groyne fields	1.9 (N)–3.4 (S)
Dungeness	Major cuspate foreland Relict barrier beach Over 5000 year sequence of beach ridges		Re-distribution within site	6.2

* Sites described in the present chapter are in **bold** typeface

Often known in Britain as 'nesses', these cuspate forelands may result from the convergence of opposing movements of sediment alongshore, as at Buddon Ness (Barry Links GCR site), Angus, or may be a horizontal wave-form that migrates alongshore progressively transferring sediment from the windward face to the opposite side. Some nesses are fringing features, for example Benacre Ness, which lies at the foot of cliffs cut in Quaternary tills and gravels. In contrast, the longshore-parallel spit at Orfordness has developed a distinct cuspate feature or ness at a point where there is a change in shore alignment combined with the effects of wave refraction by offshore banks. Most nesses are strongly associated with substantial offshore banks, although Dungeness is unusual amongst such features in lacking an associated offshore bank. These banks affect wave refraction, but it is not possible to state unequivocally whether the shoals develop as a result of offshore transport of sediment from the foreland as it aligns itself at an angle to the shore, or if their presence is a contributory factor in the development of the foreland. The forms of gravel and shingle beaches are predominantly the result of wave action, with the small-scale features responding to each individual wave. Over longer timescales, however, gravel beaches are strongly controlled by the dominance of particular wave directions and the effects on longshore transport of clasts.

Where isostatic uplift has been substantial, emerged gravel ridges occur where supply is, or has been, plentiful, such as on the west coast of Jura, in the Inner Hebrides (see GCR site report).

On many British upland coasts, gravel and shingle structures form the base upon which sand spits and dune fields have accreted, but with a few exceptions such as at Culbin, Moray (Comber, 1993) and at Central Sanday (Rennie and Hansom, 2001) (see GCR site reports in chapters 11 and 8) many of these buried gravel structures have not been interpreted. Gravel extraction from these locations may put the sand structures at risk of accelerated erosion, as happened at Spurn Point prior to the 1849 breach (IECS, 1992).

Past management also influences the ability of gravel structures to adjust to sea-level change and storminess, Porlock being a good example of a free-standing gravel structure undergoing post-management adjustment.

The conservation value of gravel and shingle beaches

In spite of reductions in sediment supply, many gravel/shingle beaches remain scientifically important, and worthy of conservation-protection measures so that they can continue to evolve and provide information about the development of coastal gravel systems, coastal form development and the effects of coastal management. It is important that such sites are managed wisely so that the systems can be allowed to develop as naturally as possible. The sites are of high conservation value because

1. internationally, they are among the most well-known coastal features of Britain, especially Chesil Beach, Dungeness and Culbin,
2. they have a distinct flora and support several endemic species of invertebrates,
3. they continue to act as sources of sediment for adjacent beaches,
4. they preserve several millennia of recent coastal deposition and changes in their form reflect variations in wave and wind climates.

In the present chapter the GCR sites (Figure 6.2 and Table 6.1) follow a sequence from shoreline or fringing beaches to the more complex forms of detached beaches.

Gravel and shingle structures as biological SSSIs and Special Areas of Conservation (SACs)

In Chapter 1, it was emphasized that the SSSI site series is constructed both from areas nationally important for wildlife, and GCR sites. An SSSI may be established solely for its geology/geomorphology, or its wildlife/habitat, or it may comprise a 'mosaic' of biological and GCR sites that may be adjacent, partly overlap, or be co-incident. There are a number of coastal SSSIs that are crucially important to the natural heritage of Britain for their wildlife value, but which implicitly contain interesting geomorphological features – such as gravel/shingle structures – that are not included independently in the GCR because of the 'minimum number' criterion of the GCR rationale (see Chapter 1). These sites are not described in the present geomorphologically focused volume.

In addition to being protected through the SSSI system for their national importance, certain types of gravel/shingle habitat are eligible for selection as Special Areas of Conservation (SACs; see Chapter 1) under the 'Habitats Directive'. The principal Annex I SAC coastal gravel/shingle habitat present in the UK is 'Perennial vegetation of stony banks', but on gravel/shingle beaches commonly fringing this habitat, the more transient 'Annual vegetation of drift lines' also occurs.

Coastal gravel/shingle SAC site selection rationale

The Habitats Directive Annex I habitat type most relevant to the present chapter is 'Perennial vegetation of stony banks'. Ecological variation in this habitat type depends on stability, the amount of fine material accumulating between clasts, climatic conditions, width of the foreshore, and past management of the site. The ridges and lows formed in gravel/shingle structures also influence the vegetation patterns, resulting in characteristic zonations of vegetated and bare gravel/shingle. The presence of the yellow horned-poppy *Glaucium flavum* and the rare sea-kale *Crambe maritima* and sea pea *Lathyrus japonicus*, all species that can tolerate periodic movement, is significant. In more stable areas above this zone, where sea spray is blown over the gravel/shingle, plant communities with a high frequency of salt-tolerant species such as thrift *Armeria maritima* and sea campion *Silene uniflora* occur. These may exist in a matrix with abundant lichens.

On the largest and most stable structures the sequence of vegetation includes scrub, notably broom *Cytisus scoparius* and blackthorn *Prunus spinosa*. Heath vegetation with heather *Calluna vulgaris* and/or crowberry *Empetrum nigrum* occurs on stable structures, particularly in the

Table 6.2 Candidate and possible Special Areas of Conservation in Great Britain supporting Habitats Directive Annex I habitat 'Perennial vegetation of stony banks' and/or 'Annual vegetation of drift lines' as qualifying European features. Non-significant occurrences of these habitats on SACs selected for other features are not included. (Source: JNCC International Designations Database, July 2002.)

SAC name	Local authority	Gravel/ shingle habitat extent (ha)
Bae Cemlyn/ Cemlyn Bay	Ynys Môn/ Isle of Anglesey	1.3
Chesil Beach and the Fleet	Dorset	96.2
Culbin Bar	Highland; Moray	122.5
Dee Estuary/ Aber Dyfrdwy*	Cheshire; Fflint/ Flintshire; Wirral	1
Dungeness	East Sussex; Kent	2266.1
Isle of Portland to Studland Cliffs	Dorset	1.4
Lower River Spey–Spey Bay	Moray	65.2
Minsmere to Walberswick Heaths and Marshes	Suffolk	8.8
Morecambe Bay	Cumbria; Lancashire	57.5
North Norfolk Coast	Norfolk	98.4
North Uist Machair	Western Isles / Na h-Eileanan an Iar	3
Orfordness–Shingle Street	Suffolk	553.3
Sidmouth to West Bay	Devon; Dorset	4.4
Solent Maritime	City of Portsmouth; City of Southampton; Hampshire; Isle of Wight; West Sussex	226.5
Solway Firth	Cumbria; Dumfries and Galloway	8
South Uist Machair	Western Isles / Na h-Eileanan an Iar	†

* Possible SAC not yet submitted to EC
† Feature is minor component of SAC
Bold type indicates a coastal geomorphology GCR interest within the site

north. This sequence of plant communities is also influenced by natural cycles of degeneration and regeneration of the shrub vegetation that occurs on some of the oldest ridges.

Vegetated stony banks are scarce. There are only a few large sites in Europe, and the UK hosts a significant part of the European resource of this habitat. Although there are only some 4000 ha of stable or semi-stable vegetated gravel/shingle around the whole of the coast of the UK, the habitat is widely distributed and also exhibits a wide range of variation. The selection of sites reflects the UK's special responsibility for conservation of this habitat type and aims to cover the geographical range and variation of the habitat type. All the largest examples with good conservation of structure and function have been selected, together with additional smaller sites to complete the coverage of range. Site selection has also favoured gravel/shingle structures that support vegetation sequences ranging from pioneer communities to heath and scrub. The selected sites represent a substantial proportion of the European resource.

The vegetation that colonizes drift lines of gravel/shingle at or above mean high-water spring tides is dominated by annual plants. The types of deposits involved are generally at the lower end of the clast-size range (2–200 mm diameter), with varying amounts of sand interspersed in the gravel/shingle matrix. These deposits occur as fringing beaches that are subject to periodic displacement or overtopping by high tides and storms. The distinctive vegetation, which may form only sparse cover, is therefore ephemeral and composed of annual or short-lived perennial species. At most sites where it occurs, the habitat is naturally species-poor, and there is a limited range of ecological variation. Many gravel/shingle beaches are too dynamic to sustain drift-line vegetation. Many of the fringing beaches supporting drift-line vegetation are small, and annual vegetation may exist in one location in one year but not another.

Therefore, although widespread around the UK, sites where this Annex I type is persistent are rare, and even the largest sites probably support less than 10 ha of this habitat. Sites have been selected to reflect the more constant occurrences of drift-line vegetation, normally found in association with larger, more stable areas of gravel/shingle structures. The selected sites represent the majority of the more persistent examples of this habitat type in the UK. They all exhibit good conservation of structure and function (i.e. they are relatively unmodified and are less prone to human disturbance) and represent the range of variation in substrate type and physical structure.

Table 6.2 lists coastal shingle SACs, and indicates which of these sites are also (at least in part) important as part of the GCR and are described in the present chapter.

WESTWARD HO! COBBLE BEACH, DEVON (SS 440 310)

V.J. May

Introduction

The cobble ridge at Westward Ho! extends northwards across the mouth of the Taw–Torridge estuary (see Figure 6.2 for general location). The major sand dunes of Braunton Burrows lie to the north. Westward Ho! ridge forms a major, classic, coastal landform, more by reason of its sediment size than of its form. Few spits are formed by large cobbles at the back of an extensive sandy intertidal zone. The cobble ridge has retreated steadily during the last 100 years (Figure 6.3) and is now much modified at its landward end by artificial gabions. There has also been some artificial beach feeding since the 1980s. Prior to the engineering works, some of the cobble material reached the beach from natural sources to the west where erosion of rock cliffs, platforms and Pleistocene pebble deposits provided a source for much of the material. Dunes form much of the northern part of the spit, known as 'Northam Burrows', but sand gravel and cobbles move towards the distal end, forming a spatulate feature within the estuary of the rivers Taw and Torridge. Both the unusual beach and the nature of its retreat have been the focus of scientific study (Spearing, 1884; Rogers, 1908; Kenyon-Bell, 1948; Stuart and Hookway, 1954; Slade, 1962; Keene, 1986, 1989, 1992, 1996). The processes and forms of the Taw–Torridge estuary have been reviewed by Comber *et al.*, (1993).

Description

The ridge described by the early writers extended from the cliffs at Westward Ho! itself to its distal end, but retreat of the spit and the threat that this posed to the village led to coast protection works (Keene, 1986, 1996) which have transformed the southern proximal part of the cobble ridge into an artificial structure. This section has been excluded from the GCR site boundary. Nevertheless, the four main elements of the original feature are still represented within the site (see Figure 7.10 for local geomorphology):

1. A single ridge of cobbles about 25 m wide, WNW-facing, which is transgressing onto blown sand overlying blue clay. The transgressional beach rests on this base at about high-water springs, but cobbles form a frontal slope to just above high-water neaps (Figure 6.3). Historically, this ridge has been associated with shoreline erosion at its landward end.
2. An area of former cobble ridges that lie at an acute angle to the present-day shoreline and represent earlier phases in its growth. They underlie much of the blown sand of Northam Burrows.
3. A spatulate distal end, known as 'Grey Sand Hill', which extends into the Taw–Torridge estuary.
4. A wide intertidal area that includes areas of sand and cobbles, particularly towards the estuary.

The tidal range at spring tides is 7.9 m. The maximum is from the Atlantic Ocean, but most waves approach the beach from slightly north of west, towards which the main ridge is aligned.

Interpretation

The origin of the ridge is not known, although Keene (1996) regards it as a comparatively recent feature. Hall (1879) described a peat and blue clay deposit about 400 m seawards of the cobble ridge and Rogers (1908) recorded remnants of former forest, a kitchen midden and a submerged pebble ridge in the intertidal area.

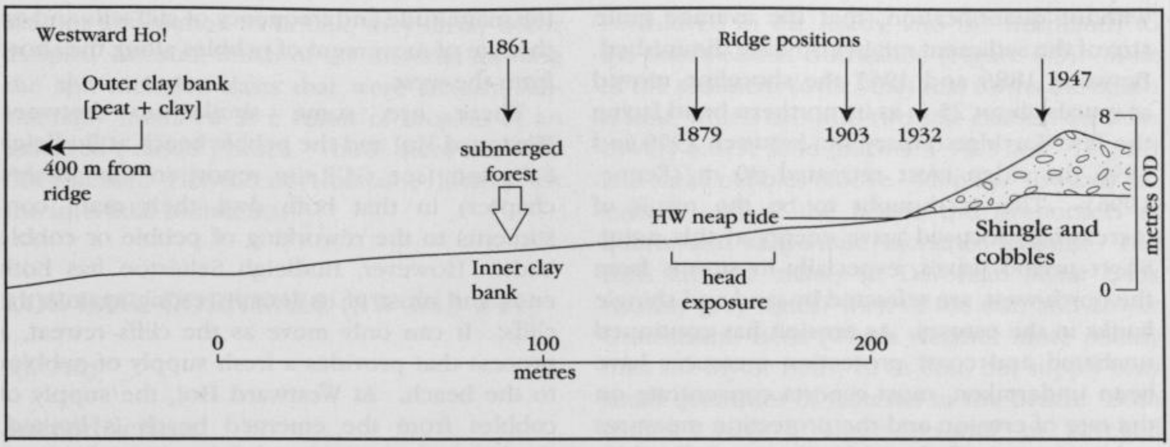


Figure 6.3 Historical recession of position of beach crests at Westward Ho! (Based on Campbell and Bowen, 1989 and Keene, 1996.)

The peat contained leaves, seeds and fruits of iris, oak *Quercus*, hazel *Corylus*, alder *Alnus*, elder *Sambucus*, sea aster *Aster tripolium*, common orache *Atriplex patula*, blackberry *Rubus*, dogwood *Cornus* and lesser spearwort *Ranunculus flammula*. Shells of oyster *Ostrea* and limpets *Patella* were also recorded. The blue clays were reported to include flint flakes (thought to be Neolithic in age), small pebbles and angular fragments of Carboniferous rocks. The mud-snail *Hydrobia ulvae* was common and bones were said to include red deer *Cervus elephas*, and Celtic shorthorn cattle *Bos longifrons*. Since most of the sediment disappeared before modern interpretation and dating techniques were available, the records have to be taken as a possible indication that the cobble ridge lay seawards of the position of the submerged forest and that the land behind it was well colonized by vegetation. The similarity with the development of the cobble ridges at Clarach and Ynyslas on the Welsh coast (see Figure 8.15) may indicate a much earlier age (about 4000 years BP) for Westward Ho!. In contrast to summer depths of sand on the western beach of up to 1.2 m, erosion during the winter of 1983–1984 exposed a thick band of head, suggesting that a wide apron of periglacial debris may have extended some distance seawards of the present-day shoreline (Keene, 1996) and this may provide an alternative explanation for the location of the submerged forest. Erosion of such deposits would provide the source for a transgressing postglacial beach. Samples from the top of the peat bed were radiocarbon dated to 6585 ± 120 year BP (Q-672; Churchill and Wymer, 1985) and 4995 ± 105 years BP (Kidson, 1977). The Holocene sequence is described in detail by Campbell (1998) following description of inner and outer peats by Balaam *et al*. (1987) which indicates that the outer peat was inundated by marine/estuarine conditions about 5200 years BP.

Steers (1946a) noted that the burrows at Northam were first recorded by Ridon in 1630. Though limited, this evidence indicates a predecessor to the present feature farther seaward than the 19th century ridge. Campbell (1998) emphasizes that Westward Ho! shows clear evidence for a transition from a lower Late Devensian sea level, followed by an intially rapid rise of the Holocene sea that swamped a coastal forest dating from about 6000 years BP. With a slowing of sea level rise, the present coastal configuration was established.

Stuart and Hookway (1954) cited Risdon's 17th century survey of Devon as the earliest reference to a ridge at Westward Ho!. The first useful cartographic evidence is dated 1861 when the ridge was 1.8 km long: in the following 100 years the cobble ridge migrated landwards some 152 m (Figure 6.3). The northern part of the ridge grew from four sub-ridges in 1884 to twelve in 1952. Accretion at the northern end was accompanied not only by retreat at the southern end but also a reduction in overall volume. Up to 1884, it is reported (Spearing, 1884) to have stood just over 2 m above the Burrows and about 6.5 m above the level of the sand beach. It was about 48 m wide. By 1954, it was about 25 m wide and stood a similar height above the Burrows. Steers (1964a) suggested, without quantification, that the average grain size of the sediment might also have diminished. Between 1886 and 1947 the shoreline moved seawards about 25 m at its northern bend facing the Taw–Torridge estuary, but between 1959 and 1996 the ridge crest retreated 30 m (Keene, 1996). This is thought to be the result of increasingly focused wave energy at this point. Short period waves, especially in storms from the north-west, are refracted by sand and shingle banks in the estuary. As erosion has continued unabated and coast protection measures have been undertaken, most reports concentrate on the rate of erosion and the protection measures to be adopted (Slade, 1962; Kenyon-Bell, 1948; Stuart and Hookway, 1954; Halcrow, 1980; Comber *et al*., 1993). The beach and intertidal changes and sediment budget estimates have been reviewed by Comber *et al*., (1993), who report a net loss of shingle and cobbles on the eroding part of the ridge of 1500–5000 $m^3 a^{-1}$. The net annual loss between 1886 (when beach volume was estimated at 300 000 m^3) and 1974 has been estimated at 1200 $m^3 a^{-1}$, and between 1981 and 1986 the local authority transported about 15 000 $m^3 a^{-1}$ of material from the distal end of the spit to re-inforce its proximal end. This has since reduced to 5000 $m^3 a^{-1}$ to offset erosional loss. Excepting Keene (1986), little attention is paid in the literature to the sources and origin of the cobble ridge, but there may be some similarities to the history of the Ynyslas ridge (see GCR site report in Chapter 8).

There appear to have been three main natural sources for the cobbles:

1. The emerged ('raised') beach and solifluction apron to the west of Westward Ho!, from which pebbles and cobbles have been and still are eroded and are subsequently moved by longshore transport to the modern beach.
2. offshore cobble ridges such as that described by Rogers (1908) and comparable intertidal cobble fields, from which waves would transport material landwards.
3. Materials derived directly from the erosion of the cliffs to the west. Much of the material forming the cobbles in both the emerged and modern beaches is derived from the Culm Measures.

The intermittent growth of the ridge is attributed by Keene to fluctuations in the supply of pebbles from the west, depending upon both the magnitude and frequency of cliff-falls and on the rate of movement of pebbles along the shore from the west.

There are some similarities between Westward Ho! and the pebble beach at Budleigh Salterton (see GCR site report in the present chapter) in that both owe their main constituents to the reworking of pebble or cobble beds. However, Budleigh Salterton has both ends and most of its length resting against the cliffs. It can only move as the cliffs retreat, a process that provides a fresh supply of pebbles to the beach. At Westward Ho!, the supply of cobbles from the emerged beach is limited. Furthermore, as the beach has migrated inland, both the distance and the area of foreshore over which the material can travel have changed. Cobbles could be moved across the intertidal area just as effectively as alongshore in the storm beach. Those within the storm beach are likely to move inland during periods of washover or to migrate alongshore towards the distal end. There is little evidence to suggest that any return towards Westward Ho!

In northern Britain, many sand spits and beaches rest upon shingle and cobble ridges. At Westward Ho!, however, insufficient sand has been transported from the intertidal zone to build up the dunes on the proximal part of the spit, which rests on clay and head above a rock platform. Beaches formed of large shingle and cobbles are found at several locations along the western coasts of England and Wales (for example, the beaches at Newgale and Llanrhystyd – see Orford, 1977). There are none that form a narrow spit such as that at Westward Ho! The presence of the spit affects the alignment of the Taw–Torridge estuary and this affects the distal development of the sand beach at Braunton Burrows (see GCR site report in Chapter 7). Carter (1988) suggests that such beaches originate as a result of either marine reworking of glacial deposits by the rising Holocene sea or primary sedimentation from adjacent eroding cliffs or seabed. At Westward Ho! the latter appears to be most important, but since the source includes clasts from an emerged beach, that deposit may be attributable to either of the origins suggested by Carter.

Conclusions

Westward Ho! is a rare and excellent example of a narrow cobble spit. Although cobble beaches are not uncommon in Britain, they rarely occur as spits. Second, much of the material forming the spit includes clasts that were already substantially modified as a result of erosion of an emerged ('raised') beach. Third, there is a well-documented Pleistocene–Holocene history for the intertidal sediments.

LOE BAR, CORNWALL (SW 643 241)

V.J. May

Introduction

Loe Bar lies about 4 km SSW of Helston, Cornwall (see Figure 6.2 for general location). It encloses a lagoon occupying part of a former ria and forms an integral part of a beach system extending from Porthleven (SW 627 254) to Gunwalloe (SW 653 223; Figure 6.4). The site is important to coastal geomorphology on two counts. First, Loe Bar is a rare example (in England and Wales) of a bay–bar, and second, it is a key member of a suite of major beaches formed and maintained by predominantly south-westerly wave regimes. The beach is formed of rounded, fine shingle and coarse sand predominantly comprising flint or chert for which there is no local source on land. Present-day inputs of sediment from the adjacent cliffs are small, and overall the beach is in deficit. The bar itself is washed over during periods of high wave-energy and a series of washover fans occurs behind the bar. The periodic breaching of the bar has been described in the literature (Ward, 1922; Toy, 1934; Steers, 1946a; Goudie and Gardner, 1985), but the origin of the bar has attracted less attention (Toy, 1934; Goudie and Gardner, 1985).

Description

The site extends for 4.3 km south-eastwards from Porthleven, but the Loe Bar itself forms only 400 m of this beach. The site is divided into three parts: the cliffs and beach known as 'Porthleven Sands', the bar itself, and the cliffs and beach known as 'Gunwalloe Fishing Cove Beach'.

The beach throughout the site forms a single sediment cell that can neither gain nor lose sediment from alongshore because of confinement by the harbour arm and rocky headland at Porthleven in the north, and the headlands to the south-east at Gunwalloe (Figure 6.5). Most of the sediment (over 90%) that forms the beach is flint, which falls into two size classes: medium to very coarse sand (between +0.5 and –1.0 phi) and small pebbles (above –4.0 phi). Towards the eastern end of the beach, the proportion of quartz and serpentine increases slightly. The cliffs are cut mainly in Devonian Mylor Beds (mainly grey slates) west of the Bar, and Lower Gramscatho Beds (which weather more readily than the Mylor Beds) to its east, but supply only small quantities of material to the beach. Both clifflines have a 'slope-over-wall' or bevelled form, the lower 'wall' being cut both into the very resistant schist and quartz that crop out at the foot of the cliff, and the much less resistant solifluction material that forms much of the cliff either side of the Bar, but especially towards Porthleven. There the local erosion has been retarded by the construction of a wall and gabions. Gabions have also been used to protect the foot of the cliff below Bar Lodge just to the west of the bar.

The Bar was described by Leland during the 16th century as having a tendency to be regularly breached by the River Cober. Borlase described the bar in 1758 as forced up against the mouth of the valley by south-west winds. From time to time, the people of Helston excavated a channel through the bar in order to reduce the risk of flooding from the lake. The cost for this activity, according to Borlase, was three halfpence! The last known artificial breaching of the bar took place in the winter of 1867–1868 after which an overflow adit was cut at the north-western end of the bar. The bar has moved inland since the mid-19th century and may have increased in height (according to Goudie and Gardner, 1985, although no details are given). Washover features can be seen on its landward side, the result of overtopping. However, because of its steep front and gentle upper and back slopes, such overtopping occurs at the extreme of the swash phase and so only limited amounts of sediment are transferred over the crest. The crest itself is only a few centimetres higher than the Loe Pool behind it. The level of water in the Loe Pool does not appear to be affected by tides and so may be perched in a similar way to that at Slapton (see GCR site report, this chapter). The orientation of the beach, and the limited amount of sediment in the beach and the bar are such that sediment

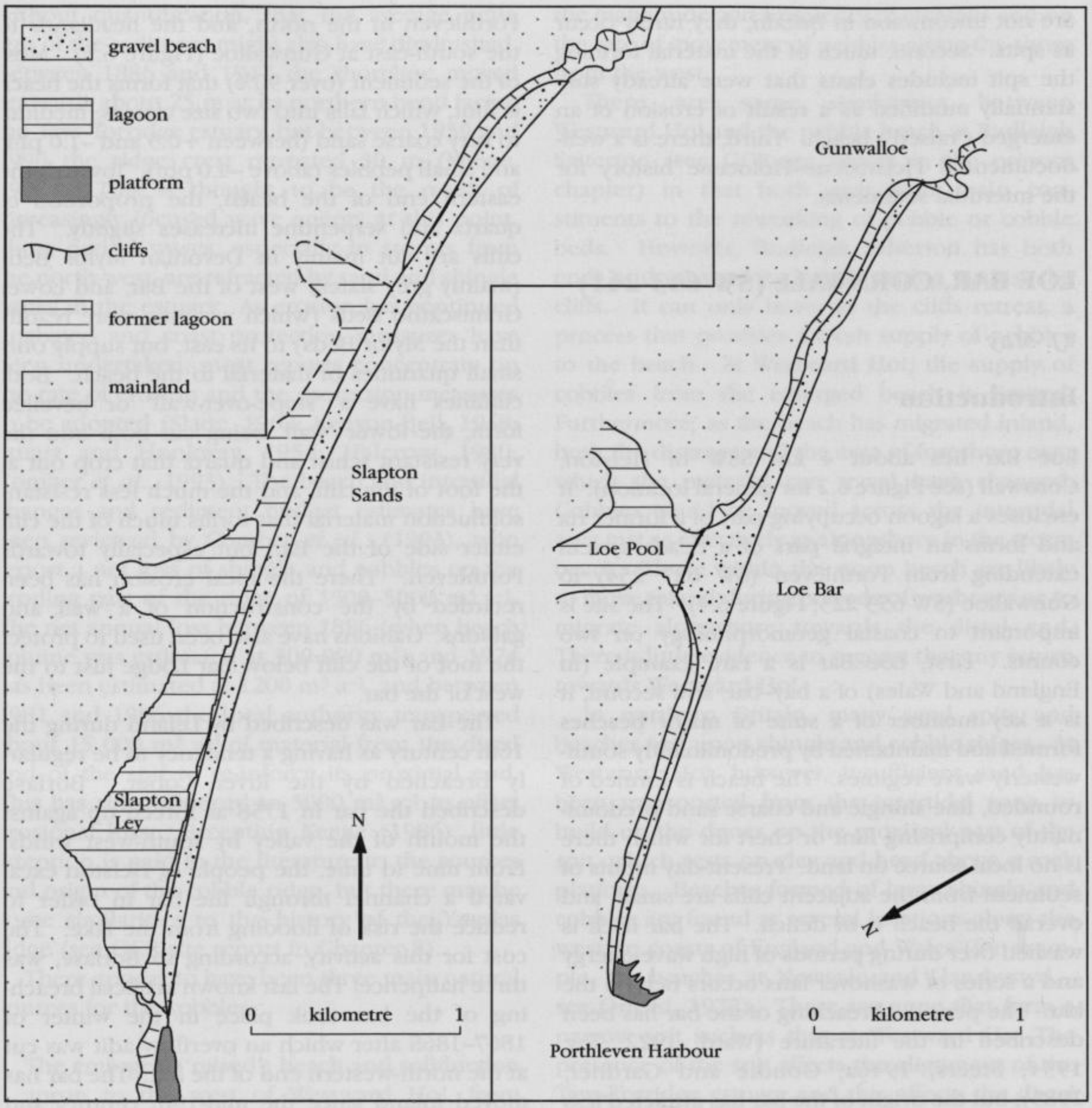


Figure 6.4 Comparison of geomorphological form between Slapton Sands and Loe Bar. Slapton Sands encloses a large lagoon, part of which has been infilled by sediment and become a brackish wetland. At Loe Bar, a cliff-foot beach confined between headlands has blocked off a narrow estuary.

tends to move towards the centre of the bay at present. This process depletes the north-western and south-eastern parts of the bay, but maintains the sand and shingle supply to the bar. The accumulation of shingle blocking the Pool appears to be considerable, for Rogers (1859) describes a boring made in 1834 that reached 22 m (just over 9 m below low-water mark) without reaching bedrock.

Interpretation

Although the Loe Bar has frequently been cited as a textbook example of the comparatively rare British case of a bar completely blocking a lake (e.g. Wooldridge and Morgan, 1937; Steers, 1946a; Monkhouse, 1965; Barnes, 1977), its formation has never been satisfactorily explained (Steers, 1981). Curiously, it is quoted more often than Slapton Ley, which has similar

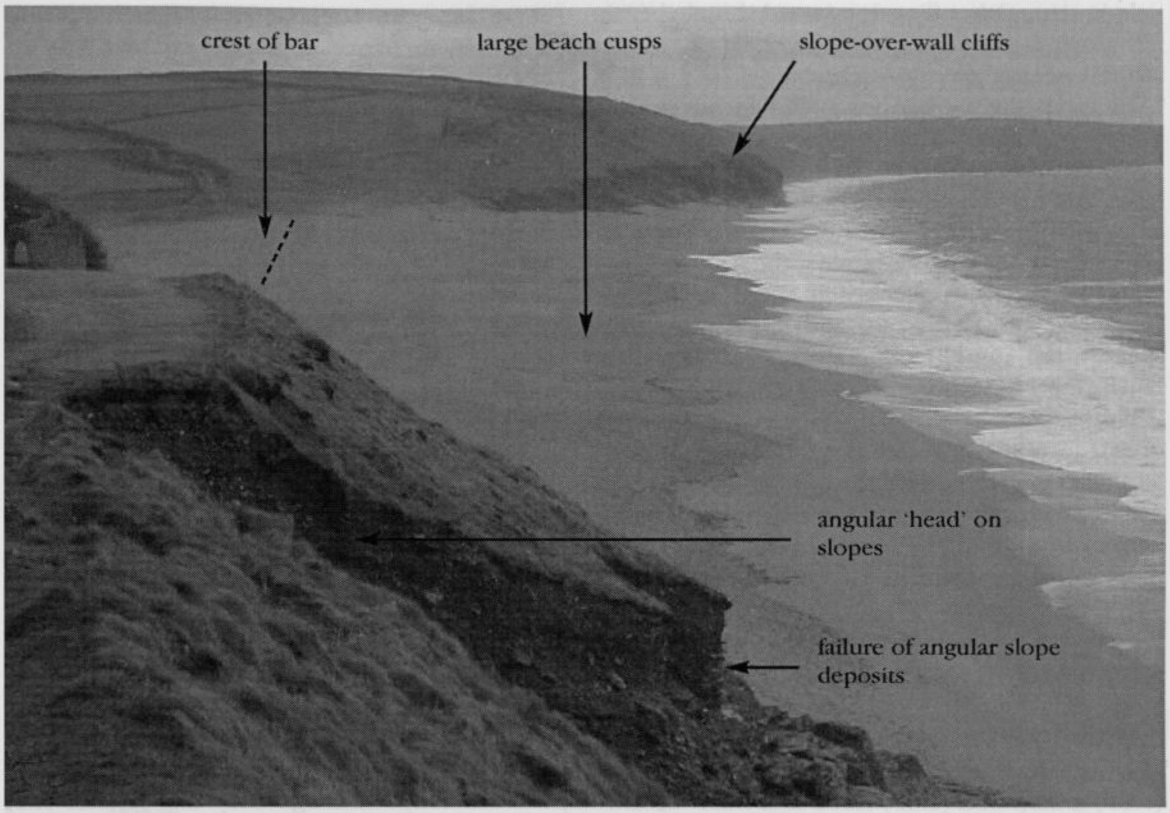


Figure 6.5 Loe Bar, looking approximately south-east, showing the bar and its washover features. (Photo: V.J. May.)

characteristics (Figure 6.4). King (1972a) regarded it as exemplifying the closure of an estuary that only occurs where discharge is very low or there is a very permeable barrier. The latter suggests general seepage through the bar, but this does not appear to be the case. Toy (1934) suggested that two spits had developed from east and west across the valley. In this model, tidal action would move sand and shingle eastwards from Porthleven towards the bar, but strong south-easterly winds would bring about a reverse movement from the east. Eddies set up by the flow of the river and tidal scour would keep the channel clear, but would encourage the growth of spits on opposite sides of the valley mouth. The bar would be completed when the gap was closed by either gradual processes, or severe waves, perhaps during storms. The so-called 'tidal wave' reported by Toy appears to have been an example of particularly large swell which pushed large quantities of sand and shingle up and over the bar. Although the occurrence of tsunamis in the North Atlantic Ocean has been considered as a possible mechanism for the formation of some beaches, there is no evidence for the influence of tsunamis here. The height of the bar is such that it would only be possible for very large run-up to carry material to the top of the bar. Steers (1946a) was convinced that the development of the bar was more likely to have resulted from wave action than from tidal action. He noted later (1981) that its position relative to wave action is comparable with Chesil Beach, i.e. 'it is only the relatively infrequent, largest waves that can carry material on to the top of the bar to build it up or even to move its crest inland.

Explaining the presence of a high proportion of flint in the beach is a problem as indeed it is at Slapton Sands. Bird and Schwartz (1985, p. 364) described the Loe Bar as a beach of flint shingle 'washed up from the sea floor (there being no flint sources in the adjacent cliffs or hinterland areas)'. King (1972a, p. 519) described the Loe Bar as made up of 'unusual material', intermediate between sand and shingle in size. Other flint-dominated shingle beaches along the southern coast of England are also found isolated from present-day sources of flint. The most likely origin for such beaches is their development as barrier beaches that were gradually moved onshore as Holocene sea level rose, the flint coming from drowned terraces of the former river flowing down the English Channel. During its movement across the sea floor in front of the cliffs the beach could have been breached and re-formed, and might even have been more spit-like. The outline of the bay is such that longshore movements could carry shingle towards the bar, and maintaining its form. Thus the beach tends to maintain an equilibrium form with the dominant and prevailing waves that means a more or less smooth curve throughout its length. The bar itself is thus a sediment sink as far as the overall beach budget is concerned. It remains unclear why the beach has such a strongly bimodal size distribution.

Conclusions

Loe Bar is a rare and excellent British example of a shingle bay–bar, and it is one that has international renown as a type example of this form. The site has additional interest because the bar blocks a ria, and the sediments in the adjacent cliffs suggest that the beach occupies an interglacial embayment. The beach is in overall deficit, but the bar is likely to survive as long as shingle remains within the main beach to maintain it. Breaching may occur from time to time, although that has not been recorded since the 1830s. The continued existence of the bar is dependent upon the maintenance of sediment supply from the wider area. As sediment is transported and, in effect, trapped in the bar, it can be replaced by sediment moving in from the main beach. As a result, the stability of the bar depends on natural and unimpeded sediment supply and transport from the wider beach system. In turn the ecological interest of the Loe Pool depends upon the unimpeded natural evolution of the geomorphological processes operating at the site.

Comparable bay–bars with a lagoon occur at Slapton and at the western end of Chesil Beach. At one time the beach at Pagham Harbour also formed a bay–bar with a lagoon, but this was breached and insufficient shingle has been supplied to it for the bar to redevelop. The differences between Loe Bar and Pagham Harbour are, first, their oceanographic context, Loe Bar being directly exposed to south-westerly Atlantic swell, whereas Pagham Harbour is affected by refracted swell, and second the tendency for the Loe Bar to rebuild after breaching, from both natural and anthropogenic causes. The evolution of Loe Bar most likely relates to migration of barrier beaches across the present-day seabed to more-or-less their present position with changes in the present-day beach mainly associated with overtopping. It is also evident that the outline of much of this coastline pre-dates present-day sea level and the Loe Bar is probably reoccupying an interglacial embayment.

SLAPTON SANDS AND HALLSANDS, DEVON

V.J. May

Both Slapton Sands and Hallsands lie on the east-facing coastline of Start Bay, south of the Dart estuary and Torbay (see Figure 6.2 for general location). Though they are distinct sites both in location and in their geomorphological features (Figure 6.6), their relationship to the past and present-day processes is intrinsically linked. This common introduction describes their setting and the debate about their origins. Site-specific details are contained in the individual site descriptions that follow.

Start Bay was described by Hails (1975a–c) as an asymmetrical embayment, about 60 km^2 in area, similar in form to many bays elsewhere for which the term 'zeta-curved bay' is frequently used. The coastline transects Lower Devonian rocks, which strike east–west and form an intermittent series of cliffs. Between Hallsands and Pilchard Cove a sequence of barrier beaches forms the main feature. These beaches are composed predominantly of granules and small pebbles (i.e. between –1 and –4 phi, with flint a dominant grain type (Gleason *et al.*, 1975). There are four main beaches, Hallsands, Beesands, Slapton Sands and Blackpool Sands, all formed of shingle coarsening to cobbles in the south. Beesands and Slapton include barrier beaches, between 100 m and 140 m wide at high tide, which impound lagoons. The crest of the barriers is generally at about 6.0 m OD ± 0.5 m. The ridges become narrower and lower towards the south, in the same direction as

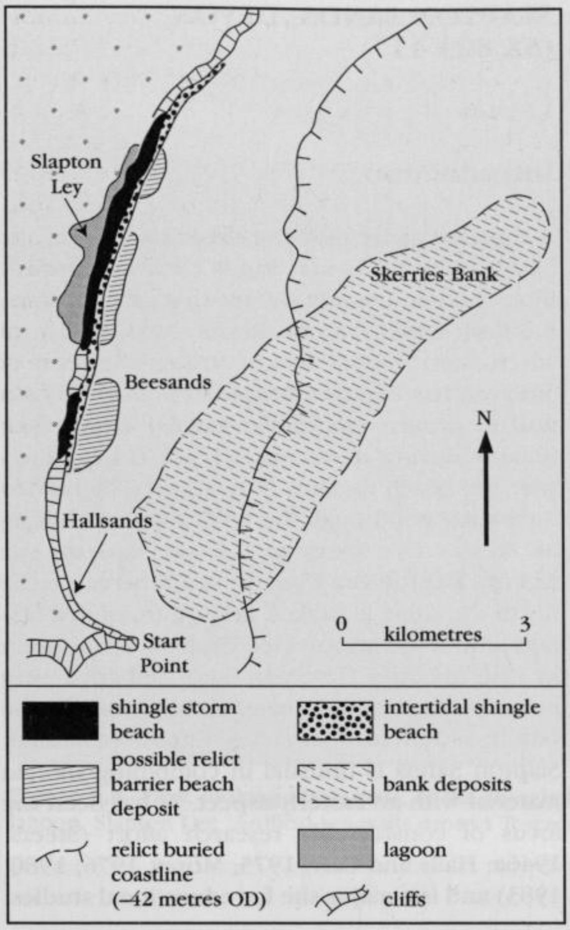


Figure 6.6 Hallsands and Slapton Sands represent parts of a once-continuous gravel beach. Offshore, there is evidence of buried shorelines and a possible former barrier beach. The present-day shingle beach is separated by rock headlands. (After Hails, 1975a.)

sediments become coarser. Worth (1904) and Hails (1975a–c) also recognized a sharp decline in sediment size offshore immediately away from the coast. Although the offshore slope varies near the beaches, depths of –14.5 m OD are encountered by 600 m offshore. There are short lengths of coast protection works at Beesands, Torcross and Blackpool Sands, but these may have little effect on the overall sediment budget of the beaches. In contrast, the commercial removal of about 640 000 tonnes of gravel at Hallsands between April 1897 and December 1902 has had a much greater impact (Worth, 1907).

Kelland and Hails (1972) have described in some detail the submerged form of Start Bay and its offshore deposits. Between 0.5 km and 3 km offshore, there is a pronounced break of slope in the bedrock slope that forms the seabed at an average depth of –42 m OD. This has been interpreted as an ancient coastline that may have been exposed during periods of lower sea level during the Pleistocene Epoch. Opposite the central part of Slapton Ley, there is a second break in the bedrock slope at about 28 m depth. An offshore bank, known as 'Skerries Bank', extends north-eastwards for over 6.5 km and appears to be linked to Start Point. Its maximum height is –4.8 m OD, but for most of its length the crest lies between –7.5 m to –9.0 m OD. Waves from all directions, except north-east, are liable to break on the bank, and it is responsible for significant refraction of waves inshore (Figure 6.7).

Kelland and Hails (1972) identified in Start Bay three lithological sub-environments as a result of grain-size analysis: 'barrier', 'bay' and 'bank' deposits. The barrier deposits consist mainly of gravels that occupy a narrow zone from the front of the barrier beaches to about 200 m offshore, except near Torcross Point and Limpet Rocks where they extend 500 m seawards. Flint and chert (40%) and quartz (46%) comprise most of these sediments, other materials including mica-schist, slate and shale, as well as rhyolite, felsite, granite, and quartz porphyry. The bay deposits are mainly medium- to fine-grained sands, with varying proportions of silt, and whole and comminuted shells.

Hails suggested that the Skerries Bank occupied its present-day position during the later part of the Holocene transgression. During the past 5000 years or so ephemeral barriers were probably constructed, destroyed and submerged and only rather limited amounts of gravel were transported landwards across the floor of Start Bay. Today, no new material is entering the bay, either alongshore or from offshore. As a result, the Bay was considered by Hails (1975a) to be a closed system under present-day conditions. Hails' analysis showed that both Hallsands and Beesands are located at points in the bay where wave energy is focused during north-easterly storms (Figure 6.7). This, combined with high spring-tides and five years of large-scale extraction of gravel, can probably be blamed for the 1917 Hallsands disaster (see GCR site report, below).

Both Hallsands and Slapton Sands thus have origins that depend to a substantial degree on the wave regimes within the bay, the effects of

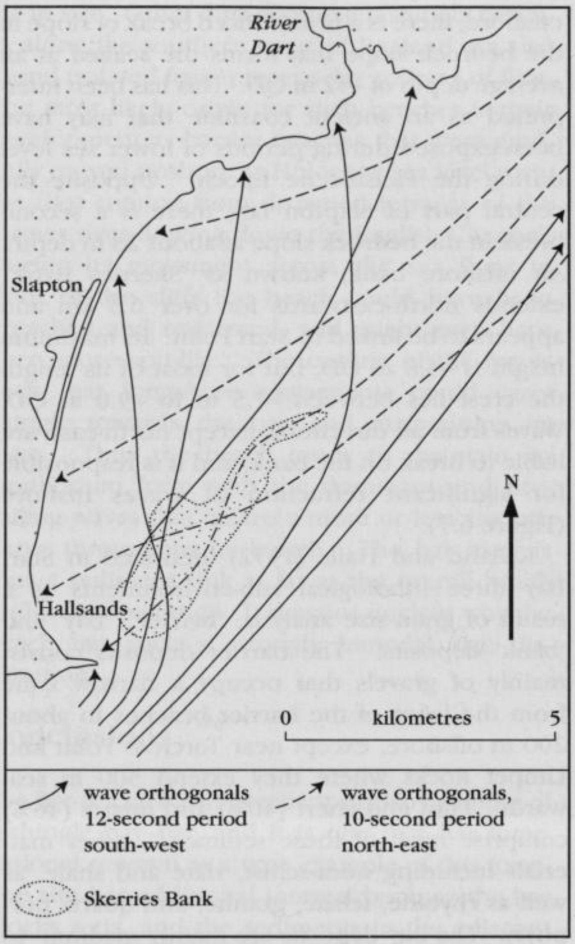


Figure 6.7 Wave refraction along the coast between Slapton Ley and Hallsands. The Skerries Bank affects waves entering Start Bay from the south-west. Wave energy is concentrated in locations such as Hallsands and Beesands during north-easterly winds. (After Hails, 1975a–c)

the Skerries Bank and the transgression of a flint–quartz barrier beach which now lacks external sources. The interest of the two sites arises from their 'closed' nature, the effects of gravel extraction on the coastal sediment budget and coastal erosion, and the development of barrier beaches.

The beaches form a single beach system, but owing to the modification of the coast at Beesands, itself separating a small lagoon from the sea, by the dumping of coastal protection material, this area was excluded from the GCR site boundary. Nonetheless, the continuity of the system is evinced by the consistent reduction in the mean sediment size from south to north.

SLAPTON SANDS, DEVON (SX 823 417)

V.J. May

Introduction

A shingle barrier beach enclosing a lagoon, the beach ridge at Slapton Sands comprises mainly flint, chert and quartz shingle that extends some 5.6 km from Limpet Rocks, just south of Torcross, to Shiphill Rock at Strete. The beach at Torcross has been artificially strengthened by a wall to protect the hamlet against wave attack notably during north-easterly gales, but otherwise the beach remains little affected by human intervention although the A379 road runs along its crest. The southern 2.2 km separate the lagoon, Slapton Ley, from the sea, whereas to the north the ridge is backed first by an infilled former arm of the lagoon (see Figure 6.4) and then by cliffs of Lower Devonian slates and grits. Very little locally derived material is found in the beach sediments. In the English Channel, Slapton Sands is unusual in combining shingle material with an easterly aspect. It has been the focus of considerable research effort (Steers, 1946a; Hails and Carr, 1975; Morey, 1976, 1980, 1983) and is a major site for educational studies.

