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The origin, classification and modelling of sand banks and ridges

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Abstract

Sand banks and elongated sand ridges occur in many coastal and shelf seas where there is abundant sand and where the currents are strong enough to move sediment, but they have a wide variety of forms. Their generation requires a source of mobile sediment, either from the local sea bed, or from coast erosion. Most appear to have been created during the post-glacial rise in sea level, but they have been subsequently modified by changing currents and waves, thus losing their relict characteristics. A descriptive classification scheme is developed to unify the approaches of marine geologists and physical oceanographers, which emphasizes the formation and present hydrodynamic setting in their long-term development. Open shelf linear ridges (Type 1) are up to 80 km long, average 13 km wide and are tens of metres in height. They are oriented at an angle to the flow, are asymmetrical and appear to migrate in the direction of their steep face. They appear to be in near equilibrium with the flow. These contrast with linear ridges formed in mouths of wide estuaries, which are aligned with the flow, and which migrate away from their steeper face (Type2A). In narrow-mouthed estuaries and inlets, tidal currents are strong only close to the mouth and waves are more dominant. The banks then form close to the mouth as ebb and flood deltas (Type 2Bi). When the coast is retreating, the ebb delta forms a primary source of sand to the nearshore region, which can become modified by storm flows into 'shore attached ridges' at angles to the coastline (Type 2Bii). Tidal eddies produced by headlands can create 'banner banks' (Type 3A), but when the headland is retreating alternating or 'en-echelon' ridges can be formed which can become isolated from the coast as it recedes (Type 3B). Coastal retreat and rising sea level can then cause the ridges to become moribund. Thus the majority of ridges rely on sea level rise for their origin. Theoretical and modelling studies of the shorter term response to present hydrodynamic forcing are generally confined to Types 1 ridges and 3A banks. The most promising work considers the coupled system of hydrodynamics, sediment transport and morphology on Type 1 ridges, and predicts features such as the ridge spacing and angle to the flow. Type 3A sand banks are clearly related to the

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flow patterns produced by the headlands, and the models can reproduce the eddy structures and sand bank extents. Nevertheless, the vital role of shoreline processes has not been fully incorporated into the models, and there is little modelling of ebb tidal deltas or other Type 2 banks. There thus appears to be a wide scope for modelling the generation, evolution and stability of sank banks under the scenario of rising sea level and coastal retreat.

1. Introduction

Sand banks and ridges are significant features on many continental shelves and coastal regions, and their location depends on the presence of tidal or other currents capable of moving the sand, and the availability of sand. Sand banks are likely to occur where the tidal currents are rotary or have low ellipticity. Sand ridges, or linear sand banks, are more likely in areas with rectilinear currents. Amos and King (1984) define ridges as having a length/width ratio exceeding 40. Here we will refer to them in general as banks, unless they are specifically elongate ridges. Banks often store large quantities of the available sand, and appear to be hydraulically maintained sand traps of a high order of efficiency. Generally banks are formed from medium or coarse sand, but when the tidal currents are sufficiently strong and there is a source of gravel, gravel banks of a very similar form can be found. Since banks are in areas with currents strong enough to move sand, there is the implication that there must be a circulation of sediment around the bank that ensures that it does not get more widely dispersed.

There is considerable confusion in the terminology of sand banks, because many have been described in terms of their present day morphology and hydrodynamic context without consideration of their geological origin and development, or their relationship with features described elsewhere. This distinction has been emphasized by Swift et al. (1991), who consider two approaches: a dispersal systems approach whereby the regional context of the ridge is considered within a sand transport system; and the hydrodynamic approach dealing with the local short-term processes.

Sand banks are present in a wide range of water depths. They are found in the mouths of estuaries, adjacent to headlands and beaches, on the exposed shelf, and close to the continental shelf edge. There is evidence that there are active and moribund banks, the active ones being in shallower water, implying that they are ideally formed and preserved under conditions of rising sea level. Indeed it is possible that rising sea level may be essential for sand bank formation. It is also possible that the processes that created the banks may be different from those maintaining them under present conditions. Theories for their origin and maintenance generally fall into two broad categories: that they are relict features created during the post-glacial sea level rise, and that they are formed as a response to the hydrodynamic and sediment regimes similar to those presently active. Both of these approaches are likely to apply to some banks, as they may have been formed during the transgression, but may have undergone active modification and have lost many of the features indicative of their origin. The elongate form of sand ridges may be an indication of the magnitude of modification by the present day hydrodynamics, as well as the source strength. Consequently, it is necessary to consider their origins and development separately from their present form. Despite their variety of situations, there are many common features that suggest there are generic relationships between different banks. A classification scheme will therefore provide a necessary context for developing predictions of long-term morphological change. An understanding of the hydrodynamical context and the present day sediment transport regime will allow prediction of short-term change.

The sand may be produced either from erosion of the sea bed, or from coastal erosion, and is transported both as bedload and as suspended load, the latter especially when waves are present. Where the coastal erosion is linked to rapid coastal recession, the bank can become isolated from its source of sediment, exposing it to a different flow regime and drastic modification as it moves towards a new equilibrium morphology. The change may eventually reduce the sand movement to the extent that it becomes moribund.

In order to produce a regional accumulation of sand there has to be a gradient in the transport rate of the sediment so that more sand is being transported into the area than is being transported out of it. The most common situations that lead to accumulation of sand are likely to be: a reversal in the sand transport direction, involving a bed-load convergence; or a reduction in the bed shear stress along a transport path until the threshold of motion is reached. Localised processes or instabilities are then required to gather the sediment into the ridges, help them to grow, and sustain them. One fundamental process appears to be the presence of mutually evasive ebb-flood channels. These cause a circulatory movement of sand over and around the bank which helps to maintain a quasi-stability. Regional and local processes may be different, and may act over different time as well as space scales. For instance, waves are likely to be an important local process, whereas they may not be on the regional scale. Nevertheless, it is obvious that the interaction between the shape of the banks and the hydrodynamics must lead to a near equilibrium configuration in order for the banks to be stable and self-sustaining when the sand is still mobile. Because of the large variability in the sand supply, and in the storm induced wave and surge currents, it will be very difficult in most cases to define how close to equilibrium the bank may be.

Sand banks are of considerable economic importance. Those close to the shore may provide a source for sediment, interacting with that on the beaches, and helping to stabilize them. Wave dissipation by and refraction around banks may be crucial in maintaining the form of the coastline and protecting some stretches from erosion in storms. Additionally banks may form hazards to shipping, and may contain exploitable reserves of sand and gravel, providing that there are no environmental consequences of extraction. Sand banks are also areas of great importance to the health of the fishing industry, as the banks are important nursery and feeding grounds for many fish species. In the geological column sand banks form potential oil and gas reservoirs. Consequently, there are important benefits from a better understanding of the processes of formation, growth and maintenance of sand banks, and how they interact with the coastline.

The aim of this review is to propose a classification scheme for sand banks, and to present evidence from the literature of the characteristics of the different types, and qualitative description of the processes forming them. This will define the long-term bank development sequence. Theories of their sedimentary hydrodynamics will be separately examined, and a number of particular cases defined for use as examples for the development and testing of mathematical models, to aid quantification of the processes and the prediction of their short-term effects.

2. Classification

There are a number of different types of bank that have been defined in terms of their location and morphology. Swift (1975) suggested that the tidal ridge fields described by Off (1963) fell into three categories. The first consisted of ridges in embayments, or the mouths of embayments, with orientations parallel to the axis of the embayment. The second occurred off capes and promontories, and tended to be coast parallel. A third was also suggested as occurring on shelf edges. Swift (1985) separated storm and tide ridges and, within each category made a distinction between those formed on marine erosion surfaces and those formed by erosional shoreface retreat. He also distinquished ridges created by ebb-flood channel interactions, and by rotary tidal currents in the lee of headlands and on the open shelf.

Amos and King (1984) suggest that there are two generic types of ridge: tidal and storm generated. They also suggest that there are three modes in ridge height; 7 m for storm ridges, 13 m for tidal current ridges, and 27 m for submerged barrier islands. Belderson (1986) agreed that there are significant morphological differences that suggest a generic difference between tide dominated sand banks, and the storm dominated sand ridges typical of the eastern American coast.

Pattiaratchi and Collins (1987) have proposed a broad grouping of sand banks into six categories. They distinguish: fields of sandbanks off open coastlines, and either parallel to the coast and oblique to the tidal current direction, or oblique to the coast; banks adjacent to convergent coasts, or straits; banks perpendicular to embayment heads or delta fronts; and isolated banks associated with coastline irregularities, or with spits. This grouping depends largely on the bank orientation relative to the coastline, but no implications are suggested for their origin.

These suggestions of regional topographical and dominant hydrodynamic forces need to be considered within a dispersal system (Swift et al., 1991) in the light of the sediment sources. The sources are either the seabed or the coastline; the strength of the latter depending crucially on the rate of retreat of the shoreface. Modification by hydrodynamics may lead to a common form. The resulting classification proposes a generic relationship between sand banks in the light of their origin and development. The different types and their characteristics will be described, and the developmental sequence will be proposed. This will be used as a context for quantitative models of their equilibrium characteristics and movement.

The classification comprises:

Type 1: Open Shelf Ridges.
Type 2: Estuary mouth
 (A) Ridges (wide mouth)
 (B) Tidal deltas (narrow mouth)
  (i) Without recession (ebb tidal deltas)
  (ii) With recession (Shoreface connected ridges)
Type 3: Headland associated banks
 (A) Banner Banks (non-recessional headland)
 (B) Alternating Ridges (recessional headland)

Most types appear to have moribund as well as active types.

Sand banks are common in the North Sea and around the UK coast, and much of the original description of tide dominated sand bank characteristics in the literature is derived from them. Fig. 1 shows their distribution and gives a categorization of them according to the proposed classification.

2.1. *Type 1*: *Open shelf ridges*

Nearly all shallow tidal seas, where currents exceed about $0.5\ \mathrm{m\,s^{-1}}$ and where sand is present, have ridges. They can be up to 80 km long, and typically average 13 km width and tens of metres in height. Their spacing tends to be proportional to their widths. The characteristics and distribution of ridges was first examined by Off (1963), who thought that the ridges were formed parallel to the tidal currents, and who showed that their spacing increased with water depth in the channels between them. The ridges are asymmetrical, having slopes of about 6° on the steeper side, and less than 1° on the gentler side. It is thought that they gradually move in the direction of the steeper side. The bank crests are flat in shallow water, but are sharp when water depth is large enough to limit wave effects. The channels between the banks are called swales, or sinuses when they end as a *cul de sac* (Ludwick, 1974).

Sediment movement may create open shelf ridges in one of three ways. These have been stated by Caston (1981) as: ridges are created by an excess of sand supply, with their growth depending on a greater supply being added to the head than being lost from the tail; ridges are a remnant of a larger deposit which is being whittled away by selective transport; or they are an expression of an equilibrium with the ridges maintained within an active sand transport path. Thus ridges on the open shelf can be created by remodelling of an extensive sand deposit (e.g. Berné et al., in press).

Some ridges appear to have a relict core of material that may be the remnant of transgressive conditions (Houbolt, 1968; Berné et al., 1994). Smith (1988a) has suggested that the Sandettie (51°15′N, 2°E) and South Falls (51°30′N, 1°45′E) Banks were formed with their southern ends attached to the ends of a gravel ridge oblique to the tidal flow through the Dover Straits. Consequently, the gap between them is ebb dominated and there is flood domination to west and east of the ridges. The South Falls Bank appears now to be detached from the ridge.

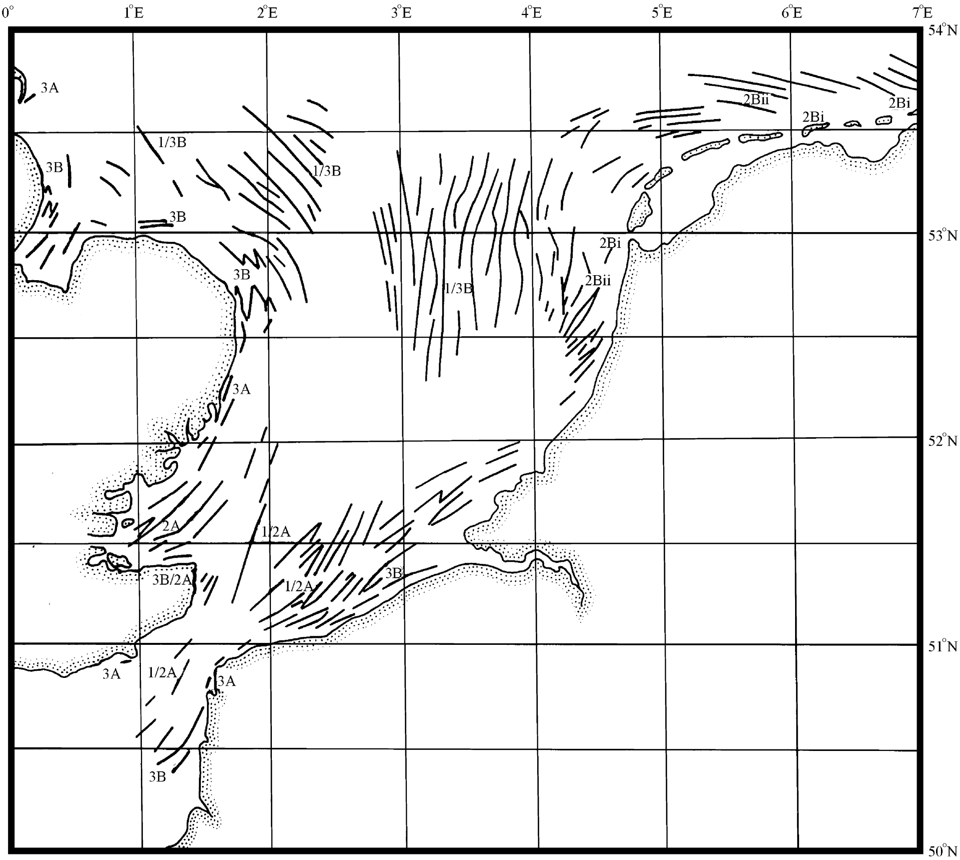

Fig. 1. Map of sand bank distribution in the North Sea and Straits of Dover showing classification and original type where appropriate.

