4. Theme 3: The modification of continental shelf archaeology by transgression and regression

The previous chapters have described the issues involved in reconstructing submerged landscapes (Section 2: Theme 1) and examined the potential of submerged archaeological material (Section 3: Theme 2). The intention of this chapter is to assess the impact that the process of transgression and regression during the sea level cycle have on both landscapes and deposits of archaeological material.

Critical to this discussion are concepts of scale. This is highlighted by the fact that continental shelves are exceedingly dynamic areas, with the capacity to change on a range of temporal and spatial scales ranging from millimetres and microseconds to thousands of kilometres and millions of years (Figure 95: Sternberg and Newell, 1999; Schwarzer et al, 2003).

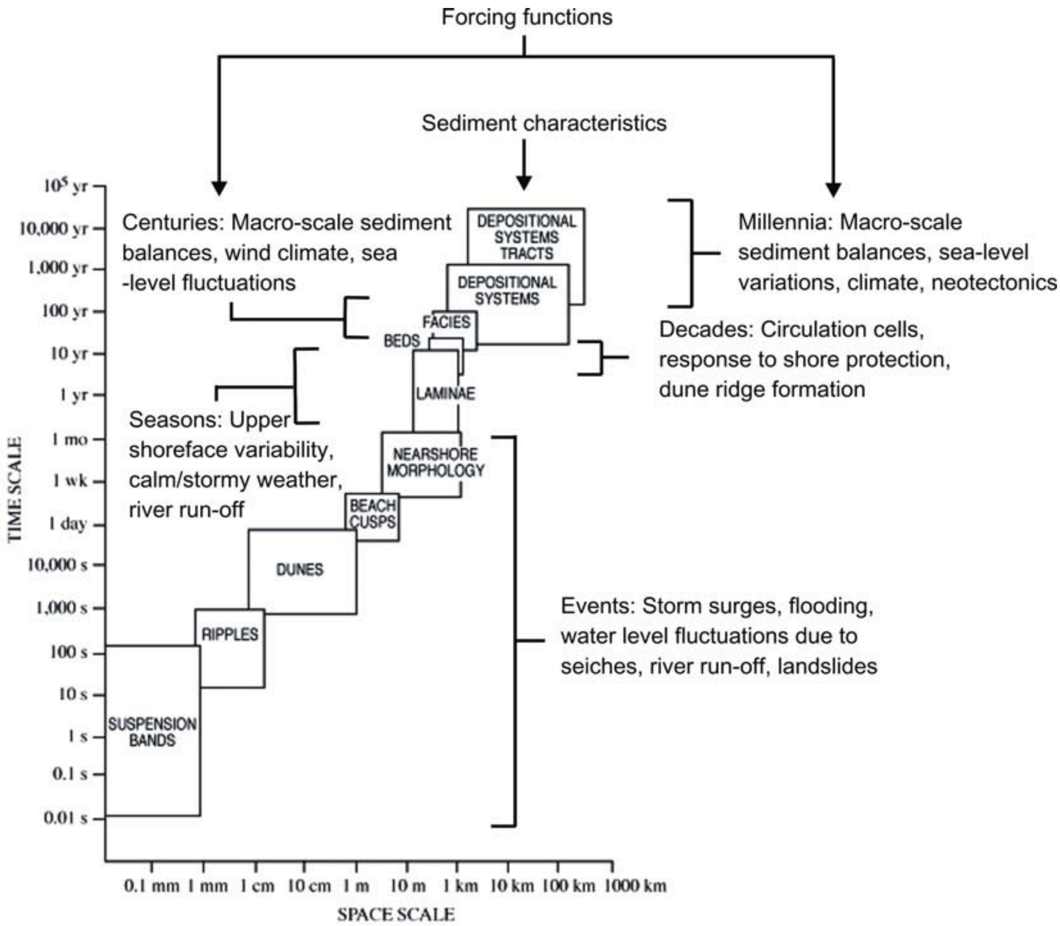

Figure 95. Timescale of forces operating in coastal evolution and the characteristic sedimentary features they create (modified from Schwarzer, 2003; Sternberg & Newell, 1999)

Processes operating at different scales have the potential to affect archaeological material in different ways and also provide information which may be suited to addressing particular research questions (see Section 3: Theme 2 for a discussion of primary versus secondary contexts).

This document is therefore divided into two parts Section 4.1 seeks to understand large scale responses of coastlines and continental shelves to marine transgression and regression while Section 4.2 examines the responses of individual deposits of archaeological material to transgressive and regressive forces. Section 4.1 should therefore assist in addressing some of the issues involved in the large scale reconstruction of submerged landscapes while Section 4.2 has a role in assessing the extent of reworking that individual deposits of archaeological material are likely to be subject to.

4.1 Shoreline and Continental Shelf Responses

4.1.1. Introduction

An examination of the available literature and techniques of palaeogeographic reconstruction have highlighted that most attempts make use of present day seabed bathymetry (Section 2.5). However, questions have been raised as to whether the present-day seabed is an accurate representation of the sub-aerial landscape at sea level lowstands (Section 2.5) or whether it has been significantly modified (buried, eroded or reworked) by cycles of sea level change. Modification of the landscape could occur as a result of processes operating sub-aerially prior to transgression, during transgression (syn-transgressive), during regression (syn-regressive) and after trans- or regression (post-trans- or regressive). This final stage is commonly referred to as a 'stillstand'. It is outside the scope of this project to consider the processes operating sub-aerially when the continental shelf is exposed because of marine regression. The aims of this chapter therefore are to:

- Determine the impact of transgressive and regressive processes on shoreline morphology both during, and after transgression and regression.
- Determine the impact of marine processes on the continental shelf, both during, and after transgression and regression.

An examination of the impact of transgressive processes on shoreline and shelf morphology should demonstrate the extent of modification, thus determining how close an analogue the present day shelf is to the sub-aerial lowstand landscape. The impact of regressive processes will also have to be looked at since sea level has undergone multiple large scale fluctuations over the Quaternary. Hence any archaeological landscapes dating from prior to the Last Glacial Maximum (LGM) lowstand may have been modified by regressive as well as transgressive processes.

The issue of trans- and regressive modification of the landscape has been recognised by a number of researchers (e.g. Garrison, 1991; Coles, 1998; Shennan et al, 2000). However, its effects have rarely been considered when reconstructing the extent of past coastlines. Section 2 (Theme 1) has highlighted the error margins in shoreline position that could result from this. Unfortunately, most reconstructions simply use present day sea floor bathymetry as an analogue to that of the past landsurface (e.g. Lambeck, 1995; Shennan et al, 2000). Again, this approach is perfectly acceptable if all that is required is a broad sense of palaeogeographic space, but we have to question whether it provides a level of accuracy sufficient for more regional and local archaeological research of continental shelves, especially in areas that may have suffered significant erosion or deposition during and after transgression.

The situation is somewhat better when it comes to reconstructing the actual pre-submergence makeup of the past landsurface (i.e. its geomorphology and topography). A number of researchers (e.g. Bridgland, 2002; Catteno & Steel, 2002; Bourillet et al, 2003) have focused intensively on this subject, and consequently there does exist a substantial body of work concerning the extent of large scale subaerially formed features on the shelf. This work is overwhelmingly geological in focus, (e.g. Cameron et al, 1992; Hamblin, 1992), but elements have been adopted by archaeologists in order to present a more realistic picture of emergent shelves at lowstands (e.g. Coles, 1998; Flemming, 2002). From a purely archaeological point of view, this situation has arisen as a result of the general lack of interest in submerged areas, and the fact that existing work has focussed on specific sites, rather than landscapes, where detailed, but very localized, reconstructions can be made (e.g. Geddes et al, 1983) or very broad studies where a rough outline of the position of past coastlines is all that is required (see Section 2.1).

This is not to say that archaeologists are unaware of, or are lagging behind the geological community. In fact, it is only fairly recently that very detailed geological reconstructions of past shelves and changes induced by transgressions and regression have come about (e.g. Bridgland, 2002; Bourillet et al, 2003; Reynaud, 2003; Van der Molen et al, 2004). Consequently, geological evidence from shelves is still incomplete and gaps do exist in the understanding of their past nature. This is illustrated by the initiation of two major International Geological Correlation Program (IGCP) Projects; Project 396 – *The record of continental shelves during the Quaternary, their interpretation, correlation and applications* - which ran from 1996 to 2000 (Yim, 2000), and its follow up; Project 494 – *Continental shelves during the Last Glacial Cycle: knowledge and applications* (2001-2005: Yim et al, 2002). Among the aims of these projects were the investigation of palaeoenvironmental changes on continental shelves and the impact of sea level change (Yim et al, 2002). Clearly, there is still a great deal of work to be done which will aid future palaeo-landscape reconstruction and which in turn will facilitate archaeological research, and furthermore, this also illustrates the value that the continental shelves have to other disciplines.

In addition to these recent studies of the impact of past sea level change on the shelf, a number of models have recently been developed to assess coastal evolution in the face of present day sea level rise (e.g. Carter & Woodroffe, 1994b; Pethick, 2001). Although present day processes or environments may not be exact analogues to those of the past, these models may provide indications of the sorts of coastal changes that could have potentially taken place in the past but which are not preserved in the stratigraphic record.

From the archaeological perspective, its is worth reviewing why the misconception that marine transgressions simply flood a landscape, leaving it broadly unchanged, thus allowing modern continental shelf topography to be used as a representation of the past landsurface, may have arisen (Stride, 1982). One reason is the terminology that surrounds continental shelves. Shelves are frequently described as 'relict'; however this should not be taken to mean that they are fossilized or unchanged landscapes. Use of the term 'relict' was first applied by Emery (1968) on the basis that shelf surface sediments appeared to have been deposited prior to the mid-Holocene highstand in sea level, and that they thus represented deposits created at times of lower sea level (Stride, 1982). 'Relict' in this case is a chronological definition and does not preclude the possibility that these sediments were reworked, both by the transgressive event itself and post-transgressive processes, thus altering the geomorphology and topography of the original terrestrial landscape (Emery, 1968; Stride 1982).

"Relict sediments can be identified by and defined by their anomalous composition, grain size, and grain surfaces, even though they may have been moved about in response to their new environments" (Emery, 1968:446)

In all fairness, structural features may be defined as 'relict' in the sense that they have been inherited from an earlier time period, and while examples of these may exist on continental shelves; it is inaccurate to think of shelves entirely in this fashion.

The difficulty lies in the fact that the difference between the two definitions is rarely explicitly expressed in the literature and consequently care should be taken. In reality, studies have indicated that deposits on continental shelves do in fact respond to modern hydrological processes, and thus the notion of the continental shelves as an untouched landscape is rendered invalid (Stride, 1982; Swift & Thorne, 1991; Nittrouer & Wright, 1994).

"Modern shelves are, to a greater or lesser extent, relict in the sense that pre-Holocene sediment is exposed and is being current-reworked" (Leeder, 1999:444)

Admittedly, there are cases in which the present-day seabed has provided a reasonably accurate analogue to the past landscape. The Danish 'fishing site' predictive model (see Section 4) was based on correlating present-day bathymetric contours with the most suitable topographic locations for fishing with standing gear and proved remarkably effective in locating archaeological material. Even so, there were occasions where the relief of the original prehistoric landscape was obscured by more recent sediments, thus reducing the effectiveness of the predictive approach based on modern bathymetry (Fischer, 1995b; Fischer & Pedersen, 1997). Furthermore, bathymetric models may be useful in sediment starved areas, such as the limestone regions of North-West Florida, where a paucity of clastic erosional material and resulting low sedimentation rate means that terrestrial features (e.g. sinkholes and river channels) can still be visually identified underwater (Figure 96: Dunbar et al, 1991).

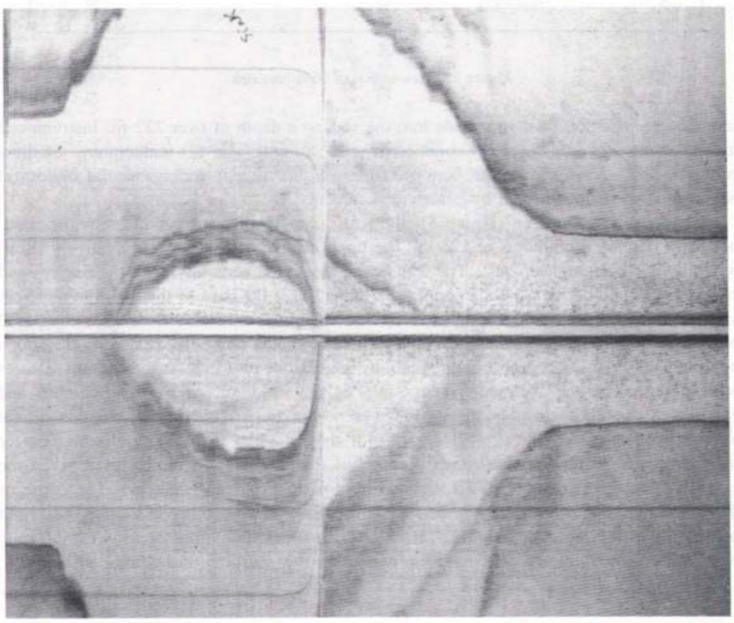

Figure 96. Sidescan sonar image of a karstic sinkhole in 175m water depth off North-West Florida (from Garrison, 1991).

Recent advances in sonar technology, and in particular swath bathymetry, means that very high resolution imagery of shelf topography is now available for palaeo-landscape and archaeological interpretation and is beginning to be used widely (e.g. Fedje & Josenhans, 2000; Mandryk et al, 2001 - Figure 97). However, the archaeological usefulness of such data is reliant upon a well preserved yet exposed pre-submergent land surface being present. It is therefore essential to identify if the sheltered archipelago environments, such as the south Scandinavian situation, or a sediment starved karstic area, as in north-west Florida, are typical of shelf conditions.

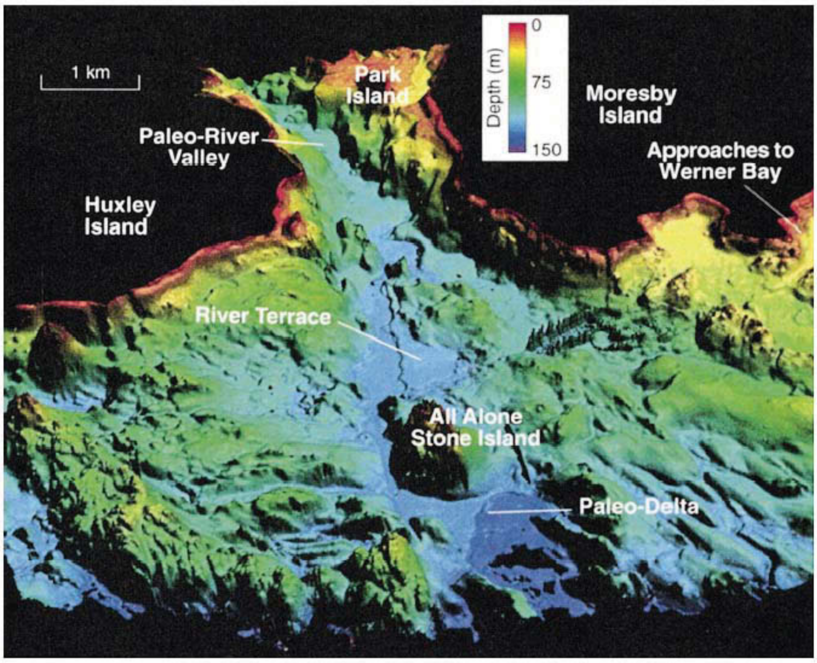

Figure 97. Multibeam bathymetry map of the southern Juan Perez Strait (British Columbia, Canada) processed into a digital terrain model. The data has a spatial accuracy of +/- 1m. Note the identification of geomorphic features created at sea level lowstands (e.g. palaeo-delta) (from Mandryk et al, 2001)

Many parts of the continental shelf are characterised by significant accumulation of syn- and post-transgressive sediments. This can be seen in some sections of the submerged English Channel river systems (see Section 3.4.3). For example Figure 98 represents a sub-bottom boomer profile from the outer section of the Arun River system, showing in excess of 17 m of unconsolidated Holocene sedimentary fill. This results in a significant smoothing of the seabed contours in this locality (Figure 99) by comparison to the obviously incised palaeo-channel described by the bedrock surface (Figure 99). The full history of this sequence is currently being explored as part of PD3363 in collaboration of the more extensive surface and sub-surface studies of upstream parts of the submerged Arun river system (PD3277) and Wessex Archaeology.

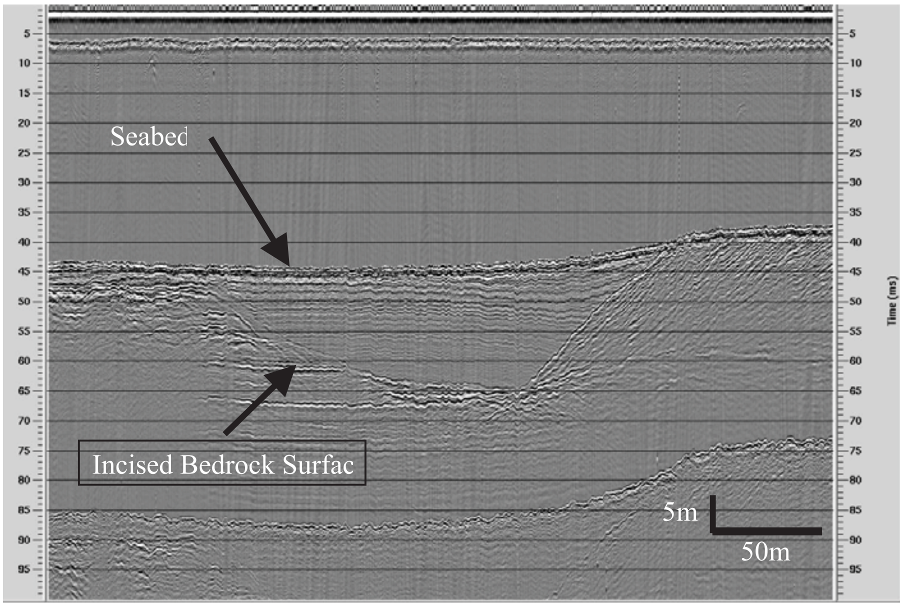

Figure 98. Seismic section of buried palaeo-channel associated with outer sections of the submerged Arun palaeo-valley showing seabed and incised bedrock horizons. Data courtesy of Wessex Archaeology.

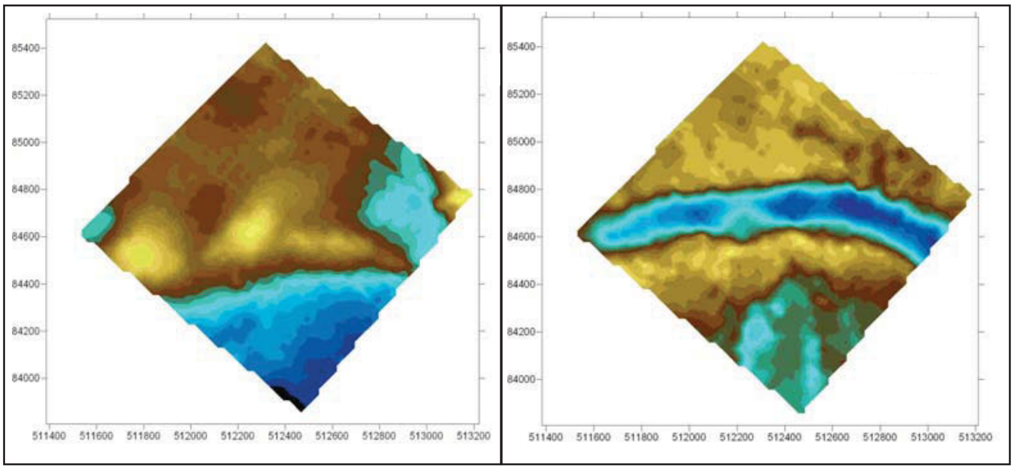

Figure 99. Contour maps based on the interpretation of seismic data from the outer section of submerged Arun palaeo-valley. The clear channel identified in the bedrock contour map on the right is not identifiable in the seabed bathymetry shown on the left. Each square represents an identical 1 km block.

