*Geomorphology and Sedimentology of Estuaries. Developments in Sedimentology 53*
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*Chapter 14*

# SEDIMENT TRANSPORT PROCESSES IN ESTUARIES

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## INTRODUCTION

Estuaries are the route by which sediment is transported from the rivers to the sea. On the way down the rivers the grain size distribution of the sediment becomes altered by continual deposition, re-erosion and transport. Much of the coarser sediment can become trapped on the flood plains of the rivers, only being released at times of flood. The finer fractions are transported into the estuary. There the estuarine processes act as a filter on the sediment input, and mixing can take place with sediment brought in from the sea. Additionally, chemical alterations can occur within the estuary that can cause the surface properties of some of the constituent particles to alter, affecting their pollutant scavenging potential, and their potential deposition.

Sediments form a crucial control in many estuarine processes. Within the estuary suspended sediment concentrations are generally high, the particles are fine, cohesive, and prone to flocculate, and they are richly organic. The energy cycles inherent in the semidiurnal and lunar tide and in seasonal changes cause continual erosion, transport and deposition. Thus, even when little new sediment is coming in, sediments can silt up harbours and navigational channels. Since the sediments are fine grained clay minerals, pollutants are absorbed on their surfaces and are transported with the sediment particles. Consequently the transport and dispersion of contaminants can only be understood through a knowledge of the movement of particles. When the suspended sediment concentrations are sufficiently high, light penetration and productivity can be limited. The muddy substrates can be host to a diverse and vigorous biological community, but this can be limited by the presence of layers of high concentration suspended sediment with low oxygen content intermittently present above the bed.

Estuaries in general are shallow, and sea level undergoes very drastic changes on the geological timescale. Thus, they are ephemeral features being fairly rapidly altered and destroyed, having an average life of probably only thousands of years. It is likely that the world is rich in estuaries at the present time because of the sea level rise of the Flandrian transgression. This inundated the valleys cut when the rivers incised to a base level, which reached a minimum level of about $-100$ m during the closing stages of the Pleistocene Ice Age. The variation in form of the resulting estuaries depends on the volumes of sediment that the rivers have contributed to fill the valleys, as well as that brought in from the sea. Deltas form where river flow and sediment discharge is high; the valleys have become completely filled and the sediment discharges directly into the sea. They are normally present in areas of high seasonal discharge, particularly in the tropics, in monsoon areas, and those with a high component of snow melt discharge. Generally deltas are best developed in areas where the tidal range is small and where the currents cannot easily redistribute the sediment the rivers introduce.

Where sediment discharge is less, the estuaries are unfilled. These drowned river valleys, or coastal plain estuaries, still retain the main features of river valleys, having a meandering outline with frequent tributaries and a triangular cross-section.

In glaciated areas the river valleys were over-deepened by glaciers. A characteristic of the resulting fjords is a rock bar or sill at their mouths that can be as little as a few tens of meters deep. Inside, however, they can be several hundred meters deep and extend hundreds of kilometres inland. The sill restricts the water circulation and isolates the interior deep basin, with obvious consequences to the sedimentation patterns.

On low coastlines, extensive shallow lagoons are often formed between the rivers and the sea. In tropical areas the lagoons can be hypersaline because of evaporation during the hot season, but almost entirely fresh water during the rainy season. During low river flow periods the mouth may even be closed by littoral drift. The mouths of these features are often called inlets.

With such a variety of estuaries it is to be expected that there will be a diverse and complex series of processes dominating the transport and deposition of the incoming sediment.

## TIDAL EFFECTS

The tidal elevation characteristics of estuaries create important distinctions in the capability of the currents to move the sediments. Davies (1964) has classified estuaries as microtidal, where tides are less than 2 m range; mesotidal, between 2 m and 4 m range; macrotidal, greater than 4 m range, to which a fourth category can be added; hypertidal, greater than 6 m range.

Within the estuary the tide can be greatly modified by the friction of the bed on the current, and by the funnelling effect of the convergence of the estuary sides. The convergence causes both a partial reflection of the wave, as well as squeezing it into a smaller cross-section, thereby increasing the height of the tidal wave. Friction increases with decreasing water depth, and with increasing velocity, thereby taking energy from the tide and decreasing its amplitude.

Where the convergence effect exceeds the frictional effect, the tidal range increases towards the head of the estuary, before decreasing in the riverine section (Fig. 14-1). This response is termed hypersynchronous. When convergence and friction are equal, the tidal range is constant throughout the estuary. This is the synchronous response. Hyposynchronous estuaries are those where friction dominates, and where the range of the tide diminishes throughout the estuary.

When the tidal range is large relative to water depth, considerable asymmetry can occur in the tidal curve and in the velocities. On the flood tide the water is flowing into a decreasing cross section as it flows up the estuary. Friction slows down the

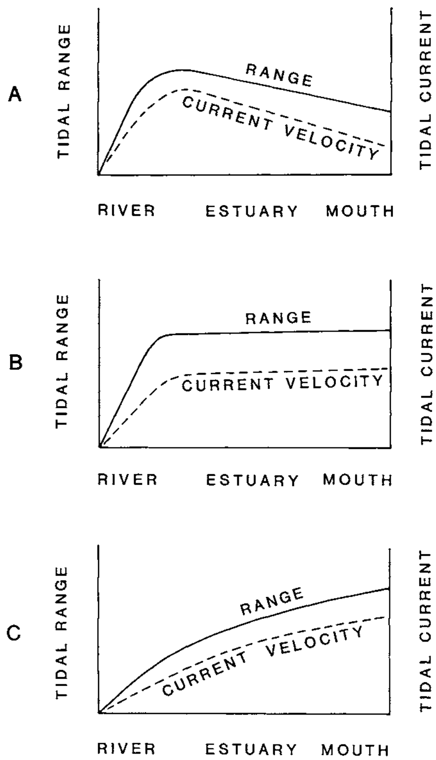


Fig. 14-1. Tidal response in estuaries. A) Hypersynchronous, B) synchronous, and C) hyposynchronous (after Nichols and Biggs, 1985).

early part of the flood tide more than later when the water depth is greater. If the convergence in cross section is rapid, then the water slope at the front of the wave increases, and the speed of propagation of the tidal wave increases. This leads to a faster progression of high water up the estuary than low water, and the time delay of high water between the mouth and the head of the estuary is smaller than that of low water. This causes the ebb tide to become longer towards the head of the estuary, and the flood tide shorter, resulting in an asymmetrical tidal current, with the flood being shorter and stronger than the ebb. This is particularly enhanced towards the head of the estuary where the estuary bed level is above the general low tide level. A flood dominant response is typical of hypersynchronous estuaries where the tidal range is large compared with the water depth (Fig. 14-2).

When convergence is not important the tidal range diminishes landward (i.e., a hyposynchronous estuary), and less water flows landward through each section between the mouth and the head of the estuary on the flood tide. If the area of the intertidal increases headwards, then filling that volume of the tidal prism occurs when the cross section area through which the water is flowing is relatively large because of the flooding tide. On the ebb, however, the emptying occurs through a cross section with a smaller area. This leads to an ebb dominant response, which is typical of

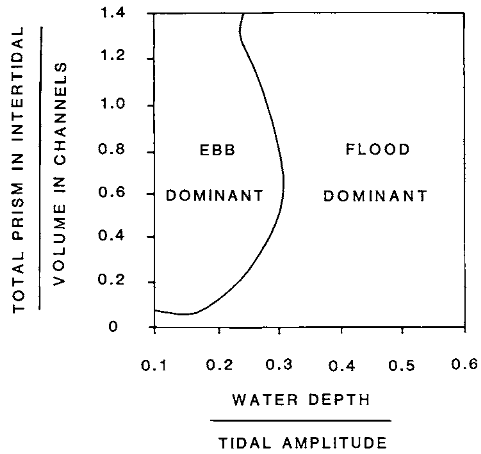


Fig. 14-2. Ebb and flood dominance in estuaries as related to their volumetric and dimensional characteristics (after Friedrichs and Aubrey, 1988).

estuaries with a large intertidal volume to channel volume ratio, as well as a relatively small tidal range (Fig. 14-2).

