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Continental Margin Sedimentation: From Sediment Transport to Sequence Stratigraphy

EDITED BY

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Preface

'*If we don't learn from history, we're doomed to relive it*'. Unlike human history, most of the events that form the record of Earth history are out of our control. However, we may still learn from them and prepare ourselves for future environmental events (e.g. storm surges, sea-level rise). Understanding continental-margin sedimentation is important for many reasons, as diverse as finding natural resources and maintaining safe navigation. In addition, the stratigraphy that results from margin sedimentation provides an extremely rich record of Earth history – including the natural processes and, more recently, the human impacts operating both on land and in the sea. Unfortunately, we cannot learn from this record until we can read it. Large portions of the following text have this purpose, and collectively provide a unique contribution to the continuing legacy of studies to unravel the secrets of margin stratigraphy.

GOALS AND ORGANIZATION

This volume is an outgrowth of the STRATAFORM programme (STRATA FORmation on Margins) funded by the US Office of Naval Research (ONR). Consequently, the goals and organization of the volume reflect those of STRATAFORM. In that programme, we set out to integrate across three major domains in our geological and geophysical examination of continental-margin sedimentation: environments, from inner shelves to distal slopes; processes, from discrete events to the long-term preserved stratigraphy; and techniques, from observations to modelling. Pieces of this integrated approach have been undertaken previously, but STRATAFORM broke new ground in its holistic investigation across such a complex matrix.

Construction of this volume has followed a similar pattern, and has experienced the same challenges. First, continental-margin sedimentation is an extremely broad field and we have had to define workable boundaries, so the scope of the volume is tractable. Future investigators and funding agencies are offered this result as a blue print for studies of margin sedimentation in other environments. Second, participants have had to think beyond their individual disciplinary specialities, so integration of results could be balanced and fair. This has not always been easy, but the consensus of the group has made it happen (and ONR programme manager, Dr Joseph Kravitz, was persuasive).

Finally, the actual mechanics of merging many people and their diverse contributions has probably been the toughest challenge of all. Rather than creating a 'project volume' with a pot-pourri of loosely related papers, we have envisioned a written document that is comprehensive and presents continua of ideas across the spectra of the research. For independent-minded scientists experienced in writing research papers in their areas of speciality, a contiguous blend of summary papers with finite boundaries and required contents is a challenge. However, we succeeded, and the results are presented in the papers that follow.

THANKS

There are many people to thank for the scientific research, operations, leadership and support that have carried the STRATAFORM programme from its inception through the completion of this volume. The research was undertaken first, and we are indebted to the legions of investigators, students and technicians at participating institutions who were involved in STRATAFORM cruises, experiments, and programming. The ONR was the funding agency, and we appreciate its commitment to this extended research effort. Among the ranks of ONR managers the greatest supporter is honoured below.

The authors created the text and the editors helped make it better. Great thanks go to the lead authors, who stuck to the task long after the programme funding ended. A diverse group of reviewers provided constructive advice, and included people outside and inside STRATAFORM, as well as the editors. Each of these receives our appreciation, and they are listed below.

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DEDICATION

STRATAFORM would not have been possible without the stalwart support of Dr Joseph Kravitz, and we express our recognition and appreciation by dedicating this volume to him.

Joe is Pennsylvania born and raised. Educated at Syracuse and George Washington Universities, he has worked his way through life with drive and determination. He managed a number of programmes over the years at ONR and NOAA, and his last was his most ambitious in terms of scientific goals and scope.

Joe provided stern, but caring leadership. He nurtured investigators in a way that allowed them to employ their best creative talents. Any successes that came from STRATAFORM were made possible by Joe. It was a special period in the professional lives of all those involved. The individuals and the science benefited from the leadership and vision he brought to the programme. Good deeds deserve recognition, and this volume is our gift, and our thanks, to Joe Kravitz.

Chuck Nittrouer, Jamie Austin, Mike Field, James Syvitski and Pat Wiberg (as co-editors)

On behalf of all scientists involved in this volume and in STRATAFORM

Dr Joseph Kravitz along the bank of the Eel River, near its mouth. (Photograph courtesy of Rob Wheatcroft.)

Writing a Rosetta stone: insights into continental-margin sedimentary processes and strata

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ABSTRACT

Continental margins are valuable for many reasons, including the rich record of Earth history that they contain. A comprehensive understanding about the fate of fluvial sediment requires knowledge that transcends time-scales ranging from particle transport to deep burial. Insights are presented for margins in general, with a focus on a tectonically active margin (northern California) and a passive margin (New Jersey). Formation of continental-margin strata begins with sediment delivery to the seabed. Physical and biological reworking alters this sediment before it is preserved by burial, and has an impact upon its dispersal to more distal locations. The seabed develops strength as it consolidates, but failure can occur and lead to sediment redistribution through high-concentration gravity flows. Processes ranging from sediment delivery to gravity flows create morphological features that give shape to continental-margin surfaces. With burial, these surfaces may become seismic reflectors, which are observed in the subsurface as stratigraphy and are used to interpret the history of formative processes. Observations document sedimentary processes and strata on a particular margin, but numerical models and laboratory experimentation are necessary to provide a quantitative basis for extrapolation of these processes and strata in time and space.

Keywords Continental margin, continental shelf, continental slope, sedimentation, stratigraphy.

INTRODUCTION

The history of processes influencing the Earth is recorded in many ways. The sedimentary strata forming around the fringes of the ocean contain an especially rich record of Earth history, because they are impacted by a complex array of factors within the atmosphere (e.g. climate), the lithosphere (e.g. mountain building) and the biosphere (e.g. carbon fluxes).

Events that occur in coastal oceans and adjacent land surfaces have great impacts on humans, because most people live near the sea and depend on the bountiful resources formed or found there. Landslides, river floods, storm surges and tsunamis are examples of processes that can have sudden and catastrophic consequences for coastal regions. Other important processes have characteristic time-scales that are longer and the processes are somewhat more predictable; e.g. sea-level rise or fall, crustal uplift or subsidence, sediment accumulation or erosion. The confluence of terrestrial and marine processes occurs in the physiographical region known as the **continental margin**, extending

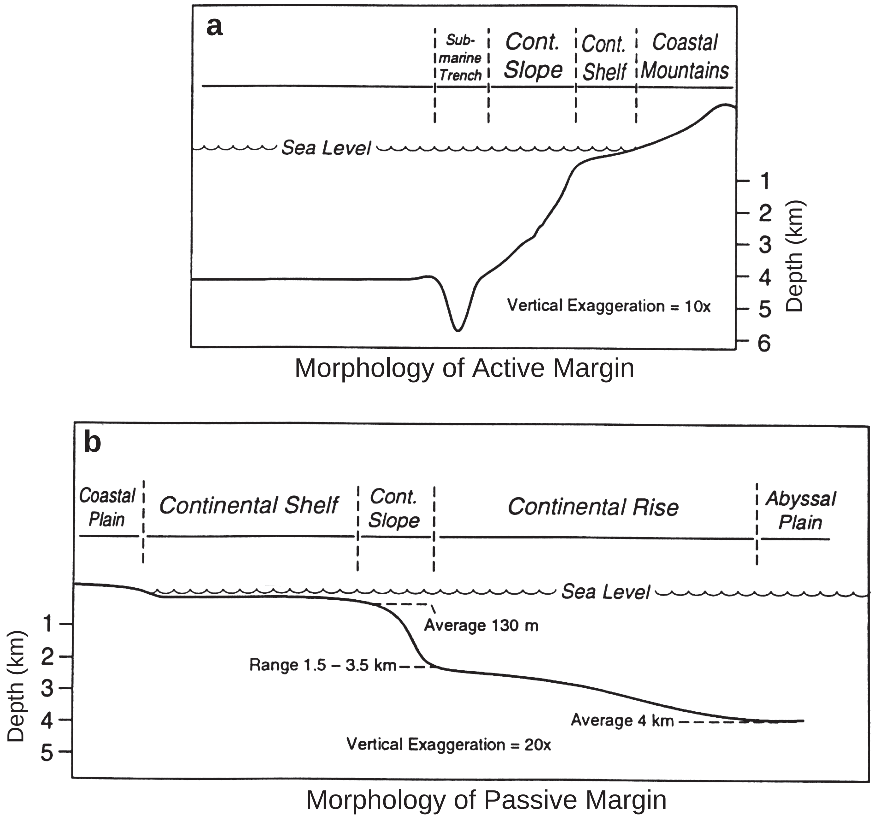

Fig. 1 Morphology of continental margins. (a) Typical morphology for a tectonically active continental margin, where oceanic and continental plates collide and subduction occurs. (b) A passive margin, where the continental and oceanic crust moves in concert. Significant distinctions include the presence of a coastal mountain range, narrow and steep continental shelf, and submarine trench (which can be filled with sediment) for the active margin. The passive margin is characterized by a coastal plain, broad continental shelf, and continental rise. (From Brink *et al.*, 1992.)

from coastal plains and coastal mountain ranges, across shorelines, to shallow continental shelves, and steeper and deeper continental slopes and rises (Fig. 1).