Description

The barrier beach is predominantly shingle despite the name 'Slapton Sands', and encloses a freshwater lagoon, Slapton Ley. Divided into two parts, the Higher Ley to the north, and the Lower Ley to the south, the lagoon occupies a former marine embayment bounded by degraded cliffs. The Higher Ley is mainly covered by reeds *Phragmites* whereas the Lower Ley is open water. It is up to 500 m wide and has a maximum depth of –4.0 m OD. The beach varies from 100 m to 140 m in width at high tide with a crest that is generally at about +6.0 m OD ± 0.5 m. The gradient of the intertidal beach persists below low-tide level to about –7.5 m OD. It then slopes gently to reach –14.5 m OD at about 600 m offshore. There are barrier beach deposits for about 500 m offshore, according to Kelland (1975), which exceed 5 m in thickness.

Within the lagoon, there are extensive sheets of washover gravels derived from the barrier (Morey, 1983) except in the central part known

Figure 6.8 View, looking north-west, of the shingle barrier beach of Slapton Sands, enclosing the freshwater lagoon, Slapton Ley. Artifical sea-walls protect Torcross in the foreground. (Photo: V. J. May.)

as 'Ireland Bay'. This is the widest part of the lagoon, and is the mouth of the Start Stream. Morey described the following sedimentary sequence in the lagoon:

1. Light-grey, silty, estuarine muds that may be an extension of the lower bay deposits (Hails, 1975a). The fauna suggests a salinity gradient with restricted water circulation behind a growing barrier or spit. The tidal entrance was not located by Morey (1983), but he suggested that it was probably in the southern part of the Lower Ley.
2. A thin brown organic silt with a sharp lower boundary, but passing upwards into fen peats. The presence of pollen of *Chenopodium* and reeds *Phragmites* has been interpreted by Morey as suggesting a local transition from vestigial saltmarsh to reed swamp.
3. Fen peats about 1.3 m thick. An early reed swamp stage gradually changes to a sedge (*Carex*)-dominated fen community. The top of the peat has been dated at 1813 ± 40 years BP and the base at 2889 ± 50 years BP.
4. A layer of muddy sand, thickening seawards.
5. The upper layer is formed by lacustrine muds of terriginous detrital origin.

The present-day lake is perched presumably on its own sediments (Morey, 1976), despite the permeability of the shingle barrier.

Interpretation

Although Ward (1922) provided an early description of the site, Steers (1946a) was emphatic that there was no very conclusive hypothesis for the origin of the beach. The sedimentary characteristics of Start Bay have been described above, but the development of Slapton Sands and Slapton Ley is based mainly upon Morey's papers (1976, 1980, 1983). During the early Holocene, a transgressional shoreline of saltmarshes, estuaries and ephemeral lagoons developed in a macrotidal environment within the shelter of the Skerries Bank. It is thought unlikely that major barriers developed until the shoreline was close to its present-day position about 5000 years BP.

The rate of transgression declined partly as a result of a reduction in the rate of eustatic sea-level rise and the proximity of the coast to a relict pre-Holocene cliffline. Major accumulations of gravel can only occur where bedrock is below modern sea level and where overwashing can spread gravels across submerged infilled Holocene valleys. Without a substantial eustatic rise, there would have been insufficient space for gravels to accumulate except at Beesands and Slapton Sands.

The site was also the location for some detailed observations of the relationships between nearshore sediment dynamics and nearshore motions of the water itself (Huntley and Bowen, 1975a). They observed edge waves with a longshore wavelength of 32 m and with a period twice that of the incident waves. Swash interaction in the narrow surf zone on this steep beach was proposed as the process generating these waves. Nearshore circulation cells were also observed here. On steep beaches (such as this one and other GCR sites, for example, Hurst Castle Spit, Loe Bar, Porth Neigwl), the short-period edge waves that are observed may be responsible for small-scale topographical features such as beach cusps (Huntley and Bowen, 1973, 1975b).

Conclusions

Slapton Sands is one of the few examples of a bay–bar in Britain. The barrier beach, which encloses a freshwater lagoon, Slapton Ley, is predominantly shingle despite the name 'Slapton Sands'. The locality has been investigated not only as geomorphologically interesting in its own right, but also as part of larger-scale studies of Start Bay as a whole. As a result of the degree of study of the site, a great deal more is known about its dynamics than in many other sites. It contrasts strongly in location with the other major example of this type of landform, Loe Bar, in that it is sheltered from the main Atlantic wave systems. It demonstrates better than many other localities around the British coast the links between seabed features and the shoreline landforms, both in their Holocene history and their effects upon modern-day wave behaviour. Together with Loe Bar, the two sites demonstrate clearly how similar coastal landforms may develop in different wave conditions.

HALLSANDS, DEVON (SX 819 382)

V.J. May

Introduction

Hallsands lies at the south-western end of Start Bay (Figure 6.6); the scientific interest in this site arises from:

1. its location at a point where wave energy is focused at the shoreline by offshore banks,
2. the catastrophic destruction of Hallsands village, and
3. the formerly buried cliff forms that were exhumed by removal of gravel and shingle during storms in January 1917.

The cliffs are cut in mica-schist and quartz-schist, an emerged ('raised') platform of which provided the foundation for part of the former hamlet of Hallsands, other parts having been built on the shingle beach itself. The site is regarded by many coastal scientists as being a classic and vivid exemplar of the dangers of beach sediment extraction, as well as having intrinsic geomorphological interest in its landforms. Work by Hails and Carr (1975) has shown, however, that the concentration of wave energy on this part of the coastline by the offshore Skerries Bank during north-easterly gales was of primary importance in bringing about the rapid localized erosion at Hallsands (as well as at Beesands and Torcross to the north). The destruction of Hallsands village has provided a stimulus for research that has led in turn to a substantially better understanding of both this site and the wider geomorphological history and hydrodynamics of Start Bay and its coastline.

Hallsands is unusual among British coastal erosion sites in being very well documented and internationally renowned (e.g. Komar, 1976; Bird, 1984, 1985; Holmes, 1965). Its renown results from the much-reported destruction of the cliff-foot hamlet of Hallsands in January 1917. Detailed surveys were carried out between 1903 and 1923 by Worth (1904, 1909, 1923), during 1956–1957 by Robinson (1961) and in the 1970s by Hails and Carr (1975). There are few erosional sites in the UK that have been studied and reported in such detail. The wider physical links with Slapton Sands and the question of the anthropogenic origins of the Hallsands disaster have ensured that it has

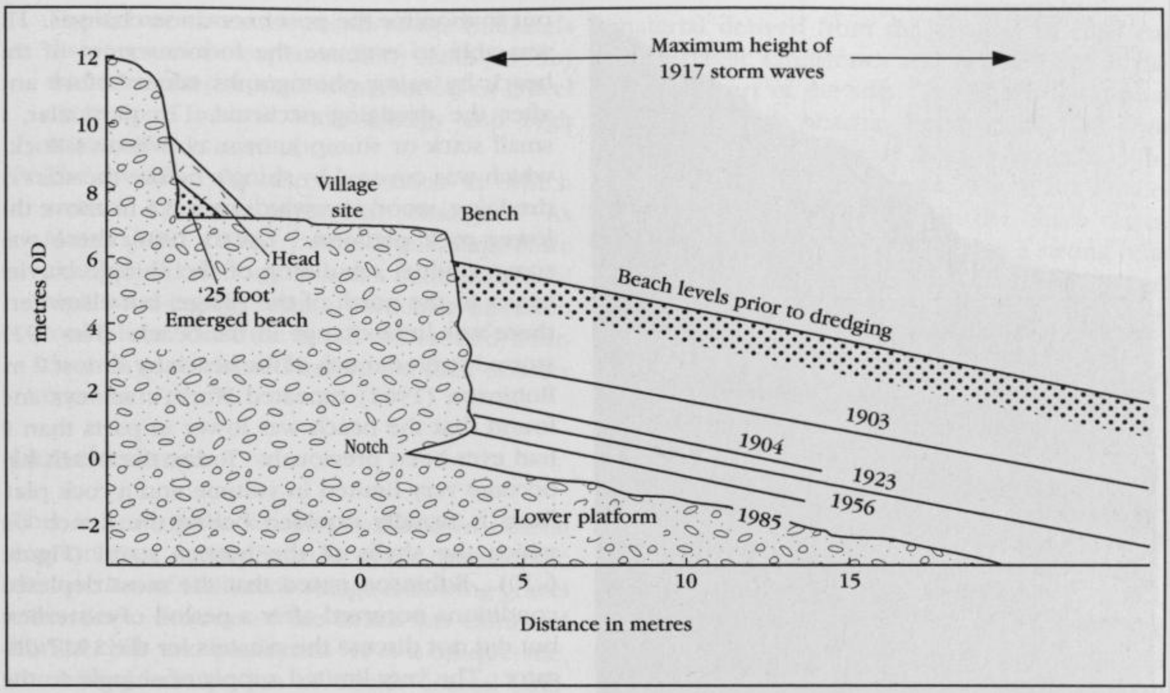


Figure 6.9 Cross-section of beach at Hallsands, showing the historic beach levels prior to dredging. (After Mottershead, 1986.)

retained its interest as a site of international importance.

Description

The site is about 500 m in length, and is part of a cliffed shoreline between Start Point and Hallsands. Although the cliffs have a slope-over-wall form, the lower wall is distinguished at locations such as Hallsands by a bench about 7 m in height above sea level (Figure 6.9). Photographs in Mottershead (1986) show that the village stood on the rock bench, about 1–2 m above the shingle beach, contrary to the impression given by the beach profiles in Worth (1904), Robinson (1961) and Goudie and Gardiner (1985) that the beach and platform were a continuous unbroken form. Robinson (1961) described the rock bench as a wave-cut bench overlain by a considerable thickness of head. The bench is discontinuous, with promontories separated by deep ravines. At two locations, emerged beach deposits with bands of rounded pebbles occur beneath the head. The steeply sloping cliffs behind the bench have been modified by solifluction since their formation.

Mottershead (1986) described a notch at the top of the lower platform that may be as much as 2 m deep and certainly pre-dates the extraction of shingle, which was removed at the end of the 19th century. The notch is at about Ordnance Datum on the promontories, but rises to +1 m OD within the ravines. These ravines can be traced in places to the base of deep gashes containing rotten rock in the upper high cliffs. Freshly fallen debris, large boulders, small stones and clay often fill the upper end of the ravines. Mottershead suggested that the ravines represent the former location of deeply weathered rock now eroded by wave action.

Worth (1904, 1909, 1923) surveyed the beach at Hallsands immediately after the cessation of commercial gravel extraction and subsequently after the disaster of 1917. In 1897, a local contractor, Sir John Jackson, was licensed by the Board of Trade to remove sand and other materials from the beach of Start Bay at Hallsands and Beesands. Up to 1600 tonnes was removed daily for the extension of the Royal Dockyard at Devonport. Worth estimated that 395×10^3 m^3 was removed before the licences were withdrawn in 1902, when the beach level had fallen by at least 3 m. During the winter of 1900–1901 storms undermined sea-walls and removed sand and shingle from the rock ravines behind. Buildings situated at these points collapsed and

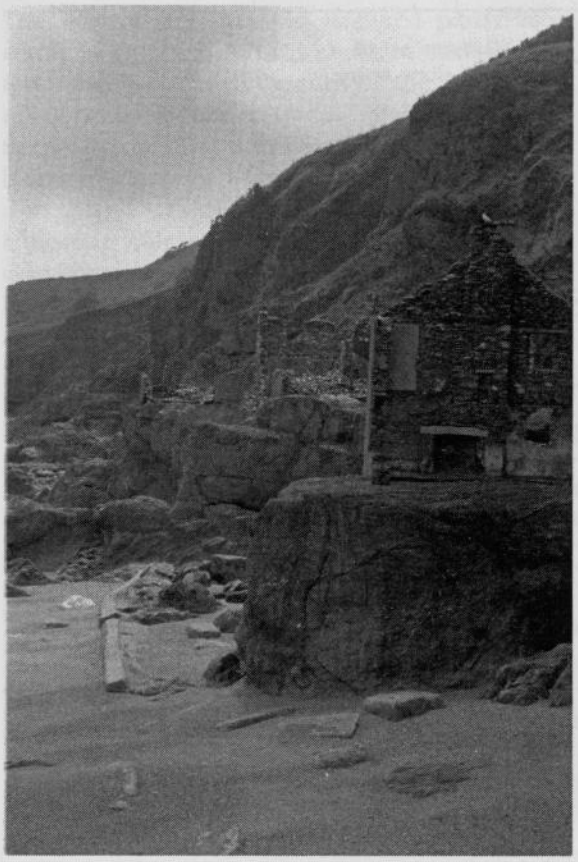


Figure 6.10 The ruins of the landward row of the houses of the former village of Hallsands. The seaward row of houses has completely disappeared. Compare with Figure 6.9. (Photo: V.J. May.)

as a result, dredging was stopped in 1902. By 1904, Worth estimated that the beach had fallen by as much as 6 m, and that 97% of the former beach volume had been removed. Continuing damage occurred with each major storm and a sea-wall was built during 1904. This appears to have been effective until January 1917, when, during a north-easterly gale, waves over 12 m in height destroyed much of the remainder of the village.

Interpretation

At the time, local opinion attributed the disaster to the effects of gravel extraction. Hails (1975a, p. 3) commented that this 'view of the reckless exploitation of shingle...has never been scientifically substantiated'. Worth (1904, 1909, 1923) had never accepted the official view that the beach would be naturally replenished and so set out to monitor the post-excavation changes. He was able to estimate the former extent of the beach by using photographs taken before and after the dredging occurred. In particular, a small stack or stump known as 'Wilson's Rock', which was covered by shingle before the start of dredging, stood afterwards over 3.5 m above the lower rock platform. Up to 1907, there was some gradual rebuilding of the shingle barrier beach to the north of the village, but elsewhere there was little change in the beach. The 1917 storm lowered much of the beach by almost 2 m. Robinson (1961) repeated Worth's surveys and found that the beach was lower in parts than it had ever been previously. Today, the beach has become very limited in volume and a rock platform is usually exposed below the bench on which the shells of the houses stand (Figure 6.10). Robinson noted that the most depleted conditions occurred after a period of easterlies, but did not discuss the reasons for the 1917 disaster. The very limited supply of shingle to the beach could not result from longshore transport since there are no sources to the south and any shingle moving southwards was probably retained by Tinsey Head about 800 m to the north.

Hails (1975a) also questioned whether the village had been constructed on bedrock that was sufficiently resistant to withstand storm wave attack. Robinson (1961) reported that at the Coastguard Cottages the cliff top retreated almost 7 m between 1907 and 1961, commenting that there had been a surprising amount of cliff recession. The schists possess many structural features that weaken the cliffs, and the sea has exploited these weaknesses. To the south of Wilson's Rock, however, there has been substantially less general cliff retreat. Mottershead (1986) described the tendency of houses that had been built on the sediment-filled clefts to collapse during storms, and demonstrated the variable strength of different parts of the village site. Nevertheless, the bedrock exhumed from beneath the shingle shows considerable resistance to erosion. The importance of this site is based on the following:

1. The detailed survey record, which is rare among coastal sites. Indeed it appears to be the longest time-series of beach profiles recorded in the British literature. (There are of course much longer series of repeated *plan* surveys.)

2. The debate about the causes of the Hallsands disaster, and the explanation by Hails (1975b) that it resulted from a combination of gravel extraction, focused wave energy and high wave and tide conditions.
3. The evidence of the timescales at which coastal systems adjust to change. As Mottershead (1986) pointed out, large-scale changes did not occur until a combination of high wave-energy from a particular direction took place. Once the beach was eroded, a further 15 years elapsed before the final disaster occurred.

Conclusions

This is one of the world's best-known coastal erosion sites, mainly because of the catastrophic destruction of Hallsands village following gravel extraction. Erosion of the beach has exhumed buried cliffs and platforms. It is a unique site, combining exhumation of earlier coastal landforms with a long record of surveys that show how this exhumation took place. It is especially important because it is a rare location in which the effects of beach erosion related to both wave conditions and gravel extraction can be demonstrated convincingly. In the context of coastal management worldwide, Hallsands is especially important because it shows the environmental impact of gravel extraction at a site where the coastal processes are not fully understood.

BUDLEIGH SALTERTON BEACH, DEVON (SY 040 801)

V.J. May

Introduction

The coastline of east Devon is characterized by several river valleys that are infilled by alluvium, and their mouths partially blocked by shingle or cobble beaches. Although the beaches at Seaton and Sidmouth have some similarities in form with the beach at Budleigh Salterton, they have suffered considerable erosion in recent years and sea defences and coast protection works have been erected to protect the low-lying towns behind them. However, Budleigh Salterton Beach (see Figure 6.2 for general location) remains largely undisturbed. The beach is formed primarily of shingle- and cobble-sized material derived from the erosion of cliffs cut into Triassic sandstones and pebble beds at the western part of the site. The beach is unusual among English beaches because it is fed with material derived entirely from erosion of cliffs, which cut into Triassic strata. There is a noticeable lack of flint and chert in the beach clasts. The plan form of the beach shows a strong relationship to the refracted Atlantic waves and more direct, but less frequent, waves from the eastern English Channel. The cliff–beach–estuary assemblage was once a very common feature of the coastline of south and south-east England, but most examples have been modified by coast protection works. At Budleigh Salterton, the relative stability of the beach and the lack of natural evolution of the estuary mouth have not required artificial protection.

Research on the site has been limited, although distinctive clasts derived from the Budleigh Salterton Pebble Beds occur in many of the beaches within Lyme Bay and are reported in the literature (Ward, 1922; Steers, 1946a; Bird, 1989; Carr, 1974; Carr and Blackley, 1974a).

In recognition of the site's importance for

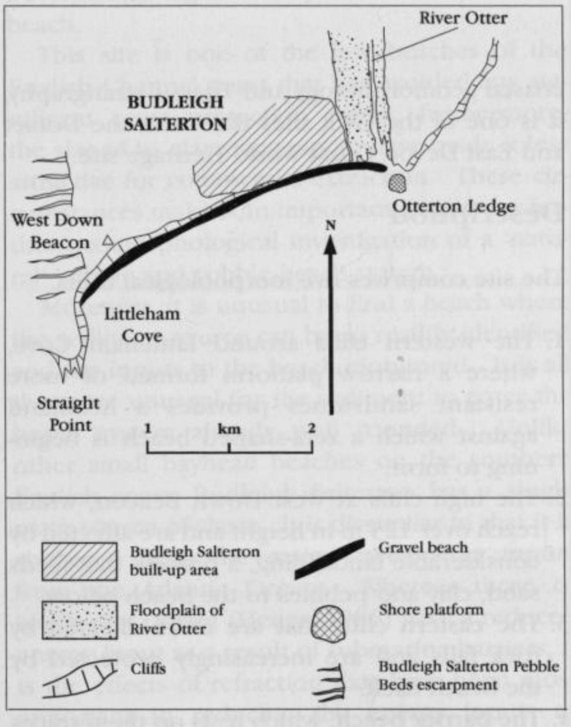


Figure 6.11 Sketch map of the Budleigh Salterton Beach GCR site.

Figure 6.12 Budleigh Salterton beach, looking west, also showing the cliffs that provide the source material for the gravel beach. (Photo: V.J. May.)

coastal geomorphology and Triassic stratigraphy, it is one of the GCR sites that form the Dorset and East Devon Coast World Heritage site.

Description

The site comprises five morphological units.

1. The western cliffs around Littleham Cove, where a narrow platform formed of more resistant sandstones provides a headland against which a zeta-shaped beach is beginning to form.
2. The high cliffs at West Down Beacon, which reach over 125 m in height and are affected by considerable landsliding, a process that feeds sand, clay and pebbles to the beach below.
3. The eastern cliffs that are more affected by rock falls and are increasingly protected by the beach itself.
4. The barrier beach, which rests on the terraces of the Otter Valley and extends across the floodplain as a spit.
5. The cliffs and platform at Otterton Point.

As erosion of the cliffs has taken place, particularly by landslipping around West Down Beacon (Kalaugher *et al.*, 1995), clasts have been added to the beach and distributed throughout its length. The beach is lower, narrower and sandier at its western end in Littleham Cove and becomes higher, wider and less sandy eastwards, especially to the east of West Down Beacon. Although the cliffs decline steadily towards the Otter valley so that the pebble beds do not crop out there, there has been sufficient input of eroded material from the western cliffs to maintain a supply, albeit spasmodically. At the eastern end of the site, the pebbles not only form a barrier across the valley, but also spread seawards around the river mouth lodging against Otterton Ledge, a rock shore-platform. The shingle bank blocked the Otter estuary by the mid-16th century (Ward, 1922). There is no record of substantial erosion or flooding at the town. It appears that, although there has been some retreat of the coastline to the west and that the barrier beach has moved slightly landwards over the valley floor, the clasts are retained within the bay and provide an effective form of coast protection.

Interpretation

The distinctive form and petrology of the clasts, which distinguish them from the predominantly flint or chert content of many English south coast beaches, make them a useful 'tracer' for studies of longshore movements of sediment, especially because the clast source is very restricted in outcrop, both at Budleigh Salterton on the coast and offshore on the seabed.

The beach is unusual in that it shows a very high degree of stability, which is probably a result of its position between two relatively stable headlands, and the fact that the beach rests for much of its length against the cliff foot.

The Triassic Budleigh Salterton Pebble Beds were described by Henson (1970) as a poorly sorted, braided river deposit consisting mainly of ellipsoidal quartzite pebbles with subordinate pebbles composed of vein quartz, 'schorl', sandstone and porphyry. They have a maximum dimension between 19 mm and 100 mm (Carr, 1974) and all show a high degree of rounding. The formation dips to the east at about 5° and forms a marked escarpment running northwards from the coast at West Down Beacon. Carr and Blackley (1974a) described more fully what they identified as metaquarzite clasts from this beach. Pebbles from this distinctive formation have been identified within many other beaches on the southern coastline of Britain, notably at Chesil Beach, Langney Point and Dungeness (Ward, 1922; Steers, 1946a; Lewis and Balchin, 1940). Steers (1946a) suggested that the sites most distant from Budleigh Salterton contain pebbles probably transported as ballast, but there is no reason to reject natural processes as a source for the Chesil Beach examples (Carr and Blackley, 1974a).

The plan form of the beach is largely controlled by the wave-energy distribution between Littleham Cove in the west and Otterton Point in the east where the beach diverts the river Otter eastwards. The alignment of the beach is also affected by the shore platform at Otterton Ledge.

The small estuary of the Otter may allow some sediment to be stored in this embayment. More important is the very high permeability of the pebble ridge at its eastern end so that the ridge is little disturbed except by the largest waves. Unlike the part of Chesil Beach formed of large clasts, Budleigh Salterton beach is not exposed to the full strength of the south-westerly waves. It is aligned towards the SSE. Periods of storm-wave attack from the more easterly directions move sediment along the beach and on occasions denude parts of it. At West Down Beacon, Kalaugher *et al.* (1995) have shown that intermittent movements in a mudslide at the base of the cliffs can be linked to large-scale collapses of the conglomerate that forms the upper cliff. In stormy conditions the mudslide is triggered at high tide. The landslides interrupt longshore transport and add new material to the beach. Bird (1989) identified this beach as the only one within Lyme Bay that is not laterally graded. Carr (1974) considered that when significant grading was observed, it occurred along the whole beach, with the smallest grade material at the eastern end. He noted that in June 1972 mean clast size increased from the centre of the beach. This observation is consistent with the patterns of sorting described by Heeps (1986) in similar confined beaches in south-east Dorset. Carr also demonstrated that not only was sediment graded in size from the centre of the bay, but it also changed in shape from the end of the beach.

This site is one of the few beaches of the English Channel coast that has avoided any significant coast protection works; furthermore, the size of its clasts appears to have made it less attractive for commercial extraction. These circumstances make it an important locality for further geomorphological investigation of a 'natural' pebble and cobble beach system.

Moreover, it is unusual to find a beach where the sediment source can be so readily identified and the inputs to the beach monitored. It is all the more unusual for the sediment to enter the beach system already well rounded. Unlike other small bayhead beaches on the southern English coast, Budleigh Salterton has a single main source of clasts. It is dissimilar in that it is sheltered from the main wave-energy inputs from the Atlantic Ocean. Whereas those of south-east Dorset (Heeps, 1986) have a reduced energy input as a result of submarine barriers, it is the effects of refraction that have been most important in reducing the energy inputs to Budleigh Salterton Beach. Large beaches of cobbles are comparatively rare on the English coast, but they are commonly found in association with large south-west fetches. This site thus provides a substantial contrast with cobble beaches at Westward Ho! and Porth Neigwl, as well as with the cobbled part of Chesil Beach (see GCR site reports in the present volume).

As a site for the investigation of sediment budgets, beach adjustment to wave conditions and the effects of beach permeability on both beach and cliff stability, Budleigh Salterton offers considerable opportunities for field investigation and coastal modelling. Owing to the distinctive clast source, the site provides a rare opportunity to observe the ways in which clasts survive or change in shape following their introduction into the marine environment. The clasts can be readily identified amidst large volumes of flint gravel, confirming their longevity. Furthermore, it is the only site that has large well-rounded clasts dominated by a single rock type other than flint and chert. The unusual quantity, hardness and shape of clasts from the Budleigh Salterton Pebble Beds have given rise to a unique beach.

Conclusions

This pebble and cobble beach is uniquely fed by pebbles and cobbles derived entirely from Triassic sediments. This is the only point where these pebbles enter the coastal system at present, although they are found in beaches along the length of the south coast of England. The lack of anthropogenic influence greatly increases the geomorphological importance of this site as one in which an intact and virtually unmodified natural system can be studied. Owing to its important geology and geomorphology, it is part of the Dorset and East Devon Coast World Heritage site.

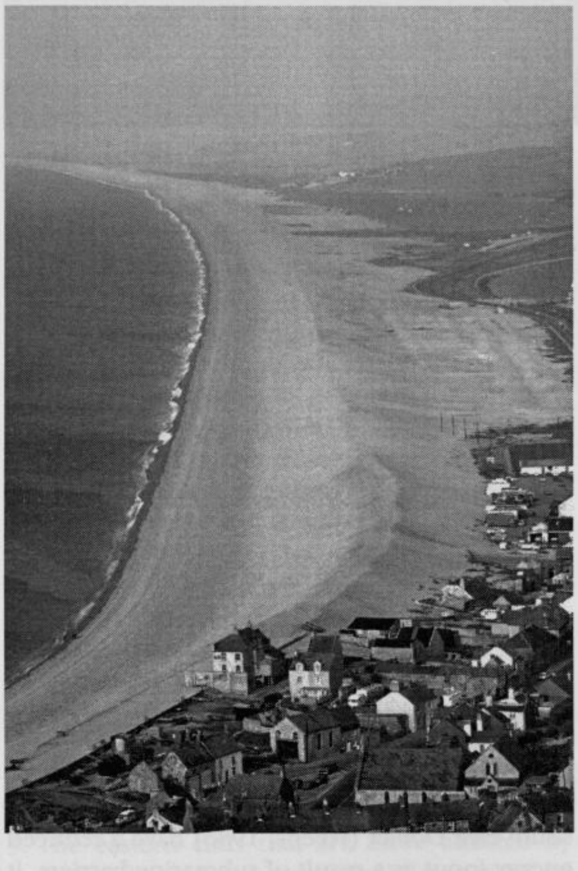


Figure 6.13 Chesil Beach. View looking north-west, from Portland, with Chesilton in the foreground. The beach reaches 14 m OD and over 150m wide at its eastern end, where limited washover still occurs in spite of artifical modifications. (Photo: J.D. Hansom.)

CHESIL BEACH, DORSET (SY 462 903)

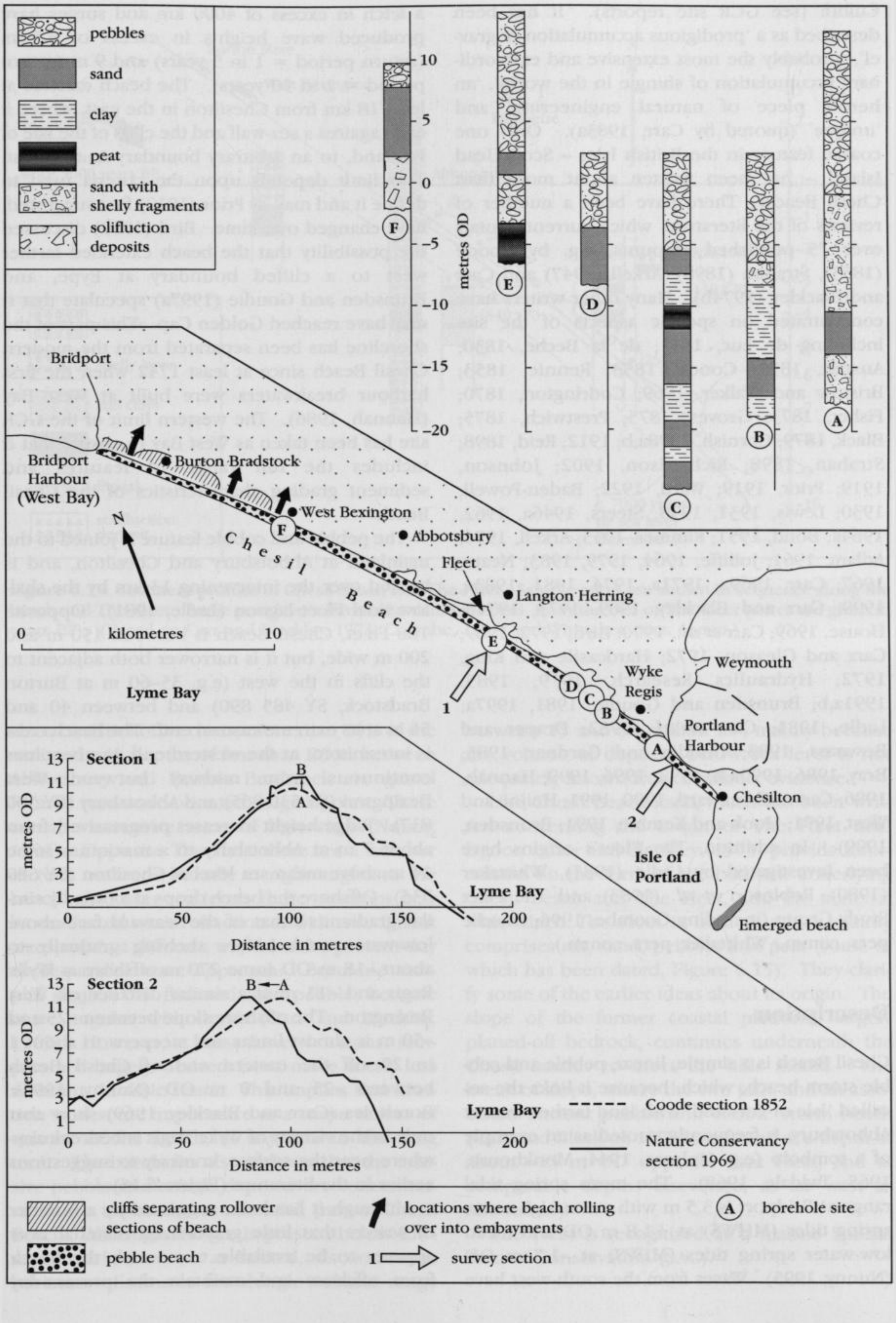
V.J. May

Introduction

Chesil Beach has been described as 'unique' – and is of considerable international renown and scientific significance. Its sheer size (over 18 km in length and exceeding 14 m in height), the systematic longshore size-grading of beach material, the evidence for the south-westerly provenance of its pebbles, the availability of historical records, and the sedimentary record in the adjacent lagoon ('The Fleet'), each contribute to the geomorphological importance of the site and help to explain why there is a vast scientific literature about it. In recognition of the importance of the site for coastal geomorphology, it is one of the GCR sites that form the Dorset and East Devon Coast World Heritage site.

Chesil Beach is one of five major gravel/shingle features along the British coast, together with Dungeness, Orfordness, Spey Bay and

Figure 6.14▸ Map and sections of Chesil Beach. For general location see Figure 6.2. (Based on borehole information in Carr and Blackley, 1969, 1973; and Carr and Seaward, 1990.)


pebbles
sand
clay
peat
sand with shelly fragments
solifluction deposits
metres OD
10
5
0
−5
−10
−15
−20
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D
E
F
Bridport
Bridport Harbour (West Bay)
Burton Bradstock
West Bexington
Abbotsbury
Fleet
Langton Herring
Chesil Beach
N
0
kilometres
10
Lyme Bay
1
Weymouth
Wyke Regis
Portland Harbour
Chesilton
2
Isle of Portland
Emerged beach
Section 1
13
11
9
7
5
3
1
0
50
100
150
200
Distance in metres
Section 2
B←A
Coode section 1852
Nature Conservancy section 1969
cliffs separating rollover sections of beach
pebble beach
locations where beach rolling over into embayments
survey section
borehole site

Culbin (see GCR site reports). It has been described as a 'prodigious accumulation of gravel', 'probably the most extensive and extraordinary accumulation of shingle in the world', 'an heroic piece of natural engineering' and 'unique' (quoted by Carr, 1983a). Only one coastal feature in the British Isles – Scolt Head Island – has been written about more than Chesil Beach. There have been a number of reviews of the literature, which currently totals over 75 published accounts, e.g. by Coode (1853), Strahan (1898), Arkell (1947) and Carr and Blackley (1974b). Many other writers have concentrated on specific aspects of the site including de Luc, 1811; de la Beche, 1830; Austen, 1851; Coode, 1853; Rennie, 1853; Bristow and Walker, 1869; Codrington, 1870; Fisher, 1873; Groves, 1875; Prestwich, 1875; Black, 1879; Cornish, 1898a,b; 1912; Reid, 1898; Strahan, 1898; Richardson, 1902; Johnson, 1919; Prior, 1919; Ward, 1922; Baden-Powell, 1930; Lewis, 1931, 1938; Steers, 1946a, 1962, 1964a; Bond, 1951; Rimmer, 1953; Arkell, 1954; Adlam, 1961; Jolliffe, 1964, 1979, 1983; Neate, 1967; Carr, 1969a, 1971a, 1974, 1981, 1983a, 1999; Carr and Blackley, 1969, 1973, 1974b; House, 1969; Carr *et al.*, 1970; Bird, 1972, 1989; Carr and Gleason, 1972; Hardcastle and King, 1972; Hydraulics Research, 1979, 1985, 1991a,b; Brunsden and Goudie, 1981, 1997a; Ladle, 1981; Carr *et al.*, 1982; Draper and Bownass, 1983; Goudie and Gardner, 1985; Bray, 1986, 1990a,b, 1992, 1996, 1999; Hannah, 1986; Carr and Seaward, 1990, 1991; Heijne and West, 1991; Hook and Kemble, 1991; Brunsden, 1999). In addition, The Fleet's origins have been investigated by Ladle (1981), Whittaker (1980), Robinson *et al.* (1983) and The Fleet Study Group (including Coombe, 1996; Goudie pers. comm., Whittaker, pers. comm.).

Description

Chesil Beach is a simple, linear, pebble and cobble storm beach, which, because it links the so-called 'Isle of Portland' with land farther west at Abbotsbury, is frequently quoted as an example of a tombolo (e.g. Holmes, 1944; Monkhouse, 1965; Twidale, 1968). The mean spring tidal range at Bridport is 3.5 m with mean high-water spring tides (MHWS) at +1.8 m OD and mean low-water spring tides (MLWS) at –1.7 m OD (Nunny, 1995). Waves from the south-west have a fetch in excess of 4000 km and surges have produced wave heights in excess of 6.5 m (return period = 1 in 5 years) and 9 m (return period = 1 in 50 years). The beach extends at least 18 km from Chesilton in the east, where it ends against a sea-wall and the cliffs of the Isle of Portland, to an arbitrary boundary in the west. This limit depends upon the criteria used to define it and may, as Prior (1919) first suggested, have changed over time. Bird (1989) discussed the possibility that the beach extended farther west to a cliffed boundary at Eype, and Brunsden and Goudie (1997a) speculate that it may have reached Golden Cap. This part of the shoreline has been separated from the modern Chesil Beach since at least 1742 when the first harbour breakwaters were built at West Bay (Hannah, 1986). The western limit of the GCR site has been taken as West Bay to ensure that it includes the full range of features and sediment grading characteristics of the Chesil Beach.

The pebble and cobble feature is joined to the mainland at Abbotsbury and Chesilton, and is backed over the intervening 13 km by the shallow tidal Fleet lagoon (Ladle, 1981). Opposite The Fleet, Chesil Beach is between 150 m and 200 m wide, but it is narrower both adjacent to the cliffs in the west (e.g. 35–60 m at Burton Bradstock, SY 485 890) and between 40 and 54 m at its extreme eastern end. The beach crest is intermittent at the western end, but becomes continuous from midway between West Bexington (SY 530 865) and Abbotsbury (SY 570 837). Ridge height increases progressively from about 7 m at Abbotsbury to a maximum some 14 m above mean sea level at Chesilton (SY 680 735). Offshore the beach drops at a broadly similar gradient to that of the seaward face above low-water mark before shelving gradually to about –18 m OD some 270 m offshore at Wyke Regis and –11 m at a similar distance off West Bexington. The offshore slope between –25 and –50 m is almost linear and steepens to about 1 in 20 off the eastern end of Chesil Beach between –25 and 0 m OD (Nunny, 1995). Boreholes (Carr and Blackley, 1969) show that only in the vicinity of Wyke Regis is bedrock anywhere near the surface, contrary to suggestions earlier in the literature (Figure 6.15).

Although it has been suggested by a number of writers that little gravel-sized material now appears to be available to nourish the beach from offshore and maintain the present-day

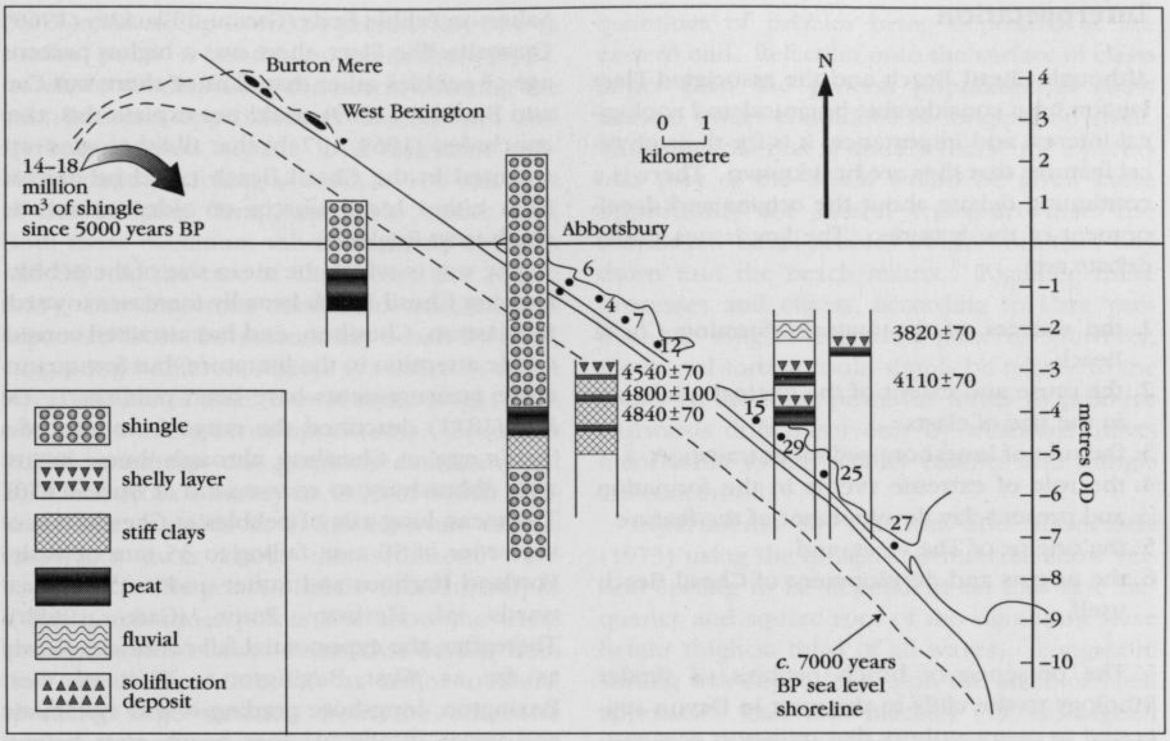

Figure 6.15 Sediment profiles of Chesil Beach and The Fleet. Sample cores are shown in sequence along the beach and The Fleet. Some peat layers have been dated in cores from the bed of The Fleet (dates are given in years BP). (Based on Carr and Blackley, 1973; Coombe, 1996 and Whittaker, pers. comm.)

characteristics of its clasts, more than 50% of the sediments in depths of less than 25 m between Abbotsbury and Burton Bradstock are gravel formed into wave-oriented mega-ripples. There are also extensive areas of bare rock (Nunny, 1995). Opposite The Fleet there are a number of rock outcrops broadly at right angles to the beach. At lower sea levels these outcrops may have affected wave refraction and thus sediment transport in a different way from the present-day shore-parallel wave approach. Borehole samples suggest that flint and chert pebbles become more angular with depth (Carr and Blackley, 1969). However, at these lower horizons, samples are largely derived from more local, less resistant, Jurassic strata. This implies that attrition is of some importance as a cause of loss of volume of the beach, at any rate in the long term. The boreholes also indicate that the massive pebble and cobble deposits are concentrated in the exposed, i.e. subaerial, part of Chesil Beach. Although shingle is present below low-water mark, it occurs as limited, discontinuous horizons. Estimates for shingle volume range between 15 and 60 million m^3, mainly because the volume of deposits below sea level is not adequately known from borehole evidence.

Like Chesil Beach itself, much has been written concerning the origin of The Fleet tidal lagoon. Carr and Blackley (1973) provide information on the form of the bedrock underlying Chesil Beach and The Fleet, and the material with which The Fleet is infilled, which mainly comprises silt, sand, pebbles and peat (some of which has been dated, Figure 6.15). They clarify some of the earlier ideas about its origin. The slope of the former coastal platform, largely planed-off bedrock, continues underneath the Chesil barrier to meet the hills inland. The break of slope, where the two join, and an associated ancient pebble and cobble storm beach (Carr and Blackley, 1973) are found at a depth of about –15 m OD opposite East Fleet, and at comparable depths elsewhere, as far west as Abbotsbury.

The Fleet is recognized as a marine Special Area of Conservation (SAC).

Interpretation

Although Chesil Beach and the associated Fleet lagoon have considerable botanical and zoological interest and importance, it is for their physical features that they are best known. There is a continuing debate about the origins and development of the features. The key issues under debate are:

1. the sources of the material forming Chesil Beach
2. the cause and extent of the distinct gradation in the size of clasts
3. the role of longshore sediment transport
4. the role of extreme events in the formation and present-day development of the feature
5. the origins of The Fleet, and
6. the origins and development of Chesil Beach itself.

The presence of beach pebbles of similar lithology to the cliffs to the west in Devon suggested to many authors that transport was eastward from the source areas by littoral drift along former shorelines (de la Beche, 1830; Fitzroy, 1853; Rennie, 1853; Pengelly, 1870; Baden-Powell, 1930; Arkell, 1947) or crossed Lyme Bay (Strahan, 1898). Prestwich (1875), however, suggested that sediments were transported north-west from a precursor of the Portland emerged ('raised') beach (SY 675 684), situated south of the Isle of Portland. Prior (1919) suggested three possible sources: an earlier Chesil-like beach from Start Point in Devon to Portland, erosion of the east Devon coast or river gravels deposited in Lyme Bay by a river of which The Fleet was once part. Bond (1951) argued for an ancient Exe–Teign river, flowing up to 10 km offshore from the present-day coast to a mouth south-east of Portland. Arkell (1947) thought shingle rafting by ice could account for some of the more exotic pebbles. The geological evidence indicates that there have been various potential sources of the pebbles and cobbles, including fluvial as well as marine deposits, and that the relative significance of these sources is likely to have varied over the long term. About 98% of the material is flint and chert that could have been derived from a number of primary (and secondary) local sources, but the remaining 2% (including Triassic quartzites) probably was derived from the south-west. 95% of the quartzite material is derived from the Budliegh Salterton Pebble Beds (Carr and Blackley (1969). Opposite The Fleet, there was a higher percentage of pebbles other than flint or chert, but Carr and Blackley (1969) could not explain this; they concluded (1969, 1974b) that all lithologies represented in the Chesil Beach could be derived from either local sources or older sources in south-west England.

The way in which the mean size of the pebbles forming Chesil Beach broadly increases towards the eastern, Chesilton, end has attracted considerable attention in the literature, but few quantitative measurements have been published. De Luc (1811) described the range from that of a 'hen's egg' at Chesilton, through 'horse beans' near Abbotsbury, to coarse sand at Burton Cliff. The mean long-axis of pebbles at Chesilton is of the order of 50 mm, falling to 35 mm opposite Portland Harbour and rather under 25 mm seawards of Herbury Point (Carr, 1969a). Thereafter, the exponential fall continues slowly as far as West Bexington. West of West Bexington, longshore grading is less systematic and varies locally, as does beach crest height. These variations may reflect minor changes in beach orientation, commercial exploitation of beach material, and/or local sources of supply from the cliff to landward. Reid (1898, 1907) explained the higher proportion of Budleigh Salterton quartzites at Chesilton by their greater resistance to abrasion. Arkell (1947) thought that their greater durability or later replenishment could account for their increased frequency here relative to the Portland emerged beach. Carr and Blackley (1969) stated that the percentage of quartzites did increase towards Chesilton but was less than appeared to be the case in the field. The quartzite pebbles tend to be flatter with a larger surface area that may make them more conspicuous (Carr and Blackley, 1969).

