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Particle trapping in estuarine tidal flows

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Abstract. Particle trapping in estuarine turbidity maxima (ETM) is caused primarily by convergent mean and/or tidal fluxes of sediment. The result is an approximately bell-shaped along-channel distribution of vertically integrated, tidal cycle mean suspended sediment concentration. Observations from the Columbia River estuary suggest that (1) strong two-layer or internal along-channel residual and overtide flows are generated by time-varying stratification and (2) correlations between the near-bed velocity and the suspended sediment fields at these frequencies are important in landward transport of sediment. A new spatially and temporally integrated form of the sediment conservation equation has been derived to analyze this trapping process. Time changes in tidally averaged sediment concentration between two estuarine cross sections can be shown to be related to the divergence of the seaward, river flow transport; the divergence of velocity shear-sediment stratification correlations for the mean flow and each tidal constituent; and net erosion or deposition at the bed. Vertically integrated variables other than seaward river transport are absent from this integrated balance. Analysis of sediment fluxes using this balance supports the idea that internal residual and overtide circulations are primarily responsible for the landward sediment transport on the seaward side of ETM found near the upstream limits of salinity intrusion. The balance also shows that attempts to represent fluxes causing trapping of sediment in an ETM as a product of a time-mean, vertically integrated, along-channel gradient and a diffusivity inevitably lead to the appearance of countergradient transport and thus a negative diffusivity on the seaward side of the ETM. This result occurs because the trapping process is inherently nonlinear and at least two-dimensional and because a one-dimensional representation is physically unrealistic.

Introduction

Trapping of particles by convergent suspended sediment transport often creates pronounced estuarine turbidity maxima (ETM) in river estuaries and narrow tidal channels (Figure 1). This concentration of particles commonly occurs near the upstream limits of salinity intrusion and has been attributed to density-driven residual circulation [Nichols and Poor, 1967; Festa and Hansen, 1978; Dyer, 1988]. Trapping may, however, occur under a variety of circumstances, and a number of internal and barotropic causal mechanisms are catalogued below. The diversity of advective mechanisms that can bring about localized concentration of sediment in ETM indicates a need for a unified methodology with which to analyze sediment fluxes and their influence on ETM sediment concentration. The primary purpose of this paper is, accordingly, to present an integral sediment conservation expression for channelized tidal flows that defines changes in ETM suspended sediment concentration in terms of bed erosion/deposition and fluxes of sediment through boundaries at two along-channel positions. This spatially and temporally integrated

sediment balance includes both horizontal (advection and spatially variable resuspension) and vertical (settling, erosion, and deposition) aspects of ETM suspended sediment behavior for each sediment size class. Nonconservative behavior such as aggregation and biological consumption or production could readily be added; therefore a modified form of this balance might also be used to analyze interactions of the suspended sediment field with biological processes in a realistic advective context.

This sediment balance and observations taken in the Columbia River estuary are then used to provide a new explanation of processes leading to particle trapping in an ETM associated with the upstream limits of salinity intrusion. Under this interpretation, ETM formation is caused by convergent shear fluxes in residual and overtide flows that occur at this location. The relevant internal residual and overtide currents are primarily caused by time-varying stratification through a process known as "internal tidal asymmetry" [Jay and Musiak, 1994], though gravitational circulation also contributes at zero frequency. Finally, we explore modeling implications of the fact that estuarine suspended sediment transport is inevitably at least a two-dimensional problem; that is, both horizontal advection and vertical processes (erosion, deposition, settling, and vertical mixing) must be considered. Neither one-dimensional formulations in the vertical, essentially Rouse balances modified to include time variation and stratification nor vertically integrated models that represent tidal transports of sediment as horizontal diffusion can be expected to yield an adequate representation of particle trapping in tidal channels. The former cannot represent advection at all, while the latter can do so only through artificial, countergradient transport. Thus estuaries pose a very different modeling problem from either continental shelves (at least those with smooth topography) or rivers. In the former, horizontal uniformity may normally be assumed, and wind-wave mean flow interactions affect vertical mass and momentum transfer but do not alter the scalar field by horizontal advection [e.g., Kachel and Smith, 1989; Glenn and Grant, 1987]. In the latter, tidal wave fluxes, salinity stratification, and horizontal circulation modes driven by density effects are all absent. Only stratification by the suspended sediment field itself need be considered, and fluvial flows may normally be treated as time invariant.

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Figure 1. Schematic of (top to bottom) mean, ebb, and flood scalar fields for two-dimensional (in the along-channel and vertical directions) estuarine turbidity maxima (ETM). Contours show the density field, and shading shows sediment concentration; the ocean is to the left, and the river is to the right. Tidal straining of the density field causes stratification to be greater on ebb than flood. The sediment field is affected by both horizontal advection and vertical processes (erosion, deposition, settling, and vertical mixing).

ETM Generation Mechanisms

The internal tidal asymmetry particle-trapping mechanism discussed below for the Columbia River estuary is found in many river estuaries [Jay and Musiak, 1994]. Nonetheless, this particular transport process is not the only means by which convergent horizontal sediment fluxes may create an ETM. Several other advective ETM generation mechanisms have been proposed, and these cover almost the entire spectrum of estuarine circulation processes. Relevant barotropic processes include tidal residual trapping over a topographic high (theoretical calculations [Ianniello, 1979]), tidally varying fluxes from tidal flats (in the Tay estuary, Scotland [Weir and McManus, 1987]), and barotropic tidal wave steepening (overtide trapping in the Gironde estuary, France [Allen et al., 1980]). Steady, two-layer gravitational circulation has long been known to cause particle trapping [e.g., Nichols and Poor, 1967]. Hamblin [1989] and Geyer [1993] have emphasized an effect of stratification different from internal tidal asymmetry. Inhibition of vertical mixing by stratification decreases the upward flux associated with vertical mixing relative to the downward (settling) flux of sediment in a way that may be correlated with horizontal advection. Several mechanisms may, moreover, be active in causing a single ETM. For example, during low-flow conditions in North San Francisco

Bay, an ETM trapped just seaward of a topographic high [Meade, 1972] appears to be maintained by a combination of tidal residual and gravitational circulation interacting with topography. Finally, multiple ETM may occur in one system. Three simultaneous ETM have been observed in the Columbia River estuary. The two landward ETM are located near the upstream limits of salinity intrusion in the two main channels and are primarily maintained by internal tidal asymmetry. The more seaward of these ETM may be associated with topographic convergence, but this hypothesis has not been tested.

Other factors in addition to convergent suspended sediment transport may also be important in ETM formation. Trapping in some systems is significantly enhanced by particle aggregation (e.g., Rotterdam Waterway and the Savannah River Estuary [van Leussen, 1988; Krone, 1972]); its occurrence on tidal time scales in other systems remains in doubt (e.g., San Francisco Bay; [Kineke and Sternberg, 1989; Kranck and Milligan, 1992]). Postma [1961], Dyer [1988], and Dyer and Evans [1989] have stressed the role of nonlinearities in deposition and erosion in ETM creation. Nonetheless, numerical models [e.g., Festa and Hansen, 1978] show that particle interaction and bed nonlinearities are not necessary to particle trapping and that the locus of particle concentration (e.g., upstream limits of salinity intrusion or a topographic feature) is determined primarily by advective processes. Fluvial sediment usually makes up the bulk of material in an ETM, but the trapping process is relatively independent of source. Fluxes of material from the river, ocean, or sources in lateral embayment may all play a role [Festa and Hansen, 1978; Weir and McManus, 1987]. The primary importance of the source lies, however, in the quality of the material supplied; it must have a settling velocity matched to the advection processes in the system. If the material supplied is too fine, it will move through the estuary as vertically uniform wash load. If it is too coarse, it will move only as bed load. In neither case can an ETM form.