The interplay of terrestrial and marine processes on continental margins creates a complex mixture of stratigraphic signals in the sediments that accumulate there. This region of Earth, however, has the largest sediment accumulation rates, which create the potential for resolving diverse signals imparted over a range of time-scales (e.g. signals of river floods, and of sea-level change). Not only are the continental margins diverse and complex, but they are also very energetic. Waves, tides and currents are strong here, and provide the means to erase as well as form sedimentary records. Continental-margin stratigraphy represents a great archive of Earth history, but the challenges of reading it are also great, and require a fundamental understanding (a Rosetta stone) for translating stratigraphic character into a record of sedimentary processes.

The goal of this introductory paper is to distill the knowledge presented in the following papers of this volume, and integrate the recent insights that have been developed regarding sedimentary processes on continental margins, their impacts on strata formation, and how the preserved strata can be used to unravel Earth history. In contrast to the following papers that isolate topics, this paper highlights the linkages that come from a multi-dimensional perspective of margins. This is a summary of continental-margin sedimentation: from sediment transport to sequence stratigraphy.

THE BOUNDARY CONDITIONS

The full range of topics relevant to continental-margin sedimentation is extensive. In high latitudes, present or past glacial processes and sediments have a strong impact on sedimentation. In some low-latitude settings, biogenic carbonate sediments and their unique mechanisms of formation (e.g. coral reefs) dominate sedimentation. However, from polar to tropical environments, rivers can be the overwhelming sediment source for strata formation on continental margins. Margins affected by fluvial sediment, therefore, are the focus of this discussion.

Rivers add to the complexity of continental-margin processes through their discharge of freshwater and solutes. Rivers are also the dominant suppliers of particulate material from land to sea (globally ~85–95% is fluvial sediment; Milliman & Meade, 1983; Syvitski *et al.*, 2003). The largest rivers create extensive deposits near their mouths (e.g. Amazon, Ganges–Brahmaputra, Mississippi), but the combined discharges of moderate and small rivers (especially from coastal mountain ranges) dominate global sediment supply (Milliman & Meade, 1983; Milliman & Syvitski, 1992) and, therefore, are important to the creation of continental-margin stratigraphy.

Fluvial sedimentation on tectonically **active** and **passive** margins (Fig. 1) can now be examined over time-scales ranging from wave periods of seconds, to the stratigraphy formed and preserved over 10^7 years. Studies can span this broad range of time-scales with new rigour because numerous instruments (e.g. acoustic sensors for particle transport) and techniques (e.g. short-lived radioisotopes for seabed dynamics) have been developed recently to provide insights into important sedimentary processes. Similarly, significant advances have been made in seismic tools (e.g. CHIRP reflection profiling, multibeam swath mapping) that allow better resolution of stratigraphic surfaces. Recent advances in numerical modelling and laboratory simulations provide the opportunity quantitatively to span the temporal gap between processes operating over seconds and stratigraphy developed over millions of years.

The continental shelf and slope are the primary targets of this discussion because they are among the most dynamic environments on Earth, and record a wealth of information about environmental processes. At the boundary between land and ocean, they are impacted by energetic events characteristic of both regions (e.g. river floods, storm waves). On longer time-scales as sea level rises and falls, shelves are flooded and exposed, and slopes switch from sediment starvation to become recipients of all fluvial sediment. The boundaries between subaerial and submarine settings (i.e. the **shoreline**) and between shelf and slope (i.e. the **shelf break)** represent two dominant environmental and physiographical transitions on Earth. The transfers of sediment across these boundaries are also of special interest, because the particles on each side experience much different processes and therefore different fates. For example, on active margins, sediment crossing the shelf break can be subducted, but sediment remaining on the shelf cannot.

In this paper, fluvial sediment supply is taken as a source function on the landward side, without extensive discussion about the myriad processes occurring on land. On the seaward side, the evaluation of sedimentary processes and their effects on the formation and preservation of strata stops short of the continental rise, and the submarine fans formed there. The goal is a general understanding of sedimentary processes and stratigraphy on the continental shelf and slope, and the complex interrelationships are highlighted through two common study areas.

THE COMMON THREADS

The discussions within this paper cascade from short to long time-scales, from surficial layers of the seabed to those buried deeply within, and from shallow to deep water. Continuity in discussions is provided through examples from two diverse continental margins, which have been studied intensely throughout the STRATAFORM programme (STRATA FORmation on Margins; Nittrouer, 1999). The continental margin of northern California, near the Eel River (between Cape Mendocino and Trinidad Head; Fig. 2), is undergoing active tectonic motions and experiencing a range of associated sedimentary processes. In contrast, the margin of New Jersey (Fig. 3) is moving passively in concert with the adjacent continental and oceanic crust, and a distinctly different history of sedimentary processes is recorded.

Eel River (California) continental margin

The Eel basin is typical for rivers draining tectonically active continental margins. It is small (~9000 km^2), mountainous (reaching elevations > 2000 m), and composed of intensely deformed and easily erodible sedimentary rocks (Franciscan mélange and other marine deposits). These conditions lead to frequent subaerial landslides, especially because the high elevations cause orographic effects that intensify

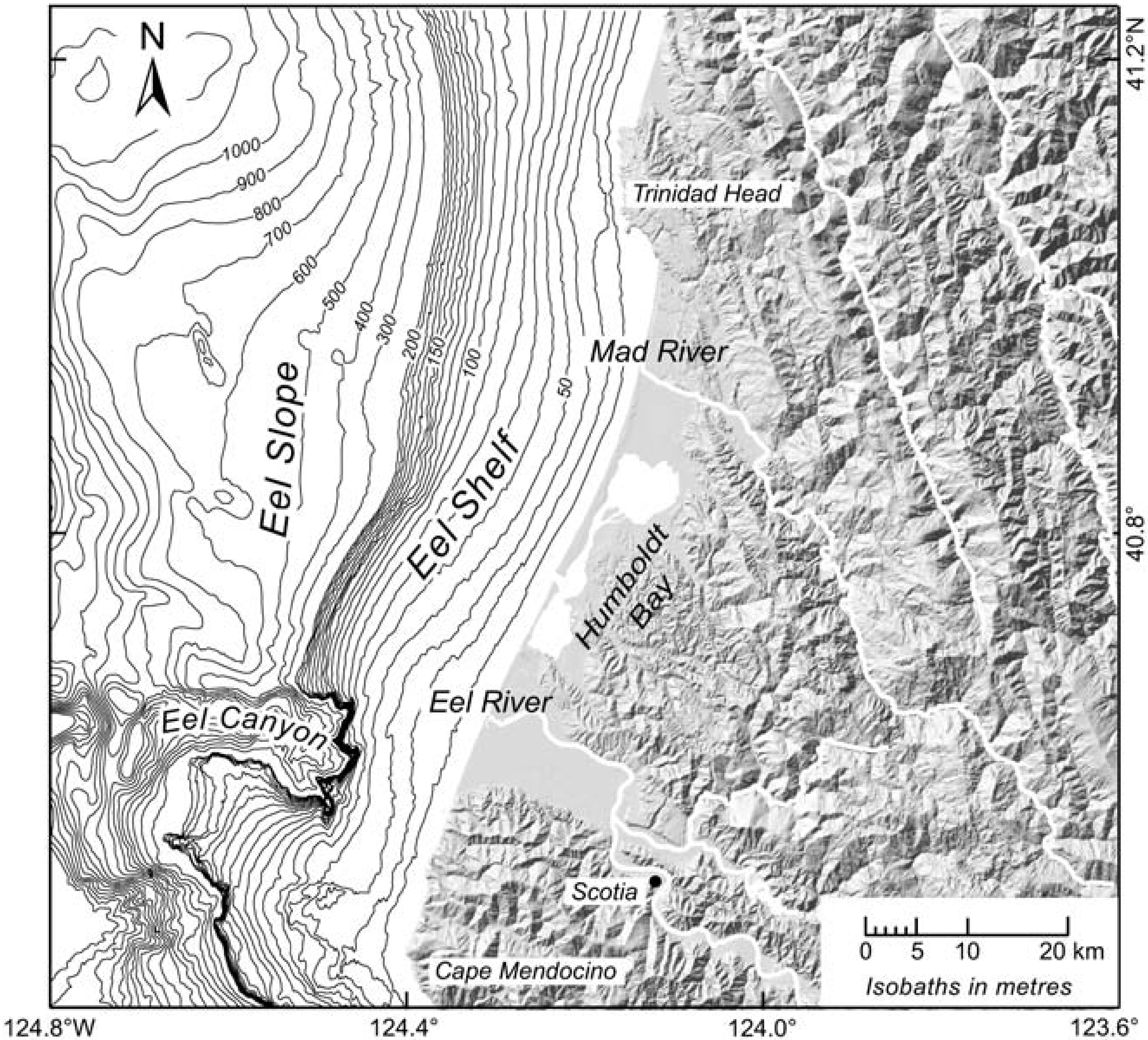

Fig. 2 The study area for the Eel margin, stretching from Cape Mendocino to Trinidad Head. The Eel River supplies an order of magnitude more sediment (~2×10^7 t yr^{-1}) than the Mad River. Below the town of Scotia (location of the lowermost river gauge), the river mouth has a small delta plain and most Eel River sediment escapes to the ocean. The shelf break is in a water depth of ~150 m, and is indented by Eel Canyon west of the river mouth. (Modified from Sommerfield *et al.*, this volume.)

