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The transport, transformation and dispersal of sediment by buoyant coastal flows

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Abstract

Rivers provide the dominant pathway of terrigenous sediment to the ocean. The density difference between riverine and salt water as well as the density anomaly contributed by sediment have important consequences on the delivery, transport and ultimate fate of the sediment issued from the land. Most fresh water outflows produce positively buoyant plumes; however, the aggregation and settling of sediment cause the route of sediment transport to diverge from the route of fresh water flow. Sediment is trapped in frontal zones, both in estuaries and on the inner shelf, often resulting in large increases in the concentration of sediment relative to the riverine source. At high concentrations, the density anomaly due to the sediment itself contributes to the vertical stability of the flow, often increasing the trapping efficiency of the frontal zone. Sediment trapping may also occur within the wave boundary layer on the continental shelf, leading to high concentrations within a layer only on the order of 10 cm thick, but which may represent an important cross-shelf conduit of sediment. The high concentration layers formed both by frontal trapping and by wave boundary layer trapping can attain excess densities large enough to initiate down-slope motion on the continental shelf, resulting in an important cross-shelf transport process. Because of these trapping processes, the formation of high-concentration layers, and the occurrence of hyperpycnal plumes, the transport of sediment in river-influenced environments is often dominated by near-bottom fluxes rather than fluxes in the surface plume.

Keywords: Coastal currents; Sediment transport; River plumes

1. Introduction

The run-off from continents produces a total riverine discharge of approximately $10^6\,m^3/s$ to the oceans (Broeker, 1974; Berner and Berner, 1996). Although this is a small number relative to the transport in ocean currents, the density contrast between fresh- and salt waters provides an important driving force for coastal circulation, both in estuaries and on the continental shelf. Continental run-off is also of immense importance to the biogeochemistry of the oceans, being the dominant transport route from the terrestrial to the marine environments. Milliman and Syvitski (1992) estimate that 10–20 billion metric tons of sediment are transported annually in the world's rivers, albeit with considerable uncertainty due to rapid historical changes in land-use and river diversions. The delivery of that sediment to the

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Nomenclature

a	coefficient in rating curve
b	exponent in rating curve
f	*Coriolis frequency*
g	acceleration of gravity
g'	reduced gravity ($g(\Delta\rho/\rho)$)
h_0	water depth
h_T	thickness of turbidity current
h_p	thickness of plume
w_s	settling velocity
C_d	drag coefficient
F_0	froude number
L_f	width of frontal zone
L_I	inertial radius
L_p	width of plume
L_s	distance sediment is advected by plume
Q_f	fresh water discharge
Q_R	river discharge
Q_s	sediment flux
T	transport number
U_0	horizontal velocity
U_p	velocity in the plume
U_R	river outflow velocity
U_T	velocity of turbidity current
α	bottom slope
$\Delta\rho$	density difference between fresh- and salt water
$\bar{\rho}$	mean density

marine environment and the ultimate fate of this sedimentary material are controlled by a sequence of fluvial, estuarine and marine processes, all of which are affected by the fresh water flow.

This review focuses on the processes influencing the transport and fate of sediment within the estuarine and marine environment, particularly those processes associated with the density difference between the sediment-laden water and the ambient seawater, i.e., the effects of buoyancy of the outflow. The review first addresses the delivery of fresh water and sediment from riverine systems to the oceans, with particular attention to the variability of input between different types of systems and the variability of sediment load with discharge. It provides a brief background on particle-aggregation processes, because of their importance in controlling the settling rate of particles in the estuarine and marine environments. Then it considers the processes influencing sediment transport as it is transported in turn to estuarine, coastal and marine environments, focusing mainly on those processes influenced by the density differences between the sediment-laden outflow and the ambient waters. As sediment is transported away from the fluvial environment, it ultimately loses its association with the fresh water that delivered it to the sea. Nevertheless, buoyancy continues to play an important role in the transport and fate of the sediment, both due to the ambient stratification of the coastal environment and the density anomaly associated with the sediment. This review considers these buoyancy effects, such as the stratification by suspended sediments and hyperpycnal flows of remobilized sediment on the continental margin. The review concludes with a discussion of directions for future research.

2. The delivery of sediment from the continents

Milliman and Syvitski (1992) estimate that approximately 20 billion metric tons of sediment per year would be delivered to the oceans, if not for the interception by dams. This is a difficult number to estimate, due to the tremendous variability of sediment load between different systems. Their analysis indicates that the size and relief of the watershed are the two most important variables in determining sediment flux, and the fresh water run-off is relatively less important. Farming and deforestation have increased sediment yields in the last 2000–2500 years, perhaps increasing the total sediment yield by a factor of 2 (Saunders and Young, 1983; Berner and Berner, 1996). Dams mitigate this increased sediment load; thus, the amount of sediment making it to the coast may be comparable to the precivilization rate (Meade and Parker, 1985), estimated by

Milliman and Syvitski to be about 10 billion tons per year.

Milliman and Syvitski (1992) also note that smaller rivers provide a sediment flux disproportionate to their size, due to steep gradients associated with the smaller watersheds of tectonically active margins relative to the relatively gentle relief of the large watersheds on passive margins. Moreover, the trapping of sediment in the flood plains of large river systems results in subaerial deposition of a significant fraction, from 30% to more than 90% of the sediment flux (Goodbred and Kuehl, 1999; Meade and Parker, 1985). The sediment delivery of small, steep rivers tends to be highly episodic, due both to the variability of runoff and the susceptibility of such watersheds to hillslope failures (Wohl, 2000, p. 29).

Wolman and Miller (1960) found that the long-term average sediment flux in rivers is determined typically by the annual maximum event, e.g., the spring freshet in a temperate watershed. Nash (1994) revises the Wolman and Miller result with extensive data on a variety of watershed sizes, indicating that there is a tremendous variability in the recurrence interval of the "effective" discharge, from 1 week to several decades. The sediment discharge tends to follow a power law

$$Q_s = aQ_R^b, \tag{1}$$

where Q_s is the sediment flux, Q_R is the river discharge and the exponent b has a value typically between 1.5 and 2.5 (Nash, 1994). This indicates that the sediment concentration increases approximately linearly with flow, producing much higher sediment fluxes during high flow periods. Depending on the configuration of the stream, the actual flux may deviate from a simple power law, for example, due to enhanced bank erosion at a certain flood stage. The power-law behavior is much more important for small streams, which exhibit a much wider range of flow magnitude, than large river systems with less variability in Q_R. For example, the Amazon's discharge only varies by a factor of 3 between its annual minimum and maximum discharge, and the sediment load, which varies by a factor of 6, is distributed commensurately through the year. In contrast, the Eel River in northern California varies by several orders of magnitude in discharge, and most of its annual sediment load is delivered in several large winter storms.

The size distribution of riverine sediment load varies as a function of the source rock characteristics (Ritter, 1978) as well as the gradient. Significant delivery of sand and coarse silt to the river mouth requires a steep gradient, usually associated with smaller rivers on active margins. Larger rivers tend to be dominated by fine sediments (Ritter, 1978).

3. Aggregation, flocculation and particle dynamics

Patterns and mechanisms of suspended sediment delivery to the ocean by plumes depend fundamentally on grain size (Syvitski et al., 1988; Orton and Reading, 1993). Grain size influences the interaction between sediment-laden river waters and ambient basin waters, and it affects the frequency, magnitude and type of sedimentation processes (Orton and Reading, 1993; Wright and Nittrouer, 1995). Grain size exerts influence through its effect on settling velocity and erodibility, which determine, respectively, the rate of sediment loss from plumes and post-depositional mobility (e.g., Syvitski et al., 1988; Wiberg and Smith, 1987).

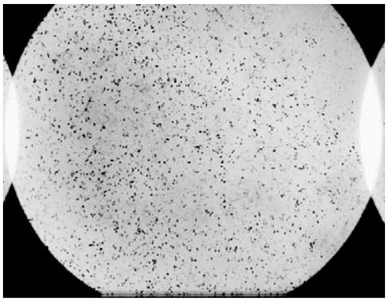
Grain size in discharge plumes depends critically on delivery from the river system. It varies widely and responds to many factors, including climate, catchment geology, basin area and relief, and the size selectivity of the dominant erosion, delivery and deposition mechanisms operating within a catchment (Walling and Moorehead, 1989; Stone and Walling, 1997; de Boer and Stone, 1999). Grain size also can vary temporally with river discharge, although in a way that at present eludes prediction. In some river systems an increase in discharge introduces coarser sediment into suspension because of increased bottom shear stress. In others, however, suspended sediment gets finer as discharge increases, likely due to enhanced delivery from hillslopes or side channels. In still others, discharge and sediment size apparently are uncorrelated (Walling and Moorehead, 1989; Walling et al., 2000).

Over the past decade it has become increasingly evident that the component, dispersed grain size distribution found in rivers differs from the in situ grain size (Walling and Moorehead, 1989; Droppo and Ongley, 1994; Slattery and Burt, 1997; Phillips and Walling, 1999). Small inorganic sediment particles are bound together with bacteria, other organisms, and organic detritus into porous agglomerations referred to either as aggregates or flocs (Ongley et al., 1981; Droppo and Ongley, 1994; Droppo, 2001). These agglomerations of particles sink many times faster than their component grains, and deliver a poorly sorted mixture of grain sizes both to riverbeds and floodplains and to discharge plumes that meet the sea (Walling and Moorehead, 1989; Warren and Zimmermann, 1994; Nichols and Walling, 1996).

Flocculation has for decades been recognized as a key process in the rapid removal of fine sediment from turbid suspensions in coastal waters (Whitehouse et al., 1960; Krone, 1962; Postma, 1967). More recently, direct in situ observations of floc size and settling velocity have shown that a substantial fraction of fine sediment in turbid discharge plumes is packaged in large flocs that are hundreds to thousands of micrometers in diameter (Fig. 1) and that sink at speeds in the range of 1 mm/s (Syvitski et al, 1985; Eisma, 1986; Gibbs and Konwar, 1986; Eisma et al., 1991; Kranck and Milligan, 1992; ten Brinke, 1994; Wolanski and Gibbs, 1994; Syvitski et al., 1995; Dyer et al., 1996; Berhane et al., 1997; Hill et al., 1998; Dyer and Manning, 1999; Sternberg et al., 1999; Hill et al,, 2000). Substantial effort has been devoted to understanding the mechanisms and rates of flocculation and deflocculation as well as the way that flocculation and deflocculation interact to produce observed in situ size distributions (McCave, 1984; Eisma, 1986; van Leussen, 1988; Eisma et al., 1991; Kranck et al., 1992; Hill and Nowell, 1995; Syvitski et al., 1995). Despite this effort, understanding has not evolved enough to produce accurate, predictive, process-based models of the kinetics of flocculation and deflocculation.

Two key gaps in understanding that impede the development of models of flocculation and deflocculation merit special attention. They are how particles stick to one another within flocs and how physical forces destroy flocs. In both of these subject areas, long-standing, widely accepted hypotheses have been challenged fundamentally by direct, in situ observations of floc size and the variables thought to control it.

