

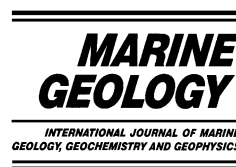


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# Post-depositional alteration and preservation of sedimentary event layers on continental margins, I. The role of episodic sedimentation

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## Abstract

The degree of post-depositional alteration and hence the preservation potential of sedimentary event beds and transient signals is determined by the outcome of a ‘race’ between biological (and to a lesser extent physical) processes that conspire to destroy a signal and sediment accumulation which advects the layer or signal out of the surface mixing zone. Preservation potential specifically depends on the relative magnitudes of the (1) biological mixing intensity (sometimes parameterized as a biodiffusivity,  $D_b$ ), (2) mixing-layer thickness, (3) layer or signal thickness, and (4) sediment accumulation rate. These terms control the dissipation time (i.e. time required to destroy a signal) and the transit time (i.e. time necessary to advect a signal through the surface mixing layer). On fine-grained, upper continental margins in general, and the Eel River shelf (northern California) in particular, biological mixing intensity is high (i.e.  $D_b$  ranges from 10 to 100 cm<sup>2</sup>/yr), mixing-layer thickness is large (> 10 cm), and sediment accumulation rates are rapid (0.1–1 cm/yr). Despite the high sediment accumulation rates, transit times through the surface mixing-layer range from decades to a century. Signal dissipation times are considerably shorter: (1) < 3 yr for beds up to 6 cm thick imaged in X-radiographs, and (2) < 15 yr for the grain-size signature of beds up to 8 cm thick. Therefore event layers and their corresponding grain-size signature have a low probability of preservation. However, short-lived episodic sedimentation events (e.g. oceanic floods) can instantaneously advect material through the surface mixing layer, thereby preserving event beds and transient signals. On the Eel River shelf the sequential timing of episodic sedimentation events has exerted a first-order control on the resultant stratigraphic record (presence/absence of layers and grain-size fluctuations). Episodic sedimentation – a hallmark of continental shelf settings – is key to understanding strata formation and preservation on margins.

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## 1. Introduction

Much of marine sedimentary geology involves a careful deciphering of the stratigraphic record – that is, a reading of the seabed recorder. Under

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ideal circumstances successive sediment transport events (e.g. storms or floods) or transient climatic signals would leave a clear, unbiased record in the sediment column, thereby permitting unambiguous matching of process and product. For reasons that involve both physical (e.g. resuspension) and biological (e.g. bioturbation) processes the real situation is far from ideal. In other words, the fidelity of the seabed recorder is often poor and always variable. All is not chaos, however. Patterns of signal alteration and preservation do exist, and elucidating such patterns is important to many aspects of marine geology.

Problems involving the fidelity of the seabed recorder are perhaps most acute on the upper slope and shelf, where important processes, such as sediment mixing and accumulation, are likely to be more variable than they are in the deep sea. Because of this variability (i.e. unsteadiness) studies of paleoceanography and paleoclimatology have often avoided continental margins. For at least two reasons this situation is now changing. First, there are many important marine processes, for example coastal upwelling (van Geen et al., 1992), that are only recorded in upper margin sediments. Moreover, signals of other phenomena, such as continental weathering or floods (e.g. Drake et al., 1972; Wheatcroft et al., 1997), may be considerably attenuated by the time they reach the deep sea, thereby necessitating a focus on proximal regions of the dispersal system. Second, continental margins generally have higher sediment accumulation rates than the deep sea. These higher accumulation rates may permit identification of high-resolution signals that otherwise are destroyed by biological processes in slowly accumulating deep-sea sediments. In short, studies of high-frequency stratigraphy on continental margins are apt to become increasingly common.

The alteration and preservation of transient sedimentary signals, especially event layers, on continental margins exhibit some subtle differences compared to the situation in most of the deep sea. This paper explores those differences and has three objectives. First, the conceptual framework regarding event-layer alteration and preservation that has been developed over the past 25 years

(e.g. Guinasso and Schink, 1975; Nittrouer and Sternberg, 1981; Wheatcroft, 1990) is re-examined in light of recent measurements on continental margins. In particular, the large number of studies that have employed radionuclides (e.g.  $^{210}\text{Pb}$ ,  $^{137}\text{Cs}$ ,  $^7\text{Be}$ ,  $^{234}\text{Th}$ ) to measure relevant terms in the problem (e.g. sediment mixing intensity, sediment accumulation rate) permit a fresh examination of potential correlations between terms, as well as their range. The hope is that a particular variable may, because of its magnitude or range, exert a disproportionately strong influence on event-layer preservation and alteration. Second, focus will be placed on the Eel River shelf, the site of the recent STRATAFORM program (Nittrouer, 1999). There, studies by several investigators (e.g. Wheatcroft et al., 1997; Drake, 1999; Sommerfield and Nittrouer, 1999; Sommerfield et al., 1999; Wheatcroft and Borgeld, 2000; Bentley and Nittrouer, 2004; Wheatcroft, in prep.) have provided specific values to important terms in the preservation problem. In particular, intensive time-series coring efforts have allowed us to track the fate of event layers over a several-year period. Third, high-resolution grain size and X-radiographic evidence will be examined to address event-layer alteration and preservation on the Eel River shelf. These two data sets are complementary because a grain-size signal, unlike bedding imaged in X-radiographs, can be mixed and remain recognizable. Bedding, on the other hand, is by definition destroyed once its fabric has been mixed (cf. Wheatcroft, 1990). The emphasis of this last section will be to assess which flood deposits have been preserved on the Eel River shelf over the past 50 years. In all three sections, the critical importance of episodic sedimentation in determining event-layer alteration and preservation on continental margins will be highlighted.

## 2. Conceptual framework

The general processes related to the preservation of strata (*sensu lato*) have been recognized for many years. For example, Moore and Scruton (1957) in their study of the Mississippi River delta clearly implicated sediment accumulation rate and

benthic biological processes to be important in determining the internal structure of marine sediments. Similarly, the many comparative studies conducted by Reineck, Howard and Frey off the coast of Georgia and elsewhere demonstrated the relative importance of biological and physical processes in shaping the sedimentary record (e.g. Reineck, 1967; Howard and Reineck, 1972; Howard and Frey, 1975; Howard, 1978; Howard and Reineck, 1981; Frey et al., 1989). These studies were qualitative in nature, however; thus the absolute rates of biological reworking and sediment accumulation remained unknown and a quantitative conceptual foundation was lacking.