rainfall from winter storm systems moving eastward off the Pacific. The annual **sediment yield** (mass discharge per basin area) is large (~2000 t km^{-2}), and although interannual discharge is highly variable, the mean value of sediment supplied to the ocean is estimated to be ~2×10^7 t yr^{-1} (Brown & Ritter, 1971; Wheatcroft *et al.*, 1997; Sommerfield & Nittrouer, 1999; Syvitski & Morehead, 1999). The grain size of the combined bedload and suspended load is relatively coarse (~25% sand; Brown & Ritter, 1971), due to the mountainous terrain and short length of the river (~200 km). Its size and orientation (generally parallel to the coastline) cause the entire basin to receive precipitation simultaneously during

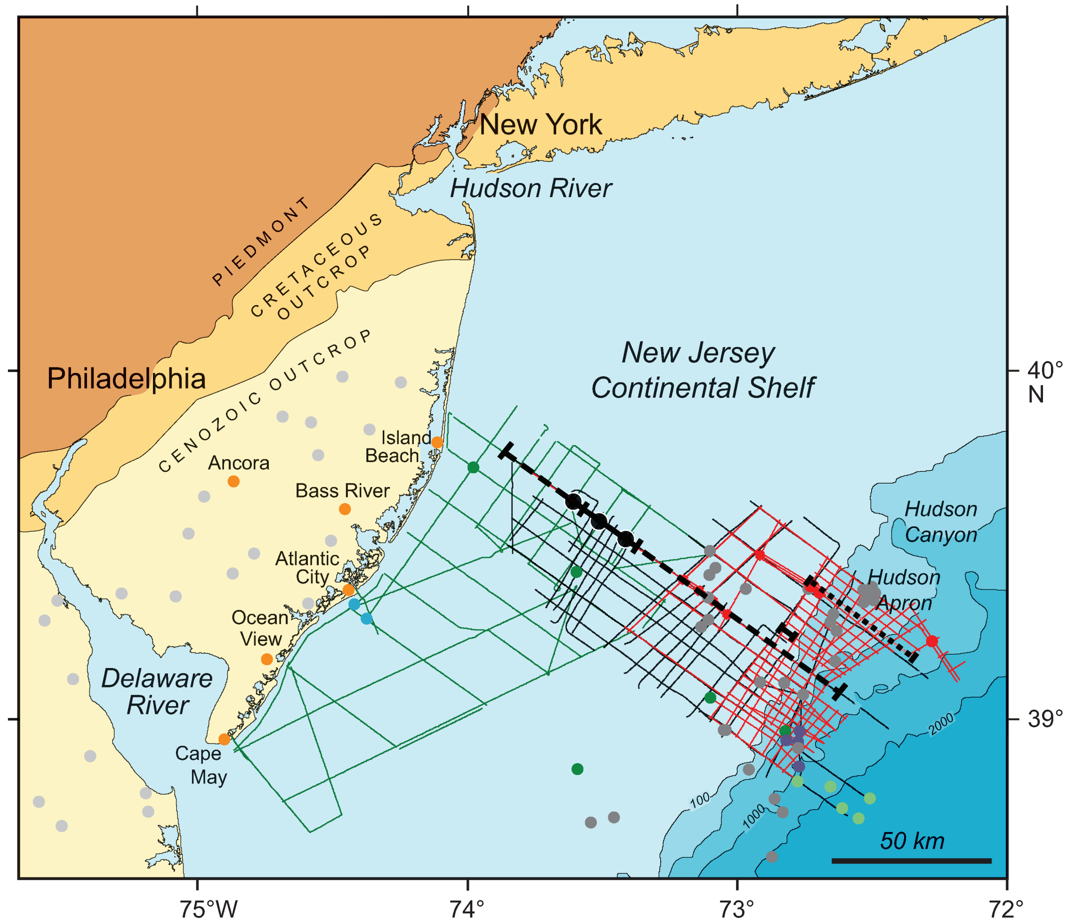

Fig. 3 The study area for the New Jersey margin, stretching between the mouths of the Delaware and Hudson Rivers. Most sediment is trapped in the estuaries at the river mouths and behind the New Jersey coastal barriers. The importance of the New Jersey margin is found in the underlying stratigraphy, which is a classic representation of passive-margin evolution. Some of the data used in this volume were collected at locations shown by the dots (drill sites) and lines (seismic profiles). Isobaths are metres. The shelf break is at ~100 m, and is indented by multiple submarine canyons including Hudson Canyon. (Modified from Mountain *et al.*, this volume.)

storms, and therefore the river discharge increases rapidly.

For the Eel River, major rainfall events commonly lead to episodic floods of the basin. Fluvial sediment discharge increases exponentially with water discharge (Syvitski *et al.*, 2000), and large floods dominate intra-annual and interannual variability of sediment transport. The mouth of the Eel River has no estuary and a very small delta plain (Fig. 2), so periods of sediment transport in the river become periods of sediment supply to the ocean. Most supply occurs during the winter (~90%; Brown & Ritter, 1971), and, for the past ~50 yr, decadal floods during the winter have had a significant impact on the river geomorphology and ocean sedimentation. The largest flood during this period was in 1964 and, more recently, a couplet of significant floods occurred in 1995 and 1997 (Wheatcroft & Borgeld, 2000).

Low-pressure cyclonic systems move eastward from the Pacific Ocean toward the west coast of North America. Commonly there is an asymmetry, such that the steepest pressure gradients are associated with the leading edges of the systems. Therefore, initial winds are strong, from the south or south-west, and Coriolis and frictional forces cause **Ekman transport** of surface water eastward toward the coast. Water elevations rise there, creating a seaward-sloping water surface that produces northward **barotropic** flow of shelf water. The eastward component of surface flows also causes **downwelling** and seaward bottom flows. The strong winds from the south and south-west create large waves approaching from those directions (as high as 10 m or more; Wiberg, 2000), and result in northward **alongshore transport** in the surf zone. This transport creates coastal landforms (e.g. spits) that direct the Eel River plume northward (Geyer *et al.*, 2000). As the low-pressure systems pass, the trailing portions of the cyclonic systems often cause winds to reverse and blow from the north.

An important aspect of sedimentation on the Eel margin is the rapid response of the Eel River to rainfall, and the occurrence of river floods during energetic ocean storms (see Hill *et al.*, this volume, pp. 49–99). These types of events can be described as **wet storms**, during which large fluvial discharges reach the ocean when sediment transport processes are strong. The river plume, coastal current, and wind waves during these periods are important dynamical processes for sediment dispersal on the Eel margin, but they are not the only processes. Energetic ocean conditions also occur without river floods (e.g. large swell waves), and these are described as **dry storms**. Tidal forcing is important on the Eel margin. A tidal range of ~2 m causes current speeds ~50 cm s^{-1} oriented primarily alongshelf. The tidal prism flowing in and out of Humboldt Bay (Fig. 2) influences shelf circulation near its mouth (Geyer *et al.*, 2000). In addition, tidal forcing in deeper water initiates internal waves that maintain suspended sediment near and below the shelf break (McPhee-Shaw *et al.*, 2004).

Sediment from the Eel River and the adjacent Mad River (~10% of the Eel discharge) is supplied to a relatively narrow continental shelf surface (~20 km wide) constrained by promontories: Cape Mendocino to the south and Trinidad Head to the north (Fig. 2). The shelf break is at ~150 m water depth and Eel Canyon incises the shelf surface just west of the river mouth. The morphological elements of the surface (e.g. narrow and steep shelf) and subsurface (e.g. structural folds and faults) are largely the result of tectonic activity. The present Eel margin is part of the larger Eel River Basin (Clarke, 1987, 1992; Orange, 1999), which became a forearc basin in the Miocene and accumulated > 3000 m of marine sediment by the middle Pleistocene (~1 Ma). At that time, the northward migration of the Mendocino Triple Junction and subduction associated with the Gorda Plate initiated modern tectonic conditions. The Gorda and North American plates are converging at ~3 cm yr^{-1} (DeMets *et al.*, 1990), and create localized uplift and subsidence with a WNW–ESE orientation. This is the tectonic framework on which Eel margin sedimentation has been imprinted for the past million years.