Formulations of ETM Sediment Transport

Two factors, the variety of processes causing particle trapping and the inherently Lagrangian nature of particle transport, suggest the use of a wave flux approach in the analysis of ETM. This methodology relates wave transport of wave period variations in a scalar property (sediment concentration in this case) to Stokes drift transport of the mean field [Andrews and McIntyre, 1976, 1978]. In the atmospheric literature it has been used to derive a variety of "nonacceleration" theorems for steady wave fields [Dunkerton, 1980]. In atmospheric nonaccelerated flows in which mixing of the scalar property may be neglected, transport by wave fluxes (i.e., by the Stokes drift) is exactly compensated for by residual flow transport, so that total Lagrangian transport is zero and the scalar field is steady. With modifications to account for vertical mixing and settling, an ETM may be treated in a similar manner.

Most existing models of sediment transport are not based on wave flux concepts and, because of the difficulties inherent in a full three-dimensional calculation of sediment transport, are either one-dimensional in the vertical or vertically integrated (two-dimensional in the horizontal). Wave flux ideas are useful for the light they shed on these conventional models. One-dimensional vertical profiles (Rouse models and the like) traditionally used in analyses of the suspended sediment field have serious defects for representation of the estuarine suspended sediment distribution. They can explain only one aspect of the

influence of tidally varying stratification, namely, its reduction of vertical suspended sediment mixing relative to settling. Neither the vital, sometimes dominant influence of horizontal mean and wave advection on the vertical distribution of suspended sediment nor the modifying effect of stratification on this along-channel transport can be considered in such a model.

The analysis that follows will show that vertically integrated models (in one or two horizontal dimensions) typically require a negative horizontal diffusivity to represent an estuarine ETM. The reason for this apparently countergradient transport is as follows. A curve representing vertically averaged sediment concentration in an ETM is approximately bell shaped in the along-channel direction. In the absence of a local source of sediment, this along-channel distribution can be created only by convergent mean and/or wave fluxes. These mean and wave fluxes are typically parameterized in vertically integrated models by a horizontal diffusivity times a horizontal gradient of a vertically averaged sediment concentration. This representation may be called a vertically integrated horizontal diffusion (VIHD) model. As shown in the analysis that follows, ETM formation requires that wave and/or mean fluxes of suspended sediment be upgradient on the seaward side of the ETM, implying a negative horizontal diffusivity in a VIHD model. For the same location, horizontal diffusivities calculated from the monotonic salinity distribution would be uniformly positive. The very different behaviors of salt and sediment stem from the fact that scalar transport in stratified tidal channel flows is inherently at least two-dimensional in the along-channel (x) and vertical (z) directions; that is, maintenance of a scalar balance against fluvial export requires correlations between shear and stratification in the various mean and wave modes. Thus there are fundamental differences between salt and suspended sediment transport that cannot easily be represented in VIHD models, whether in one or two horizontal dimensions. VIHD treatments of suspended sediment transport may give unphysical results, because these treatments incorrectly represent the processes that maintain the suspended sediment balance. It will be shown below, moreover, that even the apparent success of one-dimensional VIHD representations of the salt balance is largely fortuitous and not a result of dynamical similitude. Salt and suspended sediment are by no means exceptional in the difficulties they present for one-dimensional models. A dependence on the details of correlations between shear and stratification is a general property of scalar transport in channel flows. Thus correct representation of the shear fluxes that bring about this transport is vital, and two scalars will have the same estuarine transport properties only if they have similar spatial distributions at all relevant wave frequencies.

Representation of Scalar Wave Fluxes

One possible resolution to the difficulties with the VIHD models discussed in previous paragraphs would be not to integrate vertically. Net vector transport by wave fluxes of a nondiffusive scalar at a single point in the interior of the atmosphere or ocean may be related to a Stokes transport of the mean scalar field. This flux may be parameterized by a product of a local mean gradient times a skew tensor diffusivity that is orthogonal to the mean gradient [Middleton and Loder, 1989]. Such a representation is inappropriate in estuaries because of the dominant importance of vertical mixing and, for particles, settling.

Estuarine transport studies require a different approach. Vertical integration is needed to allow a tractable overall balance,

but effects of velocity shear and scalar stratification must be correctly represented. An earlier wave flux analysis of estuarine salt transport [Jay, 1991b] is here generalized. The previous analysis considered only dissolved scalars that, like salt, have a monotonic distribution in the along-channel direction. The present analysis expresses the change in tidal cycle mean sediment concentration between two cross sections (chosen to encompass the ETM) as a sum of three terms. They are the seaward transport by river flow, the along-channel divergence of shear stratification fluxes at mean and wave (tidal) frequencies, and fluxes to and from the bed. The sediment may originate in the ocean, the river, or both. As in the previous treatment, vertically integrated quantities, aside from the seaward river flow transport, are absent in the final result. All wave fluxes are expressed as correlations of shear and sediment stratification in each wave mode.

The theoretical development that follows employs ideas from two wave flux formalisms. The first of these is restricted to small-amplitude waves [Middleton and Loder, 1989] and is sometimes referred to as an "Eulerian" or "weakly nonlinear" theory. Some of the theorems utilized below are, however, more readily proven using a generalized Lagrangian mean (GLM) theory of wave mean flow interaction that covers even unsteady, finite-amplitude waves [Andrews and McIntyre, 1978]. The results employed herein are special cases for steady small-amplitude waves [Dunkerton, 1981]. Unlike most treatments in the atmospheric literature, vertical integration allows inclusion of vertical turbulent fluxes so long as these fluxes are specified at the sea bed and free surface. It is also necessary that the estuary be narrow: less than the smaller of a tidal excursion and the internal Rossby radius. In a wide system with complex topography, certain assumptions concerning the Stokes drift made in both wave flux methodologies employed here are not valid, because strong horizontal gradients in tidal velocity can cause a series representation of the Stokes drift to be nonconvergent [Zimmerman, 1980]; this problem is absent in narrow systems.

Correlations between stratification and shear at mean and wave frequencies replace in the present model the horizontal diffusivities employed in traditional VIHD models. These correlations are substance or size-class specific, so that, for example, a vertically integrated box model is not a property of the flow but of a particular scalar in a flow. This leads to the sensible result that different size classes of sediment have different residence times in an ETM. Similarly, dissolved substances originating in fresh water will have different flux properties and estuarine residence times than those originating in oceanic water masses. These statements are both corollaries of an important general principle: formulation of a meaningful vertically integrated model of scalar transport in a stratified channel flow requires information concerning the vertical distribution of the scalar to be modeled.

Observations of Internal Tidal Asymmetry and Wave Fluxes in the Columbia River Estuary

A cross-sectionally integrated conservation equation for individual size classes of suspended sediment is derived in the following section. This balance shows that net transport of a scalar quantity through a cross section is related to correlations of velocity shear and the scalar stratification in the mean flow and in each of the various wave modes. A frequency must be prominent

in both the velocity and scalar fields if it is to contribute to the scalar transport balance. Conversely, if it is prominent in both, it will contribute to the scalar balance unless stratification and shear are exactly 90° out of phase. Interactions between neighboring frequencies within a tidal species can cause neap-spring variations in residual suspended sediment concentration but cannot contribute at zero frequency. The frequencies that are actually present in the velocity and sediment fields are thus a key issue in understanding suspended sediment transport in tidal channel flows.

Numerous calculations of cross-sectional averaged balances of salinity, sediment and other substances have been carried out in estuaries, as reviewed by Dyer [1973], Fischer [1972] and Jay [1991b]. Several studies have found wave fluxes to play a major role in cross-sectional scalar balances [e.g., Hansen, 1965; Hughes and Rattray, 1980; Dyer, 1978; Lewis and Lewis, 1983; Winterwerp, 1983; Uncles et al., 1985; Jay and Smith, 1990]. These fluxes were invariably assumed to be at the dominant tidal frequency. No available estuarine scalar transport calculation from data has separated all the relevant tidal species, let alone the tidal constituents thereof. Robe's [1968] study was exceptional in having resolved two tidal constituents of the salinity field, the semidiurnal M_2 and the diurnal K_1 . The complexity of traditional transport expansions has precluded inclusion of additional species and constituents, including over tides. Previous research has not, therefore, resolved the frequency structure of the suspended sediment field in even a preliminary way.

