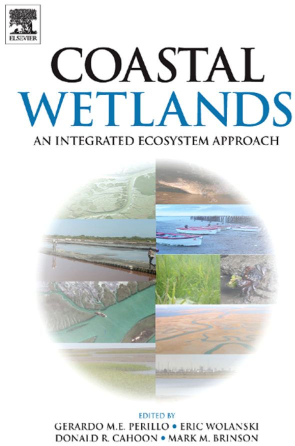

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PART III

Tidal Flats

CHAPTER 10

Geomorphology and Sedimentology of Tidal Flats

Shu Gao

Contents

1. Introduction

Tidal flats are distributed widely along the world coastlines, representing an important part of coastal wetlands. A tidal flat can be divided into three parts, according to their relation to the characteristic tidal water levels (Amos, 1995): (1) supratidal zone, which is located above the high water on springs and is inundated only under extreme conditions (e.g., storm surge events); (2) intertidal zone, located between the high water and low water on springs and is inundated periodically during spring–neap tidal cycles, and (3) subtidal zone, which is below the low water on springs and is rarely exposed in air. In literature the studies on tidal flats have been concentrated mainly on the intertidal part; some authors use the term "tidal flat" to represent the intertidal zone (which is adopted in the present study), whilst others prefer the term "intertidal flat." This chapter will concentrate mainly on the physical aspects of the intertidal flat; salt marshes will be described only briefly when necessary, because they are treated in detail in a separate chapter.

The well studied tidal flats include those along the Dutch, German, and Danish coasts (Reineck, 1972; Reineck and Singh, 1980; Pejrup, 1988), in the Wash embayment of England (Evans and Collins, 1975; Collins et al., 1981), over the bayhead areas of the Bay of Fundy, eastern Canada (Amos and Mosher, 1985) and along the Jiangsu coastlines in eastern China (Wang, 1983; Ren, 1986) (Figure 1). Generally, these tidal flat systems are characterized by accumulation of fine-grained sediments and gentle bed slopes. Tidal currents are strong on the tidal flat, resulting in high mobility of bed materials. Over the upper parts of the intertidal zone, salt marshes may be present, with water and nutrients being supplied by the tides (Zhang et al., 2004). In tropical areas, mangroves may develop on mudflats (Wells and Coleman, 1981).

Progress has been made in the understanding of the characteristics, processes, and evolution of tidal flats, which is important for the purpose of coastal wetland protection and restoration. In early times, the unique morphological features, especially zonation in geomorphology, attracted the researchers. From high water to low water marks, there are systematic changes in sediment grain size, bedforms, sedimentary structures, and biological activities, which have been studied since the 1930s (e.g. Haentzschel, 1939; Linke, 1939). Then, from the 1950s, sediment dynamic and morphodynamic studies have been carried out, in an attempt to understand the mechanisms of sediment transport and accumulation (Postma, 1954; van Straaten and Kuenen, 1957, 1958). Extensive in situ measurements

Figure 1 Coastal sections (dashed lines) and locations (triangles) along the world coastlines where extensive or detailed studies on tidal flat sedimentology and geomorphology have been undertaken (Allen, 1965; Reineck and Singh, 1980; Klein, 1985; Ren, 1986; Isla et al., 1991; Daborn et al., 1993; Perillo et al., 1996; Netto and Lana, 1997; Perillo and Piccolo, 1999; Kjerfve et al., 2002; Lim and Park, 2003; Deloffre et al., 2005; Falcão et al., 2006; Quaresma et al., 2007; Sakamaki and Nishimura, 2007; Anthony et al., 2008; Proske et al., 2008; Talke and Stacey, 2008).

have been undertaken, on the flat surface and in tidal creeks, to obtain information on tidal current, suspended sediment concentration, and the benthic boundary layer (Evans and Collins, 1975; Letzsch and Frey, 1980; Collins et al., 1981, 1998; Stumpf, 1983; Bartholdy and Madsen, 1985; Pejrup, 1988; Wells et al., 1990; Alexander et al., 1991; Gouleau et al., 2000; Andersen and Pejrup, 2001; Davidson-Arnott et al., 2002). On such a basis, quantitative models for tidal flat sedimentation and morphodynamics have been proposed (Allen, 1989, 2000, 2003; French, 1993; Allen and Duffy, 1998; Roberts et al., 2000; Pritchard et al., 2002; Malvarez et al., 2004; Temmerman et al., 2004). At the same time, studies on sedimentary sequences and associated physically and biologically induced sedimentary structures were documented in detail, to obtain information on the environmental conditions under which the deposits were formed and on the environmental changes. Recently, research has been focused on the formation of tidal flat sediment systems and the information on climate, environmental, and ecosystem changes contained in the sedimentary record (Cundy and Croudace, 1995; Dellwig et al., 2000; Gerdes et al., 2003; Gao, 2007a). Furthermore, the evolution of tidal flats in response to global climate change and intensified anthropogenic activities has become an important research area (Vos and van Kesteren, 2000).

The purpose of the present contribution is to provide a general description about the sediment distribution patterns and morphological features of tidal flats, together with an overview of the conditions and physical processes for the formation and evolution of tidal flats.

2. Basic Conditions for the Formation of Tidal Flats

Tidal flats are formed in areas where there is an important supply of fine-grained sediment (i.e., clays, silts, and fine to very fine sands), and that tides and tidal currents dominate over other hydrodynamic forces (Klein, 1985). The first condition is satisfied for most coastal environments: fine-grained sediment is delivered by rivers and discharged into estuaries and adjacent coastal areas, erosion on the seabed and cliff recession provide additional sources of sedimentary materials, and organisms living in coastal waters and salt marshes produce shell debris and particulate organic matter. The second condition determines whether or not fine-grained sediment will be deposited on the tidal flat. Several factors influence this condition.