New Jersey continental margin

The modern Hudson and Delaware Rivers bracket the New Jersey continental margin (Fig. 3), but very little sediment escapes from the estuaries at the river mouths or from behind the New Jersey barrier coastline. New Jersey is a classic example of a passive margin, and its special value comes from the stratigraphic record buried beneath its surface. The margin began to form as the Atlantic Ocean opened with rifting in the Late Triassic and spreading in the Early Jurassic (Grow & Sheridan, 1988). A range of processes typical of passive margins caused subsidence of the margin, and created space that could be filled with sediment (i.e. **accommodation space**). Through the Cretaceous, it was fringed by a barrier reef, but it became a carbonate ramp in the early Tertiary (Jansa, 1981; Poag, 1985) due to continued subsidence and sediment starvation.

Sediment accumulation rates dramatically increased (to ~10–100 m Myr^{-1}) in the late Oligocene and early Miocene, due to tectonic activity in the source area that increased fluvial sediment supply to the margin (Poag, 1985; Poag & Sevon, 1989). The resulting stratigraphic record has been examined by many seismic and drilling investigations (Mountain *et al.*, this volume, pp. 381–458). Cycles of sea-level fluctuation are recorded by repetitive sequences of strata: a basal layer of **glauconite** sand (an authigenic mineral indicating negligible sedimentation) overlain by silt, which coarsens upward into quartz sand (Owens & Gohn, 1985; Sugarman & Miller, 1997). These sequences reflect sea-level rise, followed by seaward migration of shelf and nearshore sedimentary environments. During the Miocene, most of the sediment accumulation resulted from migration on the shelf of morphological structures known as **clinoforms** (Greenlee *et al.*, 1992). These have a shallow, gently dipping **topset** region of upward growth and, farther offshore, a steeper **foreset** region of seaward growth (see below). The extent of sea-level fluctuations during the Miocene is controversial, but probably was subdued (20–30 m;

Kominz *et al.*, 1998; Miller *et al.*, 1998) relative to fluctuations that followed (> 100 m) in the Pleistocene.

Glacial erosion in the source area was largely responsible for supplying sediment to the marine environment during the Pleistocene. Earlier sedimentation had built a wide shelf with a gentle gradient, but margin subsidence had slowed and was producing little new accommodation space on the inner shelf. During lowered sea level, glacial outwash streams incised the shelf and icebergs even scraped the surface (Duncan & Goff, 2001; Fulthorpe & Austin, 2004). Generally, sediment accumulation was displaced seaward to the outer shelf and upper slope, dramatically changing the sedimentation regime (Greenlee *et al.*, 1988, 1992; Mountain *et al.*, this volume, pp. 381–458). Clinoforms were active there, and the inflection in their bathymetric gradient became the shelf break. Sedimentation on the continental slope increased significantly, which caused seaward growth of the shelf break to its present position > 100 km from shore. The slope also grew seaward, but the influx of sediment initiated localized erosional processes. Miocene **submarine canyons** and smaller erosional features (**gullies**) were buried or reactivated by the substantial sediment supply to the relatively steep slope (Mountain, 1987; Pratson *et al.*, 1994). The long history of the New Jersey margin provides an opportunity to observe how a diverse range of sedimentary processes impacts the preserved strata on a passive margin.

SEDIMENT DELIVERY

Detailed aspects of sediment delivery on continental margins have been addressed in this volume by Hill *et al.* (pp. 49–99) and Syvitski *et al.* (pp. 459–529).

General considerations

The first step in the formation of continental-margin strata is sediment delivery. The timing and content of fluvial discharge depend on many factors, such as basin character, weather, glaciation and groundwater flow (Beschta, 1987), which can be observed and modelled. Commonly, a **rating curve** is developed to relate sediment flux to river discharge (Cohn, 1995; Syvitski *et al.*, 2000). The observations needed to generate a rating curve are confounded by difficulty in making measurements over a range of flow conditions – especially during large flood events, which are important periods because much sediment is transported (Wheatcroft *et al.*, 1997). Other difficulties are imposed by changes in the curve that occur when the river basin is altered naturally (e.g. landslides) or unnaturally (e.g. land use). Asymmetry in sediment discharge is commonly associated with rise and fall of river stage, and can cause a **hysteresis** whereby different sediment fluxes occur for the same discharge (Brown & Ritter, 1971; Meade *et al.*, 1990). Over longer time-scales of climatic and sea-level changes, adjustments to the snow pack and basin size have an impact upon the timing and amount of discharge (Mulder & Syvitski, 1996). Fluctuations in regional precipitation patterns also can modify the shape of the river hydrograph and the dominance of sustained flows or episodic floods, which are conditions that affect sediment transport substantially. For example, strengthening of the monsoonal regime in the early Holocene caused the Ganges–Brahmaputra system to have more than twice its present sediment load (Goodbred & Kuehl, 2000).

Rivers supply a range of grain sizes to the ocean. Sediment in suspension (mostly silt and clay, i.e. < 64 μm) generally represents ~90% of the discharge, and the remainder is bedload (almost entirely sand; Meade, 1996). Early recognition of patterns for modern sediment distribution on continental margins provided suggestions about delivery mechanisms to the seabed. Commonly, sand is concentrated on the inner shelf, and silt and clay are found farther seaward. Potential mechanisms for dispersal of the fine sediment are:

1 a land source with high concentrations of mud that diffuse seaward through wave and tidal reworking (Swift, 1970);
2 erosion of nearshore fluvial sediment by physical processes that intensify toward shore, and advection by currents to deeper, quiescent settings (McCave, 1972);
3 resuspension of sediment in concentrations turbid enough to flow seaward under the influence of gravity (Moore, 1969).

All three mechanisms (and others) are possible, with one or another dominating under particular conditions.

The first step in sediment delivery is for particles to leave the river plume. Sand settles rapidly and reaches the seabed near the river mouth. Silts and clays sink from surface plumes within a few kilometres of the river mouth (Drake, 1976). Individual silt and clay particles settle too slowly to explain this latter observation; they must form larger aggregates that sink rapidly. One possible mechanism is **biogenic aggregation** (Drake, 1976) into faecal pellets by filter-feeding organisms, but this cannot explain broad spatial distribution of particle settling, especially in turbid plumes. Most fine particles have surface charges which, in freshwater, cause the development of large, repulsive ion clouds. In brackish water with salinities of a few parts per thousand, the ion clouds compress and allow **van der Waals' forces** of attraction to dominate, forming larger aggregates that settle rapidly. When this process occurs inorganically (e.g. glacial meltwater), it is referred to as **coagulation**. If organic molecules help bridge the gap between particles, which is common in middle and low latitudes, the aggregation process is known as **flocculation**. In addition to the mechanism of aggregation, the length of time for aggregation, the suspended-sediment concentration and the turbulence of the environment are likely to control size and settling velocity (McCave, 1984; Hill, 1992; Milligan & Hill, 1998). Despite these complexities, aggregate settling velocities are generally ~1 mm s^{-1} (ten Brinke, 1994; Hill *et al.*, 1998).

The character of the river plume has a strong impact on the delivery of particles to the seabed. Most plumes are **hypopycnal** with densities less than the ambient seawater. They flow and spread at the surface (Fig. 4), controlled by local winds,