Definitions and Mechanisms

Observations in the Columbia River estuary suggest that residual and overtide variations in sediment concentration should be large in response to maxima in shear stress on both flood and ebb (Figures 2, 3a, and 3b). This situation is probably typical of estuaries in which the tidal flow is the dominant cause of bed stress. It has long been known that the zero-frequency part of the suspended sediment field combines with two-layer mean gravitational circulation to contribute to ETM formation [Dyer,

1978; Festa and Hansen, 1978]. The relevant questions are (1) whether there is a mechanism to create substantial overtide currents within one salinity intrusion length of the mouth of river estuaries and (2) whether gravitational circulation is the primary or only internal mechanism active in the residual flow frequency band. Barotropic mechanisms can generate substantial overtide currents in the more landward parts of river estuaries, but our observations suggest that a two-layer (internal) mechanism related to salinity intrusion is likely to be a much larger factor in the lower reaches of such systems and thus to be important in ETM formation. This "internal tidal asymmetry" generates both residual and overtide currents and is caused by an interaction between tidal straining of the density field (which drives large changes in vertical turbulent momentum flux) and tidal variations in vertical shear [Jay, 1991a; Jay and Musiak, 1994]. Internal asymmetry thus contributes to ETM maintenance through both the zero-frequency and the overtide parts of the sediment balance.

The mechanism of internal tidal asymmetry can be explained as follows. Consider circulation processes associated with a single tidal frequency (as measured from the surface elevation ζ and sectionally integrated transport Q) in a hypothetical narrow estuary having a strong horizontal density gradient. Because of tidal straining of the density field (Figure 1) and the effects thereof on vertical mixing, flood currents are either bottom intensified or nearly uniform in the vertical. Ebb currents are strongly sheared, with a pronounced surface maximum (Figure 4). A Fourier decomposition of the time-varying velocity field yields, in addition to the dominant tidal M_2 component, internal Z_0 (zero-frequency), M_4 and M_8 flows of about 0.63δ , 0.42δ , and 0.08δ , respectively, where the internal asymmetry δ is defined in Figure 4 [Jay, 1991a]. Here δ and therefore all internal modes must integrate to zero over depth in our hypothetical system to satisfy continuity; only M_2 is found in ζ and Q . Thus the zero-frequency internal mode exhibits a change in sign with depth, and internal over tides show a 180° phase shift over the depth of the flow. Other internal over tides in addition to M_4 and M_8 may appear if maximum and minimum stratification do not occur exactly half a tidal cycle apart, and current meter

Figure 2. Location map showing sampling stations in the Columbia River estuary and approximate ETM positions.

Figure 3a (top left). Time series of near-bed M_2 (the largest tidal constituent), M_4 (its first overtide), and residual velocity (Z_0) and total near-bed along-channel velocity; (top right); total velocity amplitude and phase spectra; (lower left); time series of bed stress T_b and excess shear stress T_{xs} (assuming a critical shear stress of 8 dyn cm^{-2}); (lower right) amplitude and phase spectra for excess shear stress. In this artificial but typical example, M_2 is almost absent in the T_{xs} spectrum despite being the largest tidal constituent, while the Z_0 , M_4 , and M_6 components are prominent. Time on the x axis is in radians for K_1 .

Figure 3b. As in Figure 3a but based on harmonic analyses of data from a current meter 3 m above the bed near the mouth of the Columbia River estuary (station 2S in Figure 2); K_1 is also included. Again, M_2 is not a major component in the T_{xs} spectrum, despite its prominence in the current spectrum.

Figure 4. Schematic of overtide and residual current generation by a single barotropic tidal constituent in surface elevation and transport. The actual flood and ebb currents are indicated by heavy lines (U_f and U_e), and the perfectly reversing principal harmonic at tidal frequency is indicated by light lines (A). The presence of a horizontal density gradient and tidal straining of the density field combine to cause flood currents to be vertically rather uniform or bottom intensified. Suppression of vertical mixing causes ebb currents to be surface intensified, as suggested by the profiles of vertical eddy diffusivity K_{zf} and K_{ze} . The result is an ebb flood internal asymmetry (represented by δ) that integrates to zero, so that it is absent in the vertically integrated flow. A simple Fourier analysis shows that δ contributes to the motion at zero frequency and to odd overtides (M_4 , M_8 ...).

records indicate the presence of an entire internal overtide spectrum [Jay and Musiak, 1994].

Observations

The existence of the internal asymmetry mechanism was deduced initially from current meter data [Jay, 1991a], but little information concerning the cross-sectional distribution of the internal modes was available because an insufficient number of current meter records were available at that time. Acoustic Doppler current profiler (ADCP) data and conductivity-temperature-depth-optical backscatter (CTD-OBS) profiles now allow a more complete characterization of the effects of internal asymmetry and its relationship to suspended sediment advection. Figures 5 and 6 show harmonic analysis results based on time series of ADCP data collected from a moving boat on sections NC1 and NC2 (Figure 2). The M_2 current amplitude is 1.4 m s^{-1} and the top-to-bottom velocity difference in the residual flow Z_0 is about 1 m s^{-1} (Figure 5). A two-dimensional model shows that, given ambient tidal currents and density forcing, river flow and steady gravitational circulation can account for only about 40% of the observed mean shear. The remainder is associated with the internal tidal asymmetry [Jay and Musiak, 1994]. The M_4 current reaches 0.25 m s^{-1} near the surface on section NC1 and 0.4 m s^{-1} on NC2. Maximum near-bed M_4 currents are 0.10 m s^{-1} and 0.2 m s^{-1} on NC1 and NC2, respectively. These M_4 currents are an order of magnitude larger than those associated with barotropic forcing in this part of the Columbia River estuary and can be generated only by internal

asymmetry. The M_4 phase distribution shows a middepth 180° shift in the vertical on both sections (Figure 6), a definite indicator of internal motion. There is also a lateral 180° phase shift near the surface on section NC1. This may reflect the presence of a sharp bend and bifurcation of the channel about 1 km landward of the section shown. The $2M_2 - M_4$ relative phase is close to zero (within about 40° or 40 min). This means that M_4 reinforces flood currents and suggests that suspended sediment detected by both a CTD-OBS sensor and a high-frequency internal echo sounder is likely to be moved landward.

Harmonic analysis of suspended sediment concentrations inferred from CTD-OBS data (Figure 7) shows that the largest concentrations on section NC2 are at the Z_0 and the M_4 frequencies; M_2 concentrations are quite small. The greatest suspended sediment concentrations on section NC1 are at Z_0 , with M_2 and M_4 being about equal and somewhat smaller. Strong residual, upstream bottom flow clearly brings about substantial landward transport on both sections. Furthermore, the M_4 phase difference between sediment and current (Figures 6 and 7) is about 20° – 80° (20–80 min) over the northern half of the channel, where the largest overtide sediment concentrations were observed on both NC1 and NC2. The M_2 phase difference between sediment and current (Figures 6 and 7) is also small and favors landward transport on both sections. M_2 concentrations are sufficiently small on section NC2 that M_2 probably does not play a major role in landward sediment transport, whereas it may actually be the largest contributor to such transport on section NC1. Thus the ADCP and OBS results together strongly suggest that the residual and M_4 internal modes bring about landward suspended-sediment transport; M_2 may also play an important role in some cases.

Further information concerning suspended sediment transport by internal overtides can be obtained from two near-bed Aanderaa current meters equipped with transmissometers and suspended from a bridge in the turbidity maximum zone of the Columbia River estuary in July 1980. A harmonic analysis of 1 month of transmissivity data from a meter located 3 m above the bed in 18 m of water in the north channel (station CM-5b in Figure 2) showed that M_4 was the second largest component of sediment variability after the mean flow Z_0 . Harmonic analysis of along-channel current showed that M_4 was 20% as big as M_2 and nearly as large as N_2 (the third largest semidiurnal after M_2 and S_2) and O_1 (the second largest diurnal after K_1). M_4 current and sediment concentrations were in phase with each other within about 45 min. A meter placed 3 m above the bed in 12 m of water in the south channel (station CM-5a in Figure 1) during the same period showed that MK_3 and M_4 were the largest components of suspended sediment variability after Z_0 and the tidal monthly components Mm and MSf . The M_4 along-channel current was 29% of that for M_2 , making it the largest tidal current component after M_2 , K_1 and S_2 . In this case, M_4 velocity and sediment variability were in phase within 2 min. A subsequent 5-day deployment at CM-5a in October 1980 yielded nearly identical results, except that M_4 and M_2 were, after Z_0 , the major components in the transmissivity analysis. M_2 suspended sediment variability was not as prominent in any of these records as it was at section NC1.