Firstly, the tidal action should be significant. The average tidal range (R) has been used in the classification of the coast, that is, the coast can be microtidal ($R < 2$ m), mesotidal ($R = 2$–4 m), or macrotidal ($R > 4$ m) (Davies, 1964). On a microtidal coast, tidal currents are relatively weak unless the slope of the seabed is very small. In mesotidal and macrotidal environments, tidal currents tend to be relatively large compared with microtidal coasts, which favors the formation of tidal flats. It should be noted that tidal flats can be formed in microtidal areas of sheltered coastal embayments or semienclosed seas; this is because in such environments the wave action is of only a secondary importance compared with tidal currents.

Secondly, the dominance of tidal action should be understood in a relative sense: tidal flats cannot be formed where wave action dominates, even if the tidal range is large. On open coasts, waves may represent a dominant force, which break in the surf zone, causing transport of fine-grained sediment toward offshore areas (King, 1972). In this case, if the supply of fine-grained materials is not abundant, then sandy or gravelly beaches will be formed, rather than tidal flats. However, if the rate of fine-grained sediment supply is high, then the accumulation of fine-grained materials reduces the bed slope in the intertidal area. Eventually, wave breaking can rarely occur on the bed, that is, the wave energy is dissipated over the wide intertidal area due to bed friction, without causing breaking. The reduction of the bed slope will lead to enhanced tidal currents. Thus, in response to fine-grained sediment accumulation, wave action is weakened and the tidal action is enhanced. However, this observation does not imply that waves are unimportant on tidal flats; waves are important in the transport of sediment and the shaping of the tidal flat morphology. *In situ* measurements have shown that combined wave–tide action (without wave breaking) can cause intense sediment movement on the flat (Fan et al., 2006; Wang et al., 2006).

Finally, while tidal action is a dominant agent, storm events can significantly modify the tidal flat environment. For example, during a typhoon event, storm surges become temporally the dominant forcing for sediment erosion, transport, and accumulation on a tidal flat (Ren et al., 1985; Ren, 1986; Andersen and Pejrup, 2001).

The arguments outlined above imply that tidal flats may develop in sheltered tidal estuaries and coastal embayments where there is continuous supply of fined-grained sediment (although the rate of supply may be small), or on open coasts where tidal range is sufficiently large and sediment supply is abundant. A classical example of the former is the tidal flats in the Dutch Wadden Sea (van Straaten and Kuenen, 1957, 1958). Here, the tidal flat areas are sheltered by a series of barrier islands, the exchange of water between the Wadden Sea and the open North Sea being via tidal inlets cutting through the barrier islands. The sediment source is provided mainly by the North Sea (Postma, 1961; Pejrup et al., 1997), while river input represents a secondary source. For the open coast tidal flats, a typical example is the tidal flats on the Jiangsu coast, eastern China (Ren, 1986). During the Holocene period, because of the abundant sediment supply from two large rivers (i.e., the Changjiang and the Yellow rivers), extensive tidal flats have been formed.

3. Zonation in Sedimentation and Flat Surface Morphology

3.1. Vertical sediment sequences

Sediment cores taken from the upper part of the intertidal flat tend to show a "fining upward" sequence (Klein, 1985). Although the sediment deposited is ultimately determined by the source characteristics, in most cases the sediment source for tidal flats contain materials from sand- to clay-sized materials. Sandy materials tend to accumulate in the lower part of the flat, whilst muddy materials deposit over the upper part of the flat. The top part consists of clayey or muddy materials, often with high organic carbon content. Below this layer is a mud layer corresponding to the elevation near the high water, with very thin laminae as a major type of sedimentary structure. Then there is a mixed sand–mud layer, interbedded with sandy and muddy materials; such sedimentary structures are known as "tidal bedding" (Reineck and Singh, 1980). The lowest part of the core is a sand layer, corresponding to the lower part of the intertidal zone and the subtidal zone.

Such a vertical sequence reflects the spatial distributions of sediments on tidal flats. A general pattern is that along a transect from the supratidal zone to subtidal zone there are salt marshes, mud flats, mixed sand–mud flats, and sand flats (Figure 2a). On the salt marshes, the bed material is the finest, with organic matter being derived from marsh plants and organisms. The mud flat is covered with clay and fine silts; this area is often located between the high water levels on springs and on neaps. In some regions, salt marsh vegetation may extend from the supratidal zone into the mud flat. Mixed sand–mud flats are close to the mean sea level; here, sands are deposited during spring tides and muds are deposited on neaps (for details, see Section 4.2). Over the lower parts of tidal flats, well sorted sands are present, with various types of bedforms (e.g., dunes and ripples). On the tidal flats tidal creeks may be formed, consisting of small creeks and large tidal channels. Sedimentation in tidal creeks may be different from that on the adjacent flat surface,