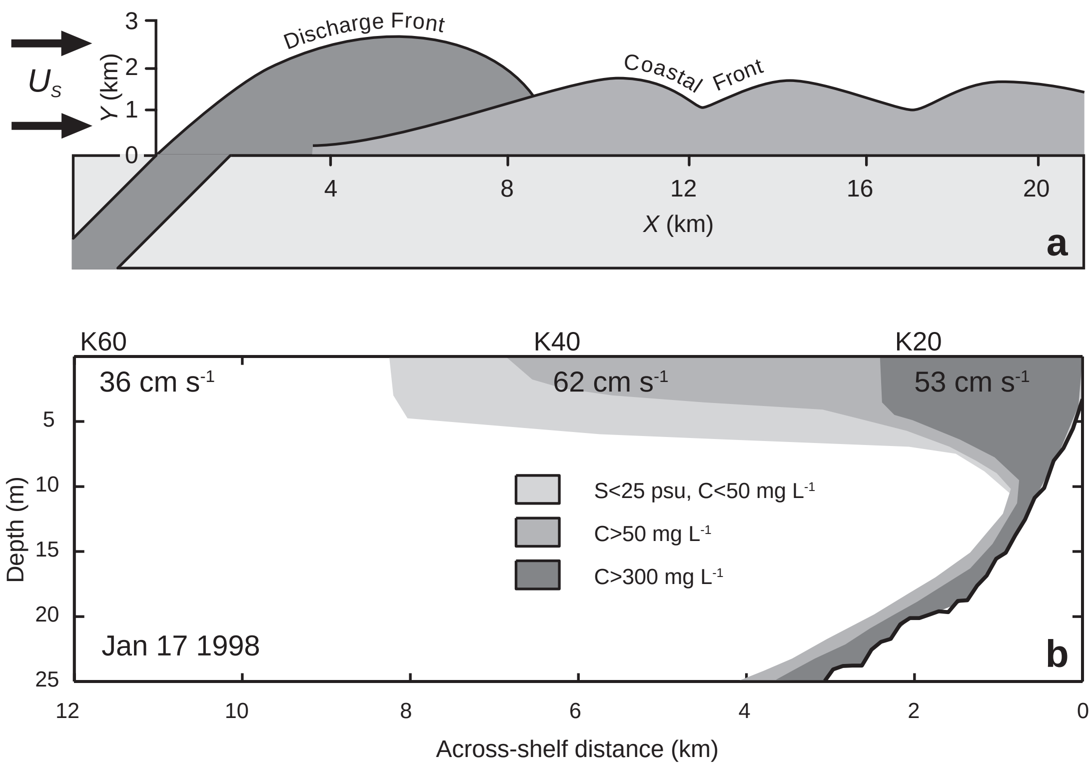

Fig. 4 Hypopycnal plumes. (a) A schematic map view for discharge of a general hypopycnal plume, with a river mouth at an angle to the shoreline. U_s is the velocity of an ambient current directed northward in the northern hemisphere. The combination of ambient current, Coriolis force, and mouth orientation causes the plume to flow to the right, creating a coastal current. (From Hill *et al.*, this volume; modified from Garvine, 1987.) (b) Cross-section (facing northward) of Eel plume on the continental shelf north of the river mouth during a period of northward winds (S = salinity; C = suspended-sediment concentration). The low-salinity and turbid river water extends offshore as a hypopycnal plume flowing northward; velocity measured 2 m below water surface shown in cm s^{-1}. Northward winds also produce downwelling against the coast and the seaward flow of bottom water with suspended sediment. (From Hill *et al.*, this volume; modified from Geyer *et al.*, 2000.)

currents, Coriolis force and the relative significance of inertial and buoyancy forces (Wright, 1977). The path of surface plumes (e.g. direction, speed) has an impact upon the trajectory of settling particles. Under special conditions, rivers can enter water bodies with similar densities, forming **homopycnal** plumes that spread throughout the water column as turbulent jets. If the density of the river plume is greater than the ambient seawater, it forms a **hyperpycnal** plume that sinks and moves near the bottom. Of special importance to this paper are conditions (e.g. floods) where freshwater has extremely high suspended-sediment concentrations ($> 40\ \mathrm{g\ L^{-1}}$) that cause the excess density. These plumes move as gravity-driven sediment flows deflected by Coriolis force and physical oceanographic conditions (e.g. currents), but primarily they follow the steepest bathymetric gradient. Although uncommon (Mulder & Syvitski, 1995), some rivers, especially those with mountainous drainage basins, can reach hyperpycnal conditions and transport massive amounts of sediment across continental margins.

During highstands of sea level, as at present, the processes of sediment delivery tend to be focused in shallow water. For fluvial systems where or when freshwater discharge is relatively weak, aggregation begins within **estuaries** at river mouths (or even within the rivers themselves) and sediment can be trapped there. This is particularly true for low-gradient rivers emptying onto passive margins, such as the Hudson and Delaware rivers. If river plumes with substantial sediment concentrations extend onto the shelves, sedimentation can occur there, and follow the mechanisms described previously in this section. Most active margins have coastal mountain ranges, steep river channels, small or no estuaries and narrow continental shelves (Fig. 1). Under these conditions, plumes can reach the continental slope. Hypopycnal plumes form surface **nepheloid layers** (diffuse clouds of turbid water), which are carried by the local currents and dissipate as suspended sediment settles onto the slope (known as **hemipelagic** sedimentation). Hyperpycnal plumes move down the steepest portions of the slope (commonly submarine canyons), and can accelerate to erode the seabed and refuel their excess density, thus becoming one of several means to create **turbidity currents**. Today, some submarine canyons extend into the mouths of rivers (e.g. Sepik River, Congo River) and gravity-driven sediment flows (e.g. hyperpycnal plumes, turbidity currents) usually dominate sediment transport (Kineke *et al.*, 2000; Khripounoff *et al.*, 2003). During lower stands of sea level, such situations were common.

Delivery of Eel margin sediment

Initial northward winds and currents, a northward-pointed river mouth and the Coriolis force cause the early stages of Eel River flood discharges (associated with winter storms) to be directed northward. The **radius of curvature** defines the turning distance of the plume at the river mouth. This radius is controlled by plume speed and the Coriolis force (Garvine, 1987), and is ~10 km near the Eel mouth. The plume turns into a northward-flowing coastal current (Fig. 4) that is restricted to regions < 40 m deep and is moving at $\sim 50\ \mathrm{cm\ s^{-1}}$ (maximum $130\ \mathrm{cm\ s^{-1}}$; Geyer *et al.*, 2000). Suspended silts and clays, which dominate the discharge, aggregate (mean floc size 230 μm; Curran *et al.*, 2002) and are largely removed from the surface plume within 10 km of the river mouth (Hill *et al.*, 2000). The correlation of discharge events and oceanic storm conditions guarantees turbulence within the coastal current. This turbulence results from wind-driven downwelling that destroys water-density stratification, and from a storm-wave surf zone that extends seaward to as far as 15-m water depth (Curran *et al.*, 2002). The intense turbulence within the surf zone keeps fine sediments suspended, providing a mechanism to resupply the coastal current. As the coastal current moves northward, it experiences some seaward transport due to **Ekman veering** in the bottom boundary layer (Smith & Long, 1976; Drake & Cacchione, 1985). When winds reverse, northward transport is slowed and the plume broadens seaward (Geyer *et al.*, 2000). For periods of low river discharge, correlation with meteorological events is not evident, and variable winds preclude a net direction of sediment transport. In some years, southward transport of shelf sediment can be significant (Ogston & Sternberg, 1999; Ogston *et al.*, 2004).

During coupled discharge and storm events, wave activity has a significant control on aggregate properties observed along the shelf, due to continual injection of particles from the surf zone into the coastal current (Curran *et al.*, 2002). Beyond the surf zone (> 15 m depth), a shelf frontal zone (Fig. 4) concentrates suspended sediment on the inner shelf (Ogston *et al.*, 2000); here, wave activity can stimulate across-shelf sediment transport. Although waves provide little net direction for sediment transport, they can create high-concentration (> 10 g L^{-1}) **fluid muds** in the wave boundary layer (< 10 cm thick) that produce gravity-driven sediment flows moving seaward at 10–30 cm s^{-1} (Traykovski *et al.*, 2000). The signature of these flows occurs within the current boundary layer (lowermost several metres of water column) where velocity normally decreases logarithmically toward the seabed. When concentrations of suspended sediment are very large, velocity increases near the bed within the wave boundary layer (5–10 cm above seabed).

These wave-supported sediment gravity flows transport much sediment mass as they move across shelf. As near-bed wave activity decreases seaward, the gradient of the shelf seabed is not sufficient to allow continued flow, and the sediment stops moving (Wright *et al.*, 2001). Within the resulting flood deposits are fine laminae (centimetre-scale **sedimentary structures**) that record pulses of sediment flux (Wheatcroft & Borgeld, 2000). The location of the gravity-flow deposits generally coincides with the convergence of sediment transport from shelf currents (Wright *et al.*, 1999; Ogston *et al.*, 2000), and together these processes create a locus of sediment deposition on the Eel shelf between 50-m and 70-m water depth and ~10–30 km north of the river mouth (Fig. 5).