Consequently, it is possible that open shelf linear ridges may express the equilibrium state towards which all banks progress, independent of their origin, once they have been separated from their source conditions.

There is coincidence in location between linear ridges and convergence in the bed load transport paths. There is normally an asymmetry in the current strengths on either side of the banks with maximum currents being in the ebb direction on one side and the flood direction on the other, causing sediment accumulation; this is discussed further in Section 3.

Bed load convergences are the result of steep horizontal gradients in tidal phase. This effect has been examined in hydrodynamic models. Pingree and Griffiths (1979) have modelled the distribution of shear stress around the British Isles for the interacting M_2 and M_4 tidal components, and the convergence zones agree fairly well with the distribution of ridges.

The ridges are generally straight or only slightly sinuous crested. They are often relatively broad and flat at one end, and narrower and pointed at the other. Caston (1981) has examined the morphology of the series of ridges that are central to the southern North Sea and are directed towards the Dover Straits (see Fig. 1). From the north these are the Gabbard Bank (51°55′N, 2°E), North and South Falls and the Sandettie Bank. These ridges rise to within a few metres of the surface, and are asymmetrical in cross section with maximum bed gradients of about 3°. She found no evidence for sand circulation round them, though sand waves faced in opposite directions on either side of the ridges, and she suggested that the broad end should be considered to be the upstream end as far as the sand transport is concerned, with sand joining the ridge at the broad end, and leaving it at the narrow end. The overall sand transport direction is consequently towards the south. However, this conclusion is at variance with that of Smith (1988a).

The long axes of most ridges are oriented at a small oblique angle to the peak tidal flow direction, generally in an anticlockwise sense in the northern hemisphere. This suggests that Coriolis force may be a significant factor in the dynamics of the banks. The Sandettie Bank seems to be the major exception, having a clockwise orientation. However, there is as yet no clear evidence as to whether the dominant orientation is reversed in the southern hemisphere. The usual angle of deflection is between 7 and 15°, but with extreme values ranging from 0° to more than 20° (Kenyon et al., 1981). This tends to make one side of the ridge more exposed to the flood flow, and the other more exposed to the ebb, thus helping to produce flood and ebb dominance on alternate sides.

Sand waves occur on active ridges, and the sand wave asymmetry is generally in different directions on either side of the ridge as a result of the differences in ebb and flood current dominance. At each end there is normally a zone of symmetrical sand waves. The sand waves are active indications of the sediment movement and they show a gradually changing orientation towards the ridge crest, indicating a substantial component of sediment transport across the ridge axis (Houbolt, 1968; McCave and Langhorne, 1982). Caston (1972) showed that sand wave orientation also varies, with angles of 40° between their crest alignment and the ridge axis on the gentle slope side, and 30° on the steep one. This implies that there is a convergence of sand at the crest of the bank. At the ends of the ridges, in the zone of symmetrical sand waves, the superimposed megaripples trend at very high angles, 55° to the sand wave crests according to McCave and Langhorne (1982), indicating transport effectively along the troughs of the sand waves and almost normal to the tidal currents. The obvious conclusion is that sediment circulates both round and over ridges and this process helps to maintain them (Smith, 1969).

From repeat hydrographic surveys Smith (1988a) has suggested that there have been changes in the volume of sand in the tail of the South Falls Bank resulting from a decrease in the sand circulation on the ridge itself, and a decrease in the supply coming from the south–southwest. He considers the ridge to be an example of a system adjusting to a new balance between sand supply and the hydrodynamics.

Ridges generally stand on a platform of less mobile gravel or rock. Their internal structure has also been revealed by geophysical surveys. Some are complex, implying erosion of older material. Others, such as the East Anglian ridges, show internal reflectors dipping at approximately the same angle as the steep bank face, and approximately parallel to it, confirming the conceptual model of Houbolt (1968), that the ridges gradually progress in the direction of the steep slope. Sand brought across the crest of the ridge tends to build the steep slope gradually outwards. Since the crests of the ridges are often at or a few metres below low tide mark, wave processes must be important in enhancing the sediment transport across the crest and limiting its vertical growth. The grain size is normally finest and best sorted at the crest, though there are exceptions to this that include Middelkerke Bank (51°10′N, 2°E) (Trentesaux et al., 1994). This may be the result of wave winnowing.

2.2. *Type 2*: *Estuary mouth ridges*

Harris (1988) in a review of large-scale bedforms has concluded that in general linear sand ridges are associated with the mouths of macro-tidal estuaries, and tidal deltas are associated with meso-tidal or micro-tidal estuaries. This agrees with the suggestions of Hayes (1975). However, Harris also points out that it is the tidal prism volume in relation to the cross sectional area of the mouth that is important, and he distinguishes a mouth width of about 10 km as separating the two types, with small widths having tidal delta deposits. Clearly, sediment availability must also be an important consideration. Hard rocky coasts do not provide sufficient sediment at the coast to generate spits to constrict the mouth and act as a sediment source. Wide meso-tidal estuaries would be unlikely to have high enough currents to move the sand, and for narrow mouth estuaries transport and deposition would be limited to a small region close to the mouth. In sediment starved environments, such as the Bristol Channel, linear ridges are only formed in association with headlands, or islands, where significant shear stress gradients and separation effects are created in the ebb and flood currents (Harris, 1988).

2.2.1. *Type 2A*: *ridges* (*wide mouth*)

Macrotidal estuaries characteristically have a rapid topographic convergence towards their heads. This creates an increasing tidal amplitude, and bed friction causes a time delay in the progression of the tide along the estuary. The tidal currents increase towards the head of the estuary with considerable asymmetry during the tide. At the mouth flood currents are greater than ebb currents in the channels, whereas ebb currents are dominant in shallow areas. There is a gradient of sand transport into the estuary, and the sand drawn into the estuary mouth increases the area of the intertidal flats. This will affect the tidal asymmetry in the currents which depends upon the changing shape of the cross section through which the water flows to fill the estuary above, the incremental volumes needed to raise the water level, and the varying friction. Towards the head of the estuary, where the intertidal areas become extensive, they are flood dominated, but in the channels ebb dominant currents develop (Fig. 2). There is a tendency for the flood dominated channels near the mouth of the estuary to avoid the ebb channels further up, leading to a bed load convergence. Banks form between them, and horizontal water and sediment circulations are

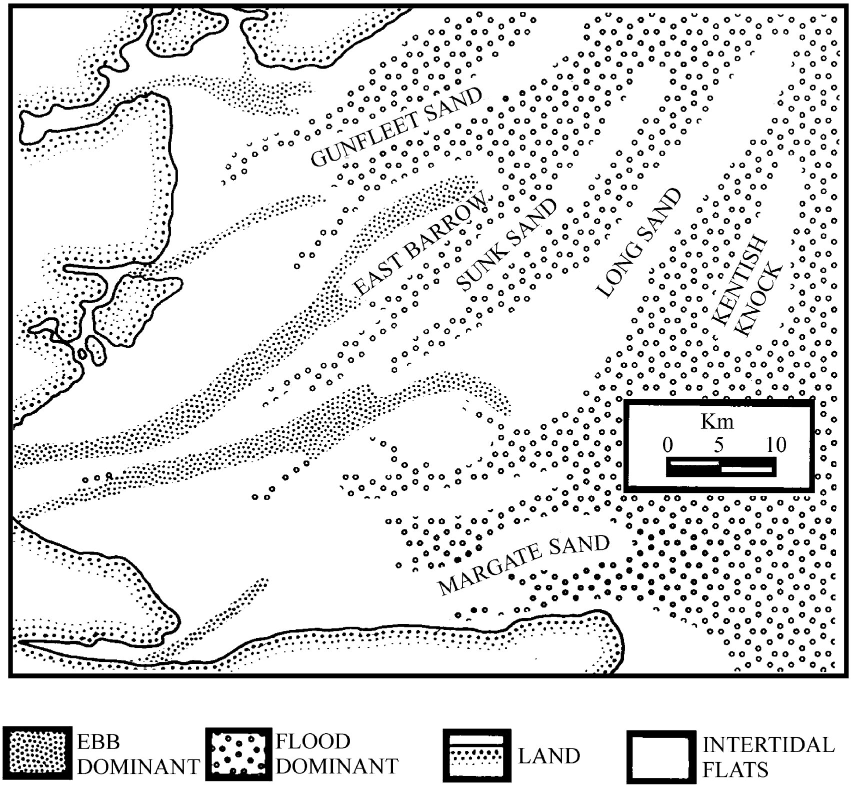

Fig. 2. The ebb and flood channels and associated banks in the Thames estuary (after Robinson, 1956).

created. Because of the lateral constraints, the banks are generally aligned with the water flow, rather than being at an angle. This process has been described for the Chesapeake Bay mouth by Ludwick (1974,1975) where the constriction locally generates tidal currents strong enough to move sand. The ridges are more or less aligned with the ebb transport streamlines, but are oblique to the flood streamlines. The flood streamlines suggest lines of convergence and divergence which alternate. The ridges are located between the mutually evasive ebb and flood dominated channels. Similar processes are likely in tidal straits.

In the Ord River Estuary in Australia, Wright et al. (1975) have shown that further upstream, because of the increasing tidal asymmetry, an increasing fraction of the ebb flow becomes concentrated in a decreasing channel cross section, causing an increase in ebb velocities in the deepest parts of the channel. The shallower parts of the cross section are flood dominated. Thus banks form at the edges of the channels.

The evolution of ridges as sea level rose in Delaware Bay have been examined by Kraft et al. (1974) and by Weil et al. (1974). As sea level rose and the volume of tidal water increased, channels were deepened, widened and eroded headward. The mud was removed in suspension and the sand was deposited as a linear bank on the channel margin, with the steep slope into the channel, and the gentler slope on the flanks. With widening of the channels, growth of the features has been from the steep side with migration towards the gentler side. There were significant variations in the intensity and direction of the currents with depth, dominant flood flow occurring in the channel, and ebb flow on the flank. It was suggested that secondary currents played a major role in causing flows up the steeper sides of the channel which eroded sediment, and deposited it outside the channel.

Thus, the two major differences between estuarine and offshore ridges is that the former are aligned with the maximum flow and, whereas the latter migrate in the direction of their steepest slope, the estuarine ridges migrate away from their steep faces. This is because in the estuarine case, the steep face in the channel is being actively eroded by the increasing discharge produced by growth in the tidal prism as sea level rises. The migration of one ridge produces interaction with the next, and there is a complex, often quasi-cyclical movement of the ridges and channels. An example of this is shown by studies on the Largo Bank in the Bahia Blanca estuary, Argentina, by Gomez and Perillo (1992), which showed ridge migration of 37 m per year.

Harris (1988) proposed a developmental sequence for estuary ridges. In the first stage, intertidal flats are created inshore, and the offshore is scoured clean of sediment. The first sand received will accumulate in patches, in association with headlands or other obstructions. As more sand becomes available the wide areas in the centre of the channel develops sand ribbons, and the intertidal areas extend to eventually occupy the whole estuary mouth. At this stage the first *en echelon* ridges occur, and evolve in such a way that mutually evasive ebb and flood channels will form. In the final stages the intertidal flats merge with the ridges, leaving only a few isolated channels. The infilling replaces much of the active volume of water in the estuary with sediment. Harris uses the Bristol Channel, Moreton Bay (Australia) and the Thames Estuary as examples of stages on the sequence. However, his analysis does not fully account for the tidal conditions within the estuary which produce the infilling, or the changes brought about by the infilling, which alter the flood or ebb dominance, and the process of attainment of an equilibrium condition.

Harris et al. (1992) have described a set of ridges in Moreton Bay on the coast of Australia, which though mesotidal has a mouth width of 14.5 km and a large tidal prism. The ridges separated mutually evasive ebb and flood channels, and were formed from sand coming from the adjacent coastline. Coring and dating showed that the ridges furthest from the shore were formed prior to 3000 yr BP, and have not been rebuilt since that time. Those closer inshore, however, are modern and undergo regular rebuilding. Harris and Jones (1988) used a sequence of aerial photographs taken over 26 yr to study the ridge movement. They found that: some migrated while others did not; migration of the crests was non-uniform, causing distortion into complex S-shapes; and two areas of rapid change corresponded with centres of sand deposition.

Examples of type 2A ridges include Long Sand, Sunk Sand and Gunfleet Sand, in the Thames Estuary (51°30–40′N, 1°10-30′E) (Fig. 2), and the banks in The Wash

(52°50′N, 0°10–30′E). It is likely that the various banks around the Dover Straits may be of the same origin.

An interesting example of ridges which have the same generation mechanism as those in an estuary mouth is shown by ridges in the south western Yellow Sea (Liu et al., 1989). These ridges spread radially from the coast, with individual ridges up to 100 km long, and 5–10 km wide. They exist from the low tide line to a depth of 20–30 m. There is a large quantity of sediment available, as it is an area in front of an abandoned mouth of the Yangtze River. The M_2 co-phase lines close at the coast, with the greatest tidal amplitude at the coast, as would occur for an estuary mouth. There is tidal asymmetry, with the flood current being greatest, and the currents are also radial. The ridges are destroyed by typhoons, but are actively rebuilt, though it is not stated whether they reform in the same places.