In terms of the normal tidal analysis procedure, the tidal distortion can be represented by a combination of the main semi-diurnal $M_2$ component and the principal overtide $M_4$, which has a quarter diurnal period. Depending on the phase relationship between these two constituents a variety of tidal curves can be produced, and each causes a similarly distorted tidal current pattern. For example, if the relative tidal phase is between 0° and 180° then the falling tide exceeds the rising tide in duration, and this produces a shorter enhanced flood current relative to the ebb current. The reverse effect occurs when the relative tidal phase is between 180° and 360°. These features have been described in a number of papers recently reviewed by Friedrichs and Aubrey (1988) and related to the shape of the intertidal basin of the estuary (Dronkers, 1986a).

A dynamic tidal equilibrium is conceptually possible between the tidal currents and the resulting sediment transport. This implies that the sediment movement causes a change in morphology that alters the tidal current regime which in turn reduces the sediment transport. The simplest equilibrium concept is that of O'Brien (1969). He found that the cross section of the mouth of an inlet $A$ was related to the tidal prism volume $P$; the volume of water that has to flow in and out through the mouth on each tide to raise the water level inside. He found $A = cP^n$. Analysis of many inlets gave $n \sim 1$ and $c = 2 \times 10^{-5}$ ft$^{-1}$. In actual fact the tidal prism volume is a volume per half tide, and the constant $c$ then has the dimensions of an inverse velocity. This results in $c$ being a half tidal mean velocity of 0.67 m s$^{-1}$, which is equivalent to the threshold of movement of sand. Thus the O'Brien relationship specifies that an increasing tidal prism leads to an increase in the velocity at the mouth of the inlet, which causes the sand to move, and the cross section to increase until the movement diminishes to the threshold value. This equilibrium tidal prism

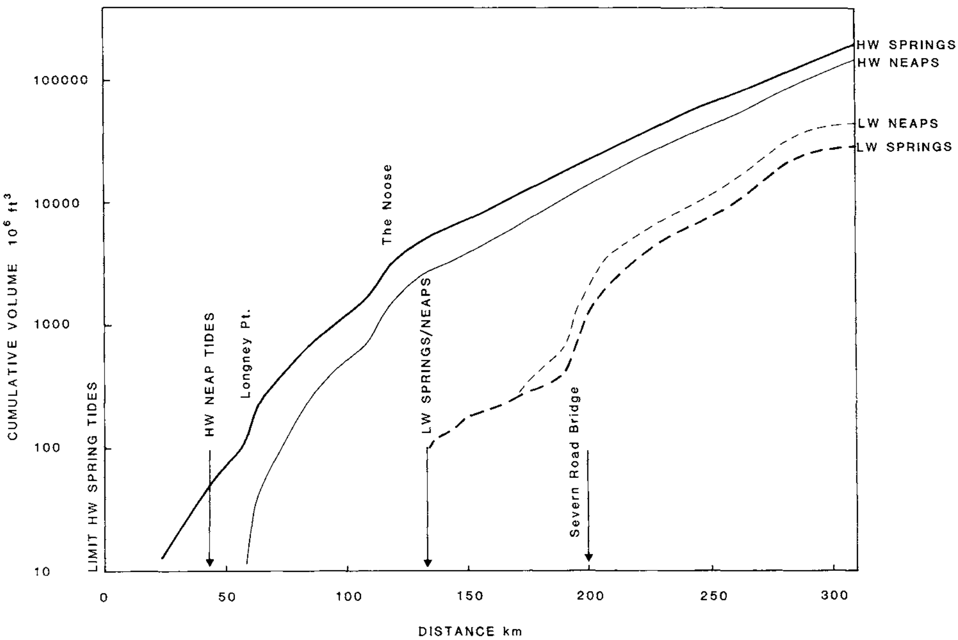


Fig. 14-3. Volumetric characteristics of the Severn Estuary.

concept has been widely applied in models of morphological development of tidal inlets and estuaries.

Similarly, empirical studies have shown relationships for other variables such as the area of the tidal flats. Van Dongeren (1992) has shown for a number of North Sea inlets that the equilibrium tidal flat area $A_f$ is related to the total basin area $A_b$ by $A_f = A_b - 0.025A_b^{3/2}$. It has also been found for many estuaries that the tidal prism and the low water volume vary exponentially with distance. Figure 14-3 shows the cross sectional area and volume variations along the Severn Estuary. Similarly, Wright et al. (1973) have shown the breadth and depth of the Ord Estuary in Australia varying in the same way. It is likely that an equilibrium estuary would have a phase relationship of either 180° or 0°, and be synchronous, though this has not been tested.

## TYPES OF ESTUARY

Within estuaries the tidal and residual water circulation patterns are important in determining the overall sediment transport. The patterns of sediment movement are different in different types of estuary, and several examples are discussed in other chapters of this volume (e.g., see Bokuniewicz or Castaing and Guilcher, this volume).

*Highly stratified estuary*

When there is little tidal motion the river flow, being less dense than the salt water, flows over it with a marked density interface between the two water masses. Because of the stable density gradient the two water masses will not mix readily together. However, if the surface layer has sufficiently high velocity the shear can create interfacial waves on the halocline. These waves break ejecting salt water into the fresher surface layer by a process called entrainment. No fresh water is mixed into the bottom layer and the mixing is entirely upwards. Thus, the bottom water loses salt gradually into the surface layer and this loss is made good by a slow inflow of salt water from the sea. The position of the salt wedge will vary with the river flow, and the tidal range is normally microtidal, ie, a range of less than 2 m.

*Partially mixed estuaries*

With increased tidal range, the whole water mass in the estuary moves backwards and forwards with a tidal periodicity. The friction between the water and the bed of the estuary creates turbulence, which mixes the water column more effectively than entrainment so that the salinity difference at the interface is considerably reduced, and there is a smaller velocity shear across the interface. Turbulent mixing not only mixes the salt water into the fresher surface layer, but it also mixes the fresher water downwards. This causes a longitudinal gradient in salinity, with salinity diminishing towards the head of the estuary, both in the surface and the bottom layers. There has to be a residual discharge of water towards the sea, but it now carries with it the salt resulting from the vertical mixing. The discharge from the surface layer can thus be an order of magnitude larger than the river discharge. Because of the requirements of continuity this discharge has to be replaced by a significant landward flow within the bottom layer. Consequently, at the estuary mouth a large volume of mixed water has to be discharged and the compensating inflow in the bottom layer has to be larger than in salt wedge estuary. This process is termed vertical gravitational circulation.

In partially mixed estuaries the tidal range is generally mesotidal, ie, between two and four metres. In this situation the tidal range can change significantly between spring and neap tides. The spring tide currents enhance the turbulent exchanges of salt and fresh water, and as a consequence the stratification can diminish considerably. This produces an increase in the vertical gravitational circulation. At times of high river flow the partially mixed estuary will become more highly stratified, and the intensity of the mean circulation should diminish. Within partially mixed estuaries there can be considerable variation in the vertical structure along the estuary, with highly stratified conditions at the head of the estuary where the water depth and the tidal range diminish, and river flow becomes comparatively more important.

*Well mixed estuaries*

When the tidal range is large relative to the water depth, especially in macrotidal conditions, there is sufficient energy in the turbulence to completely mix the water column and produce effectively vertically homogeneous conditions. In this type of estuary there can be lateral variations in salinity and in velocity, and horizontal circulation tends to develop at the expense of the vertical circulation. There can be residual flows inward on one side of the estuary and seaward on the other and separation in flood and ebb dominated channels.

## MODES OF SEDIMENT TRANSPORT

The sediment transported down the rivers is generally a heterogeneous mixture reflecting the variety of source grain sizes available within the catchment. However, some sorting takes place within the flow as a consequence of the different modes of transport of the fine and coarser material. Three modes of transport occur; wash load, suspension and bedload. The distinction between them is not a clear one since there are changes in the grain content of the modes depending on velocity. There are differences, however, between the response of non-cohesive sands and silts, and cohesive silts and clays.

Wash load comprises the finest fraction, and is normally composed of fine dispersed clay particles. They are kept in motion by turbulence, and move with the water at virtually all current speeds. The vertical profile of wash load concentration is homogeneous.