Carr (1965) took samples at 27 locations, 1.6 km apart (0.8 km apart at the eastern end), with between 3 and 11 samples per section at crest, high and low water. During 1965–1966 pebble samples were taken along a series of transects across Chesil Beach, as part of a wider research programme (Carr, 1969a). The results showed that on surface profiles between Smallmouth and the Bridging Camp, there were areas near The Fleet where pebble size was smaller, and degree of sorting greater, than nearer the beach crest. It was suggested that these samples might represent the legacy of a different beach relationship from the present-day one (a similar point is made by Brunsden, 1999). Borehole samples along the beach, covering the area between approximately Smallmouth and Herbury showed angular, local limestone pebbles at depth, reaching as much as 47% just west of the Bridging Camp, and 33% farther east. Both these maximum values occurred at about –15 OD. In the case of Carr's core (see Figure 6.15), this limestone-dominated material was separated from the present-day beach by sand, suggesting a different regime at a different depth level, and hence time (even if equal height does not always imply contemporaneity). Brunsden (1999) comments that graphical data compiled by Babtie from a survey in 1996 in which sediment was sampled at 1 km intervals reveals important facts about the sediment wave hypothesis. The printed Babtie data link only as a line graph those points that show the trend and similarities of size. Points that deviate from the overall trend are left as major outliers, revealing large variation from the otherwise westward decreasing trend. Thus there are areas where sediments are distinctively larger than would be expected from the normal trend.

Studies over very short periods of time using brickbats (Richardson, 1902) and painted pebbles (Adlam, 1961) showed net eastward wave-induced transport and preferential movement of larger materials. Carr's (1971a) experiments using clasts of a foreign or exotic lithology confirmed that larger material moved more rapidly eastwards, showing particularly strong eastward drift at Wyke Regis (SY 660 760), becoming more variable and random nearer Portland. Carr concluded that there was no consistent drift near Portland, thus explaining the absence of a large accumulation at Chesilton. Both pebble measurements and tracer experiments (Carr, 1971a) have shown that thickness (B axis) appears to be the most significant dimension affecting pebble transport alongshore by waves. Although movement seems predominantly to the east, negative correlations have also been recorded. Near Portland, travel eastwards is reduced by the more random nature of longshore movement as compared with sites farther west, where waves usually approach the shoreline more obliquely. The length of burial time of particles also varies between the two areas. These factors, coupled with the absence of sizeable amounts of new material, probably account for the lack of any permanent drift and thus of the absence of large quantities of pebbles being deposited at the eastern end. Rejection onto the surface of clasts larger than the general population is most marked under conditions of long-period swell. Thus material that is unduly coarse for a particular part of the beach would be given more opportunity for lateral transport than the remainder. Small pebbles would work their way down into the beach matrix. Together, these processes and effects, according to Carr, produce the longshore grading pattern. However, the lateral sorting could simply be related to the ability of the most powerful waves to produce eastwards drift offset only by westward movement as far as the smaller cobbles and shingle are concerned.

Experiments by Gleason and Hardcastle (1973) using the indigenous material show vertical sorting to be dependent on the wave frequency and square root of the significant wave height (highest third of all waves). Longshore sorting was dependent upon the angle of swell approach. Carr and Blackley (1974b) argued that fresh sediment inputs would tend to diminish size grading, although Bray (1990b) considered that new inputs of larger pebble sizes may be necessary to counter the continuous reduction by attrition. Carr and Blackley (1969) showed how limestone clasts derived from quarry waste from the western Isle of Portland were virtually unrepresented in the natural beach population at a distance of some 3–4 km north-west of Chesilton (a reflection of both attrition and net longshore transport).

Most explanations for the sorting concentrate on the commonly held view that:

- there is a continuous size change along the beach,
- sorting occurs by size and shape,
- rates of pebble movement depend on pebble thickness (B-axis dimension),
- different wave energy of storms from the south-west and the south-east cause the sorting, and
- different depths of water offshore and therefore differences in available energy are a fundamental cause of the clast distribution and sorting.

However, the above model may be based on simplistic notions of sediment transport with pebbles moving singly by longshore drift that assumes an open system with a continuous sup-

Figure 6.16 West Bay, Chesil Beach, showing the retreat of the shoreline and lack of sediment at the western end of the modern Chesil Beach. (Photo: V.J. May.)

ply of material for drift (Brunsden, 1999). He argues first that the present-day beach cannot be regarded as an open system, for it is closed at the western end by the harbour walls at West Bay. Even before they were constructed, restriction of longshore movement by headlands to the west existed. Brunsden suggests that on Chesil Beach, groups of pebbles move both west and east as the wave approach varies but with a net westward movement. Groups dominated by small sizes move over groups of bigger ones and along different storm ridges, forming and re-forming as conditions dictate. Such a process has also been observed on the shingle beaches at Ringstead (May, 1999) and at Spey Bay, where 'slugs' of gravel move alongshore (Gemmell *et al.*, 2001b). As pebbles are eroded from each beach ridge, remnants of pebble groups are left at different beach levels according to the severity and sequence of storms. When major storms washover the beach, surface beach material is moved over the crest regardless of the size and composition of pebbles that occur on the beach at that place at that time. As the beach moves landwards this sediment may re-emerge on the beach face many years later. As a result, Brunsden argues that any sorting model must be a complex, episodic spiral of individual movement and movement of gravel groups.

On the cliffed coast to the west, gravel inputs of approximately 5000 $m^3 a^{-1}$ have occurred over the past 4000 to 5000 years (Bray, 1990a,b). Assuming that losses by attrition and entrapment have remained constant through time at approximately 1500 $m^3 a^{-1}$, Bray estimates that between 14 million and 18 million m^3 of sediment has been supplied to the coast by landslides since 5000 years BP. The beaches between Lyme Regis and West Bay store slightly less than 1 million m^3 of shingle at present. Since the long-term mean rate is similar to present-day figures, Bray argues that current shingle budgets support the hypothesis that erosion of the west Dorset coast provided a major sediment source for the creation and replenishment of Chesil Beach. Today, however, a continuous supply of material no longer exists. Shoreline transport is regulated by landslide activity occurring at the main headlands of Golden Cap and Doghouse Hill. Bray's (1990b) model envisaged intermittent pulses of gravel bypassing these headlands at intervals of 30–50 years, most recently at Golden Cap between 1949 and 1962. Relict landslide deposits (boulder aprons) identified some 2–3 km offshore from Golden Cap (Brunsden and Chandler, 1996) may indicate the extent of past cliff erosion (Figure 6.17b). Bray (1990a) estimated that about 32 million m^3 of gravel would be supplied to the retreating shore. Some materials would also have been contributed from the East Devon coast, giving a combined total of 58 million m^3. Chesil Beach is the only significant gravel accumulation within this part of the Lyme Bay cell.

According to Bray, the hypothesis is supported by a variety of evidence. Surveys in Lyme Bay have failed to reveal alternative offshore shingle sinks. Analysis of the size and lithology of clifftop gravels in the Charmouth area indicate that the material is comparable with Chesil Beach shingle, and the size, shape and lithology of pebbles on Charmouth, Seatown, Eype and Chesil beaches support the hypothesis that the beaches were formerly contiguous (Bray, 1990b). The contemporary pattern of eastward littoral drift has probably existed over the last 5000 years. There is no evidence that the coastal orientation was significantly different in the past or that the frequency of west and south-west storms was less (Bray, 1996).

The present-day volume of Chesil Beach is estimated at 15 to 60 million m^3 (Carr, 1980). The estimated surplus by coastal landsliding is 14 to 18 million m^3 over the past 4000 to 5000 years. Although potential shingle supply from landslides is significant by comparison with the present-day volume of Chesil Beach, it is unlikely to have been the main supply (Carr, 1980). Chesil Beach had already formed by 7000 years BP (Carr and Blackley, 1973). It is therefore suggested (Bray, 1990b) that sediment supply from terrestrial sources, such as coastal cliffs, updrift or feeder bluffs/cliffs) was a mechanism by which Chesil Beach has been nourished and enlarged. The original gravel source was probably fluvial and periglacial deposits on the floor of Lyme Bay, which gradually decreased in importance as the rate of sea-level rise slowed down. Subsequent erosion of the cliffs provided a supply of flint and chert that helped to maintain Chesil Beach. Shingle supply from the west by littoral drift was possible until as recently as the mid-1860s when longshore transport was halted by the construction of the piers at West Bay. The supply process up to 1860 may have offset attrition losses and assisted in the maintenance of the unique Chesil Beach size grading (Carr, 1969a) because it ensured that a wide range of clast sizes was always available. However, mineral extraction has been important in the past, especially at Seatown and West Bay. Material has been removed from the beach at West Bay for over 700 years, with about 1 million tonnes of gravel removed between the mid-1930s and 1977 (Hydraulics Research Station, 1979). Of this, more than 470 000 tonnes were taken from East Beach, Bridport Harbour, and 370 000 tonnes from Cogden Beach. At one time, pebbles were removed from other locations, e.g. during 1905–1907 from the backslopes of Chesil Beach. Perhaps the most significant of these activities was the selective pebble picking carried out nearby, not because of the absolute quantities, varying between approximately 100 and 350 tonnes per year from 1944 to 1972, with a recorded total of some 9400 tonnes, but rather because of the removal of particular sizes and shapes. This may well have produced a disproportionate weakness in the beach as well as affecting locally the longshore grading pattern and distorting the geomorphogical processes.

Changes in crest height of Chesil Beach over the last 300 to 400 years and that at one time the crest may have been lower over most of the length between Abbotsbury and Portland. Although the total volume of beach material appeared to change very little between 1852 and 1968–1969 (Carr and Gleason, 1972), the crest height between Abbotsbury and Wyke Regis showed a substantial increase. This was of the order of 2 m at Langton Herring (SY 605 810). Between Langton Herring and east of Wyke Regis (SY 650 770) there was a rise, typically, of 1.5 m. However, near Chesilton, a drop of 0.5 m, reaching an extreme fall of 3.5 m at one point, was recorded. Carr and Gleason (1972) found difficulty in explaining this phenomenon although it gave credence to early 19th century reports that the beach used to be overtopped more frequently. A comparison of the 1968–1969 profile and associated data with that of March 1979 shows that the single winter of 1978–1979 was capable of producing the same order of change at the south-east end of Chesil Beach as that indicated between 1852 and 1968–1969. Thus at one location there was a maximum fall of 2.7 m in crest height between

September 1978 and March 1979 surveys. Coupled with the known stability of the crest between 1955 and September 1978, it suggests that one event could be enough to produce the scale of change observed over the period from 1852 to 1968–1969. Such an event appears to have occurred in 1904 under similar long-period swell conditions to those recorded on 13 February 1979. A possible mechanism to account for these height changes is that where atypically large swell waves arrive parallel to the beach, the crest is overtopped, lowered, and rolled inshore (i.e. towards Portland Harbour). Farther west, towards Abbotsbury, the same swell would arrive more obliquely so that instead of clasts moving from low-water mark, over the crest, and down the backslope, the material would simply be transferred from the face to the crest, by which time the wave energy would be expended. During this process, the crest would become higher than before and there would be some net longshore transport of clasts towards the east.

The western end of Chesil Beach shows considerable variation, which is related not only to the construction of the piers and mineral extraction, but also the reaction of the beach to prolonged periods of different wind direction. Between 1901 and 1984, based on comparisons between maps and field survey, accretion on the eastern side of the West Bay piers was not continuous. There were brief periods of erosion as for example between 1961 and 1964 (Hydraulics Research, 1979, 1985; Jolliffe, 1979). Littoral drift between West Bay (SY 462 904) and Cogden Beach (SY 504 880) was investigated using a mathematical beach transport model, and a hindcast wave climate model based on Portland wave data covering the period from 1974 to 1984 indicated mean net eastward transport at 8000 $m^3\ a^{-1}$, similar to the documented trend for accretion against the east pier at West Bay (Hydraulics Research, 1985). The analysis is open to question, however, because it ignored swell waves and waves under 1 m and used shingle transport equations that had been calibrated on other beaches. Analysis of beach profiles and aerial photographs covering the period 1977 to 1990 showed that there was a marked switch from previously recognized patterns of accretion immediately east of the pier at West Bay in about 1982, to erosion, which resulted in retreat at mean high-water level by 40 m by March 1990 (Hydraulics Research, 1991a). The wave climate changed after 1982, with fewer south-east storm waves and an increase in westerly waves, i.e. a return to the generally accepted historical prevailing pattern. Littoral drift calculations confirmed net westward drift before 1982 and net eastward drift of up to 14 000 $m^3\ a^{-1}$ after 1982 (Hydraulics Research, 1991a). It appears from these studies that net littoral drift is very delicately balanced at both ends of Chesil Beach, especially at the western end where slight wave climate and storm frequency variations may produce major reversals of drift. Beach morphometry may react relatively slowly to changes in drfit regime because of the large volume of Chesil Beach, and so trends may be identifiable only at the ends of the beach. They contribute therefore, little to an understanding of the mechanisms along the whole beach.

The environmental history of The Fleet is critical to an understanding of the origins and subsequent development of Chesil Beach. Within The Fleet, sedimentation of clays, silts and sands is evident above the –15 m level to about –3 m OD where commonly thick common-reed *Phragmites* peat layers (dated at *c.* 5000 years BP) occurred (Carr and Blackley, 1973). As peat becomes exposed it is eroded and thrown up on the beach between Abbotsbury and West Bexington. The pollen from a sample at –13.4 m OD in Carr and Blackley's Langton Herring (E) borehole (Figure 6.14) at the boundary between sand and bedrock was tentatively dated as early Pollen Zone VI. They interpreted this as showing that both the peat formation and the sedimentary infill above bedrock must have been very rapid. At the eastern end of the Narrows (SY 650 772), peat deposits, with a high pine-pollen content, were found underlying the landward side of Chesil Beach at a depth of *c.* –5.3 m and dated at *c.* 6200 years BP. Bedrock at this locality, occurred at –13.1 m OD, although it was as shallow as –7.8 m OD at a neighbouring location. It is unclear how these data fit into the evolutionary picture – does it imply, for example, that the relatively deep, narrow channel between the army Bridging Camp and Chesil Beach was cut as some sort of overflow feature? (Carr, 1999).

The detailed investigation of the sediments within The Fleet shows that its evolution has been a complex one (Goudie, pers. comm.; Whitaker, pers. comm.). Coombe (1996) provides details of 26 boreholes that include lagoonal and pre-lagoonal phases. A series of radiocarbon dates in peats at depths of –3.00 m OD, –3.60 m OD and –4.32 m OD have been dated between 4540 ± 70 and 4840 ± 70 years BP (Figure 6.15), indicating rapid accumulation such as Carr and Blackley (1973) had inferred. Two samples in the East Fleet (cores 25 and 29) at –3.00 m OD and –3.15 m OD were dated at 3820 ± 70 and 4110 ± 60 years BP respectively. Mean grain size in Coombe's cores decreased westwards, indicating that energy lessened along The Fleet, supporting the hypothesis that the infill was derived mainly from the south-east in Weymouth Bay. The pollen in a sand sample beneath the lagoonal sequence included a high frequency of pine *Pinus*, together with oak *Quercus*, hazel *Corylus* and birch *Betula*. These are similar to the Carr and Blackley (1973) findings that placed these basal sediments in Pollen Zone VI. This suggests that these sands found below the lagoonal sequence may be older than 6000–6500 years BP. In all the cores, there is evidence of saltmarsh above the sands. This predates the brackish-marine lagoonal phase, which is not older than 5000 years, and is the dominant feature today. Throughout The Fleet narrow shell beds rest on top of each peat bed. Goudie *et al.* (pers. comm.) consider that the peats either represent stillstands or slight falls in the rising Holocene sea level or indicate that The Fleet was closed by a barrier and became a freshwater lake or a reed swamp cut off from the sea. West Fleet was behaving more like an estuary than a lagoon. At West Bexington, foramifera and ostracods show that a tidal, near-marine, water body existed well to the west of the modern Fleet around 4000–5000 years BP (Whitaker, pers. comm.). However, in order to allow for such a body to exist, it is necessary to have a barrier well seaward of the solifluction slope between Abbotsbury and West Bexington.

Although there is at present insufficient evidence for this to be more than conjecture, it would help explain the nature of the West Bexington and Burton Mere environments, both of which appear to be extensions of The Fleet. A further complication in identifying the origins of The Fleet and Chesil Beach comes from the solifluction slope itself, which extends the alignment of Chesil Beach towards West Bexington. If, as Brunsden and Goudie (1997a) have suggested, the solifluction materials overlie evidence of the Portland emerged beach, the western part of an extended Fleet could be a very old feature. To the west of West Bexington, present-day cliffs probably extended farther seawards and would have acted as an early headland from which sandy materials could supply a low sandy barrier beach, it also provides a basis for some re-interpretation of the classic transgression model for the development of Chesil Beach.

The longer term evolution of Chesil Beach is still not completely understood, but the chronological sequence compiled by Carr and Blackley (1973), modified by Bray (1990b) makes it possible to put forward the following scenario for the initiation and development of Chesil Beach.

1. Around 210 000 years BP, the Portland emerged beach is formed and the slopes between Abbotsbury and the Narrows are trimmed. The beach is re-occupied about 125 000 years BP. Westwards from Abbotsbury, there is also an emerged beach and so a coastline very close to the present-day extended at least between Smallmouth and West Bexington. A forerunner of Chesil Beach may have existed as a bank well offshore of the present beach some 120 000 years BP (Carr and Blackley, 1973), contemporaneous with the development of the Portland emerged beach.
2. From the emerged beach level of between +7 and +15 m OD, sea level fell to about –120 m. The seabed is weathered by periglacial processes and a series of gravel-rich deposits, probably comprising material from the Portland emerged beach, solifluction deposits, river gravels and fluvioglacial deposits were deposited on the floor of Lyme Bay. Degraded landslides and solifluction deposits extend 2 to 3 km seawards of the modern shoreline position and mantle the former coastal cliffs.
3. From about 20 000 years BP, sea level rose by 1 mm a^{-1}, and by about 10 000 years BP sea level was at approximately –45 m. The proto-Chesil Beach approaches the former shore of Lyme Bay. The sea level then rose rapidly increasing by an average of 1.5 m every 100 years.
4. About 7000–6500 years BP, the model suggests that closer to the modern coastline, the transgressing Chesil Beach overrode existing sediments as sea level rose. A shallow lagoon that became The Fleet was rapidly filled with silt, sand, pebbles and peat from 7000 to 5000 years BP. About 7000 to 6500 years BP, with

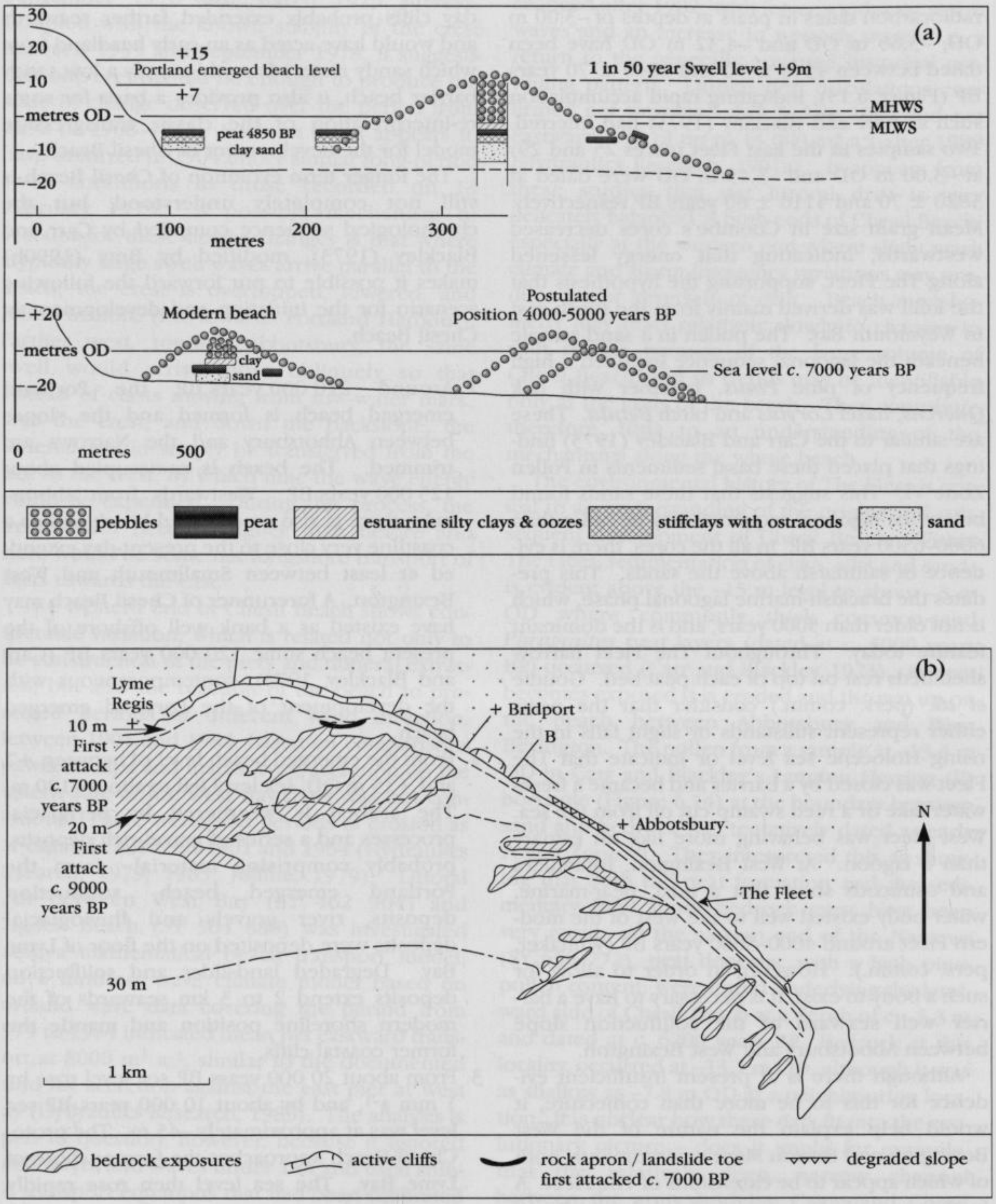

Figure 6.17 Chesil Beach (a) relationships between modern beach and dated peats and water levels (mean high-water springs and mean low-water springs, MHWS and MLWS, are shown). By c. 5000 years BP, the supply of flint was able to create a barrier beach atop an earlier sand ridge and estuarine peats. (b) Seabed features of eastern Lyme Bay and their relationship to Chesil Beach. Note the relation of bedrock exposures and seabed contours to the present shore, which probably affected the development of the earlier beach form. 'First attack' indicates the bathymetric contour representing the shoreline first attacked by the sea at the date shown.

sea level between –12 and –4 m OD, a low sand and gravel beach developed at about 1 km offshore (Figure 6.17a). Relict cliffs farther to the west, abandoned by falling early Devensian sea levels, were re-activated by marine erosion (Figure 6.17b) and large quantities of gravel were transported eastwards, to feed and enlarge the new Chesil Beach (Bray, 1990b). The Fleet forms behind this barrier beach. West of Abbotsbury, wave-energy distributions are modified by refraction and higher energies occur much closer to the eastern end. Solifluction deposits and landslide toes begin to be trimmed by the rising sea and gravels begin to travel eastwards.

5. Between 6500 and 4000 years BP, the lagoon is closed and a freshwater lake forms in the East and West Fleets. There was probably dry land between Weymouth and Portland, and The Fleet extended farther west to Burton Mere. The implicit shelter needed for peat deposition is taken as an indication that by 4000 to 5000 years BP, Chesil Beach had formed close to, or a short distance seaward of, its present position (Bray, 1990b)
6. As sea level rose, between 4000 and 2000 years BP, tidal conditions developed in The Fleet. There was now a continuous beach to Portland from Abbotsbury and also a separate beach facing eastwards into Weymouth Bay.
7. From about 2000 years BP to 1850 AD, marine erosion to the west supplies increasingly large volumes of clasts to the beach, which therefore grows in volume and height to establish the current form. The beach became a shingle ridge overlying a finer-grained, impermeable core. This continued until the mid-19th century.
8. From the mid-19th century, the building of the Cobb and West Bay harbours cut off longshore sediment transport at the same time as the beaches became more fragmented by the gradually emerging headlands at locations such as Golden Cap. Combined with the effects of beach mining and attrition, this leaves Chesil Beach as a relict feature suffering slow decline in volume.

This scenario implies that the classic model of a transgresing gravel beach could be replaced by a two-phase model in which the early Chesil Beach is a low sand and gravel barrier that provided a base upon which the more massive gravel and cobble structure was constructed as large supplies of these materials became available. Better understanding of the wave climate of Lyme Bay still leaves questions about the effect over many centuries of major events, such as those of 13 December 1978 (a 1 in 50 year event with 9 m swell) and 13 February 1979 (when 18-second period waves arrived without warning out of a moderate sea).

Despite its magnitude, it is likely that in common with many other coastal features Chesil Beach at a comparatively late stage in its evolution. Carr and Blackley (1973) suggest that during the last interglacial, any proto-Chesil Beach must have been entirely reworked by marine action. Such may well be the ultimate fate of the present-day structure.

Screening of coarse indigenous shingle for gabion and mattress-fill for the trial length completed in November 1981 could also have had an effect. It is difficult to determine the proportion of beach volume lost by extraction, but Carr (1981) estimated that between the mid-1930s and 1977, something of the order of 2% was likely to have been removed overall. Although more research needs to be carried out to determine how losses through attrition compare with this, it is very unlikely that the latter are as great on this essentially flint and chert beach.

Most anthropogenic pressures on the beach have been concentrated at the extremities (Carr, 1983a). Those at the Portland end are most critical because it is there that the beach is subject to maximum fetch from the Atlantic Ocean, and, of scientific importance, the beach crest is at its highest and the rate of change in grading is greatest along the most easterly 2 km.

There can be little doubt that Chesil Beach is in a fragile state and is finite in amount. No more material is being supplied and loss continues through attrition and removal offshore. The logical prediction must be that Chesil Beach will now steadily move onshore and break up into separate beaches and bays. It may rotate at Weymouth, breach at several places, allowing The Fleet to become saline and disappear or develop into lakes like those behind Cogden and West Bexington. The main new headland may be at the Narrows. Severe erosion will take place at East Cliff and Burton Bradstock, where the next bays will develop. The processes that happen when a barrier beach comes onshore, already seen between Lyme Regis and West Bay, will be the model for the rest of the beach.

Conclusions

Chesil Beach is a massive, linear pebble and cobble coastal barrier/tombolo backed by cliffs at its western end and a lagoon, The Fleet, in its central and eastern parts. The beach ties the Isle of Portland to the mainland of Dorset. It is renowned worldwide especially for the size grading of its constituent clasts. There are four British gravel structures that rival Chesil Beach in scale, Culbin, Spey Bay, Dungeness and Orfordness, but none displays the simplicity of form or the simple barrier shape evident at Chesil Beach. Chesil Beach is very unusual in lacking any development of recurves, even within the lagoon between Abbotsbury and Wyke Regis. The fact that it is simple in form offers enormous scope as a baseline exemplar for studying many other more complex structures.

Despite the number of cores taken through the beach and The Fleet, the internal structures remain speculative. Furthermore, the three-dimensional form of the beach, the surfaces on which it rests, the nearshore seabed and the processes that produced and maintain the beach also warrant more detailed investigation.

The origins of the beach are open to debate, but can be summarized as follows (Bray, 1990b). With sea level at the end of the Devensian about 100 m below its present-day position, a barrier beach formed as a result of the erosion of river gravels and other offshore sediments. About 7000 years BP, infilling of the Fleet began and was virtually complete by 5000 years BP (Carr, 1974). According to Bray (1990b), Chesil Beach was thus formed before there was significant erosion of the west Dorset coastline. It was then maintained by longshore sediment transport from west Dorset. An alternative view suggests a two-phase development with a sand-dominated barrier offshore upon which the cobble ridge was then established.

Even so its origins are still a matter of debate. Despite the local modifications to the beach, Chesil Beach remains a remarkable coastal landform that is regarded worldwide as having extremely high scientific value because of its form, size, composition and documentary record. Chesil Beach is included in its entirety in the Dorset and East Devon Coast World Heritage Site, declared in December 2001. Few other sites are more cited or visited by coastal scientists.

PORLOCK, SOMERSET (SS 858 484–SS 899 492)

J. Orford

Introduction

The coastal gravel barrier and beach at Porlock is the longest continuous coastal gravel barrier system on the western coast of Britain (see Figure 6.18). It is 5 km long and fronts low-lying farmland, which is being flooded daily by the tide as a result of a major breach that occurred during a severe storm in 1996. The breach has not 'healed' and has resulted in saltmarsh development and clay deposition in the back-barrier area associated with the upper elevations of a prevailing macro-tidal regime (9 m at mean high-water springs). The existence of the barrier relies upon a complex and dynamic interaction of geomorphological processes operating in an environment that is sediment-constrained and storm-wave dominated. Sporadic breaching and 'healing' events had been the natural cycle of evolution of the barrier before the onset of anthropogenic activities that attempted to stabilize the barrier and prevent it from breaching.

The barrier shows evidence of longshore segmentation, and of sediment erosion and reworking caused by the long-term failure of longshore sediment supply, which lead to well-developed swash-aligned and drift-aligned sections that are rarely found adjacent to each other on UK gravel barriers.

The geomorphological importance of, and wider interest in, the Porlock coastal gravel barrier has grown significantly over the last two decades, owing to debate about the effects of earlier coastal management strategies. There can be little doubt that the catastrophic breaching event of 1996 occurred as a consequence of barrier weakening through cumulative management activities (over several decades) intended to protect the coast. Although the barrier is returning to a natural state it is unlikely that the breach will heal, and part of its present-day interest is the development of the geomorphology following the breach.

Description

The Porlock barrier fringes the seaward edge of a coastal embayment at the eastern end of the Exmoor plateau (Figure 6.18). The crest height west of Gore Point is typically 5–6m OD, and 12 m OD to the east, towards Hurlstone Point. The local lithologies west of Porlock are red-purplish or grey fine-grained sandstones, and thick flaggy grit beds interspersed with shale and pebble beds. All of these lithologies are represented in the sediment of the barrier. The pattern of sediment facies (shape and size) differentiation across the barrier (cf. Bluck, 1967) is well developed, but has been disturbed locally by recent management activity.

The coastal edge of the Exmoor plateau (*c.* 40 km long) shows a distinctive 'hog's-back' form with a 350 m fall from the plateau top to sea level, with a steep and unstable convex coastal slope (Arber, 1911) *c.* 1 km wide. Arber commented on the thick wooded cover of the coastal slope, but during the 20th century anthropogenic disturbance of this cover resulted in an increase in the frequency of coastal landslides in the unconsolidated sediment. The sediment size arriving at the hog's-back shoreline from these slides has a considerable range from blocks (>20 m) to mud. The Porlock foreshore shows a well-developed outer boulder frame in the intertidal zone for all but the last kilometre of its proximal (eastern) end, suggesting the landward retreat of sections of the barrier. The recent increase in sediment supply from landslides has not alleviated the sediment deficit characteristic of the barrier, so it is likely that the apparent surplus of sediment west of the Porlock barrier is a recent phenomenon and not the typical situation throughout late Holocene times (i.e. over the last 2000 years), although landslides will have supplied sediment episodically in the past history of the feature.

The Exmoor coastal slope was covered by Devensian solifluction sediment, forming fans that coalesced at the foot of the plateau, which have been reworked by Holocene relative sea-level rise. The present barrier is transgressing across such a fan at Porlockford, and the western end of the barrier is developing on top of the intertidal scar of an eroded fan (Gore Point).

The rising ground of the relict solifluction fan at Porlockford cliffs controls the present-day embayed position of the barrier. Fan resistance to wave action can lead to barrier segmentation around the flanks of the fan. This western foreshore shows a mixture of boulder scars as the rising sea level has eroded its way through the old fans, and a lag of large boulders (outer boulder frame) that resist swash action and incorporation into the barrier. The frame dissipates wave energy but its low position (<–1 m OD) means that it does not protect the barrier during high tides and surges. The exposed position led to attempts to build up the barrier crest east of Porlockford and New Works in the early 1990s by reshaping the profile (making it higher to *c.* 8 m OD) but thinner (Figure 6.18C; profile P4); replacing washover sediment back onto the ridge top; and renourishing the ridge sediment volume with dredge material (gravel, sand and mud) from the entrance of Porlock Weir harbour. Attempts to export down-drift gravel back to this swash-aligned section were successfully resisted by the National Trust, the landowners of the easterly drift-aligned section. The weakness of this swash-aligned bay was the core management problem in the 1990s, as potential barrier breakdown and associated back-barrier flooding were perceived to be central issues for local economic well-being and aesthetics.

Between Porlock Weir and the tidal sluice gate (New Works; Figure 6.18B), the continuous easterly longshore depletion of sediment has reduced the volume of the barrier and accelerated its retreat onshore by rollover (where sediment is carried over the ridge top by storm washover), forcing the embayed barrier into a more swash-aligned structure. The Porlock barrier has retreated most at this section. Here, the barrier is low (*c.* 6–7m OD) and narrow (<40 m wide), commensurate with the loss of volume as the barrier is 'stretched' between Porlock Weir and New Works without new material being added.

Although New Works appears to have been built at the transition between the up-drift swash-aligned and down-drift drift-aligned sections, it is unlikely that the works *per se* were of a sufficient age to have played a part in the development of this transition point. New Works was probably sited here because of its low elevation with respect to back-barrier drainage.

After 1950, the ridge east of the New Works area was in constant need of remedial attention due to washover forcing the remnant ridge into a swash-aligned posture commensurate with the westward barrier. The barrier was rebuilt on several occasions: by regrading washover fans; adding new gravel (from the old recurves) to the upper beach; and installing groynes to reduce westward movement of sediment. All this activity reflected the continuing movement of gravel eastwards as the swash-aligned section was

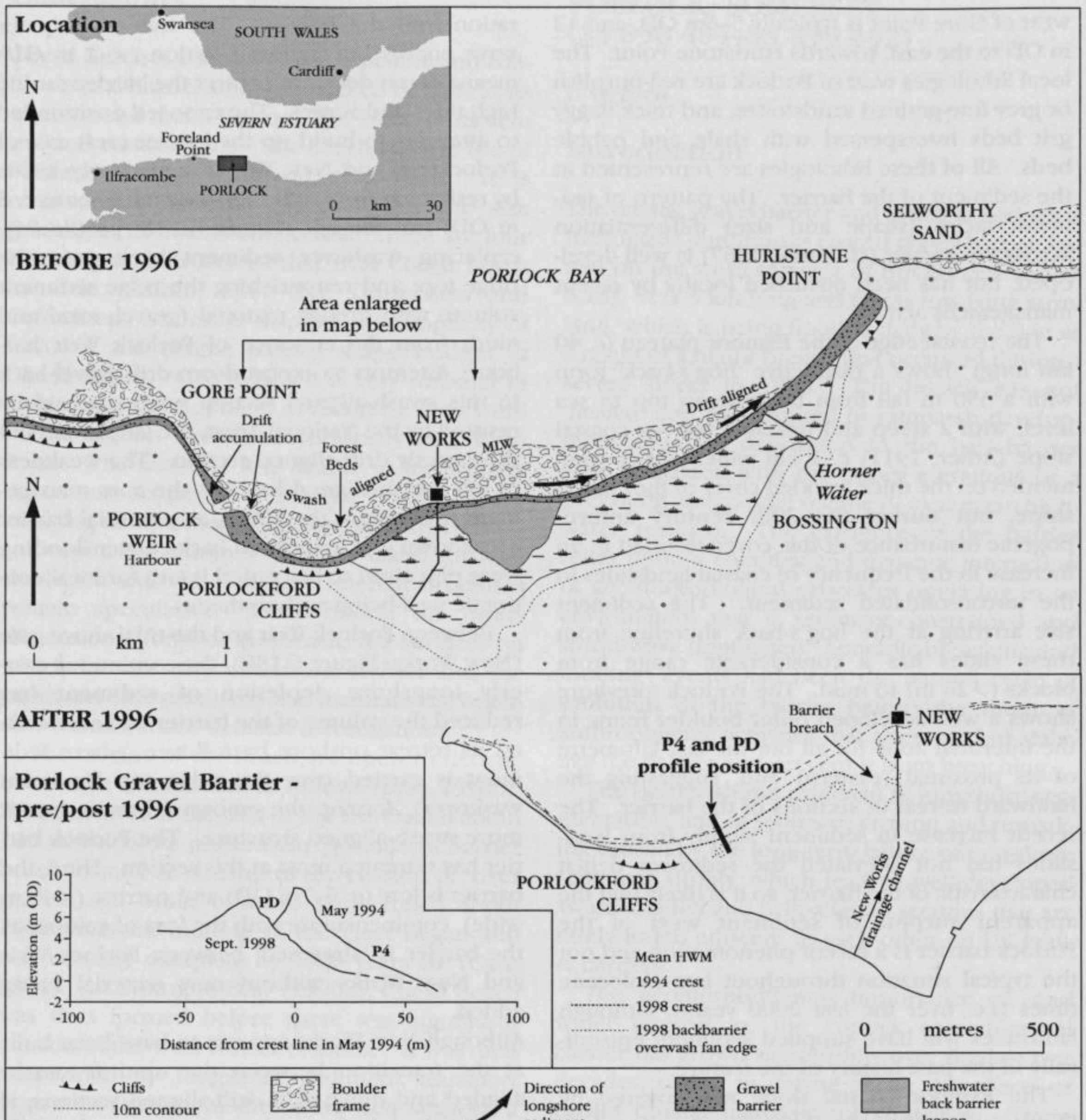

Figure 6.18 (A) Porlock barrier and back-barrier; (B) barrier crest and back-barrier changes before and after the 1996 barrier breaching; (C) barrier profile changes due to the 1996 storm.

evolving further eastwards. The barrier has retreated marginally at this section since the 1980s and is currently held by old shore-parallel stone walls now being consumed by barrier rollover. This anthropogenic intervention (stopped in 1990s) makes it difficult to determine the natural position of the barrier in the decade before 1996.

During late October 1996 Hurricane Lillie moved across the North Atlantic Ocean and degenerated into a deep depression before moving across southern Ireland and the UK. Storm surge levels superimposed on the high tide exceeded the height of the managed barrier

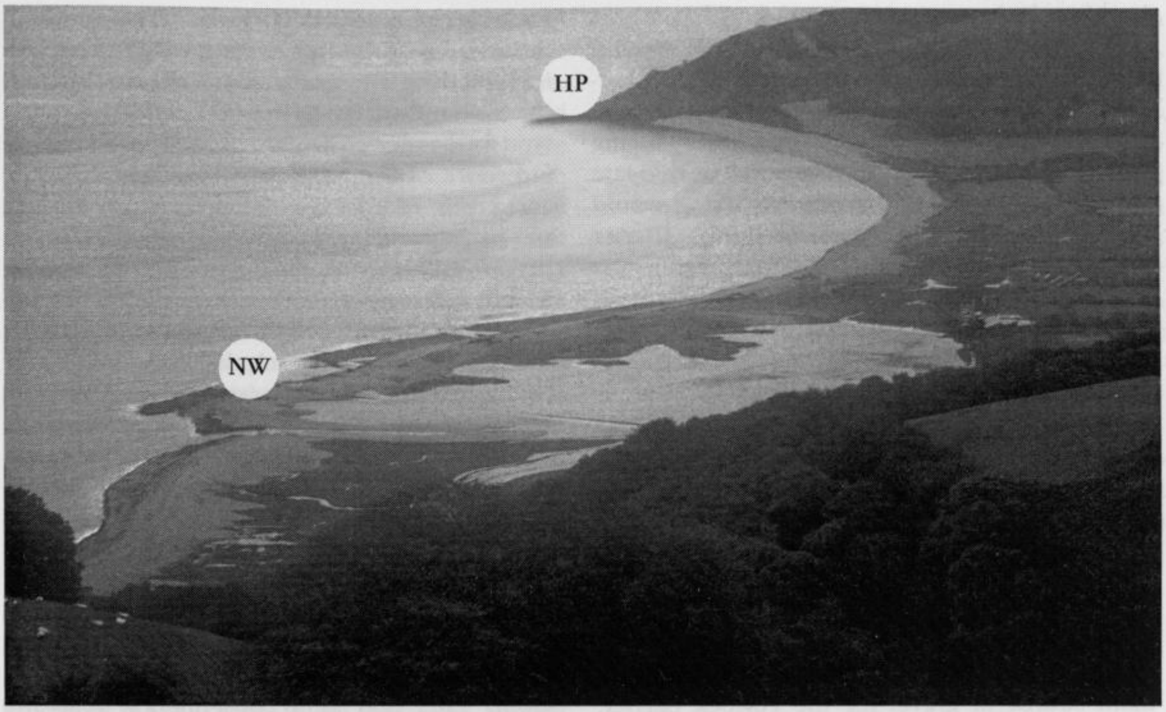

Figure 6.19 Overview of Porlock barrier (October, 1997) looking east. The 1996 barrier breach is identifiable, as are the storm generated washover fans at point NW (New Works sluice gate); HP is Hurlstone Point. (Photo: W. B. Whalley.)

between Porlockford Cliffs and New Works on October 26th. Massive overwashing demolished the barrier crest and moved gravel onto the back-barrier area. The volume of overwashing waters was sufficient to fill the back-barrier area and during the falling tide, forced a breach west of New Works. The extent to which the gravel ridge was also pushed back has been partially identified by differences in crest surveys undertaken by Orford *et al.* (2001) in July 1994 and September 1998, and in a series of detailed measurements post-1999 by Bray and Duane (2001). In outline the original managed crest of the barrier was demolished and pushed back in the form of classic washover fans (Figure 6.18B). During February 1997, a series of major westerly cyclonic storms helped to reconsolidate the beach ridge at a new position 15–25m farther landwards.

Since 1999, ridge changes west of New Works have been slight, the lack of movement and restructuring of the ridge into a broader and lower (*c.* 6–7 m OD) feature suggests a period of stability. The barrier will continue to retreat as a function of the elevation reached by extreme run-up of breaking waves, which will also increase with future storminess and relative sea-level rise. Past lowland management and land-use in the area has meant that there is space for barrier retreat without any overwhelming demand for coastal protection. In order to understand the development of gravel barriers that are not impeded by protection measures, Porlock is a key 'open air laboratory' site that will enable geomorphologists to better predict the outcome of management activities on gravel barriers. The freedom to evolve now will mean the site is once again moving towards a more 'natural' pathway and therefore a key site to see how a barrier naturally responds to past anthropogenic forcing.

Interpretation

The barrier probably initially formed as a drift-aligned spit building eastwards; remnants of the recurves from this sediment-surplus phase can be identified at positions between New Works and the old limekilns near Horner Water, but there is no available dating evidence for this barrier growth phase. As sediment supply from the hog's-back area diminished owing to a reducing rate of the rise of relative sea level in mid-late Holocene times, then barrier reworking would have taken place (Orford *et al.*, 1996). Barrier sediment from updrift positions (west) was transported to continue the spit growth at its eastern end, until reaching Hurlstone Point (which acted as a natural groyne) and sediment was trapped in major beach ridges. These latter forms have been steadily drift-aligned (20 m accretion between 1880 and 1980) into high (12 m OD) beach ridges. Their seaward growth means that there is no shore platform exposed at low water and thus bigger waves can approach closer inshore without dissipation, allowing storm waves at high tide to generate swash sufficient to move gravel to the crest top.

The west–east Exmoor coastal slope is the main source of gravel for the Porlock barrier. There is a net easterly beach drift system powered by Atlantic swell waves and depression-generated storm waves running from the north-west to west into the Severn Estuary. The slight shift in coastal alignment east of Foreland Point might indicate that the contemporary source area for Porlock is somewhat less than the whole Exmoor length. There is little fine-grained sediment in the Porlock barrier system because much of the sand-sized load is moved offshore, or deposited to the east of Hurlstone Point. It is feasible that Holocene relative sea-level rise has now eroded the low-elevation fans at the foot of the coastal slope and it is the reduction of this source that has now pushed Porlock's sediment budget into deficit. The 20th century rise in sediment supply (due to up-drift changing coastal slope management practice) seems only to be reaching as far as Porlock Weir. The cannibalization of the Porlock barrier and development of the major swash-aligned unit deep into the western end of the Porlock embayment means that sediment transport from the old western source area has now virtually ceased. This has not been the case in the past when the barrier was probably developed as a spit extension of the coastal plateau edge into Porlock embayment.