2.2.2. Type 2B: *tidal deltas* (*narrow mouth*)

Tidal deltas are formed at the narrow mouths of meso and microtidal estuaries, and the inlets through barrier islands, when there is an abundance of sand or gravel available. They are formed in pairs with ebb tidal deltas formed seaward of the mouth, and flood deltas formed landward of the mouth. The ebb tidal delta is shaped by the ebb currents issuing from the mouth, but it is modified by wave action. It relies on the sand being transported seaward from the mouth and deposited when the flow diverges. The source is the littoral transport that carries sand under wave action along the beach towards the spits that constrict the mouth. Some of this sand is recycled to the beach from the ebb delta by flood currents combined with waves. The complementary flood delta is formed as a mirror image on the inner side of the mouth, the main difference being that wave influence is less. It has been recognized (Swift et al., 1991) that the tidal inlets are a primary source for nearshore sand in areas such as the eastern Atlantic seaboard of America. Their contribution to the shelf will be linked to the rate of shoreface retreat. An important feature of the tidal delta is the deep trench formed in the mouth which can persist on the shelf after coastal retreat. There is a relationship between the volume of the sand in the ebb delta and the tidal prism of the estuary, which implies that there is a balance between the amount of sand released by coast erosion, and the fractions of that supply transported into the estuary and deposited, thus modifying the tidal prism, and that transported out of the estuary. Dean and Walton (1975) calculated that the ebb tidal deltas off Florida contain the equivalent volume to 76 years of beach retreat at the average rate.

(i) *Ebb tidal deltas*: The accepted model for ebb delta morphology is that of Hayes (1975). Sand transport in the main channel is dominated by ebb currents and the channel margins are fringed by linear bars, even when the inlet has higher flood velocities. There is a large terminal lobe at the outer end of the ebb channel where the expansion of the currents creates an area of deposition. This is the major part of the ebb tidal delta. The outer limit of the lobe generally becomes the tip of the salt wedge during ebb tides, and often also the tip during high river discharge. The crest of the shoal thus becomes a bottom current convergence zone at times of maximum sand transport.

On the edges of the terminal lobe are swash bars which are mainly subject to wave action and progress landwards across the platform. The ends of the barrier beaches are formed from these landward migrating bar complexes. Between the terminal lobe and the beaches, marginal flood channels occur within which the currents are flood dominated, and where wave-induced longshore currents are important. The presence of the bank, and also of the jet-like ebb current, are likely to refract the incoming waves, and this has the effect of locally enhancing the sediment transport towards the mouth (e.g. Fitzgerald, 1984). On the down-drift side this may lead to an area of erosion which can cause the inlet mouth to gradually progress in that direction, even though a certain amount of sand 'by-passing' occurs through the ebb delta. Also the presence of the shoal is important in modifying the tidal wave amplitude and phase as it travels into the estuary. Since this in turn has an effect on the currents, there are possibilities of a feedback between the processes of morphological modification, leading to an equilibrium.

Smith and Fitzgerald (1994) have carried out calculations of the sand transport rates around the Essex River Inlet ebb tidal delta. They show that the main ebb channel can easily remove the sand supplied by the marginal flood channels, and this implies an order of magnitude more sand is moved across the swash platform than is carried by migration of the swash bars. They conclude that far more sand is circulated within the sediment gyres than bypasses the inlet.

An example of the sand transport on an ebb tidal delta is shown for the Texel Inlet (53°N, 4°45′E) (Sha, 1989) in Fig. 3. Further examples occur at the other Wadden Sea inlets, and on the eastern seaboard of America (Boothroyd, 1978).

The ebb delta system need not be symmetrical. It is probable that the channel will be oblique to the coast when there is a dominant direction for the sediment supply, or a dominant wave direction, or when there is a phase lag between the ebb currents and the coastal current. The latter means that the current issuing from the mouth is deflected along the coast by the offshore current. Fitzgerald (1984) found cycles from 7–42 years associated with changes in the shoreline alignment and the orientation of the main channel. When the ebb delta is skewed updrift there is accretion updrift and beach progradation, coupled to erosion downdrift. This period is followed by a rotation of the channel and the delta, giving progradation downdrift and erosion updrift.

In Teignmouth Harbour (52°32′N, 3°28′W), studies have shown that there is a cyclical but erratic movement of the banks associated with the ebb delta. The cycle involves growth of the banks on the southern side of the estuary mouth, their eventual detachment and migration towards the northern side, and movement onshore. The sediment then travels towards the head of the spit and gets swept seawards onto the southern bank. This cycle is variable in length, depending largely on the wave conditions, but lasts 3–5 years with minor cycles of about a year (Robinson, 1975). It is thought that this cyclicity is a consequence of the cell of movement being almost closed.

Fitzgerald (1984) has shown that the Price Inlet in S. Carolina goes through cycles of growth and decay which cause 15–20% changes in its volume. These lasted 4–7 years. The growth phase is a coalescence of landward migrating swash bars. This

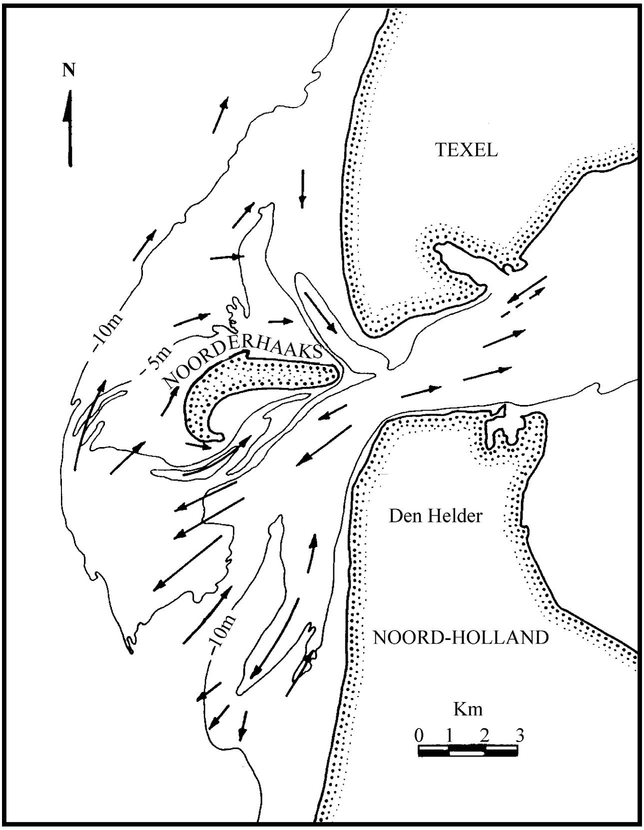

Fig. 3. The ebb tidal delta at the mouth of the Texel Inlet. The arrows show the dominant flow directions (from Sha, 1989).

refracts the waves and reverses the longshore transport direction on the downdrift shoreline so that little sediment escapes. When the bar complex reaches the shore, the sand trapping mechanism ceases and the delta reduces in volume. The processes of delta build-up then recommence. Smith and Fitzgerald (1994) found similar timescales for sand cycling in the Essex River inlet in Massachusetts.

(ii) *Shoreface connected ridges*: In areas where there is significant barrier beach recession, combined with migration of active inlets in a longshore direction, shoreface connected ridges can develop as products of the ebb tidal deltas. These and associated features are also termed shoreface attached ridges, shoreface detached ridges, storm-generated sand ridges and linear shoals. They are well developed on the Delmarva coast of the USA (Swift et al., 1973) and also occur off the coasts of Argentina (Parker et al., 1982) and the Netherlands (Swift et al., 1978). On these coasts storm effects can be more important than tides. The sand ridges are up to 10 m high, 2–5 km apart, and their crestlines can extend for tens of kilometres. Side slopes are rarely more than 1°, and the banks typically converge with the coast at angles of 25–35° along a trend that is intermediate between the dominant direction of storm wave approach and the coast parallel trend of storm currents. They are asymmetrical at their shorewards ends, with the steeper face on the landward side. The maximum side slopes decrease off shore, the offshore slope becomes the steeper one, and the cross sectional area of the banks increases. The ridges appear to build to about a third of the water depth and the troughs are likewise excavated to a third of the water depth below the mean sea bed level. Sands are coarsest in the landward trough and become finer up the landward flank to the crest, where they are well sorted, and down the seaward flank where they become increasingly fine. Thus the grain size is 90° out of phase with the topography, and this is thought to indicate erosion of the landward flank, and deposition on the seaward slope, with seaward migration of the crest (Swift and Field, 1981). The seaward migration implies an elongation of the bank and source of material at the coast. The coastal retreat and the longshore component of the ridge movement are about the same, and are in the order of 1 m yr^{-1}. Some banks are detached, in that they do not connect to the coastline, but they appear to have the same characteristics and movement as the attached ones.

Theories for their origin fall into two broad categories, the first being that they are relict features produced during the sea level rise. The second is that they are dynamic features created by modern shelf processes. Duane et al. (1972) consider three alternative hypotheses for their formation; that they are raised Pleistocene beach ridges; that they are ridges marking still-stands in the post-glacial transgression: that they are formed by the present day hydraulic regime at the shoreface. They concluded that the banks appeared to be caused by interaction of a dominantly southward transport with an intensified coastal jet during periods of strong north easterly winds. The jet intensifies towards the head of the trough which tends to promote erosion of the coast and widening of the trough until it eventually breaks through at the coast and the processes become re-established further down the coast. As the banks appear to move down coast and offshore as the shoreface retreats, this does not explain their origin, even though it may explain their maintenance and movement. Field (1980) concluded that ridges were not relict, and Swift and Field (1981) looked for generating zones on the retreating shoreface. Dolan et al. (1979) analysed shoreline periodicities in rate of change, but these did not agree with the linear shoal periodicities, nor with edge waves (Swift, 1980).

The importance of storm induced flows in the ridge movement has been emphasised by Swift et al. (1978) through a comparison between ridges in the mid-Atlantic Bight, off the east coast of South America, and off the Netherlands–North German coast. In all cases the opening angle between the coast and the ridges was about 20–50° and faced the strongest prevailing wind, and the topography and grain size distributions were very similar.

It is apparent that, though other origins are possible, with active shoreface retreat the ebb tidal deltas are the main sand source, and these features become modified into ridges by storm induced flows. A model for their origin as a combination of ebb tidal delta migration and beach retreat has been proposed by McBride and Moslow (1991). They examined the spatial distribution of active and historical (closed) inlets on the Atlantic coast of the USA, and the distribution of ridges. Both inlets and ridges are most numerous on wave dominated barrier beaches. They consider that the majority of these ridges are formed by a two-step process: (1) the sediment is deposited as ebb-tidal deltas at inlets in the barrier beaches, and (2) the sediment is reworked by shelf processes during further transgression. The ebb tidal deltas form a point source of sediment so that as the coast retreats, the inlet migrates along the coast, and the ridge is formed at an angle to the coast, the angle being related to the relative rates of movement (Fig. 4). Stable inlets do not appear to be associated with attached ridges. They found that the average angle for the Atlantic inner shelf of the USA was 29° where the beach retreat rate was 1.5–5.0 m yr^{-1}. Since many inlets become closed by storms and open elsewhere, ridges can become detached as their source is cut off, and there are likely to be more ridges on the shelf than present day inlets. They suggest a six stage conceptual model for the ridge origin and evolution, involving inlet opening, inlet migration, waning and closing of the inlet, and ridge detachment following barrier migration. However, ridge migration is not accounted for in this model. Field (1980) found that ridges migrated southwards at 2–120 m yr^{-1} and that inlets leave back-filled scour trenches rather than ridges.

Nevertheless, this origin appears to be qualitatively satisfactory for other areas of ridges (e.g. Snedden et al., 1994). Though the barrier islands of the north coast of the

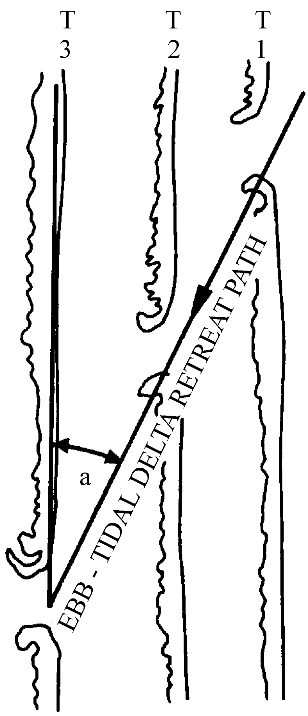

Fig. 4. Diagrammatic representation of the path of ebb tidal delta retreat following beach recession, and leading to the generation of a shore attached type 2B(ii) bank (after McBride and Moslow, 1991).

Netherlands and Germany (Fig. 1) have retreated significantly and the inlets have migrated, the ridges do not connect directly to the beach, but disappear into the shoreface sand sheet, presumably because of the higher tidal current and wave regime, and a reduction in shoreface retreat rate. The Danish coast, being 90° to the prevailing winds, has banks normal to the coastline, as predicted by this model.

Van de Meene (1994) described a series of shore face connected ridges on the western Dutch coast (52°10–50′N, 4°10–30′E). Though they are oblique to the coast they are of subdued relief, only being 1–6 m height and with slopes of 1 : 200 to 1 : 400. No definite conclusions was made of their origin, but it is possible that they are the modified relicts of shoreface attached banks created at the western entrance of the Isselmeer which formerly entered the North Sea near Ijmuiden (Beets et al., 1981), as a series of temporary inlets.

2.3. Type 3: Headland banks

On an actively eroding coast each headland, or change in coastline orientation, implies a change in the rate of longshore transport of sand. Where there is also a high variance in the direction of wave approach, the longshore transport may be in opposite directions on either side of the headland, causing a littoral transport convergence at that point. The resulting surplus of sand creates a coastal shoal, the development of which appears to depend on the rate of coastline retreat. With very slow retreat the surplus sand will be swept offshore to accumulate as a banner bank in a position of convergence. With coastline retreat, a series of alternating banks will result with each successive one more distant from the shoreline and the sediment source. As the banks grow and interact with the flow field, they become modified into elongated ridges.