Suspension occurs because of erosion of grains from the bed, and the exchange of momentum with the grains because of turbulence. The threshold for suspension of sandy grains from a flat bed can be considered to occur at about $u_* \sim 0.8w_s$, where $u_*$ is the friction velocity and $w_s$ is the settling velocity. Grains less than about 150 $\mu$m will go into suspension immediately they begin to move. For grains above about 150 $\mu$m movement as bedload occurs first, and suspension does not take place until higher velocities. As the intensity of the flow increases so does both the concentration of material in suspension and its mean grain size. The vertical profile of concentration becomes graded, with higher concentrations and a larger content of coarser grains nearer the bed.

The threshold of bed load movement for a flat bed can be approximated by $u_*^2 \sim 40D$, where $u_*$ is in cm s$^{-1}$, and $D$ is the grain diameter (mm). Initially the grains move by saltating along the bed, but ripple and dune bedforms are created with increasing flow velocity. Asymmetry in the bedforms indicates their movement downstream. For a review of sediment transport dynamics see Dyer (1986).

Within the river, sediment moves intermittently with highest transport rates during high discharge events. At low flow rates, movement is restricted to the finer sizes. Once the sediment reaches the riverine section of the estuary, where there is a tidal rise and fall of the surface and where the current velocities become oscillatory during the tidal cycle, the bed shear stresses are reversed for part of the tide, thereby reducing the mean bedload transport. Additionally, the reversal of the current gives periods of slack water during which the material in suspension will settle to the bed, though it may be re-suspended on the next phase of the tide. Nevertheless, the suspended sediment will move in the direction of the tidally averaged water flow. The bedload, however, will be affected mainly by the highest velocity, and will move in the direction of the maximum current. This is a very effective grain sorting mechanism, and fine and coarse grains can move in different directions. Additionally bedforms created in the sandy sediment at maximum current can be draped with mud at slack water. Consequently, the patterns and rates of sediment transport will vary between different types of estuary.

## MUD PROPERTIES

Fine suspended particles in estuaries are particularly important because they become trapped by the estuarine processes, their concentrations can be high, and they form distinctive sediments. Their characteristics can be very variable with time, and this becomes important in their deposition, erosion and transport.

### *Flocculation*

Flocculation of particles occurs as a result of the total surface ionic charge on the particles and the enveloping electrical double layer, the properties of which depend on factors such as pH, and organic coatings. When the particles are in close proximity there is an overall attraction which leads to the formation of aggregates of particles, or flocs. The flocculation potential of particles increases with increased concentration, but is also enhanced by Brownian motion, differential settling, grain inertia and velocity shear (Krone, 1978; McCave, 1975). Of course not all collisions will lead to flocculation as some may lead to disruption of flocs, particularly at high turbulent shears. The predominance of the various flocculation processes will vary during the tide, with differential settling being most important at near slack water, and velocity shear at times of maximum current. The most effective interaction is between large and small flocs, so that flocculation progressively removes the finer particles from suspension (Kranck, 1973).

Laboratory measurements (e.g., Owen, 1970) have indicated that salinity can be important in the flocculation of particles since it controls the intensity of their surface charges. This leads to the concept that flocculation of riverine particles occurs when they reached the salt water, and that deflocculation could occur when particles are recycled back into contact with fresh water. More recently it has become clear that in-situ particles are usually held together by organic matter (Eisma et al., 1991), and the deflocculation is unlikely by purely chemical processes. Riverine particles have variable characteristics depending on the cation content of the river water (Hunter and Liss, 1982), but it is still unclear how much the floc characteristics are determined by mineralogy.

There are two modes which contribute to the distribution of floc size: macroflocs which reach a size of the order of millimetres and microflocs of the order ten to twenty microns. Macroflocs are about the same size as the turbulent Kolmogorov microscale, and they can be readily broken down to form microflocs. Microflocs are very much more resistant to being broken up, and may form the basic unit from which flocculation takes place. There is, however, a linear relationship between the primary grain size and the microfloc size (Kranck, 1975).

It is very difficult to observe the flocculation or break-up process within estuaries because of advection, and the effects of changing concentration. However, laboratory measurements have indicated that the modal floc size is affected by both concentration and shear stress, and the primary cause of floc disruption is by three-particle collisions (Burban et al., 1989). The general principle seems to be that at low concentrations the flocs are small, and low shear increases the likelihood of floc growth due to collision. However, with increasing shear the intensity of the collisions leads to floc breaking. At higher concentrations larger size flocs are present in quiescent settling, but only a small amount of shearing is necessary to disrupt them. The floc size distribution, in terms of sorting or standard deviation, is also likely to change with shear, with a narrow size distribution at low and at high shear. At moderate shear a wide range of floc sizes should be present during floc breakup. Thus, in addition to macroflocs and microflocs, a background of primary particles of size $<2\ \mu$m, which are involved in flocculation is likely. This would comprise the estuarine wash load.

*Settling velocity*

The settling velocity of suspended material is an important parameter in determining the transport and deposition rates. For flocculated mud, the settling velocity is related to concentration. There are discrepancies between laboratory and field results largely because sampling disrupts the macroflocs. There are also considerable differences in the settling velocity/concentration relationship between estuaries which may be the result of floc density or organic content variations, and between different turbulent states during floc formation. Nevertheless at concentrations less than about 2–5000 mg $l^{-1}$ the settling velocity $w_s$ depends on concentration according to $w_s \propto c^n$, where $n$ varies 0.6 to 2.2 (Dyer, 1989). Above 10–20 g $l^{-1}$ the settling particles interfere with each other and "hindered" settling occurs (Fig. 14-4a). For modelling purposes $n$ is generally taken as unity.

*Deposition*

The product of the settling velocity and the concentration gives the settling flux towards the bed (Fig. 14-4b). This shows a maximum at a concentration of about 20 g $l^{-1}$. At any level in the flow at this concentration sediment would be settling from above faster than it is settling out beneath, and thus it builds up a layer on the bed. At this concentration the flocs are mainly separated by water. However, they collide with each other during shearing, a process that takes energy out of the flow. The suspension then becomes pseudoplastic in its rheological properties. At low shear stresses it will have a high viscosity, but as shear increases the flocs structure becomes broken down and the effective viscosity is reduced. When concentrations exceed 80–220 g $l^{-1}$

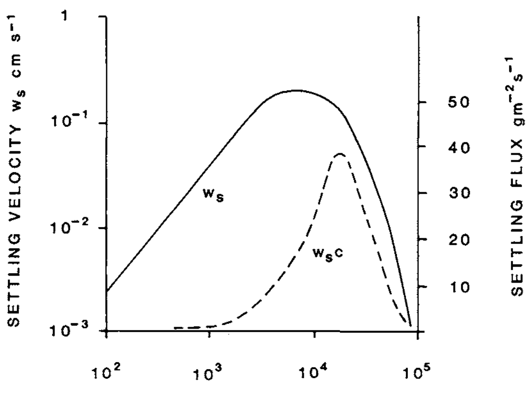


Fig. 14-4. Settling velocity and settling flux of estuarine flocs in relation to concentration.

patches occur where flocs are in contact and a framework structure gradually develops throughout the suspension, it becomes a soft bed and has the properties of a Bingham plastic (Sills and Elder, 1986). At this stage it will transmit a shear wave, and has a rigidity modulus. If left for time periods of hours the framework will gradually collapse owing to the weight of particles above, in a process known as consolidation. The pore water between the flocs is gradually forced out and the increasing particle contacts lead to an increase in the cohesive strength and density with depth into the sediment. Consequently with settling there is a gradual change from a dispersed suspension to the formation of a high concentration near bed layer, and to a soft bed which then consolidates.