4.1.2 Geomorphological Responses to Sea Level Change on Shores and Shelves

Studies of continental shelves have indicated that over geological timescales (i.e millions of years) their physical environment assumes the form of a predictable 'regime' (Swift & Thorne, 1991; Thorne & Swift, 1991). This is illustrated by the characteristic shelf-slope configuration developed by all shelves (Figure 100). This consists of relatively steep slopes close to the shoreline. They then slope gently (average gradient of 1 in 500) down to the shelf break, whereupon they drop rapidly (average gradient of 1 in 20) down the depths of the ocean basin (Pickard & Emery, 1990; Swift & Thorne, 1991). This profile is the result of a dynamic equilibrium created by the complex interactions of five variables on sub-geological timescales.

These variables are:

- Rate of sediment input from landward sources
- Type of sediment input
- Rate and direction of relative sea level change
- Rate of dispersive sediment transport
- Variations in the fluid power of shelf currents

These variables vary the rate and magnitude of sediment deposition on the shelf through sediment supply and the fluid power available to remove it. As the system is in dynamic equilibrium, any change in one variable, results in a change in the sedimentary regime, such that the change is compensated for by an adjustment of one of the other variables. Consequently, sediment accumulates up to the point at which wave energy is sufficient to mobilize it (wave base). As wave base is approached, the ratio of sediment deposited to sediment bypassed decreases. Given that vertical aggradation is no longer possible, the locus of deposition shifts seaward to the forward face of the sediment pile to form the continental slope – a steeply sloping surface dominated by gravity driven processes, thus forming the characteristic equilibrium shelf profile (Swift & Thorne, 1991). The overall seaward deepening profile is a consequence of the fact that sediment input is higher closer to the shore, as a result of fluvial discharge and coastal erosion, and that wave energy decreases as the waves propagate landward and hence wave base operates at a lesser depth than further seaward on the shelf. Continual subsidence resulting from the accumulated weight of sediment allows the continual aggradation of sediment up to wave base (Swift & Thorne, 1991).

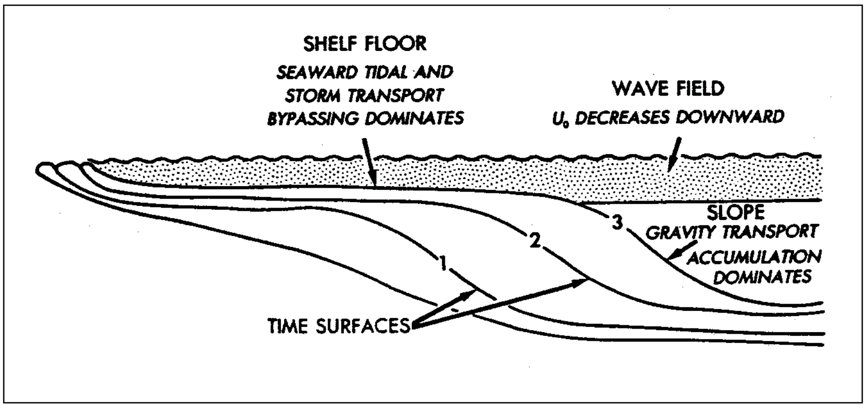

Figure 100. Schematic illustration of the long term evolution of a shelf and the components of the classic shelf profile (from Swift & Throne, 1991).

From the perspective of archaeological research, or indeed palaeo-landscape reconstruction, the temporal and spatial scales of interest are somewhat smaller than those of large-scale shelf studies, at most usually involving thousands to tens of thousands of years and tens to hundreds of kilometres. Thus, at these scales, it is the changes amongst the above five variables that are important rather than the very long term equilibrium of the shelf surface.

Generic models of stratigraphic system variability in response to the sea level cycle (and hence on a temporal and spatial scale relevant to archaeology) identifies a series of lowstand, highstand, transgressive system tracts (e.g. Woodroffe, 2003 - Figure 101).

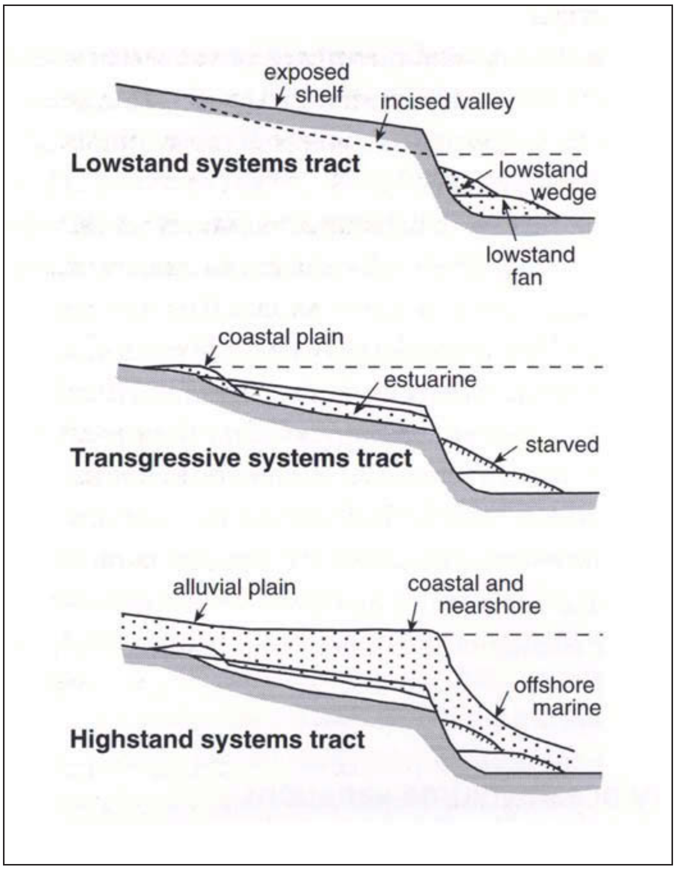

Figure 101. Illustration of the development of different depositional systems, and hence different coastal geomorphology under changing conditions of sea level (after Woodroffe, 2003).

In order to understand the process by which these system tracts develop, it is necessary to investigate the actual role of sea level change, and equally importantly, sediment supply in determining the nature of the shelf deposition regime. This interplay results in the formation of one of two basic regimes (Figure 104):

- *Supply dominated regimes (regressive depositional systems)*

Under these circumstances, the rate of sediment input is greater than the rate of sea level rise and the ability of marine processes to remove it. In essence the supply of sediment exceeds the quantity of available accommodation space. This can be created either by an excess of sediment input, or a relative sea level fall which creates a decrease in local accommodation space. In the former case, progradational regression may take place, while in the latter case forced regression takes place (see Section 2.2.5 for detail). The deposits laid down by these regimes are thick, fine grained and homogenous, and are known as regressive depositional systems.

- *Accommodation dominated regimes (transgressive depositional systems)*

In contrast, in these regimes the rate of sea level rise and effect of marine processes exceeds the rate of sediment input. Thus accommodation exceeds supply and either forced or erosional transgression may take place depending on whether the dominant factor is sea level rise or sediment supply (see Section 2.2.5 for detail). The resulting transgressive depositional systems are thin, coarse grained and heterogeneous.

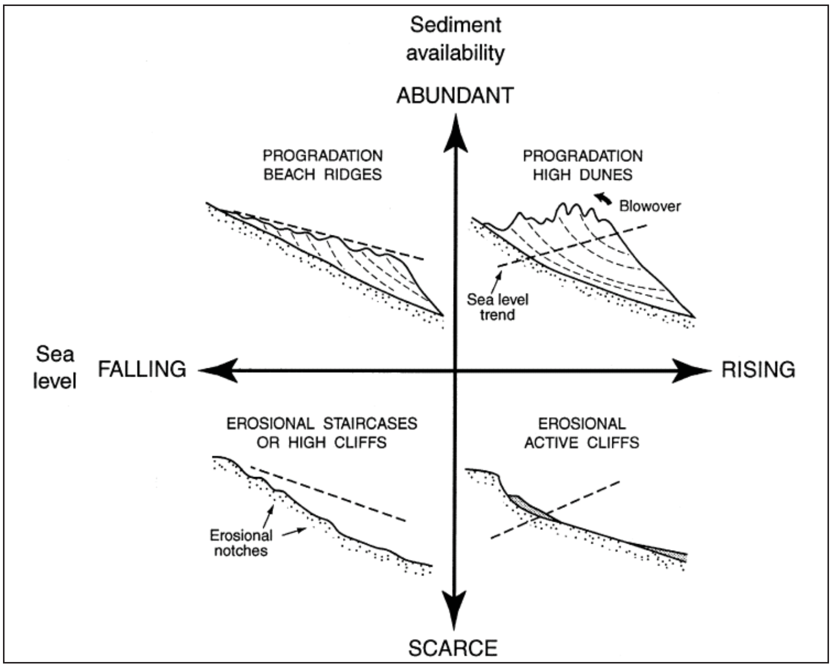

Figure 102. Response of coastlines to the interaction between sediment supply and relative sea level. Note that progradation can occur under conditions of rising sea level provided there is an abundant sediment supply, while transgression (represented here by erosion) is possible under falling sea levels if the sediment supply is scarce. The top two cases represent supply dominated regimes, while the bottom two depict accommodation dominated regimes (from Hansom, 2001).

Clearly the dynamics of sea level change in conjunction with sediment availability and accommodation space plays an important role in determining the nature of coastal and shelf sedimentation, and hence the geomorphology of shelf and coastal sedimentary bodies within the overall shelf equilibrium profile (Swift & Thorne, 1991; Thorne & Swift, 1991; Hoselmann & Streif, 2004). This has been illustrated by recent modelling experiments which investigated the impact of basin-scale bathymetric changes in response to differing rates of sea level change (Van der Molen et al, 2004). Van der Molen et al's computations focus on the basin-scale evolution of the seabed (hundreds of kilometres), span several tens of thousands of years, and represent various scenarios of sea-level change. Morphological features with length scales of tidal sandbanks and smaller were not included. Their results show that the basin will export sediment and deepen. Also, it will expand by the accumulation of eroded sediment in deeper waters. The deepening causes reduction of the flow velocities and the net sediment transport, resulting in decreasing rates of morphodynamic evolution. The feedback of the developing bed levels to the water motion is dominated by the increase in water depth, and much less by the seabed topography. Externally prescribed changes in sea level change the wavelengths of the tide and the seabed pattern and thus also change the speed of the morphodynamical evolution.

The effect of two additional factors, namely geological inheritance and climatic or oceanographic factors, also have to be considered in shelf and shoreline evolution (Carter & Woodroffe, 1994; Forbes & Syvitski, 1994; Hansom, 2001; Catteneo & Steel, 2002). Geological inheritance refers to the geological and topographic makeup of the landscape in question that has been created by various land-forming processes operating over the course of geological time (Roy et al, 1994). This includes factors such as gradient and lithology (Roy et al, 1994; Hansom, 2001). Climatic and oceanographic conditions refer to factors such as the local wave and tidal regime. The importance of these is that they control the local sediment transport pattern.

These processes (sea level change, sediment supply, geological inheritance and oceanographic conditions) operate over a series of timescales ranging from events (e.g. storm surges and landslides) to millennia (e.g. glacio-eustatic sea level change) (Schwarzer et al, 2003 - see Figure 95 for summary). The balance between these influences is such that changes in each of them have the potential to create changes in coastal morphology and shelf facies geometry (Swift & Thorne, 1991; Swift et al, 1991; Forbes & Syvitski, 1994). For instance, basin widening and increased fetch resulting from a sea level rise could potentially alter oceanographic factors, such as tidal current strength. Further modifications to current patterns could result from the changing coastal configurations and bathymetric shifts creating differing interactions between waves, sea floor and coastline. In addition, changes in sea level may also result in differential access to sediment sources and rates of supply. The aforementioned changes in current strength may in turn modify the rate of erosion (Forbes & Syvitski, 1994; Catteneo & Steel, 2002).

In summary, the dynamics of shoreline change and resulting stratigraphy of the transgressive and regressive deposits, and hence the makeup of the present continental shelf surface, is strongly dependent on the rate and sense of sea level change relative to sediment supply, basin physiography and energy distribution (Swift & Thorne, 1991; Catteneo & Steel, 2002).

The following sections will examine how coasts and shelves respond to transgression and regression, and how this may be reflected geomorphologically and stratigraphically. Although coastlines are an integral part of the continental shelf they will be discussed separately in this document. This results from the fact that marine processes and oceanographic conditions in coastal waters differ markedly from those operating further out to sea. These differences result from the relative shallowness of the water in coastal areas, the presence of the shoreline as a boundary to flow and the effects of fluvial input in coastal waters (Pickard & Emery, 1990). Further, the coastal environment will inevitably occur at all locations on the shelf at different times during the transgressive-regressive cycle and so can play a major role in the development of the whole shelf. Finally, as described in Section 1 the coastal zone may represent a key environment for Hominid evolution and hence require careful study.

For the purposes of this document, the shoreline, or coastline will be considered the point from the high water mark to the lower end of the shoreface, also known as the nearshore area (20 to 30m water depth: Leeder, 1999; Harris & Wiberg 2002), and the continental shelf the area from this point to the shelf break (Figure 103). This division has been chosen on the basis that below the depth of the shoreface, forces induced by wave action under ambient conditions cease to dominate sedimentary processes. It should be noted that the coast includes more than just beaches and cliffs, but also geomorphological features such as deltas, estuaries and lagoons.

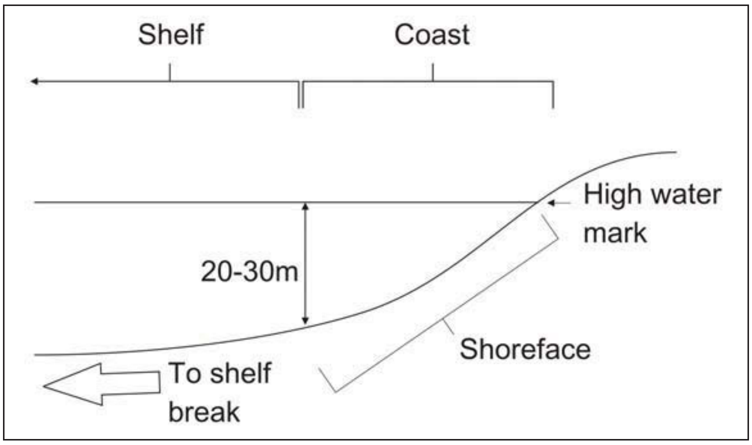

Figure 103. Division of coast and shelf adopted in this document

4.1.3 Response of Fluvial Systems to Sea Level Change

Prior to describing the effects of sea level change on coasts and shelves it is important to recognise that the effects of sea level change may reach an appreciable distance inland of the coastal margin. This effect can be particularly seen within fluvial systems which are in turn key in terms of both their importance in the evolution of terrestrial landscapes (Blum & Törnqvist, 2000), and from a strictly archaeological perspective, the advantages they afford for utilization and settlement by past humans (see Section 3: Theme 2). Blum & Törnqvist (2000) suggest that the landward limit of influence (defined as the intersection between the modern floodplain and the floodplain surface from the Last Glacial Maximum (c. 20 ka) lowstand – Figure 104) can vary from at least 300 – 400 km for low gradient, high sediment supply river systems to c. 40 km for steep gradient low sediment supply systems.

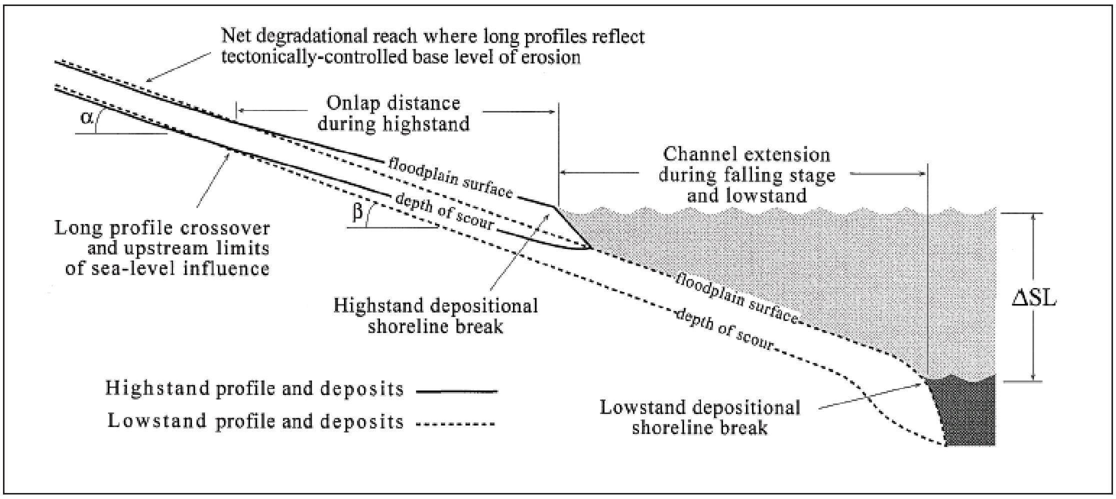

Figure 104. Schematic diagram showing the long profile response of a river system to sea level fall (from Blum & Törnqvist, 2000).

The major control of sea level change on fluvial system evolution is through its direct control on the altitude of a river systems base level; the imaginary horizontal surface to which subaerial erosion proceeds (Schumm, 1993). Riverine systems respond rapidly to changing base levels so the extent of fluvial material on the continental shelf is frequently related to the fluctuation in relative sea level. During lowstands fluvial systems (frequently enhanced by the high discharge rates associated with proglacial rivers) can extend all the way to the shelf edge (Bridgeland, 2000; 2002). This may result in the erosion of extensive drainage networks into the pre-existing shelf sediments or basement material and the deposition of associated sub-aerial facies (Gibbard & Latridou, 2003 and Section 3.4.3).

In addition, the drop in base level caused by sea level fall instigates deeper incision as the river seeks to return to its equilibrium profile (Figure 104). However, it should be noted that incision can be retarded to an extent if rates of sediment supply are high (Blum & Törnqvist, 2000). This resulting increase in gradient has also been postulated as the mechanism behind fluvial systems switching from meandering to braided patterns. In reality, studies suggest that other factors, notably increased sediment loads, higher discharge and channel bank instability may have more important, and consequently channel gradient changes cannot be seen as the sole driver behind this switch (Leigh et al, 2004).

As described by Blum & Törnqvist (2000) sea-level rise results in channel shortening, decreases in the distance over which sediments can be stored and, in most cases, flattening of the channel slope. Discharge is conserved, but reductions in slope will result in corresponding downstream decreases in stream power and sediment transport rates. These conditions result in net valley aggradation which in turn can affect channel geometry. However, even within long periods of overall net valley aggradation driven by sea-level rise (time-scales of 10^3 – 10^4 years) significant intervals of incision and/or sediment bypass can occur if sediment supply decreases relative to stream power.

In general, a degree of debate does exist over the importance of base level changes in fluvial systems. This arises from the fact that the major base level changes occurring over the Quaternary were glacio-eustatic in origin, and consequently, their effects are difficult to distinguish from those of other factors, for instance, changes in discharge and sediment load that occurred during glacial phases alongside the glacio-eustatic fall (Miall, 1996). Space does not permit an in-depth discussion of this subject, as it would require an entirely separate review, such as has been undertaken by Miall (1996) and Blum & Törnqvist (2000).

4.1.4 Coastal Responses to Sea Level Change

In this section the impact of sea level change on coasts will be presented with respect to particular coastal features and landforms.