Not all sediment discharged to the Eel continental shelf reaches the seabed; much (> 50%) continues to the continental slope. Turbid water in the bottom boundary layer of the shelf can detach near the shelf break and move seaward along an isopycnal surface within the water column as an **intermediate nepheloid layer** (INL). These layers are maintained, in part, by internal waves (McPhee-Shaw *et al.*, 2004). Eel sediment is broadcast across the slope, and rapid delivery is confirmed by the

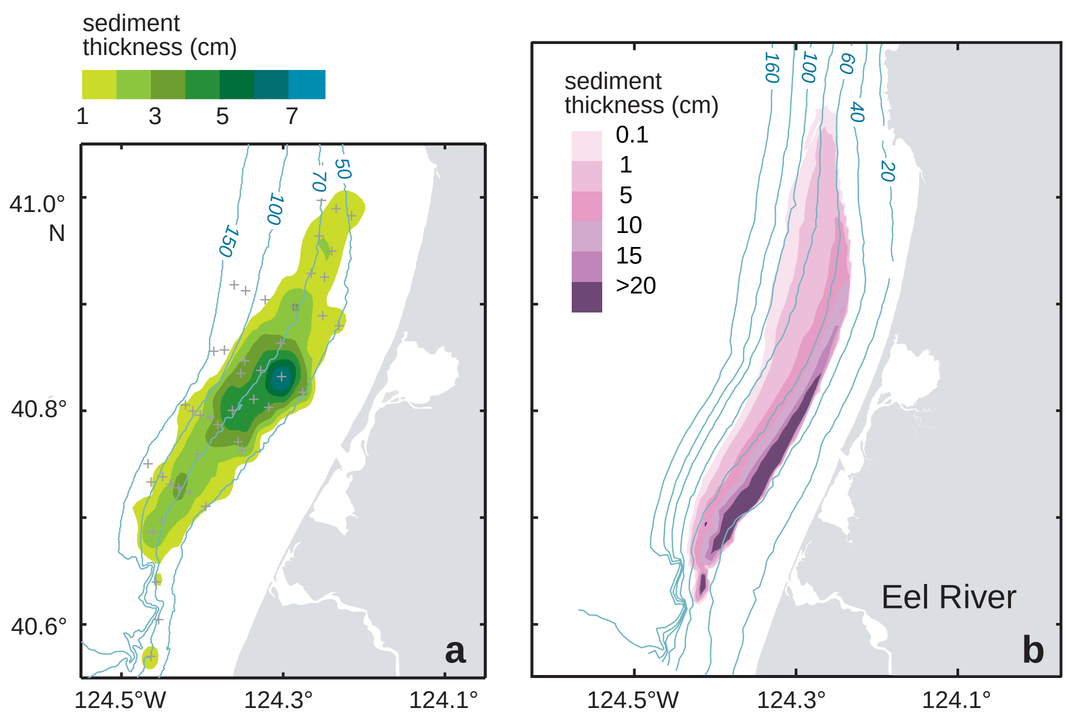

Fig. 5 Shelf sediments resulting from the 1997 flood of the Eel River. (a) Isopach map of the 1997 flood deposit. The thickest portion is found in ~70-m water depth and ~15–25 km north of the Eel River mouth. This compares well with the pattern of the 1995 flood deposit shown in Fig. 10. (From Hill *et al.*, this volume; based on Wheatcroft & Borgeld, 2000.) (b) Predicted thickness of a deposit resulting from wave-supported sediment gravity flows during the 1997 flood event (porosity assumed to be 0.75). Thicknesses are greater and extend farther north than those observed in (a), but the predicted pattern has many similarities to the flood deposit, including its shape and the location of the landward and seaward boundaries. (From Hill *et al.*, this volume; based on Scully *et al.*, 2003.)

presence of the short-lived radioisotope ^{7}Be (half-life 53 days) in sediment traps (Walsh & Nittrouer, 1999). This same isotope is found in the seabed of the open slope (Sommerfield *et al.*, 1999) and Eel Canyon (Mullenbach & Nittrouer, 2000), probably from input through intermediate nepheloid layers and other mechanisms. In the head of Eel Canyon, inverted velocity profiles (increasing near the seabed) similar to gravity-driven sediment flows on the shelf are observed (Puig *et al.*, 2003, 2004). Modelling studies indicate that substantial amounts of Eel sediment discharge are likely to be carried into the canyon by these flows (Scully *et al.*, 2003). During major flood periods (e.g. 1995, 1997), the river may become hyperpycnal, and bottom plumes may carry large fractions of the Eel discharge directly to the Canyon or the open slope north of the Canyon (Fig. 2; Imran & Syvitski, 2000). Therefore, a range of mechanisms associated with the Eel plume deliver sediment to the continental slope during the present highstand of sea level.

Modelling studies indicate that during the **Last Glacial Maximum** (LGM), the Eel basin was wetter and colder, and storm frequency was greater (Morehead *et al.*, 2001). These differences would have caused approximately a doubling of the water and sediment discharge (Syvitski & Morehead, 1999). Most discharge from the modern Eel River occurs with winter rains. For the LGM, increase in precipitation would have caused a more sustained discharge as snow pack melted during the spring and summer. Rains on low-elevation snow also would have caused more intense floods than today. These differences in water and sediment discharges and in the intra-annual variability of discharges distinguish modern and past conditions for sediment delivery to the Eel margin.

SEDIMENT ALTERATION

Processes affecting the preservation of the sediment record during deposition and the early stages of burial have been examined in this volume by Wheatcroft *et al.* (pp. 101–155).

General considerations

Sediment delivered to the seabed is altered in many ways before being preserved by burial. Especially important changes are those that alter the dynamical properties of the seabed thereby impacting lateral transfer of sediment across margins (e.g. alteration of particle-size distribution, bottom roughness, porosity), and those that cause vertical displacements of seabed particles thereby affecting stratigraphic signatures (e.g. alteration of sedimentary structures, acoustic properties). These alterations occur primarily over time-scales of days to years and over vertical length scales of millimetres to decimetres. Deposition of new particles applies a downward force (i.e. weight) to the underlying sediment. Physical processes erode and deposit particles, rearranging them based on hydrodynamic character. Macrobenthic organisms displace particles in the seabed through a wide assortment of activities, including ingestion and defaecation. Chemical processes also alter sediment after delivery to the seabed, but usually have less direct impact on transport and stratigraphy than the other processes (for summaries of chemical alteration see: Aller, 2004; McKee *et al.*, 2004).

Consolidation (also known as **compaction**) decreases **porosity**, as new overburden reduces pore space and displaces pore fluid. Initial changes occur near the surface of the seabed, such that a relatively uniform porosity is approached within a few centimetres (Fig. 6a), although consolidation continues much deeper in the seabed as overburden increases. Porosity profiles impact many properties in the seabed (e.g. bulk density, acoustic signature), and also influence sedimentary processes; high-porosity surface layers are easily eroded by weak shear stresses. Porosity profiles indicate whether the weight of overlying sediment is supported by a particle framework or by pore fluids, conditions that may ultimately determine the distribution of stresses and whether the seabed will fail. For all of these reasons, understanding the consolidation rate of natural sediment is important, as is understanding the factors affecting that rate (e.g. permeability, bioturbation). In general, sands consolidate quickly toward a minimum porosity of ~0.35 (fractional volume of pore space) and muds consolidate more slowly toward minimum values (Been & Sills, 1981; Wheatcroft, 2002). However, fluctuations in sedimentation complicate consolidation history of the seabed. Erosion of the seabed exposes sediment that is **overconsolidated** (Skempton, 1970) relative to

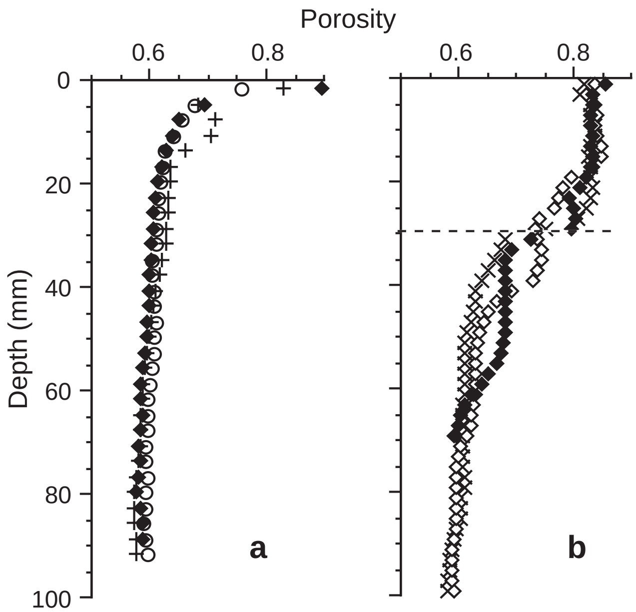

Fig. 6 Sediment porosity profiles on the Eel Shelf. (a) Replicate porosity profiles at a mid-shelf station (70-m water depth) six months before the 1997 flood of the Eel River. Relatively uniform porosity is reached within ~30 mm of seabed surface. (b) Replicate porosity profiles at the same station as (a), but 2 weeks after the 1997 flood. A uniform layer of higher porosity is observed within the upper ~30 mm, which is the thickness of the flood deposit at this location, as documented by X-radiography and radiochemistry. (From Wheatcroft *et al.*, this volume.)

what is expected at the surface. Rapid deposition of thick flood layers places sediment below the surface that is **underconsolidated** (Skempton, 1970) relative to what is expected at that depth in the seabed. Variable grain sizes and biological effects further complicate consolidation, and make modelling and prediction of porosity profiles more difficult.