A significant step toward erecting a rigorous conceptual framework of strata preservation was made following the development of radiochemical (e.g.  $^{210}\text{Pb}$ ,  $^{137}\text{Cs}$ ,  $^{234}\text{Th}$ ) techniques to quantify sediment accumulation rate and biological mixing intensity (e.g. Goldberg and Koide, 1962; Guinasso and Schink, 1975; Aller and Cochran, 1976; Nittrouer et al., 1979). Because the partial differential equations used to model and hence extract values of the above terms need to be prescribed over a limited vertical domain within the seabed, the concept of a surface mixing layer was also born (Goldberg and Koide, 1962; Guinasso and Schink, 1975). It is these three terms, sediment accumulation rate ( $S$ ), biological mixing intensity ( $D_b$ ) and mixing-layer thickness ( $L_b$ ), that were subsequently shown by Nittrouer and Sternberg (1981) to control the preservation of sedimentary strata. Nittrouer and Sternberg (1981) provided scaling arguments that suggested a non-dimensional combination of these terms,  $G$  (where  $G = D_b/L_b S$ ), could be used to identify dominance by mixing (large  $G$ ) versus burial (small  $G$ ).

Although the three terms that comprise  $G$  are indeed relevant to the preservation problem, Wheatcroft (1990) subsequently showed that inclusion of  $L_b$  in the denominator was inconsistent. That is because when  $L_b$  is large, a signal or event bed is subject to mixing for a longer period of time, hence it should have a lower preservation potential. Under the  $G$  formulation, however, a large  $L_b$  results in a small  $G$ , which favors burial or preservation. Wheatcroft (1990) went on to suggest that focus should be placed on time scales.

In this formulation ‘transit time’ ( $T_m = [L_b - L_s]/2/S$ , where  $L_s$  is event bed thickness) determines the period of time required to advect an event layer (or signal) through the surface mixing layer. Whereas ‘dissipation time’ ( $T_d$ ) represents the time necessary to destroy a signal or event bed. In cases where  $T_m < T_d$ , some portion of the event bed is preserved intact in the historical record.

Although conceptually simple and internally consistent, there are several challenges in applying Wheatcroft’s (1990) formulation based on time scales. First, empirical and theoretical knowledge of dissipation time is lacking, although recent contributions regarding the latter are encouraging (e.g. Boudreau et al., 2001; Bentley and Nittrouer, 2004). Moreover, dissipation time varies as a function of the signal in question. Second, the definition of mixing-layer thickness is not straightforward; therefore, estimation of transit time is ambiguous. Lastly, sediment accumulation rate is known to vary as a function of observation period (e.g. Sadler, 1981), therefore transit time will vary substantially depending on the choice of averaging time. These issues are explored in detail next.

### 2.1. Biodiffusivity, $D_b$

Particle displacement, ‘bioturbation’, is a nearly ubiquitous process in marine sediments. Only in areas of low bottom-water oxygen content (e.g. oxygen minimum zones, restricted basins), near-continuous physical sediment transport (e.g. beaches, tidal channels) or extremely rapid sediment accumulation (e.g. delta fronts of major rivers, proximal glacial environments) are bioturbating animals excluded. Everywhere else in marine sediments particles are displaced due to a variety of animal activities (e.g. deposit feeding, burrowing, tube building). It is this particle displacement that destroys physically produced sedimentary structures or layers and mixes transient signals to a state where they are no longer recognizable. In considering event-layer alteration and preservation it is important to obtain some knowledge of the intensity of biogenous particle displacement and its within-sediment depth dependence.

Most often bioturbation is parameterized as a

quasi-diffusive process that can be represented by a biodiffusivity,  $D_b$  (Goldberg and Koide, 1962; Guinasso and Schink, 1975; Boudreau, 1986; Wheatcroft et al., 1990). Although conceptualizing bioturbation as a diffusive process is a gross oversimplification that has many shortcomings, it is indisputable that many tracer (both impulse and continuous) profiles match diffusive profiles. It is also likely that dissipation time for a given layer or signal will scale inversely with the biodiffusivity. That is,  $D_b$  remains a valuable indicator of the intensity of biological sediment reworking irrespective of the modes (i.e. diffusive, non-local, etc.) of that bioturbation.

Biodiffusivities can also be estimated quantitatively. Over the past 25 years, many such estimates have been made in a wide diversity of oceanic environments using a variety of tracers, including naturally occurring radionuclides (e.g.  $^{234}\text{Th}$ ,  $^7\text{Be}$ ,  $^{210}\text{Pb}$ ) and deliberate tracers (e.g. glass beads, luminophores, radio-labeled particles). Recent compilations of  $D_b$  (e.g. Boudreau, 1994; Tromp et al., 1995) indicate roughly a five order of magnitude range from  $\sim 0.001$  to  $\sim 250$   $\text{cm}^2/\text{yr}$  in biologically active sediments. This wide range suggests that variations in sediment mixing intensity, and hence dissipation time, on continental margins could play an important role in determining event-bed alteration and preservation. The available evidence, however, indicates that this is not the case. The main reason for this bold statement is that biodiffusivities appear to have a strong correlation with organic C flux, and hence sediment accumulation rate (Boudreau, 1994; Tromp et al., 1995). Therefore in regions – such as the continental shelf – where sediment accumulation rate is high the biodiffusivity also increases. A compilation of sediment biodiffusivity estimates for estuarine and coastal ocean sites based on profiles of excess  $^{234}\text{Th}$  (Fig. 1) indicates a much smaller range in  $D_b$  (two orders of magnitude). In fact, most of the low biodiffusivities depicted in Fig. 1 represent winter conditions on the eastern seaboard of the United States (e.g. Long Island Sound, NY; Buzzards Bay, MA). If these are excluded, then the range decreases still further (i.e. 10 to  $\sim 100$   $\text{cm}^2/\text{yr}$ ). These results suggest that continental shelf sediments are mixed vigorously