4.1.4.1 Deltas

Deltas are seaward protruding constructional coastal landforms that represent the accumulation of fluvial sediment in river mouths (Swift et al, 1991; Dalrymple et al, 1992; Davis Jr. & Fitzgerald, 2004 – Figure 105).

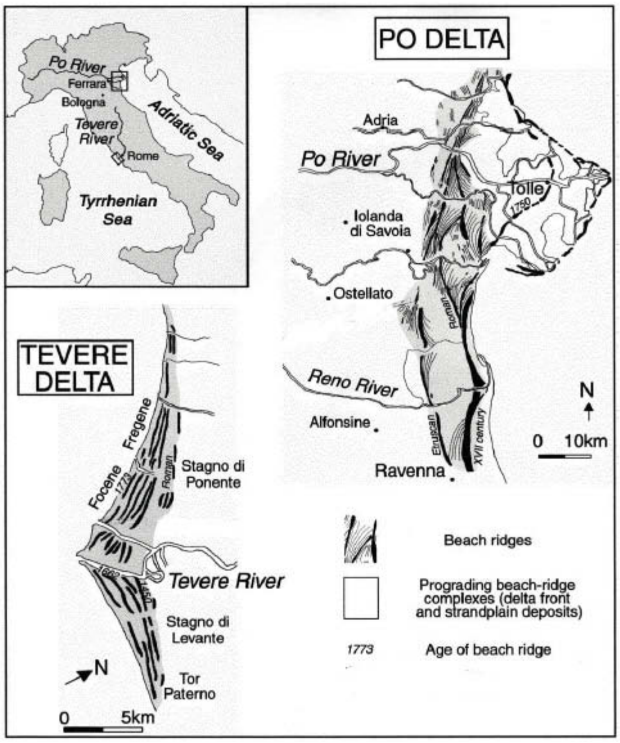

Figure 105. Plan view of two Italian delta systems, the Tevere and the Po (from Amorisi & Milli, 2001)

They form under supply dominated regimes, in that the accumulation of fluvially supplied sediment is greater than the ability of marine processes to rework and remove it, thus resulting in the progradation of shorelines (Swift et al, 1991; Suter, 1994). Hence they can be considered regressive depositional systems. The morphology of individual deltas is to a large extent controlled by the local oceanographic conditions as well as local relative sea level change, sediment supply and the fluvial regime (Figure 106).

The impact of waves on deltas systems results in the redistribution of sediment carried by the effluent jets issuing from the delta mouth. In areas of high wave power relative to river discharge, deltas are much more linear in plan form and spread parallel to the coast. Longshore transport of sand from the delta mouth may also result in the creation of barriers away from the delta mouth. Deltas in macrotidal (>4m) environments however, are shaped by the bi-directional passage of the tidal current and consequently are characterised by funnel shaped mouths and linear (i.e. shore-normal) tidal shoals and extensive tidal flats (Leeder, 1999; Davis Jr. & Fitzgerald, 2004).

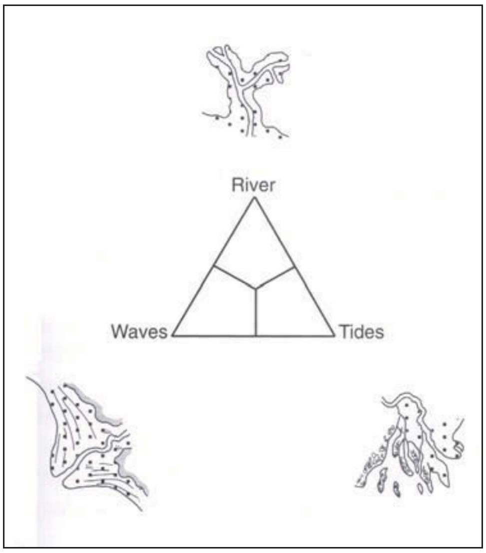

Figure 106. Generalized view of the impact of wave, tidal and fluvial processes on delta morphology (from Woodroffe, 2003).

As sea levels change, so too do deltas. In fact glacio-eustatic fluctuations are recognised as the major control on sedimentation and hence the main influence on the stratigraphic evolution of deltaic systems (Amorosi & Milli, 2001). Each particular phase of the sea level cycle is reflected in delta morphology. Falling sea levels result in rapid progradation across the shelf as accommodation space decreases. Conversely, a reduction in accommodation space results in the deposits being relatively thin and stacked progradationally. Furthermore, as shelf width decreases, so too does across-shelf wave attenuation resulting in increasing wave domination on delta morphology. Meanwhile the extension of rivers across the shelf cuts across pre-existing deltaic sediments.

At lowstands, deltas are located at or near the shelf edge. At this point in space and time, increasing accommodation space caused by the decreased rate of sea level fall, increasing subsidence and increasing sea floor gradient lead to a different depostional style. Lowstands systems tracts therefore tend to be thick and aggradationally (vertically) stacked. Downslope movements induced by turbidity currents are also common. When sea levels begin to rise again, deltas retrograde into incised valleys. They do however, prograde seaward when sea level rise slows or during stillstands. Consequently, transgressive deltas deposits are thin (<10m) and backstepping (retrogradationally stacked). In addition, the creation of embayments results in amplification of the tidal wave and hence increasing tidal domination. At highstands, deltas once again prograde seaward. This is the current global situation (Suter, 1994).

During the majority of the post-LGM period, the rate of sea level change was too great for deltas to form. Essentially, the rise in sea level created accommodation space that could not be filled by the contemporary sediment supply and thus estuaries formed where rivers met the sea. It was only when the rate of rise slowed down in the mid-Holocene that active accumulation could take place (Stanley & Warne, 1994; Davis Jr. & Fitzgerald, 2004).

4.1.4.2 Estuaries

Estuaries represent the seaward portion of drowned valley systems that are influenced by wave, tidal and fluvial processes (Dalrymple et al, 1992 – Figure 107).

They form under accommodation dominated regimes in that sediment delivered by fluvial or marine processes is insufficient to fill the accommodation space created by rising sea level, or in the case of stillstands, to compensate for removal by marine processes (Swift & Thorne, 1991). Consequently, another characteristic feature of estuaries is that they receive sediment from both fluvial and marine sources (Dalrymple et al, 1992).

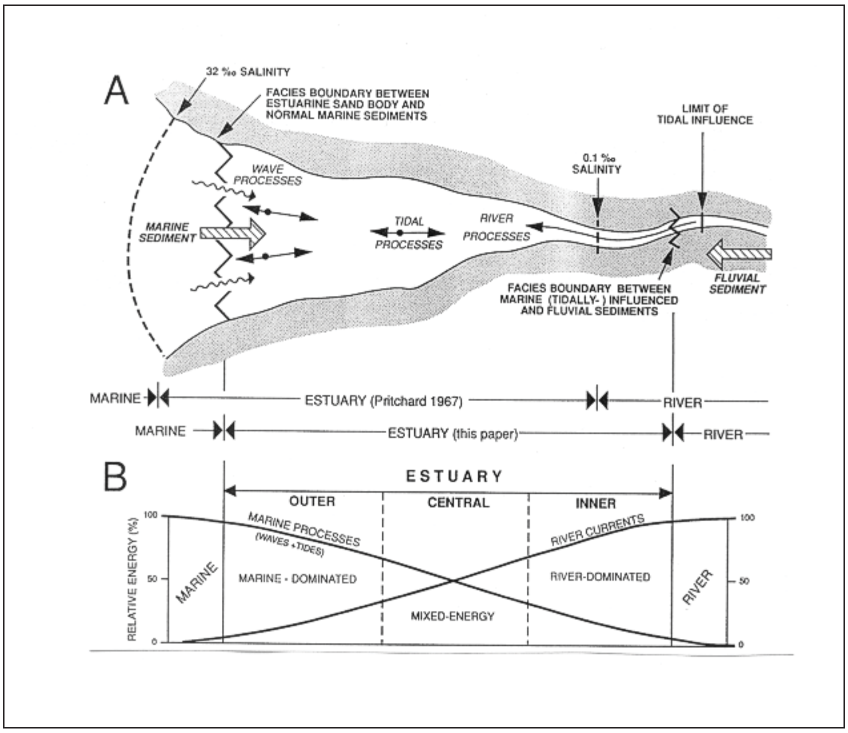

Figure 107. A) Schematic plan view of an estuary showing the generalized transport patterns. B) Schematic diagram showing the interactions between marine and fluvial processes in an estuary (from Dalrymple et al, 1992).

Estuaries tend to be classified as either tide dominated or wave dominated. The former occur along coasts with high tidal ranges and large tidal prisms while the latter are found in areas of reduced tidal range and on barrier coasts (Davis Jr. & Fitzgerald, 2004). Wave-dominated or microtidal estuaries tend to be dendritic in form while tide dominated estuaries are funnel shaped as a result of the penetration of tidal energy further up the estuary than wave energy (Swift et al, 1991; Dalrymple et al, 1992). Tide domination also results in the formation of elongate bars along the estuary mouth while wave domination leads to the formation of bars or barriers across the mouth (Dalrymple et al, 1992 – Figure 108).

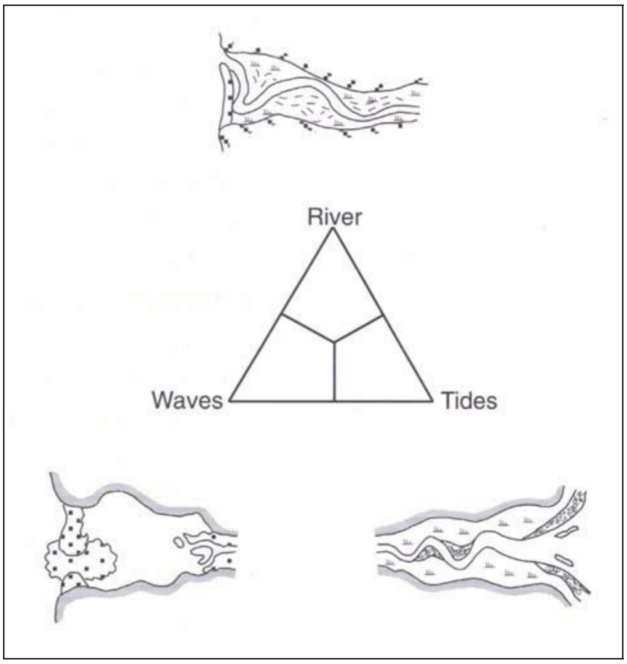

Figure 108. Generalized view of the impact of wave, tidal and fluvial processes on estuary morphology (from Woodroffe, 2003).

These 'end member cases' however, are usually modified by local factors like relative sea level rise, geological history and local oceanographic conditions such that different estuaries exhibit varying degrees of morphological deviation from the models (Dalrymple et al, 1992).

The estuarine response to constant sea level rise tends to take the form of an upward and landward 'stratigraphic roll-over' translation (Long et al, 2000; Pethick, 2001). Intertidal sediments are eroded from the outer reaches of the estuary by marine processes as wave propagation increases due to increasing water depth. This also results in the retreat of the saltmarsh/mudflat boundary (landward movement). These are then moved landward to the inner estuary and redeposited in the intertidal zone thus elevating the mudflat and marsh surfaces (upward movement). The end result is that the estuary channel and its associated landforms migrate landward as a unit with relatively little morphological change provided the sea level rise remains constant (Dalrymple et al, 1992; Long et al, 2000; Pethick, 2001). Models developed in response to present day sea level rise (6mm/yr) suggest that the rate of estuarine migration is 8 metres a year (Pethick, 2001). However, if the rate of sea level change is modified, then changes in estuary response can take place. For example, estuaries in southern England exhibited a pattern of upward and seaward (rather than landward) migration of intertidal and sub-tidal environments during the Mid-Holocene in response to a slowing in the rate of relative sea rise (Long et al, 2000).

Further changes may take place if external variables such as sediment supply, river inflow, tidal range and wave climate are modified by, or along with, the sea level change. In these instances other morphologic changes would occur in conjunction with the rollover response, such as, a local increase in tidal range leading to an estuary's morphology changing to a more tide dominated configuration (Dalrymple et al, 1992; Chappell & Woodroffe, 1994). For example, a switch from tide to wave domination during the mid to late Holocene transgression resulted in the development of coarse grained barriers and spits in the mouth of the Delaware Bay estuary (Fletcher et al, 1990).

Finally, under regressive conditions, that is if either sea level rise reverses or slows to a point at which sediment supply exceeds the rate of creation of accommodation, estuaries will infill and form deltas, if sediment is supplied fluvially, or straight prograding coasts, if sediment is supplied by marine processes (Dalrymple et al, 1992 – Figure 109).

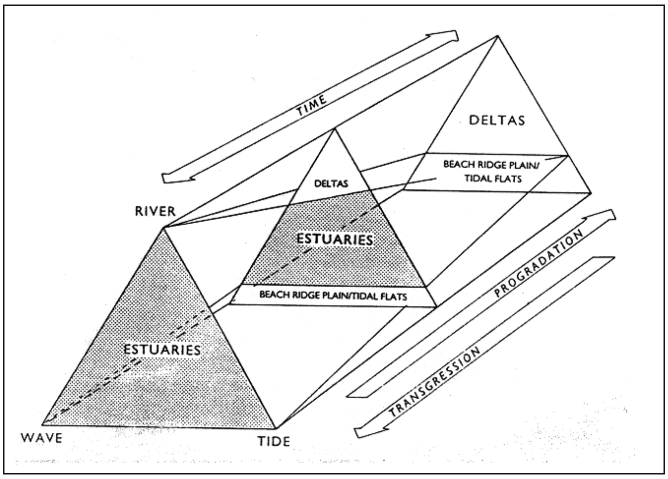

Figure 109. Diagram indicating the transition from estuary to delta or prograding coast depending on local wave and tide climates and the rate of transgression or regression (from Dalrymple et al, 1992)

4.1.4.3 Saltmarshes

Saltmarshes are defined as being the part of the high intertidal zone dominated by halophytic vegetation, such as the *Spartina* species of grasses that are regularly flooded by the sea (Allen, 1990; 2000; Davis Jr. & Fitzgerald, 2004). They are common features on low energy open coasts or in the inner protected areas of estuaries (Figure 110). The actual physical environment of a marsh depends on a number of local factors, notably local tidal range and relative sea level change (Davis Jr. & Fitzgerald, 2004). Tidal range in particular determines the altitudinal limits of the marsh and its extent landwards and seawards. Note for example that marshes tend to exist within 1m of high tide (Allen, 2000, Davis Jr. & Fitzgerald, 2004). Consequently, modifications to the tidal regime as sea level changes lead to changes in the size and extent of the marsh.

Saltmarshes represent highly effective sediment traps, and should the sediment they sequester be capable of meeting the increase in accommodation space created by a sea level rise, they can effectively inhibit forced transgression, by growing upwards at the same as sea level rise. However, rapid changes in sea level can result in their drowning. During stillstands or slow rises in sea level, organic sediment builds from the decomposition of marsh plants builds up thus prompting the growth of peat which can prograde seawards, resulting in a progradational regression of the coastline (Allen, 2000, Long et al, 2000; Streif, 2004).

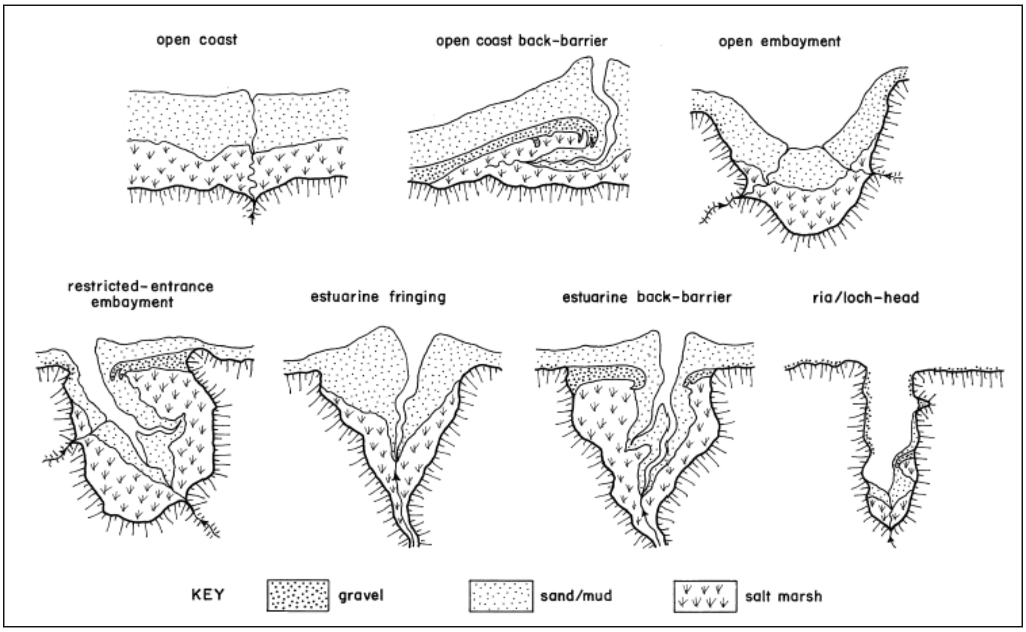

Figure 110. Geomorphological classification of salt marshes (from Allen, 2000).

4.1.4.4 Barriers

Barriers are elongate wave built accumulations of sediment that form parallel to the shoreline. They are characterised by the existence of oceanic and lagoonal shorelines and can range from less than a hundred metres to several kilometres in width, though their exact makeup varies with local sediment availability (Figure 111).

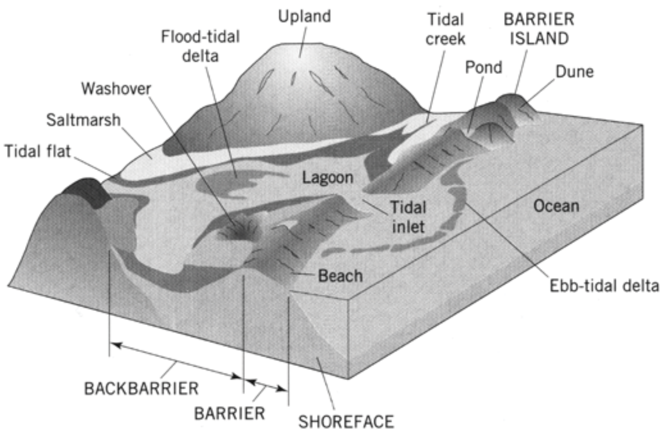

Figure 111. Idealized view of a barrier system (from Davis Jr. & Fitzgerald, 2004)

Types of barriers include spits, which are attached to the mainland at one end, welded barriers, which are attached at both ends, and barrier islands, which are totally detached (Swift et al, 1991; Davis Jr. & Fitzgerald, 2004). They form under accommodation dominated regimes where sediment supply is less than can be accommodated by compaction or rising sea levels, and can thus be reworked and transported by storm or tidal currents (Swift et al, 1991).

The development of barriers is a complex response to gradient, sediment supply, oceanographic conditions and rate of relative sea level change (Roy et al, 1994). They originate in a number of ways depending on the interplay between these factors. One way is as a result of the detachment of mainland beaches during transgression. This occurs when beaches are nourished by littoral drift and can thus keep pace with sea level rise by accreting upwards. However, the back barrier area behind them is sediment starved and starts to flood where it abuts an estuary. Over time the resulting lagoon increases in size and the beach gradually detaches to form a barrier (Swift et al, 1991). Alternatively, they can form as a result of the vertical accretion of offshore bars. Finally, spits may build up in areas of significant longshore transport where the presence of headlands results in the deposition of sediment through wave deflection and the slowing down of the longshore current. These spits may in turn be breached by storms to form individual islands (Davis Jr. & Fitzgerald, 2004).