2.3.1. Type 3A: Banner banks

In areas where the coast changes direction at an acute angle and where the headland is resistant to erosion, sand banks are likely to appear on one side or other of the headland. These are often called ‘Banner Banks’. In some cases a bank occurs on both sides of the promontory. The presence of the banks also appears to be aided by a fairly rapid deepening of water away from the coast; they are less evident off coasts with a low offshore slope. The banks are only a few kilometres in size and with an elongated pear-shaped form, the broad end being directed towards the tip of the headland. The shallower part is nearest the headland and towards the pointed end the crestal depth gradually increases. The side of the bank facing the sea has a steeper slope than that on the landward side, and the bank is separated from the headland by a deep narrow channel. Within this channel there is a very marked asymmetry of the tidal current giving a residual flow towards the headland. On the outer side of the bank there is an asymmetry in the opposite sense and a residual away from the headland. Consequently an eddy in the residual water circulation is inferred. Fig. 5 shows the flow around Start Point, South Devon and over the Skerries Bank formed in its lee. The stream-lines of the ebb and flood currents are not coincident, and the residual flow shows an eddy, with the line dividing ebb and flood dominance lying

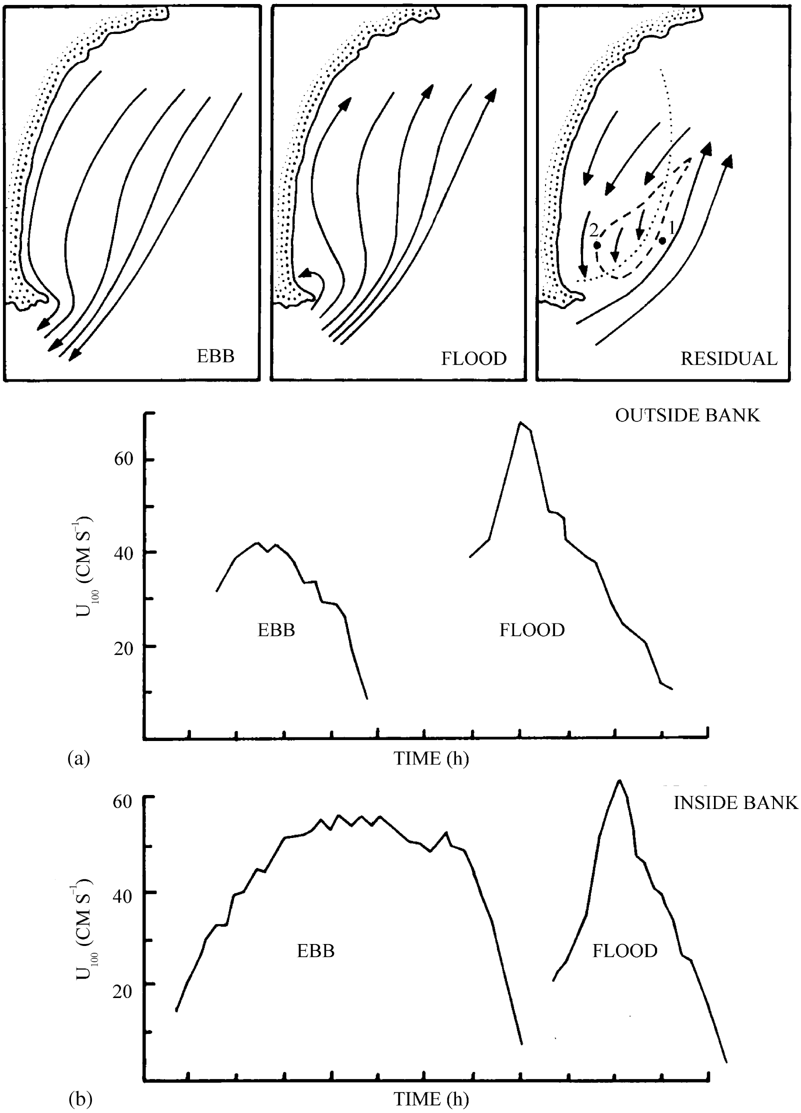

Fig. 5. The streamlines on flood and ebb tide, and of the tidal residual around a type 3A 'banner bank', The Skerries Bank, Start Bay, southern England, and the currents measured at positions 1 and 2 outside and inside the bank (from Dyer, 1986).

along the axis of the Bank. This is shown by comparison between the results of current measurements made on either side of the bank (Fig. 5). On the inshore side, the short sharp flood stage gives a higher shear stress than the longer flatter ebb, but the asymmetry is sufficient to ensure that the mean sediment transport is in the ebb direction. On the outside of the bank, the ebb current is shorter than the flood and is less strong. The asymmetry in the duration of the tidal cycle results from the current turning earlier close inshore towards the end of the flood tide, and that in magnitude is due to convergence and divergence of the flow at the headland during ebb and flood, respectively. Sand wave facing directions suggest a residual circulation of sediment corresponding to that of the water. However, consideration of the fluid dynamics of the situation shows that the residual eddies in the water flow are created by the headland, and that the banks then arise from the eddies, rather than vice versa. The eddies then trap the sediment produced by coast erosion, or arising from seafloor erosion, adjacent to the headland. The physics of this mechanism is discussed in Section 3.

2.3.2. Type 3B: alternating ridges

Where there is active retreat of the headland the sediment source also retreats, leaving the banner banks to be modified by tidal and storm forces, and creating new ones. There are many examples of groups of linear *en-echelon* or V or S shaped alternating ridges formed off large scale headlands formed by littoral drift convergences. These bank complexes have been termed shoal-retreat massifs (Swift and Sears, 1974). The authors point to four capes between Cape Romain and Cape Hatteras where the forelands have pumped sand seawards as they retreated. These have been modified into large massifs of shore parallel ridges and swales, the swales being spillways for alongshelf storm flow. It is also possible that shoreface connected ridges may form by a similar process where convergences occur in the littoral transport. At present there are two theories for the origin of *en echelon* sand ridges (Harris, 1988). They are based on: (i) shoreline retreat; and (ii) ridge multiplication. In many ways these two are related, since the source of the sand requires shoreline retreat, and the multiplication of the sand banks implies an offshore movement and modification of the outermost bank, which is aided by retreat of the shoreline.

(i) Shoreline retreat requires the detachment of headland associated sand banks, or barrier beaches, as the sea level rises, or because of shoreface erosion (Swift, 1975). As the coastline retreats, sand is swept towards the headland until there is a balance between the shape of the headland, the longshore transport rates to its tip, and the loss of sediment to the banks. The shape of the headland depends on the relative magnitudes of the transport, together with factors such as the intermittency of the processes. When the transport from one direction predominates, the resulting shore-attached ridge develops at an angle to the coast in the direction of the dominant transport. The offshore face is the steepest, and sediment is swept to the tip of the ridge. The transport in the landward trough can result from tidal currents augmented by surge currents and waves, and is similar to the mechanism for shoreface connected ridges (type 2B(ii)), the main difference between the two types being the difference in the importance of tidal currents.

A simplified conceptual picture of the growth of such ridges is shown in Fig. 6. In this instance the headland is retreating backwards by erosion from both sides. Initially, a spit is formed by transport from the left, and sand from right of the headland is also carried onto the spit when it encounters the dominant ebb current. The spit elongates, and a flood channel develops behind it. The ridge gradually

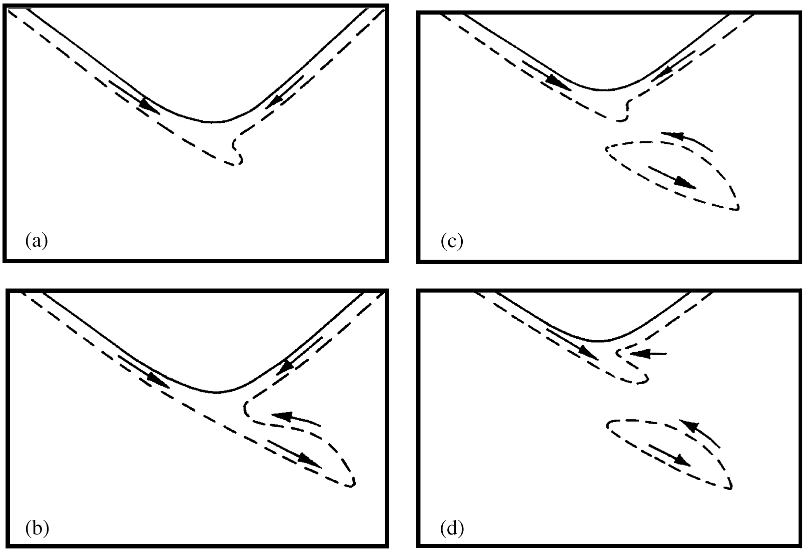

Fig. 6. Diagrammatic representation of the formation of *en echelon* banks following headland retreat. a–d show stages in the process.

extends offshore and eventually becomes separated from the retreating headland as the flood channel deepens, forming a banner bank. Continued recession forms a further spit, with ebb and flood channels. Thus, though the detached offshore ridges are initially relict, they continue to respond to the flow field resulting from tidal or storm induced currents (Swift, 1975). Varying rates of recession and differential erosion are likely to create more complex ridge distributions. The banks and shoals refract the waves and focus the energy on the headland, affecting the coast erosion rates. The formation of a ridge will tend to affect the wave erosion in its lee, and the rate of recession will be greater on the exposed part of the coast. Nevertheless, detachment could lead to renewed erosion in the area formerly protected by the ridge. The rates of erosion of the two sides of the headland may thus show considerable variation in relative magnitude as the ridge develops and separates, and it is possible that ridges may form alternately on either side, and a series of zig-zag ridges develop, which themselves may guide the flow and produce ebb-flood avoidance channels. These V or S shaped features are typified by Haisborough Bank/Winterton Shoal (52°40′N, 1°45′E), and by Skegness Middle Bank (53°10′N, 0°20′E).

Swift (1975) has described the situation for the coast of East Anglia, where the series of 'nesses' have connected ridges, formed initially as banner banks. An example of this is Sizewell Bank off Suffolk (52°15′N, 10°40′E). The channels landward of the banks are flood dominated, with strongest currents towards the ness, whereas the seaward sides are ebb dominated. Storm currents are probably also important, and can breach the connection between the bank and the headland. The nesses tend to migrate down the coast as the coast erodes back, and this appears to leave a sequence of detached ridges offshore. These ridges have probably been modified in shape and orientation to achieve approximate stability with the present flows, and so approach Type 1. The East Anglian banks are therefore considered to be alternating headland banks undergoing active formation. However, it is not clear what timescale is involved in the production of sand banks, and how that relates to variation in the physical processes and the rates of coast erosion.

Thus, banks that start as Banner Banks may become Alternating Ridges given time and further headland erosion. Once they are separated they are likely to gradually adjust to the offshore conditions so that the bank height and spacing increases steadily in a seawards direction. There is obvious similarity with the form of the Norfolk Banks.

The process of formation of alternating ridges has been described by (Swift, 1975) for Nantucket Shoals. These ridges are in an area where the flood currents reach 1–2 m s^{-1}, and ridge spacing and height increase offshore. Sandwaves are normal to the trough axes, but climb obliquely up the landward flanks of ridges to become ridge parallel at the crest. Paired ebb and flood sinuses occur, suggesting that the sense of the residual circulation is the same round each bank. Nantucket Island is the remnant of a much larger subaerial surface and Swift (1975) concludes that as the shoreface has eroded and retreated, the resulting debris has been packaged as shoreface-connected ridges, which have in turn evolved in response to the deepening flow field.

In a similar manner the Flemish Banks and the ridges in the Baie de la Somme may have been formed by separation of ridges during widening of the Dover Straits and erosion of the coastline to the east and the west of Cap Gris Nez.

(ii) The ridge multiplication hypothesis rests on an evolutionary sequence in the development of ridges suggested by Caston (1972). This is based on the form of Haisborough Sand, Hammond Knoll and Winterton Ridge, which are presumed to develop according to the process shown in Fig. 7. The principal stages appear to be:

(a) A ridge parallel to the direction of the tidal currents, probably possessing pronounced asymmetry.
(b) At this stage the ridge becomes slightly kinked, possibly because of an unequal rate of transport laterally throughout its length. It probably still possesses a constant asymmetry, but the slight double curve will be emphasised by the opposing ebb and flood currents. The Leman Bank may be approaching this stage at the moment.
(c) A more advanced stage may be typified by the Ower Bank, with the double curve developed to the extent that the asymmetry is now reversed over half of the ridge.
(d) The kink has developed into an incipient pair of ebb and flood channels.
(e) The new channels lengthen considerably, and the ridge between them becomes sub-parallel to those on either side. The Haisborough Tail – Winterton Ridge system is an example, where the parabolic terminal banks have 'blown out' leaving gaps between them.

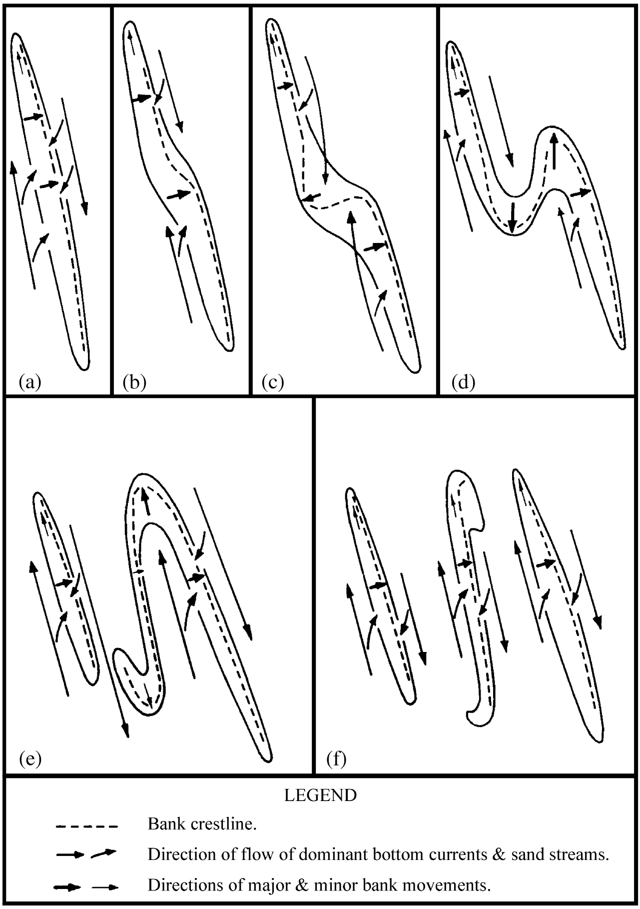

Fig. 7. The bank multiplication hypothesis of Caston (1972).