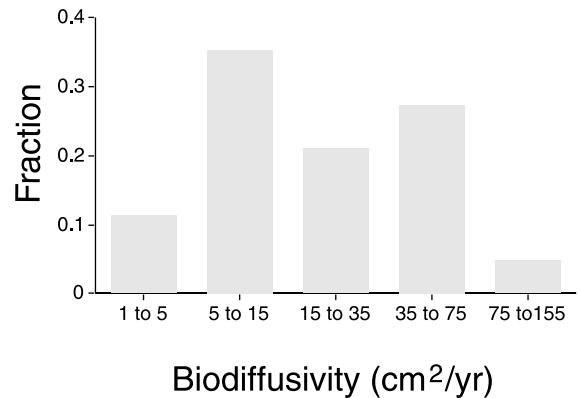


Fig. 1. Histogram of biodiffusivities based on profiles of excess  $^{234}\text{Th}$  on continental shelves and estuaries. Data sources: Aller and Cochran (1976), Long Island Sound; Aller et al. (1980), Long Island Sound; Santschi et al. (1980), New York bight; Nittrouer et al. (1983), Washington shelf; McKee et al. (1983), Yangtze River estuary; Silverberg et al. (1986), Laurentian Trough; Martin and Sayles (1987), Buzzards Bay; Wheatcroft and Martin (1996), Palos Verdes shelf; Fuller et al. (1999) San Francisco Bay.

by animals at approximately equal intensity, and hence dissipation times are likely to be relatively short and invariant.

Sediment bioturbation intensity also likely displays within-sediment depth dependence (Guinasso and Schink, 1975), whereby sediment is more intensively displaced near the sediment–water interface. Unfortunately, as Boudreau (1986) made clear, there are substantial difficulties in actually identifying and quantifying that depth dependence. These difficulties are due mainly to the rapid decay rate of biologically useful radionuclides, such as  $^{234}\text{Th}$ , and the relatively slow decrease in mixing intensity with depth. Nevertheless, it is difficult to imagine that sediment mixing intensity does not fall off with depth. The significance of this depth dependency vis-à-vis event-layer thickness is discussed in Section 2.3.

## 2.2. Mixing-layer thickness, $L_b$

The surface mixing layer is a region adjacent to the sediment–water interface in which biological and physical sediment displacement occurs. Below this layer there is no translocation of particles, and properties are maintained in their relative po-

sitions. An accurate knowledge of the vertical extent of the mixing layer and whether it varies systematically in space or time is important in determining transit time. Recent empirical (Boudreau, 1994) and theoretical (Boudreau, 1998) results suggest that  $L_b$  has a worldwide average of  $10 \pm 5$  cm ( $1\sigma$ ) and that it is relatively insensitive to sediment accumulation rate or water depth (but see Smith and Rabouille, 2002). Although it would be appealing to assign glibly a 10 cm thick  $L_b$  to all sites on the continental margin, doing so would be potentially wrong for at least two reasons. First, there are several inconsistencies in the definition of  $L_b$  and hence its measurement. Therefore, the data compilation used by Boudreau (1994) and the resultant average are questionable. Second, the concept of a surface mixing layer that is embodied in radiochemical studies is simply not appropriate for the problem at hand, namely alteration and preservation of sedimentary signals.

In the past 40 years at least three definitions of the surface mixing layer have become established in the literature. In its original guise,  $L_b$  was defined as the depth over which  $D_b$  was constant (Goldberg and Koide, 1962, p. 426). This formulation was prompted by the observation that  $^{230}\text{Th}$  concentrations in some deep-sea cores were uniform over a limited depth interval adjacent to the sediment–water interface. Berger and Heath (1968) set forth similar ideas when they developed, on theoretical grounds, a box model for cases where mixing intensity was much larger than sedimentation rate (i.e. the deep sea). In this model, mixing was instantaneous within the ‘homogeneous layer’ and zero in the ‘historical layer’. Thus, at its inception what became known as  $L_b$  had two independent meanings. On the one hand, it was a specified region where mixing intensity was assumed to be constant; on the other hand, it was a region where properties, in particular radionuclides, were mixed uniformly (i.e.  $\partial A / \partial z = 0$ , where  $A$  is radionuclide activity and  $z$  is depth). The latter definition permits estimation of  $L_b$  from radionuclide profiles. Unfortunately, subsequent workers (e.g. Thomson et al., 2000; Smith and Rabouille, 2002) have equated the depth of penetration of a particular radionuclide

with the mixing-layer thickness. In fact, many of the  $L_b$  values in table 1 of Boudreau (1994), especially those from short-lived radionuclides (e.g.  $^{234}\text{Th}$  and  $^7\text{Be}$ ), were based on penetration depth. To make matters worse,  $L_b$  – whether it is based on identifying regions of uniform radionuclide activity or penetration depth – depends on tracer half-life (Guinasso and Schink, 1975; Boudreau, 1994). Thus, longer-lived radionuclides should, because they are subject to a wider diversity of mixing events, some of which are deeper, have a larger  $L_b$  (Smith and Rabouille, 2002). These definitional inconsistencies and the tracer half-life dependence suggest that an alternative approach to determining  $L_b$  is needed.

The vertical distribution of sediment-dwelling macrofauna is more directly relevant to estimation of  $L_b$ , and hence questions of signal alteration and preservation. If this statement is accepted, then it is likely that a surface mixing-layer thickness of 10 cm is an underestimate. That is because macrofauna are found to depths considerably greater than 10 cm in continental margin settings. For example, in a study that is broadly representative, Hines and Comtois (1985) reported the presence of animals to  $>30$  cm depth in both sands and mud. Although most animals ( $\sim 75\%$ ) resided in the upper 5 cm of the seabed, roughly 5% of the animals present were found below 20 cm (Hines and Comtois, 1985). Similar results have been reported for many other shelf and slope settings (e.g. Schaffner et al., 1987; Wheatcroft et al., 1994; Flach and Heip, 1996; Wheatcroft and Martin, 1996; Moodley et al., 1998). It is important to note that although relative numbers are low in the deep subsurface (i.e.  $>10$  cm), absolute numbers are still appreciable. For example, for a depth-integrated abundance of  $\sim 2 \times 10^4$  individuals/ $\text{m}^2$ , a value typical of many continental shelf settings (Wheatcroft et al., 1994; Wheatcroft and Martin, 1996), there are  $>1000$  animals/ $\text{m}^2$  at depths  $>10$  cm. Because deep-dwelling animals are often larger than the small, tube-dwelling animals at the sediment–water interface (e.g. Hines and Comtois, 1985), their contribution to sediment reworking and hence event-layer alteration and preservation is likely to be substantial. In

summary, the thickness of the sediment mixing layer on oxygenated continental shelves almost certainly exceeds 10 cm.