On microtidal (<2m) or wave dominated coasts, barrier systems tend to be elongated and continuous. As tidal range increases, barriers shorten and tidal inlet width decreases to the point where the barriers disappear altogether on macrotidal coasts and the sediments that would form the barriers is distributed about the sea floor. In these areas coastal morphology generally consists of shore normal tidal channels and open estuaries (Swift et al, 1991). Hence modifications to the tidal regime by changes in sea level have the ability to alter barrier morphology.

Barriers tend to respond in one of three ways to sea level rise (Swift et al, 1991; Cooper, 1994); by erosion, overstepping (the barrier remains in situ) or translating (also known as 'barrier rollover' - Figure 112). Barrier rollover results from the washover of sediment from the shoreface to the backbarrier either from both storm and fair weather wave action (Swift et al, 1991). Thus the entire barrier is reworked over the course of the transgressive event, but is maintained without loss of material.

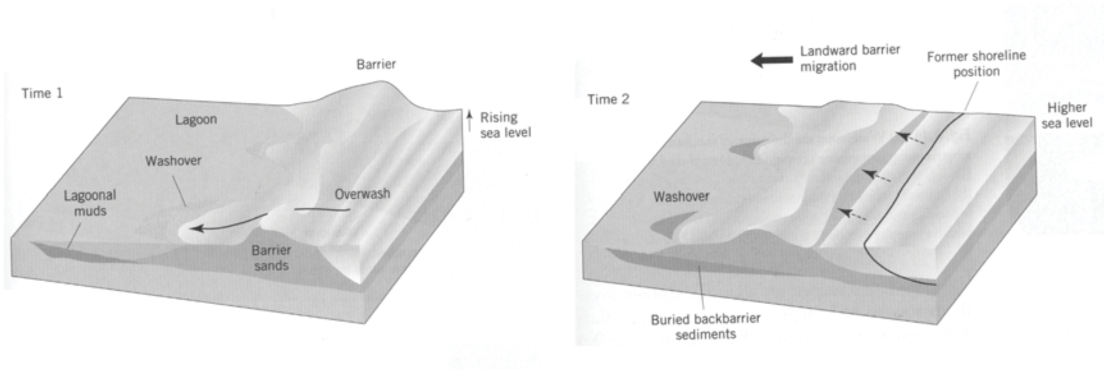

Figure 112. Response of a barrier to sea level rise by barrier rollover (from Davis Jr. & Fitzgerald, 2004)

In erosive situations, although the same cross sectional process is maintained, material eroded from the shoreface is moved seaward and deposited below wave base in the nearshore zone, resulting in the profile moving upward in relation to the sea level rise. Overstepping however drowns the barrier, thus creating a relict landform (Cooper, 1994). Integral to each of these responses is the rate of relative sea level change. Under conditions of rapid rise, overstepping and drowning are more likely, while erosional and translational responses are more likely under conditions of slower rise, which gives the barrier sufficient time to adjust to the changing regime parameters. As always, the exact nature of the adjustment will depend on local geology, antecedent topography and sediment supply (Cooper, 1994). Both overstepping and barrier migration are known to have taken place in different areas during the Holocene sea level rise. This can be illustrated by investigations of the shelf environment which reveal that in some areas barriers are replaced in the stratigraphic sequence by erosional, or ravinement surfaces, resulting from shoreface retreat (Niedoroda et al, 1985; Swift et al, 1991) while, in a number of other areas relict barriers are known to exist on the continental shelf, most likely as a result of barrier overstepping (e.g. Forbes et al, 1995, Oldale, 1995 - Figure 113).

When viewed over timescales of several thousand years, barrier migration is continuous in the face of rising sea level. However on timescales shorter than this, movement may have ceased and progradation resumed as the rate of sea level rise slowed or halted (Swift et al, 1991).

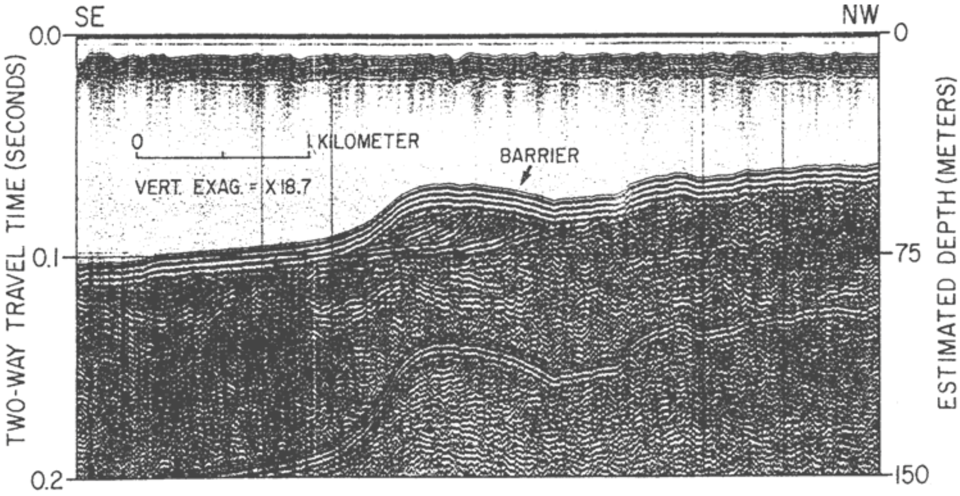

Figure 113. Seismic profile showing a drowned barrier spit located in 50-70m water depth on the North American shelf off Massachusetts. Preservation of the barrier has been attributed to its large size (>20m thick), rapid rate of sea level rise and an abundant sediment supply (from Oldale, 1985)

4.1.4.5 Lagoons

Lagoons are coastal bays that have become restricted by the presence of a barrier (see Section 4.1.4.4). They differ from estuaries in that they have little or no freshwater input and are further distinguished on the basis that the presence of the barrier reduces significant tidal changes. They form where an embayment, and a mechanism for isolating it, such as a spit or barrier, exist (Davis Jr. & Fitzgerald, 2004 – Figure 111). Lagoons themselves may be a response to sea level rise. They form when rising sea levels inundate low lying land behind ridges or dunes, or when barriers grow because of sea level rise (see section 4.1.4.4). Lagoon size and shape is largely a function of local gradient and the extent of sea level rise (Cooper, 1994). During stillstands the restrictive nature of lagoons is such that they infill with sediment, evolving to a marsh or deltaic plain through which rivers drain. Sea level change however modifies this pattern. The response of a lagoon to sea level rise depends largely on the response of the enclosing barrier to sea level rise (see section 4.1.4.4 for further discussion). A landward migrating barrier could result in the landward translation of the lagoon as well, thus maintaining a constant volume, while a vertically accreting barrier may lead to an increase in lagoon volume thus slowing the rate of infill (Cooper, 1994). However the destruction or drowning of barriers by rapid sea level rise results in the loss of the lagoon and its particular environment.

4.1.4.6 Paraglacial Coasts

Glaciated coasts are those whose coastal morphology has been sculpted by the action of glaciers, thus resulting in the occurrence of distinctive features such as drumlins, moraines, fjords, outwash sands and gravels (Davis Jr. & Fitzgerald, 2004; Forbes & Syvitski, 1994). In these situations sediment sources and supply tend to be highly localized and dependent on the former location of glaciers. Hence, the primary impact of sea level rise is that it determines the availability of sediments and the timing and location of sediment reworking. While this is a role common to all coasts, it is somewhat enhanced in paraglacial settings given the localized nature of the sediment supply (Forbes & Syvitski, 1994; Ballantyne, 2002).

It is worth noting at this point that glacial deposits (typically sub-glacial, pro-glacial and glacimarine) related to the growth and decay of ice sheets dominate many mid-latitude and high-latitude shelves and are hence worthy of discussion here. Subglacial sediments are deposited beneath grounded glaciers and ice sheets; proglacial deposits are dominated by high discharge, potentially, high sediment laden, fluvio-glacial channels either cutting into exposed terrestrial environments or directly into the ocean; and finally glacimarine sediments can be defined as being deposited from grounded tidewater ice fronts, floating glacier tongues, ice shelves and icebergs. These patterns of sedimentation vary in three-dimensions (i.e. both orthogonal and parallel to an ice front) and are controlled by the interaction of five factors: pre-existing shelf geometry; variability in the spatial and temporal pattern of ice sheet growth and decay; glacio-eustatic cycles and isostatic crustal response; and post-depositional current re-working. Simply, in proximal locations to the sediment source, steep gradients in grain-size variability can be identified in all dimensions. By comparison at distal localities, depositional environments become more homogeneous over larger areas.

This basic facies architecture described above is spatially related to the location of the ice front. Naturally over timescales of 1000's – 10,000's of years the location of these depositional environments will migrate with the growth and decay of ice sheets. Further, their potential for preservation (i.e. the combined threat of reworking or burial) will be dictated by the associated fluctuation in relative sea-level at any one site. During the major growth phases of continental ice sheets, the ice front can coincide with the shelf edge resulting in the deposition of large volumes of sediment across the shelf and can even result in active shelf progradation.

During these periods of advance individual ice streams may also be capable of excavating large cross-shelf valleys, frequently U-shaped in cross-section, several hundreds of metres deep and several kilometres across (Huuse & Lykke-Andersen, 2000 – see Section 3.4.3). Many of these features are pre-Quaternary in age and probably represent poly-phase erosion enhanced by contemporaneous uplift. These channels frequently represent local depo-centres for glacially derived sediment. During retreat proximal zone processes will replace ice-contact depositional styles (Figure 114). Further, during retreat, oceanic waters will penetrate onto the shelf, increasing long-shelf sediment dispersion whilst the increase in shelf width in response to marine incursion will increase both tidal and wind induced erosion.

It should be noted that during the Pleistocene this temporal variability in ice sheet growth and decay has been predominant on the mid-latitude northern European and

North American shelves. By comparison the high latitude, Antarctic and Greenland ice sheets, underwent very little areal or volumetric change, at least during this last glacial cycle.

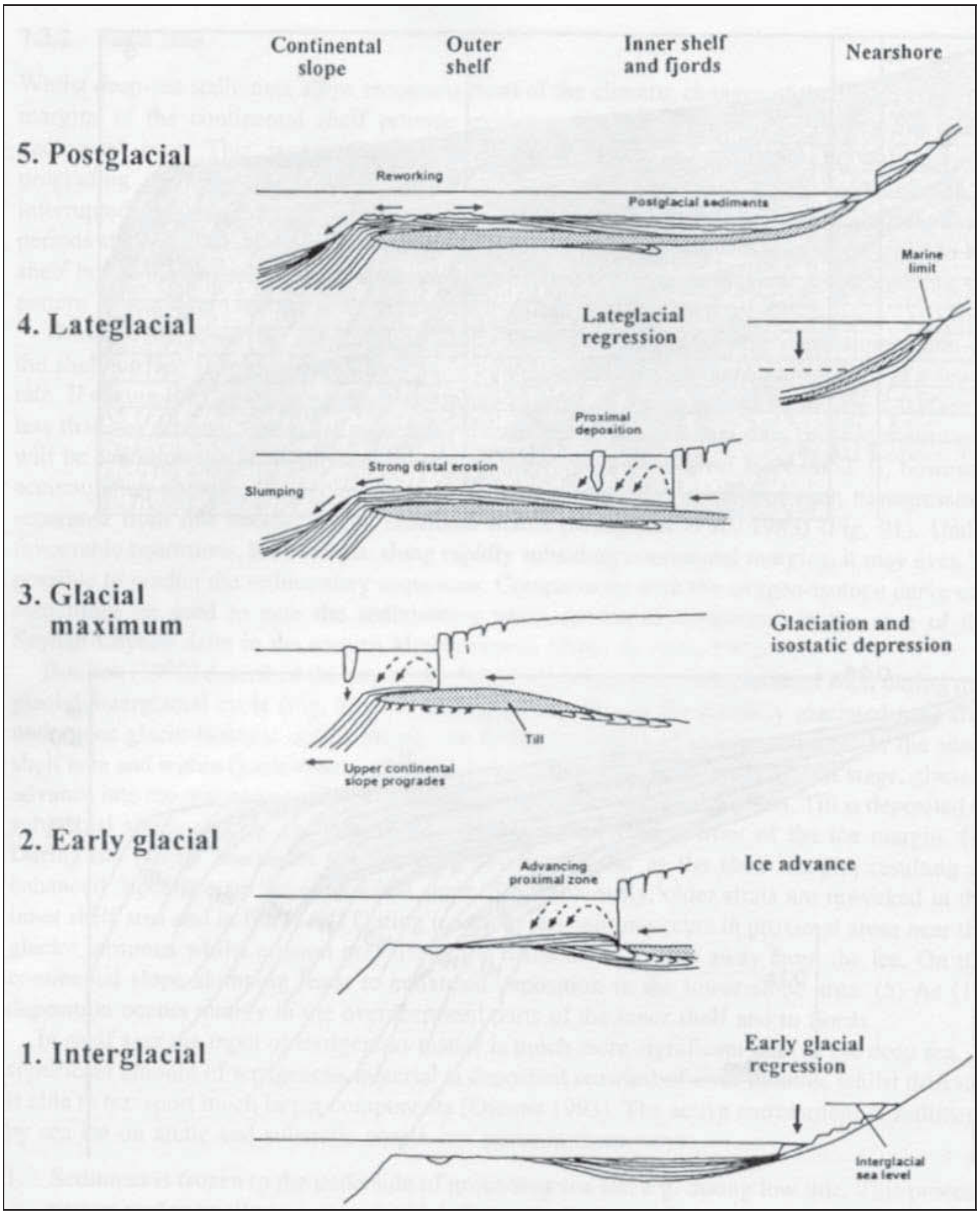

Figure 114. Model of glaciomarine depositional variation in space and time over the course of one glacial cycle. The relative changes in sea level are indicated (After Boulton, 1990).

4.1.4.7 Strandplains

Strandplains are a form of regressive depositional system that develop on the flanks of wave dominated deltas (Swift et al, 1991). They range in length from 5-100 km along shore, 1-30 km in width while their deposits are 10 to 20m thick. Although they are indicative of regressive depositional systems, they can form even when sea level is rising, provided the local sediment source exceeds the accommodation created by the sea level rise. This category is known as ‘rising sea level strandplains’ and contrasts with ‘falling sea level strandplains’ which form during sea level falls (Swift et al, 1991). The progradation of strandplains is episodic, with successive parallel beach ridges added as storms erode sediment from the nearby delta and entrain it within the local littoral currents. In areas of coarse sediment, strandplains are formed by sets of storm beaches, or gravel ridges (Swift et al, 1991).

4.1.4.8 Rocky Shores and Hard Coasts

The previous sections have focussed on depositional coastlines with a relatively abundant sediment supply. This section addresses rocky or hard coasts. These rugged shoreline types form on active continental margins characterised by seismicity, coastal mountain ranges, volcanism and narrow shelves or where the structural grain of the land is such as that it is readily exposed by the longshore removal and transport of sediment (Griggs & Trenhaile, 1994). In these situations, sediment availability is relatively low due to the difficulty of eroding the shoreline. Although erosion is difficult, it is not impossible given enough time and sufficient wave strength. Rising relative sea level therefore results in simple transgression, the effects of which depend on the gradient of the surface being flooded, the local wave climate, local geology and the rate of rise. These serve to determine the erosive impact that the sea has on the coastline in that they determine the strength of the waves, the strength of the rock and the duration for which the waves impact on the shore. This can lead to the development of wave cut notches and platforms which can indicate the position of a shoreline usually at a time of stillstands or very slow rises in sea level (Trenhaile, 2002 – see Section 2.3.2.1). It has been suggested that wave erosion may have been particularly effective during the early stages of interglacials and interstadials when sea levels were rising, and the rocky formations had suffered frost shattering during the preceding cold stages (Trenhaile, 2002). The scree formed by the frost weathering may have protected the main body of the cliff during the early stages of the transgression. The degree of protection it afforded may in turn have affected the impact that the transgression had on the rocky substrate (Figure 115).

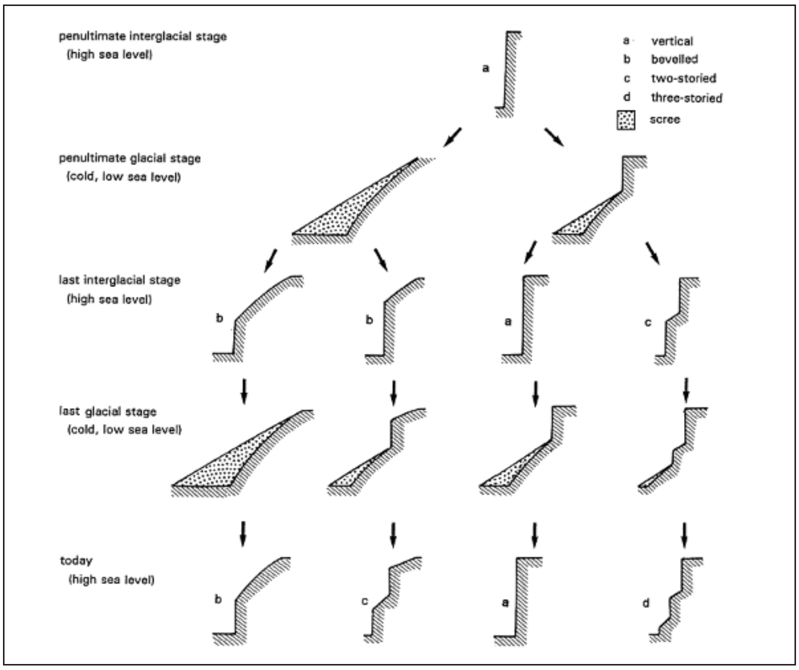

Figure 115. Diagram showing cliff evolution over the course of 2 glacial/interglacial cycles on mid to high latitude coasts. Scree formed during glacial phase was removed by marine erosion. The final morphology of the cliff is determined by the rate of wave erosion and the extent of scree cover protecting the cliff (from Trenhaile, 2002).

In fact, the degree of erosion over the course of one glacial/interglacial cycle has been estimated, on the basis of modelling, at between 1 and 3km (Trenhaile, 2002). More rapid rises are unable to rework the lithified surface and thus the existing shoreface therefore breaks up into a series of rock defended sand bodies, and bays (Swift et al, 1991). Whether or not embayments develop along these areas is determined largely by the extent and steepness of bedrock outcrops, with frequency increasing in areas with steeper and less easily erodable substrates (Roy et al, 1994; Trenhaile, 2002).

4.1.5 Continental Shelf Processes

At highstands and at offshore locations (beyond ambient wave base: see Section 4.1.2) during the transgressive/regressive phase there is the potential for reworking of shelf deposits and the deposition of new deposits. It is therefore important to appreciate the nature of the driving forces controlling sediment transport on continental shelves. This section will therefore discuss the nature of the processes operating from the shoreface down to the shelf break. This should further substantiate the view that present day sea floor bathymetry is not an unequivocal representation of the sub-aerial palaeo-landscape.

Currents on the continental shelf can be divided into three main categories; tidal currents, meteorological currents and density currents (Figure 116: Nittrouer & Wright, 1994; Leeder, 1999).