These current meter/transmissivity results cannot be translated directly into sediment transport for a number of reasons, including lack of detailed current meter profiles, uncertainty regarding the transmissivity-sediment concentration relationship, and possible distorting effects of the bridge on the flow. Nor can suspended sediment transport be calculated from the ADCP and

CTD-OBS data; the ADCP data suffer from two (partially offsetting) errors for the bottom 15% of the flow, where sediment concentrations are high: direct reflection from the bottom and an inability to track the large shear found there. Nonetheless, the ADCP/OBS and current meter/transmissivity data together strongly suggest that (1) internally generated residual and overtide currents cause landward suspended sediment transport and (2) this transport plays an important role in ETM generation.

Velocity and density observations confirming the presence of internal tidal asymmetry are available from a dozen estuaries including the Fraser River [Geyer, 1988], the Great Ouse [West *et al.*, 1985], the Tamar [West *et al.*, 1990], and the Hudson Estuary [Zhou, 1992]. Tidal stratification variations in another eight systems are suggestive of its presence [Jay and Musiak, 1994], and the Jay and Smith [1988] classification system further suggests that any estuary with both strong river flow and tidal forcing should show such an asymmetry.

In summary, tidal straining of the density field and consequent tidal cycle variations in stratification and vertical mixing cause pronounced internal residual and overtide flow modes to appear in the velocity field. Near the bed, these currents are correlated with mean and overtide components in sediment concentration. These internally generated flows therefore, play a strong role in ETM particle trapping and estuarine suspended sediment transport in general. The available observations are, however, too limited to assess spatial and neap-spring variability of ETM flux mechanisms. The substantial diurnal tidal signal in West Coast estuaries certainly introduces substantial semidiurnal variability in bed stress in these systems. This in turn should introduce a substantial semidiurnal sediment shear flux. It seems likely that this semidiurnal flux is absent in East Coast estuaries where diurnal tidal forcing is weak, but this will have to be confirmed from observations.

Formulation of an Integrated Sediment Conservation Expression

The purpose of this section is to define a spatially and temporally integrated conservation expression for the volumetric concentration C of a single size class of sediment that settles vertically, is advected horizontally, and may be deposited and resuspended at the bed. Because of the strong constraint imposed by two-dimensional (in x and z) channel geometry, a sediment balance analogous to earlier nonacceleration theorems discussed above may be derived without specification of a momentum equation. The absence of a momentum equation may be an advantage, because the resultant mass balance may be used with any two-dimensional momentum balance that yields steady waves. It is assumed in this two-dimensional analysis that the

Figure 5. Harmonic constants for sections (a) NC1 and (b) NC2 in the north channel of the Columbia River estuary about 18 km landward from the mouth (Figure 2): (top) Z_0 ; (middle) M_2 amplitude; (bottom) M_2 phase. These were calculated from 30 hours of acoustic Doppler current profiler observations collected by a moving boat and represent about 75 passes across the section, which is plotted looking landward. MLLW, is the mean of the lower low waters occurring each tidal day over a 19 year period.

Figure 6. Further harmonic constants for sections (a) NC1 and (b) NC2 computed as in Figure 5: (top) M_4 amplitude; (bottom) M_4 phase.

sediment velocity vector $U_C = U - \bar{W}_C$ differs from that of water U only by a vertical settling velocity $-\bar{W}_C$ that is a function of size class but is independent of time and spatial position for each size class. Further, x is positive landward from the ocean, and z is measured upward with $z=0$ at the mean free surface. Each tidal constituent is assumed to be steady and of small amplitude; that is, the ratio of amplitude ζ to mean depth H is in every case much less than 1. The sediment concentration is taken as small so that the concentration of water $1-C \approx 1$ everywhere in the interior of the fluid. For simplicity, only one main tidal constituent with one overtide will be considered; generalization to multiple size classes and diurnal and semidiurnal constituents is straightforward. Finally, river flow is assumed to enter only at the landward end of the system. The derivation is similar to that discussed in detail by Jay [1991b] for conservation of salt.

The analysis begins with the assumption that the velocity and sediment concentrations are the sum of wave cycle mean, tide, and overtide contributions:

$$U = \bar{U} + U' + U'' \quad C = \bar{C} + C' + C'' \quad (1)$$

where the overbar indicates a wave cycle average, the single prime indicates a tidal variable, and the double prime indicates an overtide variable. All wave variables have zero time mean. Zero-frequency integral conservation of water is expressed by

$$\frac{\partial}{\partial x} \int_{-h}^{\bar{\zeta}} \bar{U} dz + \frac{\partial}{\partial x} \int_{-h}^{\bar{\zeta}} \bar{U}^S dz + \frac{\partial \bar{\zeta}}{\partial t} = 0 \quad (2a)$$

where $-h$ is the bed elevation, total mean depth $H = h + \bar{\zeta}$, t is time, and the lowest-order approximation to the integral Stokes drift $\bar{Q}^S$ is made up of tidal and overtide contributions:

$$\int_{-h}^{\bar{\zeta}} \bar{U}^S dz = \overline{U' \zeta'} + \overline{U'' \zeta''} = \bar{Q}^S = \bar{Q}^{St} + \bar{Q}^{So} \quad (2b)$$

where an overbar followed by a superscript S indicates a Stokes correction (that is the difference between an Eulerian mean and a mean following a water parcel), and superscripts St and So indicate Stokes drift terms arising from the tidal and overtide

flows, respectively. Along-channel integration of (2a) from arbitrary x to infinity or the head of tide gives, for steady waves,

$$\int_{-h}^{\bar{\zeta}} \bar{U} dz = -\bar{Q}^S - \bar{Q}^R \quad (2c)$$

where $\bar{Q}^R$ is the river flow per unit width. Sediment conservation in two dimensions is given by

$$\frac{\partial C}{\partial t} = -U \frac{\partial C}{\partial x} - W \frac{\partial C}{\partial z} + \bar{W}_C \frac{\partial C}{\partial z} + \frac{\partial T_z}{\partial z} \quad (3a)$$

where T_z is the vertical turbulent sediment flux. No along-channel diffusion term is required in (3a), because horizontal turbulent diffusion is small relative to vertical diffusion and all wave (tidal) fluxes are explicitly included. Vertical and wave cycle integration of (3a) gives

$$\frac{\partial}{\partial t} \int_{-h}^{\bar{\zeta}} \bar{C} dz = -\frac{\partial}{\partial x} \int_{-h}^{\bar{\zeta}} (\bar{U} \bar{C} + \bar{U}' C' + \bar{U}'' C'') dz + (-\bar{T}_z + \bar{W}_C \bar{C})|_{-h} \quad (3b)$$

where a zero net flux boundary condition is assumed at $z = \bar{\zeta}$.