The high concentration layers have a lutocline at their upper surface, and this is often detectable by echo-sounders, at which stage the layers are called fluid mud. In some estuaries fluid muds have been detected as several meters thick in pond-like patches. It is obvious that these could not have formed solely by settling from the water column directly above. They must have largely been created by gravitational flow down the side slopes of the channels, even though the layers have been observed as standing on slight slopes. The erosion and re-entrainment of these fluid muds is not well quantified, but is related to the Richardsons Number at the interface (Srinivas and Mehta, 1990). However, in many estuaries they are evident at neap tides and for short periods over slack water at spring tides (Kirby and Parker, 1983). The high concentration layer, having a high viscosity, will undergo a fluid shear at the level of the lutocline which may create some erosion of the upper surface, but failure may alternatively occur at the bottom of the layer, so that the layer moves as a plug. Once it has started moving the viscosity will decrease and the layer could rapidly become dispersed within the body of the flow. There are indications that layer thickness may be crucial in separating these two modes of erosion (Odd and Rodger, 1986).

### *Erosion*

Erosion of a mud bed can occur by several processes; by erosion of individual flocs, erosion of small clusters of particles, or by failure and mass erosion of a surface layer, and these normally occur in succession with increasing bed shear stress and depth of erosion. Laboratory measurements have shown there is a correlation between density and the critical erosion shear stress for some muds, (e.g., Thorn and Parsons, 1980), but the correlation coefficient is not high and there are many other important factors such as particle mineralogy and grain size distribution, and organic content. Several studies have examined the relationship between the critical erosion shear stress and the yield stress obtained in rheological measurements (e.g., Migniot, 1968). Additionally, empirical relationships have been demonstrated between the yield stress and the rigidity modulus (Williams and Williams, 1989) and with the specific surface area at a constant concentration.

Because of consolidation of deposited sediment, there is a gradient of most physical and chemical properties with depth into the bed. Density and the critical erosion shear stress rise almost exponentially with depth. In many situations the mud surface comprises two layers at slack water; a thin fluid mud or loose fluffy layer, overlying a more rigid bed. The upper layer has a threshold shear stress of 0.06–0.1 N $m^{-2}$, and is fairly easily eroded during the tide. Erosion of this layer is often quite sufficient to explain the concentration observed throughout the water column at maximum current. The lower layer would normally only be eroded at extreme conditions, and after weakening by biological activity.

## TRANSPORT OF MUD IN TIDAL CURRENTS

Within estuaries it is apparent that there is a continual and sometimes very rapid exchange of sediment between the suspension phase and the bed. This has been described by Kirby and Parker (1983), Mehta et al. (1989), Mehta and Dyer (1990). The basic concepts are shown in Fig. 14-5. At high velocities the suspension is mobile and sustained by the turbulent shearing. Concentrations are generally of less than a few thousand ppm (mg $l^{-1}$). As the current slackens towards low water, settling creates a high concentration layer near the bed that becomes static. Initially formed at about 20 g $l^{-1}$, the layer becomes denser with time, a process that is aided by gentle shearing. Re-entrainment of the whole of the layer may occur as the current increases on the next phase of the tide. Alternatively, some part of the layer may survive over the tide and consolidate. This in particular may occur during the progression in tidal range from spring to neap tides as the currents decrease in strength. The remaining part may then survive, but become eroded on the following spring tide. Complete survival may allow the sediment to become part of the settled bed for annual timescales, or longer.

The bed shear stress at which deposition occurs is normally considered to be 0.04–0.08 Pa which is about 30% smaller than the erosion shear stress. Thus, there is a range of shear stresses during which transport occurs, but no erosion or sedimentation. This model of tidal settling and re-entrainment requires that the

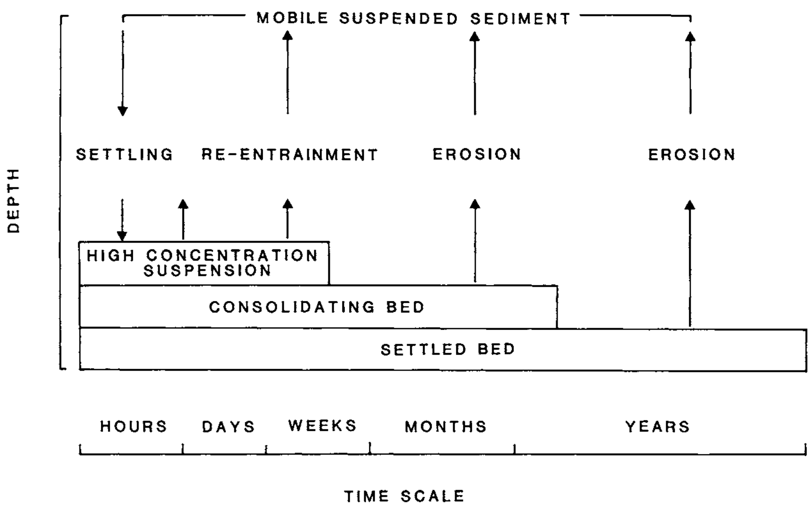


Fig. 14-5. Deposition/resuspension cycles (after Mehta et al., 1989).

concentration of suspended sediment does not start falling until the current velocity is very low, and thus there would be a large phase lag between velocity and concentration. However, this is at variance with a large number of observations which show that concentrations begin to decrease shortly after the velocity begins to decrease (e.g., Officer, 1981). Sanford and Halka (1993) have proposed an alternative model that appears to fit the observations better. They suggest that there is a continual exchange of sediment with the bed because the temporal and spatial variability in shear stress allows material to settle to the bed throughout the tidal cycle, where it is available for re-erosion.

Nevertheless, a proportion of the flocs coming into contact with the bed during the turbulent transport processes will stick and not be re-eroded. McCave (1975) has linked this process to the presence of a viscous sub-layer, which may be considerably thickened by the presence of suspended sediment. Additionally, there is the effect of differential settling of the various size classes which may produce settling of the largest flocs to the bed long before the mean settling class. Once the velocity begins to decrease it is obvious that no new material will be eroded from the bed when the critical erosion stress of the bed is no longer being exceeded.

A particular problem in the interpretation of field measurements is the separation of the effects of advection from local erosion and settling processes. Advection can produce a number of concentration peaks during the tide linked to the erosion and exhaustion of thin mud patches upstream of the measurement position. In an oscillating flow with asymmetric velocities, a particle suspended in the water will travel passively in the direction of the residual water flow. However, if the response of the particle to the flow contains a non-linearity, or a lag with respect to velocity, then the particles will have a different residual movement. The settling and erosion properties of the mud, together with movement of particles vertically in the flow can cause the necessary lags.

## TURBIDITY MAXIMUM

One of the most distinctive features of sediment transport in meso and macrotidal estuaries is the turbidity maximum. This is a zone which contains suspended sediment concentrations higher than those both in the river or further seaward in the estuary. It is generally located at, or somewhat landward of the head of the salt intrusion, where salinities are about 1–5‰. The energetic tidal flow is capable of maintaining high concentrations, and there are a number of processes that concentrate the suspended sediment, and prevent particles from dispersing.

The peak concentration of suspended sediment in the turbidity maximum varies between wide limits. Despite the differences due to sediment availability, low tidal range estuaries have maxima with concentrations of the order 100–200 ppm (mg $l^{-1}$), whereas high tidal range estuaries have much higher concentrations, of the order 1000–10,000 ppm (1–10 g $l^{-1}$).

The turbidity maximum contains a high proportion of a narrow size range of mobile fine sediment, and plays a central role in the circulation of fine sediment within the estuary, as well as probably determining the rate of transport of sediment from the river to the sea. The concentrations of sediment in the turbidity maximum appear to remain almost constant when averaged over a reasonable time, so that residence time of grains in the turbidity maximum must be considerable. Since the turbidity maximum contains several times the annual river discharge of fine sediment, the residence time is likely to be well in excess of a year.

The turbidity maximum responds to changes in river flow, with the maximum moving downstream with increasing flow. The mass of sediment in the turbidity maximum also increases. However, a movement of the turbidity maximum down estuary involves expansion into an increased cross sectional volume, and this could decrease the concentrations even though the total mass increases. In the Cumberland Basin (Canada), Amos and Tee (1989) found results which suggest that an increased mass of sediment in the turbidity maximum leads to an increased maximum concentration as well as an increased longitudinal distribution.

In the hypersynchronous Tamar Estuary (UK), the turbidity maximum is generally located in an area of locally larger currents, and is stronger when it is nearer the estuary head. Uncles and Stephens (1989) found that the magnitude of the maximum was greatest when the cross sectional area at high water at the location of the maximum was least. Additionally, the distance of the maximum from the head of the estuary varied roughly as the square root of the run-off.