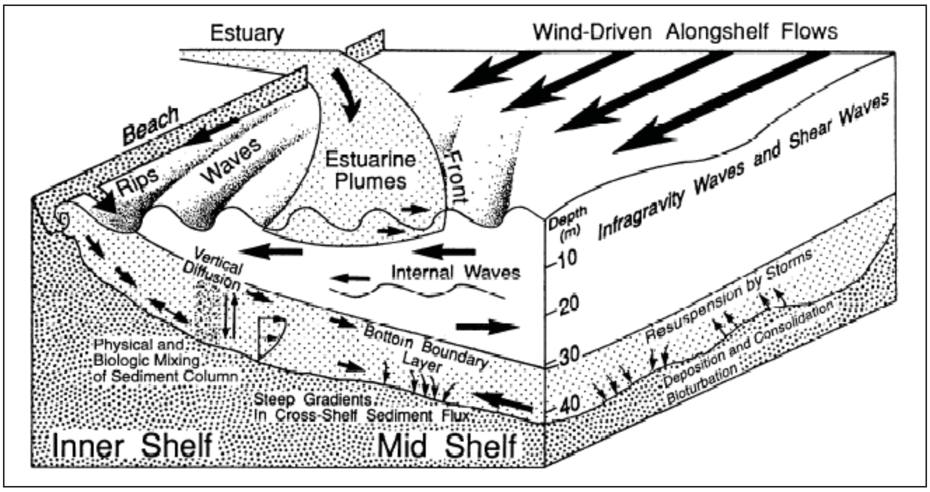

Figure 116. Summary of physical processes responsible for sediment transport on continental shelves (from Nittrouer & Wright, 1994)

4.1.5.1 Tidal Currents

Tidal currents are produced by the semi-diurnal tidal wave created by the gravitational influences of the sun and moon. As the oceanic tidal wave reaches the shelf break, it decelerates due to the reduction in water depth. This in turn increases the amplitude of the tidal wave and enhances the resulting tidal current (Howarth, 1982; Leeder, 1999). Further modifications to tidal strength also result from local basin morphology; amplification for instance may result from reductions in basin depth and width, or resonance whereby the natural oscillation of the sea coincides with the tidal period (Howarth, 1982). In general, the strength of the tidal stream decreases with depth to the point that its strength at 100m depth is half that at 1m depth (Hamblin et al, 1992), but still maintains velocities capable of re-working shelf sediments. For example, bottom current measurements in the Celtic Sea have demonstrated velocities of 1 ms^{-1} at a depth of 165m during spring tides. Furthermore, the circulation of large sand dunes confirms that these currents are mobilizing sediment (Berné et al, 1998). As tidal currents are bi-directional, it is the stronger of the peak ebb or flood tide which determines the net sediment transport direction (Stride, 1982). Shelves whose sediment transport pathways are determined largely by these forces are known as tide dominated shelves. Examples include the North Sea and English Channel areas (Reynaud et al, 2003).

Where tides dominate on shelves the tidal current transport path controls the distribution of grain size, bedforms and facies. In this case there is a generally recognised trend of decreasing grain size down the tidal current path, for example from gravel to coarse sand, to mud, all dependent on the sediment supply. In the older literature this tidally controlled distribution is known as a *nearshore sediment prism* with sand deposited in the nearshore and mud deposited in deeper waters. The complete tidal current path also shows a predictable sequence of bedforms (see Section 4.1.5.4), but the precise type depends on whether the sand supply is high or low. Bedform size decreases towards the end of the tidal current path with small sand waves, sand patches, and then mud deposition. Often the down current reduction in flow strength accompanies an increase in water depth, and in this instance, grain size decreases with increasing water depth. Although mud zones are usually located at the ends of tidal current transport paths, mud can accumulate in a variety of positions on shelves, as a response to the interaction of wind-drift and ocean circulation patterns.

4.1.5.2 Meteorological Currents

Meteorological currents are those produced by wind forcing. Wind, or weather dominated shelves are those on which sediment transport is largely controlled by wind or storm induced currents. On these shelves, tidal ranges tend to be less than 1m, and tidal currents are weak; less than 0.3 ms^{-1} (Leeder, 1999). For example, off north-eastern Denmark temporary wind induced currents can reach near surface speeds of 2 ms^{-1}, enough to induce sand wave formation in an area where tidal current speeds only reach 0.25 ms^{-1} (Johnson et al, 1982).

It is a misconception that the effect of wave induced sediment movements only takes place in relatively shallow areas (Stride, 1982). Recent research has demonstrated that storm currents can actually affect the seabed at depths of over a hundred metres (e.g. Berné et al, 1998) with storm wave generated bedforms having been observed at depths of up to 140m in the Celtic Sea (Reynaud et al, 2003) and 200m on the Oregon shelf (Leeder, 1999). However, it is only the largest storms that can affect the seabed at the outer shelf margin. In general on weather dominated shelves there is an offshore decrease in grain size, although reworking of relict sediments commonly gives a mixed sediment.

It should be noted that most shelves do exhibit a mixture of two forces in time and space (Leeder, 1999). For instance, tidal currents alone have the potential to move sand, and sometimes gravel, in strongly tidal areas of the shelf. In conjunction with storm and wind currents though, sediment can be moved over most parts of the shelf (Johnson et al, 1982). Alternatively, storm induced currents may temporarily reverse the net direction and rate of sediment transport (Johnson et al, 1982). In tidal areas, the main offshore effect of storm waves tends to be an increase in sand transport rates in the direction of the dominant tidal current (Stride, 1982). Both tide and meteorological currents may in turn alter the direction of density currents (see Section 4.1.5.2).

4.1.5.3 Density Currents

In addition to the key role played by tide and wind induced activity on the movement of sediment on the continental shelf a variety of other oceanic factors also play a role, which will vary in their importance along a given continental margin. These additional factors can be grouped into large-scale ocean currents, upwelling/downwelling events and internal tides and waves (Figure 116). The type, the occurrence and intensity of various ocean currents off any continent vary systematically in relation to the eastern and western margins of the ocean basins. Simply, the density driven ocean scale currents tend to be narrow and intense on the western boundary of oceans; e.g. the Gulf stream can locally be as little as 50 km wide and near the surface attains velocities which can range from 1 ms^{-1} to 3 ms^{-1}. Conversely, eastern boundary currents are relatively broad and weak; e.g. the California Current can be as wide as 1000 km and surface flows are typically < 0.25 ms^{-1}. The meandering of such currents adjacent to the shelf break results in both the movement of suspended material within the water column (thus controlling its final place of deposition) and the migration of bed load sediment. In addition, to displacing water by meandering, such currents produces eddies which can replace coastal water with offshore water, thus further complicating the dispersal patterns of sediment.

The influence of these currents can be demonstrated by the distribution of sediment and bedform types on the Spanish Gulf of Cadiz shelf. Here the North Atlantic Surficial Water (NASW) currents flowing southeastward across the Cadiz shelf toward the Strait of Gibraltar dominate sediment distribution. Even inner shelf sandy facies formed by wave activity are modified by the NASW currents to form a belt of sand dunes at 10-20 m water depth. The NASW also dictates the south-easterly progradation of a mid-shelf Holocene mud facies. Further, increased NASW current speeds near the Strait of Gibraltar and the strong Mediterranean Outflow Water currents that are also present at this site, result in a lack of mud deposition and the development of a reworked transgressive sand dune facies across the entire southernmost shelf. It should be noted that such ocean scale currents are significantly affected by local morphological controls (e.g. tectonically induced bedrock highs) which can result in the patchy development of sedimentary facies and bedform variability particularly on the inner shelf.

Landward upwelling and seaward downwelling events can also cause the transportation and deposition of fine grained sediments. For instance, upwelling flows of over 0.1 ms^{-1} are dominant within the bottom layer of the outer shelf off northwest Africa, sufficient to move unconsolidated muds and sands. Consequently, complex patterns of across-shelf particulate transport have been observed for upwelling systems as water motion is 3-dimensional (along-shelf, across-shelf and vertically).

Finally, internal tides and waves propagate along density interfaces and within pycnoclines and they are common in the main and seasonal thermocline of the world's oceans. If the seasonal or permanent thermocline intersects the shelf, these internal features, propagating in the thermocline, can interact with the bed and thus affect sedimentation. As these features are only present under seasonally stratified conditions for some 6-8 months per annum their effect may appear small, but the net effect over geological timescales could be significant. As flows associated with internal tides are oscillatory, both landward and seaward flows occur. When the ratio of the bottom slope is less than unity, as internal waves move across a sloping bottom, they may break and generate internal surf, which could further entrain sediment for transport. Frequently, these internally derived currents interact with tidal and/or wave activity to further enhance sediment movement. Flow velocities associated with internal waves can exceed 0.3 – 0.4 ms^{-1} and are frequently superimposed on other local currents. These velocities are sufficient to significantly affect cross-shelf transport, and may be capable of forming large-scale bedforms with wavelengths of several kilometres.

For the outer shelf of the Celtic Sea, internal waves have been observed to propagate both on-shelf and off-shelf (Heathershaw et al., 1987). These authors suggest that the shelf break actually represents a bedload parting zone. Net movements are off-shelf just below the shelf break and on-shelf immediately behind it. Further, on-shelf the predicted net sediment movements are again off-shelf suggesting a broad zone of convergence extending some 10 km behind the shelf break.

4.1.5.4 Impact of shelf currents

The existence of currents and their impact on present day seabed sediment on the continental shelf floor can be demonstrated by the existence of bedforms:

"Bedforms are an integrated response to all water movements that have been operative during quite a long period" (Johnson et al, 1982:89)

These form in areas that have a source of mobile sediment, either from the seabed, fluvial sources or coastal erosion, and currents of sufficient strength to mobilize it (Belderson et al, 1982; Johnson et al, 1982; Dyer & Huntley, 1999).

Typical features on continental shelves include: sand ribbons, dunes, sheets and ridges and banks. Sand ribbons, can be up to 20km long, 200m wide, and 100m thick and occur on coarse sand and gravel substrates in water depths of 20 - 100m. Dunes can be up to 15m high, have a wavelengths of 600m and occur in very high energy tidal environments, but tend to be absent in nearshore areas where wave activity is high as waves, especially during storms, as these tend to reduce dune height and wavelength (Leeder, 1999). Sheets are relatively thin (up to 12m) but extensive areas (up to 1000km) of sediment laid down by tidal currents. They can be formed from mud, sand or gravel with their grain size being determined by mean spring peak tidal current speeds (Stride et al, 1982). Gravel sheets in particular include clasts of up to boulder size reworked from Pleistocene terrestrial deposits by the early Holocene and transgression and later tidal current activity. The deposition of these extensive facies occurs when current strength decreases to the point at which sediment transport is no longer possible (Stride et al, 1982).

Finally, sand banks form in nearly all shallow tidal seas provided there is a supply of sand, and currents exceed about 0.5 ms^{-1}. Sand ridges are particularly elongated banks. On open shelves these features can be up to 80km long, 13km wide and tens of metres in height with wavelengths of 3 to 12km (Dyer & Huntley, 1999). They in turn may be covered by active smaller features such as dunes (Leeder, 1999). Although they tend to be composed of sand, where currents are sufficiently strong gravel banks may form. Banks and ridges form in a range of depths, from shallow water estuary mouths to depths of more than 150m on the outer shelf, and tend to occur in parallel to one another in groups (Berné et al, 1998; Dyer & Huntley, 1999; Stride et al, 1982).

A number of different classes of bank and ridge exist, including; open shelf ridges, tidal delta ridges and headland associated ridges. Morphological and evolutionary differences between these classes come about as result of the varying hydrodynamic circumstances in each situation. Changes in sea level have been frequently associated with the formation and maintenance of sand banks (see Section 4.1.5.5).

The smaller of these bedforms tend to be more indicative of short term events, such as storms (Johnson et al, 1982). Whereas the larger features highlight the effect that currents may have had on the morphology and topography of submerged palaeo-landscapes over the long term. This is because larger features respond more slowly to changes in the hydrodynamic regime and thus are a better representation of long term change.

4.1.5.5 Impact of transgression and regression on shelf processes

If sea level change accompanies transgression and regression, it may have a significant impact on regional and local tidal regimes and wave climates, and in turn these will affect the transport patterns of sediment around the coasts and on shelves, and therefore, the evolution of coastal and shelf geomorphology on timescales of thousands to tens of thousands of years (Van der Molen & van Dijck, 2000; Van der Molen, 2002). Modification of tidal regimes and wave climates by sea level change comes about because the resultant bathymetric shifts modify currents strengths by increases or decreases in bottom friction (a function of depth) while coastline changes may divert or deflect existing currents. Further impacts resulting from sea level change are changes in the tidal range and tidal prism due to modifications to basin morphology (Scourse & Austin, 1995; Shennan et al, 2000).

Numerical modelling of these forces can provide reasonably effective simulations of past changes in both tidal regime and wave climate in response to sea level change (e.g. Scourse & Austin, 1995). For example: Holocene changes in sea level around the British Isles appear to have led to relatively minor changes in areas where the tidal wave propagates progressively and without entering shallow water in regions such as Northern Scotland. Conversely, major changes are apparent further South where the opening of the straits of Dover resulted in the conversion of the Southern Bight from a quiet microtidal sea, to one with a larger tidal range (>2m) subject to tidal scouring (Austin, 1991; Van der Molen & de Swart, 2001a). Concomitant with these changes were modifications to the strength and direction of sediment transport, notably a switch from onshore transport towards the Low Countries before 6 (C^{14}) ka BP to alongshore transport by the present (Figure 117 - Van der Molen & van Dijck, 2000).

In terms of wave climate, modelling suggests that wind-wave conditions and associated orbital velocities and sand-transport patterns have changed over time as the basin geometry changed. The outputs of a model for the southern North Sea (Van der Molen & de Swart, 2001b) show that mean wave heights increased after 7.5 (C^{14}) ka BP with the largest changes occurring in the shallow water (Figure 118). Further, the dominant form of wave induced transport switched from suspended load to bedload transport after 6 (C^{14}) ka BP, while the overall east to west direction of the bed-load transport remained constant until the present day but decreased slightly in magnitude.

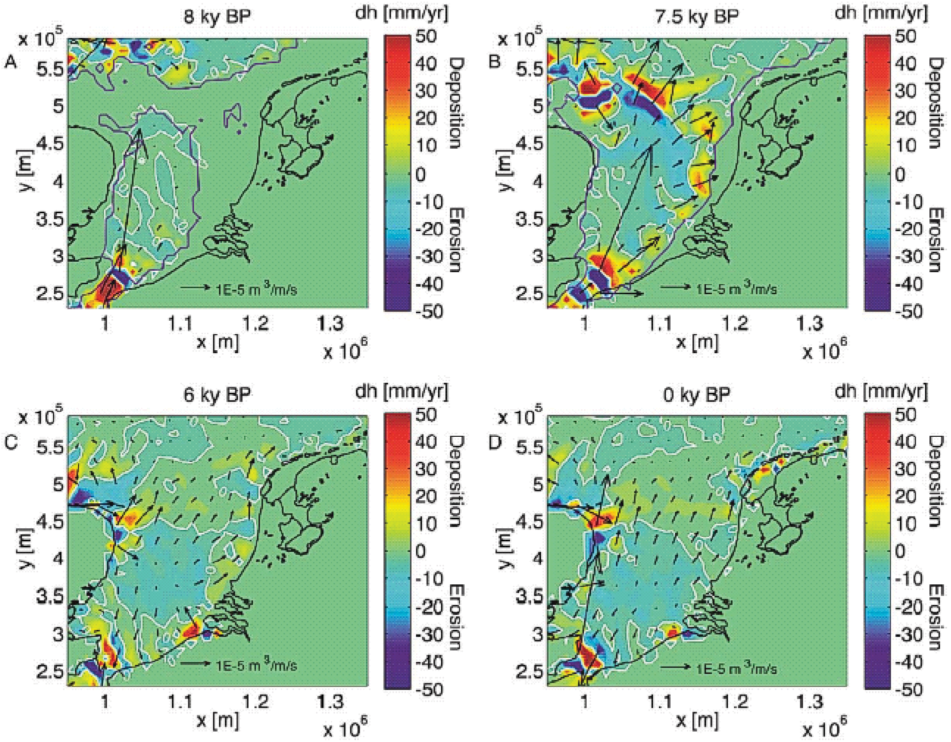

Figure 117. Sand transport patterns for the southern North Sea during the Holocene. Arrows indicate transport directions and strength (m^3/ms^{-1}), colours indicate erosion/deposition rates (mm/year) (from Van der Molen & Van Dijck, 2000).

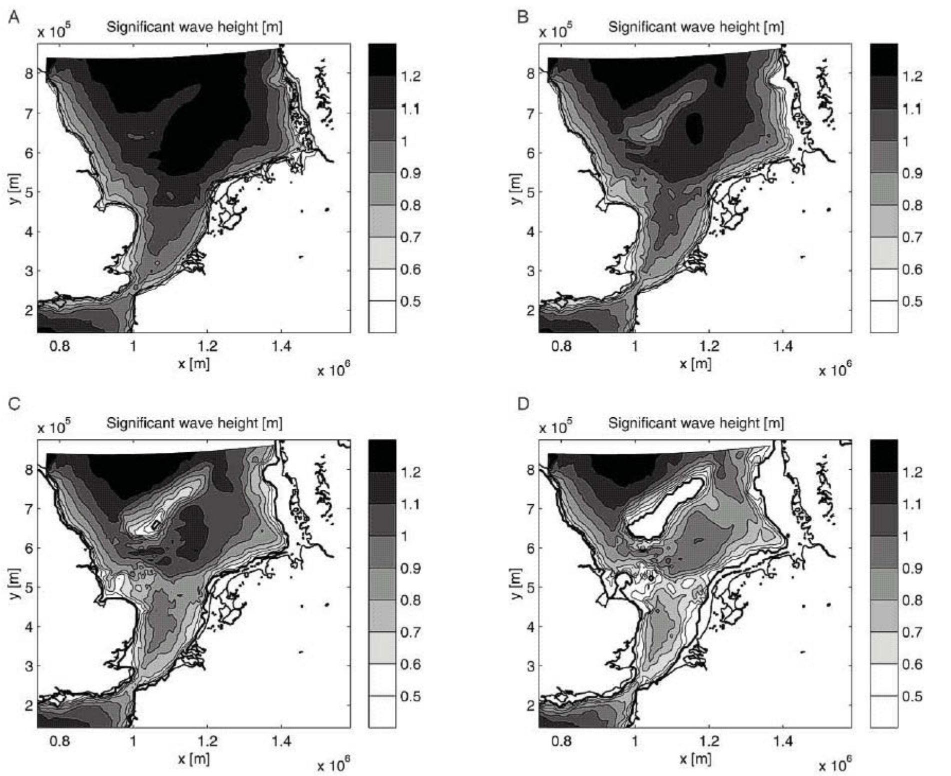

Figure 118. Results of significant wave height modelling in the North Sea for A) the present B) 6 ka BP C) 7 ka BP D) 7.5 ka BP (all dates are uncalibrated C^{14}) (from Van der Molen & de Swart, 2001b)

The products of these modifications to the hydrodynamic regime are changes in seabed morphology. Estuary mouth sand ridges for instance are interpreted as forming as a result of rising sea levels increasing tidal flow into an estuary. This increased flow in turn widens and deepens tidal channels with the eroded sediment deposited on the channel margins. Over time this sediment builds up into ridges (Dyer & Huntley, 1999). Conversely, moribund sandbanks are found in deep water, the rising sea levels of the transgression making possible the preservation of banks by removing them from zones of significant sediment movement (Dyer & Huntley, 1999).

Alternatively, a number of deep water ridges have been interpreted being created during the Postglacial transgression by the erosion and remoulding of pre-existing lowstand estuarine or deltaic deposits (Berné et al, 1998). In these cases, the eroded remains of nearshore features such as fluvio-estuarine, barrier and tidal-delta deposits are overlain by offshore dunes created by tidal action, which are in turn reworked by storm wave induced currents as the influence of constructional tidal forces wanes with depth increase caused by rising sea levels (Reynaud et al, 1999 – Figure 119).