Along-channel integration then provides a preliminary conservation expression. In the case of salt conservation, the along-channel integration was from (arbitrary) x to infinity, because the ocean at $x=0$ was the source of salt and tidal energy so that both salinity and wave velocities die out for large x [Jay, 1991b]. The case of sediment is more complex, because its distribution is not monotonic and because both oceanic and fluvial sources must be considered. The appropriate integration is from x_1 to x_2 . These limits, while otherwise arbitrary, are assumed to include the ETM. Thus:

$$\frac{\partial}{\partial t} \int_{x_1}^{x_2} \int_{-h}^{\bar{\zeta}} \bar{C} dz dx = \left[\int_{-h}^{\bar{\zeta}} (\bar{U} \bar{C} + \bar{U}' C' + \bar{U}'' C'') dz \right]_{x_2}^{x_1} + \bar{V}_{fb} \quad (4a)$$

where the flux at the bed

$$\bar{V}_{fb} = \int_{x_2}^{x_1} (-\bar{T}_z + \bar{W}_C \bar{C})|_{-h} dx \quad (4b)$$

Figure 7. Suspended-sediment harmonic constants in formazine turbidity units (FTUs) calculated from optical backscatter observations for sections (a and c) NC1 and (b and d) NC2. (a and b) (top) Z_0 ; (middle) M_2 amplitude; (bottom) M_2 phase. (c and d) (top) M_4 amplitude; (bottom) M_4 phase. Formazine is a colloid used for laboratory calibration. Field calibration data show that 1 FTU is about $1.5 \text{ mg liter}^{-1}$ for the sediment in question.

must be calculated as a boundary condition using suspended-sediment transport methods.

Further development is needed to take into account the cancellation of the landward sediment transport caused by the vertical average Stokes drift velocity and the seaward flux

associated with the vertical integral of the Eulerian Stokes drift compensation flow, that part of $\bar{U}$ in (2c) that cancels the landward Stokes drift. This requires separation of all variables into vertical average ($\langle \rangle$) and vertical deviation (subscript V) quantities:

$$\bar{U} = \langle \bar{U} \rangle + \bar{U}_V \quad \bar{C} = \langle \bar{C} \rangle + \bar{C}_V \quad (5)$$

and similarly at tidal and overtide frequencies.

A key intermediate theorem is a generalization of the small-amplitude, steady-wave relationship between wave fluxes of wave frequency scalar variations and Stokes fluxes of the mean field to include vertical mixing. For the tidal frequency wave, GLM theory shows that

$$\overline{U' \cdot \nabla C'} \equiv \bar{U}^{St} \cdot \nabla \bar{C} - \frac{\partial}{\partial z} \bar{T}_z^{St} - \bar{W}_C \frac{\partial \bar{C}}{\partial z}^{St} \quad (6a)$$

and similarly for the overtide wave, assuming that $\bar{W}_C$ is time and space invariant. The idea behind the Stokes corrections to the mean sediment concentration and mean vertical turbulent flux in (6a) is analogous to the Stokes drift $\bar{U}^{St}$. In GLM theory, all Eulerian quantities have Stokes corrections that arise as the difference between their Eulerian (at a fixed point) and Lagrangian (following a particle trajectory) means [Andrews and McIntyre, 1978].

The derivation of (6a) is straightforward. One of the consequences of GLM theory is that a time-dependent, Eulerian scalar conservation relationship such as (3a) has a corresponding Lagrangian average (indicated by an overbar followed by a superscript L) equation following water parcel trajectories [Andrews and McIntyre, 1978, equations (2.20) and (2.21)]; thus in two dimensions (x and z),

$$\frac{\partial \bar{C}^L}{\partial t} = -\bar{U}^L \frac{\partial \bar{C}^L}{\partial x} - \bar{W}^L \frac{\partial \bar{C}^L}{\partial z} + \bar{W}_C \frac{\partial \bar{C}^L}{\partial z} + \frac{\partial}{\partial z} \bar{T}_z^L \quad (6b)$$

The equivalent Eulerian tidal cycle average of (3a) yields

$$\begin{aligned} \frac{\partial \bar{C}}{\partial t} = & -\bar{U} \frac{\partial \bar{C}}{\partial x} - \bar{W} \frac{\partial \bar{C}}{\partial z} - \bar{U}' \frac{\partial \bar{C}'}{\partial x} \\ & - \bar{W}' \frac{\partial \bar{C}'}{\partial z} + \bar{W}_C \frac{\partial \bar{C}}{\partial z} + \frac{\partial}{\partial z} \bar{T}_z \end{aligned} \quad (6c)$$

Subtraction of (6c) from (6b) yields (6a), after elimination of terms that are negligible for small-amplitude steady waves [Dunkerton 1980 equations (32) to (34)].

Vertical integration of (6a) for a two-dimensional channel flow then yields

$$\frac{\partial}{\partial x} \int_{-h}^{\zeta} \overline{U' C'} dz \equiv \frac{\partial}{\partial x} \int_{-h}^{\zeta} \bar{U}^{St} \bar{C} dz \quad (6d)$$

for the tidal frequency wave; an analogous relationship pertains for the overtide wave. The necessary boundary conditions are that wave velocities vanish at the bed and that there is everywhere zero net flux at the mean free surface, that is, that the sum of the vertical settling and vertical turbulent particle flux vanishes at ζ for all x and t . Note further that the integration in (6a) and (6d) may also be used on C_V instead of C . Together with the usual rules of averaging deviation quantities, this yields

$$\langle \overline{U' C'_V} \rangle = \langle \overline{U'_V C'_V} \rangle = \langle \overline{U_V^{St} \bar{C}_V} \rangle = \langle \overline{U^{St} \bar{C}_V} \rangle \quad (7)$$

Finally, use of (2c), (5), (6d), and (7) in (4a) gives two alternate forms of integral tidal-cycle-averaged sediment conservation:

$$\begin{aligned} \frac{\partial}{\partial t} \int_{x_1}^{x_2} \langle \bar{C} \rangle H dx = & \left[-\bar{Q}^R \langle \bar{C} \rangle \right. \\ & + \left(\langle \bar{U}_V \bar{C}_V \rangle + \langle \overline{U'_V C'_V} \rangle < \overline{U'_V C'_V} \rangle \right) H \Big]_{x_2}^{x_1} + \bar{V}_{fb} \\ = & \left[-\bar{Q}^R \langle \bar{C} \rangle + \langle \bar{U}_V \bar{C}_V \rangle \right. \\ & \left. + \langle (\bar{U}_V^{St} + \bar{U}_V^{So}) \bar{C}_V \rangle \right] H \Big]_{x_2}^{x_1} + \bar{V}_{fb} \end{aligned} \quad (8)$$

Both forms of (8) express net transport into and out of a box defined by the bed, mean free surface, and horizontal positions x_1 and x_2 in terms of along-channel changes in the shear and stratification between x_1 and x_2 and the bed flux $\bar{V}_{fb}$. They differ as to whether the shear and stratification on the right-hand side are expressed in terms of tidal or mean variables. Data analysis has customarily been done with the former; model calculations might be better formulated using the latter. Both forms of (8) contain five terms: a seaward river flow flux, a mean shear transport, tidal and overtide shear transports represented either as tidal or Stokes fluxes, and a flux to or from the seabed. The tidal and overtide frequency motions will be identified with M_2 and M_4 herein. Inclusion of additional tidal constituents within each tidal species would add one more term per constituent. Consideration of interactions between neighboring constituents within a tidal species would cause additional terms at tidal monthly frequencies to appear.

The absence of vertically integrated wave flux terms in (8) stems from the fact that landward sediment transport by the vertical integral of the Stokes drift at each wave frequency is exactly counteracted by seaward transport by the vertical integral of the corresponding Eulerian Stokes drift compensation flow to the sea [(2c) and (6d)]. This also shows why material from a vertically uniform sediment size class (wash load) cannot be trapped: without enough settling to induce stratification of such material, no landward shear flux can occur. The only possibility is uniform seaward transport.

Finally, the shear-sediment stratification correlations in (8) are extremely sensitive to the vertical distributions of U and C at each frequency. Both are strongly dependent on the intensity of vertical turbulent mixing. In many cases the turbulence field will be constrained primarily by the salinity distribution [e.g., Geyer, 1993; Hamblin, 1989]. Some ETM are, however, of sufficient intensity that sediment stratification effects on the velocity profile must also be considered [Gelfenbaum and Smith, 1986].

Results: Analysis of Turbidity Maximum Fluxes

The mass conservation expression (8) derived in the previous section is sufficiently general to allow analysis of ETM created by a variety of circulation processes, but the discussion in the present section is restricted to ETM created by convergent transport near the upstream limits of salinity intrusion. Specifically, we assume that (1) the time mean position of the ETM is at the tidal average upstream limit of salinity intrusion and (2) processes at locations seaward of mid-ETM occur entirely within the salt water mass, and those on the landward side occur entirely in fresh water. The latter assumption is useful as a conceptual device but represents a simplification of reality, which is complicated by tidal excursions that may be comparable to or larger than the length of the ETM (Figure 1).