### 2.3. Event-layer thickness

Because it affects transit and dissipation time, the thickness of an event layer is of critical importance in determining its preservation potential (e.g. Larson and Rhoads, 1983; Wheatcroft, 1990). Transit time is affected by event-layer thickness because the deposition of a layer increases the sediment accumulation rate – a point discussed in more detail in the next section. Dissipation time is influenced by event-layer thickness because thin layers present less sediment volume to displace, as well as extend less far into the mixing layer. Therefore there is less of a thin bed to destroy and it spends more time in a region of high sediment displacement (this assumes there is a decrease in sediment mixing intensity with depth). Larson and Rhoads (1983) illustrated this point in a comparative study of bedding frequencies and facies in the Paleozoic (Ordovician and Devonian) of New York. There it was found that increased infaunalization (i.e. the depth and intensity of bioturbation increased) that occurred between the Ordovician and the Devonian led to a decrease in the frequency of beds < 5 cm thick.

Replicate time-lapse X-radiographs collected on the Eel margin further underscore the importance of event-layer thickness, and hence position within the surface mixing layer, in controlling preservation and alteration (Fig. 2). These X-radiographs track the fate of two flood deposits formed by the Eel River in January 1995 and 1997 (Wheatcroft and Borgeld, 2000). In the upper sequence (Fig. 2A–C) a 3 cm thick bed formed in January 1997 is destroyed over the course of a 15 month time period (i.e.  $T_d < 1.25$  yr), whereas in the lower sequence (Fig. 2D–F) a 5 cm bed that soon after formation (January 1995) was advected to 14 cm depth by the March 1995 flood persists for > 55 months. Although such differences in dissipation time could be due to variable sediment bioturbation intensity at the two sites, independent estimates of biodiffusivities

(Bentley and Nittrouer, 2004; Wheatcroft, in prep.) suggest there is little along-shelf variation in mixing intensity on the Eel shelf.

### 2.4. Sediment accumulation rate

Because the sediment–water interface is taken to be at a depth,  $z=0$ , the accumulation of sediment on the seafloor moves all material downward. Together with the mixing-layer thickness, the accumulation rate determines the transit time. Sediment accumulation rate on continental margins is a function of delivery and redistribution processes (Nittrouer and Sternberg, 1981). Because these processes operate on multiple time and space scales they impart considerable complexity to the resultant patterns of sediment accumulation rate. For example, the influence of both wind waves (e.g. Sherwood et al., 1994; Wiberg, 2000) and internal tides (e.g. Cacchione et al., 1998) on sediment resuspension is a strong function of cross-shelf position, albeit in different ways. That is, sediment resuspension from wind waves is most intense in shallow water (e.g. < 60 m), whereas internal tide resuspension is most prevalent at the shelf break. In addition to cross-isobath variability, the magnitude of sediment delivery changes along-margin. This change is due mostly to the location of rivers, but changes in shoreline geometry in relation to storm tracks and coastal lithology can also result in variable delivery.

Despite these complexities there is a growing body of data indicating that upper slope and shelf sediment accumulation rates based on  $^{210}\text{Pb}$ -geochronology fall mostly in the range of 0.1–1 cm/yr (e.g. Nittrouer et al., 1979; 1985; Carpenter et al., 1985; Frignani and Langone, 1991; Radakovitch et al., 1999; Lesueur et al., 2001). Exceptions outside this range are typically associated with large rivers, such as the Amazon, Ganges–Brahmaputra and Changjiang, where sedimentation rates may be an order of magnitude higher (Nittrouer et al., 1985; Lesueur et al., 2001). The multiple mechanisms leading to this convergence of sediment accumulation rates are uncertain, although they are likely to be related, in part, to the global eustatic sea level rise of  $\sim 0.2$  cm/yr (Douglas, 1995).