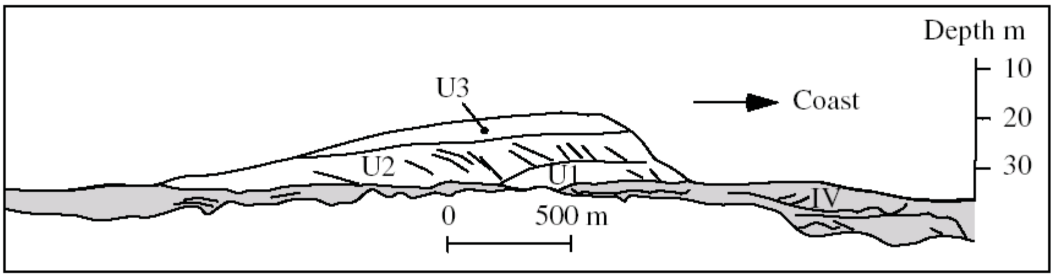

Figure 119. Cross-section of the Bassure de Baas bank (Eastern English Channel). Seismic units have been interpreted as follow: U1) Remnant of tidal delta lobes situated above and infilled valley. U2) Remnant of shoreface bank migrating coastwards under wave action. U3) Tide-dominated reworking of shoreface banks. Final stage of bank construction. Overall the bank highlights the transformation of coast and shoreface under transgressive conditions (from Reynaud et al, 2003).

4.1.6 Implications for Archaeological Studies

Archaeological work on continental shelves requires a secure grasp of palaeogeography and palaeotopography for the purposes of site prospection, cultural resource management and interpretation. This is especially important if predictive modelling is to be undertaken as many models are topographically based (see Section 5).

A review of sedimentological, geological and coastal management literature has highlighted the fact that coastline and shelf surfaces change under transgressive, regressive and stillstand regimes. Furthermore, the resulting stratigraphic record is fragmentary as in many cases processes of erosion and deposition accompanying and following transgression have obscured or destroyed much of the original sedimentary record (Streif, 2004). Therefore, there are a number of implications for archaeological research:

- Coastal and shelf surface changes under transgressive conditions are such that present bathymetry cannot be used as a direct representation of the past landscape. This is exacerbated by the modification of the shelf surface by stillstand processes.

- Care should be taken when extrapolating relatively restricted (spatially and temporally) evidence, or presenting palaeogeographic reconstructions as accurate representations.
- Sub-bottom investigations should be an essential part of archaeological investigations to enable a more secure understanding of how the landscape has changed over time.
- Topographical and morphological changes render the development of predictive models more difficult but not impossible. Certain important landscape features, (e.g. buried channels) are still present.

Future areas for research include:

- The increased integration of existing sedimentological and geological data with archaeological work to enhance existing palaeogeographic reconstructions and construct new ones.
- Increased investigation underwater using remote sensing and geophysical equipment, not simply to locate archaeological material, but to obtain information that will aid in understanding the evolution of the wider landscape.
- The possibility of using parameters other than topography to construct predictive models. Suggestions include lithic raw material (e.g. flint and chert). This approach has been tested to some extent off North West Florida (Dunbar et al, 1991). This will be discussed further in Section 5.
- This study has focused on landscape changes taking place over a single transgressive cycle. However the investigation of pre-Last Glacial Maximum deposits will require knowledge of preceding glacial and interglacial stages. A greater understanding of landscape evolution over multiple trans- and regressive cycles is necessary.

4.2 Response of Individual Deposits of Archaeological Material

4.2.1 Introduction

The previous chapters have demonstrated the possibility that archaeological material exists on continental shelves. However, as briefly outlined in Sections 3.4, 3.5 and 3.6, it is possible that syn-transgressive and regressive marine processes have modified the arrangement and makeup of the presently submerged archaeological record. This chapter seeks to further investigate the impact of these processes on individual archaeological deposits. Therefore, it aims to:

- Determine the preservation potential or likely state of the archaeological material on the seabed.
- Consider issues of temporal scale. The impact of single events (e.g. storms), versus longer-term oceanographic processes on archaeological material.

In the interests of assessing the archaeological potential of continental shelves, the extent and nature of this modification must be investigated, since, as previously stated (Section 3), deposits in primary, secondary and tertiary context have different interpretative values, in that they each are suited to addressing particular research questions. In addition, knowledge of where archaeological material is likely to have survived transgression, in a particular preservational context, may be important in designing future strategies of site prospection or cultural resource management.

A review of existing literature has indicated that approaches to the modification of archaeological material by transgression, regression and marine processes are somewhat sparse. Note for instance the following statement:

"the impact of inundation upon terrestrial sediments, a post-depositional process not hitherto relevant to Palaeolithic studies and one whose possible effects need to be considered" (Wenban-Smith, 2002:7).

The dominant perspective in much of the archaeological literature assumes that transgression will result in the destruction of sites through erosion.

"lake and sea levels have varied tremendously over the past 2 million years, and erosion during high stands has repeatedly obliterated the archaeological record where evidence for early aquatic resource use is most likely to be found" (Erlandson, 2001:300)

"with the Post-Glacial sinking and flooding of the respective coasts and deltas of the Late Glacial river systems, such diagnostic [coastal] sites may have been lost to the archaeological record due to major incision and aggregation" (Newell & Constandse-Westermann, 1996:385).

Statements such as the above tend to be based on assumptions rather than empirical evidence. However, the work that has been done on the subject of marine taphonomic processes (e.g. Flemming, 1983; Kraft et al, 1983 – Figure 120) emphasises the fact that archaeological material can survive marine transgression. This emphasis is borne out by the fact that globally, around 500 submerged archaeological sites are known (Flemming, 1998), several of which can be considered to be in primary context, for example, Tybrind Vig (Andersen, 1985).

These studies identified some basic taphonomic principles that ensured the survival or destruction of archaeological sites undergoing submergence. Destructive factors were identified as wave erosion, current erosion and ice erosion. Hence, the survival of sites occurred in situations in which these destructive influences were minimized. Key areas identified were lagoons, sheltered alluvial coasts, accumulating beaches, sea caves, karstic caves, the lee of coastal islands and coral reefs. These have been assessed in terms of wave heights, wave fetch and archaeological sites found in them (Flemming, 1983). Burial of sites to a sufficient depth in sediment, prior to transgression, such they were not affected by the passage of the erosive shoreface was also identified as a critical factor in their survival (Kraft et al, 1983). In summary, ideal conditions for preservation could be described as those that promote gentle, yet rapid burial (Dunbar et al, 1991). A further important conclusion was that while gross topography (i.e. spatial scales of 10-100km) may have created generally favourable conditions, it was the local topography (i.e. spatial scales of < 1 km) surrounding a site that determined its survival (Flemming, 1983).

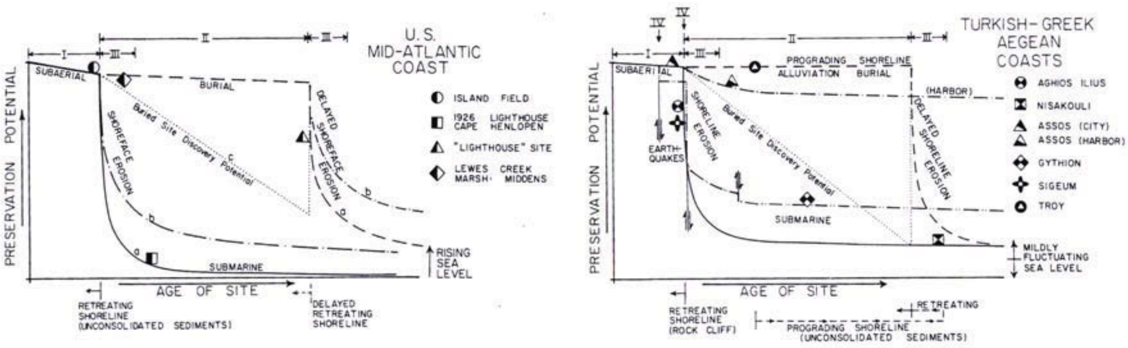

Figure 120. Kraft et al's (1983) conceptual model of submerged archaeological site preservation. Roman numerals refer to phases in the history of a coastal site. I) Subaerial degradation. II) Burial. III) Shoreline erosion. IIIa) Open ocean conditions. IIIb) Estuarine conditions. IV) Tectonic movement.

This work represented something of a starting point as prior to this, little had been done on the taphonomy of submerged terrestrial sites beyond the discussion of individual archaeological deposits (e.g. Andersen, 1980). These studies made use of a relatively small and quite diverse body of data to infer some very general principles. Kraft et al's (1983) conceptual model was based on two very different regions – the mid-Atlantic coast of the USA and the Aegean coast – while Flemming's (1983) inferences were based on a sample of sites ranging in age from the Middle Palaeolithic to the Bronze Age, and from across the globe.

This work though was not entirely comprehensive. For instance, given the small and diverse nature of the sample coupled with the temporal and spatial variability of sea level and coastline change, questions can be raised about the applicability of the general principles of site survival in all situations. Indeed, Flemming (1983) pointed out that:

"A complete study of the survival of archaeological materials would have to include the mechanism of immediate preservation at the site of occupation in the short term; the mechanism of surviving marine transgression; and the mechanism of surviving underwater for many thousands of years" (Flemming, 1983:164)

However, despite this rather promising start, the issue of marine taphonomic processes and their impact on submerged terrestrial material has not really been discussed or taken further. This paucity of recent work is illustrated by recent reviews of submerged archaeology (Flemming, 1998; 2002) which, in discussions of 'the taphonomy of submarine occupation' (Flemming, 1998:134) reference Flemming (1983) and Kraft et al (1983). This contrasts somewhat with work on site formation processes in other facets of maritime archaeology. Note for instance the updating of Muckleroy's (1978) model of shipwreck taphonomy by Ward et al (1999). It seems that there is a gap in research between studies of large scale landscape evolution (Section 4.1) and the above site specific studies that needs to be filled.

Expanding the perspective from maritime to terrestrial archaeology, some work does exist on the responses of prehistoric material to fluvial processes, such as Schick's (1986) work on East African Lower Palaeolithic material, and more recently experimental work on biface and flake distribution in Welsh rivers (Hosfield & Chambers, 2002; Hosfield, 2004). However, these sort of studies have yet to be extended to the marine situation.

As the archaeological approaches to this topic are rather limited, attention will also be turned to the areas of marine geology and sedimentology. A significant amount of research has been undertaken on the subject of sediment movement by marine processes for coastal engineering and management (e.g. Orford et al, 2002) or minerals prospection purposes (e.g. Corbett & Burell, 2001). The reasoning behind using these approaches to gain insights into the reworking of archaeological deposits is that much archaeological material, especially the lithic implements which comprise the vast majority of the prehistoric record (see Section 3: Theme 2), can be regarded simply as 'unusually shaped clasts' (Hosfield, 2004), and should theoretically respond to marine processes in similar ways to natural sedimentary particles of similar size, mass, shape and material type. Nevertheless, Schick (1986) does suggest that the unusual morphology of archaeological material relative to natural sediment, and its restricted size category within the geological range may affect its behaviour in a fluid medium, and consequently the geological studies are unlikely to provide exact analogues to the archaeological situation.

This investigation will focus primarily on coarse clastic deposits. These consist of all grains commonly classified as coarser than sand on the Wentworth scale, or greater than 2mm in diameter, ranging from granules to boulders (Pethick, 1984). This follows the terminology of Orford & Carter (1993) which, in addition to the defining 'coarse clastic', used the term gravel to refer to all material between pebble and cobble size classes (between 4 and 256 mm in diameter) (Carter & Orford, 1993; Pethick 1984). This is because bulk of the archaeological material from the periods in question, notably lithic implements, is rarely less than several centimetres in size (Schick, 1986).

Further potential departure from the situation in reality is linked to the methodology of both the archaeological and sedimentological approaches, namely the use of laboratory based flume experiments and natural field experiments. The disadvantages of using these sorts of studies is related to the general problems associated with them. Natural field experiments often have a lack of control in monitoring variables; consequently, it may be difficult to determine the exact effects of individual variables. Flume experiments meanwhile are better able to determine individual effects, but do suffer from the fact that they oversimplify complex systems (Nash & Petraglia, 1987).

Therefore, at this rather introductory stage of research, any insights drawn out of the existing body of work will not constitute unequivocal statements as to how archaeological deposits may be reworked in the coastal and shelf environments. However, they may be able to illustrate the types of general processes that should be considered when assessing the condition and location of archaeological contexts in the marine environment, and also indicate any questions which can be addressed by future research.

4.2.2 Sediment Dynamics

4.2.2.1 Basic Theoretical Principles

This section will discuss the basic concepts of sediment dynamics that are necessary to understand how archaeological material might respond under transgressive or marine conditions.

As fluid passes over a bed of sediment grains, it decelerates due to friction against the bed surface. This generates horizontal 'shear stresses' in the layer of fluid closest to the bed; the 'boundary layer'. The momentum of the moving fluid is transferred to the sediment grains by these shear stresses and exerts a force on them (Figure 121).

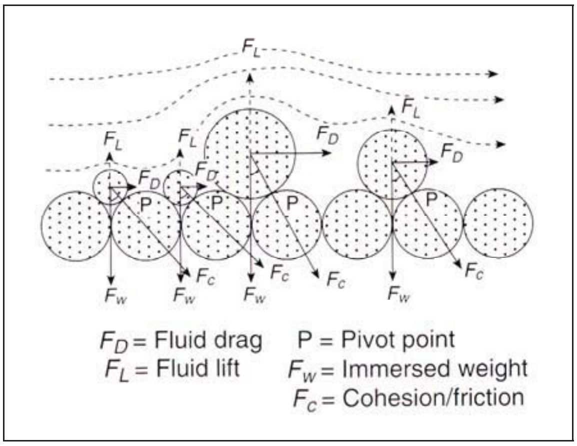

Figure 121. Summary of forces acting on sediment grains that are involved with entrainment and transport (modified from Woodroffe, 2003).

Should this force be sufficient to overcome the forces holding a grain to the rest of the bed, then it will be lifted out of the bed and over the underlying layer. The strength of these latter forces is determined largely by the grain's mass, and whether it is attracted or interlocked with other grains. The point at which grains begin to move with fluid flow is known as the 'critical threshold velocity' (Leeder, 1999). Key to this process are:

- Grain size and density
- Bottom boundary layer currents
- Type of flow
- Interactions between sediment particles

These will now be discussed in turn.

- **Grain size and density:** of each grain plays a part in determining its critical threshold velocity. Larger grains tend to have higher critical thresholds than smaller grains and hence require larger shear stresses before they move. Figure 122 provides an indication of the current velocities required to induce transport in material of different grain sizes. This effect is illustrated by the fact that around the British Isles, tidal current gravels tend to be only a few centimetres thick (though exceptions do exist in areas of strong current), while associated muds can be up to 30m thick and are potentially much more extensive, due to the weakness of currents over large areas of shelf (Stride et al, 1982). Deposition takes place as overall activity decreases to the point where grains are static for longer periods and mobile for shorter ones until they are either buried deep enough for only the most powerful water movements to affect them, or if current strength decreases sufficiently (Stride, 1982). This relatively simple relationship however, is modified by the nature of the bottom boundary currents, the nature of the flow and the effects of large scale bed morphology. The theoretical current velocity estimated to be capable of transporting coarse clastic material is of the order of c. 0.75 ms^{-1} or greater (see Figure 122 - Stride et al, 1982). It should be mentioned though, that palaeo-environmental evidence, which often forms a important component of any archaeological site, may be smaller than this coarse material. This sort of evidence includes pollen and microfossils of plants and animals. It is therefore worth bearing in mind that though a site may appear to be in primary context (i.e. the more obvious artefacts are still in situ), less visible components may have been winnowed away.

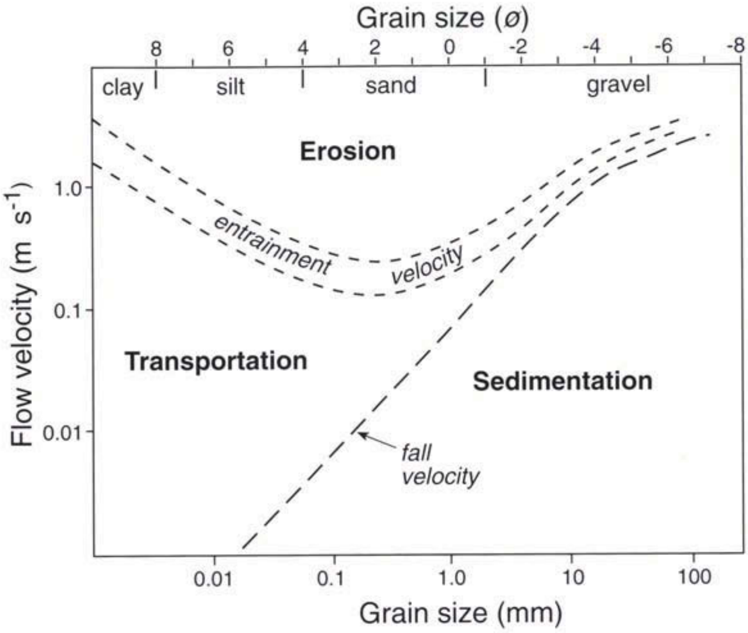

Figure 122. Hjulstroms Curve. This provides a measure of the critical entrainment velocity for material of different grain sizes (from Woodroffe, 2003).

- **Bottom boundary layer processes:** involve interactions between currents, waves, bed morphology, sediment suspension and transport. Shear stresses are a function of current velocity in that increased velocities result in higher shear stresses. Therefore shear stresses, and hence sediment movement, induced by both waves and tidal currents are depth dependent in that they decrease as depth increases (see Sections 4.1.5.1 and 4.1.5.2, Figure 123).

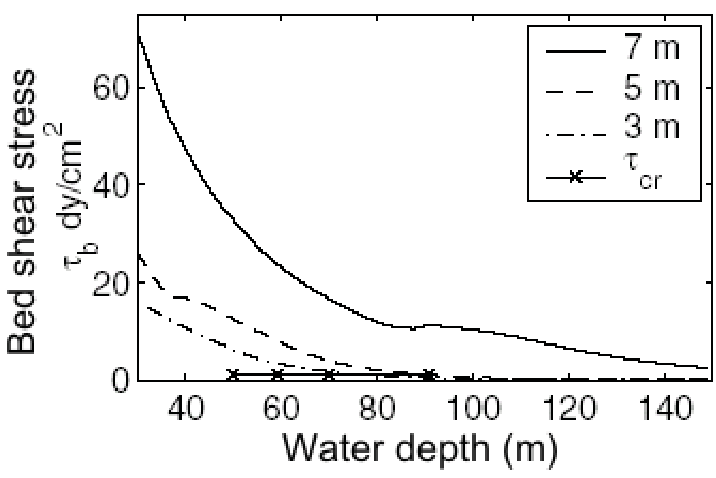

Figure 123. Bed shear stress across a continental shelf calculated for significant wave heights ranging from 3 to 7m. Note the decrease with depth in all cases (from Harris & Wiberg, 2002).

- **Type of flow:** studies of sediment movement in the marine environment suggest that the consideration of two types of flow is necessary (Leeder, 1999; Paphites et al, 2001). Firstly, unidirectional flow is essentially what was discussed at the beginning of this section. It is a flow in one direction generated by tidal or density currents that has the capacity to entrain sediment depending on current speed, sediment grain size and density. Secondly, oscillatory flow is caused by the motion of water particles under wave conditions. At the water surface, water particles move in a circular orbit. However, as depth increases, the orbit decreases exponentially (Figure 124). At the point at which depth equals one-half wavelength, there is virtually no wave induced movement of water particles. In shallow regions, the seabed may be located at less than one-half wavelength, and the orbits become progressively flattened. Consequently, at the seabed water moves in an oscillatory ('to and fro') motion (Figure 124 - Open University, 1989; Leeder, 1999; Davis Jr. & Fitzgerald, 2004).