The first hypothesis to be challenged by in situ observations is that salinity controls particle stickiness. Numerous laboratory experiments have



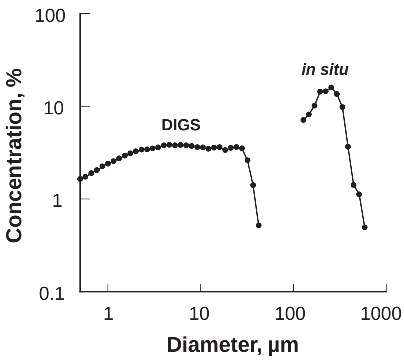


Fig. 1. Disaggregated inorganic size distribution (DIGS) and in situ floc size distribution in the Eel River discharge plume. On the left is a silhouette floc camera image of particles greater than 125 µm in the plume (see Hill et al., 2000 for description of methods). The width of the image is 7.5 cm. Dark objects are flocs. On the right is plotted the in situ size distribution of particles in the image, with object areas converted to equivalent circular diameters. Also plotted is the DIGS distribution as measured with a Coulter Multisizer IIe for a water sample collected at the same time and depth as the photograph. The component grain size distribution (DIGS) is poorly sorted and finer than the in situ floc size distribution (Figure courtesy of T.G. Milligan, Bedford Institute of Oceanography).

shown that sediment suspended in salt water sinks to the bottom of a beaker much more rapidly than chemically dispersed sediment in fresh water (e.g., Whitehouse et al., 1960; Krone, 1962; Kranck, 1980). Based on these observations and backed by the intuitively appealing and rigorous theory of electrical double layers (cf. Bockris and Reddy, 1970), the concept of "salt flocculation" in natural waters evolved. According to this hypothesis, sediment particles in fresh water are dispersed. Upon entering salt water, particles in close proximity cohere due to redistribution of the ion clouds that accumulate around the surfaces of particles in natural waters. Newly formed flocs attain a maximal size set either by deflocculation or by sedimentation (cf. Kranck, 1973). This model has been challenged fundamentally by observations that show no dependence of in situ floc size on salinity (Eisma, 1986; Eisma et al., 1991; Kranck et al., 1992), and by the numerous observations discussed previously that suggest flocculation alters the packaging of fine sediment in fresh water substantially. These observations have shifted focus to organic matter as the mediator of particle stickiness. The concentration and the composition of organic matter, which change continuously (Eisma et al, 1991; Droppo, 2001), and the configuration of organic matter at particle surfaces, which changes in response to salinity and composition (Eisma et al., 1991; O'Melia and Tiller, 1993) all alter particle stickiness. As a result, predictive knowledge of stickiness remains elusive.

The second hypothesis to be challenged by in situ observations of flocs is that maximal floc size depends on small-scale, turbulence-induced shear (Hunt, 1986). No prediction based on this hypothesis has been successful in explaining maximal size of natural flocs under turbulence levels typically found in coastal waters (Alldredge et al, 1990; Hill et al., 2000, 2001). Rather than decreasing systematically as turbulent shear increases, floc size shows little or no dependence on turbulence at low-to-moderate energy, and it decreases abruptly at some threshold level (Hill et al., 2001). This behavior may indicate that floc size is not controlled by turbulence at low-to-moderate energy (Eisma, 1986; Eisma et al., 1991; Hill et al, 2001). As an alternative, forces arising on sinking particles may limit floc size (Adler, 1979; Adler and Mills, 1979; Hill et al., 2001). This lack of understanding of deflocculation mechanisms and rates further limits our ability to predict floc size and settling velocity in discharge plumes.

Without a strong predictive framework, in situ observations of floc size and settling velocity remain the most effective means for gaining knowledge of particle packaging and flux in plumes. Such observations display remarkable similarities among environments (Eisma et al., 1991; Kranck et al., 1992; Hill, 1998). Floc sizes typically are unimodal and well sorted and fall in the range of hundreds to thousands of micrometers. Floc densities decrease as diameter increased to a power that is usually around -0.5 to -1.0 (Kranck et al., 1992; Hill et al., 1998; Sternberg et al., 1999). Direct in situ observations of floc settling velocities yield mean values of order 1 mm/s (cf. Hill, 1998). When turbulence is vigorous, floc sizes are reduced markedly (Eisma, 1986; Milligan and Hill, 1998; Hill et al., 2001).

More in situ observations are required to resolve some outstanding issues. For example, the effect of sediment concentration on floc size and settling velocity remains unclear. Some studies indicate no correlation between floc size and concentration (Hill et al., 2000), others show positive correlation (Kranck and Milligan, 1992), and still others display negative correlation (Dyer and Manning, 1999). The idea that floc settling velocity increases with increasing concentration is firmly entrenched in the literature and is based on data from settling tubes in which the clearance rate is used to infer settling velocity (e.g., Dyer et al., 1996, van Leussen, 1999). Direct observations of floc settling velocity, however, have failed to support such a relationship (ten Brinke, 1994; Dyer et al., 1996; Hill et al., 1998), suggesting that in settling tubes, particle repackaging, and not floc settling velocity, may be responsible for the observed dependence of clearance rate on concentration (Milligan, 1995; Milligan and Hill, 1998; Hill et al., 1998). Floc fraction, which is the fraction of suspended mass contained within flocs, has only been estimated in a few studies (Syvitski et al., 1995; Dyer and Manning, 1999), so it is not clear how this important variable responds to factors such as turbulence and sediment concentration. Hill et al. (2001) propose that floc size and settling velocity decrease abruptly at a turbulence threshold. This hypothesis requires more data to assess its validity and determine controls on the threshold level of turbulence. Finally, in situ devices that extend the range of observations across the full range of particle diameter from micrometers to millimetres are necessary (Eisma et al., 1991; Jackson et al., 1997).

In summary, the conventional view of sedimentation from plumes is one of dispersed sediment entering the ocean, undergoing flocculation, attaining a maximal size set by turbulence and then sinking. This view is not supported by observations. Sediment is already packaged in flocs in fresh water. Rather than forcing incremental changes in floc size, turbulence apparently produces abrupt, large changes in floc size at relatively high-energy levels. Kinetics of flocculation and deflocculation remain poorly constrained, but because suspensions typically are packaged primarily in large flocs that sink at around 1 mm/s, lack of understanding of flocculation kinetics may not impede our understanding of plume processes (Hill and McCave, 2001).

4. Estuaries: traps or conduits?

When rivers meet the ocean, the density difference between fresh- and salt water presents an impediment to the seaward transport of both the fresh water and its load of sediment. The Weser estuary (Fig. 2) has moderate fresh water outflow ($\sim 500\,\mathrm{m^3/s}$), and its salinity intrusion extends inland approximately 20 km from its mouth (Grabemann et al., 1997). Sediment is trapped at the landward limit of the salinity intrusion, producing an estuarine turbidity maximum (ETM). In larger river systems such as the Amazon (200,000 $\mathrm{m^3/s}$) the momentum of the fresh water outflow is strong enough to overcome the opposing force of the seawater, and the fresh water and sediment will escape directly into the coastal ocean (Fig. 2). The Amazon has large enough flow year round to prevent the intrusion of salt water into the river mouth (Gibbs, 1970, Geyer and Kineke, 1995), allowing unimpeded transport of sediment to the continental shelf. Other estuarine systems, such as the Mississippi, alternate between these two modes of trapping depending on seasonal variations in river flow, trapping sediment within the estuary during low flow conditions and expelling it onto the continental shelf during freshet conditions (Wright, 1971).

The balance between fresh water outflow and the density-driven inflow of salt water is expressed by the densimetric Froude number

$$F_0 = \frac{U_R}{((\Delta\rho/\bar{\rho})gh_0)^{1/2}}, \qquad (2)$$

where U_R is the river outflow velocity, $\Delta\rho$ is the density difference between fresh- and salt water, $\bar{\rho}$ is the mean density, g is the acceleration of gravity and h_0 is the water depth. Armi and Farmer (1986) showed that a steady flow with $F_0 \geqslant 1$ will arrest the salt front, thus preventing the intrusion of salt water into the river. Tides complicate the frontal conditions. For strong river outflows such as the Mississippi or the Amazon, frontal conditions still occur in the presence of tides (Wright, 1971; Geyer and Kineke, 1995), but the critical condition for the front is determined by an effective velocity that involves the combined effects of the river flow and the ebbing current. The stationary frontal condition described by Armi and Farmer is maintained during the ebb, but the front advances landward during the flood. In the Fraser River estuary, with a tidal range of 4 m, the salinity front travels as much as 18 km into the estuary during the flooding tide (Geyer and Farmer, 1989).

Sediment is trapped in estuaries due to several phenomena, but the most pervasive and generally important mechanism is the bottom convergence at the landward limit of the salt intrusion. Landward flow due to the estuarine circulation meets seaward flow due to the river outflow, trapping near-bottom sediment in the estuarine turbidity maximum zone (Postma, 1967; Schubel, 1972). The efficiency of the trapping depends on the position of the convergence zone, the strength of convergence and the settling velocity of the

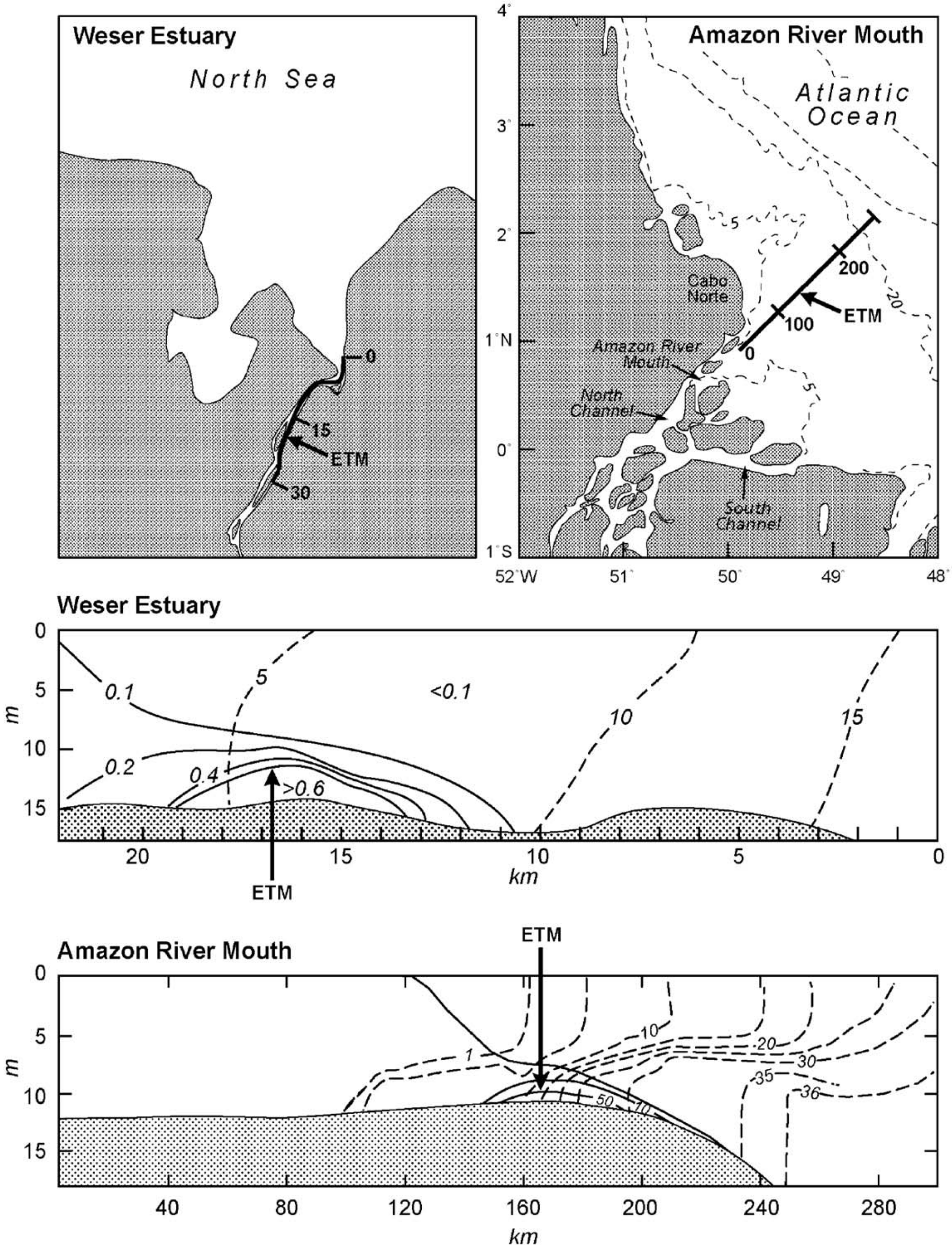


Fig. 2. Cross-sections of suspended sediment concentration in g/l (solid lines) and salinity (dashed lines) in the Weser estuary (from Grabemann et al., 1997) and the Amazon River mouth (from Geyer and Kineke, 1995). The key difference between these two environments is that the strong fresh water outflow of the Amazon has ''pushed'' the ETM onto the inner shelf.