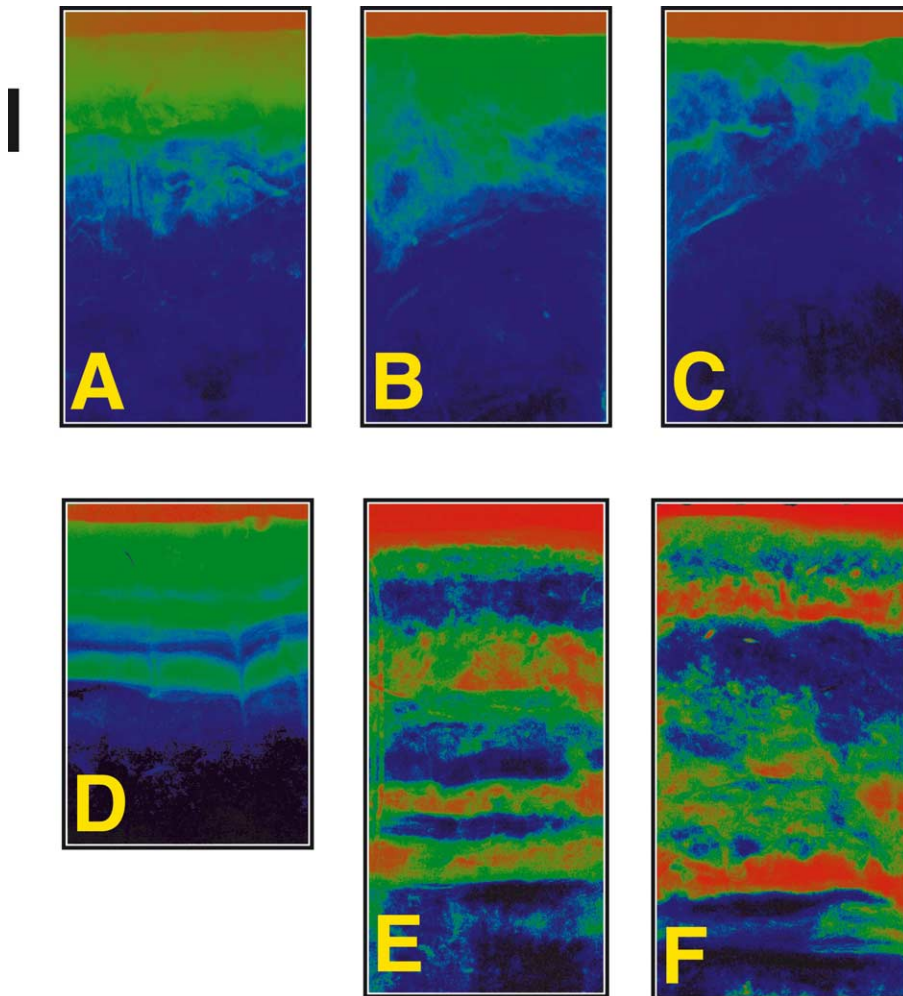


Fig. 2. Pseudocolor representations of X-radiographs taken on the Eel River shelf (1995–1999). Panels A–C document the destruction of a 3 cm thick bed formed at station C70 in January 1997 over the ensuing 10 (panel B) and 15 (panel C) months. Panels D–F document the persistence of a 5 cm thick bed formed at station S60 in January 1995 over the ensuing 14 (panel E) and 55 (panel F) months. Note that the bed depicted in panel D was advected to roughly 14 cm depth by the deposition of the March 1995 flood bed (Wheatcroft and Borgeld, 2000). Scale bar = 2 cm.

### 3. Eel River shelf

#### 3.1. Steady-state transit time and layer dissipation time

The large number of STRATAFORM studies on the Eel River shelf has yielded a comprehensive database for the investigation of event-layer alteration and preservation that is summarized in Table 1. Although few other open shelf environments have been studied in such detail, the avail-

able evidence indicates that the range of parameter values on the Eel margin is representative of many other continental shelves (again, neglecting prodelta regions proximal to major rivers). For example, the range of  $^{234}\text{Th}$ -derived biodiffusivities on the Eel shelf (Table 1) is broadly similar to that found in other shelf environments (Fig. 1), suggesting that signal dissipation times for the Eel system are likely to be representative of other continental margins. Within-sediment macrofaunal depth distributions are similar to other con-

tinental shelf settings (Wheatcroft, in prep.), and the estimated [from vertical distributions of macrofaunal (Wheatcroft, in prep.) and replicate time-lapse radiographs]  $L_b$  of  $\sim 15$  cm is not appreciably different from the worldwide average of 10 cm (Boudreau, 1994, 1998). The thickness of event layers measured over the 5 year STRATAFORM program ranges from  $<1$  to 8 cm (Wheatcroft and Borgeld, 2000), a range that is similar to the thickness of layers observed on other shelves (e.g. Aigner and Reineck, 1982). Lastly, sediment accumulation rates determined using  $^{210}\text{Pb}$ -geochronology (Sommerfield and Nittrouer, 1999) indicate a roughly seven-fold variation on the Eel shelf with an area-weighted average of 0.4 cm/yr (Table 1). This mean and range are also broadly similar to that found in many other shelf settings that are presently accumulating sediment (cf. Lesueur et al., 2001). In summary, the available data indicate that the magnitude and range of important variables (biodiffusivity, mixing-layer thickness, event-layer thickness, and sediment accumulation rate) observed on the Eel margin are broadly representative of many other shelf settings. We therefore believe that the conclusions drawn below regarding event-layer alteration and preservation have widespread generality.

Taken at face value the parameter estimates listed in Table 1 suggest that there is a low probability for transient sedimentary signals or event layers formed on the Eel shelf to be preserved intact in the stratigraphic record. That is because the transit time of, for example, a 4 cm thick event bed is 9–65 yr (accumulation rate range and  $L_b$  from Table 1). In contrast, X-radiography data from the Eel shelf indicate that *layer dissipation times* are much shorter relative to this steady-state transit time (Fig. 2). For example, the Jan-

uary 1997 flood deposit, which ranged up to 6 cm in thickness (Wheatcroft and Borgeld, 2000), was no longer recognizable in X-radiographs *anywhere* on the shelf 2.5 years later. As previously noted (e.g. Wheatcroft, 1990), however, dissipation time is a function of the signal in question, whereby signals that depend on diagnostic fabrics (e.g. layers seen in X-radiographs) are more quickly destroyed than those that are defined by some geochemical or textural attribute. Therefore, we might expect the grain-size signature of the flood deposits to persist for longer periods of time, a conjecture that is explored in the next section.

### 3.2. Grain-size dissipation time

Sediment grain size is one of the most diagnostic features of event layers in sedimentary marine environments in general, and the Eel shelf in particular. There, strong dominance by the fine silt and clay fractions ( $>95\%$ ) characterized the 1995 and 1997 flood layers (Drake, 1999; Leithold and Hope, 1999; Sommerfield and Nittrouer, 1999; Wheatcroft and Borgeld, 2000). Moreover, the grain-size distribution of both deposits was uniformly fine-grained over broad regions of the central shelf (Drake, 1999; Wheatcroft and Borgeld, 2000). Storm event layers, produced by episodes of large waves and currents that preferentially erode the finer grain sizes in the seabed (e.g. Wi-berg, 2000), generally exhibit lower percentages (5–10%) of fine silt and clay (Bentley and Nittrouer, 2004) and more regional variability, especially across isobaths.

The primary-particle (i.e. disaggregated) grain-size distribution of a layer is a conservative property that can be changed only by physically transporting particles of varying size into or out of the

Table 1  
Parameter estimates for the Eel River shelf

Parameter (units)	Average	Range	Sources
Biodiffusivity ( $\text{cm}^2/\text{yr}$ )	40	13–80	1, 2
Mixing-layer thickness (cm)	15		1, 2
Event bed thickness (cm)		$<1$ –8	3
Sedimentation rate (cm/yr)	0.4	0.2–1.4	4

1, Wheatcroft (in prep.); 2, Bentley and Nittrouer (2004); 3, Wheatcroft and Borgeld (2000); 4, Sommerfield and Nittrouer (1999).

layer, thereby changing the grain-size distribution. This transport can be accomplished by selective erosion of a subset of sizes or by addition of new sediment having a different size distribution. Bioturbation can also transport sediment of varying grain size into or out of the event layer (i.e. across the layer contacts). The rate of change in a layer's grain-size distribution due to bioturbation will be related in part to bed thickness. That is because for thick beds, a greater proportion of the mixing events will result in within-bed transport (i.e. surface-area to volume is smaller for thick beds). In an earlier communication, Drake (1999) showed that the evolution of the 1995 flood layer at a 70 m mid-shelf site (station O70) primarily involved the addition of new, coarser sediment to the top of the flood layer followed by rapid mixing of that sediment to depths of  $\sim 5$  cm by bioturbation. An inventory of grain-size fractions demonstrated that the 1995 flood layer was significantly coarsened over time by addition of medium and coarse silt, presumably from shallower regions, rather than by winnowing of the finest particles (Drake, 1999).