Sediment Fluxes in a Steady ETM

It is a reasonable long-term goal to understand the full time-dependent evolution of an ETM under variable tidal and fluvial

forcing. It is useful, however, and consonant with nonacceleration theorems derived for atmospheric flows (as summarized by Dunkerton, [1980]) to begin interpretation of the sediment balance derived in (8) by analysis of the simplest relevant situation. A steady state in which $\bar{Q}^R$ is spatially invariant and tidal-average bed flux $\bar{V}_{fb}$ and the left side of (8) vanish is therefore assumed. This yields

$$O = [-F_R + F_{SH}]_{x_1}^{x_2} = F_{NET}|_{x_1}^{x_2} \quad (9)$$

where F_{SH} is the sum of the shear-stratification correlations at all frequencies, and $F_R = \bar{Q}^R < \bar{C} >$. In this case the net sediment flux $F_{NET} = F_{SH} - F_R$ through x_1 is the same as that through x_2 . Since x_1 and x_2 are arbitrary, F_{NET} must be spatially uniform for a steady state to pertain (Figure 8). We then choose x_1 and x_2 on either side of the ETM such that $< \bar{C} >|_{x_1} = < \bar{C} >|_{x_2}$; then $F_R|_{x_1} = F_R|_{x_2}$. In the simplified,

single-size-class situation considered here, the river flow transports exactly as much material out of the ETM at x_1 as it brings in at x_2 . This emphasizes the indirect role of river flow in creation of an ETM: the river supplies part (usually most) of the material that is concentrated by along-channel gradients in shear fluxes F_{SH} . It also supplies the buoyancy flux that sets the stratification and vertical mixing that together determine the shear, but it does not in itself create the ETM through convergent horizontal fluxes.

The along-channel variations in F_{SH} and F_R may also be determined (Figure 8). This is possible because we have assumed no erosion or deposition, and thus F_{NET} , the net sediment flux to the sea, is spatially and temporally invariant. Within the ETM the amplitude of seaward fluvial flux $F_R = \bar{Q}^R < \bar{C} >$ is at a maximum because $\bar{Q}^R$ is spatially invariant and $< \bar{C} >$ is at its highest value. This means that F_{SH} is at a maximum (i.e., relatively more landward) within the ETM. This argument does not alone prove that F_{SH} is positive in sign

Figure 8. (top) Schematic sketch of vertically integrated salinity (short-dash curve) and sediment distributions (solid curve) and the negative of the along-channel gradient of the sediment distribution (long-dash curve). Positive x is landward and to the right. (bottom) Spatial distributions of total F_{NET} (solid curve), shear F_{SH} (dashed curve), and fluvial F_R (dashed curved) fluxes. The ETM extends approximately from x_1 to x_2 . Countergradient sediment transport occurs on the seaward side of the ETM (shaded). In the absence of fluxes to and from the bed, F_{NET} must be spatially uniform for the tidally averaged sediment field to be in a steady state. Both F_{SH} and F_R have their maximum amplitudes in the ETM. F_{SH} is landward, and F_R is seaward.

(landward) anywhere; it could be uniformly seaward but less so within the ETM. Consideration in the following paragraphs of the individual contributions to F_{SH} suggests, however, that F_{SH} is in fact positive within the ETM and probably everywhere.

Simple arguments may also be made concerning the along-channel forms of C' and C'' . One foundation of Eulerian wave flux methodology is an assumption that lowest-order wave variations in scalar properties stem from wave advection of the mean field (Middleton and Loder, 1989). This assumption, together with a separation of variables yields:

$$\bar{C}(x, z) = M(x)C_0(z) \quad (10a)$$

$$C'(x, z, t) = R_e \left[\frac{dM}{dx} \langle U_1 \rangle C_1(z) e^{it} \right] \quad (10b)$$

$$C''(x, z, t) = R_e \left\{ \left[A_1 \frac{dM}{dx} \langle U_2 \rangle + A_2 \langle U_1 \rangle \frac{d}{dx} \left(\frac{dM}{dx} \langle U_1 \rangle \right) \right] C_2(z) e^{2it} \right\} \quad (10c)$$

where $\langle U_1 \rangle$ and $\langle U_2 \rangle$ are semidiurnal and quarter-diurnal tidal transport magnitudes; A_1 and A_2 are unknown constants; C_0 , C_1 , and C_2 are vertical structure functions for the various suspended sediment modes; and the spatial scales of $\langle U_1 \rangle$ and $\langle U_2 \rangle$ can be assumed large compared to that of the turbidity maximum. The separation of variables employed in (10a) to (10c) assumes that the vertical structure of the sediment field (the $C_i(z)$, $i=1, 3$) is invariant in the along-channel direction at each frequency. Even if this assumption is not rigorously correct, it is useful for insights it provides into the along-channel variability of the sediment field and horizontal fluxes thereof at tidal and overtide frequencies. Assume then that the horizontal part $M(x)$ of the mean sediment field $\bar{C}$ is bell-shaped; then C' vanishes at mid-ETM (Figure 9). This is sensible, because we expect C' to be larger on flood on the seaward side of the ETM and on ebb and the landward side. It is not clear without a detailed solution whether the M_4 sediment field C'' is predominantly due to M_2 advection of C' ($A_2 \gg A_1$ in (10c)) or M_4 advection of the mean field ($A_2 \gg A_1$). The only assumption about C'' made here is that it has a maximum in mid-ETM. This follows from the fact that the bed stress is neither flood nor ebb dominant: a maximum amount of energy is at residual and overtide frequencies.

These considerations concerning along-channel variations of $\bar{C}$, C' , and C'' can be used, along with vertical distributions of U at mean and wave frequencies, to determine spatial variation of the various contributions to F_{SH} . Of considerable interest is the fact that F_{SH} is countergradient on the seaward side of the ETM (Figure 8). Although $\langle \bar{U} \rangle < 0$ (from (2c)) everywhere, $\bar{U}_V$ is landward near the bed on the seaward side of the ETM (e.g., at $x = x_1$) as a result of two processes (gravitational circulation and internal tidal asymmetry) that are directly related to the presence of a horizontal salinity gradient (Figure 10). $\bar{C}_V$ is positive near the bed and negative near the free surface for all x . The correlation of $\bar{U}_V$ and $\bar{C}_V$ causes a landward (countergradient) mean-flow contribution to F_{SH} that is strong in many systems (as in Figures 6 and 7). The along-channel distribution of F_{SH} seaward of mid-ETM is a balance between those of $\bar{C}_V$ (maximum in mid-ETM) and $\bar{U}_V$, which must decrease sharply at the upstream limit of salinity intrusion. The maximum mean shear flux should then be at or seaward of mid-ETM. On the landward side of the ETM (e.g., at $x = x_2$), gravitational circulation is symmetry decreases rapidly with increasing x , and only a weak landward, downgradient mean shear transport can occur.

Figure 9. From top to bottom, $\langle \bar{C} \rangle$ and its first two x derivatives as a function of x on the horizontal axis (positive x landward). $\langle C' \rangle$ should be proportional to the first derivative. $\langle C'' \rangle$ has terms proportional to both the first and second derivatives. Note that $\langle C' \rangle$ vanishes at mid-ETM, whereas $\langle C \rangle$ and the second-derivative contribution to $\langle C'' \rangle$ are at a maximum there.