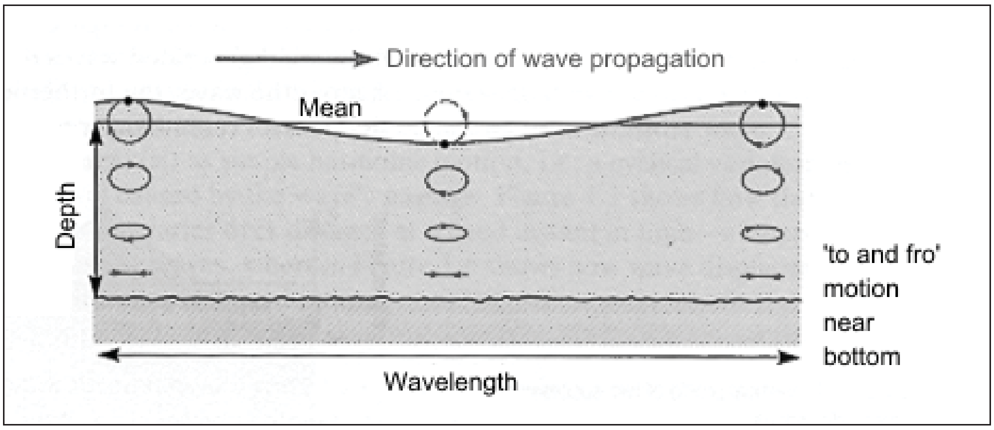

Figure 124. Diagram demonstrating the flattening of wave orbital movements in shallow water (after Open University, 1989).

Each type of flow has the potential to affect sediment grains in different ways: Assuming it is capable of entraining the sedimentary grains in question, unidirectional flow will move the grains in a single direction until it is incapable of doing so (i.e. the bed shear stress falls below the critical shear stress). Under oscillatory flow, sediment grains will move to and fro but with a net movement in the direction of propagation of the wave. This results from the fact that the orbital velocity of a wave is not the same speed in either direction (Figure 125). At wave troughs, the distance of a water particle to the seabed is reduced and concomitantly, frictional retardation by the bed is increased. Consequently, particles speeds are higher at wave crests (onshore movement) but are only maintained for short intervals of time. In contrast at wave troughs, speeds are lower (in the offshore direction) but are maintained for longer periods (Open University, 1989). This asymmetry means that both coarse and fine sediment are transported shorewards, but often the coarse fraction cannot be returned seawards (Open University, 1989). This net onshore transport of coarse material combined with net offshore transport of finer material results in cross shore segregation of sediment characterised by offshore fining (de Meijer et al, 2002).

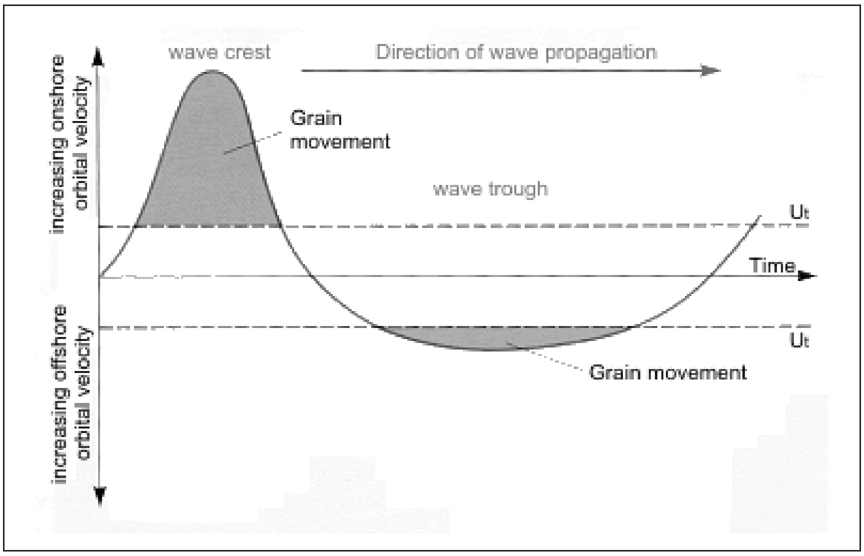

Figure 125. Asymmetry of water particle velocities associated with wave orbital motions in shallow water. Ut is the critical threshold velocity at which grains of a given size will begin to move. Grey areas represent the range of velocities for which the grains will transported (from Open University, 1989).

In most cases, these flows do not operate in isolation, but interact to produce combined shear stresses (Paphites, 2001). In general, the interaction actually reduces the threshold velocity of the sediment, and can enhance sediment transport. This happens because the oscillatory movements of waves can lift sediment into suspension at lower equivalent velocities than a steady current. Once in suspension, it can be moved by the unidirectional flows which, on their own would have been incapable of generating sufficient shear stresses to lift them off the bed (Open University, 1989). The nature of the interaction is also affected by the period of the waves. For long period waves (e.g. > 10 seconds) the two components interact in a linear fashion such that as the importance of wave action increases, the influence of unidirectional flow decreases and vice versa. Furthermore, the longer the period of the wave, the more 'developed' the interaction between flow and oscillation will be and hence will result in a lower threshold for sediment movement (Paphites et al, 2001). However, in shorter period (e.g. c. 5 seconds or less) waves, the thin turbulent boundary layer of the waves can actually suppress the impact of unidirectional flow thus increasing the threshold required for grain movement (Paphites et al, 2001).

- **Interaction between sediment particles and bed morphology:** it should be noted that grain size and density is not the sole control on transportability. Electrostatic forces may hold grains together, particularly in estuaries where the water chemistry promotes flocculation (the amalgamation of clay particles through electrostatic attraction: Leeder, 1999). Further, the transportation of small grains may be retarded if they shelter amongst larger particles (Hosfield, 2004). Artefacts may also be trapped in certain areas by localized modifications of the flow patterns, such as scour pits (Schick, 1986). The propensity of this to happen depends to a large extent on bed morphology. Sheltering for example is more likely to occur in a coarse grained bed than a fine grained one as the gaps between the larger particles can potentially trap smaller grains.

In addition, the composition of sediment may also have an impact on the resulting transport patterns. In mixed sediments for example, the threshold velocity of the coarser fraction tends to decrease, as they protrude further into the boundary layer. The resulting lower pivoting angles then appear to maximise the local lift and drag forces, thus facilitating its entrainment (Panagiotoupolous et al, 1994; Paphites et al, 2001; De Meijer et al, 2002).

4.2.2.2 Modes of Transport

Transport of particles takes place in 2 modes; bedload transport and suspended load transport. In the former, particles are supported intermittently by grain to grain contact, while in the latter they are supported by turbulence (Nittrouer & Wright, 1994; Leeder, 1999 - Figure 126).

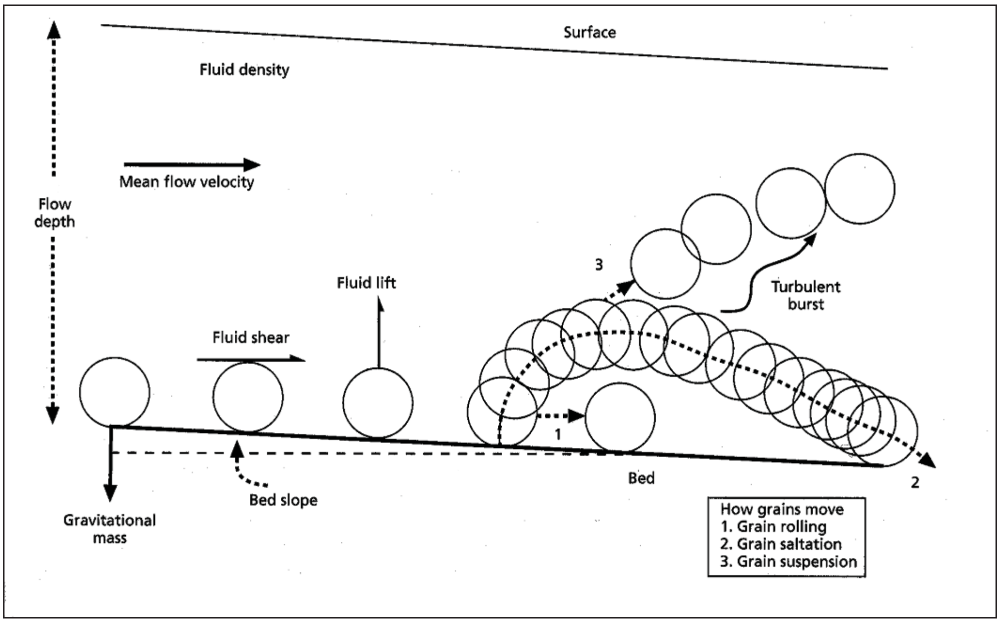

Figure 126. Grain movement under conditions of flow. Grain rolling and saltation (intermittent rising and falling of grains in conjunction with movement in a horizontal direction) constitute bedload transport. Grain suspension constitutes suspended load transport (modified from Leeder, 1999).

Coarse particles tend to be suspended less often than finer ones, due to their greater threshold velocities. The same is likely to be true of archaeological material (Schick, 1986). The movement of sediment grains tends not to be continuous, but to occur in short periods when the shear stress is sufficient to enable erosion and then transport, followed by longer periods when the particles are stationary (Hosfield, 2004). At these times it is possible for the particles to be buried beyond the range of transport processes and remain in situ (Nittrouer & Wright, 1994). This processes is complicated by the fact that differential transport of particles may occur with respect to particle size, shape and composition (de Meijer et al, 2002; Lee, 2001; Nittrouer & Wright, 1994). This means that certain particles, or particle classes may be selectively deposited, transported or buried at particular points on the shelf, thus somewhat complicating the overall pattern of sorting. Theoretically, the same should be true of archaeological material and will be discussed further in the following sections.

4.2.3 Archaeological deposits on coastlines

For the purpose of this discussion the factors to be focused on concern the proportion of sediment that is removed from the shoreline, the distance it travels, and the depth (both of water, and sediment) to which these processes operate as these should theoretically determine whether an assemblage in primary context is reworked into secondary context. The coastline or beach (Figure 127) is generally considered the most active part of the continental shelf (Emery, 1968). Forces operating here are often driven by breaking waves, each of which has the ability to move sediment (Davis Jr. & Fitzgerald, 2004). However, the diurnal tidal cycle also results in twice daily unidirectional currents flowing onshore with the flood tide and offshore with the ebb tide. Thus the exact contribution of each type of flow (i.e. the local hydrodynamic pattern) depends on the local tidal prism and local wave climate. These will be influenced by local bathymetry, shoreline morphology and geology and wind patterns (see Section 4.1.5.5).

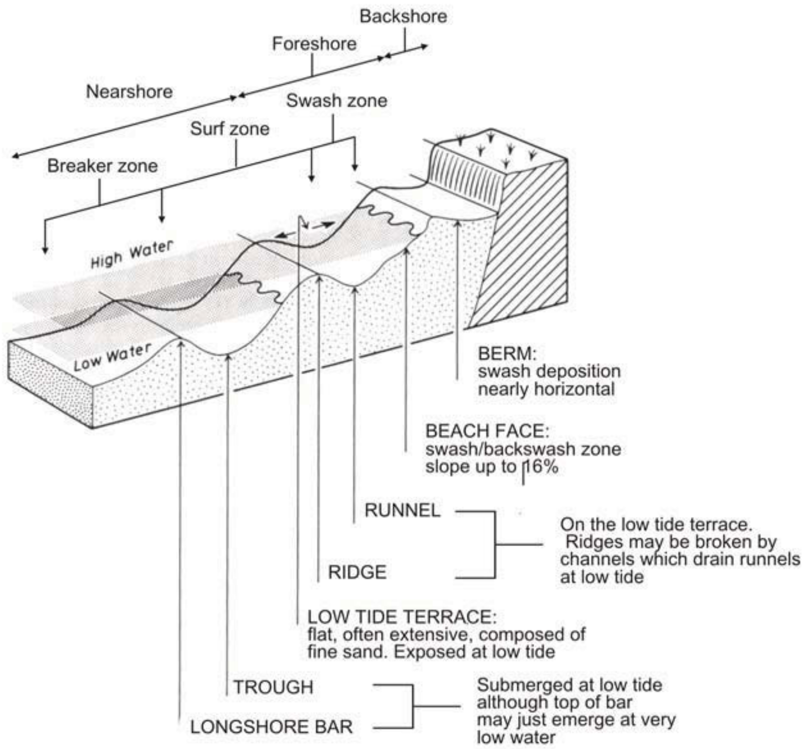

Figure 127. Idealized profile of a beach (modified from Pethick, 1984).

The gradient of the beach also influences the way sediment on it behaves. On gently sloping beaches, wave energy is gradually dissipated across a wide cross shore area, thus promoting the deposition of sediment onshore. In contrast, steep beaches tend to be erosional as the gradient reflects a significant amount of water, and hence energy, back to open water. This enhanced backwash therefore promotes removal of sediment from the beach surface (Davis Jr. & Fitzgerald, 2004). Furthermore, the gradient of the beach, in conjunction with the local tidal range, controls the amount of time for which a particular hydrodynamic zone (e.g. swash zone, surf zone) spends over a particular section of the intertidal zone. This in turn influences the degree of reworking at a given section of the beach (Jackson et al, 2002).

In terms of impact, the most obvious aspects to consider are transport of the artefacts, and damage to the artefacts, such as through abrasion or rolling (Hosfield, 2004; Schick, 1986). Transport of the artefacts results in the transformation of primary contexts to secondary contexts, and secondary contexts into tertiary contexts while damage may make the artefacts more difficult to identify, interpret and analyse.

When considering how archaeological material on the shoreline might respond to transgression, we also have to consider whether it lies on the surface, or is buried within the coastal sediment. The reason behind this lies in the fact that burial should protect archaeological deposits from taphonomic forces (Flemming, 1983; Kraft et al, 1983).

4.2.3.1 Exposed material

As individual clasts rest on a surface (e.g. the beach face, or the shoreface), they protrude into the bottom boundary layer and are subject to cyclic loading and shear stresses as waves and tidal currents pass over them (Carter & Orford, 1993; Leeder, 1999). Archaeological material is unlikely to be an exception to this rule unless buried to a sufficient depth in the sediment (see Section 4.2.3.2). Studies of coarse-grained beaches have suggested that the transport of clasts tends to be correlated with wave height, the duration of immersion, and the long axis of the clast. The main observations of these studies now follow:

On flat sand beaches or relatively planar surfaces, individual clasts may move landward by up to 35 metres within a few hours (Carter & Orford, 1993). As individual clasts tend to then collect into surface gravel patches this leads to the possibility that archaeological material of similar size and shape may accumulate in such patches (Carter & Orford, 1993). This results from the differences in critical thresholds (see Section 4.2.2.1) of differently sized particles which in turn partially determines their transport patterns and hence position within stratigraphic sequences.

On beaches, size gradients tend to result from the asymmetry between wave uprush and backwash forces (see Figure 128). This results in preferential deposition of coarse material at the beach crest, relatively fine material in the mid-beach zone and coarser material at the seaward edge of the beach. Although this pattern may be altered by factors such as tidal cycle fluctuations, all gravel beaches will exhibit some sort of cross beach variation (Lee, 2001; Orford et al, 2002).

Particle shape may also exert an influence with disk and blade shapes moving preferentially up the beach, and more rounded forms, such as spheres and rollers moving preferentially down-beach (Lee, 2001; Orford et al, 2002). As before, this is an idealized situation (Figure 130).

On beaches with a significant coarse component, extensive grain to grain collision may occur, especially during plunging wave conditions (Carter & Orford, 1993). By implication, this suggests that archaeological material on such a beach, if exposed to wave action, may be significantly abraded.

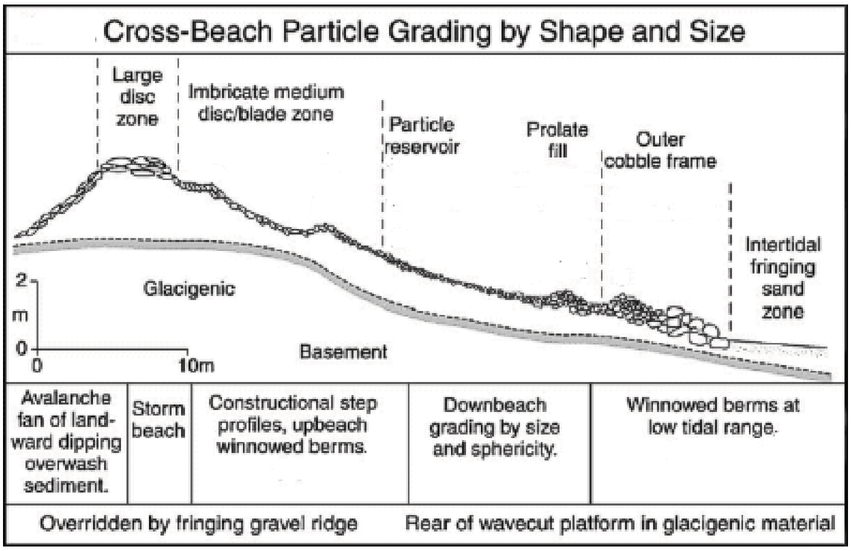

Figure 128. Idealized profile of size and shape sorting of coarse clastic particles (modified from Orford et al, 2002)

Archaeological material may however survive in relatively secure contexts on low energy beaches. In these environments, non-storm significant wave heights are less than 0.25m, beach face widths are less than 20m and significant wave heights during strong onshore winds (< 8 ms^{-1}) are less than 0.5m. This observation has been inferred on the basis that these coasts are characterised by little evidence of cyclic cross shore sediment exchange, and an inability of waves to rework micro-topographic features or beach litter (Jackson et al, 2002). By implication the same should be true of archaeological material. Note for instance that many of the areas highlighted by Flemming (1983:161-162) as being conducive to preservation of archaeological sites were characterised by 'minimal' or 'small' maximum wave heights.

Some ideas have also been advanced as to the behaviour of coarse clasts in the surf zone. In this zone, flow is dominated by breaking waves and consequently there is likely to be a large contribution from oscillatory motions (Nittrouer & Wright, 1994). While extensive research has been conducted into the dynamics of sediment motion under wave conditions, the vast majority of it has focused on fine sediment (Panagiotopoulos et al, 1994; Voropayev et al, 2003). Relatively little consideration though has been given to the movement of larger sediment particles which might more accurately mimic the behaviour of archaeological material barring some recent papers by Voropayev et al (2001; 2003) on cobble sized (6.4 to 25.6 cm diameter) material.

In general, with orbital velocities of up to 0.5 ms^{-1}, a variety of cobble responses are possible, ranging from onshore and offshore movement to steady oscillations with no net movement. The response of the cobble can be related to their size, density, initial position, the slope of the bottom, background flow and friction between the bed and the cobble. For example, where density exceeded 1.5 gcm^{-3}, no movement was observed at all in flume experiments. Complications have however been observed in the zone of breaking waves where in some instances cobbles moved onshore, but in other instances, the same cobble was trapped. This behaviour could not be explained (Voropayev et al, 2001).

Bedforms created by wave oscillations can also modify the transport patterns of individual clasts in the surf zone. For example: the migration of ripples can result in the periodic burial of clasts, effectively trapping them. This is especially the case with larger clasts (c. 20cm diameter, 3kg mass) when they are of comparable size to the ripples, though smaller ones (c. 7 to 10cm, <0.5kg mass) appear to exhibit a net onshore movement (Voropayev et al, 2003).

The cobble responses described above are not exact analogues to the behaviour of archaeological material in the surf zone, however they do provide some indications of the parameters to consider when assessing their reworking. They do highlight the fact that the interaction between the multiple factors affecting sediment movement means that the resulting distribution of particles can be difficult to predict.

A further issue to consider is that waves breaking in the surf zone set up rip and longshore currents which can move sediment parallel to the coastline. These are commonly of the order of 1 ms^{-1} or greater (Nittrouer & Wright, 1994; Open University, 1989). Observations of coarse clastic sediment on beaches have provided the following statements:

The position and form (size and shape) of coarse particles has also been argued to impact on the distance they move longshore. Experimental work suggests that particles closer to the seaward side of the beach tend to move further longshore. This is a logical conclusion since material closer to the seaward margin will be subject to a greater degree of inundation and thus more opportunities to be entrained by longshore currents (Lee, 2001).