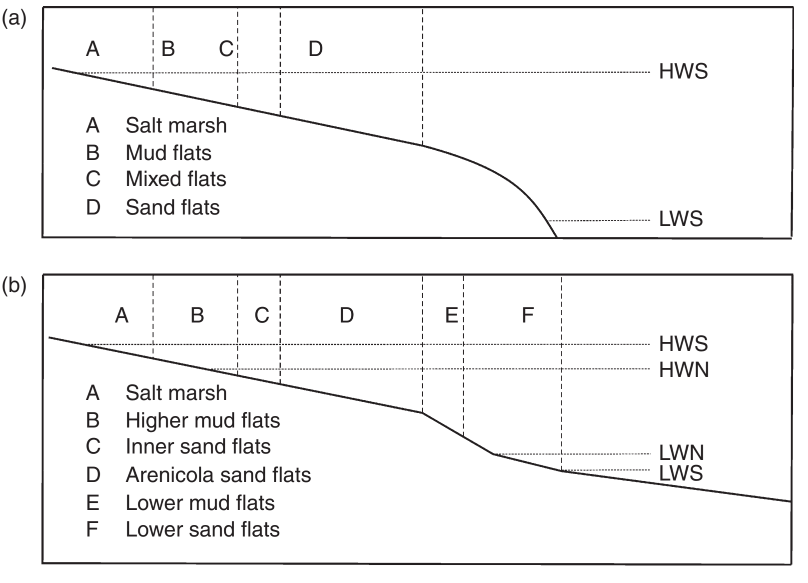

Figure 2 General patterns of (a) zonation of tidal flats (Reineck and Singh, 1980) and (b) the geomorphic and sedimentary features of the Wash, England (Evans, 1965). HWS, high water on springs; HWN, high water on neaps; LWN, low water on neaps; LWS, low water on springs.

but the general "fining-upward" pattern is also maintained for the creek systems. In some tidal flat environments, the zonation can have slightly different patterns, as indicated by Figure 2b. The reason for this is that, in addition to bed elevation, other factors such as bed slope and tidal water level also influence the transport and accumulation of sediments (see below).

The zonation of tidal flats implies that the finest materials tend to accumulate as the upper part of the sedimentary sequence. This raises a question of the thickness of the mud layer on the upper part of the tidal flat. Data sets from the different parts of the world (Table 1) show that the thickness varies considerably: in some places the entire profile is covered with muddy sediments, whilst on the other extreme mud is absent. Apparently, the correlation between the thickness and tidal range is poor (cf. data listed in Table 1). In Section 4, an explanation about the factors that control the thickness of the mud layer will be given on the basis of sediment dynamic analysis, for the tidal flats that are associated with a supply of different types of sediments ranging from clays to fine sands.

3.2. Sediment and morphology on intertidal mud flats

Generally, mud flats are located on the upper part of the intertidal zone, with very gentle bed slopes (Figure 3). Here, the sediment is the finest for the entire tidal flat, and deposition takes places because of the settling of fine-grained materials from the water column. Therefore, although there are tidal cycle changes in grain size, mud (i.e., a mixture of clayey and silty sediments) is the major component of mudflat sediments. The deposition of coarser materials, which occasionally occur in the mudflat sediment layers, is due to extreme events (e.g., storm surges). The sedimentary record consists of laminated mud, with alternating silty and clayey layers of less than 1 mm in thickness. Because resuspension of the bed material is relatively weak here, the continuity of the mud deposit is high compared with the other parts of the tidal flat. However, in some places intense bioturbation often causes destruction of the original sedimentary structure.

The landward part of the mudflat may be covered by pioneer plants extending from the salt marshes in the supratidal zone, or even covered with salt marshes. On the Jiangsu coast, eastern China, for example, extensive *Spartina* marshes are formed over the landward part of the mud flats, with an upper limit close to mean high water on springs and a lower limit slightly lower than high water on neaps (Zhang et al., 2004).

Under tidal action alone, accretion becomes progressively slower when the bed elevation is enhanced due to sediment accumulation (Pethick, 1981); thus, the high water level on springs represents the upper limit for the accretion (Amos and Mosher, 1985; Amos, 1995). However, the bed accretion continues beyond the high water level, to form the supratidal zone. There are several processes for this phenomenon. First, during storm events the water level can become much higher than during normal tidal cycles, especially when the surge coincides with an astronomical spring tide. During such events, a large amount of sediment may be transported to the upper parts of the flat and deposits

Table 1 Thickness of the sediment layer above the mixed sand–mud flat ($<R>$ = average tide range, R_S = spring tide range, R_N = neap tide range, H_m = thickness of mud layer, W = width of intertidal zone)

Location	Tide range (m)			H_m (m)	W (km)	Reference
	$<R>$	R_S	R_N			
Wadden Sea (The Netherlands)	1.3–2.8			2.0	7–10	van Straaten (1961)
The Wash (England)	5.0	6.5	3.5	1.5–2.0	1.0–6.5	Evans (1965)
Colorado River Estuary (USA)	4–5	6–8		>8		Ginsberg (1975)
Wadden Sea (Germany)		2.6–4.1	1.8–3.1	2.5	5	Ginsberg (1975)
Fraser River Estuary, Boundary Bay (Canada)	2.7	4.1	1.5	<0.5		Ginsberg (1975)
Central Jiangsu coast (China)				3	7–10	Ren (1986)
Salmon River Estuary, Bay of Fundy (Canada)	11.9	15.2	8.7	4	5	Dalrymple et al. (1990)
Haenam Bay (South Korea)	3.0	4.0	1.8	>4	2–2.5	Lim and Park (2003)
Newtownards (Northern Ireland)	3.0	3.5		0	0.5–1.2	Malvarez et al. (2004)
Baeksu Tidal Flat (South Korea)	3.9			<0.5	4–6	Yang et al. (2005)

Figure 3 Bed features of the bare mud flat, Jiangsu coast, eastern China.

there. Thick storm deposits have been identified in the upper tidal flat sequences (Ren et al., 1985). Second, animals living in the flat environment modify the bed morphology through bioturbation, generating unique sedimentary structures (Reineck and Singh, 1980). Biological activities produce uneven bed morphology (e.g., mud mounds formed by crabs and mud skippers): the elevated features may become above the high water, and the depressions will receive an increased amount of sediment during high water. Finally, the deposition of aerosol forms an additional sediment source (Li et al., 1997), which is not constrained by the high water mark.