sediment (Festa and Hansen, 1978; Allen et al. 1980). The further the salt intrusion into the estuary, the more likely sediment is to be trapped. The trapping efficiency also depends on the relationship between settling velocity and convergence rate within the salt intrusion. For typical scales in estuaries, sediment with settling velocities around 1 mm/s tend to be most effectively trapped. Slower-settling particles are carried out of the estuary in the upper layer, and coarser particles do not remain in suspension for long enough to have strong interaction with the estuarine circulation (Geyer, 1993).

The immediate source of sediment to the water column within the ETM is usually from tidal resuspension of bottom sediments (Fig. 3). The convergence within the water column leads to preferential deposition at the ETM, which provides a source of bed sediment that is readily resuspended by tidal flow (Grabemann and Krause, 1989). This mobile pool of sediment is maintained by the water column convergence processes, but changes in the hydrographic regime may occur more rapidly than the adjustment time scale of this mobile pool. As a consequence, the position of the turbidity maximum may deviate from the position of maximum water-column convergence. Over seasonal time scales, the position of the pool adjusts in response to changes in the position of the salt front. Migniot (1971) documented zones of unconsolidated mud as much as a meter thick that ranged 30 km up and down the Gironde estuary, accumulating near the mouth during high flow and moving landward tens of km during low flow. Similar seasonal variations in the Hudson estuary were observed by Woodruff et al. (2001), in which concurrent observations of the salinity field indicated that the variation in the position of the salinity intrusion was responsible for the changes in the position of sediment trapping.

Spring-neap variations in tidal amplitude often result in large variations in the concentration of suspended sediments in estuaries (Fig. 3) due to changes in the intensity of resuspension (Castaing and Allen, 1981; Geyer et al., 2001). Energetic

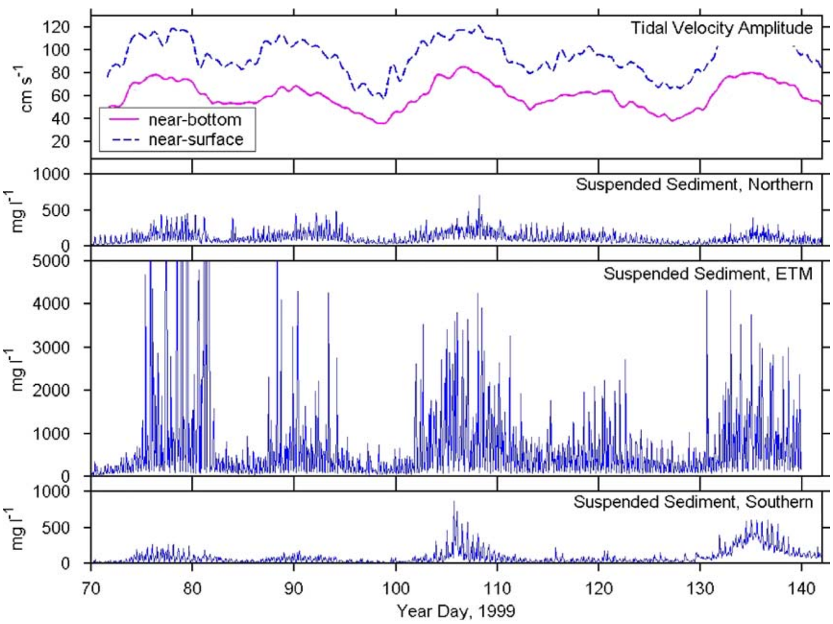


Fig. 3. Time series of suspended sediment concentration at three locations in the Hudson River estuary (from Geyer et al., 2001), showing the influence of tides on resuspension. Tidal velocity amplitude is shown in the upper panel to indicate the spring-neap variability.

tides often cause the retention of sediment due to flood-tide dominance of sediment transport (Uncles and Stephens, 1989; Allen et al., 1980). However, the increased suspended load during spring tides may also lead to significant sediment export, if the spring tides correspond to periods of strong fresh water outflow, as demonstrated by Castaing and Allen (1981) in the Gironde estuary. They showed that suspended sediment is distributed higher in the water column during spring tides, where it is carried seaward by the net surface outflow. They also noted that tidal dispersion augments the export of sediment during high flow and spring tide conditions, when the turbidity maximum is pushed closer to the mouth. Geyer et al. (2001) note that the timing of the spring freshet with respect to the spring-neap cycle may be as important as the actual magnitude of the freshet in determining whether sediment is exported during high flow events.

Large and deep estuaries such as Chesapeake Bay and Puget Sound never have strong enough flow to push the salt front to the vicinity of the mouth (Schubel, 1972). The seaward escape of sediment is limited to the fine fraction that remains in the surface outflow during high flow events (Stumpf, 1988), which is typically a small fraction of the input due to flocculation and settling. Such estuaries trap most of the fluvial sediment input, and, in many instances, the input of sediment from the sea as well (Meade, 1972). Given a long enough time interval, the shoaling would lead to seaward movement of the salinity front, the increase in export of sediment and the establishment of a morphological equilibrium. However, the variability of sea level during the Holocene has outpaced the ability of many estuarine systems to achieve this dynamic equilibrium (Dyer, 1995).

5. River outflows

5.1. Near-field and frontal zone

Riverine outflow onto the inner continental shelf may be either stratified or well mixed, depending on the strength of the outflow relative to the intensity of mixing in the estuary and inner shelf. Outflows with shallow receiving waters and strong tide- or wind-driven mixing exhibit a well-mixed region on the inner shelf adjacent to the river mouth. The Huanghe (Yellow River) enters a shallow shelf with water depths on the order of 5 m and strong tidal currents that produce a well-mixed zone. (Wright et al., 1990, Wiseman et al., 1986). The Amazon river enters an inner shelf zone 5–10 m deep with tidal currents of 1–2 m/s (Fig. 2), producing a well-mixed zone that extends approximately 150 km seaward of the river mouth (Geyer and Kineke, 1995). The large volume of river outflow renders this mixed zone nearly fresh. In contrast, the Huanghe has typical salinities in the well-mixed zone of 5 psu during high flow periods and 20 psu during low flow (Wright et al., 1986). The Atchafalaya outflow in Louisiana also enters a shallow inner shelf, approximately 5-m deep. Although tidal currents are weak, strong winds intermittently produce a well-mixed plume (Murray 1997; Allison et al., 2000) (Fig. 4).

Intense sediment resuspension occurs within the well-mixed zone, due to the influence of tidal- or wind-driven currents, and the absence of stratification allows turbid sediment to mix through the water column. The associated turbidity plumes are clearly evident in satellite images of large river mouths as broad regions of high reflectance (Curtin and Legeckis, 1986; Wright et al., 1986; Walker and Hammack, 2000) (Fig. 5). Sediment transport tends to be highly variable in the well-mixed zone, due to varying contributions of tidal currents, wind-driven motions causing wave resuspension and riverine outflow. Wiseman et al. (1986) and Wright et al. (1990) found that tidal currents produced along-coast dispersion of Huanghe sediment in the well-mixed inner shelf, although the spatial extent of that transport was limited to the shallow water adjacent to the mouth. Sediment accumulation rates tend to be low on the seabed beneath well-mixed plumes, due to the absence of accommodation space (Nittrouer, pers. comm., 2001) and the high energy for resuspension. However, Kuehl et al. (1996) and Jaeger and Nittrouer (1995) do observe rapid sediment deposition and erosion over tidal and seasonal time scales.

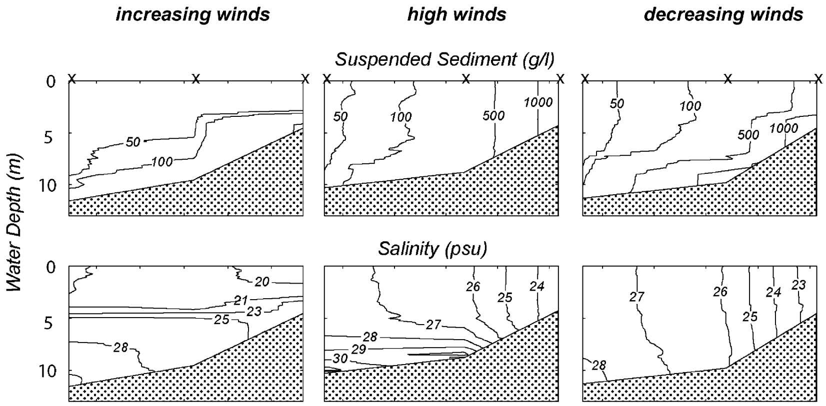


Fig. 4. Cross-sections of suspended sediments (upper panels) and salinity across the inner shelf adjacent to the mouth of the Atchafalaya River, showing the influence of winds in generating well-mixed conditions on the inner shelf (Allison et al., 2000).

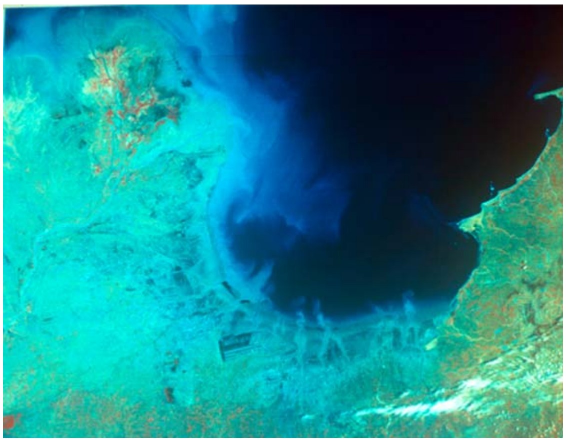


Fig. 5. Satellite image of the Huanghe outflow (courtesy of L.D. Wright), showing the high suspended sediment concentrations in the shallow waters of the inner shelf and the limited penetration into deeper water.