Internal tidal asymmetry also contributes to the M_4 shear flux on the seaward side of the ETM. As with the steady, density-driven circulation, U_V' is "landward" (i.e., M_4 is nearly in phase with M_2) near the bed and "seaward" (i.e., M_4 is out of phase with M_2 velocities) at the surface. This pattern of overtide currents combined with strong M_4 variations in near-bed sediment concentrations that occur at or shortly after times of peak flood and ebb flow near the bed (Figure 3) causes strong landward (countergradient) transport, as shown figuratively in Figure 11. Here C_V'' and U_V'' are probably both at a maximum in mid-ETM. Thus the overtide contribution to F_{SH} should also have a maximum in mid-ETM. This is consistent with the observations presented in Figures 6 and 7, which were collected at sites seaward of mid-ETM in the Columbia River estuary. Internal tidal asymmetry must drop rapidly to zero landward from mid-ETM as salinity is present for a smaller and smaller part of the tidal cycle. Finite-amplitude wave steepening and frictional typically shorten and sharpen the flood and lengthen and weaken

Figure 10. Schematic vertical distribution of mean flow variables (left) $\bar{U}$ and $\bar{C}$ and (right) $\bar{U}_v$ and $\bar{C}_v$. $\bar{C}$ increases monotonically with depth, so $\bar{C}_v > 0$ at the bed, as is also true of the sediment stratification at other frequencies. $\langle \bar{U} \rangle < 0$ from (2c); this and the change of sign with depth of $\bar{U}$ leads to $\bar{U}_v > 0$ at bed. $\bar{U}_v$ and $\bar{C}_v$ are positively correlated.

the ebb, should in the absence of large tidal flats bring about landward (downgradient) transport [Aubrey, 1986]. This transport will normally be weak, because barotropic M_4 generation within one salinity-intrusion length of the mouth of most narrow estuaries is small.

This leaves only the tidal frequency contribution to the F_{SH} to be considered. Assume, consistent with the sense of near-bottom mean flow, that the phase of the tidal frequency variations in C' is such that C' and C'_v are at a maximum on flood seaward of the ETM and on ebb landward thereof (Figure 12). In the absence of a salt wedge, the tidal flow decreases monotonically with depth. Then U'_v is negative near the bed, where C'_v is positive, and positive near the surface, where C'_v is negative, leading to a net seaward (downgradient) transport that is likely to be weak, particularly near mid-ETM, where the residual flow changes sign and C' is small. The presence of a strong salt wedge could cause a positive near-bed U'_v and might bring about a landward tidal shear flux. It would probably be weak near mid-ETM. The results of Figures 6 and 7 indeed suggest

landward transport by the M_2 tide at locations seaward of mid-ETM. At one station (NC1) this landward transport may have been large relative to those at Z_0 and M_4 associated with internal tidal asymmetry. Because C' is at a maximum on ebb on the landward side of the ETM, and because salt wedge flow is by definition absent, transport will be landward (downgradient) but again probably weak near mid-ETM. Results of the above discussion of fluxes in a time-invariant ETM are summarized in Table 1.

Transient ETM Sediment Fluxes

Time-dependent development of an ETM after a neap tide and its deterioration after a spring tide may also be considered using (8). It is again assumed that x_1 and x_2 can be chosen such that $\langle \bar{C} \rangle_{x_1} = \langle \bar{C} \rangle_{x_2}$ over the period of consideration, so that the river flow does not contribute directly to ETM time variation. Time changes in the total amount of suspended sediment between x_1 and x_2 may then appear as the result of two processes: net erosion or deposition at the bed and convergence/divergence of

Figure 11. Schematic peak flood vertical distributions of overtide variables (left) U'' and C'' and (right) U''_v and C''_v . The spatial distributions of overtide shear and stratification are similar to those at zero frequency. Their phase relationship is such that U''_v and C''_v are positively correlated. In this case, however, U'' integrates to zero, so that it does not influence the barotropic flow.

Figure 12. Schematic peak flood vertical distributions of tidal variables (left) U' and C' and (right) U_v' and C_v' . Because U' increases monotonically from the bed, $U_v' < 0$ at the bed; U_v' and C_v' are thus negatively correlated.

horizontal shear fluxes in the mean, tidal, and overtide flows. After a period of weak tides, an ETM will develop as long as the sum of bed erosion and convergence terms in (8) is positive. Erosion is likely to occur because of increased bed stress. If convergence occurs, it requires transport that is countergradient on the seaward side of the ETM. The ETM will cease to develop when no more net erosion of bed material is possible given ambient shear stresses and/or the along-channel maximum in $\bar{C}$ becomes so pronounced that countergradient transport on the seaward side of the ETM (and thus a convergent horizontal sediment flux) is no longer possible. Conversely, ETM deterioration after a period of strong tides can be expected when the horizontal shear flux becomes divergent (probably down-gradient on both sides of the ETM) and/or bed stress is insufficient to keep net deposition from occurring. Tidal monthly variability in the position of the tidal-average salinity intrusion limit may further complicate transient ETM processes, because the entire ETM will typically move landward and seaward with the salinity intrusion, unless it is fixed in one location by topographic influences. Thus time-dependent ETM evolution is inherently more complex than the steady state summarized in Table 1, and model calculations would likely be needed to determine the spatial distributions of the various flux terms.

Finally, the distinct along-channel transport properties of salt and suspended sediment deserve comment. These dissimilarities arise because salt and sediment have different phase relationships with the along-channel velocity field. Salinity has a maximum close to high-water slack and is at a minimum around low-water slack. The largest components of salinity variability are usually at Z_0 and the dominant tidal frequency (M_2 in the present case). In contrast, suspended sediment concentration has maxima on both flood and ebb, and the largest components of sediment variability are at Z_0 and the first overtide frequency of the dominant tidal wave (M_4 in this case). The exact timing of the sediment concentration maxima is a function of size class and may be strongly influenced by horizontal advection. In some systems, furthermore, two or more suspended sediment concentration maxima are seen on each flood and each ebb [Lang *et al.*, 1989], because of variability of stratification during each half tidal cycle and because of the presence of M_4 in the velocity field. These two factors result in M_8 variability in the excess shear stress and thus in the suspended sediment field [Jay and Musiak, 1994]. From a wave flux perspective, however, it is the

settling and erosion properties of suspended sediment that cause the details of the time-dependent suspended sediment distribution and that thus differentiate the transport behavior of all sediment classes (except wash load) from that of salinity.

Discussion: Implications for Models of Reduced Dimensionality

Scalar transport problems are often considered in models of reduced dimensionality, both to limit computational complexity and to provide insight into transport processes. The transport formalism developed above has definite implications for the utility of such models. These are developed separately for vertical and horizontal models, because the two approaches have different strengths and weaknesses.

Vertically Integrated Models

It is a long-established and convenient practice to use a VIHD model to calculate salinity intrusion length and the vertically averaged salinity in tidal channel flows [e.g., Harleman, 1966]. After careful calibration, such models may accurately reproduce known conditions [e.g., Uncles and Stephens, 1990; Robinson, 1983; Holloway, 1981]. Conceptually, however, this procedure is unattractive because it is impossible to maintain a two-dimensional (in x and z) salt balance against outward river flow without landward transport by shear-stratification correlations

TABLE 1. Estuarine Turbidity Maximum Fluxes

Frequency	Seaward Side	Landward Side
Residual	++ (density driven, countergradient)	+
Tidal	+/-	+
Overtide	++ (internal, countergradient)	+(barotropic)

Here plus is landward; minus, seaward.

[Jay, 1991b], just as it is impossible to trap wash load in an ETM. Thus along-channel salt diffusivity is a function of precisely those variables not calculated by VIHD models. For example, tidal monthly transitions in circulation and stratification alter the basic mechanisms (the shear fluxes) maintaining the salinity intrusion, and because these transitions occur at a tidal amplitude that is a function of river flow, salinity intrusion length may be a complex function of tidal and fluvial forcing. In the Columbia River estuary, the present-day river flow cycle management is such that neap-tide salinity intrusion exceeds that on spring tides most of the year, while the opposite situation prevails in an apparently very similar system, the Fraser River estuary [Geyer, 1988; Jay and Smith, 1990]. This suggests that even a solution such as that of Kranenburg [1986], in which tidal average salinity is a function of both x and $\bar{Q}^R$, will not in general work without a great deal of tuning. VIHD models are thus limited as predictive tools, even for salinity.