The dissipation time of the grain-size signal of the January 1995 and January 1997 flood layers was determined by comparing the peak wt% of the  $< 20 \mu\text{m}$  size fraction to the mean wt% of that size fraction at the site. When the peak value is less than two standard deviations above (for flood layers) or below (for storm layers) the mean of the  $< 20 \mu\text{m}$  fraction at the site, it is no longer distinguishable from ambient sediment and thus is classified as dissipated. The ambient grain-size distribution was determined from piston or box core samples. For example, at some sites (e.g. O60 and O70), piston cores provided up to 4 m of sediment and the mean and standard deviations for the  $< 20 \mu\text{m}$  fraction in over 100 samples in these cores provided an accurate and stable (i.e. there was little down-core variability) estimate of the ambient grain size. When no piston cores were available at a site, the mean and standard deviation were computed using the available box core profile samples.

As long as a flood layer remained in the surface mixing layer there was an easily measured temporal change in the peak concentration of the  $< 20$

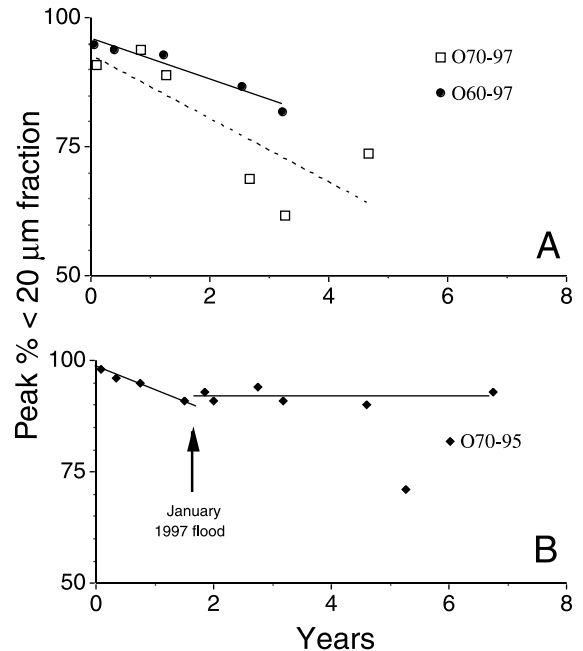


Fig. 3. Examples showing the change in the peak wt% of the  $< 20 \mu\text{m}$  fraction as a function of time (years) since formation of the January 1997 (A) and January 1995 (B) flood deposits. Sloped lines in panels A and B represent least-squares regressions, whereas the horizontal line in panel B has a slope = 0 (i.e. signal dissipation has stopped).

$\mu\text{m}$  fraction as the signal progressed toward destruction. In general, the rate of that change was inversely related to the initial thickness of the flood layer. Layers that were 1–3 cm thick and remained in the mixing layer tended to be completely destroyed or were very close to being destroyed in a few years. The grain-size signature of thicker layers, such as the January 1995 flood layer at station O70 (8 cm) or the January 1997 flood layer at station O60 (5 cm), were still recognizable at the time of the most recent sampling (October 2001). By plotting peak concentrations in an event layer as a function of time it was possible to determine the trend of the layer toward destruction (Fig. 3). To do this, a linear regression was fit to the data and the trend line extrapolated to the point representing two standard deviations above the long-term mean (e.g. 55% for O70 and 52% for O60). In this way grain-size signal dissipation time was determined for the January 1995 and January 1997 flood layers at nine separate sites

where the flood bed thickness ranged from 1.5 to 7.5 cm, and the results plotted as a function of initial layer thickness (Fig. 4). The strong positive correlation ( $r^2 = 0.95$ ) between layer thickness and signal dissipation time for the two flood beds at several widely distributed mid-shelf sites suggests that the mechanisms that are responsible for destroying the beds operate with similar intensities on a regional scale. This result is in agreement with the available bioturbation data that show little spatial variability in sediment mixing intensity on the Eel shelf (Wheatcroft, in prep.).

As expected, dissipation time for the grain-size signature of an event layer is appreciably longer than the period of time the layer is recognizable in X-radiographs. For example, the grain-size signature of a 3 cm thick flood layer persists for roughly 5 years (Fig. 4), whereas the layer itself is not recognizable after  $\sim 1.3$  years (Fig. 2D–F). Despite the longer persistence of the grain-size signal, it is still short relative to the steady-state transit time (10–100 yr). Stated a different way, the steady-state sediment accumulation rate must be quite rapid to preserve the grain-size signature of flood layers on the Eel shelf. This idea is illustrated in Fig. 5, which shows preservation potential curves for the grain-size signal as a function of initial bed thickness and subsequent sediment accumulation rate. The different curves represent a surface mixing-layer thickness of 10, 12.5 and 15

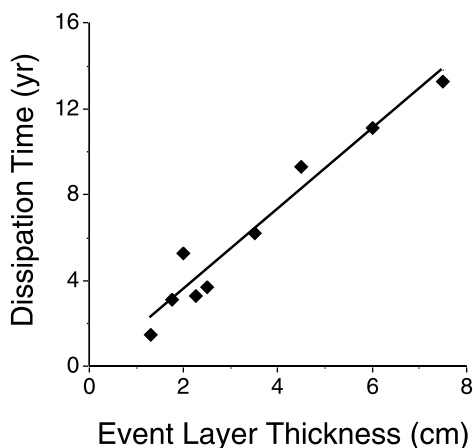


Fig. 4. Grain-size signal dissipation time (years) as a function of initial event-layer thickness (cm) for flood layers on the Eel River shelf. Line is a least-squares regression ( $r^2 = 0.95$ ).