(f) The complete cycle leaves three ridges where there was one, but, as the innermost ridge cannot shift inshore to any extent, there is a resulting fairly massive transfer of sand offshore.

One difficulty with this conceptual model is that the middle ridge of the three has initially a reversed asymmetry according to the ebb and flood dominance. However, this theory does not account for the initial generation mechanism of the banks, but may explain part of the adjustment of the banks to the offshore hydrodynamic regime. The stability of an offset kink in the North Hinder Bank over a 135 year period has been examined by Smith (1988b). He found no evidence to support Caston's model, but considered it more likely that the kink would eventually cause separation into two *en echelon* ridges.

McCave and Langhorne (1982) have constructed an illustrative box model to describe the main features of sediment transport around Haisborough Sand. It assumes that the ridge is effectively a closed system, that the highest fluxes are about 250 $m^2\ y^{-1}$, and that the cross ridge flux implies a lateral migration rate for the ridge of 2.5 $m\ y^{-1}$. In order to have equal transport along the two sides, the steeper northern side requires much higher fluxes per metre than the southern side, which agrees with general observations. Using the model, it was estimated that it takes 550 years for sand to travel round the ridge.

2.4. Moribund ridges

Moribund ridges are found where the present peak currents are insufficient to move the sea bed sand. They do not have large sand waves on their flanks, and have more round crested cross sections. Their slopes are generally only 1° or so. They are separated by sandy or muddy floors, rather than clean gravel, and their crests do not appear to have as much wave disturbance due to the deeper water. Examples are shown by Indefatigable and Viking Banks (53°30–40′N, 2°10–30′E).

Other ridges are conjectured to have been formed at a lower sea level at positions formerly occupied by estuary mouths. An example is found just to the west of the Dogger Bank in the North Sea. They form a set of linear ridges collectively called East Bank which are considered to have been formed at about 9000 yr BP when sea level was about 50 m lower than present day. At that time the Dogger Bank was exposed, and there was a tidal strait to the west with much stronger currents than the present 50 $cm\ s^{-1}$. Typically the ridges are 40 km long and 3 km wide. They rise about 25 m above the general seabed of about 90m. The south-eastern flanks are slightly steeper than the northwestern ones. The general form and internal structure have been described by Davis and Balson (1992). They found that one ridge is composed of fine, well sorted and well packed sand. Even though the ridge appeared to be homogeneous from seismic surveys, there was more than one period of formation shown by coring, which revealed stacked coarsening-upwards sequences. There were fine laminations and ripple cross-bedding, and bioturbation was restricted to the upper metre or so.

Further moribund ridges are present on the edge of the continental shelf of the Celtic Sea. These are north-east trending features normal to the shelf edge. The largest is about 200 km long, 55 m high and 15 km wide. They have a separation of 20 km or so. The ridges towards the south-east are smaller, being 40–70 km long, 4–7.5 km wide, and 20–50 m high, and with a spacing of about 15 km. Slopes are less than 1° and the crests are rounded. Many of the banks have no clear asymmetry, though some of them have internal structures indicating asymmetry originally towards the south-east. Surface currents are only about 50 $cm\ s^{-1}$. However the southeastern ridges experience currents of about 75 $cm\ s^{-1}$, and these seem to have a partial cover of sandwaves and may be experiencing continuing present day modification. Modelling using a sea level 100 m lower than present, suggested currents would have been 1.5–2.5 times as strong as those at present (Belderson et al., 1986). Additionally, the tidal ellipse would have been rotated in a clockwise direction, thus giving ridge/current orientations agreeing with present day examples. These ridges are considered moribund, and it is likely that they were formed in the entrance to a major estuary system comprising all of the rivers discharging into the English Channel and Celtic Seas.

Very similar features are the shore-normal ridges and swales which converge towards the shelf edge on the North Carolina shelf. Swift and Sears (1974) suggests that they are of estuarine origin, but have been remoulded during transgression. Also, Yang and Sun (1988) have documented a series of ridges off the Chinese coast that stretch from water depths of over 100 m inshore into depths of 45 m. They consider that the ridges were formed progressively at the mouth of the Changjiang River as it retreated during the sea level rise. They distinguish four distinct groups which were formed between 13,500 and 11,000 BP.

Amos and King (1954) describe moribund ridges on the Canadian shelf which they ascribe to submerged barrier islands, and Berné et al. (1999) have investigated contrasting features in the Celtic Sea and the Gulf of Lions. It seems that with a fast rise in sea level, moribund forms of all ridge types can be preserved.

3. Theories for quantitative prediction of sand bank formation and maintenance

In the previous sections a classification of types of sand bank has been presented and some essentially descriptive ideas for their formation have been discussed. In this section we look more closely at theories for sand bank formation and maintenance which involve quantitative, or potentially quantitative, prediction of sand bank characteristics.

It is striking that, in the present state-of-the-art, most of the *descriptive* explanations for sand banks consider the water and sand movements as part of a single inter-related system, whilst, as we shall see, most theories attempting *quantitative* prediction start with the water movement and assume that sand banks develop as a result of spatial variations in hydrodynamic conditions. An important exception to this is the 'seabed stability analysis' approach exemplified by Huthnance (1982a), in which a coupled system of sediment- and hydro-dynamic equations is essential to prediction of sand bank characteristics.

Existing theories to explain the existence of sand banks fall into three generic types: those based on secondary flows assumed to produce a convergence of sediment transport towards a sand bank crest; those considering long period wave motion with length scales similar to sand bank scales; and those based upon stability analysis of the coupled hydrodynamic and morphodynamic system. Each of these types is briefly reviewed in the following sections. The general conclusions are that secondary flows may be relevant to the formation and maintenance of type 3a sand banks, seabed stability analysis appears to be very promising as an explanation for type 1 sand banks, but long period flows are unlikely to be relevant for sand banks. Banks in estuary mouths (type 2) have not yet received full theoretical examination, though some preliminary ideas will also be discussed. The effects of rising sea level are not incorporated explicitly in existing theories, so that type 3b sand banks are excluded from discussion.

3.1. Secondary flows

3.1.1. Helical flows

Off (1963) and Houbolt (1968) suggested that type 1 linear sand banks lying essentially parallel to a tidal or mean flow are the result of counter-rotating helical flows in the water column with their axes aligned in the flow direction. Their qualitative hypothesis was that these spiralling secondary flows extended to the scale of sand banks, creating regions of flow convergence near the bed, with a consequent accumulation of sediment to form a sand bank. Asymmetric sand banks could result from spirals of unequal strength.

Field evidence for this hypothesis is very mixed. Some observations of smaller-scale bedform migrations (Caston and Stride, 1970) and grain size distributions (Swift et al., 1973) do suggest convergence of sediment transport towards sand bank crests, but other observation of bedforms (Swift and Field, 1981) and current measurements (McCave, 1979; Soulsby, 1981; Howarth and Huthnance, 1984) find no such evidence.

The existence of transverse secondary circulation for flow over a rough bed is to be expected; it is analogous to Langmuir circulation at the water surface due to wind stress. However, the dominant scale of such circulation is now considered to be much smaller than the scale of sand banks. It may be responsible for low-relief sand ribbons in the presence of strong flows but not features of the scale of offshore sand banks. Transverse secondary flows induced by a pre-existing sand bank also appear to be very small (e.g. Loder et al., 1992), at least in the absence of water column density stratification.

3.1.2. Phase lag between bed stress and bed topography

Smith (1969) considered the case of bedforms at an appreciable angle to the main flow direction, and showed theoretically that the component of flow across the bedform should create a bed shear stress whose peak would occur *upstream* of the bedform crest, due to the advective term in the momentum equation. Since bedload sediment transport rate is expected to be related to some power of the bed stress, this displacement of the stress maximum relative to the bedform crest should cause erosion upstream of the stress maximum and deposition downstream of this stress maximum, i.e. in the vicinity of the bedform crest. Thus, an initial bedform would be expected to migrate downstream. In a reversing tidal flow the tide-averaged flow on either side of the bank crest would be towards the crest near the seabed, and the sediment transport would therefore have a net convergence towards the crest.

This explanation was originally applied to transverse sand waves, but it was also assumed that it would be relevant to larger sand banks on continental shelves. Swift and Field (1981) found it to be compatible with grain size distributions across sand banks on the US east coast and other researchers have also found it consistent with the forms and geology of sand banks in the North Sea (McCave, 1979; Swift et al., 1978).

More recent work, however, suggests that this lag between stress and topography, and the resulting longitudinal secondary flows, are relevant at sand wave scales but become negligible for the larger sand banks. Hulscher (1996) developed a three-dimensional model of seabed stability under oscillatory tidal flow and showed that the advective effect is only relevant for smaller scale features. Soulsby (1983) suggested that the appropriate criterion for the depth-dependence to be significant (a ratio of advection to friction terms) is

$$\frac{4\pi\Delta H h}{C_D \lambda^2} > 0.1 \tag{1}$$

where ΔH and λ are the height and wavelength respectively of the bedform, h is the mean water depth and C_D is the quadratic drag coefficient. For typical sand bank geometry this criterion is far from being fulfilled (e.g. for $\Delta H = 10$ m, $\lambda = 6$ km, $h = 20$ m, $C_D = 0.0025$, the ratio $= 3 \times 10^{-2}$). Thus for sand banks, the flow pattern is essentially depth-independent, with maximum flows over the crest at all depths. The conclusion is that Smith's (1969) theory cannot account for sand banks, though the stress distribution and associated residual longitudinal circulations may have a role for sand wave migration on the flanks of sand banks.

3.1.3. Vorticity from shoreline irregularity – Headland eddies

The secondary circulations described so far should occur for straight flows in open coast waters, but, as we have seen, type 3 sand banks are found close to the coast, associated with irregularities in the coastline (e.g. Skerries Bank, The Shambles). Tee (1976,1977) Pingree and Maddock (1977,1979), Pingree (1978) and Maddock and Pingree (1978) have studied the generation of eddies by headlands and other coastal irregularities, and suggested that type 3 sand banks are formed by, and centred on, these eddies.

Even in the absence of friction potential flow theory predicts an acceleration of the flow around a coastal promontory. The effect of inertia (the tidal advection terms in the momentum equations) means that the tidal flow tends to 'overshoot' the promontory on the downstream side on both flood and ebb, with the result that the tidal flows are asymmetric on either side of the promontory. With friction included, such a promontory will create vorticity which will be advected away from the promontory on both the flood and ebb tides, creating a pattern of eddies on either side of the promontory (Pingree and Maddock, 1977 (Portland Bill); Pingree, 1978 (Shambles); Pingree and Maddock, 1979 (Skerries Bank)). Fig. 8 shows the tidal-residual flows modelled by Pingree (1978) for the vicinity of Portland Bill. Robinson (1981) deftly explained this vorticity generation, pointing out the two separate mechanisms which arise (Fig. 9). In Fig. 9b the increased flow near the promontory creates a proportionately larger frictional force inshore, for a quadratic (or any non-linear) friction, and in Fig. 9c a shallower depth inshore results in greater depth-averaged friction inshore, both mechanisms creating a similar sense of vorticity. The 'overshooting' of the flow downstream of the promontory advects this vorticity away from the coast. There is also the possibility of separation of the flow from the coastline as it flows around the promontory, but Robinson (1983) concluded that this effect is probably relatively unimportant because it operates on a small scale (O (1 km)), though it might be responsible for some bedforms very close to the coast (Belderson et al., 1982).

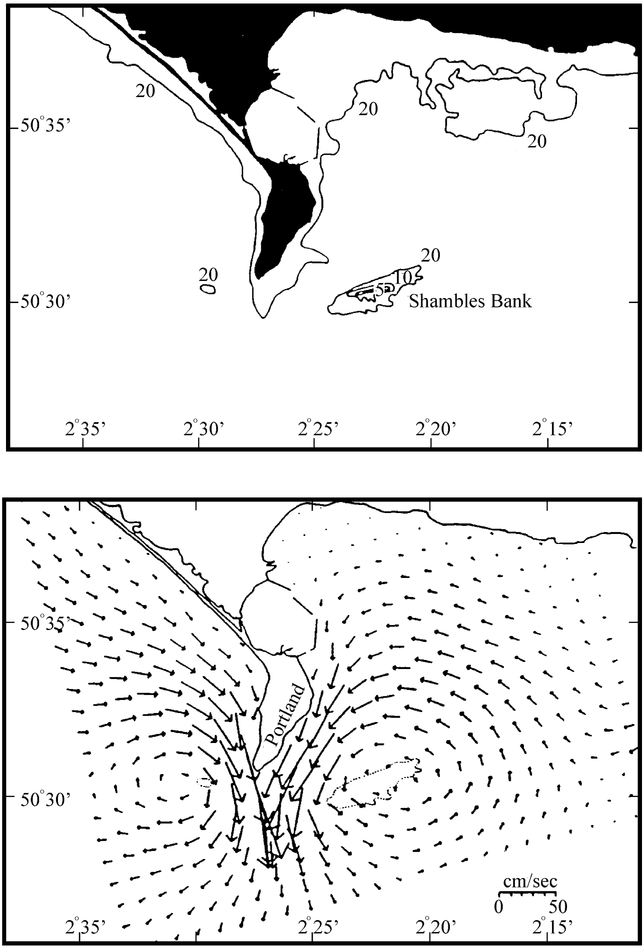

Fig. 8. Modelled tidal residual flows in the vicinity of Portland, southern England, showing the relationship between the residual flow gyre and the Shambles Bank (from Pingree, 1978).