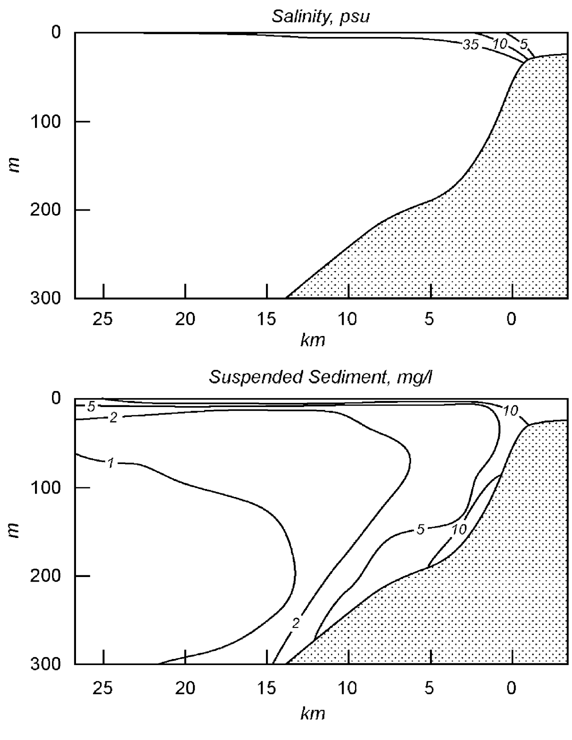


Fig. 6. Cross-section of salinity and suspended sediment in the Zaire (Congo) River (from Eisma and Kalf, 1984). The salinity front is extremely abrupt, owing to the sharp change in depth at the head of the sumarine canyon. Trapping of sediment occurs in the canyon but not at the front itself.

In outflows with deep receiving waters or weak tides, the outflow may transform abruptly from a fluvial regime to a stratified plume, with no well-mixed inner shelf zone. The Zaire (or Congo) and Sepik River outflow in Papua New Guinea both have canyons that extend landward into their river mouths. They both exhibit a sharp salinity front at the head of the canyon, and the outflow enters the ocean as a highly stratified plume (Eisma and Kalf, 1984; Kineke et al., 2000) (Fig. 6). The frontal zone is extremely abrupt, extending for as little as a few hundred meters between the fluvial-dominated river and the marine environment. The Mississippi is notable both for its strong outflow and weak tides. During freshet conditions, the seawater is pushed out to the mouths of its distributaries, where strong fronts occur (Wright and Coleman, 1974). During lower discharge periods, a strongly stratified salt wedge extends into the lower reaches of the river (Wright, 1971). Hydraulic transitions are found at the mouths of the distributaries, where the halocline rises sharply over the rivermouth bar to form a plume (Wright and Coleman, 1971). A sharp frontal transition does not require weak tidal flows; the Fraser River has a similarly abrupt transition, occurring where the fluvial regime meets the deep water of the Strait of Georgia, with tidal currents greater than 1 m/s. The front only resides at the mouth of the Fraser during the latter part of the ebb, and it is advected into the estuary as a salt wedge during the flood (Geyer and Farmer, 1989).

All major river outflows eventually encounter a frontal transition to stratified conditions at some distance from the mouth, even if they are well mixed in the shallow inner shelf. The position and extent of the frontal zone depends on the bathymetry and tidal conditions of the receiving waters. Strong tidal currents and shallow receiving waters result in a broader frontal zone, at greater distance from the mouth, than weak tidal currents and deep receiving waters. An extreme case is the Amazon River, in which the frontal zone is 150 km seaward of the river mouth (Geyer and Kineke, 1995; Curtin and Legeckis, 1986). The frontal zone extends approximately 20 km seaward between the 10- and 20-m isobaths. The frontal zone at the mouth of the Changjiang River is similarly located between the 10- and 20-m isobaths, approximately 50 km seaward of the mouth (Beardsley et al., 1985).

The position of the frontal zone is governed by a dynamical balance between the outflow velocity and the density gradient, leading to a critical Froude number condition as represented in Eq. (2). The combination of river outflow and ebb tidal velocity typically amounts to 1–2 m/s, which constrains the front to be in the vicinity of the 10-m isobath, given the density difference between fresh- and salt water. The width of the front is a more complicated function of the tidal mixing intensity and bottom slope in the vicinity of the front. If the topography is abrupt, as in the Fraser or Zaire river mouths, the front is commensurately abrupt (Fig. 6). However, if the depth slopes more gently, and particularly if tidal currents are strong, the frontal zone may extend over 10's of km. The gradient of near-bottom salinity in the Amazon frontal zone extends over 50 km in the cross-shore direction (Fig. 2) (Geyer and Kineke, 1995). The Amazon frontal zone has a similar cross-shore salinity distribution as an estuary; in fact, it has dynamics and kinematics similar to an estuary, except for the absence of lateral boundaries, thus permitting transport in the along-front direction. From a dynamical point of view, inner-shelf frontal zones can be regarded as estuaries that have been displaced onto the shelf due to the combination of strong river outflow and ebb tidal currents.

The frontal zone may be an effective sediment trap, due to the same mechanisms that make an estuary an effective sediment trap, i.e., convergence of near-bottom flow (Postma, 1967) and separation of the outflow from bottom-generated turbulence (Geyer, 1993). On the landward side of the front, the sediment is distributed throughout the water column, maintained in suspension by vigorous bottom turbulence. Across the front, stratification increases, suppressing turbulence in the upper part of the water column, even if tidal currents are strong. The sediment-laden fresh water is advected over the saline layer into the plume, but with the shut-off of bottom turbulence, the settling is no longer balanced by resuspension, and sediment begins to rain out of the plume. Enhanced flocculation may occur in the frontal zone due to a decrease in turbulence, which increases the settling velocity and further promotes trapping of sediment.

Whether or not there is significant sediment trapping in the frontal zone depends on the width of the frontal zone and the distance that sediment can be transported before it settles. If L_f is the frontal width, significant frontal trapping will occur for

$$L_f \geqslant \frac{U_0}{w_s} h_0. \tag{3}$$

The Amazon frontal zone exhibits intense frontal trapping of fine sediment; with characteristic horizontal velocities of 1.5 m/s, settling velocities of 1 mm/s and a frontal zone depth of 10 m, the horizontal scale for settling of sediment is 15 km, compared to a 50 km width of the frontal zone. The trapping of sediment in the frontal zone of the Amazon leads to concentrations high enough to produce fluid mud (concentration greater than 10 g/l), which is the concentration at which the settling velocity starts to be impeded by inter-particle interactions. These high concentrations also produce excess densities that are dynamically significant, both with respect to the stratification of the water column and horizontal pressure gradients. The consequences of these high concentrations for subsequent transport are discussed in Section 6.

For river outflows with narrow frontal zones and/or very fine sediments, the horizontal scale for settling may greatly exceed the width of the front, and then the sediment will bypass the frontal zone. The Sepik River outflow is an example of an abrupt frontal zone, in which the trapping scale for fine sediment exceeds the frontal width, which is only several hundred meters wide. Whereas sand settles rapidly enough to accumulate at the front, fine sediment is transported over the front in the plume. Settling of aggregates causes settling out of the plume close to the plume lift-off, but rather than collecting in the frontal zone, much of the fine sediment falls into the canyon at the mouth of the river.

An important difference between frontal trapping of sediment and bypassing relates to the potential for remobilization of the sediment and its concomitant influence on the sediment chemistry. The shallow depths of the frontal zone have adequate energy due to tides and waves to resuspend the sediment, either periodically with tidal oscillations or episodically, with wind and wave events. Thus, the sediment that is trapped in frontal zones may be remobilized multiple times before its ultimate burial. The remobilization of sediment, and particularly its cycling through a broad range of oxidation–reduction conditions, causes more complete chemical processing of the sediment than the sediment that bypasses the frontal zone (Aller, 1998).

5.2. River plumes and coastal currents

The river plume is the continuation of the fresh water outflow beyond the frontal zone. The vertical scales of plumes typically range from 1 to 10 m, but the horizontal scales vary over several orders of magnitude, from hundreds of meters to hundreds or even thousands of km for the largest river outflows. The processes affecting the dynamics of river outflows vary as a function of scale; even within a particular plume the processes vary with distance from the mouth as the effective lengthscale of the plume increases. For small plumes and close to the mouth for larger outflows, the inertia of the outflow is a dominant dynamical variable. The Connecticut River plume (Garvine, 1974) is an example of an inertia-dominated plume, in which the conditions at the mouth markedly affect its structure and trajectory. The scale of influence of inertia is set by the inertial radius $L_I = U_p/f$ (where U_p is the velocity in the plume and f is the Coriolis frequency); this scale tends to be around 10 km at mid-latitudes. At larger distances, the influence of the earth's rotation becomes dominant. Even within the inertial zone, the earth's rotation is often evident when the plume is not strongly forced by winds or along-shelf currents—this is evident in an anticyclonic turning region (to the right in the Northern hemisphere) over scales of approximately 10 km (Chao, 1988).

At scales greater than the inertial radius, the dynamics of the plume are more strongly influenced by the earth's rotation, winds and ambient currents. If wind-forcing and ambient currents are weak or directed in the downwelling-favorable direction (to the right looking offshore in the Northern Hemisphere, called "downcoast" henceforth), the plume becomes a coastal current, flowing parallel to the coast. The cross-shelf density gradient becomes geostrophically balanced with the along-shelf velocity shear, leading to the approximate relation

$$U_P = \frac{g' h_0}{f L_P}, \tag{4}$$

where h_0 is the water depth where the plume intersects the bottom at its inshore end and L_P

is its width and g' is reduced gravity, $g(\Delta\rho/\rho)$ (Fig. 7). This relation defines the baroclinic component of the transport—that part due to the density gradient associated with the plume. An additional component of velocity can be imposed barotropically, due either to along-coast wind stress or larger-scale currents. Typical baroclinic velocities are on the order of 20 cm/s for moderate-sized, mid-latitude plumes, with values of g' of around M S^{-2}, h_0 around 10 m and L_P around 10 km (Fig. 8). The widths of plumes (and the associated along-shelf velocity) may vary considerably due to wind forcing—strong downwelling-favorable winds will make the plume narrower and faster, and upwelling winds will broaden and slow down the plume, or reverse its direction by setting up a barotropic gradient that exceeds the baroclinic gradient (Berdeal et al., 2002).

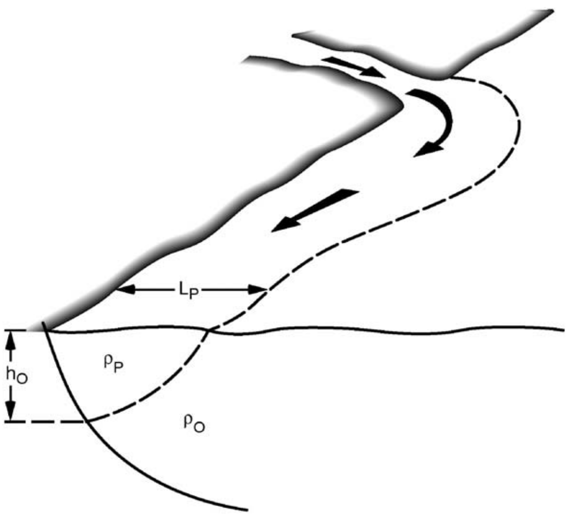


Fig. 7. Schematic of coastal current in the Northern hemisphere.