The movement of the maximum is well illustrated for the Gironde Estuary by Fig. 14-6. The more or less steady position at high river discharge indicates that the estuarine response tends to be buffered by the increasing stratification. Consequently, as river flow increases, the stratification gradually decouples the upper and lower layers, so that an increasing proportion of the river borne suspended sediment passes straight through the estuary in the upper layer, with the possibility of loss of material into the coastal zone. The seasonal changes in river flow suggests that sediment can accumulate in the upper estuary in spring and summer, and be redistributed down estuary in autumn and winter.

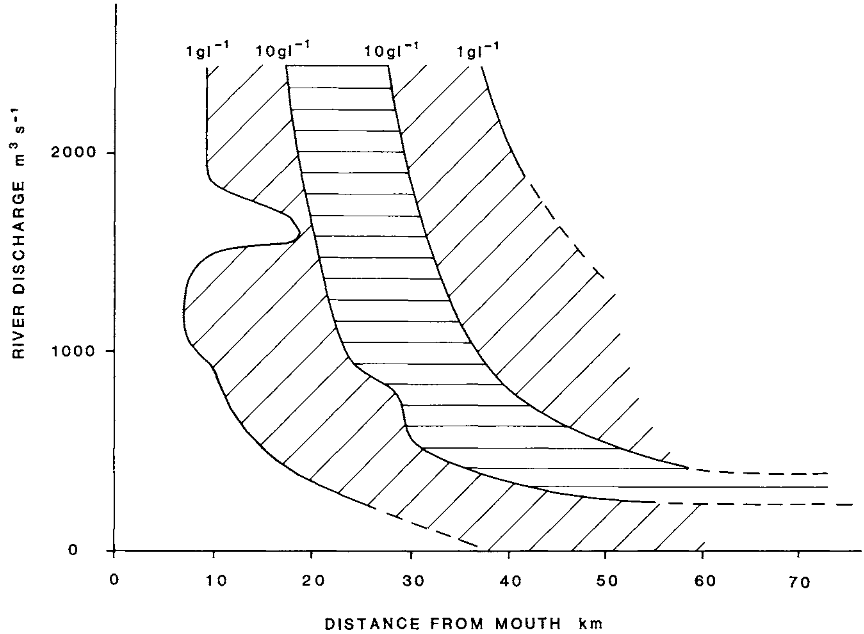


Fig. 14-6. Movement of the turbidity maximum in the Gironde Estuary (after Allen, 1973).

Additionally, the turbidity maximum moves with the tide. At about high water the maximum is located well up the estuary, and concentrations are relatively low because of settling over slack water. During the ebb tide the maximum is moved seaward, sediment is entrained from the bed and concentrations increase. At low water the maximum is further down estuary, and over slack water some settling of material occurs.

The structure of the turbidity maximum revealed by measurements at a single station is also shown for a position near the head of the Tamar Estuary in Fig. 14-7 (McCabe et al., 1992). The depth is shown relative to the sea surface, and the measurements spanned just over ten hours at the end of the ebb tide and the beginning of the flood. The rapid deepening at about 1700 hours, when the velocities during the early flood tide reached 1 m $s^{-1}$, illustrates the asymmetry of the tide. The ebb tide is longer in duration, with a peak velocity of only 0.6 m $s^{-1}$. The turbidity maximum passage coincides with the velocity maxima, and this occurs at a time when the salinity intrusion is seaward of that location. The majority of the suspended sediment is being entrained from the bed, with a velocity of about 0.4 m $s^{-1}$ at 0.5 m above the bed being the critical erosion threshold. On the ebb tide the lower current velocities erode less sediment and the peak concentrations only reach 400 mg $l^{-1}$. Further down the estuary the maximum velocities occur when the tip of the salt intrusion is landward and the turbidity maximum appears within the saline water (West and Sangoyodin, 1991).

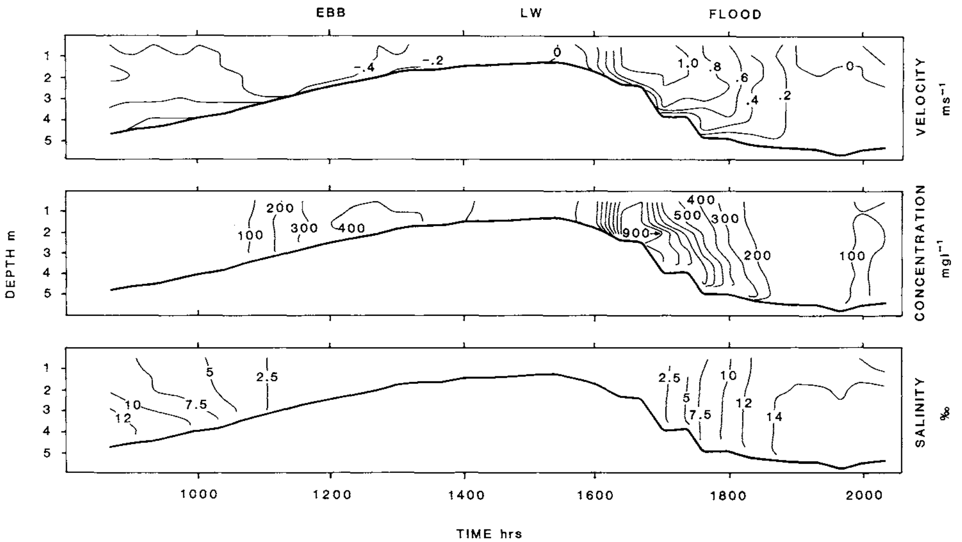


Fig. 14-7. The movement during a spring tide of the turbidity maximum in the Tamar Estuary and its relationship with velocity and salinity (after McCabe et al., 1992).

Because of the flood dominance, over the tidal cycle there is a very significant landward sediment flux; the strong flood turbidity maximum returns on the ebb at a much lower concentration, deposition having occurred over high water. This landward flux occurs at low river flow, and is balanced by a downstream re-distribution of sediment at high river flow (Uncles and Stephens, 1989).

A feature of macrotidal estuaries is the large difference in tidal range between spring and neap tides. Because of the considerable variation of velocities there are changes in position and magnitude of the turbidity maximum (Allen et al., 1980; Gelfenbaum, 1983). The estuary may be partially mixed at neap tides, but during the increasing tide there can be a sudden change to well mixed conditions. Though the estuary may be well mixed at spring tide, at neap tides it can be partially mixed, or even stratified. At spring tide the turbidity maximum has its highest concentration, as the currents are able to erode and sustain more sediment in suspension, and it will be located further up the estuary. This is due to the fact that there is a higher mean sea level in the upper estuary at springs than at neap tides, arising because the increased range at spring tides involves a large extra volume of water at high tide, but only a slight volume difference at low tide, relative to the neaps. During decreasing tidal amplitude towards neaps, the peak currents decrease, and less material is capable of being re-eroded and suspended. Additionally, the durations of slack water increase, enhancing deposition.

In the Severn Estuary the mean total mass in suspension varies 5–30 kg m$^{-2}$ between neap and spring tides. However, at one station tidal variations were 15–100 kg m$^{-2}$ and 70–160 kg m$^{-2}$ at the two tidal states, respectively (Kirby, 1986).

PROCESSES FORMING THE TURBIDITY MAXIMUM

The turbidity maximum is a dynamic feature that involves interaction of the tidal flow with erosion and deposition of sediment. There are a number of processes that operate to concentrate the fine sediment at the upper end of the estuary, and to keep it there.