Tidal creeks are present on the mudflat. On the bare mud flat, the tidal creeks have a relatively small depth to width ratio, with lateral migrations.

At the boundary between the mudflat and salt marshes, cliffs or scarps tend to occur, with a height of less than 1 m (Reineck and Singh, 1980). In some places much higher scarps can be formed, for example, those found in the Severn Estuary, UK, which has been interpreted as periodic erosion and accretion cycles in response to changes in hydrodynamic and sediment supply conditions (Allen, 1989). However, disagreement exists with regard to the significance of the low cliffs that are more widely distributed. Some researchers believe that the cliffs represent an indication of coastal erosion, but others argue that they result from localized scour. Observations show that because the accretion on the marsh that is more rapid than the adjacent bare flat (the flat with plants traps more sediment than the bare flat, and the organic matter further adds to the sedimentary materials in the marsh), the bed elevation gradually becomes different (Amos, 1995). Then, the edge of the marsh becomes progressively steeper, causing concentration of wave energy (the small waves would otherwise not break at this location).

In coastal embayments, such low cliffs are often associated with low hydrodynamic forcing and low sediment supply. For example, in Christchurch Harbour, southern England, where the deposition rate has been higher on the marsh than on the bare flat. Subsequently, the small waves formed in the estuarine waters (with a fetch of only 1 km) started to erode part of the materials at the marsh edge, forming low cliffs of 0.4–0.6 m in height (Gao and Collins, 1997). Numerical model output (Gao and Collins, 1997) indicated that the action of small waves, which break at the marsh–bare flat boundary because of the enhanced bed slope at this location, is responsible for the localized scour and the cliffs can be stable or retreat slowly. In the German and Danish Wadden Sea, marsh cliff erosion was measured by Pejrup et al. (1997) and similar results were obtained. Elsewhere, measurements in Rehoboth Bay, USA, revealed that the rate of cliff recession is also small, ranging between 0.14 and 0.43 m/year (Schwimmer, 2001). Hence, it is the evolution of the tidal flat itself that creates the condition for the formation of the cliff.

3.3. Sediment and morphology on mixed sand–mud flats

Mixed sand–mud flats are characterized by alternating deposition of muddy materials on neaps and sandy material on springs. The spatial distribution of this morphological unit on the tidal flat depends upon a number of factors, including tidal regime, sediment supply, and suspended sediment concentration of seawater. On the Jiangsu coast, where the tidal currents during the flood and ebb maximum periods exceed the threshold for bedload transport over the entire spring–neap tidal cycle, the mixed flat is located between high water on neaps and mean sea level (Zhu and Xu, 1982).

In the sediment sequences, typical "tidal bedding" (i.e., interlayered relatively coarse- and fine-grained sediments), is found on the mixed sand–mud flats. These beddings contain information on the sedimentary processes during flood–ebb cycles and spring–neap cycles (Dalrymple et al., 1990). A well-preserved sediment record at this location will show that the thickness of the sand layer decreases with decreasing tidal range from the springs toward the neaps. Likewise, the mud layer increases its thickness toward the neap tides. However, the preservation potential for the sand–mud flat is usually not high, and most of the sedimentary record formed in tidal cycles is destroyed subsequently by reworking or resuspension of the bed materials (Fan et al., 2002; Deloffre et al., 2005, 2007; Gao, 2007a). On the bed, scour features are present, resulting from sediment reworking (Figure 4a,b). Generally, on the lower part of bare mud flats and mixed mud–sand flats, a part of the muddy material accumulated on neaps can survive until the next neap tidal phase. This is the reason for the formation of the scour features. It should be noted that such scour does not indicate long-term net erosion of the bed; on the contrary, the accretion rate on the mixed sand–mud flat is actually higher than the other parts of the tidal flat system (Gao and Zhu, 1988). The scour represents only short-term effects within a long-term accretion trend.