The role of river plumes in sediment transport is strongly influenced by whether or not they are attached to the bottom. Yankovsky and Chapman (1997) showed that a key parameter influencing the attachment of the plume to the bottom is the "transport number"

$$T = \frac{2Q_f f}{g' h_0^2}, \tag{5}$$

where Q_f is fresh water discharge. Yankovsky and Chapman showed that if $T < 1$, i.e., deep receiving waters and/or low to moderate discharge, the plume remains detached from the bottom as it develops a down-coast, geostrophic flow. For shallow receiving waters and large discharge volumes in which $T > 1$, the plume extends from the surface to the bottom in the far field. This condition cannot be satisfied even for the biggest river outflows with large values of g', i.e., when there has been little mixing in the plume between the riverine and ambient salt water. However, if there is significant dilution of the outflow in the estuary or nearfield, e.g., due to tidal mixing, then the outflow Q_f is enhanced, and g' is reduced, leading to surface-to-bottom plumes. Most major

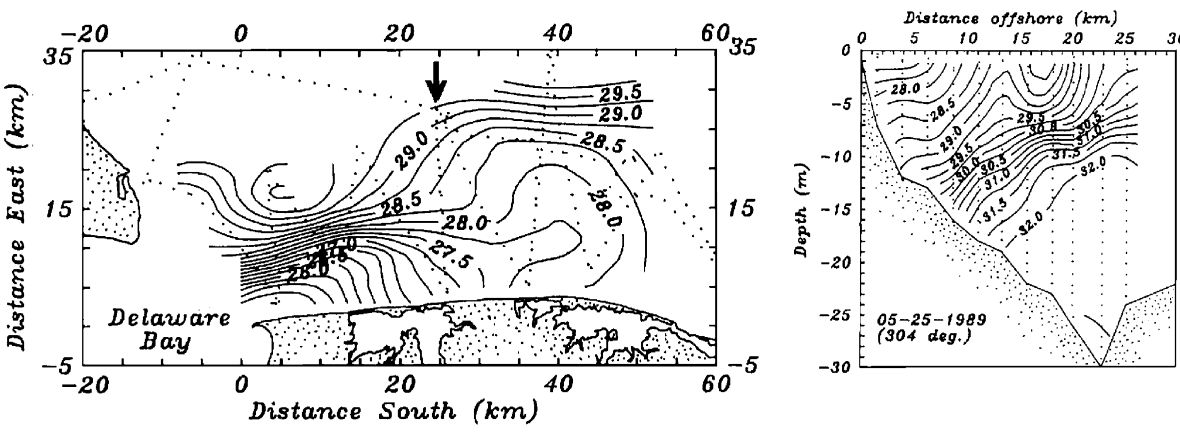


Fig. 8. Plan view and cross-section of the Delaware coastal current during high flow (from Munchow and Garvine, 1993).

river outflows, including the Amazon, the Changjiang, the Ganges-Bramaputra fit into the intermediate category of the Yankovsky and Chapman classification, in which the inshore end of the plume is bottom-attached, but the plume lifts off from the bottom close to the 10-m isobath, with appreciable transport occurring in a surface plume.

The mechanisms of far-field sediment transport by the buoyant outflow differ markedly between bottom-attached and surface plumes. In surface plumes, there is no mechanism of vertical flux to keep sediment in the plume, thus it rains out at a rate determined by the settling velocity and plume thickness. The distance that the sediment is advected by the plume can be estimated from the settling velocity and the speed and thickness of the plume (similar to Eq. (3)):

$$L_s = \frac{U_p}{w_s} h_p, \tag{6}$$

where U_p is the vertically averaged velocity in the plume, w_s is the settling velocity, and h_p is the thickness of the plume (Hill et al., 2000). For a 5-m thick plume at 1 m/s, flocculated mud with a settling velocity of 1 mm/s would be carried 5 km by the plume. Fine sand, with a settling velocity of 1 cm/s, would be transported only 500 m. Unflocculated sediment, with settling velocities of 0.1 mm/s or less, could be carried 50 km or more. However, only a small fraction of the total sediment inventory is usually unflocculated (Hill et al., 2000) so the transport of even fine sediment by surface plumes is generally limited to kilometers from the "lift-off" point.

Bottom-attached plumes can potentially carry sediment further than the limits dictated by settling, because bottom resuspension can maintain sediment in the water column at large distances from the river mouth. The limiting constraints on the far-field transport for bottom-attached plumes are the energy for resuspension and the along-shore velocity in the bottom boundary layer. In the absence of additional forcing variables such as along-shelf winds or large-scale pressure gradients, the baroclinically driven velocity associated with the bottom-attached plume tends to go to zero at the bottom (Chapman and Lentz, 1994). Thus the plume neither provides a significant advective contribution nor does it generate appreciable bottom stress. Tidal currents often provide the energy source for resuspension in the shallow, inner shelf environments where bottom-trapped plumes occur, for example, the Huanghe (Wiseman et al., 1986; Wright et al., 1990). However, the strong dissipation in shallow, tide-dominated environments inhibits along-shelf advection. Tides themselves do not provide significant advective transport beyond their excursion scales of about 10 km.

There are examples of dispersal of sediment a thousand km or more in coastal "mud streams", notable examples being the South American coast north of the Amazon, and the Texas–Louisiana coast to the west of the Missisippi. The Amazon mud stream extends 2000 km to the mouth of the Orinoco, leading to the accretion of numerous "mud-capes" along the coasts of French Guiana, Surinam and Guiana (Fig. 9). Allison et al. (1995) investigated the sediment transport processes responsible for the formation and evolution of these mud capes. Most of the sediment transport is confined to a zone very near the coast, in water depths of several meters or less. The water has a salinity anomaly due to the fresh water influence of the Amazon, but there is no salt stratification and only minor dynamical influence of the salinity gradient in the shallow, nearshore waters. The northward transport of sediment is due mostly to the wind-driven current, which is strongest when the southeasterly trades are maximal. Both surface waves and tidal currents contribute to resuspension of sediment. The seasonal variability of wave energy results in temporary deposits in the nearshore zone that are remobilized during the energetic periods. Allison and colleagues found that the total quantity of sediment transport in this coastal mud stream is a small fraction of the total Amazon sediment flux. Most of the Amazon's sediment is trapped in the frontal zone within a few hundred km of the mouth and deposited along growing foreset beds in water depths of 30–60 m.

Large river systems produce widespread areas of anomalously low salinity, not just along the coast but often extending well into the interior of the

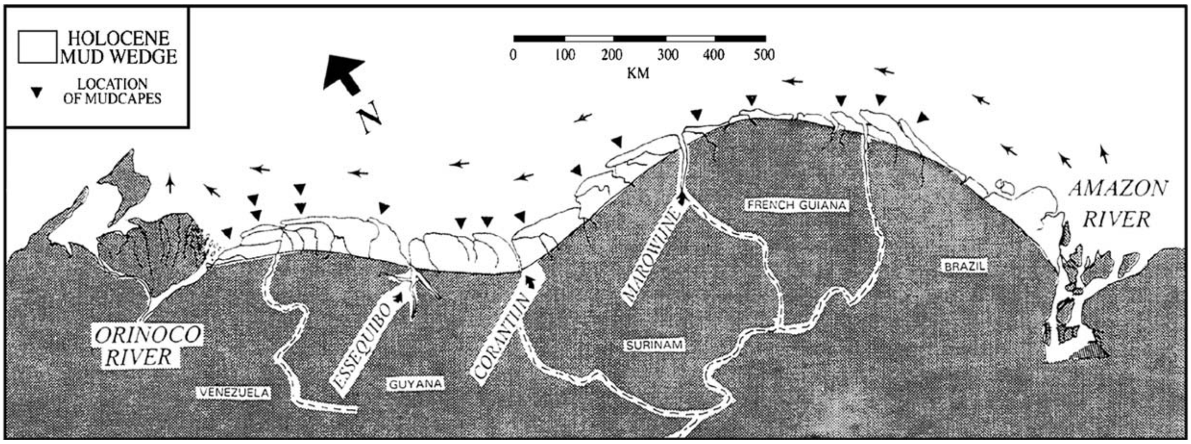


Fig. 9. Coastal mud stream extending from the Amazon River mouth past the coast of Guianna (from Allison et al., 1995).

ocean (Gibbs, 1970; Dinnell and Wiseman, 1986). However, the far-field transport of fresh water in large river systems is largely uncorrelated with the sediment transport, because the bulk of the fresh water is transported in water far too deep to maintain sediment in suspension. Silt and flocculated clay fall out of the plume within km of the mouth, thus, only disaggregated, clay-sized particles can be transported large distances by the fresh water plume. Far-field suspended sediment concentrations in the Amazon plume are on the order of 10 mg/l or less, contributing less than 5% of the total load of the Amazon. In spite of the small load, the Amazon does provide a significant far-field impact on the ecology of the equatorial Atlantic surface waters, due to its contributions of nitrogen and trace nutrients (Muller-Karger et al., 1995). The Mississippi, like the Amazon, has a large far-field salinity plume with modest suspended sediment content. Salisbury et al. (2001) used SeaWIFS satellite data to demonstrate that the dissolved organic matter provides a signal of the Mississippi plume as far as the Texas–Mexico border, whereas the suspended sediment plume is evident only in the vicinity of the delta.

6. Hyperpycnal transport processes

Bates (1953) defined three categories of outflows: hypopycnal, homopycnal and hyperpycnal, corresponding to outflows that are less dense, similar density, or more dense, than the ambient waters. Whereas hyperpycnal outflows are common where rivers enter fresh water systems (Lambert and Hsue, 1979), they are extremely rare in the coastal ocean, due to the density difference between fresh- and salt water. Suspended sediment concentrations of 40–50 g/l are required to produce densities of the outflow that exceed the density of seawater. Numerous authors have inferred the occurrence of hyperpycnal outflows based on sedimentological evidence (e.g., Foster and Carter, 1997; Mulder et al., 1997; Normark et al., 1998; Wright et al., 1990); however, there have been few direct measurements of fluvial suspended sediment concentrations high enough to produce hyperpycnal conditions. The increase of suspended sediment concentration with discharge (Eq. (1)) indicates that hyperpycnal outflows could occur during rare, extreme events, most likely in small river basins with highly variable discharge (Mulder and Syvitski, 1996).

Hyperpycnal flows are not limited to the case of hyperpycnal river discharges. Turbidity currents are the best known examples of hyperpycnal, sediment-laden flows (Middleton, 1993). The simplest balance of forces for a turbidity current is represented by the Chezy equation (the same as for river flows) as

$$C_D U_T^2 = \alpha g' h_T, \tag{7}$$

where C_D is the drag coefficient (accounting for both bottom and interfacial drag), U_T is the velocity of the turbidity current, α is the bottom slope, g' is the reduced gravity associated with the excess density of sediment and h_T is the thickness of the hyperpycnal layer. In the classic models of turbidity currents on slopes, the flow is initiated by a slope failure (for example, due to an earthquake), which initiates a high-concentration suspension that starts moving downslope due to gravity. The bottom-generated turbulence associated with the flow is strong enough to erode the bottom sediment, thus providing an "auto-suspension". This condition requires a steep slope and a supply of erodible sediment from the bed—a set of conditions that is attainable in submarine canyons on the slope but may not be relevant to the more gentle slopes of the continental shelf.