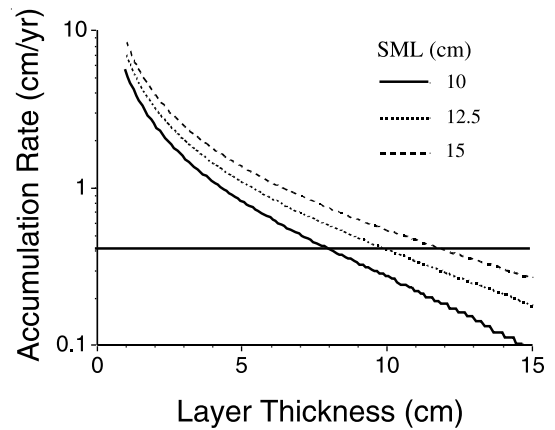


Fig. 5. Plot of accumulation rate necessary for the transit time to be less than the dissipation time for beds of varying thickness. The horizontal line represents the area-normalized sediment accumulation rate on the Eel shelf (Sommerfield and Nittrouer, 1999). See text for detail.

cm, whereas the horizontal line is the area-averaged sediment accumulation rate determined using  $^{210}\text{Pb}$ -geochronology (Sommerfield and Nittrouer, 1999). Thin layers (e.g.  $< 4$  cm) must be buried at rates that are rarely achieved on the Eel shelf, whereas beds thicker than  $\sim 7$  cm stand a chance of partial preservation. The latter is confirmed by the existence of recognizable flood layer grain-size signatures at multiple depths in several piston cores on the Eel shelf (e.g. Sommerfield et al., 2002).

### 3.3. Episodic sedimentation and transit time

The available evidence for continental shelf environments in general, and the Eel shelf in particular, indicates that: (1) there is an extremely low probability of preserving intact layers (i.e. those recognizable in X-radiographs), and (2) grain-size signatures of all but the thickest layers (i.e.  $> 7$  cm) will be destroyed as well (Fig. 5). The reason for this situation is that steady-state transit time is considerably longer than dissipation time. These conclusions are incomplete, however, because they neglect the important role of episodic sedimentation events. Short-lived events, such as the January 1995 and 1997 oceanic floods, can deposit several centimeters of sediment over hours to days (Wheatcroft and Borgeld, 2000), thereby

rapidly advecting layers and signals through the surface mixing layer. Transit times can decrease by more than an order of magnitude due to episodic sedimentation events, a key factor in determining preservation potential.

### 3.4. Preserved flood layers

Further appreciation of the role of episodic sedimentation in event-layer preservation is obtained by examining stratigraphic sequences formed during the past 50 years on the Eel shelf. During that time period the Eel River has had major floods roughly once a decade (1955, 1964, 1974, 1986, 1995 and 1997). Of these floods, the 1964 and 1955 events had the highest discharge ( $2.1 \times 10^4$  and  $1.5 \times 10^4$  m<sup>3</sup>/s, respectively), whereas the others were roughly equivalent ( $\sim 1 \times 10^4$  m<sup>3</sup>/s). If all things were equal, then one might expect that the largest events would have left the most obvious record, whereas evidence for the others would be of lesser extent and approximately equal in magnitude.

To test this conjecture, X-radiographs taken from more than 100 box and Kasten cores collected in September 1995 (cruise W9509A) were used to form a comprehensive picture of strata in the upper 50 cm of the Eel shelf. If analysis is restricted to the sites along the 70 m isobath, then areas where physical bedding is present extend along-shelf for roughly 50 km. Intact layers, typically 1–5 cm thick beds with relatively sharp lower contacts and biologically disturbed upper contacts, were also distributed fairly evenly over the upper 40 cm of the sediment column (Fig. 6A). [Note that essentially identical results are obtained if data from other sites or cruises are used in the analysis.] Two pieces of information were used to assign ages to the layers preserved in the seabed. First, the thickness of the January and March 1995 flood deposits (Wheatcroft and Borgeld, 2000) was subtracted from the depth data. Second, sediment accumulation rates determined using <sup>210</sup>Pb-geochronology (Sommerfield and Nittrouer, 1999) for sites along the 70 m isobath were used to convert to elapsed time. Although there is some uncertainty associated with <sup>210</sup>Pb-derived accumulation rates, the resultant age distributions

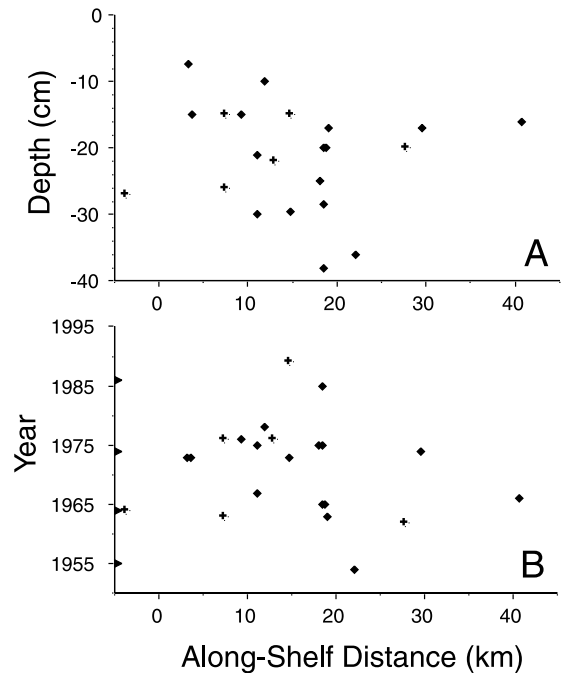


Fig. 6. Depth (A) and age (B) of physical layering observed in X-radiographs taken along the 70 m isobath on the Eel River shelf in September 1995 (cruise W9509A). The along-shelf position of the river mouth is referenced to zero on the abscissa. Solid diamonds denote obvious layers (1–4 cm thick), whereas pluses represent vague contacts. See text for detail.

provide an unambiguous message. Namely that in contrast to the roughly even distribution of layers within the sediment column, the corresponding ages of the layers cluster into essentially two groups that are close in age to the 1974 and 1964 floods (Fig. 6B).

Given its extreme magnitude, one might expect that the 1964 flood layer would remain recognizable in the sediment column. However, it is surprising that the 1974 flood layer is readily apparent, especially considering that there is little record of the 1955 and 1986 floods, which were of equal or greater magnitude. It is possible that the thickness of the flood layers does not scale directly with discharge magnitude. Thus, the 1955 and 1986 floods may not have produced sizeable deposits in the first place. Although it is true that flood magnitude is not a perfect predictor of flood bed thickness (Wheatcroft and Bor-

geld, 2000), it is highly unlikely that a flood deposit was not produced in 1955 and 1986. That is because even floods of substantially lesser magnitude, for example the March 1995 event that had a peak discharge  $< 6 \times 10^3 \text{ m}^3/\text{s}$ , have been observed to result in relatively thick flood deposits on the mid-shelf region of the Eel margin (Wheatcroft and Borgeld, 2000).