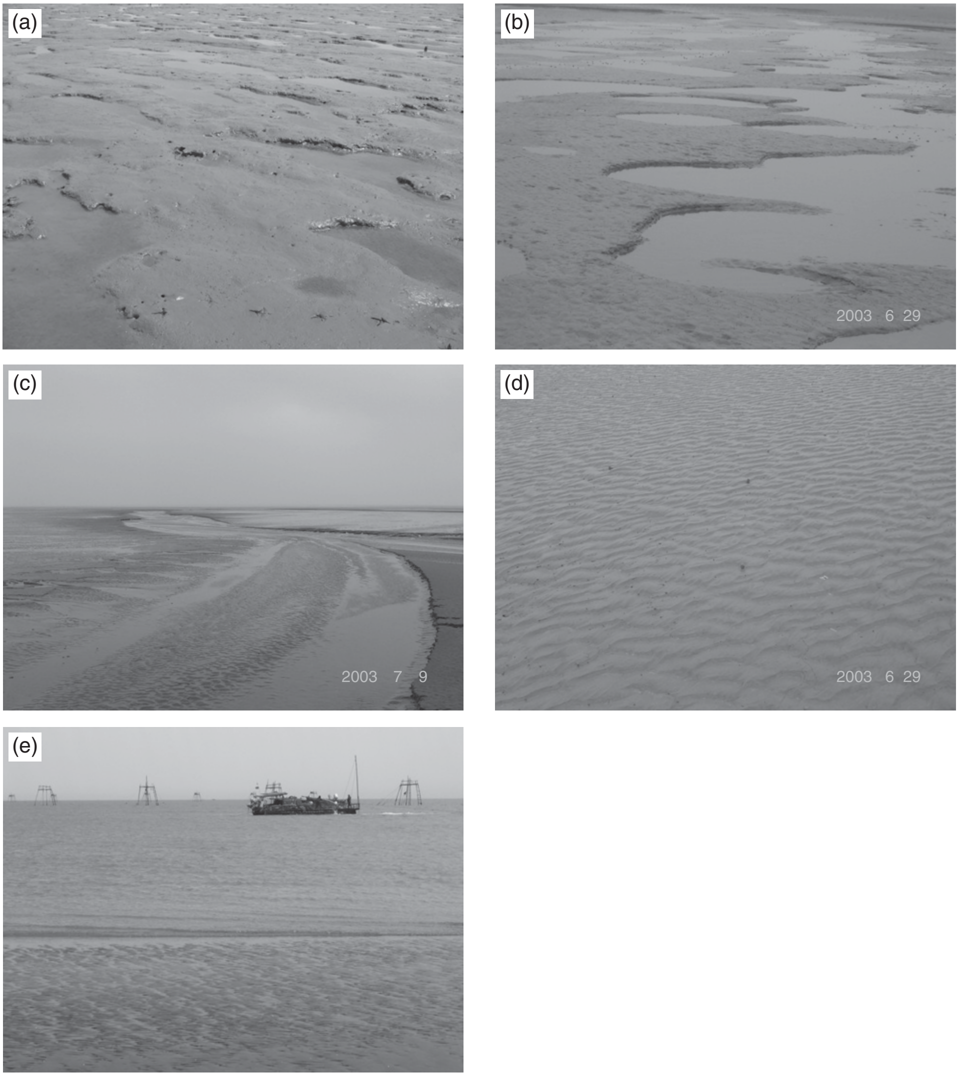

Figure 4 Geomorphic and sedimentary features of the middle to lower parts of the intertidal zone on the Jiangsu coast, eastern China: (a) scour features on mud flats; (b) scour features on mixed mud–sand flats; (c) a tidal creek on mixed sand–mud flats; (d) sand flat bed, with current ripples; and (e) sand flat near the low water mark.

Tidal creeks are highly dynamic on mixed sand–mud flats. For instance, on the central coast of Jiangsu Province, China, the strong currents in combination with the silty sediments on such flats result in rapid lateral migration of the tidal channels. Channel migration generates new tidal creeks, which can extend toward the upper part of the intertidal zone rapidly, in the form of headward erosion. During storms, the tidal creeks are even more active; significant deepening of the channel bottom, development of many new channels, and rapid migration may be observed (Ren, 1986).

3.4. Sediment and morphology on sand flats

In a typical tidal flat system with abundant sandy sediment supply and strong tidal currents, sand flats occupy the lower part of the intertidal zone. The sands are normally well sorted. Tidal creeks on the sand flat are relatively wide, with a small depth to width ratio (Figure 4c). The migration of the tidal channels is active, and is influenced by bedload transport and accumulation. Lateral migration can be caused by bedload movement in the longshore direction (i.e., transport parallel to the shoreline). Cross bedding is a common type of sedimentary structure in tidal creek sequences.

During the late stages of the ebb, before the bed is exposed to air, the tidal currents are too strong for the fine-grained, suspended sediment to settle onto the bed, and the material settled during the flood slack tends to be suspended during the ebb. Numerical calculations show that net accumulation of mud on the sand flat is possible only when the suspended sediment concentration is extremely high (Amos, 1995), in which case sand flats with pure sand deposition cannot be formed.

Where the bed slope is sufficiently large (1.0×10^{-3} or greater) to allow rapid draining of water mass, the geometry of bedforms (dunes and ripples) is preserved and exposed during low tide (Klein, 1985; Dalrymple et al., 1990). The dunes have a wavelength (or spacing) of 0.6–6 m, whilst the ripples are small bedforms (wavelength <0.6 m). Compared with the dunes, ripples are more extensively distributed. Field observations show that during the high water slack ripples are formed by combined wave–current action; during the ebb the upper part of the sand flat will be exposed rapidly and the morphology of wave ripples may be preserved, but the lower sand flat can be modified by ebb currents to form ebb-oriented, flow-generated ripples (Amos and Collins, 1978). If the bed has an extremely gentle slope, then these bedforms may be modified by water flow generated by seepage from the substrate; because the flow depth is very small (i.e., $<10^{-1}$ m) and the Froude number is large (i.e., close to 1), plane bed without bedforms can be formed in a short period of time (i.e., 1–2 h). On the Jiangsu coast, eastern China, the central parts of the sand flat (with a mean grain size of 0.06–0.1 μm) have a bed slope of around 0.0005, and the slope increases to more than 0.001 toward either the low water mark or the boundary between the sand flat and the mixed sand–mud flat (Zhu and Xu, 1982). Therefore, during low water on springs, wave ripples are present near the boundary between the sand flat and the mixed sand–mud flat, plane beds are present over the central part of the sand flat, and current ripples are found over the lower sand flat (Figure 4d,e).

4. Factors and Processes

4.1. Influences of quantity and composition of sediment supply

The evolution of present-day tidal flats began when sea level reached its highest elevation during the Holocene period, from a base that was left from the last glacial period. Thus, the accommodation space for the tidal flat deposit is confined by the high water mark, the original topography/bathymetry and the tidal flat profile. While the high water level and the topographic baseline are influenced by sea-level changes and crustal movements, the rate of shoreline advancement is determined by the amount of sediment supply, and the shape of the profile is influenced by the grain size composition of the materials supplied.