Observations on the continental shelf have indicated that hyperpycnal flows can occur there, even without slope failures or hyperpycnal river inflows (Wright et al., 1990; Kineke et al., 1996). At the mouth of the Huanghe, Wright et al. (1990) found that tidal and wave-induced resuspension of previously deposited, riverine sediment in the well-mixed inner shelf produced concentrations of about 3000 mg/l. The waters had been mixed to salinities of about 20 psu, with relatively weak cross-shore gradients. These sediment concentrations would not be high enough to produce a hyperpycnal flow in fresh water entering seawater, but because the water was already diluted with seawater, the excess density of the sediment was adequate to generate a hyperpycnal flow (Fig. 10).

Observations by Kineke and Sternberg (1995) in the Amazon frontal zone reveal that frontal trapping can also produce hyperpycnal conditions (Fig. 11), again with waters that are close in salinity to ambient. Concentrations initially increase due to the convergence of near-bottom flow; this process is augmented by increased stratification in the frontal zone both due to salinity and suspended sediment, which suppresses turbulence. As concentrations increase, above 10,000 mg/l, hindered settling further augments sediment trapping. Kineke et al. (1996) measured currents within the core of a fluid mud zone indicating offshore transport, counter to the prevailing onshore direction of the overlying flow. Salinity anomalies within the fluid mud also suggest that the near-bottom flow was generally advected in the offshore direction. Partial burial of a tripod at the 60-m isobath (Cacchione et al., 1995) probably occurred as a result off an offshore transport event of fluid mud originating in the frontal zone. The specific conditions required for the episodic transport of fluid muds appears to include both the occurrence of high concentrations (supplied by intense tidal resuspension, wave resuspension and riverine supply) and the relaxation of onshore flows of the ambient bottom waters. The latter condition may be related to wind-driven current fluctuations, but very few observations were available to draw conclusions. The offshore fluxes

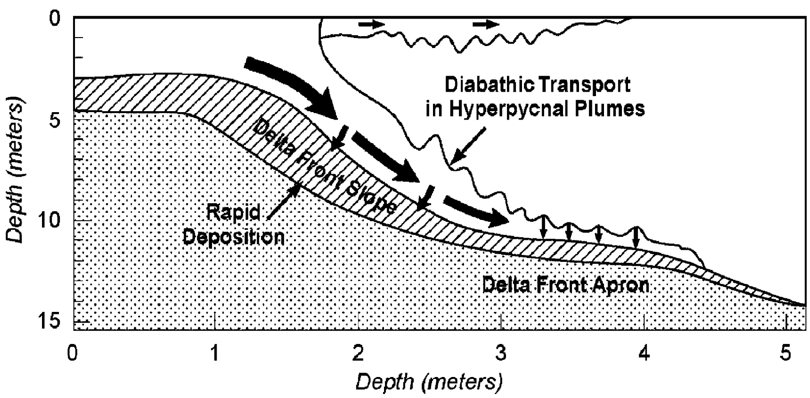


Fig. 10. Schematic of hyperpycnal flows in the Huanghe (from Wright et al., 1990). Note that the dense suspension separates vertically from a buoyant plume over the delta front slope.

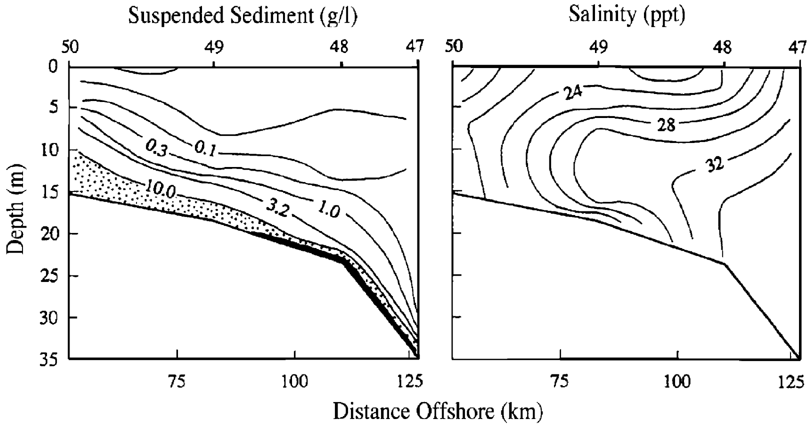


Fig. 11. Suspended sediment and salinity cross-sections across the Amazon shelf during high discharge (from Kineke and Sternberg, 1995). Fluid mud concentrations are stippled. Note the low salinity water within the fluid mud layer, providing evidence of seaward advection.

associated with these hyperpycnal flows appears to be the dominant mechanism of cross-shelf transport from the topset beds to the foreset beds on the Amazon shelf (Kineke et al., 1996).

Observations on the Eel River shelf by Traykovski et al. (2000) provide another case of hyperpycnal transport due to in situ increase in suspended sediment concentration. Unlike the Amazon, the trapping occurs in the wave boundary layer of the inner shelf rather than in a frontal zone. During observations in 1997 and 1998, observed concentrations in the plume never exceeded 1200 mg/l (Geyer et al., 2000), and the dynamics of the effluent were strongly hypopycnal. However, the large amount of sediment that rained out of the plume into the bottom waters between 15- and 40-m water depth provided a source for dense suspensions within the wave boundary layer. Energetic surface waves with periods of 14–16 s produced a turbulent wave boundary layer with a vertical scale of 10–15 cm. Acoustic backscatter measurements within this thin, near-bottom layer at the 60-m isobath indicated intermittent appearance of dense suspensions, contemporaneous with offshore transport in the near-bottom waters. Although there were no direct observations in the "source region" under the plume, the occurrences of these wave boundary layer events followed periods of significant loading from the river plume, and they are consistent with trapping of sediment in the wave boundary layer.

The dynamics of hyperpycnal, sediment-laden flows are significantly different from those on the continental slope, due to the important role of the ambient currents in contributing to turbulence production. Thus, the Chezy equation [7] needs to be modified to include the contribution of the ambient tidal or wave-orbital motions. Traykovski et al. (2000) and Wright et al. (2001) developed dynamical models for hyperpycnal flows on the continental shelf, incorporating the influence of the ambient currents in the wave boundary layer. These models indicate that the maximum transport of sediment occurs with the combination of high wave energy and large sediment supply. The wave orbital motions provide the energy to maintain resuspension, and the large sediment supply not only provides the gravitational force for net seaward transport, it also produces intense stratification at the top of the wave boundary layer, which suppresses mixing with the overlying fluid. Both models were applied to the Eel River shelf for the floods of 1997 and 1998, using observed wave and current forcing conditions and estimated sediment loading from the flooding river. Both the models compared favorably with observations of near-bed velocity and suspended sediment distributions, and they indicated that dense suspensions in the wave boundary layer could provide the major mechanism of cross-shelf sediment flux on the Eel margin. The model of Wright et al. (2001) was also applied to the Huanghe and the Mississippi continental shelf, indicating that these dynamics apply across a range of sediment loading and wave energy conditions.

7. Conclusions and directions for future research

Because of aggregation, settling, trapping and resuspension, the route sediment takes from a river mouth is far more complex than that of the fresh water. Aggregation and trapping processes are particularly important in redirecting fine sediments, the major constituents of large river systems, as they enter coastal waters. Recent research has demonstrated the importance of frontal dynamics and wave boundary layer processes in the trapping of fine sediment, leading to highly concentrated suspensions that are dense enough to generate hyperpycnal flows. The association of sediment trapping on the continental shelf with the generation of hyperpycnal flows is an important mechanism for cross-shelf transport of fine sediment, capable of extending the deposition of sediment beyond the range of transport by surface plumes. Although the fresh water signatures of surface plumes extend thousands of km beyond the mouths of large rivers, the extent of significant sediment transport is much more limited, due to settling of sediment out of the plume after it detaches from the bottom boundary layer. The exceptions are coastal mud streams, which may extend for thousands of km from their riverine sources, but only occur in very shallow water where resuspension can maintain significant sediment loads.

Notwithstanding the recent progress in identifying new mechanisms of sediment transport on the continental shelf, we are a long way from the point at which sediment fluxes and deposition patterns can be predicted, given knowledge of the supply, the geometry and forcing conditions. The crucial importance of settling velocity on the fate of sediment makes the aggregation processes particularly important. The size distributions of aggregates and their variation through the fluvial, estuarine and marine environments need to be better quantified, and these observations need to be used to constrain and refine models of sediment aggregation. Riverine outflows are inherently time variable, with a complex, three-dimensional structure that presents considerable challenges to conventional observational programs. Three-dimensional models can provide an effective complement to observational studies, both an aid in interpreting observations and a means of testing hypotheses about mechanisms. Novel observational approaches are also warranted, in order to better resolve temporal and spatial variability of the structure of river outflows. High-frequency radar (Barrick et al., 1985) may hold promise for spatial resolution of near-surface currents in plumes, although the temporal resolution of existing systems is too coarse for all but the largest plumes. Satellite remote sensing provides very coarse spatial resolution and/or inadequate temporal resolution for most plume environments; however, aircraft remote sensing could provide the appropriate spatial and temporal scales.

Surface observations miss the largest part of the sediment transport signal—in fact, recent studies on the Eel indicate that the bottom-most 15 cm of the water column may hold most of the information with respect to cross-shelf transport. Thus, future studies must increase the resolution of suspended sediment and flow within the wave boundary layer. The processes occurring on these small vertical scales are complex; suspended sediment gradients impact turbulence and settling velocity is affected by particle–particle interactions. Because these processes occur on vertical scales of 10's of cm, they are extremely difficult to examine in the field, but they may be amenable to laboratory studies. Ultimately, the combination of innovative field methods, laboratory experiments and numerical modeling will lead to the understanding and quantitative prediction of the transport of sediment in the boundary layer in regions of intense sediment loading.

Understanding of the sediment transport mechanisms is a prerequisite to understanding the morphodynamics of river mouths and adjacent coastal regions. Future research will focus increasingly on the mechanisms controlling the patterns of erosion and deposition, and how they result in the evolution of the morphology of these environments. Previous geomorphological studies of these environments have by necessity simplified the influence of sediment transport processes as well as reduced the dimensionality of the problem (e.g., Steckler et al., 1999), although there have been recent attempts to examine morphological evolution in a three-dimensional context (Driscoll and Karner, 1999) and with realistically parameterized sediment transport models (Harris and Wiberg, 2001). As models and observations improve, morphodynamic studies will improve commensurately, opening up opportunities to better link sediment transport processes with geomorphology and with the geological record.

Finally, although this review focuses on sediment, the most important issues facing society relate not to sediment, but to materials associated with sediment such as contaminants, nutrients and organic carbon, which have major impacts on both the local and global environment. Better understanding of sediment transport across continental margins will improve our estimates of the rate of remineralization of organic carbon and its burial in sediments. Similarly, the distribution of contaminants in the estuarine and marine environment is closely tied to sediment transport processes; the long-term management and mitigation of contaminant exposure requires a fuller understanding of these sediment transport mechanisms.

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References

Adler, P.M., 1979. A study of disaggregation effects in sedimentation. A.I.Ch.E. Journal 25 (3), 87–493.

Adler, P.M., Mills, P.M., 1979. Motion and rupture of a porous sphere in a linear flow field. Journal of Rheology 23, 25–37.

Alldredge, A.L., Granata, T.C., Gotschalk, C.G., Dickey, T.D., 1990. The physical strength of marine snow and its implications for particle disaggregation in the ocean. Limnology and Oceanography 35, 1415–1428.