Similar conclusions have been made for particle size in that large clasts move greater distances longshore. This seemingly simple relationship though may be modified by wave energy to the extent that it may be accentuated with increasing wave energy, but reversed with decreasing wave energy. This has been related to the sheltering of smaller grains by the larger particles which reduce their longshore velocity (Lee, 2001). However, when wave energy is reduced, larger clasts cannot be entrained, thus even with a reduced velocity smaller grains will move further. In terms of the distances moved the largest movement of gravel that has been recorded was 2km in 30 days (Lee, 2001). However, this is an exceptional result and most rates are an order of magnitude less, while in some instances virtually no movement was noted (Carter & Orford, 1993).

Particle density may also have a role as it has been observed that clasts of lower density tend to have a higher longshore transport velocity than their denser counterparts (Lee, 2001). It should also be noted that when accumulations of material occur, consideration such as group imposed controls (e.g. bed acceptance or rejection, contact stresses) should be brought to bear as these may influence subsequent sorting and entrainment (Carter & Orford, 1993).

The deposition of material entrained by longshore currents occurs where the velocity of the current decreases to the point where it is no longer to exceed the threshold of the material it carries. In many cases this occurs in particular topographic or geomorphic situations. For examples, studies of placer (heavy mineral) deposits have demonstrated that they are often best developed at points of shoreline curvature (e.g. headlands), barriers and also near river mouths where currents slow and wave action diminishes (Browne, 1994; de Meijer et al, 2002).

The above observations have been drawn out of studies which have examined material affected by multiple tidal cycles. However, very rapid events, chiefly storms can also have a significant impact, though studies suggest they often result in the rearrangement of material on a beach rather than the removal of significant quantities offshore. This rearrangement often takes the form of the removal of sediments from the beach crest and their redeposition on lower portions of the beach (Van Wellen et al, 1997). However, other authors suggest they can also cause severe beach and dune erosion, and result in significant offshore sediment transport (Schwarzer et al, 2003). The kind of impact that extreme wave erosion could have on archaeological material is graphically illustrated by the fact the concentrations of gravel, peat lumps and fossil wood from 10 to 25m water depth are often found after storm events on the barrier island beaches of German North Sea coast (Hoselmann & Streif, 2004). In either case (i.e. rearrangement or removal) the spatial and structural integrity of an archaeological site will be lost. In all cases the survival of a site will depend on the unique local interaction of wave action, tidal current, coastal substrate and morphology, weather and nature of the archaeological material itself.

In terms of the impact of transgression on exposed material, it will essentially alter its duration of inundation and thus the time it spends in the swash, surf, breaker and finally offshore zones (see Section 4.2.4). It may thus have an effect on the final sorting pattern of a deposit, though the exact effect is difficult to quantify at this point in time.

There have been some observations on the behaviour of in situ archaeological material exposed on beaches. For example, in Langstone Harbour (southern England), archaeological material ranging from Mesolithic flint implements to Roman pottery has been observed eroding out of low salt marsh cliffs on the edge of the upper foreshore (Figure 129). These have been interpreted as being in primary context when incorporated in the cliff deposit (Stage 1: Figure 129). However, cliff retreat due to wave erosion releases the artefacts onto the upper foreshore. At this point, the finer resolution of the assemblage will have been lost as some minor rearrangement has taken place but the basic horizontal spatial integrity, (though not the vertical stratigraphic relationships), remains (Stage 2: Figure 129). Wave action then results in the winnowing out of the smaller artefacts, and limited movement of the larger pieces (Stage 3: Figure 129). As coastal recession of the saltmarsh continues, the intertidal foreshore migrates shoreward. More rigorous wave action and swash then results in further reworking (Stage 4: Figure 129) with the end result that the artefacts are removed by tidal action, probably into the tidal channels of the harbour (Stage 5: Figure 129 - Allen, 2000).

In general, it seems that artefacts may survive for millennia in primary context in the cliffs. Once released from this context, they have moderate residence time at the foot of the cliffs before they are deposited on the wave cut upper foreshore platform, often with other, similarly sized material. As the foreshore is cut down, the artefacts are rapidly removed from the foreshore in into the tidal channels (Allen, 2000).

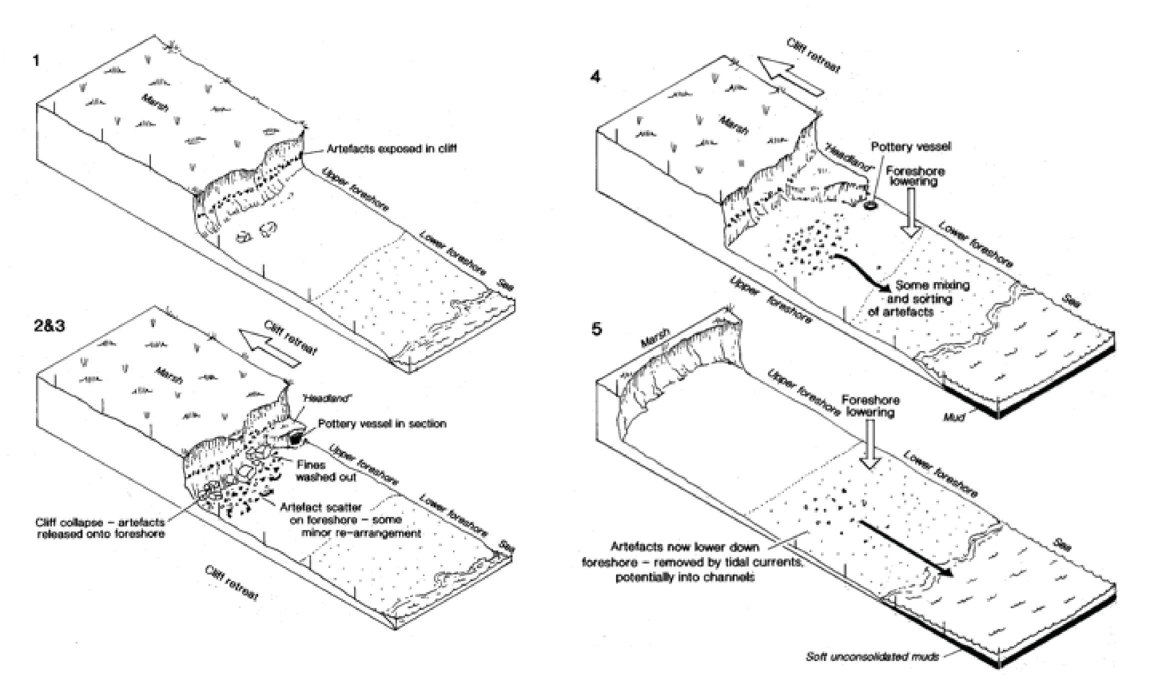

Figure 129. Diagram showing sequence of events in the reworking of primary context assemblages at Langstone Harbour. See above text for sequence of events (from Allen, 2000)

Tidal range in Langstone Harbour is around 3m (Allen, 2000). Consequently it is worth considering that there may be a significant unidirectional contribution to the local sediment transport regime. Experimental work on lithic flake scatters under conditions of unidirectional flow (river system) has demonstrated a pattern, which may also influence the sequence of events depicted in Figure 131. Freshly knapped, or in situ, scatters are characterised by a high concentrations of flakes and debitage within a relatively limited area. Consequently, high levels of interaction between the flakes, could retard transport. Therefore, the first instance of transport is characterised by reduced reworking (flakes transported <10m), but after that, each subsequent transport event increases the spread of the material, thus reducing the interaction between particles, which in turn increases the overall spread of the scatter (flakes transported >80m: Hosfield, 2004).

4.2.3.2 Buried material

In addition to material exposed on the shoreline, we must also consider the effects of transgression on material buried within beach sediment. Indeed one of the main factors when considering the likely preservation of buried deposits, is the depth to which sediment may be mobilized by wave processes in the surf zone. In sedimentological terms this can be related to the 'sediment mixing depth', which is can be thought of as follows:

"The vertical thickness of a layer where active sediment exchange takes place, below which lies an immobile bed" (Ferreira et al, 2000)

The maximum mixing depth is therefore the limit of detectable erosion of the bed and controls the truncation of antecedent sedimentary structures (Sherman et al, 1994), which may contain archaeological material. The mixing depth tends to be controlled by processes that operate over the course of a few hours (Sherman et al, 1994). However, the continual operation of these short term processes on archaeological material over a longer period of time, for instance, from days to years, certainly has the potential to disrupt a primary context assemblage. Indeed, as the previous section (4.2.3.1) has highlighted, movement of coarse clasts in the surf zone can be very rapid, so even a very short term event may have a significant impact on the spatial integrity of a deposit of archaeological material.

Experimental studies have indicated that mixing depth is related to breaking wave heights and foreshore slope, with greater breaking wave heights causing increased mixing depths, and increases in slope also resulting in increased mixing depths. On steeper slopes, plunging and collapsing breakers dominate, thus resulting in higher wave energy dissipation over a relatively restricted cross shore zone, while on gentle slopes spilling breakers dominate. These dissipate the same amount of wave energy, but over a large cross-shore area, thus resulting in reduced mixing depths (Ferreira et al, 2000). Mixing depths in everyday situations tend be of the order of several centimetres to several tens of centimetres depending on breaker height and foreshore slope (Table 14 - Sherman et al, 1994; Ferreira et al, 2000; Lee, 2001).

Breaking wave height (m)	Beach face slope	Min. mixing depth (cm)	Max. mixing depth (cm)
0.37	0.11	10.6	15.0
0.34	0.11	10.6	15.0
0.37	0.11	10.6	15.0
0.64	0.10	9.9	15.0
0.49	0.10	10.3	20.0
0.80	0.14	22.0	25.0
0.49	0.11	10.7	19.0
0.60	0.10	16.0	28.5
0.81	0.10	15.3	34.7
0.61	0.12	14.4	32.0
0.85	0.14	17.2	34.6

Table 14. Experimental results from Portuguese beaches showing wave height, beach slope and maximum and minimum mixing depths (after Ferreira et al, 2000).

The influence of burial depth on the movement of coarse clasts is clearly shown by the fact that particles are 3, 5 and 66 times as mobile at the beach surface than the lowest moving sediment under conditions of high, intermediate and low prevailing wave energy conditions respectively (Lee, 2001). Clearly archaeological material buried below the maximum sediment mixing depth will be unaffected by wave and tidal processes, except under extreme storm wave conditions which may erode sediment to a greater depth.

Finally, the type of sediment in which the archaeological material is buried may also be an important consideration. Peat and associated fine grained inorganics that are buried beneath the oxic zone in particular, (Streif, 2004; Behre, 2004) are regarded as providing excellent protection, both in terms of organic preservation, and maintaining the spatial relationships between artefacts by packing them in fine grained cohesive sediment (Flemming, 2002). Key to the good preservation (certainly of wood and other fibrous material) is the inhibition of macrofaunal and microfaunal species that are so effective at degrading exposed materials. In the near anaerobic conditions of the shallow section, only soft rot and microbial erosion bacteria are a threat but these do tend to be the slowest forms of faunal degradation. This is evidenced by the dramatic preservation in peat and silt of a Bronze Age (c. 4 ka BP) timber circle ('Seahenge') in the intertidal zone on the Norfolk coast (Figure 130), and the presence of what appear to be in situ artefacts eroding out of the peat layer in the vicinity of the circle (Champion, 2000).

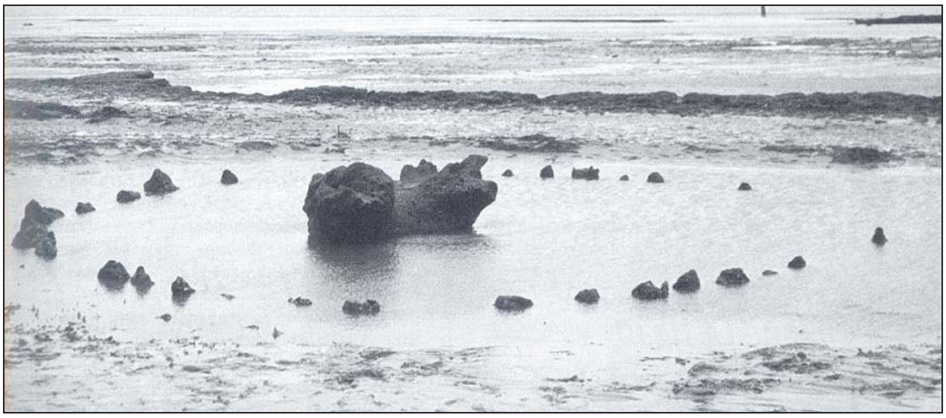

Figure 130. Bronze Age Timber circle ('Seahenge') eroding out of peat deposits on the Norfolk coast (from Champion, 2000)

4.2.4 Archaeological deposits on the shelf

As explained in Sections 4.2.2.1, more energetic waves result in higher bed shear stresses, thus resulting in entrainment from a deeper layer of sediment, in a given area, and also entrain more sediment from deeper waters than less energetic waves (Harris & Wilberg, 2002). For given surface wave conditions, the intensity and magnitude of wave induced bed shear stress diminishes rapidly as water depth increases (Harris & Wiberg, 2002). Thus, a point is reached below which the impact of waves on sediment transport is minimal; this point is known as the 'wave base'. It can be located as much as 100m in depth depending on the period of the wave (Swift & Thorne, 1991; Leeder, 1999), though does not usually extend below 20 or 30m depth (Harris & Wiberg, 2002;). Consequently, bottom boundary layer processes on the mid and outer shelf are dominated by unidirectional tidal and density currents, though storm waves may occasionally induce movement at great depths (Nittrouer & Wright, 1994).

As has been described in Section 4.1.5.4 the bulk of sediment transported on shelves is sand sized. However, where current speeds are strong enough, such as at extreme spring tides, gravel can also be moved (Johnson et al, 1982). Bedforms comprised of this material range from gravel waves ranging from several centimetres to several metres in height, which form when peak near surface mean spring current speed is greater than 0.65 – 0.7 ms^{-1} while current speeds of greater than 1.5 ms^{-1} can create 30m wide by 1m deep furrows in gravel substrates (Belderson et al, 1982; Stride et al, 1982). Furthermore, 1m high gravel waves of about 10m wavelength are known in areas of strong tidal currents and contrast with the lag gravels formed in areas of weaker current due to the winnowing away of the finer fraction. In these areas though, storm waves are capable of forming smaller gravel ripples of 20-25cm in height and 125cm in wavelength (Stride et al, 1982).

Thus, sediment transport on shelves is controlled by grain size, current velocity and depth. The movement of gravel on the shelf also suggests that archaeological deposits are unlikely to be immune from further sorting and disturbance as they move further seaward. The sort of reworking that archaeological material may encounter may be more similar to the unidirectional flows encountered in fluvial environments. Flume experiments by Schick (1986) have shown that below speeds of 0.15 ms^{-1} there in no movement in artefacts greater than 0.5 cm in diameter. At about 0.40 ms^{-1} artefacts of 1-2 cm in diameter move intermittently, while larger pieces make in place stabilizing adjustments. Increasing the speed to 0.6 ms^{-1} results in the movement, be it continuous or intermittent, of all pieces, including those up to 8cm long. Finally speeds of greater than 0.75 ms^{-1} were found to be capable of entraining all artefacts up to 15cm long and moving them continuously or nearly continuously. On this basis of these results, there is no reason to assume that archaeological material will be unaffected by the tidal currents described above.

However, these experiments also demonstrated that the movement of artefacts is not simply controlled by the velocity of the flow and their size and shape. Both Hosfield (2004) and Schick (1986) have postulated that interaction between particles (both natural and man-made) and bed morphology modifies this pattern to the extent that Hosfield (2004) observed that distribution of larger clasts is essentially random, while Schick (1986) noted that it was possible for regular patterns to be distorted. Both studies though, made the point that the movement of material is episodic rather than continuous and phase of burial or trapping may remove individual artefacts from the sedimentary regime, thus modifying the composition of the final deposit.

This observation will almost certainly apply to material on the continental shelf, for the simple reason that sediment transport is not homogenous across its surface as flow regimes vary due to the action of different forces. For example, protrusions on the seabed may redirect flow, or alternatively depressions or channels may act as conduits for particle flow (Nittrouer & Wright, 1994). Consequently certain areas may be surfaces of net erosion and others of net deposition. Indeed depressions and hollows tend to be preferential areas of deposition (Dunbar et al, 1991; Roy et al, 1994).

Consequently, archaeological material on the shelf is not going to exist in a simple system of depth dependent, and size sorting but may occur in patches dependent on localized conditions of deposition and erosion.

4.2.5 Implications for Archaeological Material

The above sections have highlighted the possible processes operating on archaeological material as it undergoes transgression. The impact of these means that the submerged continental shelf is likely to be a palimpsest of preservational variation. The implications are as follows:

- Archaeological material exposed on a beach is likely to be moved about by wave action. Hence sites exposed on a beach surface are unlikely to survive in primary context.
- On the shelf, even at great depth they may be reworked by unidirectional current flows and occasional storm wave action.
- If sites are buried to a sufficient depth of sediment they stand a far better chance of surviving in situ. However, this reduces the possibility of their discovery compared to exposed material. Recently exposed material may also have the advantage that the spatial relationships between artefacts are not too disturbed.
- Small pieces of evidence, such as palaeoenvironmental data (e.g pollen, microfauna), may well be winnowed away even if larger artefacts are only minimally disturbed.
- Secondary and tertiary assemblages are likely to be much more common. There is also a high potential that they will occur as patches of archaeological material sorted by size and type. However, this is not entirely certain given the complex responses of material to bed morphology, other particles and the combined impact of unidirectional and oscillating flows.
- Certain areas, (e.g. hollows or scour pits) may be preferential areas of deposition and trapping, and hence may represent potential areas of investigation.
- There is also a significant possibility that much of this material will be abraded and damaged and by collisions and bomardment with other particles.

Potential areas for future research to investigate could include:

- The archaeological potential of marine secondary contexts and tertiary gravels. Both Schick (1986) and Hosfield (2004) alluded to robust patterns observable in the distribution of secondary contexts material that allowed the investigation of the taphonomic processes affecting them and which may in time allow further interpretations to be made of hominid behaviour. Likewise, Allen (2000) noted that that the patterns seen in the scatters of material on the foreshore at Langstone Harbour could be compared to known primary context assemblages of the same period and the differences between them used to ask questions about 'cultural activity, disposal practices, taphonomy and possible rearrangement by erosion' (2000:198). Similarly, if marine secondary and tertiary contexts are commonplace in the shelf environment, then effective use will have to be made of them. Key to this will be determining research questions to which they are appropriate.
- Targetting site prospection on the basis of preservational condition. This could either take the form of investigation of areas of present erosion on the basis that it is only in these areas that archaeological material will be subject to further disturbance. Alternatively, areas of likely preservation such as deeps and hollows subject to preferential infilling could be targeted.
- Greater quantification and investigation of sediment mixing depths, especially in relation to depth of burial in archaeological sites.
- Experimental work on taphonomic processes operating on archaeological material on beaches and in the surf zone. This follows on from sedimentological

research in coastal environments (e.g. Lee, 2001) and archaeological research in terrestrial environments (e.g. Hosfield, 2004; Schick, 1986) which has been able to demonstrate that robust patterns are deducible from disturbed archaeological material. In addition, a PhD dissertation has recently been submitted which assesses taphonomic processes on the basis of artefact abrasion and damage (Chambers, 2004). Although the full details of this were not available at the time of writing, this type of approach could be valuable in examining the marine reworking of secondary and tertiary contexts.

This preliminary investigation has highlighted that the study of marine taphonomic processes still has a long way to go. At the one end of the spectrum exists a number of specific locales inferred largely in the basis of single sites (Flemming, 1983). At the other end exist studies of coarse grained sediment movement under marine conditions. The next step will bring the two together into an analytical middle ground which may be of use to constructing models of preservational 'hotspots'.