As shown in a simplified shore-normal transection for the geometry of a tidal flat sediment system (Figure 5), the increase of the tidal flat sediment body through time is associated with upward accretion and the advancement of the tidal flat profile toward the sea. The vertical accretion rate of a tidal flat can be related to the portion of sediment supply that contributes to the tidal flat formation, by:

$$D = \frac{\Delta S}{\Delta L} \frac{\sin(\beta - \alpha)}{\sin\alpha \sin\beta} \tan\beta \tag{1}$$

where D is the vertical accretion rate averaged over the entire profile, ΔS is the proportion of sediment supply that is deposited on the flat per unit time over unit length of the shoreline, α is the slope angle of the original topography, β is the average bed slope angle of the flat profile, and ΔL is the distance of shoreline advancement during a unit period of time.

In addition to the accretion rate, the thickness of the mud layer of the upper part of the sequences is also an important parameter for the tidal flat. Here, the mud layer represents the deposits of the mud flat and the mixed sand–mud flat (see Sections 3.2 and 3.3). Assuming that most of the muddy materials are deposited on the upper part of the tidal flat, as is the case for the Jiangsu coast, this thickness can be expressed approximately as (cf. the geometric relationship shown in Figure 5):

$$H_{\mathrm{m}} \approx \frac{\Delta S_1}{\Delta S} (L + \Delta L) \tan\alpha \tag{2}$$

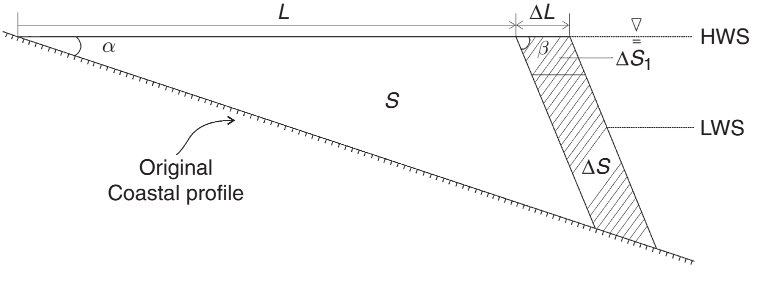

Figure 5 Schematic diagram showing a shore-normal cross section of tidal flat sedimentation (S = volume of sediment accumulated over unit length of the shoreline, L = the width of coastal plain formed by tidal flat sedimentation, ΔS = the amount of sediment supply per unit time over unit length of the shoreline, ΔL = the distance of shoreline advancement during a unit time, α = the slope angle of the original topography, and β = the average bed slope angle of the flat profile).

where H_m is the thickness of the mud layer, ΔS_1 is the mud fractions within ΔS, and L is the total width of coastal plain formed by tidal flat sedimentation. Equation (2) implies that the thickness is related to the composition of sediment supply and the stage of the tidal flat evolution (since it is a function of ΔS_1 and L), but it is not directly related to tide range. Observations of the world tidal flats have provided support to this inference (cf. Table 1): the thickness of the mud layer does not increase with the increasing tide range. In an extreme case (Malvarez et al., 2004), the profile is covered entirely with mud, because in this system the supply of sandy sediments is very small. On the Jiangsu coast, the sediment input from the Yellow River has relatively stable composition; as a result, there is a trend of increase in the thickness of the mud layer during the tidal flat development (Gao, 2007a). Such an observation is also consistent with the prediction by Equation (2).

The thickness of the mud layer determines the boundary on the tidal flat where sandy and muddy deposits are divided. At this boundary, the bed slope can be expressed as a function of the threshold for the initial motion of the sands, as demonstrated below. At any site over the intertidal flat, the instantaneous tidal current velocity is a vector, which consists of an onshore–offshore component, u, and a longshore component, v (Anderson, 1973; Perillo et al., 1993; Wang et al., 1999). The onshore–offshore component is controlled by the water level changes and the bed morphology, on the basis of the principle of mass conservation (Zhu and Gao, 1985; Wang et al., 1999):

$$u = \frac{1}{\tan\beta}\frac{\mathrm{d}h}{\mathrm{d}t} \tag{3}$$

where $\tan\beta$ is the average bed slope over the inundated section and h is the tidal water level.

At the sand–mud boundary, maximum current speeds during the flood should not exceed the threshold for initial bedload motion; otherwise, sandy materials will be transported across the boundary further toward the upper tidal flat. At the same time, it should not be smaller than the threshold; otherwise, sandy materials cannot be transported to this location. Hence, there is only one possibility: the sand–mud boundary is located where the maximum shore-normal current speed is equal to the threshold:

$$u_{\mathrm{max}} = u_{\mathrm{cr}}\sin\theta = \frac{1}{\tan\beta_{\mathrm{b}}}\frac{\mathrm{d}h_{\mathrm{b}}}{\mathrm{d}t} \tag{4}$$

where θ is the angle between the current direction and the longshore direction, the subscript b denotes the sand–mud boundary, and $\mathrm{d}h_{\mathrm{b}}/\mathrm{d}t$ denotes the rate of water level change when the flow reaches the boundary. Thus, the slope at the sand–mud boundary can be defined by

$$\tan\beta_{\mathrm{b}} = \frac{1}{u_{\mathrm{cr}}\sin\theta}\frac{\mathrm{d}h_{\mathrm{b}}}{\mathrm{d}t} \tag{5}$$

The threshold current speed for initial sediment motion is a function of near-bed shear stress. Equation (5) implies that a steep bed slope at the sand–mud boundary is associated with a large value of dh_b/dt, or a low elevation for the sand–mud boundary (since the rate of water level change cannot be large near the high water mark). Because the thickness of the mud layer is the vertical difference between the elevation at the boundary and the high water mark, the steep slope also means a lager thickness of the mud deposit in the tidal flat sequence. This observation may explain the phenomenon that on a tidal flat with a thick mud layer maximum slope gradients occur on the mixed sand–mud flat (Zhu and Xu, 1982).