These eddies are expected to generate secondary flows, with convergence towards the centre of the eddies at the seabed and divergence at the surface, with a strength which depends upon the balance between pressure gradient, Coriolis and inertial forces (Pingree, 1978). For rapid residual flows of large curvature the circular motion of the eddy is maintained by an inward pressure gradient balancing the outward inertial (centrifugal) force, and this pressure gradient forces a flow towards the centre of the eddy near the bed, where friction reduces the strength of the centrifugal force. In this case both clockwise and anticlockwise eddies would have equal inward secondary flows. For smaller flow speeds or curvature the Coriolis force becomes more important. The relative importance of the Coriolis and centrifugal forces is given by the

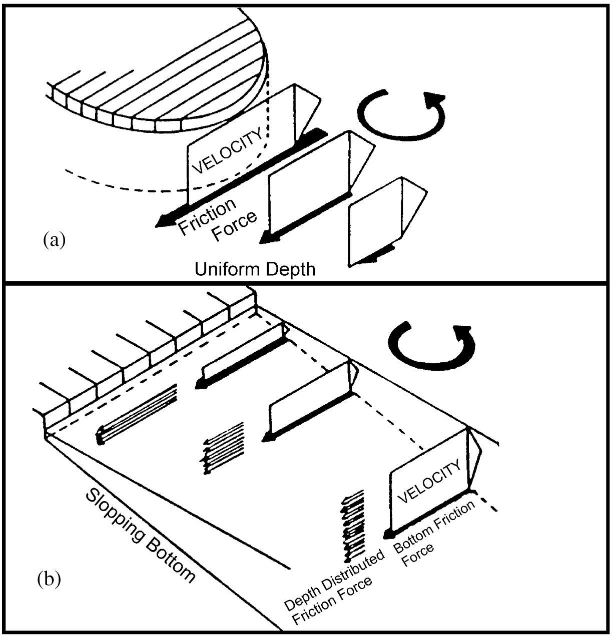

Fig. 9. Vorticity generation mechanisms for flow near a promontory. (a) Gradient of friction force due to gradient of flow speed. (b) Gradient of depth-averaged friction due to changing water depth (from Robinson, 1981).

Rossby number, R_o, of the flow:

$$R_o = \frac{U}{fR}, \tag{2}$$

where R is the radius of curvature of the flow. For small R_o (significant Coriolis effect) in the northern hemisphere anticlockwise eddies experience enhanced convergent flow at the bed, while clockwise eddies have a reduced convergence, or even a divergence of secondary flow at the seabed at the centre of the eddy. Pingree (1978) suggested that the secondary convergent flows generate sand banks at the centres of the eddies. He calculated $R_o = 1$ for Portland Bill and $= 5$ for Lundy Island, showing consistency with the more symmetrical development of sand banks observed on either side of Lundy.

Takasuki et al. (1994) proposed that eddies generated by flow past irregularities within an estuary can be advected to the coastal waters at the estuary mouth where they can generate sand banks. They show examples of sand banks in the Neko Seto Sea in Japan which they explain by this mechanism.

One problem with explaining sand banks through convergent secondary flows is that these flows are likely to be much smaller than the primary tidal flows. The residual circulation around an eddy is itself predicted to be an order of magnitude less than the tidal flow, and the secondary flows generated by the residual flow will be smaller still. Even where secondary flows are enhanced by density stratification of the water column their predicted strength is only of the order of 0.01 m/s (Tee, 1985). Another problem is highlighted by the current measurements made by Geyer and Signell (1990), using a shipboard ADCP, in the vicinity of a headland in Vinyard Sound, Massachusetts. They show very clearly the expected acceleration of the tidal flow around the headland and the associated vorticies being advected away from the headland, but they also show the highly transient nature of the flow over the tidal cycle, with the strength and location of the eddies varying rapidly. Thus, simple ideas about the generation of headland sand banks from *residual* circulation patterns alone are clearly inappropriate.

Nevertheless the residual flows, the secondary flows, and the tidal harmonics (e.g. the M4 tidal constituent) which are created by the same mechanism as the steady residuals, combine to modify the basic tidal flow and thus may create banks related to the predicted eddies. Field measurements, particularly those related to bed stress which would be expected to drive sediment transport, tend to confirm this suggestion. Soulsby (1981) made direct measurements of the seabed stress by the Reynolds stress method, through measurement of turbulence. He shows that the mean stress at a station near Skerries Bank is inclined about 6° to the residual flow direction, more towards the bank crest, probably due to steering of the residual flow by the bottom contours. He hypothesises that this current veering may act to sort sediment so that fine sediment is carried preferentially towards the bank crest. Other studies deduce bed stress from current (tidal and residual) measurements. Mardell and Pingree (1981) also study stresses associated with Skerries Bank, using a simple quadratic estimate of bed stress, and conclude that M_4 flows dominate the stress offshore while mean flows are important inshore, thus influencing the net stress pattern and inferred sediment transport near the bank. Similar studies by Pattiaratchi and Collins (1987), Collins et al. (1995) and others, use simple drag laws or sediment transport algorithms based upon bed stress to deduce net transport patterns which generally appear to converge towards bank crests.

Finally it is worth considering the extent to which curvature of the primary tidal or mean flows, for example around the ends of a linear bank, might generate secondary flows converging towards the centre of curvature at the bed. Taking a cue from Soulsby (1981; equation 82) and Pingree (1978), a suitable dimensionless parameter to determine whether secondary circulation due to flow curvature is significant should be the ratio of the centrifugal inertial force to the frictional force. Where the centrifugal force dominates, secondary circulation is likely, but where the friction force dominates, the turbulence generated at the bed will suppress any depth-dependent secondary circulation. Parametrising this ratio gives

$$\frac{\text{Inertia force}}{\text{Friction force}} = \frac{h}{RC_{\mathrm{D}}}, \tag{3}$$

where h is the water depth, R is the radius of curvature of the flow and C_{D} is the quadratic friction coefficient. With $C_{\mathrm{D}} = 0.0025$, a water depth of 20 m, for example, would require flow curvature of radius less than 8 km for significant secondary circulation, with proportionally smaller radii in shallower depths. This radius is not atypical of radii of the ends of sand banks, suggesting that 'centrifugal secondary flow' might occur there. It may also be representative of flow curvature around sharp coastal promontories, though larger radii might be more usual, perhaps implying a minor role for this kind of secondary circulation for sand banks as a whole.

3.2. *Long period waves*

3.2.1. *Standing long period edge waves*

Dolan et al. (1979) linked large-scale shoreline periodicities and associated offshore linear sand banks to long period (100s seconds) edge waves. Edge waves, phase-locked to each other so as to produce fixed alongshore patterns of flow, have been suggested as the explanation of a range of nearshore and beach bedforms with some alongshore regularity, from beach cusps to crescentic bars and more complex nearshore topography (Holman and Bowen, 1982). For a given beach slope, the alongshore and cross-shore scales of edge waves increase as the square of the wave period, and at each wave period a family of edge wave modes can exist, with the alongshore scale and the number of bars in the cross-shore direction increasing with the mode number. The formation of bedforms is assumed to be the result of steady drift velocities formed in the boundary layers of the edge waves, with convergent and divergent patterns similar to the edge wave structures. Edge waves which are progressive in the alongshore direction are expected to produce shore-parallel offshore sand bars, and edge waves which are standing alongshore as a result of phase-locking between components (due to reflection from a headland for example) are expected to produce crescentic patterns of offshore bars, with horns which point landward and with alongshore wavelengths similar to the edge wave motion.

The existence on beaches of edge waves of period much longer than wind waves is no longer in doubt (e.g. Huntley et al., 1981), and the link to topographic features of scales of a few hundred metres is strongly suggestive even if positive evidence is still elusive. However, the extension of these ideas by Dolan et al. (1979) to explain coastal bedforms on the scale of many kilometres is highly speculative since there is no evidence that edge wave motion of the required very long periods (100s of seconds) exists with sufficient magnitude to create sand banks. The postulated alongshore regularity of such large-scale features is also open to question. Perhaps most limiting, the edge wave hypothesis fails to provide a clear explanation for the fact that most offshore sand banks lie oblique to the coastline; as mentioned, edge waves are generally expected to produce either shore-parallel or crescentic bars.

3.2.2. *Instability of progressive long period waves*

Boczar-Karakiewicz and Bona (1986), Boczar-Karakiewicz et al. (1990), and Boczar-Karakiewicz et al. (1991) have postulated that offshore sand banks are the result of non-linear propagation of long-period swell waves (15–20 s), infragravity waves (50–100 s) or internal waves (15–20 min) over a shoaling continental shelf. Regular non-linear waves exhibit 'recurrence', in which energy flows cyclically back and forth between harmonics as the waves progress, resulting in a regular spatial variation in the form of the waves. This spatially varying pattern in the direction of propagation, the 'recurrence length', can then provide the basis for a spatially varying transport of sediment, and hence the development of bedforms on the scale of the recurrence. The recurrence length decreases with mean water depth, in qualitative agreement with the observation that spacing between offshore sand banks often decreases as depth decreases.

This hypothesis has been used to explain sand banks at several locations. On Sable Island Bank off the east coast of Canada, for example, there is a system of small sand banks 5–10 m high, with wavelengths 1600–3000 m (both features increasing with depth) in water depths of 20–80 m. Based upon current meter data from the area, Boczar-Karakiewicz et al. (1990) propose that these sand banks are caused by infragravity waves of approximately 50 s period propagating onto the continental shelf. These waves themselves have wavelengths of around 1200 m in deep water, but the recurrence length over the continental shelf is calculated to be 6000 m in 70 m water depth and 2000 m in 40 m water depth, providing good agreement with the spacing of the observed sand banks. Assuming propagation of the infragravity waves from the east, the sand bank orientation is found to be approximately transverse to the refracting wave propagation direction, as expected.

This hypothesis appears to provide a self-consistent explanation for type 1 sand banks on an exposed, wave-dominated coast. However there are a number of features of the explanation which require investigation before it could be generally accepted. One important problem is the need for the progressing waves consistently to exhibit the same recurrence pattern at the same locations over the continental shelf. Boczar-Karakiewicz et al. (1990) argue that locking of the recurrence pattern occurs relative to the location of the shelf edge, where the long period waves propagate rapidly from deep water to shallow water conditions. However, in reality any incident long-period waves are likely to be highly stochastic in nature, and the form of the recurrence is strongly dependent on the form of the initial non-linear waves. There is also the problem that, although it is possible in any particular case to find long-period waves with a recurrence length which matches observed sand bank scales, the choice of the period of these waves is not well constrained by direct observations. In a situation where the recurrence length is sensitive to the chosen long period, the proposed pattern of sediment convergence and divergence can become rapidly smeared by a broadening of the frequency band of long-period waves, particularly as distance from the shelf edge increases (assuming locking of the recurrence pattern at this location). The hypothesis also predicts that sand banks will be aligned with crests perpendicular to the propagation direction of the long period waves. In shallow coastal zones refraction will generally ensure that long period waves travel essentially perpendicular to the coastline, thus predicting shore-parallel sand banks rather than the typical oblique orientation. De Vriend (1990) found the changing orientation of sand banks with decreasing depth along the Dutch coast to be opposite to that predicted by long-wave refraction.

The conclusion must be that neither of these long-period wave hypotheses can account for the existence of sand banks.

3.3. Seabed stability analysis

Theories discussed so far consider sand banks to be the result of sand response to hydrodynamic effects. However, a number of recent theories consider the interacting system of water and sand, the characteristics of the sand response being an integral part of the theory.

An early example of this approach is the suggestion of Postma and Stride (see Stride, 1974; Swift and Ludwick, 1976) that a lag between the maximum of suspended sediment transport and the maximum flow in a tidal ellipse results in transport oblique to the principal axis of the flow. If this principal axis is parallel to the bank axis, and there is asymmetry between flood and ebb currents, a net transport onto the bank will occur. Unfortunately, this theory is probably inappropriate because the field evidence suggests that periods of dominant suspended load transport, occurring during storms, tend to erode rather than build sand banks.

Huthnance (1973) appears to have been the first to recognise the dynamical implications of the observation that sand banks are generally inclined at an angle to the main axis of tidal flows. He used depth-averaged hydrodynamic equations to show that tidal flows over a sand bank would create an anticyclonic residual (tide-averaged) flow around a sand bank inclined at an angle of 30–60° cyclonically relative to the major axis of the flow.

In a further important development, Huthnance (1982a) used the same depth-averaged equations in combination with a simple bedload transport equation which included a bed slope term, and studied the stability of this coupled morphodynamic system to perturbations in the seabed elevation. He was able to show that the fastest growing perturbations were those with depth contours inclined at 28° to the major tidal axis, though with a broad maximum so that the actual angle may be susceptible to other influences, such as the proximity of the coast. He also predicted a spacing (wavelength) of 250 times the mean water depth. These predictions are in at least qualitative agreement with observed type 1 sand banks in European (e.g. Pattiaratchi and Collins, 1987; Swift et al., 1978), North American (Figueiredo et al., 1981) and South American (Parker et al., 1982) coastal waters.

The simplest way to explain the growth of sand banks which are inclined relative to the flow direction is to note that the cross-bank component of the flow is constrained by continuity to increase over the crest, but the alongcrest component will be reduced by the influence of friction. The current vector therefore turns towards the crest as the flow approaches the crest, resulting in crestward sediment transport. On the downstream side of the bank the flow speed is reduced due to friction over the bank. Thus sediment transport is towards the bank on the upstream side, and for a reversing tidal flow there will be convergence towards the crest.

This explanation is reminiscent of the Smith (1969) explanation, in that it postulates transport up the upstream side of the bank, but the mechanism is quite different, being the result of friction in the Huthnance theory as opposed to advection in Smith's (1969) theory. Huthnance (1982a,b) does not include depth-dependent hydrodynamics and therefore would not predict growth for perturbations normal to the flow direction.