Allen, G.P., Salomon, J.C., Bassoullet, P., Du Pengoat, Y., De Grandpre, C., 1980. Effects of tides on mixing and suspended sediment transport in macrotidal estuaries. Sedimentary Geology 26, 69–90.

Aller, R.C., 1998. Mobile deltaic and continental shelf muds as suboxic, fluidized bed reactors. Marine Chemistry 61, 143–155.

Allison, M.A., Nittrouer, C.A., Faria Jr., L.E.C., 1995. Rates and mechanisms of shoreface progradation and retreat downdrift of the Amazon River Mouth. Marine Geology 125, 373–392.

Allison, M.A., Kineke, G.C., Gordon, E.S., Goñi, M.A., 2000. Development and reworking of a seasonal flood deposit on the inner continental shelf off the Atchafalaya River. Continental Shelf Research 20, 2267–2294.

Armi, L., Farmer, D.M., 1986. Maximal two-layer exchange through a contraction with barotropic net flow. Journal of Fluid Mechanics 164, 27–51.

Barrick, D.E., Lipa, B.J., Crissman, R.D., 1985. Mapping surface currents with CODAR. Sea Technology 26, 43–48.

Bates, C.C., 1953. Rational theory of delta formation. Bulletin of the American Association of Petroleum Geology 37, 2119–2162.

Beardsley, R.C., Limeburner, R., Yu, H., Cannon, G.A., 1985. Discharge of the Changjiang (Yangtze River) into the East China Sea. Continental Shelf Research 4, 57–76.

Berdeal, I.G., Hickey, B.M., Kawase, M., 2002. Influence of wind stress and ambient flow on a high discharge river plume. Journal Geophysical Research 107 (C9), 3130.

Berhane, I., Sternberg, R.W., Kineke, G.C., Milligan, T.G., Kranck, K., 1997. The variability of suspended aggregates on the Amazon continental shelf. Continental Shelf Research 17, 267–286.

Berner, E.K., Berner, R.A., 1996. *The Global water Cycle: Geochemistry and Environment*. Prentice-Hall, Englwood Cliffs, NJ, 397pp.

Bockris, J.O'M., Reddy, A.K.N., 1970. Modern Electrochemistry, Vol. 2. Plenum Press, New York, 1432pp.

Broeker, W.S., 1974. Chemical Oceanography. Harcourt Brace Jovanovich, New York 214pp.

Cacchione, D.A., Drake, D.E., Kayen, R.W., Sternberg, R.W., Kineke, G.C., Tate, G.B., 1995. Measurements in the bottom boundary layer on the Amazon subaqueous delta. Marine Geology 125, 235–257.

Castaing, P., Allen, G.P., 1981. Mechanisms controlling seaward escape of suspended sediment from the Gironde: A macrotidal estuary in France. Marine Geology 40, 101–118.

Chao, Shenn-Yu, 1988. River-forced estuarine plumes. Journal of Physical Oceanography 18, 72–88.

Chapman, D.C., Lentz, S.J., 1994. Trapping of a coastal density front by the bottom boundary layer. Journal of Physical Oceanography 24, 1464–1479.

Curtin, T.B., Legeckis, R.V., 1986. Physical observations in the plume region of the Amazon River during peak discharge—I. surface variability. Continental Shelf Research 6 (1–2), 31–51.

de Boer, D.H., Stone, M., 1999. Fractal dimensions of suspended solids in streams: comparison of sampling and analysis techniques. Hydrological Processes 13, 239–254.

Dinnell, S.P., Wiseman Jr., W.J., 1986. Fresh water on the Louisiana and Texas shelf. Continental Shelf Research 6, 765–784.

Driscoll, N.W., Karner, G.D., 1999. Three-dimensional quantitative modeling of clinoform development. Marine Geology 154, 383–398.

Droppo, I.G., 2001. Rethinking what constitutes suspended sediment. Hydrological Processes 15, 1551–1564.

Droppo, I.G., Ongley, E.D., 1994. Flocculation of suspended sediment in rivers of southeastern canada. Water Research 28, 1799–1809.

Dyer, K.R., 1995. Sediment transport processes in estuaries. In: Perillo, G.M.E. (Ed.), Geomorphology and Sedimentology of Estuaries. Developments in Sedimentology, Vol. 53. Elsevier, Amsterdam, pp. 423–449.

Dyer, K.R., Manning, A.J., 1999. Observation of the size, settling velocity and effective density of flocs, and their fractal dimensions. Journal of Sea Research 41, 87–95.

Dyer, K.R., Cornelisse, J., Dearnaley, M.P., Fennessy, M.J., Jones, S.E., Kappenberg, J., McCave, I.N., Pejrup, M., Puls, W., Van Leussen, W., Wolfstein, K., 1996. A comparison of in situ techniques for estuarine floc settling velocity measurements. Journal of Sea Research 36, 15–29.

Eisma, D., 1986. Flocculation and de-flocculation of suspended matter in estuaries. Netherlands Journal of Sea Research 20, 183–199.

Eisma, D., Kalf, J., 1984. Dispersal of Zaire river suspended matter in the estuary and the Angola Basin. Netherlands Journal of Sea Research 17, 385–411.

Eisma, D., Bernard, P., Cadee, G.C., Ittekkot, V., Kalf, J., Laane, R., Martin, J.M., Mook, W.G., van Put, A., Schuhmacher, T., 1991. Suspended-matter particle size in some West-European estuaries; Part II: A review of floc formation and break-up. Netherlands Journal of Sea Research 28, 215–220.

Festa, J.F., Hansen, D.V., 1978. Turbidity maxima in partially mixed estuaries: a two-dimensional numerical model. Estuarine and Coastal Marine Science 7, 347–359.

Foster, G., Carter, L., 1997. Mud sedimentation on the continental shelf at an accretionary margin—Poverty Bay, New Zealand. New Zealand Journal of Geology and Geophysics 40, 157–173.

Garvine, R.W., 1974. Physical features of the Connecticut River outflow during high discharge. Journal of Geophysical Research 79, 831–846.

Geyer, W.R., 1993. The importance of suppression of turbulence by stratification on the estuarine turbidity maximum. Estuaries 16, 113–125.

Geyer, W.R., Farmer, D.M., 1989. Tide induced variation of the dynamics of a salt wedge estuary. Journal of Physical Oceanography 28, 1060–1072.

Geyer, W.R., Kineke, G.C., 1995. Observations of currents and water properties in the Amazon frontal zone. Journal of Geophysical Research 100, 2321–2339.

Geyer, W.R., Hill, P.S., Milligan, T.G., Traykovski, P., 2000. The structure of the Eel River Plume during floods. Continental Shelf Research 20 (16), 2067–2093.

Geyer, W.R., Woodruff, J.D., Traykovski, P., 2001. Sediment trapping and transport in the Hudson River. Estuaries 24, 670–679.

Gibbs, R.J., 1970. Circulation in the Amazon River estuary and adjacent Atlantic Ocean. Journal of Marine Research 28, 113–123.

Gibbs, R.J., Konwar, L., 1986. Coagulation and settling of Amazon River sediment. Continental Shelf Research 6, 127–149.

Goodbred, S.L., Kuehl, S.A., 1999. Holocene and modern sediment budgets for the Ganges–Brahmaputra river system: evidence for highstand dispersal to flood-plain, shelf and deep-sea depocenters. Geology 27, 559–562.

Grabemann, I., Krause, G., 1989. Transport processes of suspended matter derived from time series in a tidal estuary. Journal of Geophysical Reserch 94, 14,373–14,380.

Grabemann, I., Uncles, R.J., Krause, G., Stephens, J.A., 1997. Behaviour of turbidity maxima in the Tamar (UK) and Weser (R.R.G.) Estuaries. Estuarine, Coastal and Shelf Science 45, 235–246.

Harris, C.K., Wiberg, P.L., 2001. A two-dimensional, time-dependent model of suspended sediment transport and bed reworking for continental shelves. Computers and Geosciences 27, 675–690.

Hill, P.S., 1998. Controls on floc size in the sea. Oceanography 11, 13–18.

Hill, P.S., McCave, I.N., 2001. Suspended particle transport in benthic boundary layers. In: Boudreau, B.P., Jorgensen, B.B. (Eds.), The Benthic Boundary Layer: Transport Processes and Biogeochemistry. Oxford University Press, Oxford, pp. 78–103.

Hill, P.S., Nowell, A.R.M., 1995. Comparison of two models of aggregation in continental-shelf bottom boundary layers. Journal of Geophysical Research 100, 22749–22763.

Hill, P.S., Syvitski, J.P., Cowan, E.A., Powell, R.D., 1998. In situ observations of floc settling velocities in Glacier Bay, Alaska. Marine Geology 145, 85–94.

Hill, P.S., Milligan, T.G., Geyer, W.R., 2000. Controls on effective settling velocity in the Eel River flood plume. Continental Shelf Research 20 (16), 2095–2111.

Hill, P.S., Voulgaris, G., Trowbridge, J.H., 2001. Controls on floc size in a continental shelf bottom boundary layer. Journal of Geophysical Research, 106, 9,543–9,549.

Hunt, J.R., 1986. Particle aggregate breakup by fluid shear. In: Mehta, A.J. (Ed.), Estuarine Cohesive Sediment Dynamics. Springer, Berlin, pp. 85–109.

Jackson, G.A., Maffione, R., Costello, D.K., Alldredge, A.L., Logan, B.E., Dam, H.G., 1997. Particle size spectra between 1 um and 1 cm at Monterey Bay determined using multiple instruments. Deep-Sea Research 44, 1739–1767.

Jaeger, J.M., Nittrouer, C.A., 1995. Tidal controls on the formation of fine-scale sedimentary strata near the Amazon river Mouth. Marine Geology 125, 259–281.

Kineke, G.C., Sternberg, R.W., 1995. Distribution of fluid muds on the Amazon continental shelf. Marine Geology 125, 193–233.

Kineke, G.C., Sternberg, R.W., Trowbridge, J.H., Geyer, W.R., 1996. Fluid mud processes on the amazon continental shelf. Continental Shelf Research 16, 667–696.

Kineke, G., Woolfe, K.J., Kuehl, S.A., Milliman, J., Dellapena, T., Purdon, R.G., 2000. Sediment export from the Sepik River, Papua New Guinea: evidence for a divergent dispersal system. Continental Shelf Research 20, 2239–2266.

Kranck, K., 1973. Flocculation of suspended sediment in the sea. Nature 246, 348–350.

Kranck, K., 1980. Experiments on the significance of flocculation in the settling of fine-rained sediment in still water. Canadian Journal of Earth Sciences 17, 1517–1526.

Kranck, K., Milligan, T.G., 1992. Characteristics of suspended particles at an 11-hour anchor station in San Francisco Bay, California. Journal of Geophysical Research 97, 11,373–11,382.

Kranck, K., Petticrew, E., Milligan, T.G., Droppo, I.G., 1992. In situ particle size distributions resulting from flocculation of suspended sediment. In: Mehta, A.J. (Ed.), Nearshore and Estuarine Cohesive Sediment Transport, Coastal and Estuarine Studies Series, Vol. 42. American Geophysical Union, Washington, DC, pp. 60–75.