*Residual circulation*

In partially mixed estuaries the vertical gravitational circulation produces a residual landward bottom flow, and a seaward surface residual flow. This has long been thought to be the main mechanism for maintaining the turbidity maximum (Schubel and Carter, 1984). Because of the residual downstream flow in the river, there is a convergence in the bottom flow at a null point near the head of the salt intrusion, in salinities of about 1–5‰ (Fig. 14-8). Suspended sediment is brought into the estuary by the river, and in the upper estuary energetic tidal mixing enhances sediment resuspension, and transfers the sediment between the surface and lower layers (Kostaschuk and Luternaeur, 1989). The surface layer transports sediment downstream to the middle estuary where the particles settle into the lower layer, only to be carried headwards on the residual flow, together with particles brought in from lower down the estuary. Consequently, the maximum concentration of suspended sediment occurs at the bottom near the null point. This circulation process can lead to a turbidity maximum without the need for consideration of sediment properties other than settling velocity, and without any sediment erosion or deposition. Also it is a mechanism for sorting the flocs, since a change in the settling velocity leads to a variation in the suspended sediment concentration (Festa and Hansen, 1978). The fact that the turbidity maximum sometimes occurs landward of the salt intrusion indicates that the vertical gravitation circulation is not always dominant.

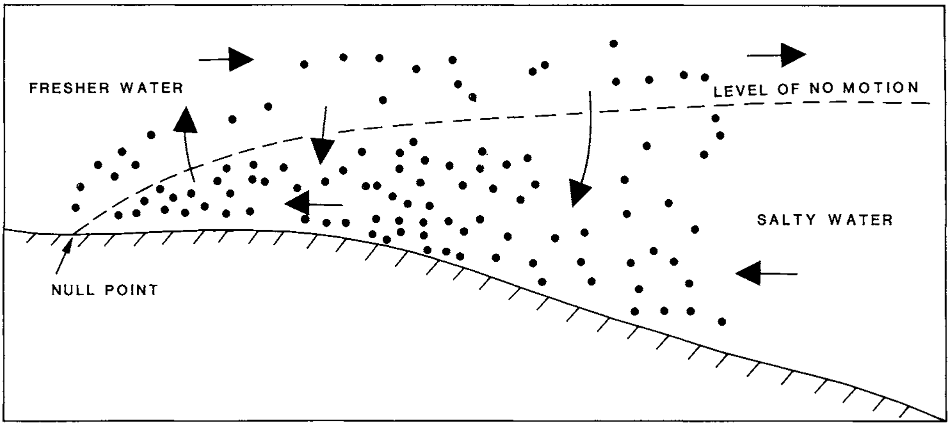


Fig. 14-8. Diagrammatic representation of the formation of the turbidity maximum by vertical gravitational circulation.

### *Lag effects*

As has been described above, if the sediment particles were a passive tracer which responded instantaneously to the flow, the movement of the sediment would be in the residual flow direction. However, the sediment response lags the flow. This phase difference between the suspended sediment concentration and the water velocity can produce a residual flux of sediment even when there is no residual movement of water, providing the currents are asymmetrical. Lags can be produced by a variety of causes, with different processes being important for mud and for sand.

#### *Threshold lag*

A lag can be produced by the presence of a threshold for sediment movement. In an asymmetrical tidal current, sediment movement can take place for a longer time on one phase of the current compared with another. As an example, in a tidal cycle with an intense short flood current and a long lower velocity ebb current, the duration of movement on the ebb tide will decrease more rapidly than that on the flood tide with an increase in the threshold of sediment movement (Fig. 14-9). In the extreme, the current on the ebb tide may not reach the threshold velocity, and all of the movement then occurs on the flood tide. The asymmetry in the sediment discharge, or transport rate, caused by this effect will be even more marked if the transport rate has a non-linear relationship to the current velocity. In practice measurements have shown that the sediment transport rate for bed load is normally proportional to the 3/2 to 7/2 power of the bed shear stress. As the asymmetry of the tide increases towards the head of the estuary the increasing magnitude of the flood current causes a transport of sediment towards the head of the estuary. This has been suggested by Allen et al. (1980) as a major process in macrotidal estuaries for creating the turbidity maximum.

#### *Erosion lag*

In mud, consolidation causes an increase in the critical erosion shear stress with depth. Thus, as the current velocity increases, erosion takes place to a depth where the ambient fluid shear stress equals the critical erosion value for the mud. Once the current diminishes after the maximum, no more erosion takes place. Thus, there will be an asymmetry created in the suspended sediment concentration over and above that present in the current. Modelling of the consolidation properties is an important aspect of mud transport in estuarine models (Hayter, 1986).

#### *Scour Lag*

Once sand particles are in motion they can be kept moving at velocities below the threshold of initial motion. Consequently, between the threshold of erosion and the threshold of deposition material is kept in motion, but no new erosion takes place. This produces a scour lag (Postma, 1967).

For mud, scour lag can be defined as the time taken for sediment, when entrained from the bed, to disperse to higher levels in the flow (Nichols, 1986a). This means that once material is eroded from the bed it is only gradually mixed through the water flow as it moves downstream. Initially the sediment will move in the near bed

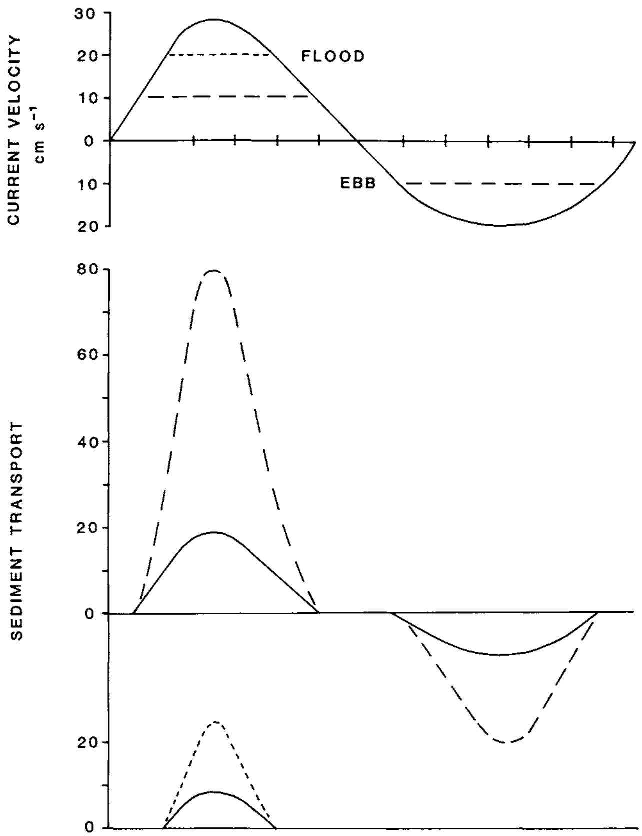


Fig. 14-9. Variation of current velocity and sediment transport (relative units) during a tidal cycle. Top: current velocity. Centre: sediment transport rate for a threshold of 10 cm $s^{-1}$, full line: assuming a linear relationship with excess velocity, dotted line: assuming transport proportional to 3/2 power of excess velocity. Bottom: sediment transport rate for a threshold of 20 cm $s^{-1}$, full line: linear relationship, dotted line: power relationship.

layers at a velocity lower than the depth mean current. Consequently, at higher levels in the flow the suspended sediment concentration will lag behind the concentrations being produced at the sea bed.

*Settling lag*

On the decreasing tide the particles will start to settle once the turbulence in the flow is incapable of maintaining them in suspension. As the particles settle they are moving along on the waning current so that they eventually reach the bed some distance from the point at which settling commenced. This effect is settling lag, and a qualitative model describing these effects was developed by van Straaten and Kuenen (1958) and Postma (1961).