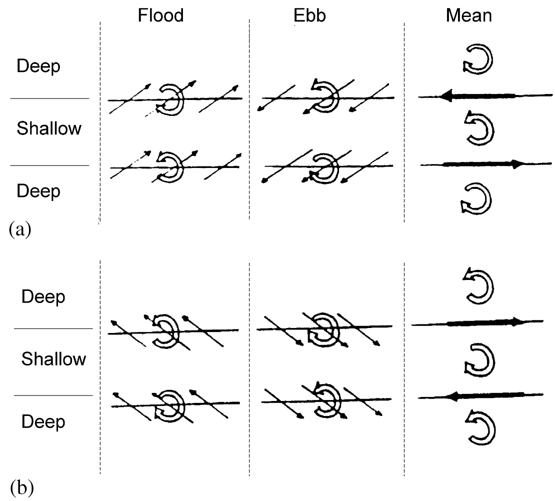

Fig. 10. Vorticity generation by differential friction acting on tidal flow over a sand bank (from Robinson, 1983).

The combined effect on the cross-bank and alongbank flows is often described in terms of vorticity. Robinson (1983) provides perhaps the clearest description, though the approach appears to have originated with Zimmerman (1978). Fig. 10 shows the vorticity which occurs as a result of the frictional torque on the tidal flows, with the stronger retarding effect of friction on the shallower side of the flow. The curvature of the vorticity is such as to bend the flow vectors in the direction shown in Fig. 11.

The vorticity approach also allows a ready description of the additional effect of the Coriolis force (Fig. 12), in which the shortening of water columns passing over the sand bank induces anticyclonic vorticity through the principle of conservation of potential (local + planetary) vorticity. This effect will tend to enhance the 'refraction' effect for flow over banks which are cyclonically inclined to the tidal flow direction (Fig. 10b) and reduce it for banks which are anticyclonically inclined (Fig. 10a). In the Huthnance (1982a) theory for sand bank growth this extra refraction leads to the prediction that cyclonically inclined banks will grow more quickly, and hence be more prevalent. However, since friction depends upon the square of the flow speed whilst the Coriolis effect increases only linearly with speed, the relative importance of the Coriolis term decreases for faster flows. The ratio of the Coriolis term to the friction term can be written (Huthnance, 1982b):

$$\frac{hf}{C_D U} \approx \frac{h}{C_D U T} < 1, \tag{4}$$

where f is the Coriolis parameter, U is a typical flow speed and T is the tidal period. Huthnance (1982b) showed that there is also a dependence on the angle of obliquity of the bank, but this is typically $O(1)$. Thus, for the strong currents and relatively shallow water typical of coastal sand banks the Coriolis effect is relatively small. For example,

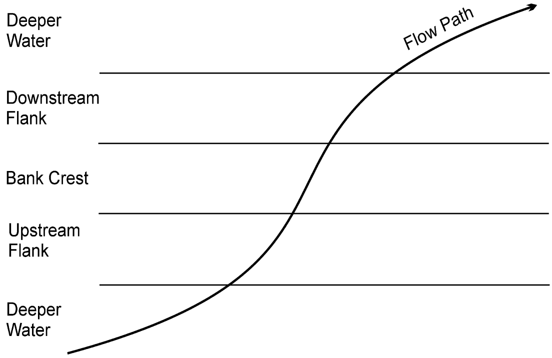

Fig. 11. Diagrammatic representation of the expected refraction of a current flowing over a sand bank.

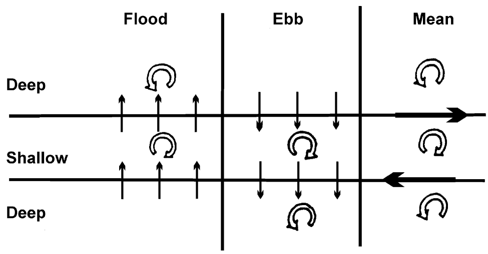

Fig. 12. Vorticity generation by water columns stretched and squeezing over a ridge (from Robinson, 1983).

in 10 m depth and a flow velocity of 1 m/s, with $f = 10^{-4}$ and $C_D = 0.0025$, the ratio is 0.4, small enough to make the bias towards cyclonically inclined banks weak (Huthnance, 1982b). Nevertheless, as pointed out in Section 2.1, cyclonically inclined banks do appear to predominate in the northern hemisphere, though the situation in the southern hemisphere is less clear.

Huthnance's (1982a) first paper considered only straight parallel depth contours, but the subsequent paper (1982b) dealt with the development of two-dimensional perturbations of the seabed. For axisymmetric bumps the evolution was found to be complex, with development both along and across the tidal flow direction, but elliptical bumps, with a flow direction inclined between 18 and 48° relative to the major axis of the bump, became more elongated in the major axis direction, with a tendency also to rotate towards the 27° inclination found to give the greatest growth rate in the straight parallel contour case.

De Vriend (1990) has extended the Huthnance (1982a,b) analysis to include wind waves (mild slope equation, including refraction, shoaling and diffraction, but ignoring reflection, non-linearity and randomness). His motivation was to attempt to explain the morphology of type 1 sand banks in shallow coastal water where surface wave effects cannot be ignored. The results were rather speculative, but de Vriend concluded that wave effects were primarily sediment 'stirring', acting through the

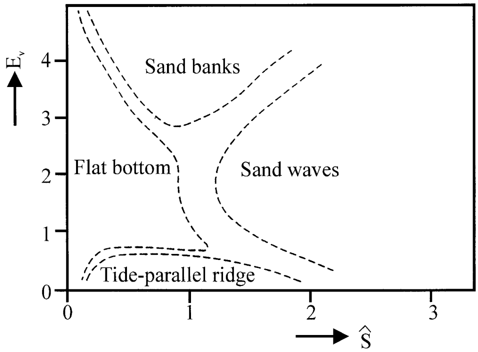

Fig. 13. Characteristic bedforms predicted by a three-dimensional shallow water model, as a function of the non-dimensional bottom slip parameter $S = 2s/\sigma H$, and Stokes number $E_V = 2A_v/\sigma H^2$, where A_v is the eddy viscosity, s is a seabed slip or resistance parameter, H is the water depth and σ is the tidal frequency. The scale on this schematic diagram is only approximate (after Hulscher, 1996).

sediment transport algorithms, with other wave effects (asymmetry, boundary layer streaming) being hard to assess at present.

Hulscher (1996) has studied the stability of the seabed through a three-dimensional hydrodynamic model linked to a sediment transport model similar to Huthnance (1982a,b). The theoretical approach considers tidal flow, with an arbitrary degree of ellipticity, and a variable seabed boundary condition, from no-slip to perfect slip, to allow for different bed roughnesses and hence bed stresses. The three-dimensional theory provides prediction of the growth of sand waves, with crests essentially perpendicular to the flow, as well as sand banks obliquely aligned to the flow. The growth rate and angle of the bedforms is shown to be a function of a Stokes number (dimensionless vertical eddy viscosity, scaled by the water depth and tidal frequency), reflecting the fraction of the water column influenced by the boundary turbulence, and a normalised slip parameter representing the importance of bed friction. Fig. 13 shows the resulting predicted bedforms schematically. Large values of Stokes number (E_v) mean that phase lag between bed stress and depth-averaged current is small, and small to intermediate slip values mean that the influence of Coriolis is significant compared to the influence of bed friction. Thus, sand banks are the result of a balance between Coriolis and friction effects at large values of Stokes number where the three-dimensional flow tends to 2-D vertically homogeneous. Hulscher (1996) predicts that the angle of the sand bank relative to the flow direction (major axis) will be up to 60° cyclonic if the Coriolis effect dominates (low slip parameter) but will reduce with increasing influence of the bed friction. The approach shown schematically in Fig. 13 has been quantified and applied with some success by van den Brink (1998) to the prediction of the distribution of sand waves in the southern North Sea.

Trowbridge (1995) has attempted to extend these instability ideas to type 1 sand banks found on storm-dominated continental shelves such as off the Atlantic coast of the USA. He considers the instability of an alongshore-flowing current over a seaward-sloping seabed, with a simple linear relationship between sediment flux and current strength. Flow over an incipient ridge causes a seawards deflection of the flow, and with a transport rate which decreases in the seaward direction, sediment deposition occurs thus causing growth of the ridge. He finds that the predicted scale and orientation of the fastest-growing ridges are in good agreement with the observed scales of the eastern US sand banks, which occur in a region of predominantly southwards flow along the coast. However, the conditions for growth in his simplified model are relatively restrictive. He concludes that wind-driven flows alone cannot account for the observed ridges, and suggests that longshore flows due to obliquely incident breaking waves must also contribute. There is certainly scope for further work, both theoretical and observational, to follow up this interesting theory.

It is clear that the morphodynamic stability models first proposed by Huthnance (1982a) are proving very productive, providing at least qualitatively accurate predictions for type 1 sand banks on tidally, and perhaps storm-dominated shelves. However, quantitative comparisons, though encouraging, are as yet relatively crude and require better field measurements of hydrodynamics and sediment dynamics in the field. Even the basic premise that tidal flows are accelerated and 'refracted' over sand banks is proving remarkably difficult to confirm in the field. Pattiaratchi and Collins (1987) describe measurements from moored current meters in the vicinity of Scarweather Sands, a type 3A sand bank on the northern side of the Bristol Channel (9.6 km long, 1.6 km wide in a typical mean depth of 15 m) which are typical of those from a limited number of fixed moorings. Their measurements show that, far from accelerating over the bank, the 'flow avoids the shallower region of the bank' at each of their three moorings and tends to run along the regional depth contours. Howarth and Huthnance (1984), reporting current measurements near Well Bank, a type 3B sand bank in the Norfolk Sand Bank system, find a similar tendency of the flow direction to align to local and regional depth contours, and point out that curvature of irregular depth contours may have an important influence on tidal current structure; they state that 'curvature $> (6\ \mathrm{km})^{-1}$ renders centrifugal forces important for the $O(0.7\ \mathrm{m/s})$ tidal currents'. For Well Bank they find that the major axis of the essentially rectilinear semi-diurnal tidal flow is inclined at $O(5^{\circ})$ (probably clockwise) to the bank axis, an angle rather less than suggested by earlier observations, and hardly exceeding the uncertainty in the bank axis. They speculate that this small obliquity may be the result of the bank having been formed at a lower sea level when the tidal ellipse was at a greater angle. Some confirmation of current refraction was found by Venn and D'Olier (1983) from current meter measurements over a long narrow Type 1 sand bank (South Falls Bank, southern North Sea). Harris et al. (1992) also find evidence for acceleration of the tidal flow over the crests of sand banks in Moreton Bay, Australia, using an array of six current meter moorings. These sand banks are within a relatively restricted bay with ample supplies of sand, and appear more like estuary-mouth sand banks than offshore sand banks.

Van de Meene (1994) used a combination of ship-board ADCP measurements and moored current meters to study tidal hydrodynamics over shoreface-connected ridges off the central Dutch coast. The spatial detail of the ADCP results did enable him to observe acceleration of the crossbank flow during the ebb phase of the tide, but

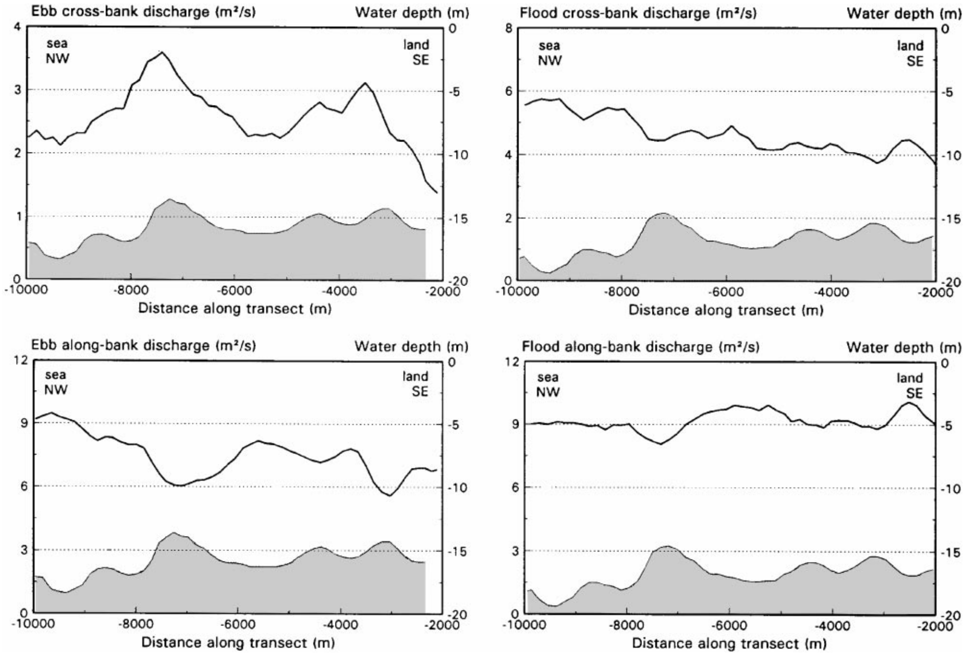

Fig. 14. Ebb and flood cross- and along-bank components of total water discharge along a transect through shoreface-connected ridges on the Dutch coast (from Van de Meene, 1994).

confusingly no such acceleration was found for the flood tide (Fig. 14). He concludes that the assumption of crossbank continuity is not valid for his banks on either the flood or ebb phases, perhaps because of the finite length of the bank, or because of regional topographic features (the proximity of the shoreline and a regional slope of the seabed), or because the assumption of a fixed sea level (rigid lid) is invalid. Moored current meter data shows that the major axis of tidal flow is inclined to both the shoreline and the bank, with an angle of around 30° relative to the bank axis. This is of a similar magnitude to that predicted by Huthnance (1982a) but, with the observed banks inclined *clockwise* relative to the tidal flow, is not in the sense predicted for the Coriolis effect. Given the average depth (15 m) and the relatively slow tidal flows (0.25–0.5 m/s) this is surprising since the dimensionless parameter of Eq. (4) gives the value 1.2–2.4, suggesting a significant Coriolis influence. However, Huthnance (1982b) shows that the main influence of this term is in producing an effective tidal current asymmetry between flood and ebb. In this coastal region tidal asymmetry arising from regional effects which are not local to the bank may well swamp the local Coriolis effect.