Physical reworking adds and subtracts particles from locations on the seabed, often removing fine particles (i.e. **winnowing**) and coarsening (i.e. **armouring**) the surface. The fine particles (silts and clays) possess interparticle forces of attraction and, where these sediments deposit, the seabed develops **cohesion**. With consolidation, cohesive forces increase, and the fluid velocities needed for resuspension also increase. Armouring inhibits resuspension by developing a coarse surface layer, and cohesion causes an abrupt decrease in erodibility just below the surface of muddy deposits.

The extent of physical reworking depends on the strength of operative processes (e.g. surface waves, coastal currents) as well as seabed properties (e.g. grain size, porosity). Under special conditions (e.g. equatorial settings with sustained trade winds, shallow tide-dominated coastal areas), reworking can be relatively continuous (Nittrouer *et al.*, 1995). However, most continental margins are dominated by cyclonic storms, which cause episodic physical reworking that is largely the result of surface waves (Komar *et al.*, 1972; Drake & Cacchione, 1985). Waves impact the seabed in water depths less than about half their wavelength, and large waves can rework bottom sediments to depths of 100–200 m in extreme events (Komar *et al.*, 1972). The near-bed wave orbital velocities increase toward shore, and are additionally dependent on wave height and period (Komar & Miller, 1975; Madsen, 1994; Harris & Wiberg, 2001). A velocity of ~14 cm s^{-1} has been observed as the critical value needed for resuspension of muddy shelf deposits by waves (Wiberg *et al.*, 1994, 2002) but this value is influenced by many factors, including grain size and consolidation state.

In non-cohesive sandy sediment, the **active layer** of moving sediment can be a few centimetres thick but, where bedforms develop and migrate, it is comparable to their height (~5–10 cm). In cohesive muddy sediment, the active layer is dependent on the thickness of high-porosity surficial sediment. Erosion and redeposition of sediment create a **graded** deposit (i.e. fining upward) within the active layer (Reineck & Singh, 1972; Nittrouer & Sternberg, 1981). Subsequent to deposition, benthic organisms alter the seabed through a range of activities. Ingestion of particles and formation of faecal pellets change the effective grain size of sediment. Together with formation of mounds and burrows, these processes increase seabed roughness (Jumars & Nowell, 1984) and alter porosity, all of which influence sediment transport. The mucous that glues animal faecal pellets is similar to organic substances produced by microalgae on seabed surfaces, and adhesive coatings from both sources tend to bind the seabed and reduce physical reworking. Feeding, locomotion and dwelling construction

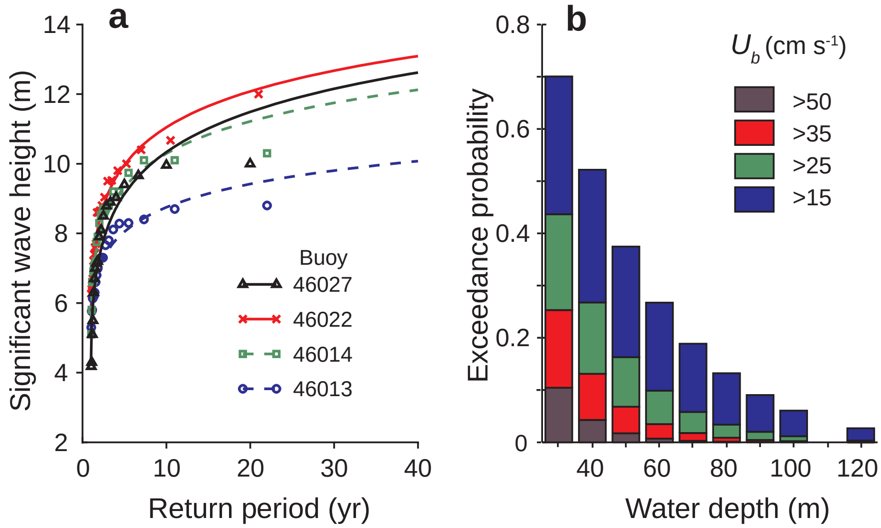

Fig. 7 Wave energy on the Eel margin. (a) Spatial variation of wave characteristics measured by buoys (NOAA National Buoy Center) along the northern California coast (north of San Francisco). Buoy 46022 is located near the Eel River, and has the most energetic wave climate, as shown by the return period for a given significant wave height. (b) The probability is shown of exceeding various near-bed orbital velocities (U_b) at different water depths across the Eel shelf. For the mid-shelf deposits (~50–70 m water depth), a velocity of ~15 cm s^{-1} is likely to erode the surface sediment. (From Wheatcroft *et al.*, this volume; modified from Wiberg, 2000.)

are processes by which benthic organisms stir sediment within the seabed (i.e. **bioturbation**), destroying physical sedimentary structures and creating biological structures. These processes occur within a region known as the **surface mixing layer**, which is ~5–20 cm thick.

Alteration of Eel margin sediment

The Eel margin is an instructive place to investigate seabed alteration, because the relevant processes operate intensely and cause the seabed to be dynamic. Floods of the river create thick layers of high-porosity sediment of variable grain size on the continental shelf. Energetic oceanic storms cause reworking of that sediment. An abundant and well-adapted benthic community rapidly mixes the seabed.

Beyond the inner-shelf sands (> 60 m depth), steady-state porosity profiles asymptotically approach values of 0.6–0.7 several centimetres below the seabed surface (Fig. 6a; Wheatcroft & Borgeld, 2000). The floods in 1995 and 1997 added significant perturbations, creating layers of uniform porosity many centimetres thick (up to ~8 cm) with values of 0.8–0.9 (Fig. 6b). The consolidation rate of this sediment had an important control on the erodibility of the seabed. The upper centimetre returned to steady-state porosities within months (< 4) and made the seabed resistant to erosion, even though a couple of years were needed for deeper flood sediments to reach the lower values (Wheatcroft *et al.*, this volume, pp. 101–155). Concurrent bioturbation imposed significant spatial variability on these general observations.

The Eel margin has the greatest wave energy along the northern California coast (north of San Francisco), with waves reaching heights > 10 m (Fig. 7a; Wiberg, 2000). The inner-shelf region (< 50 m depth) experiences relatively long durations when the near-bed wave orbital velocities exceed the critical value (totalling > 40% of the time; Fig. 7b). These events are sufficient to winnow most mud (silt and clay), and create a seabed dominated by sand. Farther seaward, mud becomes a substantial portion of the seabed (> 50%) and adds cohesion as a relevant property. Despite the energetic wave regime experienced by the Eel margin, the thickness of the active layer is surprisingly small. For a strong wave event estimated to have a 10-yr recurrence interval (December, 1995), erosion occurred to ~2 cm within the seabed at 50-m water depth (Wiberg, 2000). The estimated thickness increases to 5 cm for a 100-yr storm and to 10 cm for a 1000-yr storm. For most storms, however, the active layer is millimetres thick, especially in water depths > 50 m. In addition to redeposition of local sediment, some areas

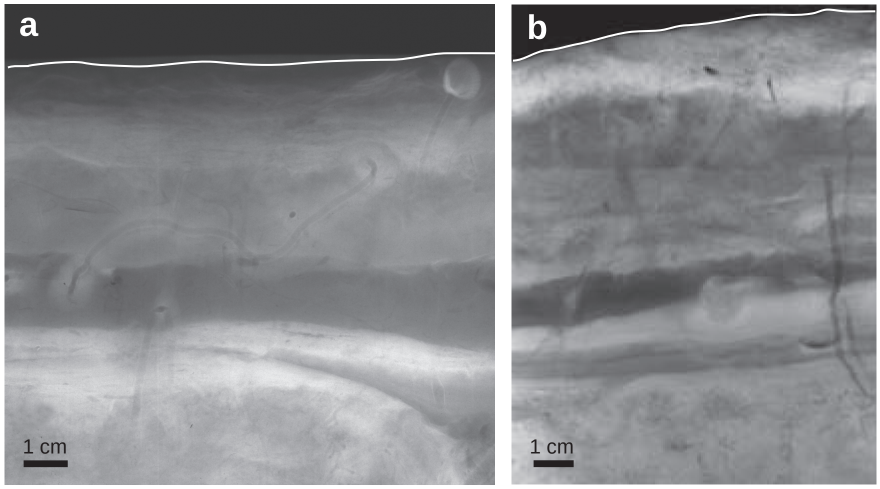

Fig. 8 X-radiograph negatives of sediment cores collected from the mid-shelf about (a) 15 km and (b) 25 km north of the Eel River mouth, illustrating various biogenic structures. (a) Collected from ~70-m water depth during February 1995, showing the January 1995 flood deposit. The burrow extending from middle left to upper right is most likely to be an escape structure of the bivalve mollusc at the sediment–water interface (upper right). (b) Collected from ~60-m water depth during July 1996. The physical sedimentary structures near the base of the radiograph are coarse silt and clay layers in the 1995 flood deposit. Bioturbation has partially destroyed the records of the flood and has imparted a general mottling to the sediment. In addition, animals have created discrete burrows that extend tens of centimetres into the seabed. In 18 months following the 1995 flood event, new sediment was added to the seabed above the flood deposit, a process that favoured preservation of the deposit. (X-radiographs are courtesy of R.W. Wheatcroft, Oregon State University; see also Wheatcroft *et al.*, this volume.)

can experience a convergence of sediment flux during dry storms (e.g. transfer from inner-shelf to mid-shelf depths) adding millimetres to 1 cm of sediment (Zhang *et al.*, 1999; Harris & Wiberg, 2002). The resulting storm deposits are graded, but bioturbation destroys them within weeks (Fig. 8; Harris & Wiberg, 1997; Bentley & Nittrouer, 2003; Wheatcroft & Drake, 2003).