To illustrate the effect consider the simple situation shown in Fig. 14-10. The symmetrical tide has a decreasing maximum current towards the head of the estuary,

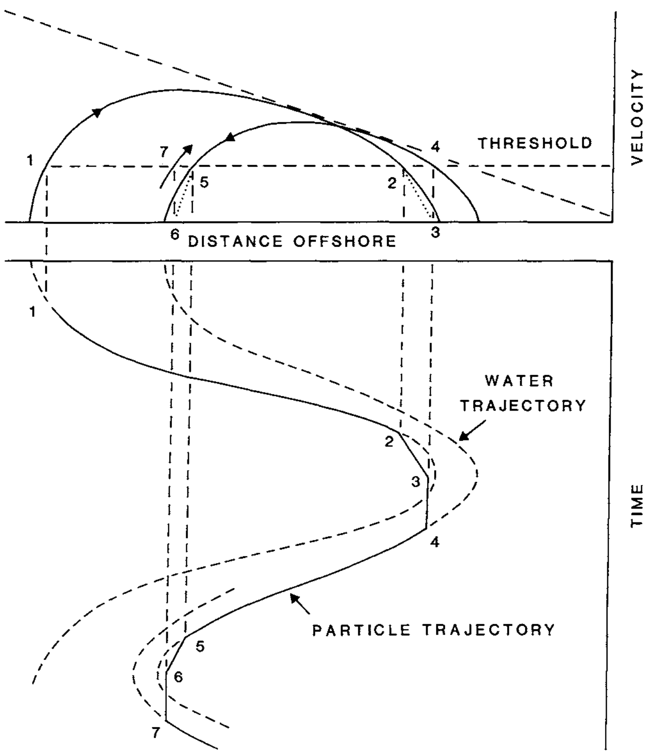


Fig. 14-10. Schematic illustration of settling lag in an oscillatory current reducing in magnitude towards the head of the estuary (after Postma, 1961).

and the water and sediment particles undergo a changing velocity with distance along the channel during the tide. A particle on the bed at 1 will be lifted into suspension as the threshold velocity is exceeded. It then travels with the water until 2, at which point it starts to settle. Because of settling lag it reaches the bed at 3. On the following ebb tide it will not be entrained until later in the tide when the threshold velocity is reached at that position, and it travels with the water until deposition at low water at position 6. Consequently, the particles gradually migrate shorewards to deposit in the area where the maximum velocity during the tide equals the threshold velocity of the grain. Settling lag will sort the particles according to their threshold characteristics and settling velocity.

Dronkers (1986b) considered the time interval during which sediment particles can settle at slack water and remain on the bottom until resuspended, and concluded that the magnitude and direction of the residual sediment flux is mainly determined by the current velocities around low water and high water slack. The slack water period at high water generally exceeds that at low water. Additionally, when there are extensive intertidal areas at high water the average water depth can be less than that at low water. Thus, settling is a more efficient process at high water. In the Humber Estuary there is a marked delay of about an hour in the rise in concentration at mid depth relative to near bed during the flood tide. At the end of the flood tide the concentrations are sustained by sediment settling from higher in the flow, and the time lag is reduced to about half an hour. In this case it appears that a combination of scour lag and settling lag produces the depth variation.

## HORIZONTAL FLUXES

The lag effects occur simultaneously and their effects will be difficult to separate. The relative importances of their contributions to the horizontal fluxes can be considered from examination of the temporal and spatial variations of velocity and suspended sediment concentrations during the tidal cycle.The instantaneous flux of suspended sediment through a vertical element of an estuary is given by:

$$F = \int_0^h uc\,\mathrm{d}z$$

where $h$ is the water depth. Averaging over the tidal cycle gives:

$$\overline{F} = \overline{h}\,\overline{u}\,\overline{c} + \overline{hU_A}\,\overline{c}_A + \overline{hC_A}\,\overline{u}_A + \overline{h}\,\overline{U_A C_A} + \overline{hU_A C_A} + \overline{h}(\overline{u}_d\overline{c}_d)_A + \overline{h}(\overline{U}_d\overline{C}_d)_A$$

where the overbar denotes a tidal average, and subscript $A$ a depth average. $U$ and $C$ are the tidal cycle fluctuations in velocity and concentration, and subscript $d$ are the deviations with depth from the depth mean values (for the full derivation see Dyer, 1978).

The first two terms on the right hand side are the downstream advection on the river flow, $\overline{u}$ being the non-tidal drift, and term 2 the Stokes Drift. Terms 3–5 are fluxes due to phase differences between the depth mean velocity, concentration, and the water depth, and arise mainly because of threshold and erosion lags in the sediment response to the tidal current asymmetry. These terms contribute to what is known as tidal pumping. Terms 6 and 7 known as the shear effect, arise because of variations in the vertical of the profiles of velocity and concentration. For term 6, a negative (upstream contribution) is produced if a large velocity at the surface is associated with a small concentration, together with a small velocity with a large concentration near the bottom. When averaged over the tide the result is the response to the vertical gravitational circulation. Term 7 arises from the different form of the velocity and concentration profiles during the tidal cycle, due to entrainment and settling lags.

This approach has been applied to several estuaries by Dyer (1978; 1988), Uncles et al. (1984; 1985), and Su and Wang (1986). The main difficulty is the assumption that over the tidal cycles of observation, the estuaries are in steady-state. Nevertheless, in all cases the cross-sectional fluxes produced by tidal pumping were larger than those produced by residual gravitational circulation. Consequently, erosion and suspension of sediment during the tide is a major factor in generating and supporting the turbidity maximum.

At the seaward end of the turbidity maximum advection from upstream of eroded material leads to maximum concentrations appearing close to low water, and to phase relationships creating an upstream tidal pumping. At the upper end of the turbidity maximum the reverse happens, with maximum concentrations occurring near high slack water, producing a downstream tidal pumping component. At locations near the peak of the turbidity maximum the tidal pumping term is likely to be a minimum. These effects are likely to be mainly coincident with the asymmetry in the flow; flood dominated in the saline intrusion, and ebb dominated in the riverine section. The vertical gravitational circulation is likely to be a minor contribution to the turbidity maximum, though it may help to sharpen the peak, and the concentration gradients.

There are also significant differences laterally across the estuary in tidal pumping. Uncles et al. (1984) have shown that near the head of the Tamar Estuary landward pumping of sediment occurred in the central channel, whereas other sections had weak landward pumping in shallow water and seaward pumping in deeper water.

## ESTUARINE TRAPPING

Within the estuary the riverborne sediments become trapped by the tidal pumping and residual circulation, and mixes with material brought in from the sea. Meade (1969) has argued that the majority of the sediment in estuaries of eastern North America is derived from the sea, despite the high river discharge. This conclusion seems to be valid for many temperate estuaries.

Riverborne and marine sources of sediment can often be distinguished from examination of clay mineralogy, heavy mineral content, and radioactive and stable isotope tracers, e.g., Nichols (1972), Song et al. (1983), Mulholland and Olsen (1992). Figure 14-11 illustrates this, showing that a large percentage of marine derived material can be present right up to the head of the salt intrusion. The process of mixing involves continuous erosion, deposition and exchange of sediment within the estuary; the fine sediment cycling through the turbidity maximum and coarser sediment cycling round the ebb–flood channel systems. Individual particles may spend a considerable time moving within the system before being finally deposited, or passing through to the sea.

The residence time of particles can be defined as the number of particles inside the estuary divided by the number leaving per unit time (Martin et al., 1986). Some of the particles entering from the river will remain in suspension and pass through the estuary fairly quickly particularly at times of high river floods. However, a significant proportion will undergo many cycles of deposition on the bed followed by resuspension, with the deposition occurring at a number of points along the estuary which form temporary sinks for the sediment particles operating for a variety of timescales. Consequently the mass of particles in suspension in the turbidity maximum comprise proportions of particles that may have ages (time since input) lasting from a few days to possibly years. Little is known concerning particle residence times in estuarine turbidity maxima.

The trapping efficiency of the estuary is the ratio of the fluvial sediment input, to that accumulated in the estuary. For partially mixed estuaries it can exceed 100% (Nichols, 1986b), since the fluvial sediment is only part of that accumulating. Some of that drawn in from beyond the estuary mouth is likely to be fluvial material exported

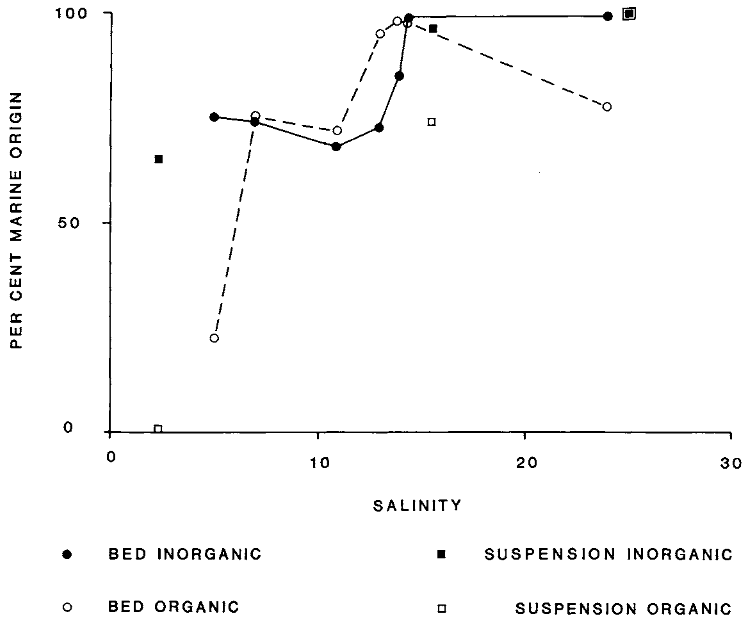


Fig. 14-11. Mixing of sediments of marine and freshwater origin in the Savannah Estuary (after Mulholland and Olsen, 1992).