The best demonstration of refraction of tidal currents over sand banks comes from recent measurements of surface flows over Middelkerke Bank off the Belgian coast during the EU-funded CSTAB (Circulation and Sediment Transport Around Banks) Project (O'Connor and the CSTAB Group, 1995). Surface currents were measured

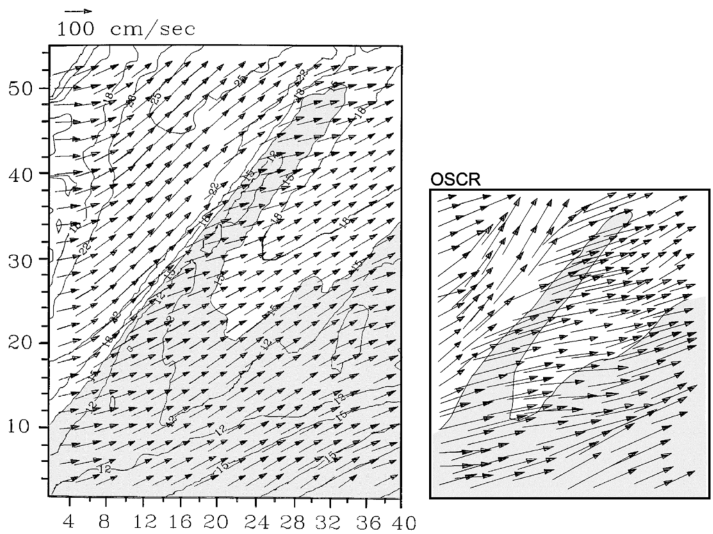

Fig. 15. Comparison between OSCR measured surface currents and modelled currents over Middelkerke Bank, Belgian coast, demonstrating the observed and modelled refraction of tidal currents over the bank. The shaded area represents depths less than 15 m, and Middelkerke Bank is the finger of shallow water pointing towards the top right-hand side of the diagrams. The flows are for low water, a 6 m s^{-1} SE wind and waves of significant height 0.5 m and period 3.4 s. (after O'Connor and the CSTAB Group, 1995).

using the NERC OSCR II system, an Ocean Surface Current Radar system measuring currents through the Doppler shift of the surface return. Measurements cover a rectangular grid of 1 km^2 cells over approximately 100 km^2, and are integrated over a 20 minute period, resulting in an accuracy of about ± 4 cm/s. Fig. 15 shows measured flows at low water for a 4.5 m tide and a 6 m/s SE wind, with waves of 0.5 m significant height and 3.4 s period. The refraction to a more bank-perpendicular direction over the bank crest is clearly shown. Fig. 15 also shows good agreement with 3D numerical modelling of the flows (O'Connor and the CSTAB Group, 1995).

These observations highlight the very localised nature of the refraction, perhaps explaining why limited moorings can miss the effect. Indeed Williams (O'Connor and the CSTAB Group, 1996, Chapter 8), commenting on the results from moorings around Middelkerke Banks, finds that the tidal analysis 'shows clearly that despite the proximity of Middelkerke Bank, the phase and direction of a given (tidal) constituent is very similar irrespective of location or depth'.

This observed local nature of refraction over banks is supported by an approximate dimensionless analysis. Since the basic physical balance of forces near the sand bank is likely to be between the advective term and a pressure gradient term, a suitable criterion to determine whether flow goes over or around a bank might be the ratio of the over-bank to round-bank advective terms. Soulsby (1983) gives an estimate of the over-bank advective term, and the around-bank term will be determined by the centrifugal force due to flow curvature. We can therefore estimate that flow *around* a sand bank will become significant if

$$\frac{\lambda^2}{4\pi\Delta HR} > 0.1, \tag{5}$$

where R is the radius of curvature of the flow around the bank. For a bank of height 10 m and width (wavelength) 6 km, this criterion implies a radius of flow curvature less than about 28 km, suggesting that flow around rather than over many sand banks is to be expected.

Theory also predicts a tidally-induced residual circulation around sand banks which, for small Rossby numbers and hence cyclonically inclined banks, should be clockwise in the northern hemisphere and anticlockwise in the southern hemisphere. Again, however, field confirmation is not universal. A clockwise circulation has been inferred from measurements by Caston and Stride (1970), Caston (1972), Harvey and Vincent (1977), Heathershaw and Hammond (1980), Pattiaratchi and Collins (1987) and Collins et al. (1995). Other studies have failed to reveal any such circulation (Harris et al., 1992; Klein and Mittelstaedt, 1992; Van de Meene, 1994). Butman et al. (1983) and Howarth and Huthnance (1984) do observe residual flows of the predicted circulation, but deduce that at best only half of the magnitude of the observed flows is due to the predicted tidal rectification, with wind and stratification effects readily swamping any short-term attempt to measure the tidally induced residual. There are very few measurements from the southern hemisphere from which residual flows around sand banks can be determined, though Parker et al. (1982) show vectors of sediment transport calculated from current meter records on sand banks on the Argentinean continental shelf which are consistent with an anticlockwise circulation; unfortunately flow speeds are not published so estimates of the Rossby number cannot be made.

Further theoretical developments are also needed since, in principle at least, the present linear stability analysis becomes invalid once the banks grow to a finite height. The relevance of the theories to storm-dominated coasts also needs to be investigated further.

3.4. Ebb and flood channels in estuary mouths

Estuary mouth sand banks (types 2A and 2B) have been widely studied in terms of descriptive geomorphological processes (Section 2) but have received little quantitative attention. However, some recent work has clarified some of the controls determining ebb and flood dominance and the resulting banks and channels.

Asymmetry of the tidal currents flowing through an estuary mouth depends on the shape of the cross-section of the mouth, the changing volumes of water needed to fill the estuary as the tide rises, and the nature of bed friction. Where shallow areas are still covered at low water, the flooding tide is mainly concentrated in the main channel, but in the shallower water the ebbing current exceeds the flood. Thus, the channels are flood-dominated and the shallows are ebb-dominated. However, where there are extensive intertidal areas, the combination of asymmetry in the tidal curve and the exposed areas at low water produces a flood residual current over the intertidal areas, and the compensation for the resulting Stokes Drift enhances the current in the channels, which become ebb dominant. Some of these effects have been explored by Friedrichs and Aubrey (1988). The consequence is that there is typically a change of dominance of the currents as one progresses into an estuary, with a tendency for the flood dominated channels near the mouth to avoid the ebb channels further up. Banks form between them, of type 2A, and horizontal water and sediment circulations are produced. These ideas have not yet been developed to the point of providing quantitative prediction of sand bank forms, but they may point the way ahead.

The extent to which the theory of Huthnance (1982a) may be applicable to type 2A banks is clearly worthy of further study. As discussed in Section 2, the sand ridges observed in Chesapeake Bay by Ludwick (1974,1975) lie essentially parallel to the ebb tidal flow but are oblique to flood steamlines, as required by the Huthnance mechanism. It is important to establish whether such obliquity is a common feature of type 2A banks. Even if this proves not to be the case, type 2A banks are manifestly the product of strong interaction between flow and topography, and the developments in morphodynamic stability models exemplified by the work of Huthnance (1982a) and Hulscher (1996) provide a promising basis for further research.

Type 2B sand banks, for which surface wave effects are very important, have also received extensive qualitative study (Section 2) but little quantitative study based upon processes. For shorelines subject to oblique wave incidence, stability analysis of the combined bedform and wave-driven longshorecurrent system, discussed for example by Damgaard-Cristensen et al. (1995), may be productive, but has as yet only been applied to smaller scale sand bars.

3.5. Summary

Of the theoretical ideas for sandbank growth and maintenance discussed here, the most advanced are the stability theories based upon coupled hydrodynamic and sediment dynamic equations. They provide reasonable quantitative predictions of the wavelengths and orientations of type 1 sand banks in tidally dominated regions, although they as yet relate primarily to the initial development of a sand bank and cannot predict the ultimate height of such features. The relevance of the frequency-dependent theory to storm-dominated environments is still to be assessed, though there are indications (e.g. Holloway, 1987) that similar effects may occur for a wide spectrum of wind-driven and other time-varying flows. Theories based on the instabilities of steady coastal flows (e.g. Trowbridge, 1995) also need further investigation.

Theories dependent upon convergence of residual (tide-averaged) circulation are less satisfactory, largely because these residual flows will generally be much smaller than the tidal flows. This and the strongly non-linear dependence of sediment transport on instantaneous flow speed means that there is no guarantee that average sediment transport follows average flow. (Indeed the theory of Huthnance (1982a) predicts convergence of sediment towards the sand bank crest with no such residual flow convergence.) Thus, although the headland eddy explanation for type 3A headland sand banks clearly correctly predicts the location and scale of these sand banks, detailed prediction requires the inclusion of a satisfactory sediment transport model for this case. The recent theoretical work of Hulscher (1996) is also of considerable interest in showing that, for the offshore case at least, vertical circulation (cf. the Smith (1969) theory) generally works on the scale of sand waves and shorter, rather than on sand bank scales.

Both of the long-period wave theories appear to be flawed. Neither can explain the commonly observed oblique orientation of sand banks relative to the principle flow direction or to the orientation of the shoreline. They both also appear to require specific choices of long period wave motion, while observations, at least to date, do not support such specific choices.

Theoretical ideas relevant to type 2 estuary mouth sand banks are still very primitive. Banks dependent on recession of the shoreline are not yet treated by any quantitative theoretical approach.

The conclusion from this study of existing theories must be that the state-of-the-art is far from being able to predict the scale and evolution of sand banks in most cases. Whilst type 1 banks are closest to being predictable, other types are at best only qualitatively predictable.

4. Discussion and conclusions

It is apparent from this review that there has been a lot of confusion in the literature concerning sand banks and ridges, their origin and maintenance. This appears to have arisen because banks have been viewed from two different perspectives. The first, which has been mainly taken by geologists and geographers is a non-quantitative one. This emphasises the origin of the banks under the scenario of the post-glacial sea level rise. It considers the regional setting within which the banks have developed, and because the rate at which the banks develop and move is long compared with the observation period, short-term sediment movement indicators are assumed to be indicators of the processes that generated them. This approach considers that the banks interact with the sediment source, and with each other as part of a transport path. All of the origins depend on the source of sediment being created by coastal retreat and coastal erosion. The contribution of eroded sea bed sediment is not normally considered to be large, though morainic material may form the cores of some ridges.

The alternative view that has been taken by physical oceanographers is based on assumptions that the shape of the ridge is in equilibrium with the present day processes, and that effectively each is separate and does not affect its neighbour. This takes no account of recession of the coastline, and assumes that the sediment source is the bank itself and the immediate surrounding sea bed, with the implication that the bank is more or less in the same position as when it was formed. Consequently, even if sea level rise may be crucial to the formation of banks, it is not necessary to explain their present day characteristics.

It is apparent that there are several possible origins for sand banks. It is also possible that banks may lose the characteristics of their origin once they are separated from the initial sediment source, and tend to converge on an equilibrium with the tidal hydrodynamics. The Huthnance model is an attractive one as it has the prospect of being a generic model that describes the equilibrium situation. Its most powerful result is that it explains the angular discordance between the flow and the bank orientation. However, it is limited to constant water depth, and to situations where neighbouring banks do not interact hydrodynamically or sedimentologically. It models the type 1 (Open Shelf Linear Banks), and this seems to be the ultimate state to which banks in the type 2B(ii) category (Tidal deltas, with recession-shoreface connected ridges) and the type 3B (Headland banks with recession) converge. How quickly that convergence takes place, and how quickly the sand bank loses its 'memory' of its original conditions is a matter of conjecture.

The other generic model that describes the equilibrium conditions is that of Pingree for the Type 3A Banner banks. In this case the banks are likely to have migrated little, and may be in an equilibrium situation that is constrained by a low rate of input of sediment and a low coastal retreat rate. Given further input of sediment, they may evolve to a new steady state. Again, the time rate of change is difficult to measure.

There are several bank types that cannot be modelled using Huthnance's approach, or its derivatives, or the Pingree model. These are the linear banks in wide estuary mouths (type 2A), and the ebb tidal deltas (type 2B(i)). For both of these, the dependence of the banks on the details of the surrounding banks, coastline and inlet precludes the development of generic models. However, the banks are included as topographical features in many estuarine and coastal models, but without the objective of trying to elucidate the processes governing their formation and movement. The first step in obtaining general conditions for the linear banks to be found in macrotidal estuaries may be by investigating empirical relationships between, for instance, bank number and estuary width. This can then be followed by taking an existing regional model of a situation, such as the Thames Estuary, and adjusting the number and shape of the banks to see how the tidal and sediment transport conditions adjust. The equilibrium for this situation is likely to be one where there is a constant sediment transport rate along the estuary, i.e. no divergence in the sediment transport rate, so that though there may be tidal transport of sand, there is no net erosion or deposition.

An alternative and potentially fruitful approach is to perform numerical experiments with a flow model coupled to a sand transport model, and a sand continuity equation that causes the bed to evolve. Such numerical experiments could be performed for a range of simplified topographies, and physical processes could be deduced from sensitivity analyses which explore the importance of, for example, different non-linear hydrodynamic terms, wind and tidal flows, sand transport algorithms (suspension and bedload) and vertical flow structure. The rapid increase in computing power is making such numerical experiments a valuable means of exploring the range of processes responsible for the formation and maintenance of sand banks.

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