Krone, R.B., 1962. Flume Studies of the Transport of Sediment in Estuarial Shoaling Processes. Hydraulic Engineering and Sanitary Engineering Research Laboratory, University of California, Berkeley, pp. 1–110.

Kuehl, S.A., Nittrouer, C.A., Allison, M.A., Faria, L.E.C., Dukat, D.A., Jaeger, J.M., Pacioni, T.D., Figueiredo, A.G., Underkoffler, E.C., 1996. Sediment deposition accumulation, and seabed dynamics in an energetic fine-grained coastal environment. Continental Shelf Research 16, 787–815.

Lambert, A., Hsue, K.J., 1979. Non-annual cycles of varve-like sedimentation in Walensee, Switzerland. Sedimentology 26, 453–461.

McCave, I.N., 1984. Size spectra and aggregation of suspended particles in the deep ocean. Deep-Sea Research 31, 329–352.

Meade, R.H., 1972. Transport and deposition of sediments in estuaries. The Geological Society of America 133, 91–120.

Meade, R.H., Parker, R.S., 1985. Sediment in rivers of the united states. US Geological Survey Water-Supply Paper 2275, 49–60.

Middleton, G.V., 1993. Sediment deposition from turbidity currents. Annual Review of Earth and Planetary Sciences 21, 89–114.

Migniot, C., 1971. L'evolution de la Gironde au cours des temps. Bulletin of the Institute of Geology Bassin Aquitaine 221–279.

Milligan, T.G., 1995. An examination of the settling behaviour of a flocculated suspension. Netherlands Journal of Sea Research 33, 163–171.

Milligan, T.G., Hill, P.S., 1998. A laboratory assessment of the relative importance of turbulence, particle composition, and concentration in limiting maximal floc size. Journal of Sea Research 39, 227–241.

Milliman, J.D., Syvitski, J.P.M., 1992. Geomorphic/tectonic control of sediment discharge to the ocean: the importance of small mountainous rivers. Journal of Geology 100, 525–544.

Muller-Karger, F.E., Richardson, P.L., McGillicudy, D., 1995. On the offshore dispersal of the Amazon's Plume in the North Atlantic: comments of the paper by A. Longhurst, "Seasonal cooling and blooming in tropical oceans. Deep-Sea Res 42, 2127–2137.

Mulder, T., Syvitski, J.P.M., 1996. Climatic and morphologic relationships of rivers: implications of sea-level fluctuations on river loads. Journal of Geology 104, 509–523.

Mulder, T., Savoye, B., Syvitsky, J., Parize, O., 1997. Hyperpycnal turbidity currents at the head of the Var Canyon? Hydrological data and geological observations. Oceanologica Acta 20, 607–626.

Munchow, A., Garvine, R.W., 1993. Dynamical properties of a buoyancy-driven coastal current. Journal of Geophysical Research 98, 20,063–20,078.

Murray, S.P., 1997. An Observational Study of the Mississippi–Atchafalaya Coastal Plume: Final Report. OCS Study MMS 98-0040. US Department of the Interior, Minerals Management Service, Gulf of Mexico OCS Region, New Orleans, La, 513pp.

Nash, D.B., 1994. Effective sediment-transporting discharge from magnitude–frequency analysis. Journal of Geology 102, 79–95.

Nichols, A.P., Walling, D.E., 1996. The significance of particle aggregation in the overbank deposition of suspended sediment on river floodplains. Journal of Hydrology 186, 275–293.

Normark, W.R., Piper, D.J.W., Hiscott, R.N., 1998. RN Sea level controls on the textural characteristics and depositional architecture of the Hueneme and associated submarine fan systems, Santa Monica Basin, California. Sedimentology 45, 53-70, 1998.

O'Melia, C.R., Tiller, C.L., 1993. Physicochemical aggregation and deposition in aquatic environments. In: Buffle, J., van Leeuwen, H.P. (Eds.), Environmental Particles, Vol. 2. Lewis Publishers, Boca Raton, FL, pp. 353–386.

Ongley, E.D., Bynoe, M.C., Percival, J.B., 1981. Physical and geochemical characteristics of suspended solids, Wilton Creek, Ontario. Canadian Journal of Earth Sciences 18, 1365–1379.

Orton, G.J., Reading, H.G., 1993. Variability of deltaic processes in terms of sediment supply, with particular emphasis on grain size. Sedimentology 40, 475–512.

Phillips, J.M., Walling, D.E., 1999. The particle size characteristics of fine-grained channel deposits in the River Exe Basin, Devon, UK. Hydrological Processes 13, 1–19.

Postma, H., 1967. Sediment transport and sedimentation in the estuarine environment. In: Lauff, G.H. (Ed.), Estuaries, American Association for the Advancement of Science Special Publication, Vol. 83, 158–179.

Ritter, D.F., 1978. Process Geomorphology. Wm C. Brown Co., Dubuque, IA 602pp.

Salisbury, J.E., Campbell, J.W., Meeker, L.D., Vorosmarty, C., 2001. Ocean color and river data reveal fluvial influence in coastal waters. EOS, Transactions, American Geophysical Union 82, 221.

Saunders, I., Young, A., 1983. Rates of surface processes on slopes, slope retreat, and denudation. Earth Surface Processes and Landforms 8, 473–501.

Schubel, J.R., 1972. Distribution and transportation of suspended sediment in Upper Chesapeake Bay. The Geological Society of America 133, 151–167.

Slattery, M.C., Burt, T.P., 1997. Particle size characteristics of suspended sediment in hillslope runoff and stream flow. Earth Surface Processes and Landforms 22, 705–719.

Steckler, M.S., Mountain, G.S., Miller, K.G., Christie-Blick, N., 1999. Reconstruction of Tertiary progradation and clinoform development on the New Jersey passive margin by 2-D backstripping. Marine Geology 154, 399–420.

Sternberg, R.W., Berhane, I., Ogston, A.S., 1999. Measurement of size and settling velocity of suspended aggregates on the Northern California continental shelf. Marine Geology 154, 43–54.

Stone, P.M., Walling, D.E., 1997. Particle size selectivity considerations in suspended sediment budget investigations. Water, Air, and Soil Pollution 99, 63–70.

Stumpf, R.P., 1988. Sediment transport in Chesapeake Bay during floods: analysis using satellite and surface observations. Journal of Coastal Research 4 (1), 1–15.

Syvitski, J.P.M., Asprey, K.W., Clattenburg, D.A., Hodge, G.D., 1985. The prodelta environment of a fjord: suspended particle dynamics. Sedimentology 32, 83–107.

Syvitski, J.P.M., Smith, J.N., Calabrese, E.A., Boudreau, B.P., 1988. Basin sedimentation and the growth of prograding deltas. Journal of Geophysical Research 93, 6895–6908.

Syvitski, J.P.M., Asprey, K.W., Le Blanc, K.W.G., 1995. In situ characteristics of particles settling within a deep-water estuary. Deep-Sea Research II 42, 223–256.

ten Brinke, W.B.M., 1994. Settling velocities of mud aggregates in the Oosterschelde tidal basin (the Netherlands), determined by a submersible video system. Estuarine Coastal and Shelf Science 39, 549–564.

Traykovski, P., Geyer, W.R., Irish, J.D., Lynch, J.F., 2000. The role of density driven fluid mud flows for cross-shelf transport on the Eel River continental shelf. Continental Shelf Research 20, 2113–2140.

Uncles, R.J., Stephens, J.A., 1989. Distributions of suspended sediment at high water in a macrotidal estuary. Journal of Geophysical Research 94, 14,395–14,406.

van Leussen, W., 1988. Aggregation of particles, settling velocity of mud flocs: a review. In: Dronkers, J., van Leussen, W. (Eds.), Physical Processes in Estuaries. Springer, New York, pp. 347–403.

van Leussen, W., 1999. The variability of settling velocities of suspended fine-grained sediment in the Ems estuary. Journal of Sea Research 41, 109–118.

Walker, N.D., Hammack, A.B., 2000. Impacts of winter storms on circulation and sediment transport: Atchafalaya–Verilion Bay Region, Louisiana, USA. Journal of Coastal Research 16, 996–1010.

Walling, D.E., Moorehead, P.W., 1989. The particle size characteristics of fluvial sediment: an overview. Hydrobiologia 176/177, 125–149.

Walling, D.E., Owens, P.N., Waterfall, B.D., Leeks, G.J.L., Wass, P.D., 2000. The particle size characteristics of fluvial suspended sediment in the Humber and Tweed catchments, UK. Science of the Total Environment 251, 205–222.

Warren, L.A., Zimmermann, A.P., 1994. Suspended particulate grain-size dynamics and their implications for trace-metal sorption in the Don River. Aquatic Sciences 56, 348–362.

Whitehouse, U.G., Jeffrey, L.M., DeBrecht, J.D., 1960. Differential settling tendencies of clay minerals in saline waters. In: Swineford, A., (Ed.), Clays and Clay Minerals, Proceedings of the Seventh National Conference, pp. 1–79.

Wiberg, P.L., Smith, J.D., 1987. Calculations of the critical shear stress for motion of uniform heterogeneous sediments. Water Resources Research 23, 1471–1480.

Wiseman, W.J., Yang, Z.-S., Bornhold, B.D., Keller, G.H., Prior, D.B., Wright, L.D., 1986. Suspended sediment advection by tidal currents off the Huanghe (Yellow River) delta. Geo-Marine Letters 6, 107–113.

Wohl, E., 2000. *Mountain Rivers*. American Geophysical Union, Washington, DC, 320 pp.

Wolanski, E., Gibbs, R.J., 1994. Flocculation of suspended sediment in the Fly River Estuary, Papua, New Guinea. Journal of Coastal Research 10, 105–109.

Wolman, M.G., Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology 68, 54–74.

Woodruff, J.D., Geyer, W.R., Sommerfield, C.K., Driscoll, N.W., 2001. Seasonal variation of sediment deposition in the Hudson estuary. Marine Geology 179, 105–119.

Wright, L.D., 1971. Hydrography of South Pass, Mississippi River. J. Waterways, Harbors and Coastal Eng. Div., Proc. Am. Soc. Civil Engineers, Vol. 97, WW3, pp. 491–504.

Wright, L.D., Coleman, J.M., 1971. Effluent expansion and interfacial mixing in the presence of a salt wedge, Mississippi River Delta. Journal of Geophysical Reserach 76, 8649–8661.

Wright, L.D., Coleman, J.M., 1974. Mississippi River Mouth processes: effluent dynamics and morphologic development. Journal of Geology 82, 751–778.

Wright, L.D., Nittrouer, C.A., 1995. Dispersal of river sediments in coastal seas: six contrasting cases. Estuaries 18, 494–508.

Wright, L.D., Yang, Z.-S., Bornhold, B.D., Keller, G.H., Prior, D.B., Wiseman Jr., W.J., 1986. Hyperpycnal plumes and plume fronts over the Huanghe (Yellow River) delta front. Geo-Marine Letters 6, 97–105.

Wright, L.D., Wiseman Jr., W.J., Yang, Z.-S., Bornhold, B.D., Keller, G.H., Prior, D.B., Suhayda, J.N., 1990. Processes of marine dispersal and deposition of suspended silts off the modern mouth of the Huanghe (Yellow River). Continental Shelf Research 10, 1–40.

Yankovsky, A.E., Chapman, D.C., 1997. A simple theory for the fate of buoyant coastal discharges. Journal of Physical Oceanography 27, 1386–1401.