Jennings *et al.* (1998) explored the relative sea-level rise identified by a palaeo-ecological reconstruction of organic deposition found exposed in the foreshore, and in the pre-1996 back-barrier marsh at Porlock. Their results fit into the broader relative sea-level rise envelope identified by Heyworth and Kidson (1982) for the Severn Estuary. Tree roots bedded into a fen-type freshwater environment are exposed on the foreshore of the western swash-aligned barrier, and cores taken from the pre-1996 seasonally wet back-barrier area (now tidally flooded) identify a mixture of environments related to whether the back-barrier area was open to intermittent tidally induced flooding and mud deposition, or was closed such that a freshwater fen environment was generated. The thesis of Jennings *et al.* (1998) was that barrier coherence (hence barrier strength to resist breaching) and potential for tidal incursions into the back-barrier area was related to the rate of relative sea-level rise. They suggested three domains of barrier activity related to relative sea-level rise, noting that as the rate of rise decreased there would be an increasing tendency for spatial stability of the barrier, a decreasing longshore sediment supply rate, and as a consequence a greater potential for the barrier to be cannibalized and disturbed sufficiently to allow storm breaching sufficient for tidal incursion. Prior to 7000 years BP, relative sea-level rise rates of *c.* >6 mm a^{-1} were likely to generate high longshore sediment supply but reduce the ability of the barrier to maintain any longshore coherency sufficient to act as a barrier to tidal influence in the back-barrier area – hence the evidence of fine-grained, marine, back-barrier deposition. Between 7000 and 5000 years BP, relative sea-level rise rates of 6–2mm a^{-1} allowed sufficient sediment to seal any barrier breaching, while the spatial translation of the ridge was so reduced that the barrier maintained its coherency to act as a buffer to saline influences, thus letting a back-barrier freshwater regime operate in which fen-deposition predominated. When storm breaches were intermittently open, the back-barrier area was exposed to saline waters, thus allowing thin marine mud-sequences to be inter-digitated with organic fen materials. By late Holocene times (<2000 years BP) decelerating rates of relative sea-level rise (<1 mm a^{-1}) reduced the longshore sediment supply allowing barrier breaches to remain open and the fen to be replaced by tidally dominated back-barrier sedimentation.

The 1996 breach is unlikely to be sealed by the existing low longshore transport rates. It has been widened and deepened with headwater erosion by tidal flows into the consolidated

Holocene clays of the old back-barrier. Bray and Duane (2001) have mapped the retreat of the breach and the extension of the inlet sidebars. They have also monitored tidal elevations and associated sedimentation within the now active upper tidal frame behind the barrier. The high turbidity of the water column in the Severn Estuary and its macro-tidal regime ensure that fine-grained sediment enters behind the barrier on almost all tides. Annual deposition rates of 10 mm a^{-1} measured during 1999–2000, suggest that this is a site of great potential for saltmarsh development. Whether this sedimentation rate is the initial result of a forced change in the system and will decline as the back barrier adjusts to the new tidal regime is uncertain. However the lack of coastal squeeze at this site suggests that this will be an important test site for evaluating saltmarsh growth and adjustment to accelerating relative sea-level rise in future decades. Bray and Duane (2001) also underline the potential for barrier change immediately east of New Works. The implications for further breaching in this area are intriguing, though current inlet efficiency may be reduced if more breaches occur, thus limiting more persistent breaches.

The historical state of the barrier is unknown, though the 'stabilized' barrier during the 20th century produced a local view of back-barrier stability that has not been the norm for most of its Holocene existence. The tidal sluice-gate system (New Works) indicated the measures taken to ensure the drainage of freshwater from the back barrier to allow pasture development. Even then, the pre-1996 intermittent wetlands identified a problem in fully draining the back-barrier area and the resultant small mere and fen that did develop provided the interest for the original Porlock SSSI biological designation. This past anthropogenic intervention has flavoured perspectives as to how Porlock barrier should be managed, although the latest storm-forced changes to the managed barrier section show that a natural mode of barrier evolution is now appropriate, and that this should be of prime concern in future management strategy.

Conclusion

The Porlock barrier is central to geomorphological studies into how a freestanding gravel barrier responds to relative sea-level rise and storminess changes. The barrier is likely to remain a centre of coastal interest for its combination of evolving swash-aligned and drift-aligned longshore sections; its post-management adjustment to a stable cross-barrier profile in relation to relative sea-level rise and storm-wave climate; its rollover dynamics and washover response; its breaching behaviour; its developing tidal inlet control on barrier segmentation and longshore transport impedance; its back-barrier fine-sediment deposition and saltmarsh development. It represents one of the best UK examples of how managed 'stabilized' barriers are non-sustainable at the decade-timescale. It also exemplifies the likely mode of barrier failure, if coastal gravel-dominated barriers are not allowed to adjust freely to changing relative sea level.

HURST CASTLE SPIT, HAMPSHIRE (SZ 310 900)

V.J. May

Introduction

Hurst Castle Spit extends the shingle fringing beaches at the eastern end of Christchurch Bay across the western arm of the Solent (see Figure 6.2 for general location). Its seaward end is marked by Hurst Castle, constructed in the mid-16th century, and threatened from time to time by erosion since the mid-19th century. The spit protects Keyhaven Marshes (see separate GCR site report in Chapter 10 of the present volume).

Following the seminal paper of Lewis (1931) in which he argued that beaches align themselves at right angles to the direction of approach of dominant waves, Hurst Castle Spit is often used as an example of a multi-recurved spit (Johnson, 1919; Wooldridge and Morgan, 1937; Steers, 1946a; Sparks, 1960; Bird, 1968; King, 1972b; Komar, 1976; Bird and Schwartz, 1985). King and McCullagh (1971) used Hurst Castle Spit as the basis for an early computer model – 'Spitsim'). More recently, Clark and Small (1967), Clark (1974), Nicholls (1984, 1985) and Nicholls and Webber (1987a–c, 1989) re-examined its features. Nicholls and Webber suggest that this is not a complex recurved spit, but owes its detailed form to variations in local sediment supply and to changes in sea level that had not been considered in previous work.

During the winter of 1989–1990 the whole ridge was overtopped and moved inland by up to 80 m. The risk to the recently completed coast protection works at Keyhaven and Lymington, the ecologically important saltmarsh behind the ridge and low-lying residential areas from flooding was judged to be so severe that major coast protection works were put in place during the late summer and autumn of 1996. 120 000 tonnes of imported Norwegian rock (with boulders up to 1.5 m across) were placed along a 550 m-long section of the proximal end of the spit. 500 000 tonnes of shingle dredged from the Shingles Bank were placed along the remaining length of the barrier beach to double its width and raise it to a height of 7 m (see Figure 6.21). A rock revetment 100 m in length was constructed to protect the western wing of Hurst Castle and regrettably shingle was excavated from the most recent distal recurves to be placed along the remainder of the frontage of the castle. As a result, despite the recognition of the importance of this GCR site, much of its natural interest has been removed. Although much weakened at its proximal end by erosion combined with a reduction in the littoral supply of shingle as a result of coast protection works, the spit nevertheless retains its characteristic form.

Description

The natural spit had two main parts (Figure 6.20):

1. Hurst Beach, a single transgressive shingle ridge orientated towards the dominant south-westerly waves of Christchurch Bay. Along much of its inner length, the shingle rests upon saltmarsh and earlier gravels which are occasionally exposed on the foreshore. Its maximum height varies between 3 m and 5 m OD. The direction and energy distribution of the waves approaching the beach is affected by both the Isle of Wight and a shallow offshore shingle shoal, known as the 'Shingles Bank'.
2. An active recurve, behind which are three groups of preserved recurves, which may pre-date the construction of Hurst Castle (AD 1541–1544). The recurves are aligned towards the dominant north-easterly waves of the western Solent.

The beach is formed mainly of subangular to subrounded flint pebbles with subsidiary fine- to medium-grained sand, derived from the erosion of Pleistocene sandy gravels farther to the west. Net littoral drift is towards the east, although much of the sand is lost offshore. Nicholls and Webber (1987a) reported a littoral sediment transport sub-cell boundary at Hordle Cliff to the west. This, they believed, means that the spit is much more dependent upon local sources of sediment than earlier writers had supposed. At Hurst Point, much of the shingle is lost offshore into Christchurch Bay or the western Solent. The nearshore slope is steep and tidal energy in the western Solent is high, the tidal streams between Hurst Point and the Isle of Wight attaining 2.3 m s^{-1} at spring tides and so capable of moving small shingle. Dyer (1970) reported that much of the material forming the Shingles Bank is similar in composition to that forming Hurst Beach and most writers agree that it is likely that much of the sediment lost from Hurst Point reaches the Shingles Bank.

The shingle ridges forming the recurves are generally about 1 m lower in altitude than the transgressive Hurst Beach and the present active recurve, except over the last 100 m at its distal end. Within the main area of shingle north of Hurst Castle it is still possible to identify two strongly recurved distal ridges, which suggest that this area includes at least three ridge groups. The innermost of these must pre-date the construction of the castle, and the second may pre-date the extension of the castle between 1861 and 1873. The second group of ridges are not continuous, having been removed in their central section by erosion. At present the active recurve appears to be marked by zones of accretion at Hurst Point and its distal end with a zone of erosion between them (Figures 6.20 and 6.21).

The landward end of the spit was most recently armoured by boulders in 1996 and both beach replenishment and reshaping of the ridge have been adopted as measures to prevent breaching of the ridge. The beach fronting the castle has been subject to considerable erosion and the foundations of the outer works of the castle have been undermined. A series of groynes had been in place for several decades, but these had decayed badly and were renewed by English Heritage during the late 1980s together with a programme of beach replenishment using shingle from the zone of accretion at Hurst

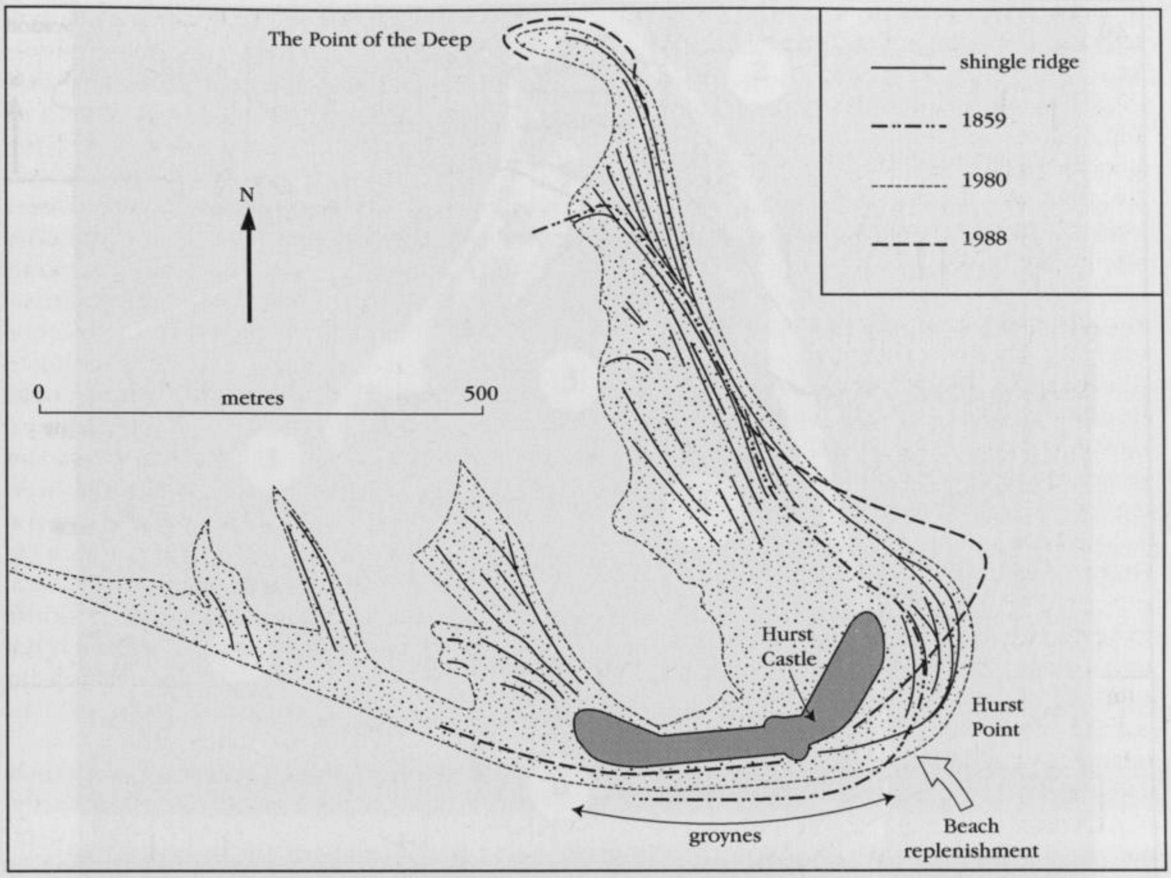

Figure 6.20 Distal recurves at Hurst Castle Spit – the history of geomorphological development. (After Nicholls and Webber, 1987a.)

Point. Further replenishment was needed by 1996.

The natural spit was affected by several processes. Hurst Beach is transgressive (Figure 6.21), moving over the saltmarsh and Pleistocene gravels, which were reworked, providing a local source of shingle. The transgressive ridge is affected by overwashing and seepages, both of which can lower the ridge crest. Some shingle is moved along the spit with littoral drift increasing towards Hurst Point where it attains a maximum of 15 000 m^3 a^{-1} for the shingle fraction alone (Bray *et al.*, 1992). Shingle is moved seawards from Hurst Point onto the Shingles Bank, which, as it changes shape and height, affects the wave energy distribution along Hurst Beach (Nicholls and Webber, 1989). North of Hurst Point, the active recurve continues to grow towards the Point of the Deep, but this appears to be at the expense of the central part of the active recurve.

The spit protects a large area of saltmarshes, known as 'Keyhaven Marshes' (see separate GCR site report in Chapter 10 of the present volume) which are drained by an intricate pattern of creeks dominated by three major creeks: Mount Lake, alongside the spit, Keyhaven Lake and Hawkers Lake. The first two merge and drain into the Solent after being diverted by the modern recurves of the spit. Active marsh-edge beaches (cheniers) are formed mainly of shells and shingle with a low sand content. Much of the saltmarsh edge is being eroded rapidly, resulting in patches of unvegetated mud. The surface of the marshes is characterized by a high proportion of eroded marsh, pans, and broad channels. There are only small areas of higher-level, species-rich saltmarsh, located mainly

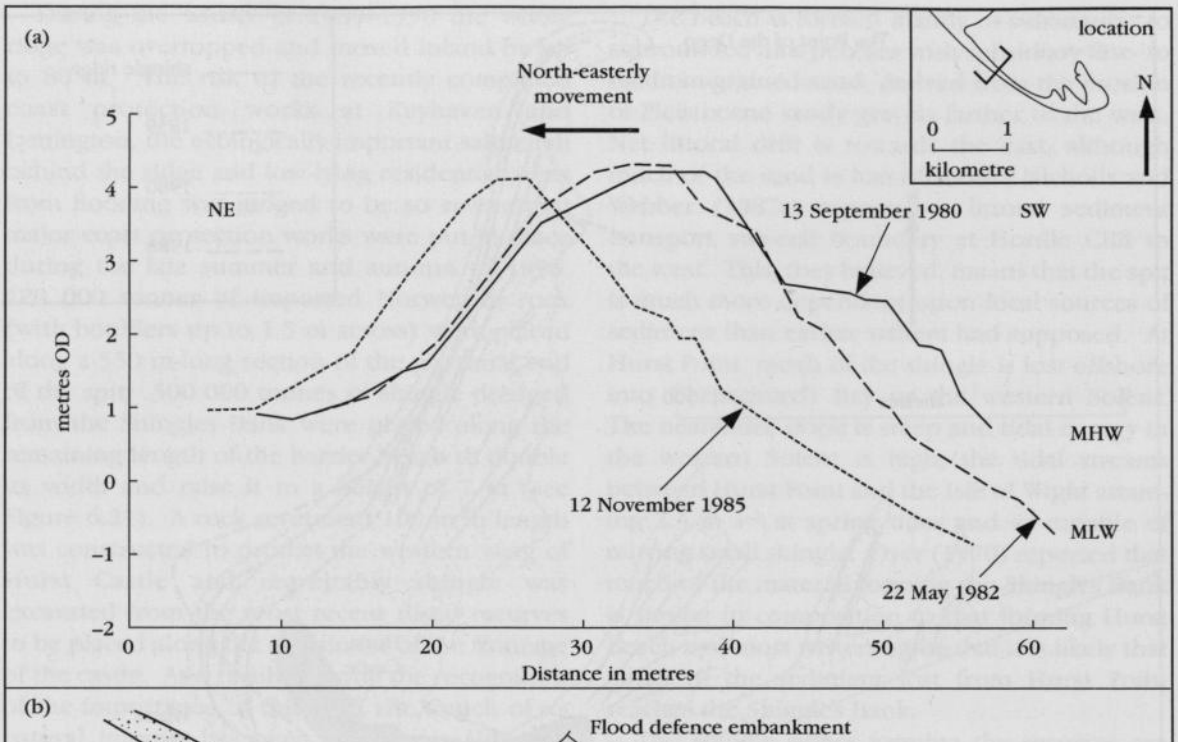

Figure 6.21 (a) Changes in the profile of Hurst Castle Spit. (After Nicholls and Webber, 1987a.) (b) 1996 coast protection works at Hurst Castle Spit. The pecked line in (b) delimits the saltmarsh edge.

close to the spit and on its older recurves. Sea purslane *Atriplex portulacoides*, common sea-lavender *Limonium vulgare*, sea plantain *Plantago maritima*, sea meadowgrass *Puccinellia maritima*, annual sea-blite *Suaeda maritima*, samphire *Salicornia* spp., and sea aster *Aster tripolium* are common throughout these higher marshes. In contrast, the lower more extensive marshes are species-poor and dominated by common cord-grass *Spartina*

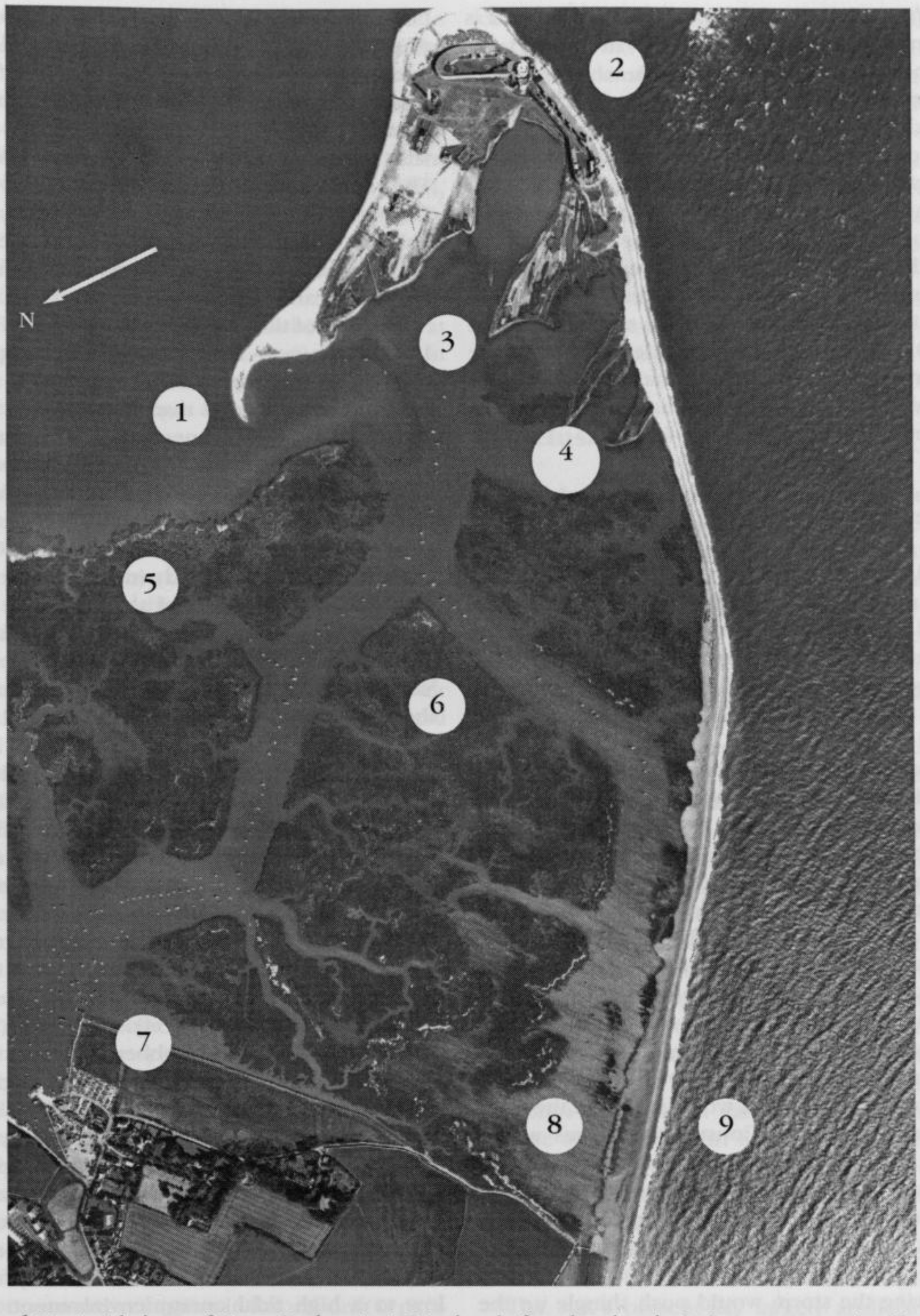

Figure 6.22 Aerial photo of Hurst Castle spit. 1. Distal end of modern beach; 2. Groynes protecting Henrician (16th century) castle; 3. and 4. earlier recurves; 5. saltmarsh – the seaward edge of saltmarsh is undergoing retreat; 6. *Spartina anglica*-dominated saltmarsh, declining in area; 7. coastal defences at Keyhaven; 8. most commonly overtopped and artificially rebuilt section of beach ridge; 9. waves approaching from south-west. For discussion of the saltmarsh features, see GCR site report in Chapter 10 for Keyhaven Marsh. (Photo: courtesy Cambridge University Collection of Aerial Photographs, Crown Copyright, Great Scotland Yard.)

anglica. The intertidal area close to the spit is often a stony mud (see GCR site report for Keyhaven Marsh in Chapter 10).

Before the late 19th century, much of this marsh lay as much as 1 m lower and was dominated by eel grass *Zostera*. Colonization by *Spartina anglica* following its hybridization from the native *Spartina maritima* and the introduced *Spartina alterniflora* in Southampton Water led to a rapid build-up of the saltmarsh surface. The area of *Spartina*-dominated saltmarsh reached a maximum about 1930, after which the area declined. As the recurves of the modern spit have extended into the westernmost creek, they have increased local accretion of mudflats.

Interpretation

Ward commented (1922, p. 114) that 'for many centuries, the spit has in size and position suffered no substantial permanent change, so that accretions of fresh shingle must, now that the period of growth is over, be balanced by wastage from the spit…'. Both the history and the future development of Hurst Castle Spit depend upon the balance of sediment supply and loss. Lewis (1931, 1938) showed that the main ridge was aligned at right angles to the direction from which storm waves approach along the English Channel with a fetch extending across the Atlantic Ocean from where the strongest winds and waves also come. The recurves face the waves from the north-east along the Solent. Lewis suggested that the spit grew towards the south-east as shingle was moved by oblique waves alongshore from the eroding coastline of Christchurch Bay. Large waves from the direction of maximum fetch would build up the main part of the ridge. Constructive waves would build the beach up to high tide level and storm waves would deposit shingle onto the crest of the ridge, building it above the reach of normal waves. Storm waves would also comb shingle down onto the lower beach. Constructive waves following the storm would push shingle up the beach. The result of this combined wave action would be a gradually consolidated spit probably moving slowly towards the north-east. Lewis considered that, as the spit curved more towards the east and the angle of waves to the beach changed, the rate of drift would increase. Shingle would move rapidly around Hurst Point, but travel more slowly along the recurves. As water depth increased, more shingle would have been required to extend the spit. Refracted waves would move some shingle along the recurves, while north-east waves in the Solent with their longer fetch could build up the ridges. The distinctive sharp angle at Hurst Point (Figure 6.22) is the result of the main ridge-building waves being restricted by the shelter of the Isle of Wight to two main directions, the south-west and the north-east.

The role of tidal streams was thought by Lewis 1931 to be very small, though affecting wave height and the angle of wave approach to the beach. He concluded that the tidal role in the growth of the spit was negligible. This was contradicted by Williams (1960) who claimed that the effect of east–west tidal streams in Christchurch Bay could explain the eastward growth of the spit. Clark and Small (1967) concluded from a drifter study in Christchurch Bay that seabed movements of shingle could occur in depths as great as 20 m both directly onto the spit and across the floor of the bay towards Mudeford, from where they could move alongshore back towards Hurst Castle. King (1968b) and King and McCullagh (1971) followed Lewis' (1931, 1938) hypothesis for the growth and shape of the spit in their development of a computer model to simulate the effects of different wave directions, refraction and increases in the depth of water. Clark (1974) questioned whether the spit was either a complex recurved spit, as had been commonly accepted, or merely a storm beach resting on the edge of one of the partially submerged former terraces of the Solent River (Everard, 1954). Nicholls (1984, 1985) and Nicholls and Webber (1987a–c, 1989) reviewed the history and present development of Hurst Castle Spit arguing that previous models of its evolution emphasized longshore growth at the expense of other factors, especially changes of sea level. The submergence of a Pleistocene valley system along the line of the present-day western Solent brought about a major transformation of Christchurch Bay from a low to a high tidal energy environment. The Shingles Bank was particularly important because, in refracting waves crossing it, it influenced wave energy along Hurst Beach. However, the date of its formation remains unclear. If it acts as a sink for shingle from the spit then it may be relatively recent. Substantial local supplies of sediment were also available from Pleistocene sandy gravels, which lie both in the cliffs at Milford and beneath the root of the spit.

Nicholls and Webber (1987b) suggested that the second youngest recurve formed during a possible period of sea-level still-stand between about 4500 years BP to 3000 years BP (West, 1980; Devoy, 1982; Nicholls and Clark, 1986). Nicholls and Webber (1989) acknowledge that sub-shingle sediment compaction may be important in affecting ridge height and follow Lewis (1931, 1938), Lewis and Balchin (1940: at Dungeness) and Carr (1970: at Orfordness) in their interpretation of increasing height with decreasing age of ridges as indications of a rising sea level.

On Hurst Beach, there are two types of overwashing: crest-maintaining overwashing and throat-confined overwashing (Nicholls and Webber, 1989). The backslope of the beach ridge is affected by collapse features associated with seepage, but this does not appear to affect the development of washover throats as has been reported elsewhere (Eddison, 1983b; Carter *et al.*, 1984). Nevertheless, the washover throats are significant forms. The landward movement of shingle during a single storm was estimated by Nicholls and Webber (1989) as twenty-five times greater in throat-overwashing than in crest-maintaining overwashing. Washover throats have been observed to have a preferred longshore spacing, and wave run-up has exhibited regularly-spaced maxima along the shore, implying edge waves or some similar effect.

The present-day processes described by Nicholls and Webber at Hurst Castle Spit suggest that much of the spit should be regarded as a barrier beach rather than as a spit in the traditional sense, because it has not arisen simply as a result of longshore transport. Nevertheless, the alignment of the transgressive beach and the recurves depend upon the two different sets of dominant waves from the south-west and the north-east, as described by Lewis. Sediment was derived from erosion of the shore to the west. The present-day reduction in littoral drift is partly a result of the protection of the coastline at Milford and the construction of groynes. Longshore input of sediment has now been reduced and some beach replenishment has been undertaken. Beach nourishment will have to be repeated in the future, as the decision to protect Hurst Castle itself requires stabilization of that part of the beach. Much of the surface of the shingle has been disturbed and the recurves have been damaged by vehicles. Parts have been affected by the various phases of castle and lighthouse construction. The interpretation of the history of the spit during the Holocene Epoch will depend upon investigation of the recurves, and considerable care will need to be taken to avoid further damage. The main ridge is now almost entirely managed, but provides an interesting situation for case-studies of the effects of diminished littoral transport, and barrier management by beach replenishment and rock armouring.

Conclusions

Although Hurst Castle Spit is commonly regarded as a classic example of a complex recurved spit, recent work suggests that it was initiated in response to a combination of processes, among which a rise in sea level appears to have been most important. It may thus be better interpreted as a form of barrier beach. It is nevertheless a very important site because of its seminal role in coastal studies and its simplicity of form when compared to other shingle beach sites such as Orfordness and Dungeness. Its place in coastal studies rests upon the classic studies of Lewis (1931, 1938). The revised model of Nicholls and Webber is a development of earlier work and offers a basis for understanding of other shingle beaches. Studies of the detailed forms and processes on Hurst Beach have brought about a better understanding of shingle beach development. Hurst Castle Spit, Orfordness and Dungeness each display different aspects of shingle spit and barrier beach development, but the similarities of ridge patterns and their relationships to sea-level variations make them a unique suite of sites in Europe. The spit provides shelter for the intricate system of creeks in the very important Keyhaven saltmarshes where *Zostera*-dominated marshes have been colonized by *Spartina anglica.* The site forms part of a Ramsar site and both an SAC and an SPA.

The spit's present-day largely artificial form provides an opportunity to analyse changes from a natural feature to a mainly anthropogenically influenced landform, and research into its re-establishment of equilibrium conditions will be of significant interest, particularly when compared with the formerly managed Porlock GCR site.

PAGHAM HARBOUR, WEST SUSSEX (SZ 880 960)

V.J. May

Introduction

Pagham Harbour (see Figure 6.2 for general location) is the easternmost of a series of drowned river valleys and shallow estuaries that characterize the coastline of southern central England. They include Poole Harbour in the west, the Solent, and Langstone and Chichester harbours. With the exception of the Lymington River and the Medina estuary at Cowes, they all have sand or shingle spits at their mouths, and many are distinguished by double spits extending from both sides of the estuary. The shingle spit across the mouth of Pagham Harbour comprises a series of sub-parallel shingle ridges and recurves, which mark different phases of extension and accretion. Shingle reaches the beach via the intertidal zone. The behaviour of the spit and the so-called 'Pagham Delta' (an area of deposition associated with the mouth of the estuary) are intimately linked with water and sediment circulation around the Selsey peninsula. The area also provides an excellent example of the role of weed-rafting of shingle in coastal sediment budgets.

Ward (1922) and Steers (1946a) described the main features of the development of the estuary and the spit; Robinson (1955) compared the development of the Pagham site with the spits at Christchurch and Poole. Kidson (1963) challenged the hypothesis for double spit development. May (1964) showed how the spit had developed over a period of several years. The development of double spits is not uncommon at the mouths of shallow estuaries, but Pagham Harbour is distinguished by having changed from spit to bay-bar to spit again. Its double form is probably a result of breaching rather than a result of opposing directions of longshore sediment transport. The supply of shingle to the spit has been and continues to be dominated by transport from the direction of Selsey Bill (Harlow, 1979; Hooke *et al.*, 1996), supplemented by kelp-rafted pebbles (Jolliffe and Wallace, 1973).

Description

The site comprises a shingle beach that extends from the east side of Selsey Bill (SZ 870 941) to the Pagham Beach Estate (SZ 895 975), part of the Pagham Harbour estuary, and its extensive intertidal gravels. The intertidal gravels occur as irregular extensions of the beach at Inner Owers and a bank known as 'The Spit', but form a distinctive delta-like form at the mouth of Pagham Harbour. At Church Norton, the beach is formed by a number of ridges or 'fulls' (as they are called locally). Shingle is characteristically larger locally on the ridges. The shingle spit extends across the mouth of the estuary, with a series of short recurves marking periods of advance. The estuary forms the easternmost part of a valley system which extends to the west of Selsey Bill where it has been truncated by the retreat of the coastline and is now only prevented from periodic inundation by a shingle ridge. In 1910, this western ridge was breached and the sea flowed through the low-lying land into Pagham Harbour, breaching the shingle ridge, which until then completely closed the harbour entrance.

The present-day spit across the mouth of Pagham Harbour is a much smaller version of a longer feature that grew north-eastwards from at least the mid-17th century and gradually forced the outflow from the estuary farther towards Pagham, until it was breached in 1910. By 1934, two separate ridges extended from each side of the estuary. The intertidal gravels diverted the outflow towards the north-east, but ten years later the outflow was located at the south-western end of the southern ridge (Robinson, 1955). This spit gradually decayed and was replaced by a newer structure that grew from the south-west (Figure 6.23). This not only diverted the outlet north-eastwards but also changed the wave patterns within the entrance to the harbour and thus the alignment of the decaying former spit.

The intertidal area at the harbour mouth extends over 800 m seawards and is largely composed of gravel locally derived. The outflow from the harbour appears able to maintain a channel through these deposits. The intertidal area has had much the same width throughout its history, but both the outlet and the form of the delta have altered their positions. Thus, as the spit has changed its shape, so it has affected the form of the intertidal area. As a result, wave refraction has also altered.

Severe erosion along the eastern side of Selsey Bill has meant that the present area of the harbour mouth has been exposed increasingly

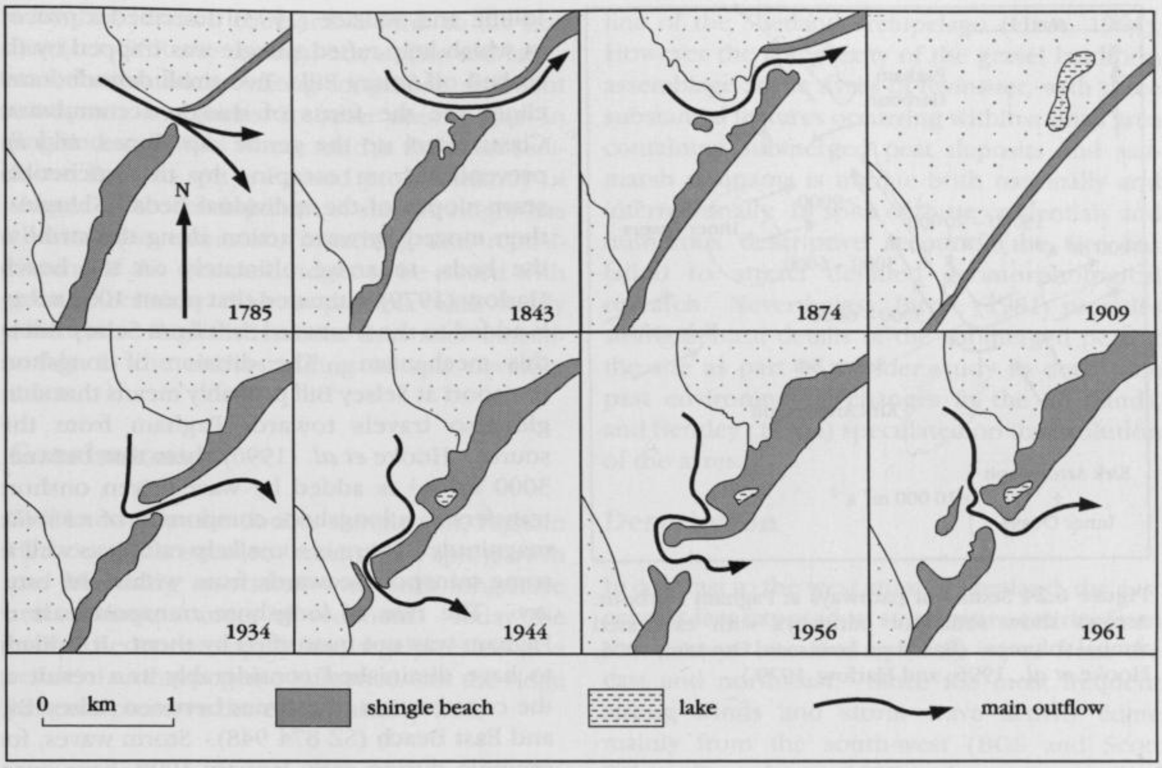

Figure 6.23 Historical changes at Pagham Harbour 1785–1961. (After Robinson, 1955.)

to waves approaching from the south. The limited documentary evidence suggests that Selsey was connected to the mainland only on its south-western side at Medmerry. Gradual silting of the harbour and its land-claim was associated with the growth of the beach across the mouth of the harbour by 1909.

Interpretation

Steers (1946a) noted the different sized spits, the larger extending from the south-west, the much smaller from the north-east. He suggested that they appeared similar to the opposing spits at the mouth of Poole Harbour and were probably of similar origin. It was not clear, however, if a local counter-movement of shingle from the north-east was responsible for the smaller spit. Robinson (1955) considered the double spits not only at Poole and Pagham, but also at Christchurch. From a detailed consideration of cartographic and field evidence, he argued that the spits resulted from unidirectional drifting followed by breaching. This model depends upon longshore transport that forms, maintains and usually extends a spit across an estuary and also diverts the river outflow. At Christchurch and, Robinson believed, elsewhere, wave and river conditions would bring about breaching, such that the distal end of the spit would eventually be attached to the mainland and its proximal end modified so that it assumed the form of a spit. At Pagham Harbour (Figure 6.24), it can be shown that the spit has normally grown from the south-west fed by the very large quantities of material eroded from the cliffs at Selsey. Between 1956 and 1961, such a phase of growth was accompanied by gradual decay and transgression of the former north-east spit on to the saltmarsh within the estuary (May, 1964). Although Kidson (1963) challenged the general applicability of Robinson's unidirectional view, he acknowledged that in some areas growth was predominantly from one direction and that, since breaching could be shown to have taken place, attached beaches could co-exist on both sides of an estuary. Robinson's discussion focused on Poole Harbour as well as Pagham Harbour, and the debate is covered more fully in the description of South Haven Peninsula GCR site in Chapter 7 of the present volume.

Davies (1972, p. 140) suggested that 'discus-

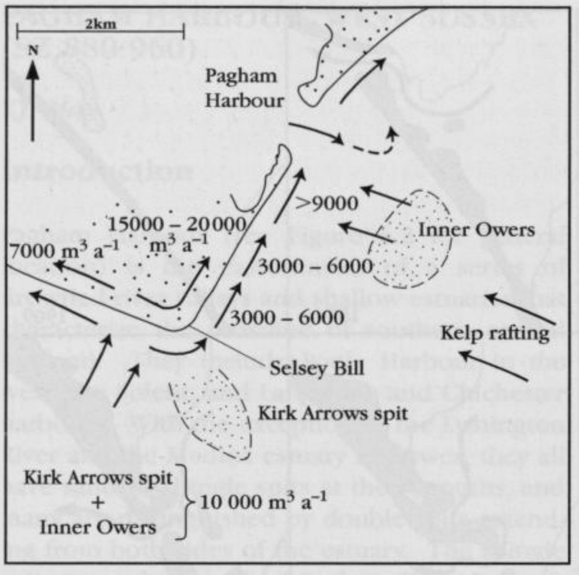

Figure 6.24 Sediment pathways at Pagham Harbour. Arrows show sediment pathways with estimated annual volumes. (Based on Lewis and Duvivier, 1976; Hooke *et al.*, 1996; and Harlow, 1979.)

sion of apparently anomalous inlet locations on the British coasts for instance (Robinson, 1955; Kidson, 1963) may possibly have been clouded by lack of consideration of the swash deflection process'. Bascom (1954) argued that, on beaches where drift is minimal, the position of inlets is determined by berm height. This in turn is determined by the distribution of wave energy along the beach. Where wave refraction is greatest, the berm is lowest and therefore a breach is most likely. At Pagham, there is considerable refraction of long waves crossing the delta with the result that at high tide they approach the beach from a more south-easterly direction than at low tide. Nevertheless, drift is considerable. The mechanism proposed by Bascom does not appear to apply here because drift is substantial, and berm height has tended to be lowest towards the distal end of the spit.

Until the construction of coast protection works at Selsey Bill in the late 1950s, erosion of the emerged ('raised') beach deposits had contributed about 4000–5000 m^3 sand and shingle annually to the westwards drift (Harlow, 1979). Larger amounts probably travelled towards Pagham. With annual cliff retreat in excess of 6 m a^{-1} between 1932 and 1951, the eastern cliffs at Selsey Bill were probably supplying about 9000 m^3 a^{-1} to the beach leading to Pagham. Jolliffe and Wallace (1973) described a process in which kelp-rafted shingle was trapped by the seabed off Selsey Bill. Two small denuded anticlines are the focus of shingle accumulation. Clasts travel up the gentle dip slopes, and are prevented from escaping by the ratchet-like scarp slopes of the individual beds. Shingle is then moved by wave action along the strike of the beds, to arrive ultimately on the beach. Harlow (1979) estimated that about 1000 m^3 a^{-1} is added to the westward drift from Selsey Bill by this mechanism. The division of longshore transport at Selsey Bill probably means that shingle also travels towards Pagham from this source. Hooke *et al.* (1996) show that between 3000 m^3 a^{-1} is added by wave-driven onshore transfers to a longshore component of a similar magnitude. There is some kelp-rafting as well as some transport seawards from within the estuary. The rate of longshore transport east of Pagham was not quantified by them. It is likely to have diminished considerably as a result of the construction of groynes between Selsey Bill and East Beach (SZ 874 948). Storm waves, for example during early January 1998, have overtopped the shingle ridge and moved the main crest landwards.

The historical evidence for the Pagham site demonstrates that a double spit form can result from unidirectional sediment movement. Breaching of the spit, nevertheless, produced a feature upon which the smaller-scale structures, for example small recurves, are a result of local longshore movement contrary to the general regional pattern. Thus the larger feature is the result of one set of processes, but its detailed form is a result of the modification of the smaller-scale processes. The importance of different time and spatial scales is well exemplified by the site. Pagham spit is the best-documented member of a considerable number of small paired spits in southern England, which together enhance our understanding of estuary-mouth sediment dynamics. It is the only such site where the sequence from unidirectional shingle spit to breaching and the resultant formation of a double spit have been documented definitively. In other cases, one spit has been described but the other ignored (e.g. at the mouth of Chichester Harbour), or there has been no investigation at all (e.g. the spits at the mouth of Newtown Harbour on the Isle of Wight), or coast protection works have radically altered one or both of the spits (e.g. Christchurch Harbour). As a bay-bar, it was a comparable form to the Loe Bar (see GCR site report), but unlike the latter was dominated by strong longshore sediment transport. In contrast to other double spits in England and Wales where sand is the main sediment, Pagham spit is formed predominantly of shingle. The development of shingle ridges has allowed the extension, breaching and repositioning of the detached ridges to be traced with greater certainty than is possible with sandy structures. Pagham Harbour thus adds considerably to the understanding of spit development.

Conclusions

Well known for the double shingle spit, Pagham Harbour is an excellent example of spit growth and breaching associated with both longshore and offshore sources of sediment. Today the natural sediment supply has largely ceased as a result of anthropogenic influence, but the ridge patterns preserve the earlier history well.

THE AYRES OF SWINISTER, SHETLAND (HU448 723)

J.D. Hansom

Introduction

The gravel beaches of the Ayres of Swinister ('ayre' is a local Shetland name for a spit or barrier) together form an exceptional example of a barrier complex, connecting the north-east mainland of Shetland to the small offshore island of Fora Ness (Figure 6.25; see Figure 6.2 for general location). Of the three gravel barriers, only the South Ayre forms a complete connecting tombolo. The other two extend out from the mainland but do not reach Fora Ness. However, they are classic examples of bay-head and mid-bay barriers (Shepard, 1952). A tidal basin called 'The Houb', lies below mean low-water springs between the South Ayre and North Ayre and contains submerged peat deposits, the dates of which have provided important information concerning the Holocene sea-level history of the area (Birnie, 1981). This site is a classic example of a coastline undergoing submergence.

Tombolos, bay-head and mid-bay barriers are relatively common features of the inner coastline of the Shetland archipelago (Flinn, 1964). However the complexity of the gravel landform assemblage at the Ayres of Swinister, with three substantial features occurring within a small area containing submerged peat deposits and saltmarsh remnants is unique both nationally and internationally. In spite of these credentials and numerous descriptive accounts, the site has failed to attract detailed geomorphological research. Nevertheless, Birnie (1981) provides stratigraphical details of the submerged peat at the site as part of a wider study to determine past environmental changes in the Shetlands, and Bentley (1996a) speculated on the evolution of the ayres.

Description

In contrast to the west coast of Shetland, the east coast is less exposed to storm-wave activity from the west yet remains open to waves from the east and north-east. Since the most frequent strong winds and storm wave activity come mainly from the south-west (BGS and Scott Wilson Consultants, 1997a), the eastern voes of Shetland and the inner reaches of Dales Voe, where Fora Ness is located, are relatively sheltered and subject mainly to locally generated waves. On account of the shelter provided by Fora Ness, sea conditions adjacent to the North Ayre are benign enough for the safe mooring of

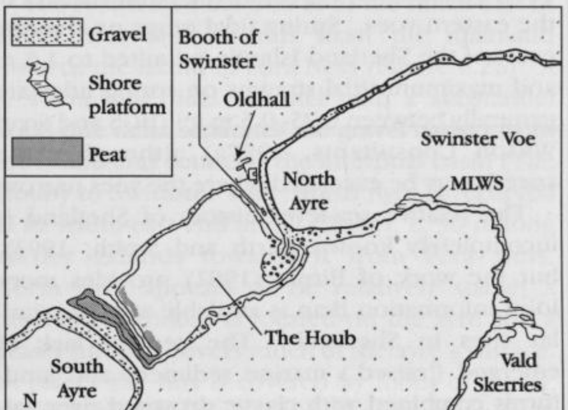

Figure 6.25 The Ayres of Swinister: a triple gravel barrier. Only the southern barrier is a true tombolo, the others are spits that enclose The Houb, a tidal basin. For general location, see Figure 6.2.