The alternative explanation is that variable transit times following the floods cause differences in preservation. To illustrate that point, the cumulative suspended sediment load of the Eel River was computed for 3 years following each flood using a previously established rating curve (Wheatcroft and Borgeld, 2000). The results indicate three different patterns (Fig. 7). First, following the 1974 and 1995 floods, subsequent high-discharge events during the same rainy season introduced significant amounts of fine-grained sediment into the coastal ocean where it likely buried the initial event bed. This deposition was followed by additional high-discharge events in subsequent years so that by the end of the 3 year period substantial accumulation had likely occurred (we know it did following the January 1995 flood). Second, the 1964 and 1997 flood layers were subject to a year's worth of biological reworking before subsequent high discharge. The 1964 flood bed, perhaps because of its large thickness, was still recognizable, whereas the 1997 event was completely destroyed. Third, the 1986 flood was

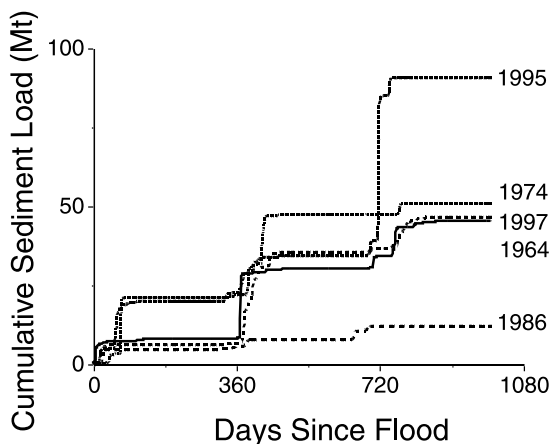


Fig. 7. Cumulative suspended-sediment load as a function of days since major floods in 1964, 1974, 1986, 1995 and 1997.

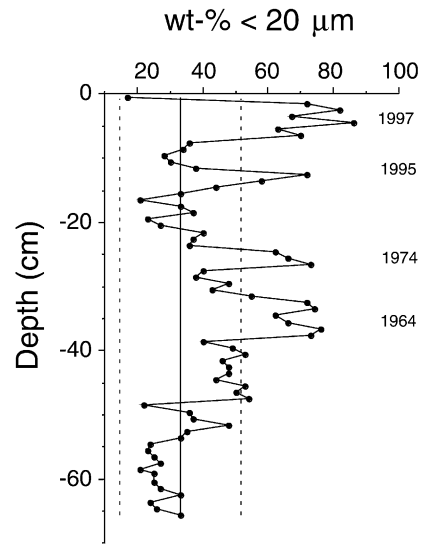


Fig. 8. Percentage of particles  $< 20 \mu\text{m}$  as a function of depth in the upper 60 cm of a piston core collected at station O60 in August 1999. The solid and dashed lines are the average,  $\pm$  two standard deviation wt% of particles  $< 20 \mu\text{m}$  determined over the full depth of the piston core. See text for detail.

followed by a multi-year period of moderate to low discharge. Thus, the 1986 flood deposit remained in the surface mixing layer for a long enough period that it was destroyed.

Insight obtained from the X-radiographs and the river discharge history can be used to predict the fate of the flood-layer grain-size signal over the past 50 years. In particular, the 1974 flood deposit is expected to be more readily recognizable (i.e. be less degraded) than the 1986 flood bed. Because the sub-cores used for the grain-size sampling were typically  $< 15 \text{ cm}$  long, and therefore too short to reach the earlier flood beds (buried  $> 20 \text{ cm}$ ), the available data set to test the above prediction is small. However, at two sites for which deep grain-size samples exist, the data match our expectations. Stations O60 and O70, located roughly 20 km north of the Eel River mouth near the center of mass of the 1995 and 1997 flood deposits (Wheatcroft and Borgeld, 2000), reveal a detailed record of grain-size fluctuations over the past millennium (Sommerfield et al., 2002). Focusing on the upper 60 cm of the sediment column at O60 there are four

regions of fine-grained sediment with  $< 20 \mu\text{m}$  peak values appreciably greater than two standard deviations (Fig. 8). On the basis of age control data from  $^{210}\text{Pb}$  and  $^{137}\text{Cs}$  (Sommerfield and Nittrouer, 1999; Sommerfield et al., 2002), as well as information on the thickness of 1995 and 1997 flood deposits (Wheatcroft and Borgeld, 2000), we can estimate the approximate age of each fine-grained zone. Thus, the upper two layers correspond to the 1997 and 1995 flood deposits, whereas the lower two layers (at  $\sim 26$  and  $35$  cm depth) correspond to the 1974 and 1964 flood deposits, respectively (Fig. 8). The region between 15 and 23 cm is roughly where the 1986 flood bed should reside, but instead there is a zone of relatively coarse sediment. The likely reason for this pattern is that the transit time of the 1986 flood bed was greater than its dissipation time; hence its grain-size signature was destroyed completely.

#### 4. Summary

Unlike the deep sea, where the balance between sedimentation and bioturbation is strongly in favor of the latter, the two terms display approximate parity on continental shelves. Therefore, circumstances that shift the system out of balance will have an important impact on the alteration and preservation of transient sedimentary signals and event layers. In this paper, it has been shown that on many fine-grained, upper continental margins around the world bioturbation intensity is high ( $10\text{--}100 \text{ cm}^2/\text{yr}$ ), hence signal dissipation times are typically short ( $< 10$  yr). Conversely, despite relatively high sediment accumulation rates ( $0.1\text{--}1 \text{ cm}/\text{yr}$ ) compared to the deep sea, transit times through the mixing layer are long (decades to centuries). Therefore, under conditions of steady sediment accumulation, event layers and their sedimentary signals (e.g. grain size) stand little chance of preservation on most continental margins. Sediment accumulation on continental margins is not steady, however. Instead, short-lived events lead to episodic sedimentation that can instantaneously advect layers or signals through the surface mixing layer, thereby leading to preservation. Explicit recognition of the

importance of episodic sedimentation is key to deciphering the stratigraphic record of continental margins.

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