The elevation associated with the critical current speed differs between the spring and neap conditions. As a result, there will be two critical elevation values, one for neap tides and the other for spring tides. Between the two critical elevations, mud is deposited on neaps and sand is deposited on springs. Therefore, the thickness of the mud layer (consisting of mud and mixed mud–sand deposits) is related to the lower critical elevation for the neap tides, and the mixed sand–mud flat itself is a result of spring–neap tidal cycles.

4.2. Sedimentation during tidal cycles

On a tidal flat, why are the sediments transported toward the land? Two mechanisms have been identified, one for suspended load and the other for bedload. The movement of the suspended load is related to a physical mechanism known as "settling and scour lag effects" (Postma, 1954; van Straaten and Kuenen, 1957, 1958). The basic condition for these effects is the particular patterns of current speed variations over the tidal flat. During a tidal cycle, the rate of water level changes and, according to Equation (3), the current speed also changes. Minimum rates of water level change occur during high and low water periods, and maximum rates appear at the middle of flood or ebb phases. Consequently, only the middle and lower parts of the intertidal zone will experience large tidal currents, and the upper part is associated with weak currents. In such an environment, it will be difficult for the suspended material originated from subtidal areas to permanently stay on the middle and lower parts of the tidal flat, unless the concentration is sufficiently high (Amos, 1995).

The "settling and scour lag effects" explain why a suspended particle is carried by currents toward the upper tidal flat, as shown in Figure 6. On the tidal flat, P represents the location above which the tidal current speed never exceeds the critical value for resuspension, and below which the speed is below the critical value only during the slack periods. Let us suppose that a water particle and a suspended sediment are located at the same site, A, at the beginning of a flood tide (Figure 6a). During the flood, when the current speed decreases to the threshold at location B (at this time the water level reaches the elevation of site P), the sediment particle starts to settle, but it cannot reach the bed at location B (Figure 6b). It may reach the bed at a site, C, which is further toward the high water mark (Figure 6c). Such an effect is called "settling lag." During the ebb, when the water particle reaches the location C, its speed is still below the threshold for resuspension, because at this time the water is above site P (Figure 6d). Resuspension for the sediment in consideration does not occur until the water particle arrives at B

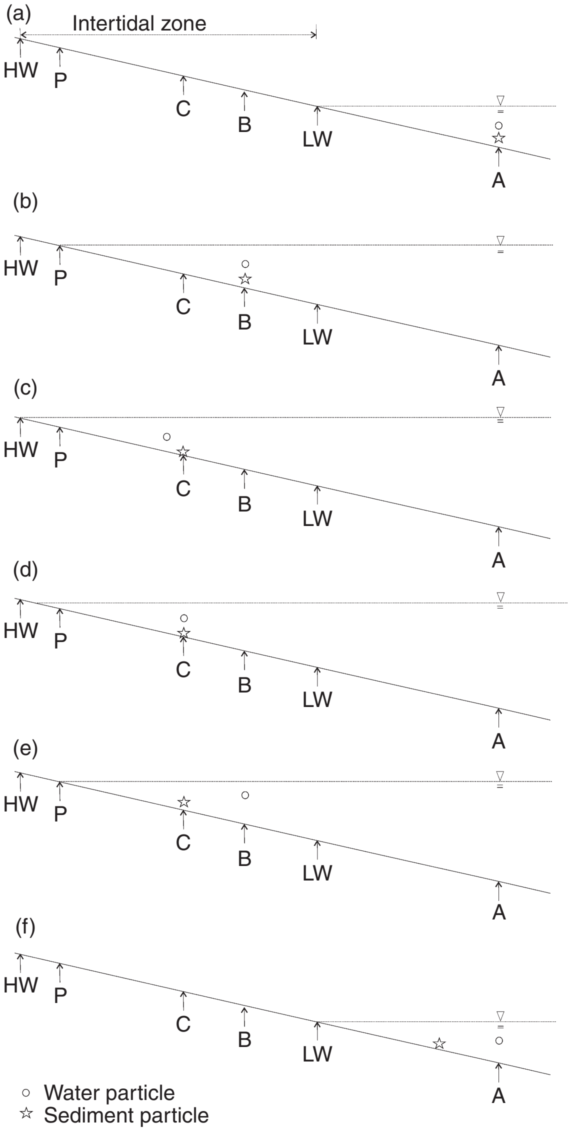

Figure 6 Diagram showing the settling and scour effects, by comparing the tidal cycle movement of a water parcel and a sediment particle at the stages of (a) low water (representing the beginning of the tidal cycle); (b) water level reaching the point *P* when the current speed is reduced to the critical value for bed erosion; (c) high water; (d) the water parcel reaching the site C where the sediment particle is settled to the bed; (e) water level reaching the point *P*; and (f) low water (the end of the tidal cycle).

(Figure 6e). At the slack water of the ebb, the positions for the water parcel and sediment particle becomes different (Figure 6f). The process described in Figure 6d–f is known as "scour lag." Thus, during each tidal cycle, the net transport is directed toward the high water mark. Although some biological and biogeochemical processes also influence the transport of suspended materials (Neumeier and Amos, 2006), the lag effects represent a basic physical mechanism.