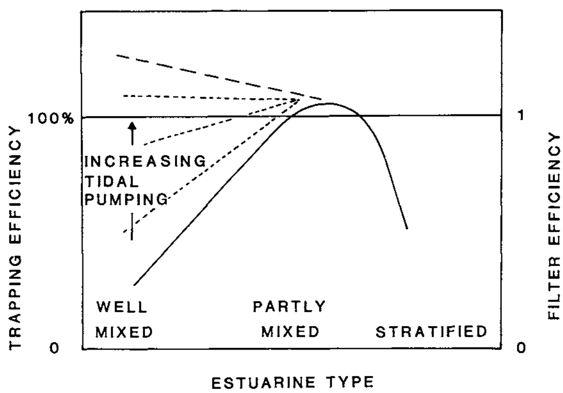


Fig. 14-12. Variation of trapping and filter efficiency of various estuary types.

at higher river flow stages, but much will be of coastal or marine origin. Additionally, in well mixed estuaries tidal pumping becomes significant in transporting sediment up-estuary into the turbidity maximum (Fig. 14-12), with the degree of tidal pumping depending on the tidal characteristics, as well as those of the sediment.

A filter efficiency has also been defined by Schubel and Carter (1984). This takes account of the additional accumulation of sediment of marine derivation. However, tidal pumping is not included in their analysis.

The trapping characteristic is likely to be fairly sensitive to the topography of the estuaries, in its effect on the tidal velocity field, and on the river discharge which effects the stratification and the gravitational circulation. The trapping thus undergoes considerable short term variability. For instance, the Tay Estuary exports sediment to the sea on neap tides but imports on the spring tides (Dobereiner and McManus, 1983). This situation is likely to be most apparent in estuaries near sedimentary equilibrium.

Nevertheless, the sediment particles can be continually cycled from one part of the estuary to another through the turbidity maximum. The major sites of this interchange are the intertidal areas which often show deposition rates of the order of a centimetre per year.

The surface of salt marshes above neap tide high water mark, and the upper part of the intertidal flats, show regular sedimentation, with layering and lamination in core samples. However, the outer edges of the salt marsh often show erosion by "cliffing", the undercutting and erosion of small blocks of compacted salt marsh sediment. Additionally, gullies and meandering channels cross the flats and show active erosion of the banks and migration of the meanders. As the channels meander across the mudflats they transform the horizontally stratified sediments into sequences showing laminations inclined at 7–15°. These are produced by deposition on the inside of bends in the gullies while erosion occurs on the outside of the bends (Bridges and Leeder, 1976). Within the Humber Estuary there are short term (1–30 year) cycles of mudflat and marsh edge erosion which appear to be related to periodic shifts in the low water channel (Pethick, 1988).

Consequently, one can envisage a continual cycle with the mudflats building up to a particular level, and then being attacked and eroded by shifts in the channels and by gullying. The eroded sediment is exchanged via the turbidity maximum to other areas of temporary deposition.

There is an important seasonal cycle in the build up of sediment on the intertidal areas within which plants play an important role. However, the response is somewhat different between the exposed mudflats, and the salt marshes. Frostick and McCave (1979) have measured a 5-cm accretion of mudflats between April and September because of trapping by algae and erosion in the winter. A surface layer of benthic diatoms cause a large increase in the critical erosion shear stress (Paterson, 1989) and this must reduce erosion by waves, as well as enhancing deposition. However, other seasonal variations are possible. Kraueter and Wetzel (1986) have shown stable sediment conditions occurred between December and March, but increased benthic activity in the summer caused increased water content, a decrease in sediment shear strength, and increased suspended sediment concentration.

Orson et al. (1992) has illustrated the processes of salt marsh accretion by seasonal trapping by plants. Erosion on the mudflats can be effected by ice, rain and waves (Anderson, 1983). Small amplitude waves can increase suspended sediment concentrations in the shallow water by a factor of three. In high turbidity the waves are modified into solitary waves (Wells and Coleman, 1981). The forward velocities under the crests are greater than the backward motion and cause a preferential shoreward motion. On the flood tide the suspended sediment is transported onto the higher tidal flats where some can be trapped. During the ebb tide the flow becomes quickly concentrated into the gullies, and is ejected as plumes into the main channels, where it becomes part of the turbidity maximum. The processes of sedimentation on Korean mudflats has been described by Wells et al. (1990).

On the intertidal flats the maximum rate of sedimentation occurs on the outer edge of the intertidal flats about mid tide level (Dieckmann, et al., 1987), so that the flats build outwards and upwards, with the consequence that the active volume of the estuary gradually decreases, reducing the sedimentation rate.

On the salt marshes the maximum accretion rate occurs near the high water spring tide elevation and this has been modelled by Allen (1990).

## SUMMARY

There is a sea level rise of about 1 mm $yr^{-1}$ occurring worldwide. This is predicted to accelerate and give a total sea level rise of about 50 cm by 2050. It is to be expected that estuaries will respond to this rise, though with possible lags. Stevenson et al. (1980) examined fifteen American estuaries and found a strong correlation between mean tidal range and accretionary balance, with high range estuaries showing accretion exceeding sea level rise. However, Nichols (1989) has examined the response of twenty two American lagoons to rising sea level, and found that the majority of them had accumulation rates equivalent to the local sea level rise. Many other equilibrium estuaries appear to be infilling at a rate consistent with sea level rise. Whether this will still hold with an accelerated sea level rise rate will depend crucially on the response of the sedimentary sources. Rising sea level will produce enhanced coast erosion or barrier beach retreat, though this may be limited by coastal defence works. Littoral transport will convey much of this material to the estuary mouths. There the coarser material will become trapped in the ebb and flood tidal deltas. The finer material will be carried into the turbidity maximum, by a combination of tidal pumping and gravitational circulation, where it will join material coming down the rivers.

A rise in sea level would normally reduce the rate of sediment input into the estuary, because of preferential deposition in the lower flood plains of the rivers. However, global warming is likely to increase the storminess of the weather. The increased incidence of floods are likely to flush these sediments into the estuary.

Within the estuary deposition would produce an expansion of the intertidal flat levels, especially if the inner edges of the salt marshes were allowed to encroach onto the surrounding low lands. If sedimentation on the marsh surface was insufficient to keep up with sea level rise, there would be a progressive narrowing of the vegetational zones, which may lead to a further reduction in the sedimentation rate. The deeper water in the channels would lead to more active wave attack on the intertidal zone, as well as a change in the tidal regime. It is possible that the estuary may change from being flood to ebb dominated.

Prediction of the future sediment patterns in estuaries and the infilling rates depends on a complex of interacting processes. Predictive models require a good

hydrodynamic basis coupled to specification of the erosion thresholds and rates, settling velocities, and consolidation of the sediment. The modification of turbulence and shear stresses by high concentration layers is an important feedback between the sediment and the water flow. Flocculation is also an important process whereby there is direct interaction between the flow and the sediment properties. On the tidal flats waves and currents will together be important, and models would need to be three dimensional.

Because of the influence of time in the sedimentary reactions to the flow, tidally averaged models will only be of restricted use. Consequently, estuary sedimentation is a challenging area of interest where direct collaboration between the disciplines, and combined field, laboratory and modelling work is essential.

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