Figure 6.26 South Ayre and North Ayre at high tide looking north-east towards Swinister Voe, showing the very sheltered nature of the site. Fish farms can be seen at the North Ayre (upper left of the photograph) (Photo: J.D. Hansom.)

floating pontoons associated with a fish farm (Figure 6.26). In spite of this, severe storm events from the north and east can affect the east coast of Shetland, especially the outer reaches of the eastern voes. Spring tidal range on the east coast of the Shetland Islands is limited to 1.5 m and maximum tidal streams on spring tides are generally between 0.25–0.5 m s^{-1} (BGS and Scott Wilson Consultants, 1997a) although these speeds may be exceeded where the voes narrow.

The relative sea-level history of Shetland is incompletely known (Firth and Smith, 1993), but the work of Birnie (1981) provides more local information than is available at other similar sites in Shetland. The general lack of emerged ('raised') marine sediment and landforms combined with classic drowned river valleys (the voes) and a local tradition of marine submergence in historical times, have long been accepted as evidence for continuously rising sea-level since the decay of the late Devensian ice-sheet (Mykura, 1976). This view is supported by observations of now submerged freshwater peats in many of the sheltered voes (Hoppe, 1965; Birnie, 1981).

The Ayres of Swinister consists of a tombolo (South Ayre) which connects the mainland to Fora Ness, a bay-head barrier (unnamed) and a mid-bay barrier to the north-east (North Ayre); all of which are mainly composed of angular and subangular gravel (Figures 6.26 and 6.27). The bay-head barrier and mid-bay barrier are both hinged on the mainland side of the voe and enclose the shallow lagoonal area (The Houb), which lies below MLWS and does not dry out. A smaller lagoon is enclosed between South Ayre and the bay-head barrier, although this dries out at low tide.

The gravel tombolo of South Ayre, which is 300 m long and *c.* 50 m wide connects the island of Fora Ness to the north-east mainland of Shetland (Figure 6.25). The outer (south-west) side of the tombolo is exposed to waves from the south-west and Dales Voe, and the gravel beach maintains a smoothly curving swash-aligned arc, whereas the inner (north-east) side is characterized by a grassy turf, which is affected by spray and waves from the south-west

Figure 6.27 Washover lobes of gravel on the tombolo of South Ayre (to the right), with Fora Ness in the distance. Intertidal peats, which extend subtidally, are exposed at low tide in the lagoon between South Ayre and the unnamed barrier to the north-east (on the left). (Photo: Lorne Gill/SNH.)

(Bentley, 1996a). The inner side is characterized by a series of well-defined washover lobes and, in places, the colonizing vegetation has been subsequently partly or wholly buried by gravel. It is not known whether the peat exposed in The Houb to the north-east of the South Ayre extends to any extent under the feature itself, since it is absent from the exposed intertidal foreshore on the south-west.

A narrow, *c.* 20 m-wide lagoon separates South Ayre from the unnamed bay-head barrier immediately to the north-east (Figure 6.27). Indeed, as a result of the proximity of these two gravel features, South Ayre has often been described as a double ayre (Smith, 1993). However, the bay-head barrier is breached by a narrow tidal channel at its recurved southern tip connecting the lagoon to The Houb. The north-east side of the barrier describes a series of small arcs in response to the low-energy waves and shallow waters within The Houb. This narrow lagoon has been described as an intertidal peat flat (Smith, 1993) on which drying cracks readily develop at low tide. The peat substrate, exposed beneath the intertidal muds of the lagoon, extends between the South Ayre and the bay-head barrier.

The North Ayre lies 500 m to the north-east of South Ayre and extends from the mainland towards the island of Fora Ness (Figure 6.26). A *c.* 40 m-wide tidal channel with a substantial flood-tide delta separates the gravel barrier from the island and connects the intertidal basin (The Houb) to Swinister Voe. North Ayre is recurved at its south-east end and a smaller, *c.* 30 m-long barrier extends towards it from Fora Ness. North Ayre appears to be relatively stable; a derelict crofthouse is located on the ayre and a grassy turf mat covers much of the ayre's surface. In places the turf is covered by lobes of gravel, which have been deposited by waves from both north-east and south-west directions.

On the eastern shore of The Houb, hill peat overlying till and weathered bedrock extends below the high-water mark and floors the intertidal basin. The submerged peat contains in-situ tree stumps complete with stems and roots (Smith, 1993). The dissected peat apron (parts of which have been subject to peat cutting operations in the past) has in places been transformed into pseudo-saltmarsh. Coring of one of the uncut peat areas in the lagoon showed approximately 3 m of organic material overlying a grey, gritty clay, and yielded a radiocarbon date on *Betula* (birch) fragments 1.6 m below the surface of 4586 ± 40 years BP (Birnie, 1981). The intertidal zone on the western shore of The Houb is characterized by a low-gradient slope of sands and gravels that extend south-eastwards from the flood-tidal delta at the North Ayre.

Interpretation

Individual tombolos, bay-head and mid-bay barriers are characteristic features of the submerging inner coastline of the Shetland archipelago (Flinn, 1964), however the Ayres of Swinister provide a unique assemblage of all three of these landforms in close association. It seems likely that given its angularity and match with local glacial tills, the sediment that comprises the North and South Ayres has been eroded from the flanks of Swinister Voe by waves refracting from the north-east and south-west. Waves approaching the North Ayre comprise both swell and storm waves generated in the larger Voe in the east and thus energy levels on the northern shore are likely to be relatively higher than on the south shore. As a result the Ayre is wider and higher on this side, and it tapers to the south where the tidal prism of The Houb has given rise to flows strong enough to keep open the tidal narrows. On the other hand, waves approaching the South Ayre are unidirectional having been generated wholly within Dales Voe and travelling north-east. As a result, energy levels are evenly distributed across beach and a uniformly wide barrier has been constructed, which joins the mainland and Fora Ness to form a tombolo.

Over a longer period of time the development of The Houb must have been linked to changes in sea level. The Holocene period in Shetland is generally considered to have been characterized by a rising relative sea level (Firth and Smith, 1993). Modelling of sea-level change seems to support this view of an early and rapid rise in sea level at Shetland sites, slowing towards about 6500 years BP, but at no time undergoing relative sea-level fall (Figure 6.28; Lambeck, 1993) . During the early phase of rapid Holocene relative sea-level rise, shoreline migration in

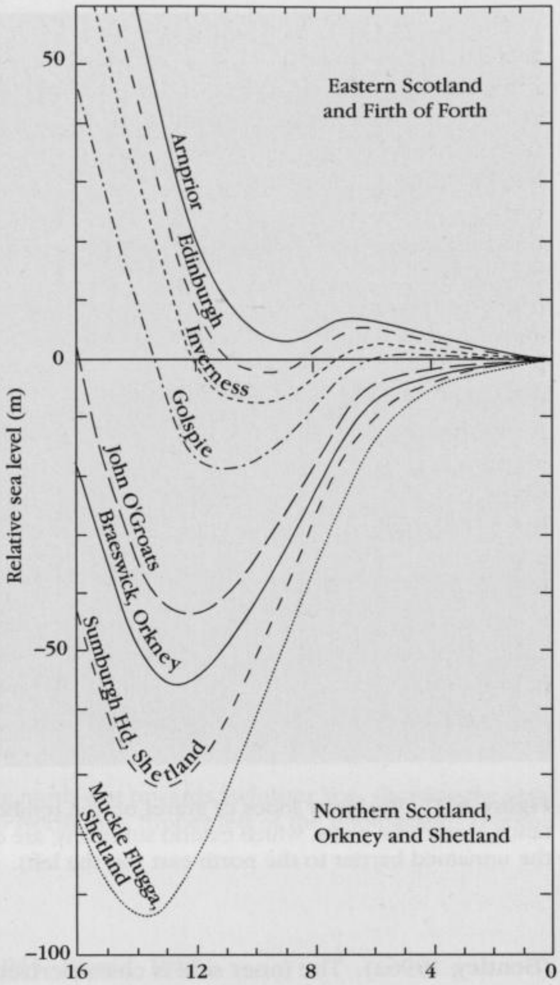

Figure 6.28 Graphs of modelled relative sea level against time over the last 16 000 years, along a south–north transect from Shetland to the Firth of Forth. (After Lambeck, 1993; Hansom, 2001.)

Scotland was probably too rapid for significant shoreface modification (Hansom, 1999, 2001) and so it seems reasonable to suggest that these constructional features were not emplaced until after the rate of sea-level rise began to slow down at some time around 6500 years BP. A radiocarbon date of 4586 ± 40 years BP (Birnie, 1981) from the underlying peat in The Houb suggests that although subsequent sea-level rise flooded The Houb and arrested further peat development, it is possible that earlier rising sea levels drove pre-existing gravel barriers onshore over any pre-peat backshore wetland. Barrier growth before about 4600 years BP may have resulted in the water table alterations that led to peat growth.

Unfortunately, the outcrops of intertidal peat that might be expected to be eroding out of the foreshores of both the North and South Ayres, are nowhere to be seen. However, the existence of peat between the South Ayre and the bay-head barrier (Smith, 1993) strongly suggests that it is continuous below the latter and that the bay-head barrier was constructed after the peat developed and thus it may post-date the development of both the South Ayre and North Ayre, although its relationship with the latter is less clear. This would suggest that the bay-head barrier may be the product of sediment eroded from the underlying glacial till and reworked alongshore. Further, the low-energy undulating planform of the feature suggests that it has never been subject to strong swash-aligned wave action from the north-east, although, if the Houb were open, this would be the direction of approach of storm waves. It may, however, be subject to waves of sufficient power to move sediment alongshore within The Houb to construct the bay-head barrier.

An alternative explanation is provided by Bentley (1996a) who has proposed the construction of the South Ayre and bay-head barrier at a high relative sea level, and the North Ayre at some time later as a result of sea-level fall. Sea-level rise then consolidated the North Ayre, flooded The Houb at some time after 4600 years BP and reworked all three features. A slowly rising sea level is suggested to have resulted initially in the development of a tombolo (South Ayre) either by normal tombolo constuction or by 'roll-over' as the gravel barrier mounted a peat-covered gap connecting Fora Ness to the mainland. A bay-head barrier then formed on the north-east side of the tombolo as a result of gravel moved from the north-east into the newly formed bay. Prior to *c.* 4600 years BP a suggested fall in sea level resulted in shoreline migration to the north-east and a new bay-head barrier (North Ayre) formed allowing peat to develop in the dry area (The Houb) between the two barriers. A later rise in sea level resulted in the breaching of the North Ayre, the flooding of The Houb, the submerging of the peat and gravel washover of the structures.

The above two evolutionary models of the Ayres of Swinister are speculative and raise several unanswered questions. For example, are the levels of longshore wave power and sediment supply of Model 1 sufficient to result in construction of the inner barrier? Model 2 requires a sea-level fall before 4580 years BP followed by a rise, for which there appears to be no other evidence in Shetland. Much scope remains for further research at this site and the triple assemblage of gravel landforms together with the associated peat beds rich in plant remains provide an excellent site for detailed research in evolutionary chronology and in stages of submergence of the Shetland coastline. Establishing these details would enhance the scientific interest of the site. Nevertheless, this triple assemblage of a gravel tombolo, bay-head barrier and mid-bay barrier, unique in Britain, forms a classic example of gravel construction on a submerging coastline.

Conclusions

Although individual barriers and tombolos occur around much of the Shetland Isles, the Ayres of Swinister is a rare example in Britain of a triple gravel tombolo-barrier complex where a tombolo, bay-head barrier and mid-bay barrier exist in close association. In addition, the site includes important peat deposits that record the sea-level history of the area and have already provided an important constraining date on sea-level rise in this part of Shetland. For these reasons, the site is justly regarded as a classic example of a submerging coastline and is of international geomorphological importance. The site is also important for its biological interest, including lagoonal flora and fauna (Thorpe, 1998).

WHITENESS HEAD, MORAY (NH 802 587–NH 842 568)

J.D. Hansom

Introduction

Whiteness Head on the south shore of the Moray Firth, north-east Scotland (for general location see Figure 6.2), is a classic example of an actively prograding spit of *c.* 3.5 km long. It is composed entirely of gravel ridges whose ends recurve southwards into an area of saltmarsh and intertidal sandflat (Figure 6.29). The gravel ridges are capped by small amounts of wind-blown sand. Westward sediment transport along this shoreline has resulted in the progressive westerly growth of Whiteness Head (Stapleton and Pethick, 1996; Ramsay and Brampton, 2000c), the historical evolution of which has been well documented (Ogilvie, 1923; Steers, 1973; Smith, 1974; Halliwell, 1975; Stapleton and Pethick, 1996; Bentley, 1995). From a short gravel bar in the 1880s, the end of the spit at Whiteness Head has migrated in a westerly direction at rates of between 10 and 30 m a^{-1} over the last 150 years. Such downdrift growth of the distal end of the spit is currently fuelled by updrift erosion of its proximal end (Ritchie, 1983; Bentley, 1995). The Whiteness Head SSSI comprises one area on the eastern side of the Ardersier Platform Construction Yard (an oil rig construction facility), and another area on the west side of the yard. The GCR site lies entirely within the eastern part of the SSSI.

Description

The gravel spit at Whiteness Head protrudes out from the southern shore of the inner Moray Firth, in a north-westerly direction. At its broadest the 3.5 km-long gravel spit is *c.* 200 m wide, although it narrows to less than 15 m close to the updrift end, and encloses *c.* 1 km^2 of saltmarsh and intertidal sandflat on its southern side (Stapleton and Pethick, 1996). To the west, extensive mudflats extend to the glacial till headland of Ardersier, whereas to landward, the spit and saltmarsh are backed by a complex of emerged Holocene sand and gravel deposits backed by a prominent, 10–12 m-high, emerged cliff (Ritchie *et al.*, 1978).

Whiteness Head spit consists of at least seven recurved gravel-spit complexes, each one partially truncated at its proximal end by the succeeding one (Ritchie *et al.*, 1978) (Figure 6.30). They lie at altitudes of between 3 m and 5 m OD and are separated by saltmarsh and dry gravel slacks. In general, the highest ridges form the active beach and lie on the seaward (northern) margin of the spit. The upper beach face consists of a gravel storm-ridge with a steep seaward slope of over 10°, increasing to 18° in the centre of the spit (Bentley, 1995). The lower beach face is characterized by a low-gradient sand slope with intertidal bars. The beach face is also punctuated by well-developed, but ephemeral, suites of rhythmic cuspate forms. The offshore hydrography indicates a gentle offshore gradient passing into deep water beyond a narrow sloping shelf with no evidence of a submarine bar system in the nearshore zone.

Extending along the entire length of the spit

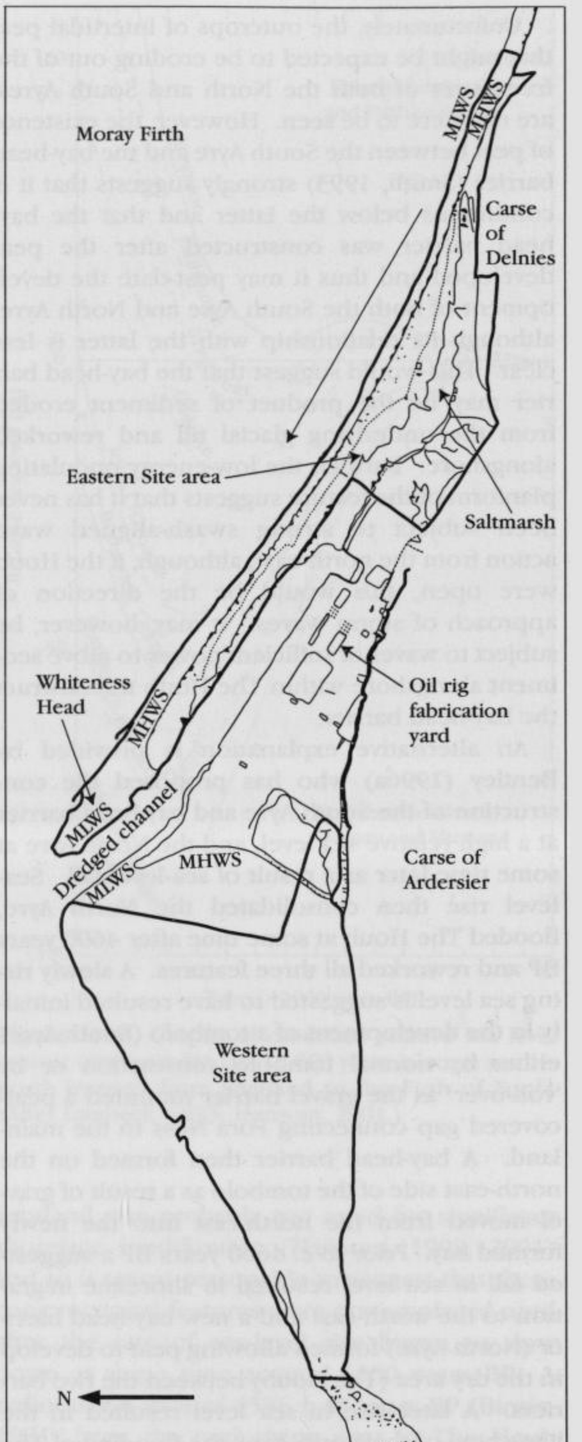

Figure 6.29 Sketch map of the Whiteness Head area. The GCR site lies entirely within the eastern site area; the arrow indicates direction of net longshore sediment movement.

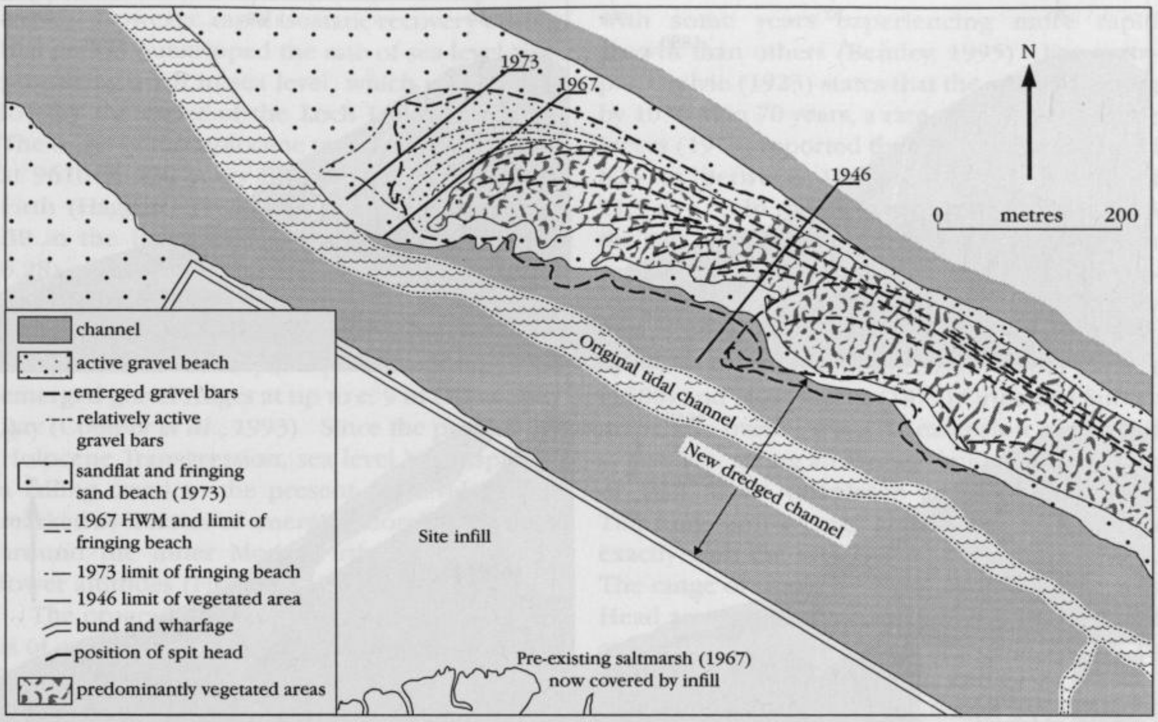

Figure 6.30 Map of historical changes in the Whiteness Head Spit between 1946 and 1973. (After Smith, 1974.)

parallel to the outer coast, a prominent gravel ridge marks the limit of storm-wave action. This laterally mobile storm ridge is almost 2 m higher than the older recurved ridges and is overtopped by waves during storm-wave activity (Bentley, 1995). It declines in height eastwards towards the proximal end of the spit where a low plateau of gravel extends landwards into the saltmarsh as a series of lobate washover fans. This has imparted a crenulate inner edge to the spit at its narrowest point. Known locally as 'the Gut', the gravel storm ridge forms the only barrier preventing waves from entering the saltmarsh behind, although wave over-topping and erosion is common during storm activity (Bentley, 1995). This eastern part of the spit is also characterized by erosion of the older gravel ridges, which trend at an angle to the present-day coast. This results in truncation of the seaward ends of ridges and slacks to form an undulating gravel cliff whose height is usually less than 0.7 m (Ritchie *et al.*, 1978; Bentley, 1995). Accretion and growth occurs at the distal (western) end of the spit where the storm ridges of the outer beach form a series of landward-trending gravel recurves, which are rapidly colonized by pioneer plants. Ongoing westward extension is now curtailed by maintenance of the tidal channel at the distal end by dredging.

Although the spit is composed almost entirely of gravel, the distal (west) and inland portions are locally overlain by a thin cover of blown sand, seldom exceeding 2 m in thickness (Bentley, 1995). The dune forms in this location are subdued as a result of episodic deflation and wave overtopping and the underlying gravel base is exposed in several places. In the place of a true foredune system there is a rather irregular sand mantle of amorphous mounds capped by marram grass *Ammophila arenaria*, although small areas of narrow linear dunes capped by blue lyme-grass *Leymus arenaria* have formed atop the gravel ridges, particularly at their distal ends. In the east, a dense cover of *Ulex-Calluna* (gorse–heather) communities occurs but this changes westwards through *Ulex–Ammophila* communities to *Ammophila* and eventually to a dominance of *Leymus* on the most recent embryonic dunes at the extremity of the spit. For a fuller description of the vegetation of Whiteness Head see Currie (1974).

The inner (southern) beach of the spit is characterized along its length by small bays that have formed between the recurved ends of the

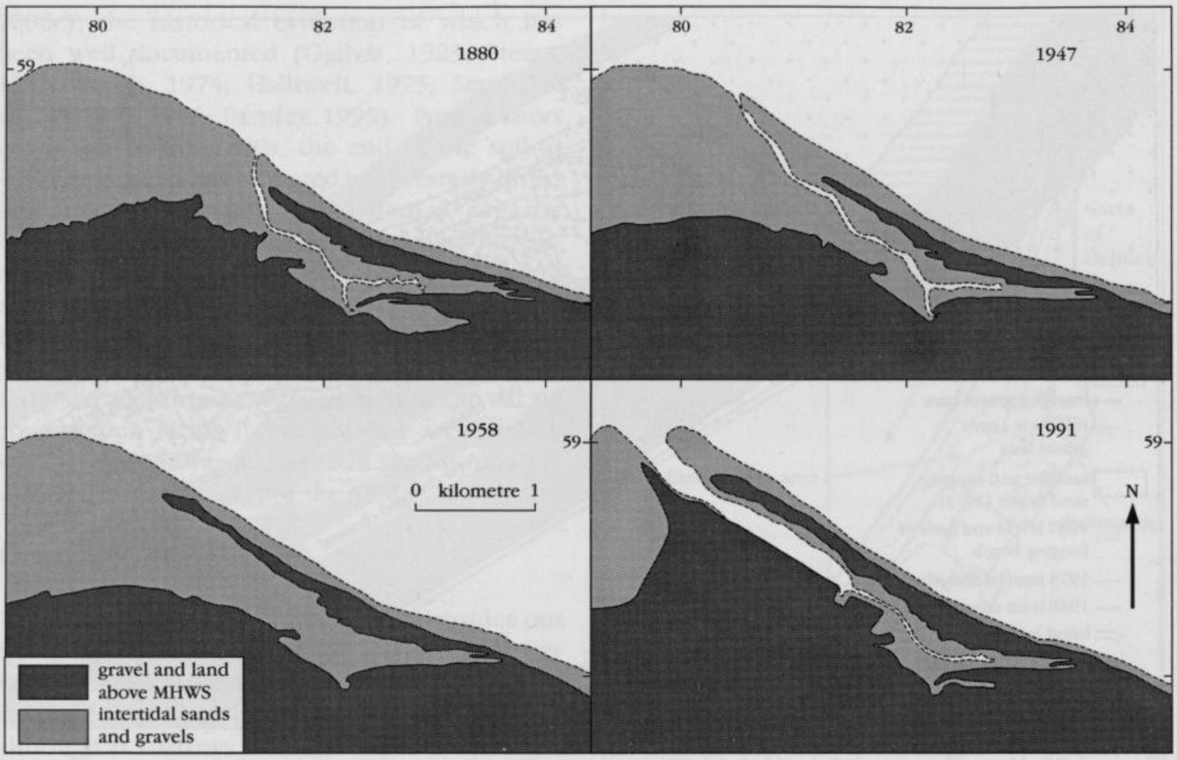

Figure 6.31 Historical evolution of the Whiteness Head Spit between 1880 and 1991. In 111 years the spit has lengthened considerably and the creek morphology has changed. Note the pronounced change between 1958 and 1991 when the McDermott construction yard was built and a prominent channel was dredged on the south side of the spit. (After Stapleton and Pethick, 1996.)

relict gravel ridges marking former positions of the distal end of the spit (Figure 6.30). Along this inner edge a wide intertidal sand-flat is exposed at low tide although this merges landwards into gravels that underlie saltmarsh. The saltmarsh is drained by a system of creeks and there is a marked absence of saltpans. Dominant species on this marsh include samphire *Salicornia*, common saltmarsh-grass *Puccinellia maritima*, sea aster *Aster tripolium*, sea plantain *Plantago maritima*, greater sea-spurrey *Spergularia media*, red fescue *Festuca rubra*, sea-milkwort *Glaux maritima*, thrift *Armeria maritima* and cord-grass *Spartina anglica* (Stapleton and Pethick, 1996).

Interpretation

The Moray Firth is landlocked except in the sector 000°–090° and so north-easterly swell and wind waves dominate. Although smaller wind waves generated in other sectors within the firth can be significant, the dominant direction of wave energy is from the north to north-east sector and incident waves meet the shoreline obliquely. This has strongly influenced the alignment of coastal features, the net westward drift of sediment being manifest in the orientation of spits along the middle and inner Firth including at Speymouth, Lossiemouth, Findhorn Bay, Buckie Loch, The Bar at Culbin and Whiteness Head. Erosion of the coast is thus at its most severe under north and north-easterly storms. Tides in the Moray Firth lie in the range 3.5 m at springs, but they generate relatively weak currents because, rather than being directed into the firth, the tidal wave crosses its entrance (BGS, 1996). However, the co-incidence of a north-easterly gale and high spring tides can elevate water levels considerably along the coast, producing locally significant erosion along the southern Moray Firth.

The sea-level history of the Moray Firth is relatively well known and newly deglaciated areas were being flooded by the sea at *c.* 13 000 years BP (Firth, 1989; see Figure 6.28, Inverness curve). However, rapid isostatic recovery during this period outstripped the rate of sea-level rise, producing a fall in sea level, which was already low by the onset of the Loch Lomond Stadial. The onset of the Holocene transgression is dated at 9610 ± 130 years BP from the inner Moray Firth (Haggart, 1986, 1987), and at 8800 years BP in the Beauly Firth near Inverness (Figure 6.28). The culmination of this rise at *c.* 6400 years BP was marked in the Beauly and inner Moray Firths by the formation of the Main Postglacial Shoreline (MPS) and a series of emerged gravel ridges at up to *c.* 9 m OD at Spey Bay (Comber *et al.*, 1993). Since the peak of the Holocene Transgression, sea level has displayed a falling trend to the present-day level, and is marked by a series of emerged shoreline features around the inner Moray Firth at successively lower altitudes (Firth and Haggart, 1989).

The prograding gravel spit of Whiteness Head is of outstanding geomorphological significance, and Whiteness Head, together with Spey Bay (see GCR site report below) and Culbin (see GCR site report, Chapter 11), form an important suite of GCR sites on the southern shore of the Moray Firth. Wave-induced westerly longshore drift predominates along this coastline (Ramsay and Brampton, 2000c) and is reflected in the coastal landforms of all three GCR sites with westerly deflected spits at Speymouth, Findhorn Bay, Buckie Loch, The Bar at Culbin and Whiteness Head. Although extensive coastal protection works at Nairn, 6 km east of Whiteness Head, has undoubtedly reduced the natural supply of longshore sediment to Whiteness Head, it has been suggested that the spit represents the early stages in the development of a 'flying barrier' similar to The Bar at Culbin (Ritchie *et al.*, 1978; Ritchie, 1983; Bentley, 1995).

The historical evolution of Whiteness Head spit is relatively well documented (Figure 6.31). The spit has been developing since before 1853 (Ogilvie, 1923), although no information is available concerning the time of its initiation. There are reports that in 1823 no gravel was present at Whiteness Head, and although Steers (1973) considered this statement to be highly unlikely, his reasons are unclear. There is great variation in the published figures of the rate of westerly growth at Whiteness Head, which probably results from either differences in the methods used to calculate changes, or in the actual growth of the spit, which may have been pulsed, with some years experiencing more rapid growth than others (Bentley, 1995). For example, Ogilvie (1923) states that the spit had grown by 1070 m in 70 years, a rate of 15.2 m a^{-1}, while Steers (1973) reported the spit to have extended 823 m between 1868 and 1937, a rate of 11.9 m a^{-1}. In the 10 years between 1955–1965 the spit is reported to have extended 304 m, an annual rate of 30.4 m a^{-1}, whereas in the following 10 years (1965–1975) this rate of westerly growth had slowed to 18.2 m a^{-1} (Halliwell, 1975). Based on a comparison of the earliest (1880) and most recent (1977) Ordnance Survey maps, the overall westerly extension of the spit is 1.48 km, giving an average annual growth rate of 15.2 m a^{-1} (Stapleton and Pethick, 1996). This long-term average growth rate corresponds exactly with the rate quoted by Ogilvie (1923). The range of westerly growth rates of Whiteness Head are remarkably similar to those measured at The Bar, Culbin, which has been shown to extend westwards at a mean rate of 14.6 m a^{-1} between 1976 and 1989 (Comber, 1993). Both spit features can be seen to be predominantly moving to the west rather than landwards, although proximal erosion is a feature of both.

As occurs elsewhere in the Moray Firth, westerly longshore drift of sediment has been the driving force behind the creation and continued migration of the gravel spit at Whiteness Head, its westward growth resulting from updrift erosion fuelling gravel accretion at its distal (western) end where newly deposited recurved gravel barriers develop. Most of the sediments responsible for the construction of the features of this coast were originally derived from rivers and offshore sources accumulated in the Moray Firth basin during Lateglacial and Holocene times, and plentiful before about 6500 years BP (Hansom, 1999). However, in common with coasts elsewhere (Carter, 1988, 1992), the amount of sediment has since diminished and on the Moray Firth coastline there has been a progressive reduction in both gravel and sand supplies to the coast. In a situation of sediment starvation coastal development proceeds via internal re-organization of existing sediments and so current spit extension is probably entirely fuelled by the truncation of older ridges at the proximal end and recycling of sediment downdrift (Hansom, 2001). Continued growth of the spit is thus at the expense of updrift erosion.

The impact of this process on coastal alignment can be identified in the erosional truncation of the older gravel ridges, which trend at an angle to the present-day ridge orientation. There has also been a north-eastwards movement of the whole system as the spit builds out seawards and an increase in intertidal width at the distal end of the spit (Stapleton and Pethick, 1996). As a result of updrift erosion and downdrift accretion, the spit has 'rotated' clockwise over time from an east–west alignment to a south-east–north-west alignment (Bentley, 1995). Where sediment supply is restricted, this rotation is consistent with the concept of a wave-driven progression in the evolution of beach alignments from an original state of drift-alignment towards an equilibrium state of swash-alignment (Davies, 1972). As there is no reason to assume that the gravel shortage is a temporary phenomenon, it has been suggested that a continuation of the present mode of spit development (i.e. proximal erosion fuelling distal accretion) will eventually result in the severance of the spit near its hinge-point and the creation of a detached barrier similar to that at Culbin (Smith, 1974; Ritchie *et al.*, 1978; Ritchie, 1983; Bentley, 1995).

A significant proportion of the previous extent of the saltmarsh was reclaimed during the building of the Ardersier Platform Construction Yard in 1973. The yard lies to the landwards of the GCR site and occupies a 120 ha site of former saltmarsh and emerged mudflat (carse) (Smith, 1974). Land-claim was achieved by suction-dredging of sand from the intertidal zone landwards of the spit and deposition into settling ponds behind sand bunds, which were infilled to a level of *c.* 4 m OD (Smith, 1974). A dredged channel is now maintained to a depth of up to 12 m and the yard edge adjacent to the channel is protected with a steel wall (Smith, 1974) (Figure 6.29). Erosion is currently a problem at the eastern end of the wall and parts of the claimed land are suffering erosion (Bentley, 1995). Although the geomorphological interest of the intertidal flats and saltmarsh behind Whiteness Head has been diminished by the dredging and land-claim operations, the outer beach remains a classic example of a rapidly prograding gravel spit complex. It is an important site for the study of the processes of active spit development and migration, in a context of sea-level fall, relative sediment starvation and active longshore drifting. As a result, the site is of importance to the understanding of the long-term evolution of dynamic gravel spits.

Conclusions

Whiteness Head on the south shore of the Moray Firth, north-east Scotland, is a classic example of an active, rapidly prograding gravel spit. The *c.* 3.5 km-long dynamic spit complex, which has been migrating westwards at rates of 10–30 m a^{-1} over the last 150 years is of outstanding national geomorphological importance. The relatively well-documented historical evolution of Whiteness Head adds to the scientific interest and provides a key to understanding the long-term evolution of gravel features in areas of active longshore drift. Today, in a period of relative sediment starvation and sea-level fall, updrift erosion of the spit is currently fuelling downdrift accretion and driving the continued westerly migration of the spit complex. It is possible that a continuation of the present-day processes will result in the severance of the spit from the mainland creating a detached barrier similar to that at Culbin (see GCR site report in Chapter 11).

SPEY BAY, MORAY (NJ 264 688–NJ 388 642)

J.D. Hansom

Introduction

Spey Bay (see Figure 6.2 for general location) is one of the most important gravel coastal geomorphology sites in Great Britain for several reasons. The extensive and well-developed gravel ridge complex is recognized as the finest in Scotland, providing examples of dynamic coastal processes and active fluvial supply of gravels that are unparalleled in the British context (Comber *et al.*, 1994; Gemmell, 2000; Gemmell *et al.*, 2001a,b). In addition, the active coastal margin is backed by a magnificent emerged strandplain of Holocene gravel ridges that record the progressive history of coastal development and sea-level fall in this part of Scotland over the last 10 000 years (Ogilvie, 1923; Steers, 1973). The Speymouth delta and related forms have a complex and well-documented history of dramatic change (Grove, 1955) and provide an excellent example of the fluvial-coastal interaction and sediment interchange of an actively braided gravel-bed river entering a high-energy coastal environment (Gemmell *et al.*, 2001a,b). Indeed, the lower River Spey has also been selected as a GCR site in its own right on account of its unique fluvial geomorphology (see GCR site report in the GCR volume *Fluvial Geomorphology of Great Britain* (Gregory, 1997)). Nowhere else in the UK is there such a dynamic example of an actively braiding gravel-bed river delivering sand and gravel to a wide coastal gravel beach and backed by a suite of emerged gravel shorelines (Figure 6.32). The wealth and scale of juxtaposed features provide a unique insight into the Holocene development of this part of the Scottish coastline.

The GCR boundary follows the Spey Bay SSSI boundary along mean low-water springs (Figure 6.33) for 13 km from Porttannachy (Portgordon) in the east to 2.5 km east of Lossiemouth in the west, and is paralleled by a landward boundary along tracks and fencelines up to 250 m inland from MLWS. The site is selected for its gravel features and although the western limit is located some 2.5 km east of Lossiemouth, the ongoing migration of gravels beyond this boundary is set to continue. As a result, the obvious geomorphological boundaries of this stretch of coast lie at the rocky headlands at Porttannachy (Portgordon) and Lossiemouth.

Description

As already noted in the description of the Whiteness Head GCR site above, the Moray Firth is landlocked except between north and east and so north-easterly waves dominate and their oblique incidence produces a westerly movement of sediments. Erosion of the coast is at its most severe under northerly and north-easterly storms. For further details on waves, tides and

Figure 6.32 The extensive gravel ridges and emerged coastal and fluvial terraces of the Spey mouth in 1963. At this time, the river was diverted west by over 1 km, threatening the village of Kingston in the right centre of the view. (Photo from Gemmell *et al.*, 2001b)

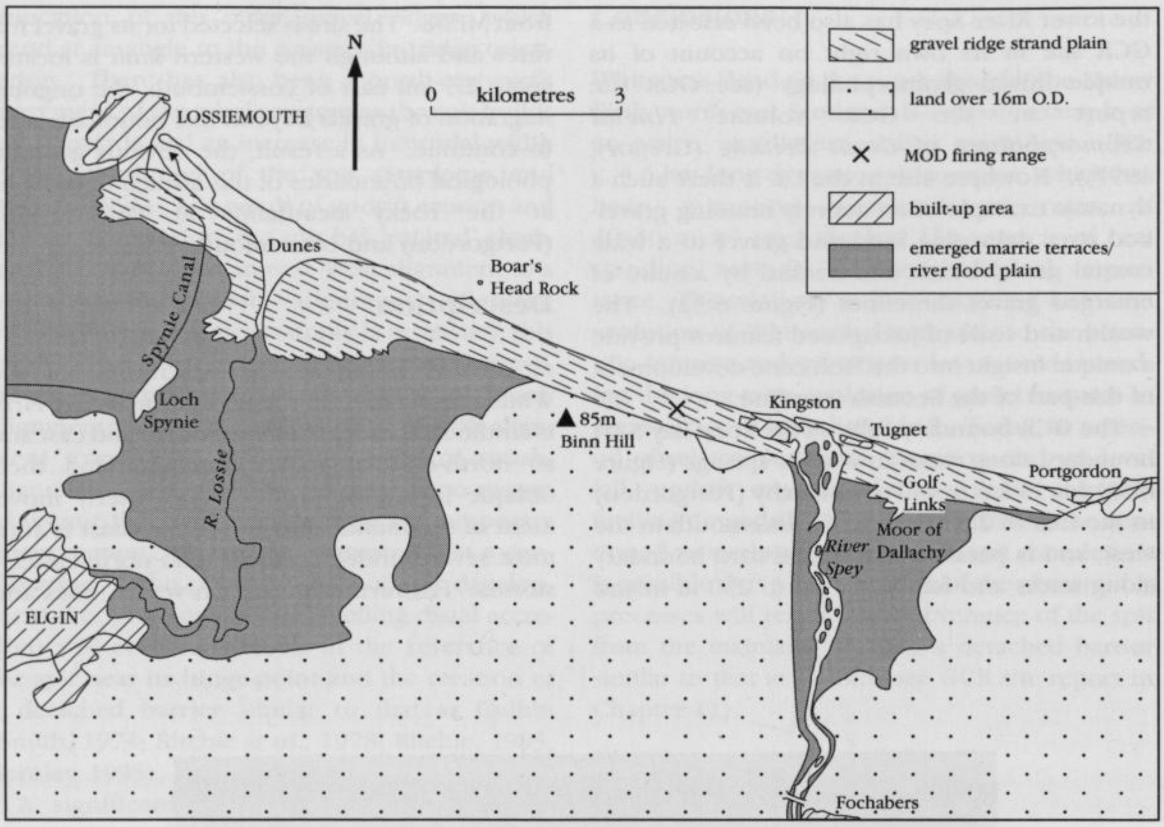

Figure 6.33 Spey Bay showing coastal gravel strandplain backed by emerged marine (and fluvial) terraces. Land over 16 m is mainly glaciofluvial sands and gravels. MoD is a Ministry of Defence weapons testing range. (After Ritchie, 1983.)

general sea-level changes, see site GCR report for Whiteness Head (present chapter).

Spey Bay lies on the southern shore of the Moray Firth and extends for a distance of *c.* 16 km between rock headlands at Lossiemouth in the west and Porttannachy/Portgordon in the east (Figure 6.33). The present-day coastline trends WNW–ESE and is juxtaposed against a magnificent emerged strandplain of gravel ridges (Figure 6.34) which have isolated a series of glaciogenic deposits, glacio-fluvial and recent river terraces and residual pockets of low marshy ground (Ritchie, 1983). The contemporary gravel beach extends for *c.* 13 km from Porttannachy/ Portgordon in the east to about 2.5 km east of Lossiemouth, where an abrupt transition occurs to a low-angled sand beach backed by sand dunes. The gravel beach is punctuated between Tugnet and Kingston by a complex of gravel ridges that mark the delta of the River Spey (Figure 6.32). Along much of Spey Bay, the average altitude of the main gravel ridge crest is *c.* 6 m OD, although this varies considerably. Crestal overtopping of the gravel storm ridge is common along much of the coastline, particularly west of Porttannachy/ Portgordon and west of Kingston, and coastal recession is a constant concern. The width of the gravel beach varies considerably, from *c.* 10–15 m immediately west of Kingston, to up to 70 m wide, west of Boar's Head Rock.

West of the Speymouth delta, close to the Ministry of Defence firing range, several low altitude vegetated ridges are truncated by the present gravel beach, and curve gently landwards. These low altitude recurves occur continuously from this point to the westward extent of the gravel beach beyond Boar's Head Rock and become higher and more prominent westwards. West of Boar's Head Rock up to five well-defined gravel ridges curve gently landwards from the rear of the present-day active ridge. In the west, the ridges reach *c.* 6 m and stand up to 2 m above the adjacent intervening troughs. To landward, the ridges are sparsely vegetated by mosses and grasses and are eventually buried by

Table 6.3 Westerly extension of the active gravel beach (West Spey Bay). (From Gemmell *et al.*, 2001b.)

Time period	Westerly growth (m)	Growth per annum (m a^{-1})
1870–1903	1360	41
1903–1967	2090	33
1967–1994	720	27
July 1994–December 1995	30	20
1870–1995	4200	34

high sand dunes and forest. The full landward extent of ridges is known to be substantial, continuing into the Spynie area to the south of Lossiemouth and as far west as Burghead Bay (Gemmell *et al.*, 2001a,b). Within Spey Bay, the junction between these younger, more recently deposited ridges and the ridges of the emerged gravel strandplain, is marked by a 1–2 m rise in altitude and a distinct break in slope, best seen 2–3 km east of Kingston.

On the seaward face of the gravel beaches, cusp forms of different wavelengths are well developed, the size and spacing of these ephemeral features altering in response to short-term processes that vary with wave and tidal conditions. Beach-face slope angles, the degree of sediment sorting and crest elevations also alter in response to wave and tidal conditions. Sediment sorting is well developed down the beach face, with finer-grained, well-sorted gravel lying in the intertidal zone whereas larger calibre, but more poorly sorted, gravel, occurs at or above high-water mark or in the horns of the cusps. However, there is no obvious alongshore trend in beach sediment size, until the abrupt transition from gravel to sand close to Lossiemouth. The median grain size of the gravel varies from 30 mm to 50 mm along the beach, whereas the sand has a median grain size of 0.22 mm (Gemmell, 2000; Gemmell *et al.*, 2001b).

Along much of its length, the gravel ridge is subject to washover during storms and at several places washover throats occur in the main gravel ridge that allow coarse gravel lobes to accumulate landwards of the main ridge. This roll-over effect is widespread along the coast. Gravel is also being moved westwards under the influence of westerly waves. According to Grove (1955) 'the most recent gravel bank on the west side [of Spey Bay] appears to have grown steadily along the beach towards Lossiemouth over a distance of one and half miles (2.4 km) since 1870' at an average rate of westerly extension comparable with that of the gravel spits that grow across Spey mouth (Grove, 1955). Using map and field evidence the total westerly extension of the gravel beach was 4.2 km between 1870 and 1995, an average annual extension rate of 33.6 m a^{-1} (Gemmell *et al.*, 2001a,b; Table 6.3). Where there was no gravel present in 1903, today there is a 60 m-wide gravel beach, consisting of an active beach ridge of *c.* 4 m OD behind which lie several landward-recurving ridges at about the same altitude.