For the bedload, the time–velocity asymmetry patterns caused by the deformation of tidal waves favor landward transport. Over the tidal flat, shallow tides (known as "over tides") are generated by seabed friction. As a result, the flood duration becomes shorter and the ebb duration becomes longer on the flat surface (Zhang, 1992). At the same time, the peak current speed during the flood becomes larger than during the ebb. In addition, the transport rate for bedload has a nonlinear relationship with the current speed (Hardisty, 1983; Wang and Gao, 2001). Thus, it has been shown mathematically that the transport capacity during the flood is greater than during the ebb (Zhu and Gao, 1985). This indicates that, if sufficient bedload is available, then net transport will be directed to landward during a tidal cycle. However, such a pattern may not be observed if the sites for measurements are close to tidal creek systems.

The landward transport of bedload is important for the maintenance of accretion of tidal flats. Otherwise, the accumulation of fine-grained materials over the mudflat will lead to a narrower intertidal zone and reduce the average bed slope. This will, in turn, reduce the tidal current speed according to Equation (3). If the tidal dominance disappears, then development of tidal flats cannot continue. Thus, it is crucial for sands to be deposited over the lower parts of the tidal flat; only in this way can the small bed slope of tidal flats be maintained.

4.3. Long-term accretion–erosion cycles

Sediment supply is a necessary condition for the growth of tidal flats. When the supply is cut off, or it becomes too small, erosion will occur. The coastline near the old Yellow River Delta, in northern Jiangsu Province, China, is a typical example which indicates the effect of sediment cut off. Before 1855, the sediment discharge of the Yellow River formed a large delta, and tidal flats were well developed. Then, in 1855 the Yellow River shifted its course to discharge into the Bohai Sea in northern China. Since then the shoreline in northern Jiangsu has been retreating. The original tidal flat has been modified in terms of sediment distribution and profile morphology. The width of the intertidal zone has been reduced to less than 2 km, sandy materials and mud pebbles are found near the high water mark. The overall profile shape is approaching to a wave-dominated beach profile (Gao and Zhu, 1988).

Such responses can be explained by sediment dynamics. Because the supply is reduced, the landward transport capacity cannot be satisfied. However, at this time, the transport capacity during the ebb remains (i.e., the tidal flat surface is now transformed from a sediment sink to a sediment source). The removal of sediment from the lower part of the flat results in reduction in bed slope and in the strength of tidal currents on the flat. Eventually wave action becomes a dominant factor. Wave breaking takes place on the shore face, and the fine-grained material is transported toward deeper waters, just as observed on a sandy beach. During shoreline recession, if the sediment strata contain shells and other coarse-grained materials, then this debris may form cheniers (Augustinus, 1989; Wang and Ke, 1989). Thus, the presence of cheniers in tidal flats is indicative of coastal erosion periods. Often such cycles are related to sediment supply changes.

Even if the sediment supply is maintained, there may still be a limit of growth for the tidal flat sedimentary system. The growth of a river delta may represent an analogue: in response to reduction in the Sediment Retention Index (SRI, the ratio of the sediment permanently retained in the system to the total sediment input provided by fluvial, marine, atmospheric, and sources) when the delta progrades toward deeper water areas, the rate of delta growth will decrease (Gao, 2007b). It is likely that similar processes are associated with the tidal flats, that is, they have a limited space for its development and after their growth over the Holocene period they may be already approaching the growth limit.

4.4. Tidal creek systems

Tidal creeks have a different function compared with the flat surface. During the flood there exist several velocity maxima in a creek, in response to rapid enlargement of the inundated area on the flat adjacent to the creek, and during the ebb extra water enters the creek from the surface or through seepage from the bed (Bayliss-Smith et al., 1979; Wang et al., 1999). As a result, water balance in the creek can be asymmetric: the water discharge during the ebb is larger than during the flood. Observations demonstrate that net sediment transport in creeks tends to be seaward, especially during spring tides (Yang et al., 2003). This pattern may be further enhanced when storm occurs. Therefore, it can be inferred that tidal creeks reduce the overall accretion rate of the tidal flat.

Each tidal creek system occupies a "drainage basin" on the tidal flat (Zhang, 1992). In macrotidal environments, tidal creeks may migrate laterally intensively, forming various sizes of point bars. On the sand flat, the creek channel migration is affected by bedload transport in the longshore direction, as observed on the Jiangsu coast (Wang et al., 2006). However, the migration of the creek is confined with the drainage basin.

5. Summary

The basic sedimentological and geomorphological characteristics of tidal flats may be summarized as follows.

1. Tidal flats are formed under the condition that tides dominate over other hydrodynamic processes. They have a significant pattern of zonation, with a general pattern that from the supratidal zone to subtidal zone salt marshes, mud flats, mixed sand–mud flats and sand flats are distributed. Such a pattern may be modified by other factors such as tidal creek formation and sediment supply. In a sediment core from an upper part of the tidal flat, the zonation is reflected by a "fining-upward" sequence.
2. Settling and sour effects are responsible for the transport and accumulation of muddy sediments over the upper parts of tidal flats, whilst the deformation of tidal waves causes landward transport of sandy materials, with the upper limit

of sand accumulation being controlled by the tidal current speed during the flood. In addition to these physical mechanisms, biological effects on the suspended sediment transport are also important.
3. Some characteristics and/or parameters of tidal flats (e.g., the bed slope gradient, sedimentary structures, salt marsh cliffs, scour over the mixed sand–mud flat, the thickness of the mud layer, the tidal creeks, and the sediment retention index) contain important information on the system behavior and evolution of tidal flats. Such information may be obtained by means of sediment dynamic analysis.

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