Vertically integrated models in one or two horizontal dimensions are poorly suited to description of the suspended sediment field and its transport, because each size class of sediment will have distinct transport characteristics and because apparently countergradient transport will occur in the vicinity of maxima in the longitudinal distribution. Horizontal sediment diffusivities in such a model must either be tuned empirically for each size class or be determined from the distribution of some other scalar such as salinity. The former will lead to negative diffusivities in the circumstances defined in previous sections and limits the role of such a model essentially to interpolation of observations. The latter necessarily leads to incorrect representation of wave flux transports of sediment.

One-Dimensional Vertical Models

Most calculations of suspended sediment profiles, whether steady or time-dependent, have assumed horizontal uniformity, so horizontal divergence effects could be neglected. Although the above analysis has not investigated the details of the vertical distributions of sediment at the various wave frequencies, it has emphasized that ETM dynamics must be viewed as a balance between local vertical processes and horizontal advection. Including local acceleration effects in calculations of sediment profile creates a phase shift between maximum bed stress and peak sediment concentration at points removed from the bed [Smith, 1977]. This lag represents the time required for sediment to diffuse vertically in the water column. Horizontal advection adds another factor; the maximum sediment concentration may lead or lag the local maximum bed stress because advection from locations of higher mean concentration (resulting from greater bed stress, greater bed availability, or both) will, at points removed from the middle of the ETM, partially determine the timing of maximum concentration. In the absence of a complete two-dimensional (in x and z) theory of sediment distribution for tidal channels, it is difficult to assess the relative role of time-varying stratification, nonlinearities in the erosion/deposition process and horizontal advection in creating the settling and erosion lags discussed by Dyer [1988].

The ETM in both the Columbia and the Hudson River estuaries are considerably less than one tidal excursion in length. In such river estuaries, where horizontal gradients are sharp, there is no justification for excluding tidal advection from calculations of suspended-sediment profiles. It can further be predicted that the larger $\bar{W}_C$ is for the sediment trapped in the ETM (i.e., the larger the prevailing bed stress), the more important advection is

to ETM dynamics. That is, any increase in the ratio of settling velocity $\bar{W}_C$ to horizontal advection will bring about an ETM that is more focused at the point of transport convergence. This enhances horizontal gradients on either side of mid-ETM and therefore the effects of advection.

Summary and Conclusions

Conservation of mass has been used to derive a new, two-dimensional (in x and z) suspended-sediment balance (8) for estuarine turbidity maxima (ETM). This balance is simplified using the constraints of two-dimensional geometry and small-amplitude tides and a relationship between wave transport of tidal fluctuations in scalar concentration and Stokes drift transport of the mean scalar field. The resulting balance shows that time changes in tidal-mean sediment size-class concentration between two estuarine cross sections are caused by the sum of seaward river flow transport, landward transport associated with correlations between shear and stratification in the mean flow and at the various tidal frequencies, and fluxes to and from the bed. Vertically integrated wave modes (e.g., correlations of vertical averages $\langle U' \rangle$ and $\langle C' \rangle$ of along-channel tidal velocity U' and sediment concentration C') cannot transport material landward, because this landward transport is exactly counteracted by seaward transport associated with the Eulerian residual excited by the landward Stokes drift.

This balance emphasizes the joint role of horizontal advection and vertical processes (settling, deposition, vertical mixing, and erosion) in determining suspended sediment distributions in estuarine systems. Thus appearance of an ETM at a given location does not necessarily indicate a local maximum in resuspension at that location. It represents instead a balance between local resuspension and convergent horizontal advection. If in fact there is a resuspension maximum in mid-ETM, this strong local resuspension likely reflects convergent transport at some earlier time, either during a tidal month or a season. Advection from a location of greater bed concentration provides, furthermore, another explanation (in addition to nonlinearities in erosion/deposition at the sea bed, acceleration, and time-varying stratification) of why peak sediment concentration does not occur at the time of peak bed stress. The importance of advection varies directly with the settling velocity of material trapped in the ETM. Thus advection is likely to be most important in systems with strong bed stress that traps relatively coarse, rapidly settling material.

The sensitivity of (8) to the details of U and C profiles at zero and at all tidal frequencies also emphasizes the importance of accurately calculating these profiles. Such calculations require careful attention to stratification effects on vertical mixing caused by salinity. In some cases the self-stratification of suspended sediment must also be considered. Furthermore, the balance also illustrates quite simply why washload cannot be trapped: in the absence of vertical variations in sediment concentration, only uniform advection to the sea is possible.

Conclusions may also be drawn concerning processes leading to ETM creation and maintenance. In the simplest situation in which vertically averaged, tidal-mean sediment concentration $\langle \bar{C} \rangle$ for a given size class in the ETM is invariant from one tidal cycle to the next and the integrated river flow $\bar{Q}_R$ is steady and uniform in the along-channel direction, the local maximum of $\langle \bar{C} \rangle$ in the ETM causes a local maximum in the seaward transport by the river flow $F_R = -\bar{Q}_R \langle \bar{C} \rangle$. Because a steady

state situation implies a spatially and temporally constant net sediment flux F_{NET} toward the sea, landward transport of sediment by the various shear-stratification modes (F_{SH}) must be at a maximum in the ETM so that $F_{NET} = -F_R + F_{SH}$ is spatially uniform. If F_{NET} were not spatially invariant, then sediment would accumulate or be removed at some location. When represented in terms of a one-dimensional, vertically integrated horizontal diffusion model, F_{SH} is countergradient on the seaward side of the ETM. Consideration of vertical distribution of the various flow modes suggests that it is primarily the residual and overtide shear fluxes that are responsible for the countergradient transport. The presence of diurnal currents in West Coast estuaries may also, however, through quadratic bed stress nonlinearities, bring about the presence of a substantial semidiurnal contribution to ETM sediment fluxes.

It is also important to relate the present analysis to the idea advanced by Hamblin [1989] and Geyer [1993] that inhibition of vertical mixing by density stratification contributes to near-bed variations in suspended sediment concentration in a way that is likely to be correlated with tidal advection. In the context of the present analysis, the suppression of upward mixing of sediment by density stratification serves to trap sediment near the bed on the seaward side of the ETM, where it may be moved landward by gravitational circulation and/or internal tidal asymmetry. Geyer is correct in pointing out that convergent two-layer mean flow has been emphasized (and possibly overemphasized) in earlier studies as the primary trapping agent. The present analysis shows, moreover, that density stratification may enhance particle trapping through its influence on the vertical distributions of both suspended-sediment and velocity at a variety of frequencies, because it is correlations between velocity shear and suspended-sediment stratification at residual and various tidal frequencies that traps particles. To take these arguments one step further: the effect of tidally varying vertical scalar mixing may drive overtide and residual variations in particle concentration through the vertical turbulent flux term in the sediment conservation equation. The mechanism is exactly analogous to the time-varying vertical momentum flux that gives rise to internal tidal asymmetry in the along-channel flow field.

The present analysis also has interesting implications for box and other vertically integrated horizontal diffusion models. The absence of $\langle U \rangle$ and $\langle C \rangle$ from the integral balance (except in F_R) shows that the channel flow transport scalar is inherently two-dimensional and can be reduced to a one-dimensional diffusion problem only under special circumstances. Mean and overtide sediment transports on the seaward side of the ETM are countergradient (for a steady state and in terms of the vertically integrated sediment concentration), even though salt transport is downgradient at the same location. Landward transport occurs for both salt and sediment, because both are concentrated near the bed, where transport is landward. That a vertically integrated horizontal diffusion model can often be made to describe an along-channel scalar distribution is convenient, but it is not a reflection of fundamental physics. Thus vertically integrated horizontal diffusion models should be applied with care to stratified tidal channel flows.

The different along-channel transport properties of salt and suspended sediment arise out of the details of their respective relationships to the tide and overtide velocity fields. Salinity is high at the end of flood and low at the end of ebb. Suspended sediment concentration usually has maxima on both flood and ebb in response to flood and ebb maxima in bed stress. More fundamentally, the suspended sediment and salinity fields behave

differently because settling and deposition/erosion introduce additional degrees of freedom into the sediment balance, and tidal variations in sediment concentration are not exclusively the result of wave advection of a mean field.

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