Changes in the position of mean high-water springs (MHWS) and MLWS at Spey Bay between the first (1870) and the latest current Ordnance Survey (1970) reveal that the eastern side of Spey Bay has been eroded over the intervening 100-year period. This erosional trend declines to the west beyond the Spey delta until it gives way to accretion *c.* 4 km west of the delta (Gemmell *et al.*, 2001a,b). Recession rates since 1975 of 1–1.5 m a^{-1} have been recorded both east and west of the river exit (Riddell and Fuller, 1995). The replacement of sand by gravel, discussed above, is reflected by accretion over the 1870–1970 period in the west of Spey Bay along a 4 km stretch in the vicinity of Boar's Head Rock. Farther west, the sandy beach and dunes at Lossiemouth are wholly erosional over the map period, a trend which continues today (Gemmell *et al.*, 2001a).

At the mouth of the Spey, complex fluvial and coastal processes interact to create a dynamic and highly active system (Figure 6.35). Historical records (Grove, 1955) suggest that: 'the mouth of the Spey alters more rapidly from year to year than almost any other section of the coastline of Britain and ... the position of the mouth of the Spey has fluctuated violently throughout the last two centuries'.

Changes in the position of the river mouth

Figure 6.34 The emerged gravel ridges of Spey Bay descend in a 'staircase' from 9–10 m OD to the present-day beach. The greatest extent of the unvegetated gravel occurs to the west of Kingston (see Figure 6.35), where this picture was taken. (Photo: J.D. Hansom.)

between 1726 and 1995, documented in Gemmell *et al.* (2001a,b) show a natural tendency for the river mouth to shift westwards towards Kingston, driven by the wave-driven westward migration of the spit across the mouth. If this natural process is uninterrupted, the mouth migrates by up to 1.2 km west of its 'central' (*c.* 200 m west of Tugnet) position. According to local tradition the Speymouth spit was *c.* 5 km long in 1798 (Hamilton, 1965). The river has always returned to a central location through natural breaches of the gravel spit, as recorded in 1829 and 1981 (Riddell and Fuller, 1995), but recent breaches have been artificially engineered to a 'central' location in order to reduce the threat of flooding and erosion at Kingston. In spite of this, there are also several documented examples (e.g. 1870, 1989 and 1995) of temporary easterly drift at Speymouth, which, although generally short-lived, demonstrate the sensitivity of the longshore drift system to local variations in the wind and wave climate. A complex suite of gravel ridges is present both to the east and west of the Spey outlet, enclosing tidal lagoons. The orientations and recurves of these ridges relate to the interaction of coastal and fluvial processes during the varying positions of the river over time, allowing former positions of the Spey mouth to be identified.

As a result of the erosional nature of much of Spey Bay protection measures have been implemented, mainly in the east. A vertical sea-wall fronts the village of Porttannachy/Portgordon in the east, a 400 m-long section of rip-rap backs the beach immediately to the east of the river mouth at Tugnet and a programme of beach replenishment using gravel excavated from the ridges at the delta mouth has recently been implemented along a 2 km stretch at Kingston (Gemmell *et al.*, 2001a,b).

Interpretation

The contemporary coastal development of Spey Bay is juxtaposed against a fine suite of older emerged Holocene landforms and provides a unique site for the study of former sea levels and how these relate to both past and contemporary sediment budgets. In the UK context, Spey Bay

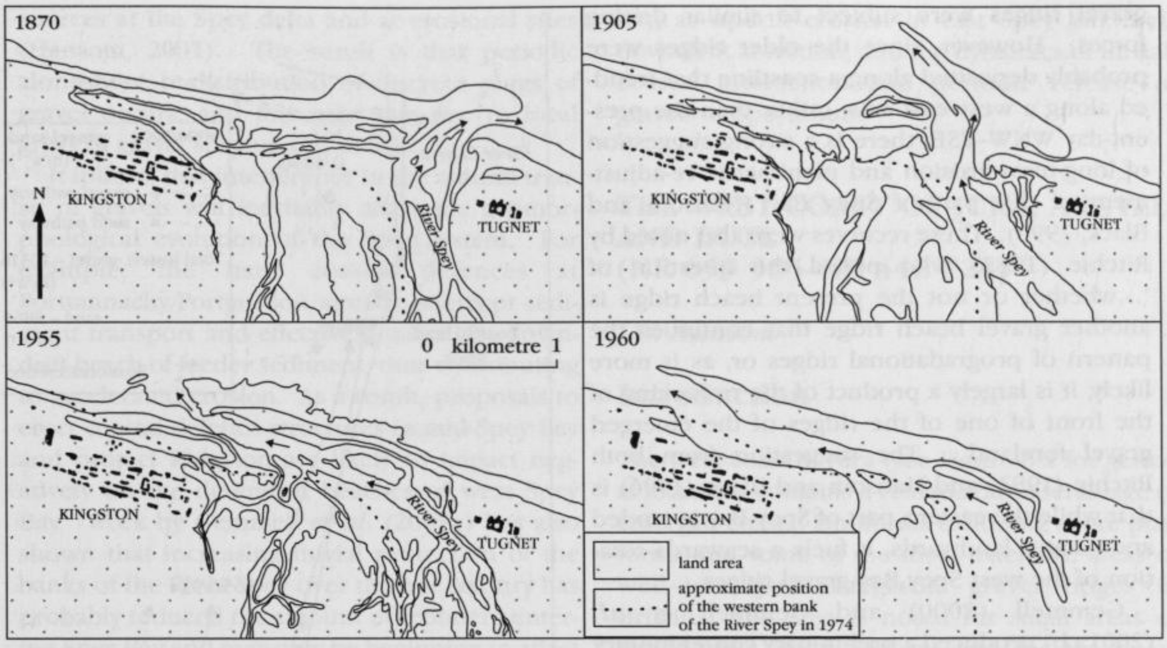

Figure 6.35 Movement of the River Spey mouth between 1870 and 1960. (After Grove, 1955 and Gemmell *et al.*, 2001a.)

also provides a unique insight into the under-researched area of deltaic processes that occur as a large gravel-bed river exits into a highly dynamic open coast situation.

Ogilvie (1923) provided the first interpretation of the gravel strandplain of Spey Bay in its wider regional and Holocene context, although the most recent and detailed work is that of Riddell and Fuller (1995), Gemmell (2000), and Gemmell *et al.* (2001a,b). Initial emplacement of the Spey Bay ridges against the foot of the Holocene cliff probably began before the peak of the Holocene Transgression (*c.* 6500 years BP), when rising sea levels delivered large quantities of offshore material to the coast. Together with River Spey gravels, this material infilled the low and flooded areas south of the present-day coast in the area of the Moor of Dallachy, almost as far as Fochabers upriver and to the south of Kingston along the foot of Binn Hill (Gemmell *et al.*, 2001a,b; Figure 6.33). Along its entire length, the Holocene coastline was marked by a very prominent cliff, last occupied *c.* 6500 years BP, and now fronted by a seawards-falling staircase of younger emerged beaches (Figure 6.34). Progressively, the bay of the lower Spey was closed off by gravel storm ridges across the mouth between Porttannachy/Portgordon and Kingston and infilled by fluvial accretion behind. Extending westwards from Kingston, gravel ridge accretion extended beyond the lower Lossie–Spynie area, which was an inlet of the sea, as far as Burghead Bay and the Findhorn area. However gravel accretion and sea-level fall progressively cut off the inlet of the Spynie area and River Lossie from the open sea. Eventually an 800 m-wide swathe of ridges developed to separate Binn Hill from the sea. Altitudes of the ridges suggest that the majority of the sequence was deposited after the peak of the Holocene transgression, with levelled transects displaying a stable and then rapidly falling trend in altitude (Comber, 1993; Gemmell *et al.*, 2001a,b).

There remains great potential for further research at this site and elsewhere on the Moray Firth to produce a quantified Holocene sediment budget for the Firth. For example, research at Culbin *c.* 40 km west of the Spey (see GCR site report for Culbin, Chapter 10) recognized the Holocene contribution of sediment from the River Spey and Findhorn to the development of the large gravel ridge strandplain at Culbin (Comber, 1993).

Contemporary coastal processes and landforms are dominated by a strong westerly movement of gravels and truncation of the gravel recurves in west Spey Bay by the present-day beach suggests that the preceding generations of gravel ridges were subject to similar driving forces. However, since the older ridges were probably deposited along a coastline that trended along a west–east axis, rather than the present-day WNW–ESE, there is a strong suggestion of long-term erosion and planimetric re-adjustment of this part of Spey Bay (Hansom and Black, 1996). These recurves were also noted by Ritchie (1983) who posed the question of '...whether or not the present beach ridge is another gravel beach ridge that continues the pattern of progradational ridges or, as is more likely, it is largely a product of the reworking of the front of one of the ridges of the emerged gravel foreland'. The suggestion from both Ritchie (1983) and Hansom and Black (1996) is that while the eastern part of Spey Bay is eroded and rotates landwards, it fuels a seawards rotation of the west Spey Bay gravel ridges.

Gemmell (2000) and Gemmell *et al.* (2001a,b) produced a preliminary contemporary sediment budget for Spey Bay (Figure 6.36), using a combination of map and field evidence and results from published computer modelling. Sediment input into the contemporary system derives from updrift erosion of the present-day beach and from the River Spey itself, which from modelling studies contributes an estimated 8000 $m^3\ a^{-1}$ to the coast (Riddell and Fuller, 1995). These input sediments contain approximately 80% gravel and 20% sand, and so the annual input of *gravel* into the system from the river is *c.* 6400 m^3 (Figure 6.36). When added to the amount contributed by recession of the adjacent shore (9496 m^3), some 15 900 m^3 of gravel is input to the Spey system annually. However, the annual accretion of gravel in the ridges at Boar's Head Rock and beyond only accounts for approximately 5600 m^3 and so there is an apparent annual net loss of 10 316 m^3 of gravel from the Spey system. This apparent loss may be a function of error in the amount and periodicity of material delivered or in the estimate of gravels held at Boar's Head Rock. If losses are occurring then the final destination of the gravels is unknown. There is also an apparent annual net loss of *sand* from the Spey Bay system that amounts to 8864 m^3. The final destination of the sand is probably to sand dune accretion behind the beach and infill of the Lossie saltings, however an unknown amount is almost certainly lost to the offshore zone and may bypass the local sediment cell boundary (Ramsay and Brampton, 2000c) at Lossiemouth.

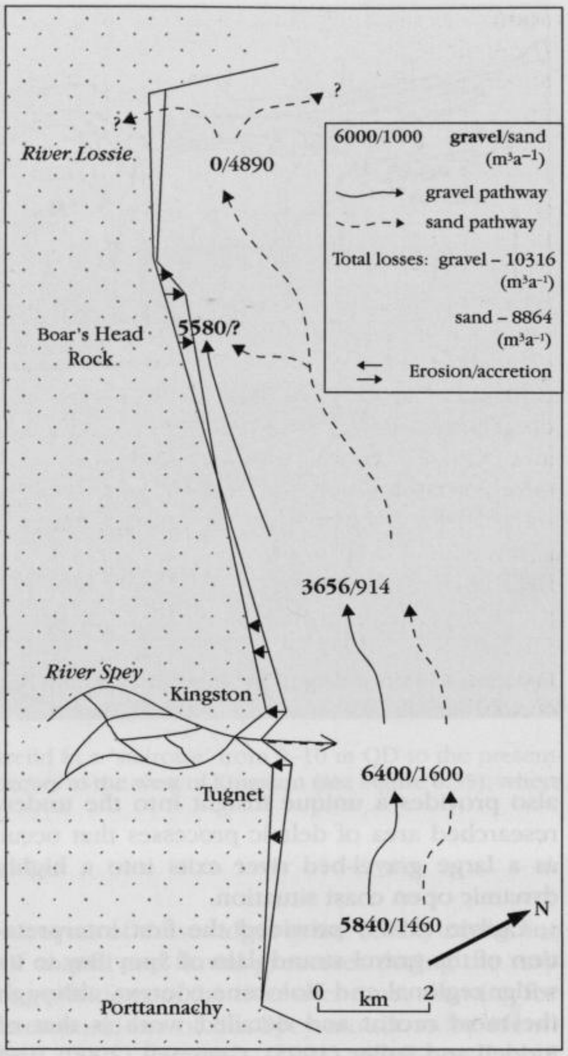

Figure 6.36 Diagrammatic representation of the Spey Bay sediment budget. Scale approximate. (After Gemmell *et al.*, 2001a.)

The above sediment budget, although preliminary and under revision, is instructive because it supports the geomorphological evidence and indicates that the entire length of Spey Bay functions as a discrete sedimentary unit, with erosion of one section influencing accretion at another. However, the supply of gravel alongshore is not constant and is subject to pulsing depending upon fluvial supply and storm events (Gemmell *et al.*, 2001a). Further, since sediment supply from the offshore to Scottish coasts is now much diminished, the supply to the Spey system is no longer added areally to the shoreface as before. Instead, reduced volumes of sediment are now added mainly as point sources at the Spey delta and at erosional sites (Hansom, 2001). The result is that periodic alongshore re-distribution of discrete plugs of gravel occurs, and this may give rise to local areas of gravel surplus and deficit.

It follows that interference in the natural transit of gravels will inevitably affect the geomorphological evolution of the Spey system. For example, the hard coastal defences at Porttannachy/Portgordon already interrupt sediment transport and effectively starve the downdrift beach of feeder sediment, thus contributing to accelerated erosion. As a result, proposals to erect coastal defence structures in mid-Spey Bay and protect Kingston are likely to impact negatively on the downdrift beaches of west Spey Bay. Work by Gemmell *et al.* (2001a) has also shown that increasing fluvial protection of the banks of the River Spey over the last century has probably reduced the amount of sediment entering Spey Bay and may now be beginning to affect the natural geomorphological evolution of the coastal system.

Conclusions

Spey Bay is an important site for coastal geomorphology and of particular interest because the large input of fluvial gravels is unusual in a UK context. The active gravel storm ridges of the present-day coast are some of the finest in Scotland and their constant adjustment to waves in Spey Bay demonstrates both short- and medium-term dynamic coastal processes. The active coastal margin is juxtaposed against a magnificent emerged strandplain, with a suite of gravel shorelines relating to the progressive history of coastal development within the Moray Firth as adjustments took place in Holocene sea levels. Additionally, the site is important on account of the unique fluvial–coastal interaction displayed at Speymouth, where the coarse sediments of the dynamic and actively braided River Spey enter a high-energy, open-coast gravel beach.

At Spey Bay, there is great scope to provide more accurate contemporary and Holocene sediment budgets for the river and coast. In addition, the contemporary development of Spey Bay has three unique features, all of which have great potential for future study: the loss of the sand beach at Lossiemouth and the progressive replacement by the westerly accretion of gravel; the gradual change in coastal orientation of Spey Bay, as updrift erosion in east Spey Bay fuels downdrift accretion; and the dynamics of fluvial-coastal interaction and periodic release of gravels at Speymouth.

THE WEST COAST OF JURA, ARGYLL AND BUTE (NR 659 985–NR 442 724)

J.D. Hansom

Introduction

The west coast of Jura (see Figure 6.2 for general location) contains a remarkable assemblage of emerged coastal landforms including shore platforms and some of the most extensive areas of well-developed Lateglacial gravel ridges in Britain. The area is noted for small areas of machair-like dune surfaces and for the finest example of a medial moraine in Great Britain at Sgriob na Caillich (NR 475 765). The emerged coastal landforms provide valuable information for understanding changes in Late Devensian and Holocene relative sea level.

Description

The 37 km-long stretch of the west coast of Jura, between Glengarrisdale Bay (NR 659 985) in the north, and Rubha Aoineadh an Reithe (NR 448 751) in the south, together with a small area (0.2 km^2) at Inver (NR 442 724), is one of the classic localities in Great Britain for emerged coastal landforms. Spectacular unvegetated spreads of Late Devensian and Holocene emerged beach gravel are juxtaposed with excellent examples of three emerged shore platforms, the High Rock Platform, Main Rock Platform and Low Rock Platform. The emerged gravel beaches were described by Ting (1936, 1937), however, the first major study of the emerged beaches was by McCann (1961, 1964), who sought to describe and explain the origin of the western Jura gravels by relating them to Lateglacial relative sea-level change. More recently, the emerged shorelines of western Jura have been investigated in detail by Dawson in several papers, most of which are reviewed by Dawson (1993) in the *Quaternary of Scotland* GCR volume.

The High Rock Platform and associated cliff that extend continuously between Shian Bay

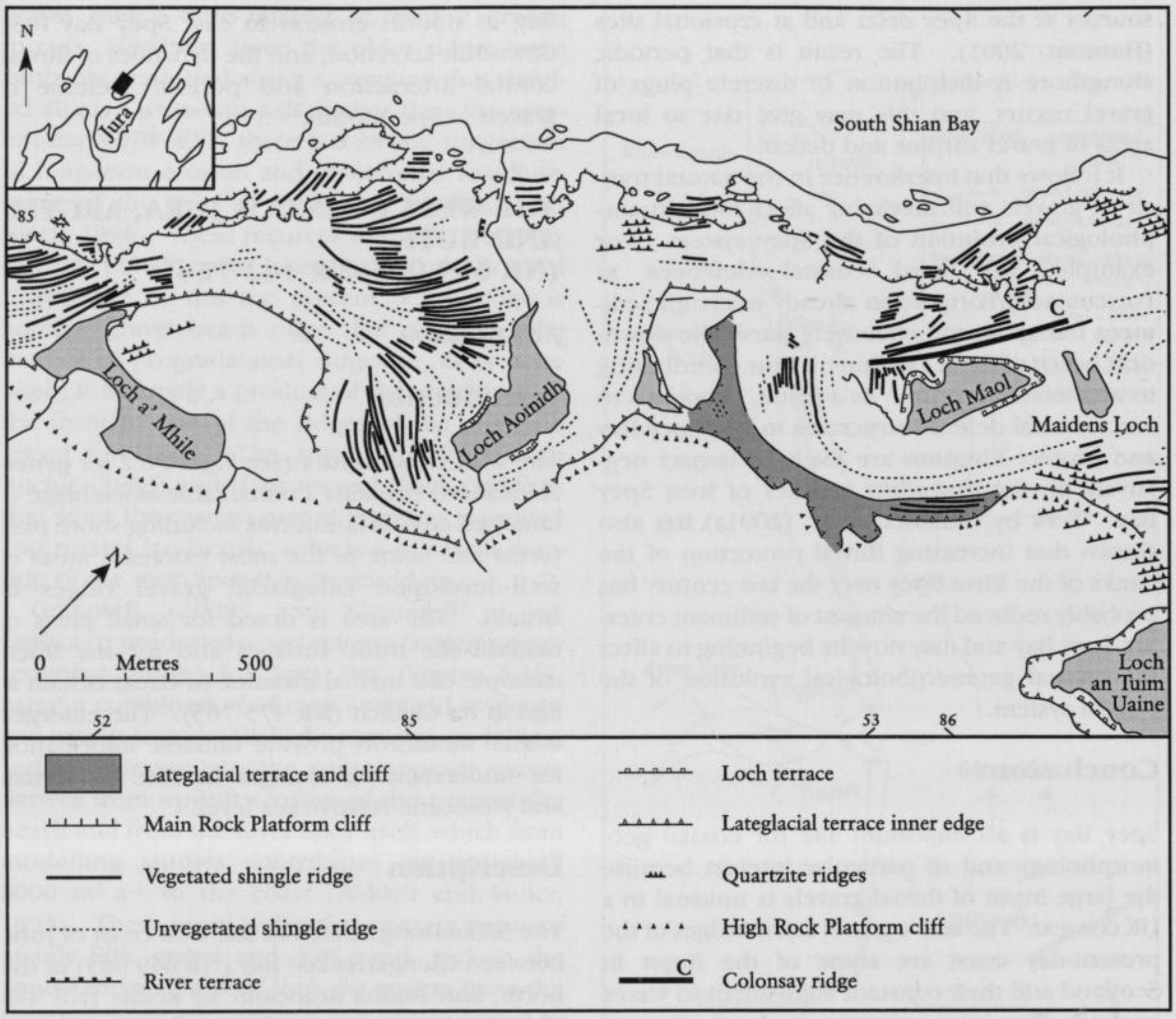

Figure 6.37 Geomorphology of western Jura in the area of South Shian Bay, showing the 'staircase' of emerged gravel ridges. (After Dawson, 1993.)

(NR 530 875) and Ruantallain (NR 505 833) are are discussed extensively by Dawson (1993) and so only a brief account is included here. The platform has an average width of 350 m but reaches 600 m and the backing cliffs are usually 5–15 m high. The inner edge of the platform lies at an altitude of 34.1–32.1 OD and its seaward slope is about 4°. The surface is covered in places by gravel spreads, which overlie not only rock platform but also patches of lodgement till. Elsewhere the surface is ice-moulded and a striated bedrock sea stack occurs immediately north of Loch a' Mhile (NR 514 850).

Between Shian Bay and Ruantallain the seaward edge of the High Rock Platform forms the cliff of a lower platform, the Main Rock Platform. The lower platform is 50–150 m wide and the inner edge occurs at 3–5 m OD. It is locally overlain by Holocene emerged beach sediments, and the crenulate cliffs of 10–15 m high are indented by numerous emerged sea caves. The platform has a serrated and uneven surface with no signs of glacial moulding. The platform is also continuous between Shian Bay and Glendebadel Bay.

The intertidal rock platform fragments of the Low Rock Platform are conspicuous along long stretches of the Jura coast. Typically 100 m wide, they are best developed on the foreshore between Rubh' Aird na Sgitheich and Allt Bun an Eas. The platform surfaces are locally ice-moulded and in most places pass inland beneath till. Between Rubh'Aird na Sgitheich and Glenbatrick the platform is overlain by up to 15 m of Late

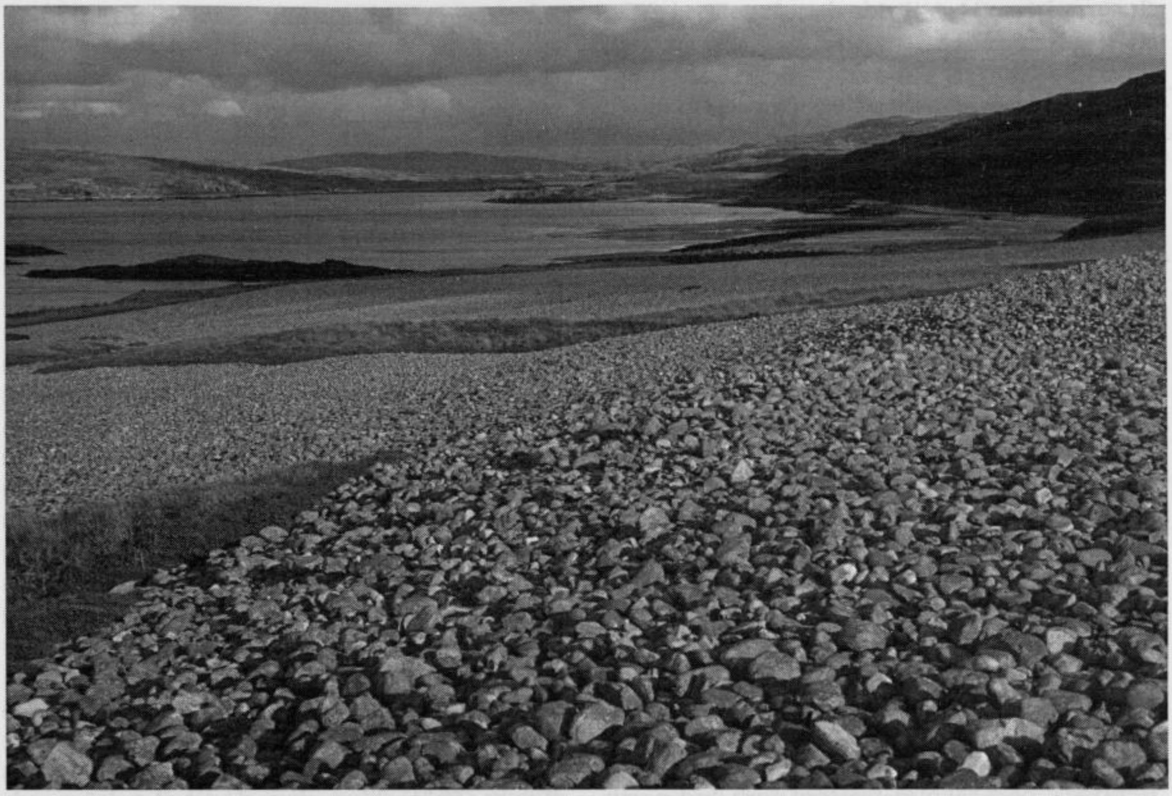

Figure 6.38 An unbroken 'staircase' of unvegetated emerged gravel beaches falls from *c.* 30 m OD to sea level on the West Coast of Jura. Looking eastwards towards Glenbatrick. (Photo: J.D. Hansom.)

Devensian emerged beach gravels.

The coast of western Jura is dominated by conspicuous emerged beach terraces and 'staircases' of unvegetated beach ridges. Although discussed by Ting (1936, 1937), the most detailed studies of these emerged coastal features are by McCann (1964, 1968) and Dawson (1979, 1982). There are two main sets of gravels: a higher suite and a lower suite. The higher suite of emerged coastal terraces can be traced almost continuously southward from Shian Bay (Figure 6.37) to Inver but other areas occur at Corpach Bay, Glendebadel and in the Glenbatrick area (Figure 6.38). In most cases the emerged marine deposits are ridges of unvegetated quartzite gravels that decline in altitude from north-east to south-west, from 40 m OD at Corpach Bay to 24.5 m OD at Inver. The altitudes of the higher suite of beaches suggests the existence of two shorelines, the higher of which declines in altitude from 40 m OD at Corpach in the north to 34 m OD near Ruantallain in the south-west, a regional gradient of 0.56 m km^{-1}. A separate and lower set of ridges occurs in the south-west, from 31 m OD at Glenbatrick to 24 m OD at Inver, a regional gradient of 0.53 m km^{-1}. Along most of the west coast of Jura the higher suite of emerged ridges terminate at the cliff of the Main Rock Platform and so it is unusual to find them below 20 m OD. However, at South Shian Bay (Figure 6.37), emerged beach gravels descend to 11 m southwest of Loch Maol, probably on account of the unusually low altitude of the rock platform on which they sit (Dawson, 1993). The detailed pattern of ridge crests on both sets of gravel beaches bears a close relationship to the intricacies of the rock platforms on which they sit and of the rock headlands that separate individual beach units.

The lower suite of beach gravels is widespread throughout the west coast of Jura and falls in altitude from between 10 and 12.3 m OD to merge with the gravel ridges of the modern beach. Only a few distinct coastal terraces exist, but emerged gravel banks and beach accumulations everywhere mantle the rock surfaces of the Main and Low Rock Platforms. Spectacular staircases of emerged gravel ridges commonly occur, the best example of which is present north of

Inver where 31 individual unvegetated ridges descend from 12.3 m OD to the present-day beach.

Small areas of windblown sand occur in several of the small pocket beaches. Of these, Corpach is of most geomorphological interest in that it contains a small area of rare cliff-foot dunes that resemble a machair surface. These dunes mantle the Main Rock Platform and are locally banked in great ramps of bare sand against the emerged cliff landwards. Although showing active deflation in places, the dunes are mostly vegetated and contain buried palaeosols together with several features of archaeological interest including grave burials in the emerged gravels. The dunes at Corpach, together with dune areas at Shian, Bagh Gleann nam Muc and Glengarrisdale show some of the geomorphological attributes of machair, with a close sward of grasses and an absence of the normal dune grasses such as marram *Ammophila.*

Interpretation

The High Rock Platform of the southern Inner Hebrides, best seen in western Jura and northern Islay, was considered by Wright (1911) to be 'pre-glacial' in age on account of the emerged beach gravels that rest between its surface and a superficial cover of till. McCann (1968) suggested instead that it was 'interglacial'. Dawson (1979) accepted an interglacial origin but considered that the shoreline had been warped by neotectonic activity. Sissons (1982) proposed that the platforms that comprise the High Rock Platform were produced by cold-climate shore erosional processes and that the various platform fragments are part of a series of glacio-isostatic tilted shorelines. However, the altitudes from the West Coast of Jura (and northern Islay) do not demonstrate any platform tilt and so it is probably a single feature. At present there exists no general agreement on platform origin or age. Dawson (1983) argues that formation of the western Jura platform by cold-climate shore erosion would have taken a minimum of 8000 years and that such a lengthy period of relative sea-level stability during a single period of cold climate was unlikely. It would therefore appear that the western Jura High Rock Platform may represent the product of several periods of Pleistocene coastal erosion (Dawson, 1993).

The inner edge of the platform, which abuts the lower cliff of the High Rock Platform, occurs at 3–5 m OD and constitutes part of a glacio-isostatically tilted shoreline that declines in altitude to the south-west, from 6 m OD in northern Jura to sea level in northern Islay at a regional gradient of 0.13 m km^{-1}(Dawson, 1993). Together with its altitude, jagged nature and freshness of form, this gradient suggests that the platform conforms to the Main Rock Platform identified elsewhere in the Inner Hebrides (Gray, 1978). Dawson (1980a) reports the platform to be unglaciated although where the regionally tilted Main Rock Platform merges with and crosses the regionally horizontal intertidal Low Rock Platform the distinction becomes blurred (Dawson, 1979, 1980a; see below). The intertidal ice-moulded rock platforms that occur in north-west Jura and northern Islay are referred to by Dawson (1980a) as the 'Low Rock Platform', its regional horizontality explained as having been produced by marine processes during interglacial periods.

Since the lower set of platform fragments has been ice-moulded, they were produced prior to the last glaciation. This platform, first noted by Wright (1911) as a '... preglacial plain of marine denudation...', termed the 'Low Rock Platform' by Dawson (1979). Dawson noted that its presence as an ice-moulded intertidal feature, unaffected by glacio-isostatic tilting along many parts of the Scottish coastline, implied interglacial origins. Sissons (1981) argued that the glaciated intertidal features represented a set of platform fragments of different ages that had been subject to glacio-isostatic deformation and then exhumed in the intertidal zone as a result of present-day marine activity. According to this hypothesis, the rock platform features were initially produced by cold-climate, shore-erosion processes. The higher platform fragments were considered part of the glacio-isostatically tilted Main Rock Platform that was regarded as having been produced during the cold climate of the Loch Lomond Stadial (Dawson, 1993). This shoreline is generally considered to pass below sea level on Islay, owing to its glacio-isostatic deformation (Dawson, 1993).

The regional variations in altitude of the highest emerged beach terraces on the west coast of Jura suggest the existence of two shorelines. The older emerged shoreline at 34–40 m OD is also thought to occur in northern Islay and has a regional gradient of 0.56 m km^{-1} (Dawson, 1993). A separate and slightly younger shoreline is present in south-west Jura declining in altitude to the south-west from 31 m to 24 m OD at a regional gradient of 0.53 m km^{-1}. Dawson (1982) inferred from these that both south-west and north-west Jura remained ice-covered while the higher shoreline formed between Corpach Bay and Shian Bay. Deglaciation of south-west Jura took place at a slightly later date and was accompanied by the formation of the lower shoreline and its gravel ridges. The regularly declining ridge-crest altitudes of the western Jura gravel 'staircases' indicates that, although stillstands may have occurred during the fall in the sea level from 35 m to 20 m OD, no major sea-level oscillations occurred. Most of the west Jura gravel spreads below this altitude terminate at the cliff of the Main Rock Platform and so patterns of sea-level change below 20 m OD cannot be established except at South Shian Bay where McCann (1964) proposed that a prominent gravel spit at 19 m OD called the 'Colonsay Ridge' (Figure 6.37) represented a pause in the overall fall in sea level.

The presence on the west coast of Jura of extensive spreads of emerged gravel is primarily due to the glacio-isostatic uplift of the higher shoreline and its altitudinal relationship with the till-covered High Rock Platform. These relationships indicate that both high and low gravels are of Late Devensian age. On deglaciation, the maximum sea level along this coast (34–40 m OD) stood several metres higher than the inner edge of the High Rock Platform. Wave erosion of the till cover resulted in extensive gravel deposition, a process enhanced by the gentle sloping nature of the underlying platform surface and its open exposure to westerly waves.

The trend of falling sea levels in the Lateglacial was reversed later in the Holocene with the highest ridges of the extensive suites of lower beach gravels probably representing the culmination of this rise. The culmination of the sea-level rise at about 6500 years BP in the west coast of Jura produced ridges that now lie at 12 m OD. Subsequently a fall of sea level to its present-day level, the result of ongoing isostatic uplift, deposited a staircase of 31 individual gravel ridges and indicates that about 12 m of uplift has occurred in the area over the last 6500 years (MacTaggart, 1998a).

The cliff-foot dunes at Corpach are of interest in that they, along with similar features in other bays along the west coast of Jura, have been interpreted as machair (Ritchie and Crofts, 1974). The dunes mantling the Main Rock Platform are mostly stable and vegetated but bare sand has been blown into highly dynamic climbing dunes up the emerged cliff behind. The dunes at Corpach, together with dune areas at Shian, Bagh Gleann nam Muc and Glengarrisdale, all show some of the geomorphological attributes of machair, with a close sward of grasses and an absence of dune grasses (e.g. marram *Ammophila*). However, their status as machair has been questioned from a botanical perspective because few of the dune areas have even moderate amounts of carbonate sand and do not support classic machair vegetation communities (Dargie, 2000; Angus, 2001). Without this supporting evidence, it seems more appropriate to regard the blown sand deposits of the west coast of Jura as dune systems rather than machair.

Conclusions

The west coast of Jura is outstanding for its assemblage of emerged coastal landforms and gravel beach deposits. Both the range of features and their extent and degree of development are exceptional and include not only fine examples of the three major rock platforms recognized in western Scotland, the High, Main and Low Rock Platforms, but also extensive spreads of unvegetated Lateglacial and Holocene gravel beach ridges unparalleled elsewhere in Britain for the length of their morphological record of sea-level changes.

BENACRE NESS, SUFFOLK (TM 532 824–TM 535 831)

V.J. May

Introduction

Benacre Ness (see Figure 6.2 for general location) is a good example of a ness formed in shingle and associated with rapid coastal retreat of part of the beach and of nearby cliffs. Williams (1956) outlined the history of the ness (which had been referred to by Ward (1922) under the name 'Covehithe Ness'), showing how it had progressively moved from south of Covehithe to its present-day position (see Figure 6.39); almost 6 km in 200 years. The site comprises three landform units: namely cliffs cut mainly in fluvioglacial sand with a fringing beach of sand and shingle, a beach ridge fronting Benacre Broad and The Denes, and Benacre Ness itself, formed of sand and shingle ridges. Although the evidence from longshore sediment transport is that material moves towards the south, the ness form itself has moved northwards. As well as being a classic landform, therefore, Benacre Ness is of considerable importance for studies of coastal form-process dynamics (Steers, 1946b; Russell, 1956; Williams, 1956, 1960; Hardy, 1966; Cambers, 1975; McCave, 1978b; Carr, 1981; Onyett and Simmons, 1983).

Description

The southern part of this site is formed by the cliffs at Covehithe, which are undergoing very rapid erosion. Cambers (1973) estimated the historical (100-year) rate of retreat at up to 4.25 m a^{-1} and it has exceeded 6 m a^{-1} since 1929. The estimated loss of beach volume during a single 24-hour period at Covehithe was 300 000 m^3 (Williams, 1956). The rapidity of cliff-top retreat is shown vividly by the truncation of the lane leading from Covehithe itself and the loss of autumn-sown crops by the next spring. The cliffs decline northwards to Benacre Broad. Here the beach is a single fringing ridge that blocks the mainly infilled Benacre Broad (between TM 532 824 and TM 535 831). This ridge has similar rates of retreat to the cliffs to the south, but the rate of change declines along The Denes until Benacre Ness where accretion at rates up to +2.46 m a^{-1} has occurred since 1880. The northern part of the ness is marked by erosion, but at significantly lower rates (up to 0.36 m a^{-1}) than the southern part. Steers (1981) has drawn attention to the very high rates of retreat that have occurred during storm surges. Between 19 March 1977 and 11 March 1978, at three separate points, 9 m, 8 m and 14 m were lost. During 1979 and 1980, the same points lost 2.7 m, 4.6 m and 8.8 m. Even larger values of 12 m and 27 m were recorded by Williams (1956) during the storm surge of early 1953. A surge in 1990 caused overnight recession of 35 m (K. Clayton, pers. comm.).

Interpretation

Much of the early description of the ness and its changes was based upon cartographic evidence (Ward, 1922; Steers, 1946b; Williams, 1956). Both Williams (1956) and Steers (1964a) found inconsistencies in the cartographic record, and Steers inclined towards the view that the ness has only existed in a form similar to present since about 1826. Steers (1946a) suggested that the feature had originated as a spit across the Kessingland River but offered no evidence for this or for the direction of the spit's growth. Williams (1956) suggested that a northward transport of sediment would occur immediately following surges. The large quantities of material eroded from the cliffs would be transported 'under the action of an abnormal north-going pull as the level of the sea falls'. Sediment would accumulate seawards of the ness and would then be gradually pushed up the beach by wave action. Steers accepted that this might be the case today, but could not see how this process could occur at earlier stages of ness formation when the ness was protecting the cliffs. The beach throughout the site is composed mainly of flint shingle but there are varying amounts of sand, except at the ness, which is formed almost entirely of shingle. Whereas the beach fronting Benacre Broad and The Denes is tending to move landwards, the ness itself has moved progressively northwards. Onyett and Simmons (1983) described Benacre as moving rapidly to the north, but were unable to provide evidence as to the net change in its volume.

In his discussion of Ward's paper, Russell (1956) suggested a simple mechanism for the apparent conflict between southward movement of sediment and northwards movement of the ness. The alignment of the northern face of the ness would mean that drift was zero, whereas the alignment of the southern face would increase longshore transport. Material would be accreted at the northern face, but the southern face would be eroded (Figure 6.39). As a result the ness would move northwards. The same mechanism was proposed by Cambers (1975) at Winterton Ness. Williams (1960) accepted Russell's suggestion, adding that the rate of movement of the ness is probably slowing as the amount of material available for transport reduces. However, Russell's concept does not explain how the ness forms in the first instance. At some point along the coast, wave direction must have been affected by refraction and the alignment of the coastline in such a way that the beach aligned itself to face the dominant waves. If, as Steers (1946a) suggested, there was a spit along this coast, the slight change in alignment

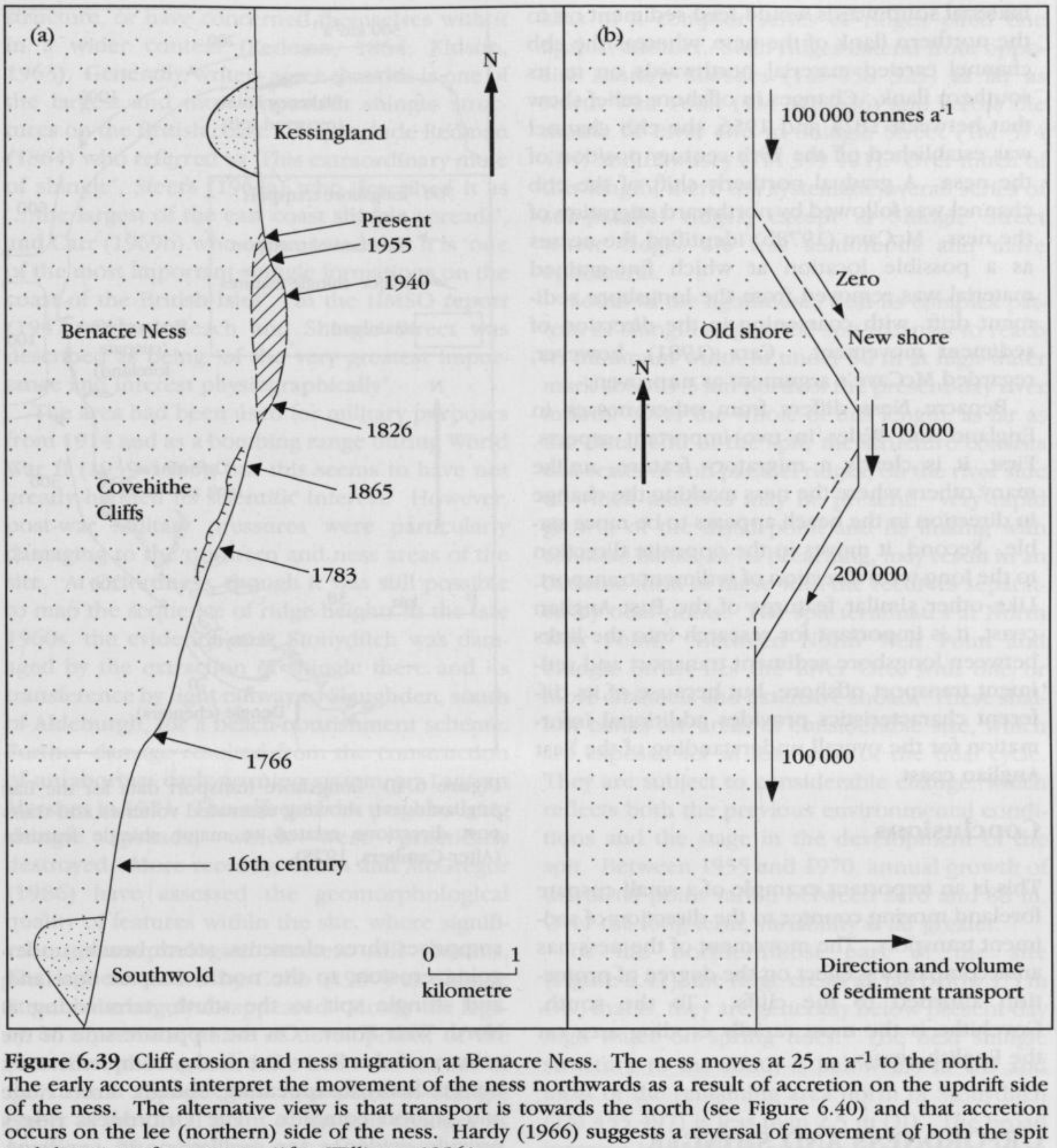

Figure 6.39 Cliff erosion and ness migration at Benacre Ness. The ness moves at 25 m a^{-1} to the north. The early accounts interpret the movement of the ness northwards as a result of accretion on the updrift side of the ness. The alternative view is that transport is towards the north (see Figure 6.40) and that accretion occurs on the lee (northern) side of the ness. Hardy (1966) suggests a reversal of movement of both the spit and direction of transport. (After Williams, 1956.)

of the coastline south of Benacre could have been sufficient to have caused the beach to change its alignment slightly to face the waves. Once this had occurred, the mechanism proposed by Russell would ensure the maintenance of the ness and its movement northwards.

However, Hardy (1966) demonstrated that the ness had migrated southwards since the beginning of the 20th century, in association with a complex pattern of erosion and deposition in neighbouring areas. A northward-trending offshore bank had developed since 1945 that was aligned away from the coast in the immediate vicinity of the ness. It was likely that a flood channel had developed between the ness and this bank that would allow sediment to come ashore in the area of the ness. An ebb channel carried most material offshore on the eastern side of the north–south-trending bank. Changes in the foreshore appeared to have occurred before the formation of the bank. Robinson (1966) argued that a flood channel carrying material southwards would feed sediment on to the northern flank of the ness, whereas the ebb channel carried material northwards on to its southern flank. Changes in offshore relief show that between 1824 and 1956, the ebb channel was established off the 19th century position of the ness. A gradual northerly shift of the ebb channel was followed by northward migration of the ness. McCave (1978b) identified the nesses as a possible location at which fine-grained material was removed from the longshore sediment drift, with coarsening in the direction of sediment movement. Carr (1981), however, regarded McCave's argument as unproven.

Benacre Ness differs from other nesses in England and Wales in two important aspects. First, it is clearly a migratory feature, unlike many others where the ness marking the change in direction in the beach appears to be more stable. Second, it moves in the opposite direction to the long-term direction of sediment transport. Like other similar features of the East Anglian coast, it is important for research into the links between longshore sediment transport and sediment transport offshore, but because of its different characteristics provides additional information for the overall understanding of the East Anglian coast.

Conclusions

This is an important example of a small cuspate foreland moving counter to the direction of sediment transport. The movement of the ness has an accompanying effect on the degree of protection afforded to the cliffs. To the south, Covehithe is the most rapidly eroding area on the English coast.

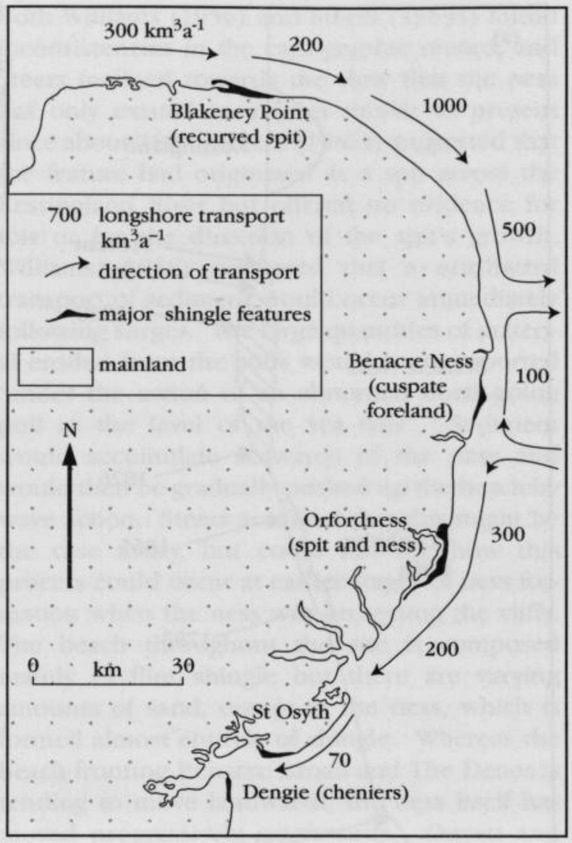

Figure 6.40 Longshore transport data for the East Anglian coast, showing estimated volumes and transport directions related to major shingle features. (After Cambers, 1975).

ORFORDNESS AND SHINGLE STREET, SUFFOLK (TM 358 400)

V.J. May

Introduction

The shingle ridges that form Orfordness (see Figure 6.2 for general location) extend about 15 km south from Aldeburgh on the Suffolk coast and divert the River Ore for a similar distance (Figure 6.40). South of the mouth of the river, the shingle ridges at Shingle Street continue southwards towards Bawdsey. Orfordness comprises three elements: storm beach, undergoing erosion, to the north; cuspate foreland; and shingle spit to the south, terminating at North Weir Point. On the opposite side of the estuary of the River Ore is the complementary, but gradually disappearing, Shingle Street ridge and lagoon complex. The Orfordness ridges provide evidence for oscillations in sea level, and research work on the spit and in the estuary has helped clarify many of the processes that are relevant in spit development worldwide.

The site has been well documented (Redman, 1864; Redstone, 1908; Steers, 1926a; Grove, 1953; Cobb, 1957; Kidson *et al.*, 1958; Kidson and Carr, 1959; Kidson, 1961, 1963; Carr, 1962, 1965, 1967, 1969b, 1970, 1971c, 1972, 1973, 1986; Carr and Baker, 1968; Randall, 1977; Green and McGregor, 1986) but with the exception of Steers (1926a) and Carr (1965, 1967, 1969b, 1970, 1971c, 1972, 1973, 1986), most have considered either only part of the shingle structure, or have concerned themselves with it in a wider context (Redman, 1864; Kidson, 1963). Generally, writers agree that this is one of the largest and most important shingle structures on the British coast. They include Redman (1864) who referred to 'This extraordinary mole of shingle', Steers (1964a) who described it as '...the largest of the east coast shingle spreads', and Carr (1969b) who commented that it is 'one of the most important shingle formations on the coast of the British Isles'. In the HMSO report (1947) Orford Beach and Shingle Street was described as being 'of the very greatest importance and interest physiographically'.

The area had been used for military purposes from 1914 and as a bombing range during World War II (1939–1945), but this seems to have not greatly harmed its scientific interest. However, post-war military pressures were particularly damaging to the northern and ness areas of the site. At Orfordness, though it was still possible to map the sequence of ridge heights in the late 1960s, the evidence near Stonyditch was damaged by the extraction of shingle there and its transference by light railway to Slaughden, south of Aldeburgh, for a beach-nourishment scheme. Further damage resulted from the construction of an abortive early-warning system on Lantern Marshes in 1971. This affected the neighbouring shingle spreads, which were practically destroyed. More recently, Green and McGregor (1986) have assessed the geomorphological quality of features within the site, where significant geomorphological interest still remains. The lagoons noted by Cobb (1957) at Shingle Street have largely disappeared. Mostly the legacy of a previous phase of development of the River Ore, the lagoons have been the victims of the natural changes that take place at Shingle Street as the distal point of the spit changes its position. The southern part of the spit is outstanding, for it is virtually undisturbed.

Description

Orfordness comprises three elements: a storm beach undergoing erosion in the north, together with some intermittent shingle spreads; an extensive spit in the south where the shoreline is either accreting or being slowly eroded; and linking these, the ness proper. This is a cuspate foreland situated where there is a change in the orientation of the beach from approximately north–south to north-east–south-west. It consists of a complex series of ridges piled one against another. Such ridges extend from opposite Lantern Marshes (TM 458 525) as far as North Weir Point (TM 376 436) and across the mouth of River Ore to Shingle Street (TM 374 434) and Bawdsey (TM 359 401). Over much of this length, there is a systematic overall series of sub-parallel ridges, except at Shingle Street where ridges are less continuous and more ephemeral.

South of the lighthouse, with its complex pattern of shingle ridges, the spit narrows to reach its minimum width of under 50 m at high-water mark about 1.5 km north of the present-day river mouth. Over the whole of this length, as far as the distal end of the spit, the structure consists of a series of sub-parallel ridges, on the river side of which recurves may be present. Very rapid growth of the distal point and its linking with estuarial banks, or its breaching, may result in an extreme form of these with the recurves separated by tidal pools. The spit terminates at North Weir Point. Between North Weir Point and Shingle Street lies the River Ore, with one or more channels and extensive shoals. These shallow banks are areas of considerable size, which are exposed for at least part of the tidal cycle. They are subject to considerable change, which reflects both the previous environmental conditions and the stage in the development of the spit. Between 1955 and 1970, annual growth of the distal point varied between zero and 88 m. Over the long-term, variability is far greater.

In the northernmost part of the site (Figure 6.41), the ridge crests all fall below 1.5 m OD, that is, they are generally below present-day high water on spring tides. The next shingle structure to the south is below 2.0 m OD and most of the remaining area north of Stonyditch (TM 455 497) is less than 2.5 m OD. The broad increase in height continues immediately south of Stonyditch where most of the ridges fall into the ranges 3.0 m to 3.5 m and 3.5 to 4.0 m OD. There is an extensive area towards Stonyditch Point, where the ridge crests are below 3.0 m OD. Most of these ridges, which fall in the range 2.5 to 3.0 m OD are either cut by ridges on the seaward side or are truncated near The Crouch, but one or two extend along the spit as far as the present-day Havergate Island where the relationship between the ridges again becomes more complex.

Apart from the fairly extensive area of ridges in the range 3.0 to 3.5 m OD on the ness, where

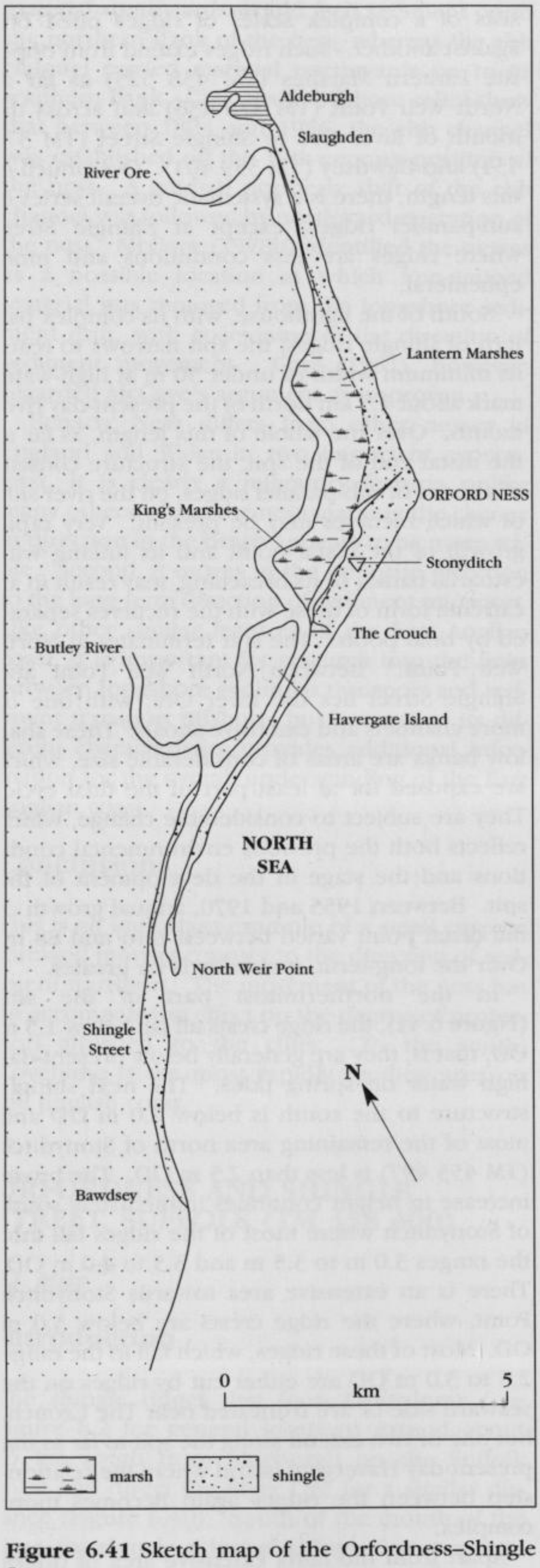

Figure 6.41 Sketch map of the Orfordness–Shingle Street area.

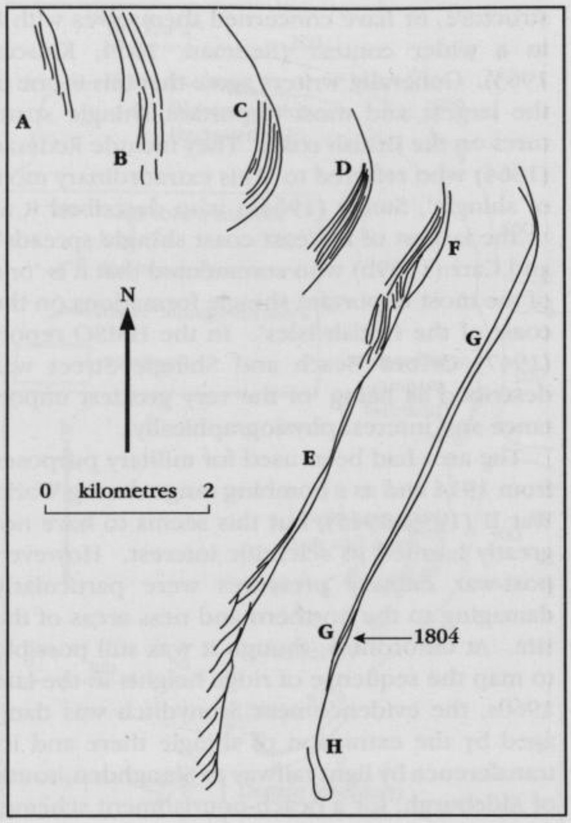

Figure 6.42 Schematic diagrams of shingle ridge groups representing developmental phases of Orfordness (A–B) spit development; (C–D) ness development; (E) extended spit with distal recurves; (F) additions to ness; (G–H) storm beach additions to spit. Each diagram portrays the north–south position accurately; the east–west position is arbitrary. (Based on Carr 1969b, 1972, 1973.)

they fall in height to seaward, only small areas of similar height occur elsewhere along the length of the spit. At the ness, both 2.5 to 3.0 m and 3.0 to 3.5 m groups are truncated on the seaward side by ridges between 3.5 and 4.0 m OD (Figure 6.43). These, together with the highest category (greater than 4.0 m OD with a maximum of approximately 5.0 m OD), extend throughout the spit to the distal point.

The height between the shingle ridges ('fulls') and the intervening hollows ('swales' or 'lows') varies. Generally there is about 0.3 m difference between adjacent ridges and hollows but occasionally this reaches a maximum of 1.5 m. The base of the shingle under the ness, and much of the nearby area, lies between −0.2 and −10.7 m OD. It appears to rest on marine planation surfaces of various ages. Shingle forms the river

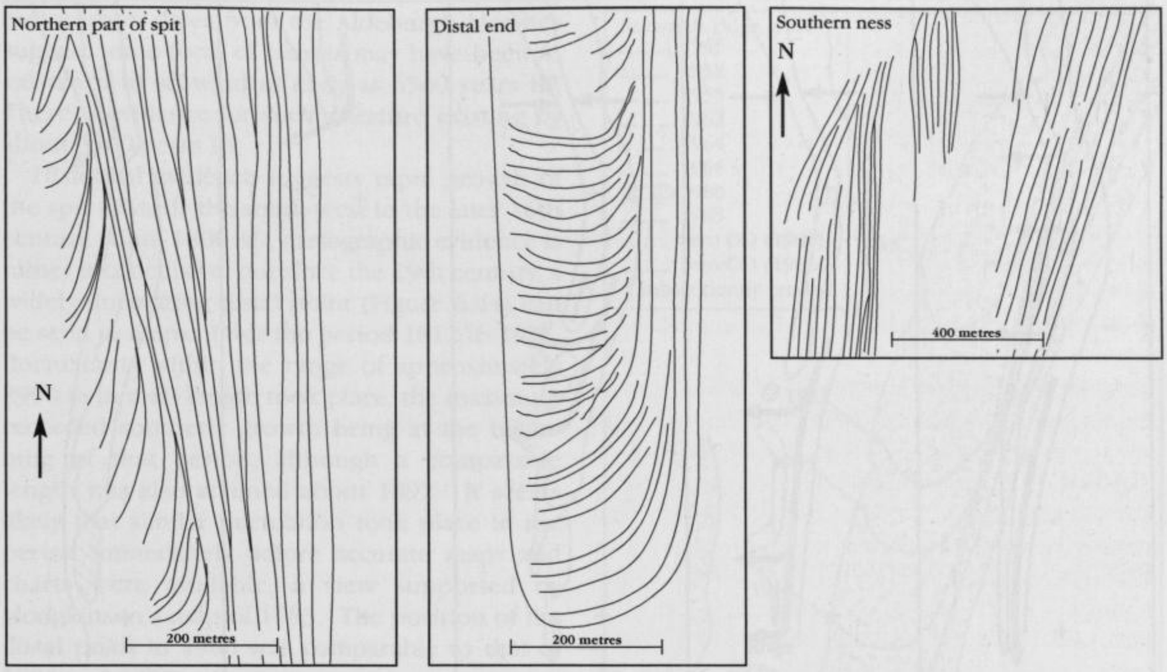

Figure 6.43 Variations in ridge patterns of Orfordness, in the southern part of the ness, the northern part of the spit with an earlier recurved spit fronted by individual shingle ridges, now largely destroyed, and also at the distal end of the spit, showing recurves. (Based on Carr, 1973; Green and McGregor, 1988.)

bed at The Narrows (TM 420 475), and one borehole records 'stone' under the estuarine clays on the northern part of Havergate Island from –10.1 to –14.0 m OD.

The well-rounded clasts of the ness and the spit range between 4 mm and 75 mm in length. Over 99% are flint. Elsewhere in the site, if pebbles are present, they are found in small quantities only: they are frequently rectangular rather than rounded, and are slightly more varied in composition. Most occur in a primarily sandy matrix.

The River Ore runs roughly parallel to the spit from Slaughden to North Weir Point (Figure 6.41). Its bed ranges in depth from –5.0 m to –12.5 m OD. Except near the present-day river mouth, the maximum depth at any given place has remained almost constant for the last 160 years and probably longer. The mouth of the Ore appears to have been displaced towards the south as the shingle structure lengthened in that direction. This impression is correct only in part. Displacement is likely both immediately south of Aldeburgh and south of The Crouch, although at the latter site it was complicated by the precursors of Havergate Island and the Butley River. Elsewhere, existing creeks, which ran approximately parallel to the coastline, were joined together as their exits became blocked.

Marshes that have been the subject of landclaim are present on the landward side of the River Ore throughout its whole extent. They vary in width from 0.4 to 2.4 km. Marshes also occur at Havergate Island (the interest in which is now almost entirely ornithological), and from Slaughden to The Crouch on the southern side of the river. On the spit farther south, there are only small areas of saltings. The reclaimed marshes are at –0.3 m to +0.6 m OD and the saltmarshes at 1.2 m to 1.5 m OD. King's Marshes are separated from the shingle of the ness by a tidal creek (Stonyditch), which runs approximately north-east–south-west. The truncated head of this creek rises in an area of saltmarsh between two series of ridges, so that northward of this point the shingle ridges and marshes are adjacent. This is the only instance where the major shingle structure abuts the marshes. Borehole logs suggest that at least some of the shingle and estuarine clay were deposited contemporaneously. The differing depths at which bands of shingle occur both in nearby boreholes and within the same borehole indicate how complex the sequence of events must have been. Although the ridges south of The Crouch are about 1000 years old or less,

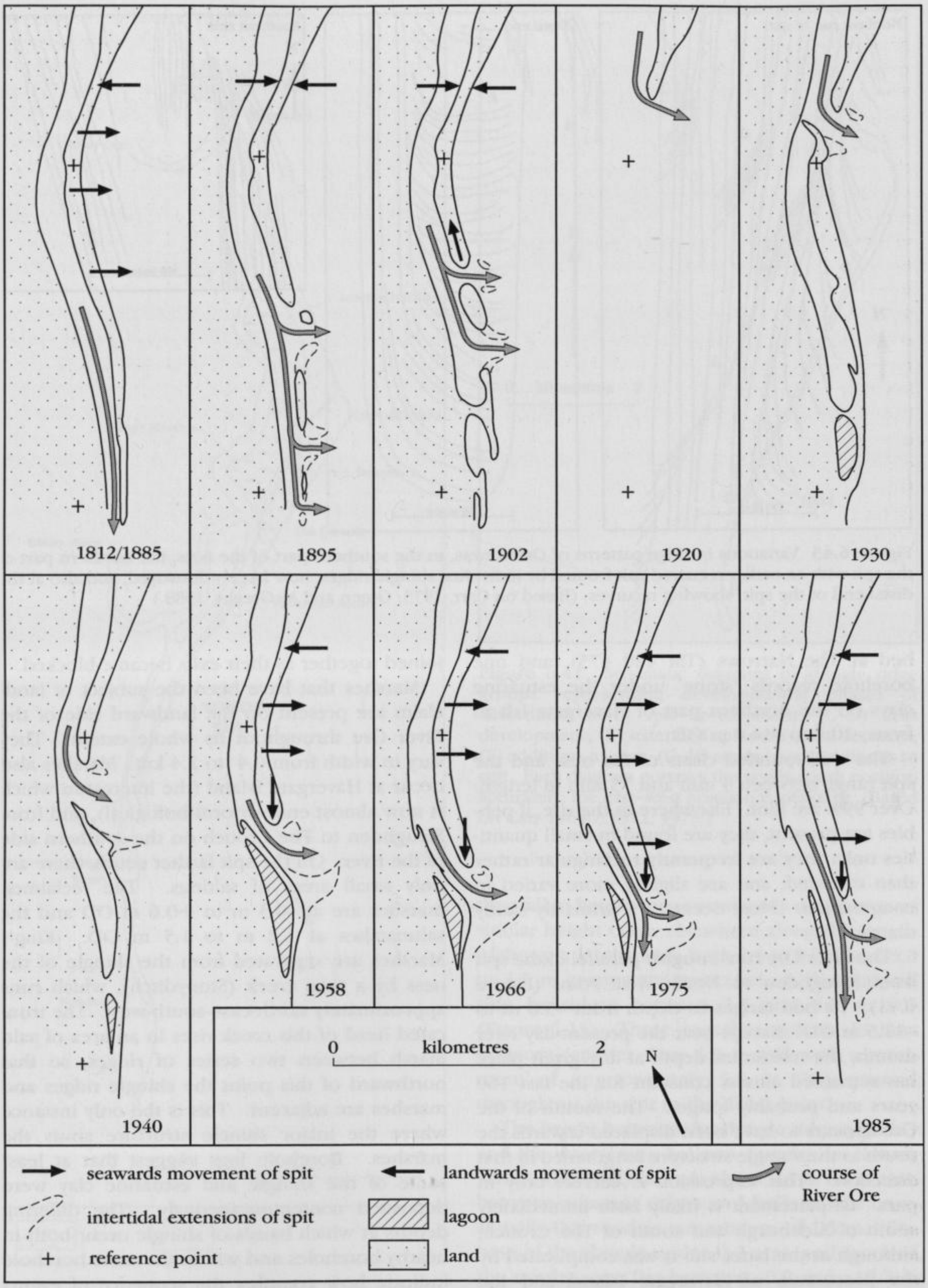

Figure 6.44 Historical changes in the position of distal features at Orfordness. (After various authors, mainly Carr, 1965; and Green and McGregor, 1988.)

radiocarbon dates from the Aldeburgh Marshes suggest some form of barrier may have been in existence to seaward as early as 6500 years BP. There is evidence for such a feature existing by about 3500 years BP.

Historical evidence suggests rapid growth of the spit towards the south-west in the later 16th century. Until 1800 AD, cartographic evidence is rather inconclusive, but since the 19th century, a widely fluctuating distal point (Figure 6.44) can be seen in maps. Over the period 1812 to 1921, fluctuations within the range of approximately 2900 m in total length took place, the maximum recorded southerly growth being at the beginning of that period, although a comparable length was also attained about 1892. It seems likely that similar fluctuation took place in the period immediately before accurate maps and charts were available, a view supported by Hodgkinson's map of 1783. The position of the distal point in 1980 was comparable to that of 1804 and 1902.

Interpretation

The stretch of coast that comprises the Orford shingle spit and the estuary of the River Ore has been the subject of extensive geomorphological research, especially during the period 1955–1970, and several new concepts or modifications of previous ideas about such environments have resulted (e.g. the 'counter-drift' concept of Kidson, 1963). In Britain, Carr has argued that it is the only remaining natural, dynamic and sustainable cuspate foreland, as well as being an outstanding example of a shingle spit.

Carr (1962, 1965, 1967, 1969b, 1970, 1971c, 1972, 1973, 1986) demonstrated the path of the shingle across the estuary, the absence of supply of material from offshore, the way in which new ridges may be melded onto earlier ones without obvious trace, and the inter-relation of spit, bar, banks and the Shingle Street features. The southward progression of the distal point results in the landward recession of the shoreline at Shingle Street. Nevertheless, each sequence of spit development has occurred in the same lateral position, probably due to the artificially constrained channels of the River Ore, which prevent landward migration of the position of the spit. In this respect, the spit differs from many other sites, such as Spurn Point, which have a history of spit breaching and lateral displacement.

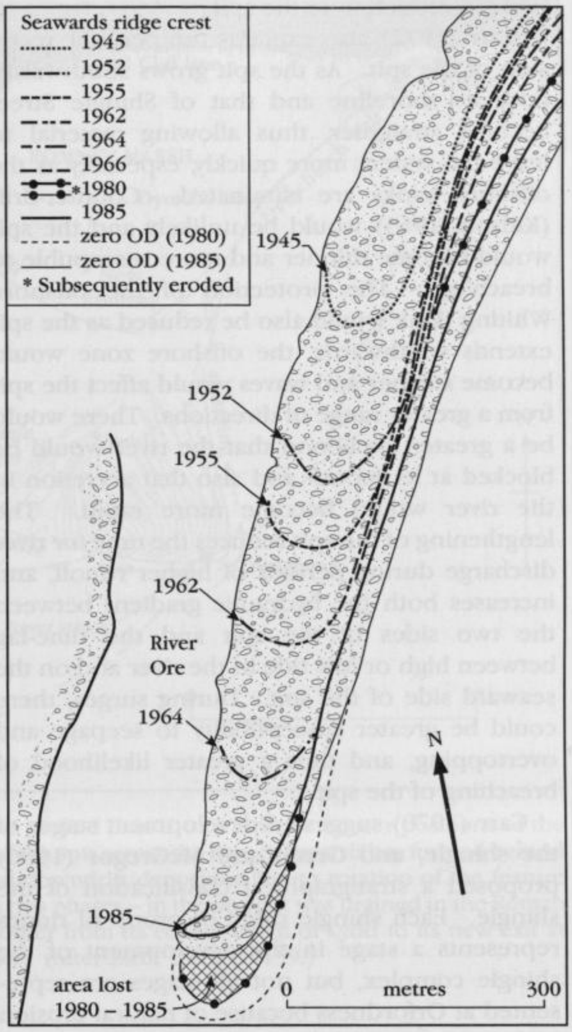

Figure 6.45 Historical distal changes at Orfordness. showing development of major ridge crests.

There appears to be no direct relationship between the extension of Orford spit and wave incidence, for instance, there is no correlation between annual southward growth and the prevalence of winds from a north-easterly quarter, a relationship that might have explained the longshore movement of shingle along the spit (Steers, 1926a). Carr (1986) suggested that the changing position of the distal point followed a cyclic pattern of development (Figure 6.44). It is not possible to confirm this model because of a lack of early records and a gap in the recent record. Carr suggests that the last breach occurred in 1920, but there is no mention of this in Steers' account. Carr's model would predict a rapid southward extension of the spit in the near future followed by breaching and an equally rapid reduction in the spit.

Carr (1972) also explains the periodic recession of the spit. As the spit grows southwards, both its shoreline and that of Shingle Street become straighter, thus allowing material to leave the system more quickly, especially as the offshore banks are eliminated. Counter-drift (Kidson, 1963) would be unlikely and the spit would become thinner and more susceptible to breaching. The protection of the offshore Whiting Bank would also be reduced as the spit extends southwards, the offshore zone would become steeper and waves would affect the spit from a greater range of directions. There would be a greater likelihood that the river would be blocked at its mouth and also that accretion in the river would become more rapid. The lengthening of the spit reduces the time for river discharge during periods of higher runoff, and increases both the hydraulic gradient between the two sides of the spit and the time-lag between high or low tide in the river and on the seaward side of the spit. During surges, there could be greater susceptibility to seepage and overtopping, and thus a greater likelihood of breaching of the spit.

Carr (1970) suggested development stages of the shingle, and Green and McGregor (1986) proposed a stratigraphical classification of the shingle. Each shingle ridge or group of ridges represents a stage in the development of the shingle complex, but not all stages are represented at Orfordness because of natural erosion and human activity. Green and McGregor (1986) argue that, in any area of Orfordness, an individual ridge or group of ridges can be classified in terms of the extent to which the stage that it represents is found elsewhere in the complex. The continuity and relationships of groups of shingle ridges can be traced using aerial photographs, and the shingle system can be divided into development stages and sub-stages. These are either individual ridges partly or entirely isolated in marshland, or groups of ridges separated from one another by a major erosional contact. In several cases, sub-stages consist of several groups of ridges (usually 5–10 ridges per group) with similar alignments, but separated from each other by erosional contacts. Green and McGregor (1986) believe that their interpretation confirms, and in places refines, the pattern of development stages proposed by Carr (1970).

Where successive stages are in direct contact with each other, the sequence and relative age of the ridges can be ascertained. Continuity of development is poorly represented in contrast to Dungeness, reflecting the limited extent of the shingle and its greater susceptibility to natural change. In particular, the limited seaward progradation of the ness itself over less than 1 km has allowed only 50 to 60 ridges to develop in contrast to more than 500 at Dungeness. Ridge continuity has been interrupted by natural erosion, usually as a consequence of the narrowness of the spit and its extreme elongation. The ness is also migrating towards the south. Green and McGregor (1986) have regarded overlapping groups of ridges in erosional contact as being separate development stages whenever their age relationships are uncertain. The small, frequently isolated, areas of shingle on the landward side of the estuary are also regarded as separate development stages.

Conclusions

Orfordness comprises three elements: a storm beach in the north, the ness itself, and a spit in the south. One of the largest and most important shingle structures on the British coast, it is an outstanding example of a shingle spit and shingle-spit cuspate foreland complex. Orford Beach and Shingle Street have the very greatest physiographical importance and interest. The international reputation of Orfordness rests primarily upon the history of scientific investigation and the existence of detailed records of its early growth and present-day dynamics.

DUNGENESS AND RYE HARBOUR, KENT AND EAST SUSSEX

V.J. May

Introduction

The large shingle cuspate foreland of Dungeness and the associated beaches at Rye (see Figure 6.2 for general location and Figure 6.46) provide a record of the development, partial destruction and reconstruction of a very large shingle barrier beach and spit system. Former shorelines, which are thought to have formed during approximately the last 5500 years, include both exposed shingle ridges and a large number of

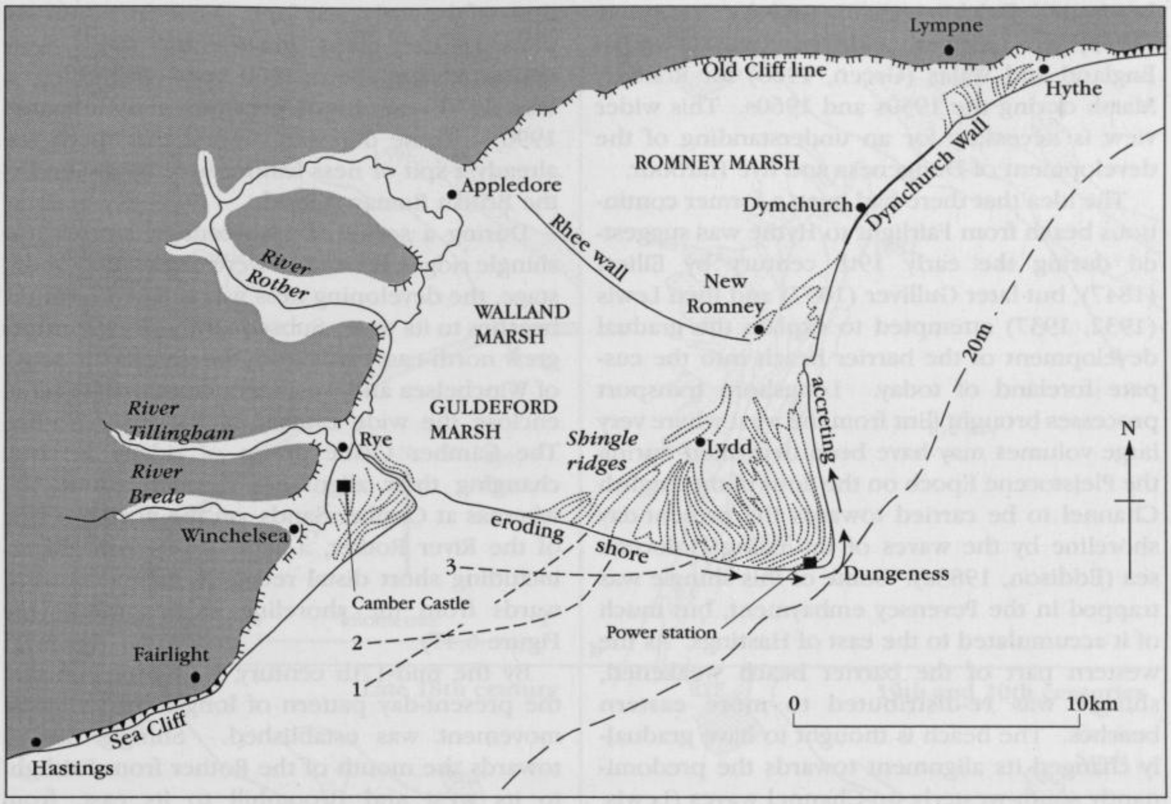

Figure 6.46 The cuspate foreland, Dungeness, Kent. The pecked lines 1 to 3 indicate former positions of the original spit over time, showing the downdrift extension of the spit across the bay. Saltmarsh has formed behind the outer shingle barrier. Over time, updrift erosion and downdrift deposition led to rotation of the feature from position 1 to 3. Land-claim of the marsh occurred in two phases – in the north it was drained in the Roman period, and in the 13th century diversion of the River Rother from its course north of Lydd to its new exit at Camber Castle led to the draining of the southern marshes. (After Bird, 1984, p. 159.)

buried ridges, Dungeness itself being especially noteworthy for its sequence of some 500 progressively younger, but increasingly cuspate eastwards, ridges. Bare shingle occurs at Rye Harbour and Camber, Dungeness and Hythe. At Rye Harbour lies the westernmost group of shingle beaches, which extend across the former Romney Marsh embayment between the Fairlight Hills east of Hastings and the former sea cliffs at Hythe. Unlike the beaches of Dungeness and Hythe, the development of the beach ridges at Rye Harbour has taken place mainly since the 16th century. At Camber Castle, the exposed shingle ridges generally post-date the mid-16th century (Lovegrove, 1953), whereas at the western end of the Lydd Ranges, they may be over 3500 years old (Eddison, 1983a,b). Between New Romney and Hythe (see Figure 6.46), a large number of ridges at high angles to the present-day shoreline were described by Elliott (1847) and have been detected in the distribution of the Beach Bank soil series and in the exposed shingle ridges west of Hythe. Both dunes and beach ridges are found at the foot of former sea-cliffs at west of Hythe, where there is considerable archaeological interest in the relationship between the Roman and Saxon forts at Lympne and the nature of a navigable inlet behind the shingle ridges (Shackley, 1981). Much of the human history of the marshlands has involved land-claim and drainage. The development of the marshes, land-claim and drainage within Romney Marsh (*sensu lato*) are discussed in Eddison (1995), Eddison and Green (1988) and Eddison *et al*. (1998).

Although the development of the features at Rye, Camber, Dungeness, Romney and Hythe are interrelated, they form separate physiographic units. Eddison (1983a,b) summarized the main phases of the evolution of the barrier beaches, expanding upon the detailed surveys carried out during the 1930s by Lewis (1932, 1937), by

Lewis and Balchin (1940), and by Lovegrove (1953) for Camber, and the Soil Survey of England and Wales (Green, 1968) for Romney Marsh during the 1950s and 1960s. This wider view is necessary for an understanding of the development of Dungeness and Rye Harbour.

The idea that there had been a former continuous beach from Fairlight to Hythe was suggested during the early 19th century by Elliott (1847), but later Gulliver (1897) and then Lewis (1932, 1937) attempted to explain the gradual development of the barrier beach into the cuspate foreland of today. Longshore transport processes brought flint from the west where very large volumes may have been deposited during the Pleistocene Epoch on the floor of the English Channel to be carried towards the present-day shoreline by the waves of the rising Holocene sea (Eddison, 1983a). Some of this shingle was trapped in the Pevensey embayment, but much of it accumulated to the east of Hastings. As the western part of the barrier beach weakened, shingle was re-distributed to more eastern beaches. The beach is thought to have gradually changed its alignment towards the predominantly south-westerly up-Channel waves (Lewis, 1931).

Eddison (1983a,b) suggested that the earliest features, a submerged forest at Cliff End and the low-lying shingle deposits at Broomhill and Sandylands, represent an early barrier beach, probably dating from between 5500 and 4000 years BP. The barrier progressively extended towards Hythe by a series of recurves, identified by Elliott (1847). Thus with the shingle emplaced close to the present-day sea level, longshore transport gradually moved much of the shingle into the Romney embayment. The eastern end of the barrier beach extended towards, but never joined, the former cliffs around Lympne, so providing shelter for both Roman and Saxon vessels approaching the fort at Lympne, a site now about 3 km from the coastline.

Between Jury's Gap (TQ 993 180) and Dungeness itself, there are about 500 individual ridges in four main groups: their alignment changes by about 10° between each group. Eastwards from Galloway's Lookout (TR 045 172), the ridges are characterized by increasing curvature and the preservation of a ness form. Some of the western exposed shingle (for example, Jury's Beach, The Forelands and Holmstone) probably represents recurved sections of the early ness form. Near the power stations (Figure 6.46), gravels and sands were deposited for about 1900 years beginning at least 3270 years BP (Greensmith and Gutmanis, 1990). These deposits suggest that there was already a spit or ness feature here by the end of the British Roman period.

During a series of 13th century storms, the shingle ridges at Camber were destroyed. At this stage, the developing ness was isolated from the beaches to its west. Subsequently, shingle ridges grew north-eastwards from the area to the south of Winchelsea and westwards from Broomhill to enclose the wide estuary of the River Rother. The Camber Castle group of ridges fan out, changing their alignment through about 50°, whereas at Camber Sands, on the northern side of the River Rother, a series of narrow ridges, including short distal recurves, extended westwards from the shoreline at Broomhill (see Figure 6.47).

By the mid-17th century, it is probable that the present-day pattern of longshore sediment movement was established. Shingle moved towards the mouth of the Rother from Fairlight to its west and Broomhill to its east, from Broomhill along the southern shoreline of Dungeness to the ness itself and thence northwards to Greatstone, and from St Mary's Bay southwards towards Littlestone. At Hythe, shingle moved eastwards towards the Lympne inlet, as appears always to have been the case. Much of the barrier beach at Romney has been built upon and the ridges at Hythe retain little of their original form. The modern shoreline is reinforced by sea-walls at Winchelsea Beach, Broomhill, Littlestone and Dymchurch, with artificial beach feeding at Cliff End, Jury's Gap, the Dungeness power station and St Mary's Bay, with over 110 000 m^3 of material being added in 1979 (Eddison, 1983a), mostly by re-distribution from within the site. Nevertheless the southern shoreline continues to be eroded and shingle added to the eastern shoreline, the landward movement of the southern shoreline often taking place in stormy periods when it may move several tens of metres inland.

The present-day features of Dungeness and the associated beaches at Rye, Romney and Hythe thus result from changing sediment transport rates and deposition over some 5000 years, the different alignments of the shingle ridges and the buried beach ridges demonstrating gradual development from a barrier beach to the

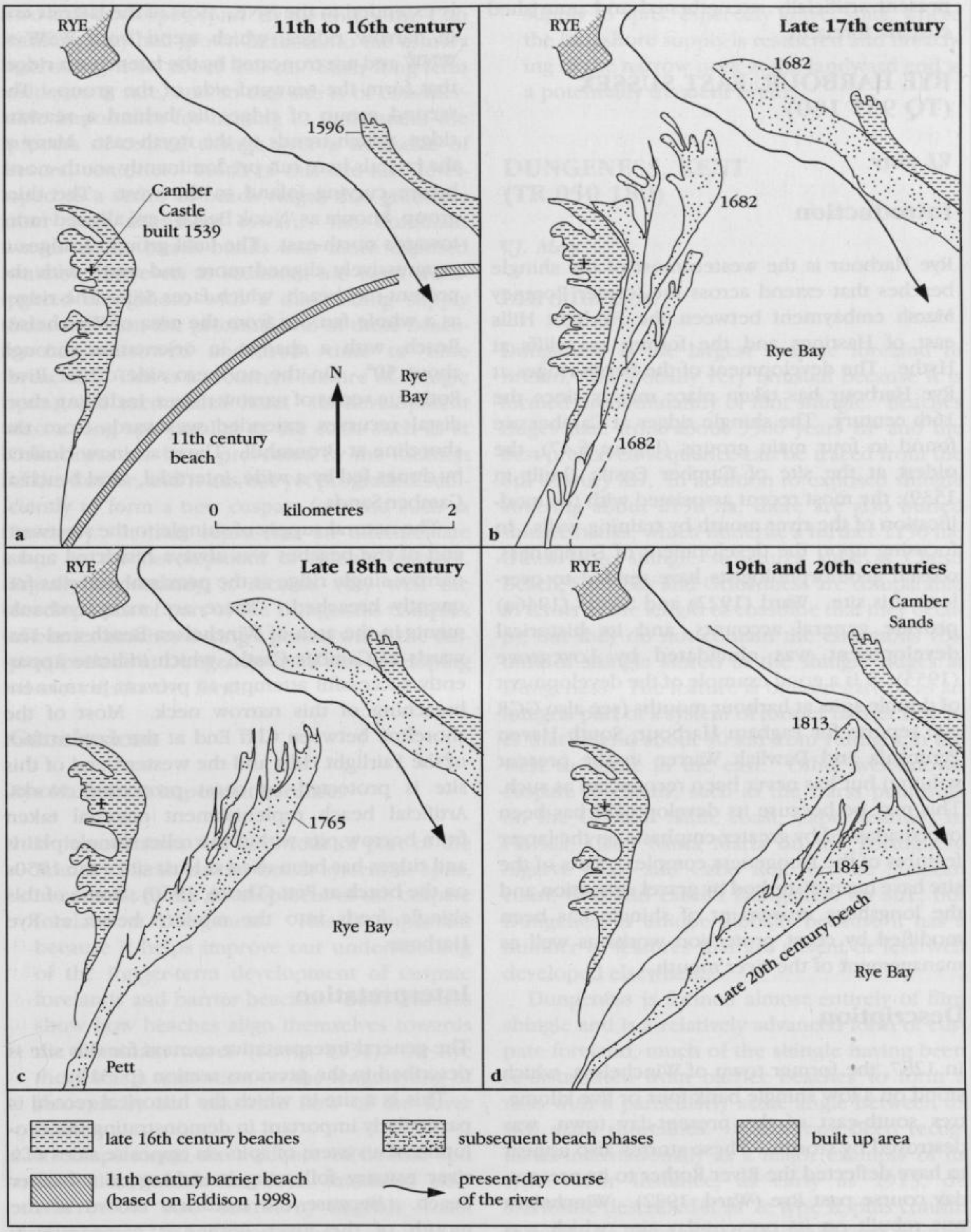

Figure 6.47 The historical evolution of Rye Bay. Dates indicate shoreline and beach area from contemporary maps and charts. (After Lovegrove, 1953; and Eddison, 1998.)

present artificially strengthened and nourished cuspate foreland.

RYE HARBOUR, EAST SUSSEX (TQ 935 180)

V.J. May

Introduction

Rye Harbour is the westernmost of the shingle beaches that extend across the former Romney Marsh embayment between the Fairlight Hills east of Hastings and the former sea cliffs at Hythe. The development of the beach ridges at Rye Harbour has taken place mainly since the 16th century. The shingle ridges at Camber are found in four main groups (Figure 6.47), the oldest at the site of Camber Castle (built in 1539), the most recent associated with the modification of the river mouth by training walls. In focusing upon the development of Dungeness, coastal geomorphologists have tended to overlook this site. Ward (1922) and Steers (1946a) provide general accounts, and its historical development was elucidated by Lovegrove (1953). It is a good example of the development of double spits at harbour mouths (see also GCR site reports for Pagham Harbour, South Haven Peninsula and Dawlish Warren in the present volume) but has never been recognized as such. This may be because its development has been overshadowed by greater emphasis on the larger features of the Dungeness complex. Parts of the site have been disturbed by gravel extraction and the longshore movement of shingle has been modified by coast protection works as well as management of the river mouth.

Description

In 1287, the former town of Winchelsea, which stood on a low shingle bank four or five kilometres south-east of the present-day town, was destroyed by storms. These storms also appear to have deflected the River Rother to its present-day course past Rye (Ward, 1922). Winchelsea was rebuilt on its present-day site, which was then accessible by sea. Shingle ridges grew north-eastwards from the area to the south of Winchelsea and by the mid-16th century provided the site for Camber Castle (Figure 6.47).

The Camber Castle group of ridges are mainly recurving distal features of a spit that gradually extended to the NNE. Most of the laterals end in narrow ridges, which trend from WSW to WNW and are truncated by the later storm ridges that form the seaward side of the group. The second group of ridges lie behind a seaward ridge, which trends to the north-east. Many of the laterals here run predominantly south–north before curving inland to the west. The third group, known as 'Nook Beach', are aligned more towards north-east. The final group of ridges is progressively aligned more and more with the present-day beach, which faces SSE. The ridges as a whole fan out from the area of Winchelsea Beach, with a change in orientation through about 50°. On the northern side of the River Rother, a series of narrow ridges, including short distal recurves extended westwards from the shoreline at Broomhill. These are now cloaked by dunes fed by a wide, intertidal, sand beach at Camber Sands.

The natural supply of shingle to the westward end of the beaches was always restricted and a narrow single ridge at the proximal end was frequently breached. There are many embankments in the area of Winchelsea Beach and seawards of Camber Castle, which indicate apparently successful attempts to prevent permanent breaching of this narrow neck. Most of the shoreline between Cliff End at the eastern foot of the Fairlight Hills and the western end of this site is protected by coast protection works. Artificial beach replenishment material taken from borrow pits within the relict shingle plains and ridges has been used at least since the 1950s on the beach at Pett (Thorn, 1960). Some of this shingle feeds into the modern beach at Rye Harbour.

Interpretation

The general interpretative context for this site is described in the previous section (p. 310).

This is a site in which the historical record is particularly important in demonstrating the evolution of a system of spits on opposite sides of a river estuary following breaching of a barrier beach. Because the historical record at the mouth of the River Rother is comparatively good, it is possible to recognize the growth of two beaches, terminating in recurved spits, into the estuary (Figure 6.47). By the end of the 16th century, according to the cartographic evidence examined by Lovegrove (1953), the Rother estuary was about 2 km across and was bounded by two well-developed spits. By the end of the 17th century, they had grown farther into the estuary narrowing it to about 400 m. Such long-term evidence is rare, and so this site is of considerable importance for helping to understand the process of coastal change over a timescale of several centuries. Much of this site has developed as a series of beach ridges that gradually alter their alignment towards the dominant waves as the beach builds into more exposed waters. However, as sediment has been transported alongshore with a diminishing supply from the west, the proximal end of these beaches has narrowed and from time to time breached. This is a recurrent feature of shingle spits, well exemplified here. Its development into a long spit parallel to the coast such as at Orfordness has been prevented by the location of the Rother, and it has not yet prograded sufficiently to form a new cuspate foreland such as Pevensey. It thus represents an intermediate stage in the development of shingle features. Equally importantly, it records very well the development of part of the Dungeness complex of barrier beaches during a period when the cuspate foreland at Dungeness itself was developing its most distinctive form.

Conclusions

Rye Harbour is important because:

1. It demonstrates the behaviour of part of the Romney Marsh barrier beach system as spits, in contrast to the development of the cuspate foreland at Dungeness. This is important because it helps improve our understanding of the longer-term development of cuspate forelands and barrier beaches. Both features show how beaches align themselves towards the dominant waves (Lewis, 1931). At Rye there was a restriction on the lengthening of the spit by the estuarine flow of the River Rother.
2. It complements the evidence of double spit formation, which has been examined elsewhere on the southern English coast (Robinson, 1955; Kidson, 1963).
3. It provides a good example of beach plain development where the dominant forms are large numbers of laterals, unlike eastern Dungeness where successive storm beaches form the beach plain.
4. It exemplifies the problem of the sediment supply to spits, especially gravel spits, where the longshore supply is restricted and breaching of the narrow neck at the landward end is a potentially frequent event.

DUNGENESS, KENT (TR 050 180)

V.J. May

Introduction

Dungeness is the largest cuspate foreland in Britain, and globally very unusual because it is formed predominantly of flint shingle. Beaches ridges date from about 5500 years BP and the best-preserved sequence can be traced from the 8th century AD. In addition to exposed shingle covering about 2158 ha, there are also buried shingle banks, which underlie a further 1150 ha. Other large shingle structures such as Chesil Beach, Spey Bay and Orfordness are comparable in terms of the length of coastline that they occupy, but they do not contain the enormous volume of shingle stored in the shingle ridges at Dungeness. The feature is often regarded as an integral part of a system of former barrier beaches that extend about 40 km from Fairlight in the west to Hythe in the east. Other well-known cuspate forelands, such as the Darss peninsula on the German Baltic coast, Cape Kennedy in Florida, Cabo Santa Maria on the Portuguese Algarve coast and Cabo Rojo on the Mexican coast, rival and exceed Dungeness for size, but Dungeness is unique globally because it has a number of features that are absent or less well developed elsewhere.

Dungeness is formed almost entirely of flint shingle and is a relatively advanced form of cuspate foreland, much of the shingle having been re-distributed from barrier beaches to form a ness with a particularly acute angle between its two main shorelines. It has long been recognized internationally as a major example of its type. For instance, as early as 1913, de Martonne described it as 'le type le plus connu: la pointe de Dungeness'. Standard texts from all parts of the world refer to Dungeness as the best-known example of a cuspate foreland (e.g. Holmes, 1944, 1965; Zenkovich, 1967; Bird, 1968, 1984; Paskoff, 1985).

No area inland of beaches to have been occupied and land-claimed over so long a period of time (about 1200 years) has been documented so intensively as Dungeness, and the documentary record extends over a far longer period than for any comparable site.

Finally, in contrast to many similar features, it lacks an offshore shoal that might extend its form seawards.

The Soil Survey of England and Wales (Green, 1968) has shown that shingle ridges often extend many hundreds of metres beyond the area of exposed shingle, the Beach Bank soil series representing the distal parts of successive beach ridges. Parts of the Lydd soil series also lie above shingle, while the Lydd series itself and parts of the Greatstone series are dominated by sand and loamy sand, which may be derived from sandy beaches associated with the shingle beaches in much the same way as sandy beaches are found today on the eastern shoreline of Dungeness. Recent archaeological and geomorphological studies have built on the work of the Soil Survey.

Large areas have been damaged by gravel extraction, vehicle tracks, military training areas and the construction of the Dungeness group of nuclear power stations. Detailed assessments of the damage have been made by Fuller (1985) and Green and McGregor (1986), both reports being drawn upon extensively in the assessment of this GCR site.

Description

The present-day shoreline at Dungeness is formed by a southern beach that faces SSW and is gradually moving north and inland over older relict beach ridges, the acute bend of the ness itself, which is migrating SSE, and an eastern beach, which has gradually migrated eastwards as the ness has grown. Much of this eastern beach is fronted by a wide intertidal sand beach. Landward of the present-day beach, there is a sequence of buried and exposed shingle ridges, which become both younger and more curved towards the east. Waller (1993, 1994) suggests that peat dating from 1100 to 2000 years BP helps date the development of both Romney Marsh and Dungeness. The oldest beaches, buried at Broomhill and Sandylands, have been tentatively dated between 5500 and 4000 years BP (Eddison, 1983a). Between Jury's Gap (TQ 993 180) and Dungeness itself, about 500 indi-

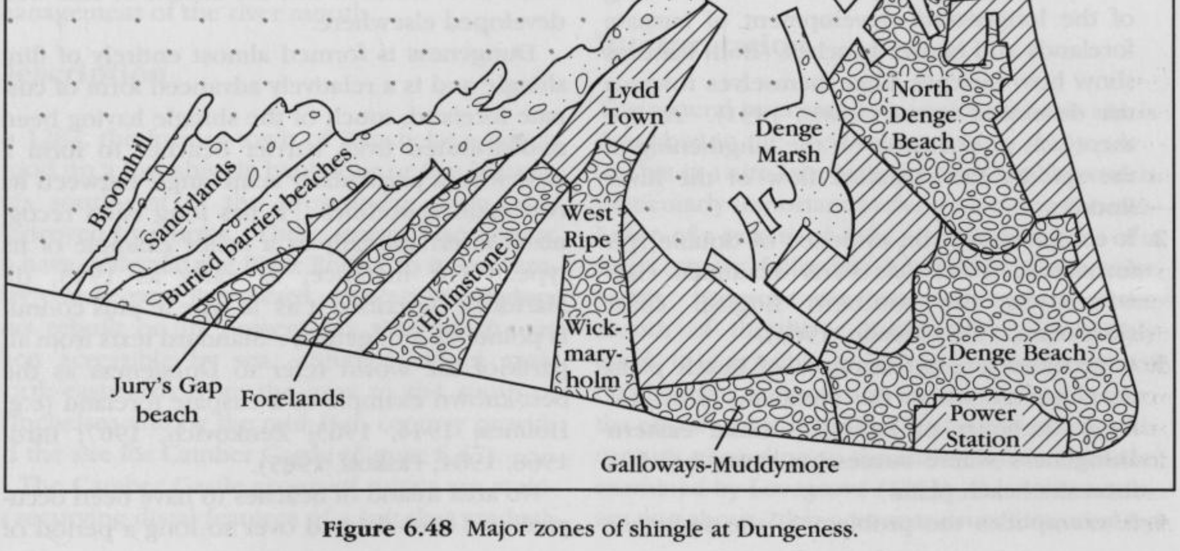

Figure 6.48 Major zones of shingle at Dungeness.

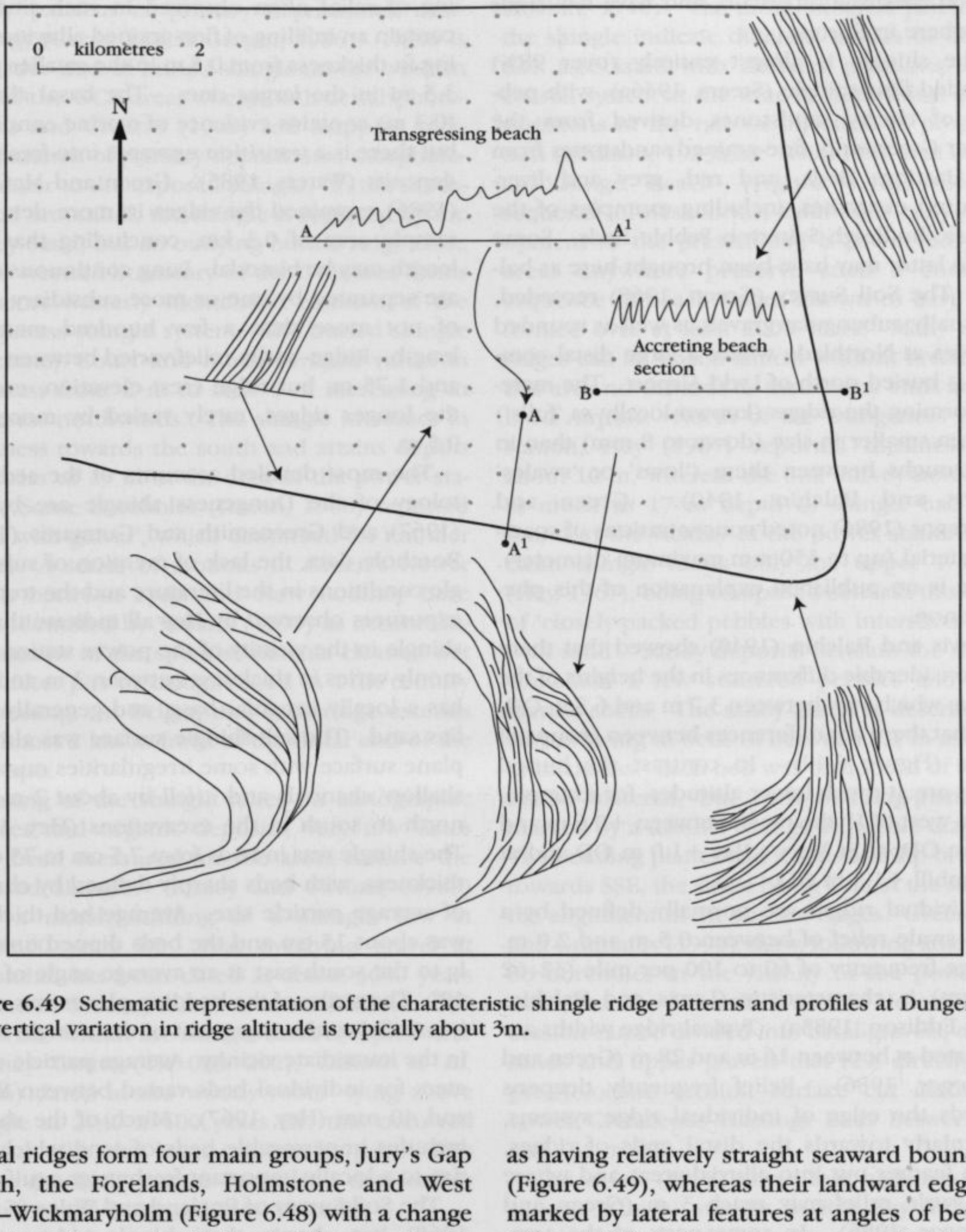

Figure 6.49 Schematic representation of the characteristic shingle ridge patterns and profiles at Dungeness. The vertical variation in ridge altitude is typically about 3m.

vidual ridges form four main groups, Jury's Gap beach, the Forelands, Holmstone and West Ripe–Wickmaryholm (Figure 6.48) with a change in ridge alignment of about 10° towards the north-west between each group. Eastwards of Galloway's Lookout (TR 045 172), the ridges become more curved and preserve earlier ness forms. At their seaward (southern) edge, the ridges are truncated by a single continuous ridge that forms the present-day shoreline. Inland, many of these curved ridges are buried, but represented by the Beach Bank soil series, which is described as associated with 'old inland pebble banks and beaches' (Green, 1968). The mapped distribution of these old beaches portrays them as having relatively straight seaward boundaries (Figure 6.49), whereas their landward edges are marked by lateral features at angles of between 30° and 40° to the main ridges, probably representing recurves in the distal part of the beaches. Only in the area around Open Pits (TR 073 180 and TR 074 185) are distal recurves completely preserved in the bare shingle. To the north of the Open Pits and the present-day ness, the shingle is found in over 100 sub-parallel south–north-trending ridges that extend northwards to very short, buried, distal forms. Although parts of this area have been disturbed by gravel extraction, the undisturbed parts remain the finest example of a gravel strandplain on the coastline of Britain, and have few rivals elsewhere in Europe.

The shingle is almost entirely (over 98%) rounded flint shingle (Steers, 1946a), with pebbles of cherty sandstones derived from the Upper Greensand, fine-grained sandstones from the Hastings Beds, and red, grey and liver-coloured quartzites, including examples of the Triassic Budleigh Salterton Pebble Beds. Some of the latter may have been brought here as ballast. The Soil Survey (Green, 1968) recorded, unusually, subangular gravels as well as rounded pebbles at Northlade where a large distal complex is buried north of Lydd Airport. The material forming the ridges (known locally as 'fulls') is often smaller in size (down to 8 mm) than in the troughs between them ('lows' or 'swales' (Lewis and Balchin, 1940). Green and McGregor (1986) noted concentrations of coarser material (up to 150 mm maximum diameter). There is no published explanation of this phenomenon.

Lewis and Balchin (1940) showed that there are considerable differences in the heights of the ridges, which vary between 3.7 m and 6.3 m OD, and that there are differences between groups of ridges (Figure 6.49). In contrast the buried ridges are at much lower altitudes, for example, ridges west of Hythe reach between +0.6 m and −1.0 m OD, at St Mary's Bay +1.0 m OD and at Broomhill +1.5 m OD.

Individual ridges are normally defined by a ridge–swale relief of between 0.5 m and 2.0 m. A ridge frequency of 60 to 100 per mile (32–62 per km) is characteristic (Lewis and Balchin, 1940; Eddison, 1983a). Typical ridge widths are estimated at between 16 m and 28 m (Green and McGregor, 1986). Relief frequently deepens towards the edge of individual ridge systems, particularly towards the distal ends of ridges, which feather out into alluvial areas and where ridge-swale relief may reach 3 m (Green and McGregor, 1986). In some parts of the area, individual ridges are rarely continuous over long distances (Lewis and Balchin, 1940). Parasitic ridges are common, inter-ridge recurves also occur in many sectors, and natural pits have often formed where shingle-branching or recurve patterns give rise to enclosed hollows. Irregular, anastomosing ridge patterns also occur (Green and McGregor, 1986). These features increase in frequency towards the distal end of individual ridge systems. Natural pits whose formation is encouraged by the deepening of relief often observed in such situations contain an infilling of fine-grained alluvium varying in thickness from 0.5 m in the smaller pits to 3.5 m in the larger ones. The basal alluvium (0.3 m) contains evidence of marine conditions, but there is a transition upwards into freshwater deposits (Waters, 1985). Green and McGregor (1986) examined the ridges in more detail in a sample area of 0.3 km, concluding that ridge length may be bimodal. Long continuous ridges are separated by one or more subsidiary ridges of not more than a few hundred metres in length. Ridge–swale relief varied between 0.5 m and 1.75 m, but ridge crest elevation, even on the longer ridges, rarely varied by more than 0.5 m.

The most detailed accounts of the sedimentology of the Dungeness shingle are by Hey (1967) and Greensmith and Gutmanis (1990). Borehole data, the lack of mention of sub-shingle conditions in the literature and the trenched exposures observed by Hey all indicate that the shingle in the vicinity of the power station commonly varies in thickness between 3 m and 7 m, has a locally irregular base, and generally overlies sand. The sub-shingle surface was almost a plane surface with some irregularities caused by shallow channels and it fell by about 2 m from north to south in the excavations (Hey, 1967). The shingle was in beds from 7.5 cm to 75 cm in thickness, with beds sharply defined by changes of average particle size. Average bed thickness was about 15 cm and the beds dipped uniformly to the south-east at an average angle of 8° to 10°. The strike of the bedding planes was exactly parallel to the alignment of the shingle ridges in the immediate vicinity. Average particle diameters for individual beds varied between 8 mm and 40 mm (Hey, 1967). Much of the shingle includes impermeable beds of sand which give rise to a locally important freshwater aquifer.

The Soil Survey of England and Wales (Green, 1968) has shown that shingle ridges often extend many hundreds of metres beyond the exposed shingle, the Beach Bank soil series representing the distal parts of successive beach ridges. Parts of the Lydd soil series also lie above shingle, while the Lydd series itself and parts of the Greatstone series are dominated by sand and loamy sand, which may be derived from sandy beaches associated with the shingle beaches in much the same way as sandy beaches are found today on the eastern shoreline of Dungeness. The ridges often display outlines that indicate clearly patterns of primary depositional morphology (Green and McGregor, 1986). There is a broader tract of buried shingle on the western side of the GCR area, where both aerial photographs and Green's (1968) soil mapping indicate patterns of primary depositional morphology, similar to the exposed shingle. To the north-east of Broomhill, the shingle is replaced by the Midley Sand, but an outlying buried ridge close to the western boundary of the GCR area forms the most westerly identifiable element of the Dungeness shingle system. The buried shingle at Scotney Court and north of Lydd varies in thickness from 2 m to over 5 m increasing in thickness northwards. The shingle increases in thickness towards the south and attains depths in excess of 15 m in the area of the power station. Some boreholes (Green, 1968) reported 'sand with gravel', which continued for another 12 m. A series of mounds in Green's Newer Marsh between Lydd and New Romney have been identified by Vollans (1995) as accumulated remains of the spent sediments cleaned out from filter pits or troughs used in 11th century salt-making. At Belgar, one such ridge extends for almost 2 km in front of the distal end of the Lydd spit.

Dating of the shingle relies on cartographic sources and organic deposits; very few dates have been measured for the areas outside the exposed shingle. Tooley and Switsur (1988) date a marsh infilling of a shingle low at 3410±60 years BP. Peat overlying gravel at Broomhill has been dated at about 3600 years BP and in Scotney Marsh at around about 4000 years BP. Within the shingle-bank complex near Scotney Court (TR 023 202), Callow *et al.* (1966) dated in-situ woody roots lying above shingle at 2740 ± 400 years BP: this occurred beneath silty clay loam overlain by peat. This places the Early Barrier Beach at a date earlier than between 3100 and 2300 years BP.

The development of the foreland has been described (Lewis and Balchin, 1940; Eddison, 1983a,b) by using the trend patterns of the shingle ridges, assuming that each ridge is a former storm beach and so represents a former position of the shoreline. The rate of progradation of Dungeness has been estimated at between 4.1 m a^{-1} and 5.5 m a^{-1}, both Redman (1852) and Hey (1967) estimating the higher value for the period from the early 17th century to the early 19th century. The lower value was estimated for the period 1878 to 1938 by Swallow (Lewis and Balchin, 1940). The morphological patterns of the shingle indicate different modes of deposition associated with different positions in the coastal system at the time of deposition and/or variations in the rate or direction of progradation (Eddison, 1983a,b). Within the area known as 'Denge Beach' (Figure 6.48), the full sequence of ness forms, from their early development to the present-day, is preserved. Few areas anywhere preserve such a complete sequence of beach ridges known to have been formed over at least 2000 years. Many of the ridges can be traced almost without break from this area northwards to their distal ends around Lydd Airport. North of the Dungeness power station, Hey (1967) reported thicknesses of about 10 m, whereas the Soil Survey noted that as much as 17 m depth of shingle had been found in the vicinity of the power station. The clean shingle forms only the upper 1–1.5 m (Hey, 1967), being composed beneath this depth of 'closely-packed pebbles with interstices filled with sand'. Sandy deposits beneath the shingle contained a few scattered pebbles and some marine shells. The sandy gravel is described by Hey as being in beds of between 0.1 m and 1 m in thickness. Each bed was composed of similar mixed material, but each bedding-plane was marked by a distinct change in pebble diameter. The bedding planes had a constant dip of 8°–10° towards SSE, the strike being almost the same as the alignment of the surface ridges. Greensmith and Gutmanis (1990) show, following analysis of 80 boreholes in the vicinity of the power stations, that the 40 m-thick marine Holocene succession can be divided into basal gravels, middle sands and upper gravels that rest directly on a pre-Holocene erosion surface cut across the Lower Cretaceous Hastings Beds between –32 and –35 m OD.

Dix *et al.* (1998) show that high resolution seismic (Chirp) surveys in Rye Bay indicate a dominant seaward-prograding shelf sand body (SSB) with only minor amounts of gravel. The presence of buried gravel beaches at Broomhill dating from the mid-Holocene (Tooley and Switsur, 1988; Long and Innes, 1995b) and studies of drowned Holocene barriers elsewhere (e.g. Forbes and Boyd, 1987; Oldale, 1985; Browne, 1994; Forbes *et al.*, 1995) pointed to the possible preservation of early Holocene barrier structures in Rye Bay. The bedrock surface undulates between –25 m and –35 m OD (Lake and Shepherd-Thorn, 1987; Greensmith and Gutmanis, 1990; Long *et al.*, 1996). NW–SE-trending channels with maximum depths of *c.* –45 m OD may be offshore extensions of the former valleys of the Rother, Tillingham, Brede and Pannel (Dix *et al.*, 1998).

At Dungeness point, Greensmith and Gutmanis (1990) and Basa *et al.* (1997) describe a basal gravel (0.5 to 1.0 m thick) overlain by very fine- to fine-grained, moderately well-sorted sands (20–30 m). Their upper surface is channelled. These 'Middle Sands' are capped by gravel up to 5 m in thickness. At the Open Pits (TR 073 180 and TR 074 185), the ridges are very short and are truncated by a single south–north ridge suggesting that spit extension was more important here than the formation of individual storm beach ridges. There is no other part of the Dungeness beach-complex where distal features occur other than at the landward end of the very long linear ridges (Fuller, 1985; Green and McGregor, 1986).

The area of North Denge Beach broadly occupies the area between the former Southern Railway line from Lydd to New Romney, Lydd Airport and the residential buildings along the coast. It is a fine example of a shingle beach-plain, comprising over 100 sub-parallel shingle ridges, which run northwards to end in very short buried distal features. Some parts of this landform have been excavated for gravel, but a complete set of the ridges straddles the track from the Water Tower (TR 068 202) to Lade (TR 083 208), with only the most recent ridges being obscured by housing and the coastal road. Most of the ridges post-date the mid-8th century shoreline postulated by Lewis and Balchin (1940), the distal features south of Greatstone having marked historically the south side of the gradually silted and reclaimed area known as 'Romney Sands'.

Interpretation

The general interpretative context for this site is described above in the previous section (p. 310 ff).

There are three major issues to be addressed at Dungeness: the description and interpretation of the pattern of shingle ridges, the age of the features and their relationship to the development of the beaches, and the relationship between marsh sediments and the shingle structures.

A summary of the phases of development of Dungeness is presented in Table 6.4.

The earliest discussion of the formation of Dungeness (Elliott, 1847; Gulliver, 1897; and Lewis, 1932, 1937) regarded the ness as having evolved from barrier beaches crossing Rye Bay. These beaches were regarded as having aligned towards the dominant south-west waves and grown by redistribution of sediment from proximal and seaward areas to the recent locations straddling the mouth of the River Rother. Lewis's (1931) conceptual model has provided the basis for the early evolution of the beaches. However, Dix *et al.* (1998) argued that there is little evidence offshore to support Lewis' view.

The western shingle structures described above appear to represent the barrier spit extending towards Lydd and Hythe. The change in growth direction of the ness towards the south-east has not been explained adequately, and some of the western exposed shingle (for example, Jury's Beach, The Forelands and Holmstone) probably represents recurved sections of the early ness form. Green and McGregor (1986) show that these shingle areas are separated by alluvium, which often attains depths of more than 2 m within 10 m of the shingle margin. They consider that these areas of alluvium imply rapid eastwards growth of the ness, which would not have allowed sufficient time for closely spaced recurves to develop. The former southern shore of the ness is first identifiable where it is intersected by the modern southern shoreline about 1 km east of the Galloways Lookout. Northern parts of the sharply curved shingle ridges represent proximal areas of recurves, with, in several places, deep natural pits separating the curved ridges at the point of greatest inflection. This probably indicates shorter periods of rapid eastward growth of the ness (Green and McGregor, 1986), but it may also reflect reduced supplies of shingle from the west and reduced wave energy inputs to the southern shore. Most of the northern extremities of the recurves forming Denge Beach appear to have ended in deep water. The distal parts of the recurves gradually changed alignment towards the north from an early orientation of about 310° to 320° to a modern beach alignment of about 340° to 350° (Figure 6.50). The Holocene sediments in the vicinity of the power stations are consistent with a prograding, upwards-coarsening, barred shoreline, laid down under mixed wave-tidal conditions and predominantly rising sea levels (Greensmith

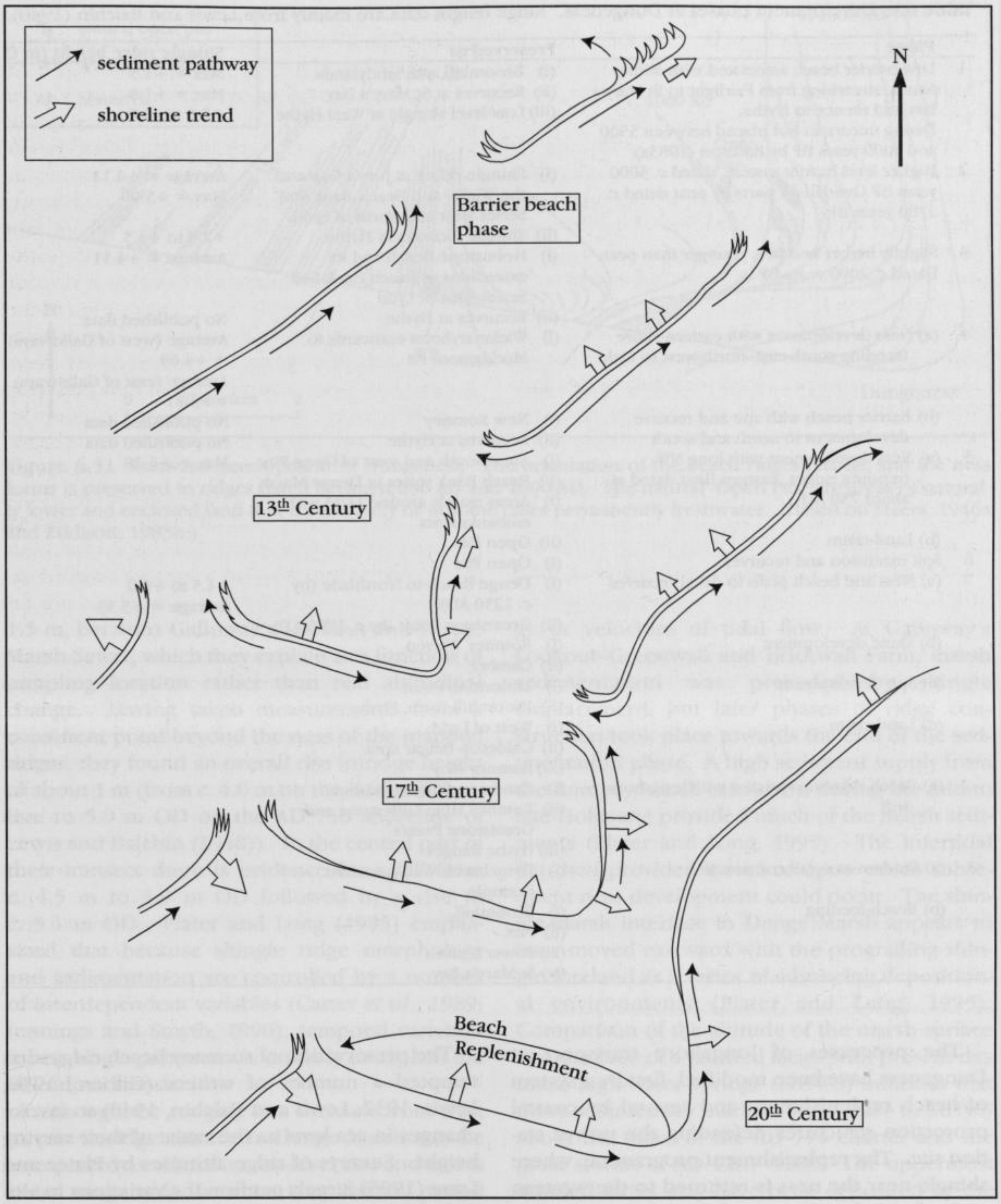

Figure 6.50 Historical sediment pathways and development at Dungeness. Each schematic map shows the probable sediment movements associated with the erosional and accretional trends in the shoreline.

and Gutmanis, 1990). The basal gravels and middle sands were deposited over a period at least 1900 years, a process that began at least 3270 years BP. Radiocarbon dates (1370–3270 years BP) from levels between –32 and –34 m OD are regarded as anomalously young and interpreted as arising from intertidal and high subtidal shells being swept, probably during prolonged stormy period, into depths greater than 20 m.

Table 6.4 Development phases at Dungeness. Ridge height data are mainly from Lewis and Balchin (1940).

	Phase	Preserved as	Shingle ridge height (m OD
1	Low barrier beach associated with Midley Sands, stretching from Fairlight to St Mary's Bay and thence to Hythe.	(i) Broomhill and Sandylands	Max = +1.5
		(ii) Recurves at St Mary's Bay	Max = +1.0
		(iii) Low-level shingle at West Hythe	+0.6 to –1.0
	Dating uncertain but placed between 5500 and 4000 years BP by Eddison (1983a)		
2	Higher level barrier system, dated *c.* 3000 years BP Overlain in parts by peat dated *c.* 2700 years BP	(i) Shingle ridges at Jury's Gap and the Wicks, and Beach Bank Soil Series west and north of Lydd	Average = +4.11 Max = +5.00
		(ii) Shingle recurves at Hythe	+2.8 to +3.5
3	Slightly higher beaches, younger than peat. Dated *c.* 2000 years BP	(i) Holmstone Beach and its extensions as Beach Bank Soil Series west of Lydd	Average = +4.31
		(ii) Recurves at Hythe	No published data
4	(a) Ness development with eastern shore trending south-east–north-west to Lydd	(i) Wickmaryholm eastwards to Muddymore Pit	Average (west of Galloways) = +4.69 Average (east of Galloways) =+3.81
	(b) Barrier beach with spit and recurve development to north and south	(i) New Romney	No published data
		(ii) Recurves at Hythe	No published data
5	(a) Ness development with long NW-trending ridges. Eastern limit dated at about 750 AD.	(i) Areas south and west of Open Pits	Max = +6.28
		(ii) Beach Bank series in Denge Marsh	
		(i) Areas mainly around Lydd within embankments	
	(b) Land-claim	(ii) Open Pits	
6	Spit extension and recurves	(i) Open Pits	
7	(a) Ness and beach plain to distal recurves	(i) Denge Beach to Northlade (by *c.* 1250 AD)	+4.5 to +6.0 Average = +5.33
		(ii) Greatstone Point (by *c.* 1800 AD)	
	(b) Dune development	(i) Romney Warren	
		(ii) Camber	
	(c) Spit development	(i) Littlestone Point	
		(ii) Broomhill Farm, Hythe	
	(d) Land claim	(i) West of Lydd	
		(ii) Caldecot–Belgar area	
		(iii) Romney Hoy	
	(e) Beach ridges associated with longshore drift	(i) Camber and Rye Harbour	
		(ii) Romney Hoy: Littlestone and Greatstone Points	
		(iii) Hythe Ranges	
8	(a) Modern sea-wall construction	(i) Dymchurch Wall is earliest example	
	(b) Beach-feeding	(i) Broomhill,	
		(ii) Pett	
		(iii) Power Station	
		(iv) St Mary's Bay	

The processes of longshore transport at Dungeness have been modified, first by a system of beach replenishment and second by coastal prorection structures defending the power station site. The replenishment programme, where shingle near the ness is returned to the western end of the beach near Brommhill, has been operating since the 1950s (Thorn, 1960) and is one of the longest running schemes anywhere. Figure 6.50 offers an interpretation of the probable sediment pathways both now and at earlier stages of beach development. They warrant further investigation to evaluate the effects of wave climate, storm events, different sea levels, and changes in sediment supply.

The preservation of so many beach ridges has tempted a number of writers (Gilbert, 1930; Lewis, 1932; Lewis and Balchin, 1940) to invoke changes in sea level as the cause of their varying height. Surveys of ridge altitudes by Plater and Long (1995) largely confirm the variations in altitude reported by Lewis and Balchin (1940). Plater and Long do not, however, agree with the earlier interpretation. They recognized that the altitude of the ness could be as much as 1.2 m below both the adjacent west–east (proximal) ridges and the south-east–north-west (distal) ridges. The latter also fell towards their north-western ends. Plater and Long (1995) observed an overall rise in ridge-swale altitude of about

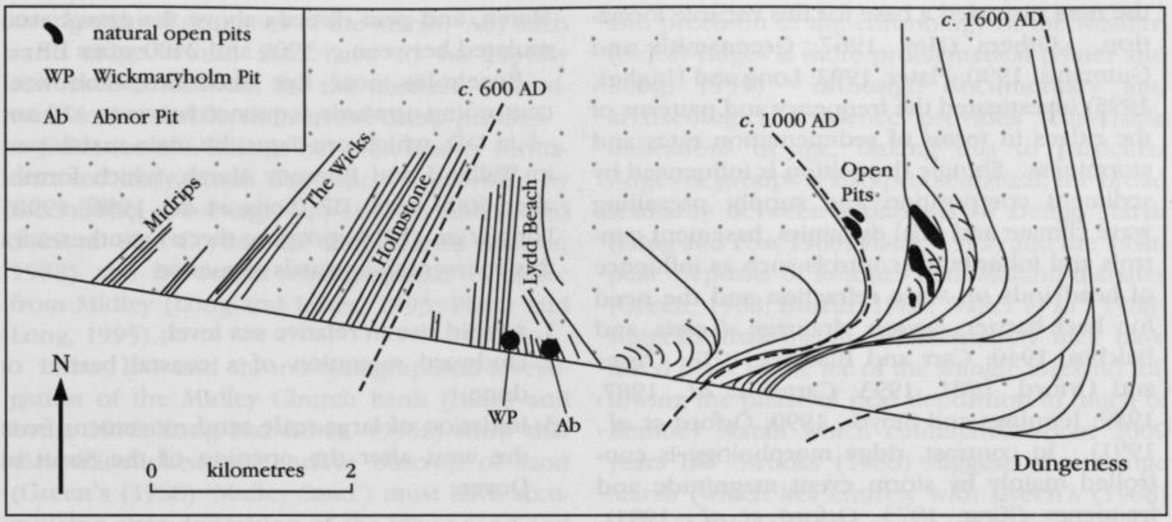

Figure 6.51 Eastward development of Dungeness. The orientation of the beach ridges change and the ness forms is preserved in ridges dated between 600 AD and 1000 AD. The natural 'Open Pits' are areas of naturally lower and enclosed land that is seasonally or in some cases permanently freshwater. (Based on Steers, 1946a and Eddison, 1983b.)

1.5 m, between Galloway's Lookout and Denge Marsh Sewer, which they explain as a function of sampling location rather than real altitudinal change. Having taken measurements from a consistent point beyond the ness of the mapped ridges, they found an overall rise in ridge height of about 1 m (from *c.* 4.0 m on the Roman shoreline to 5.0 m OD on the AD 750 shoreline of Lewis and Balchin (1940)). In the central part of their transect there is evidence for a fall from *c.* 4.5 m to 3.9 m OD followed by a rise to *c.* 5.1 m OD. Plater and Long (1995) emphasized that because shingle ridge morphology and sedimentation are controlled by a number of interdependent variables (Carter *et al.*, 1989; Jennings and Smyth, 1990), temporal variation in any single parameter is unlikely to explain the altitudinal trends.

Although the roughly 1 m increase from west to east in ridge altitude in Denge Marsh may be interpreted as related to sea-level rise and storm event magnitude between the Roman period and the mid-8th century, along-profile morphology accounts for much of the variability in ridge altitude (Plater and Long, 1995). Their stratigraphical, magnetic and diatom evidence indicates relatively uniform and widespread phase of marsh sedimentation. They propose that sedimentation took place on a surface extending from lower marsh to intertidal mudflat. Coarser laminations resulted from increased wave energy or velocities of tidal flow. At Galloway's Lookout–Greenwall and Brickwall Farm, marsh sedimentation was preceded by shingle emplacement, but later phases of ridge construction took place towards the end of the sedimentation phase. A high sediment supply from the Romney Marsh catchment during the mid-to late-Holocene provided much of the marsh sediments (Plater and Long, 1995). The intertidal flat then provided a surface upon which subsequent ness development could occur. The shingle-marsh interface in Denge Marsh appears to have moved eastward with the prograding shingle foreland as a series of advancing depositional environments (Plater and Long, 1995). Comparison of the altitude of the marsh surface and the base of a mottled facies with present-day mean high-water springs (MHWS) indicates that these sedimentary markers were close to MHWS about the times of the AD 774 charter and the great storms of AD 1287–1288. The uppermost mottled facies may have been deposited by the 13th century storms onto 8th century marsh surfaces (Plater and Long, 1995). This, according to them, contradicts the north-easterly younging trends at Brickwall Farm and Denge Marsh.

Thus long-term sea-level rise may have driven progressive tidal sedimentation (Plater and Long, 1995), taking advantage of storm-induced recurve emplacement and consequent back-barrier deposition. The intertidal flat seaward of the ness provided a base for this recurve formation. Others (Hey, 1967; Greensmith and Gutmanis, 1990; Plater, 1992; Long and Hughes, 1995) investigated the frequency and patterns of the ridges in terms of sedimentation rates and storminess. Shingle deposition is influenced by sediment composition and supply, prevailing wave climate and tidal dynamics, basement controls and inheritance controls such as influence of headlands on wave refraction and the need for back-barrier lagoon drainage (Lewis and Balchin, 1940; Carr and Blackley, 1973; Carter and Orford, 1981, 1993; Carter *et al.*, 1987, 1989; Jennings and Smyth, 1990; Orford *et al.*, 1991). In contrast, ridge morphology is controlled mainly by storm event magnitude and frequency (King, 1973; Orford *et al.*, 1991). Ridge orientation is largely affected by sediment budget and transfers alongshore and wave climate (Lewis, 1931, 1933).

Marsh accretion results from incremental deposition of tidal lag sediments (settling at high tide) (Pethick, 1981; Allen, 1990a). Development of the marsh-shingle complex thus depends on processes at different ends of magnitude–frequency scales. Long and Hughes (1995), for example, argue that alternating gravel and marsh sediments result from changes in storm incidence and rates of gravel supply. Plater (1992) suggests that argillaceous and arenaceous sediments above buried shingle ridges in Denge Marsh result from storm breaching of the shingle complex. Most ridges (in their final form) are the result of major storms and so may indicate changing patterns of storminess, but changes in sea level have undoubtedly also been involved. Wass (1995) considers, on the basis of an investigation of sediments and microfauna, that the channel mapped by Green (1968) was a sheltered arm of a tidal inlet in which low-energy conditions prevailed. He concludes that this is inconsistent with the Rother (or any major distributary) crossing the northern part of Romney Marsh since the peat formed there about 3000 years BP. Plater and Long (1995) coupled stratigraphical investigation of Denge Marsh with diatom, mineral magnetic and radionuclide analyses to attempt to establish a chronology of marsh development, the nature of the palaeoenvironments and the primary sediment sources. Spencer *et al.* (1998a) utilized 3400 boreholes and pollen, diatom and radiocarbon dating to interpret the sedimentary record of Walland Marsh. Gravel lies beneath much of Scotney Marsh, and peat directly above the gravel accumulated between *c.* 3900 and 2400 years BP.

Boreholes near Rye show a pronounced coarsening-upwards sequence between –12 and –4 m OD, which pre-dates the main marsh peat in Walland and Romney Marsh, which formed after 6000 years BP (Long *et al.*, 1996, 1998). Long *et al.* (1996) propose three hypotheses for this coarsening-upwards sequence:

1. a rapid rise in relative sea level;
2. landward migration of a coastal barrier or dune;
3. initiation of large-scale sand movement from the west after the opening of the Strait of Dover.

Dix *et al.* (1998) argue that the Rye Bay sand body has many similar characteristics to shelf sand bodies (SSBs) of south-east Australia (Roy *et al.*, 1994). Rye Bay lies in a similar high-energy environment, has a steeper (more than 1°) shoreface and was affected by stable relative sea-level rise and may have had large-scale sand transport: all features of SSBs. Dix *et al.* (1998) argue that there is no evidence from their Chirp survey to suggest landward-migrating barriers in the early Holocene Rye Bay. Rather the evidence points to seaward progradation, with gravel largely absent. Thus Lewis' (1932) former extrapolated positions of the shoreline may not have existed. Dix *et al.* (1998) argue that any barriers probably existed much closer to the present-day shoreline. They identify a need for further work on the processes that allowed SSB progradation during a marine transgression and in-situ examination of the buried intertidal and subtidal stratigraphy of Rye Bay. They consider that an early Holocene complex could have been reworked by relative sea-level rise during the mid- and late Holocene and that this stopped close to the upper depositional surface of the underlying sandy body. A second possibility is that Rye Bay was too deep to allow early Holocene inter- and supertidal gravel deposits to accumulate. Such barriers would only form much closer to shore and rapid SSB progradation occurred with the slowing of relative sea-level rise after about 6000 years BP. A third hypothesis suggests that the early Holocene barrier was sand not gravel and that this is represented by the Midley Sand. Long and Innes (1993) have shown, however, that the Midley Sand is one of the youngest elements of the stratigraphical sequence of the marsh. Any such sand ridge would also have to be rapidly reworked to account for the absence of landward-inclined reflections in the Chirp profiles.

Evidence of a linkage between barrier formation and early marsh deposition is provided by the infilling of a shingle low at Broomhill (dated 3410 ± 60 ^{14}C years BP (Tooley and Switsur, 1988) and by recent stratigraphical evidence from Midley (Long and Innes, 1993; Plater and Long, 1995).

Litho-, bio- and chrono-stratigraphical investigation of the Midley Church bank (Innes and Long, 1992; Long and Innes, 1995a) show that the near-surface and surface outcrop of sand (Green's (1968) 'Midley Sand') must have accumulated after deposition of the lower sand and the younger marsh sediments. Peat began to accumulate beneath the Midley Church bank as marine influence declined from about 3700 years BP until about 2700 years BP, after which there was a gradual return to marine conditions; peat accumulation had ceased by *c.* 2200 years BP. Cereal-type pollen, other herbs and ruderal pollen types within the peat may indicate local Bronze Age farming. Long and Innes (1995a) suggest that the Midley Church bank was either an aeolian deposit or a water-lain sandbank (possibly within a former course of the Rother).

According to Eddison's (1983a,b) model, the first of the high-level shingle ridges were emplaced by about 3000 years BP with appromately 500 ridges over 10 km deposited owing to the co-incidence of storms and high tides. Discrete populations of ridges with similar orientation can be identified within the Dungeness complex. These might be the result of extended periods of high storm-frequency rather than individual events. Greensmith and Gutmanis (1990) also note this phase of shingle deposition on the seaward flanks of the ephemeral Midley Sound barrier complex from *c.* 3400 years BP. This could be linked to a more regional scale control on shingle deposition via changes in wave climate between 5000–300 years BP (Jennings and Smyth, 1990). Eddison (1983b) implies progressive or pulsed development, other authors have proposed much shorter periods of time for upper shoreface and storm beach deposits near the ness (i.e. 750 years – Greensmith and Gutmanis, 1990, and *c.* 350 years – Hey, 1967).

The more recent (eastern) ridges can be dated from cartographic evidence, but the accuracy and precision of the chronology of the western (older) ridges is more problematical (Plater and Long, 1995). Although documentary and archaeological evidence provides reasonable indications of age, linking this to particular ridges or groups is also problematical, the broad similarity between sediments of Denge Marsh (Long and Fox, 1988; Plater, 1992) and the 'post-peat' deposits of Romney and Walland marshes (Green, 1968; Burrin, 1988; Waller *et al.*, 1988) suggests that marsh sedimentation may have taken place in the lee of the shingle foreland following the phase of peat deposition in much of Romney Marsh which culminated about 2000 years BP. Brooks (1988) suggests that Denge Marsh (which lies entirely with Green's (1968) 'New Marshland') was emplaced by Saxon times. However, Bronze Age axes in shingle north of Lydd (Needham, 1988) and evidence of Roman occupation of shingle west of Lydd (Cunliffe, 1988: Green, 1988) imply earlier marsh deposition. An alternative view (Cunliffe, 1980, Lamb in Eddison, 1983b) is that the most recent sediments of Denge Marsh were deposited during a series of storms in the 13th century. This largely confirms the view of Lewis and Balchin (1940) and Steers (1946a) that the marsh at Denge was largely deposited around AD 744 (Table 6.4).

Understanding of the development of the Dungeness foreland over time has a very practical application today. At the Public Inquiry held in 1958 into the proposed siting of a nuclear power station, it was pointedout that the new construction would be on an eroding shore. In spite of this the construction proceeded and, together with subsequent development, now requires to be protected from frontal erosion of the beach by the annual addition of up to 30 000 m^3 of gravel (Summers, 1985). The gravel is sourced from the accreting east side of Dungeness and transported artifically to nourish the south side where the reactors are sited (Figure 6.46).

Conclusions

Dungeness is a large, complex and geomorphologically important site, first because of the shingle ridges, and second for the shingle foreland. Beach ridges such as those found at Dungeness are not confined only to cuspate forelands, shingle ridges with recurved distal ends being found at many scales around the British Isles (for example, Blakeney Point, Orfordness, Hurst Castle

Spit, and Pagham). The complex overlapping – and associated truncation – of sets of ridges that can be dated is extremely well-developed at Dungeness, where it occurs on a large scale over a known timescale.

Shingle structures of such complexity are unusual globally. Dungeness is a cuspate foreland of intermediate size in global terms, but features the size of Dungeness are rare on the coasts of Britain.

Although none of the individual geomorphological features of Dungeness is unique, their association together gives the site its special interest. The considerable damage to much of the original feature (Fuller, 1985) has not obliterated the most important features and every part of the sequence of ridges is still preserved at some point. The as yet little-analysed archival and archaeological evidence provides a potentially rich field for further interpretation of the development of this large and complex feature.