Thicknesses of event deposits are greater during wet storms, due to the influx of new river sediment (Fig. 5). These deposits have relatively high clay contents (Drake, 1999), and can be easily identified by their physical sedimentary structures (Wheatcroft & Borgeld, 2000), radiochemical signatures (presence of ^{7}Be and low level of ^{210}Pb; Sommerfield *et al.*, 1999) and terrestrial carbon composition (Leithold & Hope, 1999). Subsequent to the formation of clay-rich flood deposits, the seabed coarsens by the addition of silts and fine sands from the inner shelf (Drake, 1999). In addition, animal bioturbation gradually destroys physical sedimentary structures and creates discrete biogenic structures (Fig. 8). Polychaete worms dominate macrofauna on the Eel margin, and most of the abundant species are subsurface-deposit feeders (Bentley & Nittrouer, 2003; Wheatcroft, 2006). They produce many small burrows (millimetres diameter) within the upper 3–5 cm and build a few larger burrows (1–10 cm diameter, some with reinforced lining) extending down as much as 15–20 cm (Fig. 8). On the Eel shelf, the dominance of subsurface-deposit versus surface-deposit feeders minimizes the importance of faecal pelletization at the seabed surface (Drake, 1999). Biogenic seabed roughness is important seaward of ~60 m depth (Cutter & Diaz, 2000), but monitoring observations in these deeper shelf locations (Ogston *et al.*, 2004) demonstrate significant temporal variability as storm events form ripples, even on substrates of silt and clay.

Subsurface bioturbation can be quantified from seabed profiles (upper 4–8 cm) of the short-lived radioisotope ^{234}Th (half-life 24 days; Aller & Cochran, 1976; Wheatcroft & Martin, 1996). The **biodiffusion coefficient** is moderately high (3 cm^2 yr^{-1}

to > 100 cm^2 yr^{-1}, mean 20–30 cm^2 yr^{-1}; Bentley & Nittrouer, 2003; Wheatcroft, 2006) on the Eel margin. It reveals substantial small-scale variability over tens of metres, but also demonstrates a decrease between the shelf and the deeper continental slope (water depth > 500 m; Wheatcroft *et al.*, this volume, pp. 101–155). Most interesting is the temporal variability in bioturbation. Organism abundance shows an increase during summer and autumn, and a decrease in winter due to annual cycles of recruitment and growth (Wheatcroft, 2006). Although the extreme flood of January 1997 caused a subsequent drop in abundance, the mortality that year was comparable with other winters without major floods, and was consistent with weak seasonal changes in biodiffusive mixing intensity (slight increases in autumn). Winter is normally a period of low numbers of benthic organisms and low bioturbation activity in the seabed. Therefore, the Eel margin benthic community is well adapted to seasonal cycles in storm reworking and flood deposition.

The dominance of the subsurface-deposit feeders controls the preservation of sedimentary signals on the Eel margin. Important factors are thickness of event signals (storm reworking, flood deposits), thickness of the surface mixing layer, intensity of bioturbation (biodiffusion coefficient) and the sediment accumulation rate. Knowledge of these terms allows evaluation of the **transit time** for a signal to pass through the surface mixing layer, and the **dissipation time** for destruction of the signal (Wheatcroft, 1990). For the Eel shelf, the transit time is 9–65 yr and the dissipation time is ~2 yr; therefore, most signals are destroyed before they can be preserved (Wheatcroft & Drake, 2003). This is particularly true for physical sedimentary structures, which are lost due to particle mixing with overlying and underlying sediment. Event layers > 5 cm thick can be preserved, but those < 3 cm cannot. Other event signals (e.g. increased clay content, decreased ^{210}Pb activity, increased terrestrial carbon) are smeared vertically, but are still recognizable in preserved strata (Sommerfield & Nittrouer, 1999; Blair *et al.*, 2003; Wheatcroft & Drake, 2003). The timing of subsequent events can have an impact on preservation. For example, emplacement of the 1997 flood deposit effectively decreased the transit time for the 1995 flood deposit and allowed its partial preservation. Without such benefit, the 1997 flood deposit was destroyed in 2.5 yr (Wheatcroft & Drake, 2003).

SEDIMENT DISPERSAL SYSTEM

The dispersal of sediment on continental margins has been reviewed in this volume by Sommerfield *et al.* (pp. 157–212).

General considerations

Fluvial sediment is delivered to the seabed, where it undergoes alteration that influences its burial or transport to more distal locations. The integrated result over decades and centuries (i.e. longer than the transit time through the surface mixing layer) is a sedimentary deposit stretching along a succession of hydraulically contiguous sedimentary environments. This succession of environments is a **sediment dispersal system** (Sommerfield *et al.*, this volume, pp. 157–212) and the marine portion is just part of a longer system stretching from terrestrial sources. The expansion of time-scales brings new factors into the consideration of margin sedimentation. The slowing of **eustatic** (i.e. global) sea-level rise ~5000 yr ago (from ~5 mm yr^{-1} to ~2 mm yr^{-1}) has allowed some rivers to fill their estuaries, to extend sediment dispersal systems to the continental shelf and slope, and to form deposits with significant morphological expression (e.g. subaerial and subaqueous **deltas**). As such deposits build toward ambient sea level, they consume the space available for sediment accumulation (i.e. accommodation space). On active margins, vertical **tectonic motions** cause subsidence and uplift that adds or subtracts space for further sedimentation. Changes in accommodation space can put the seafloor into or out of energetic environments reworked by physical processes (e.g. surface waves), and can lead to displacement of sedimentation along a dispersal system.

The increased time-scale also brings climatic variability into consideration. Fluctuations in global precipitation patterns have caused periods, lasting from many decades to centuries during the late Holocene, when North America was wet and flood-prone (Ely *et al.*, 1993; Knox, 2000). On shorter time-scales, ENSO (El Niño–Southern Oscillation) and PDO (Pacific Decadal Oscillation) events have impacted fluvial discharge (Inman & Jenkins, 1999; Farnsworth & Milliman, 2003). Land use by humans has compounded the climatic impacts, both increasing sediment discharge (farming, logging) and decreasing discharge (damming, diverting). Global sediment budgets indicate there has been an anthropogenic increase in fluvial transport (from 14 to 16 Gt yr^{-1}). They also suggest ~30% trapping of this sediment landward of the coast, so that the net discharge to the ocean is ~10% less than natural levels (12.6 Gt yr^{-1}; Syvitski *et al.*, 2005). Global budgets mean nothing to individual rivers, where the scales of perturbations, the mechanisms associated with sediment routing and the storage capacity of the basin determine the impact of perturbations (Walling, 1999). Generally, these factors lead to anthropogenic impacts being greatest (and commonly most conspicuous) on rivers of small to moderate size.

The diversity and intensity of processes operating on continental margins creates the rich record of events preserved in the deposits of sediment dispersal systems. However, these same processes cause erosion and time gaps (i.e. **hiatuses**) in the record over a range of scales (e.g. storm erosion, sea-level change). In this regard, the metric for quantitatively evaluating sedimentation is the mass flux into the seabed, averaged over some time-scale. Ephemeral placement on the seabed is **deposition**, but the sediment is subsequently impacted by **erosion**. The integrated sum of deposition and erosion through time is **accumulation**. The relevant time-scales for deposition rate and accumulation rate can be set for any processes of interest (McKee *et al.*, 1983). As described for this discussion of sediment dispersal systems, deposition refers to sediment placement over days/months and accumulation is the net growth of the seabed over decades/centuries. The distinction is important, because mass flux into the seabed is inversely related to the time-scale of integration (e.g. Fig. 9; Sadler, 1981; Sommerfield, 2006), as the result of more and of more severe hiatuses impacting strata formation over progressively longer time-scales.

Fortunately, a range of natural and artificial **radioisotopes** is found in terrestrial and marine environments, and they can serve as chronometers tagged to sediment particles. The large surface area (per gram of sediment) and the surface charges of silt and clay particles allow them to adsorb large concentrations of particle-reactive chemical components, including radioisotopes. Analytical techniques typically limit sedimentological use of radioisotopes to a time-scale < 4–5 half-lives. Of

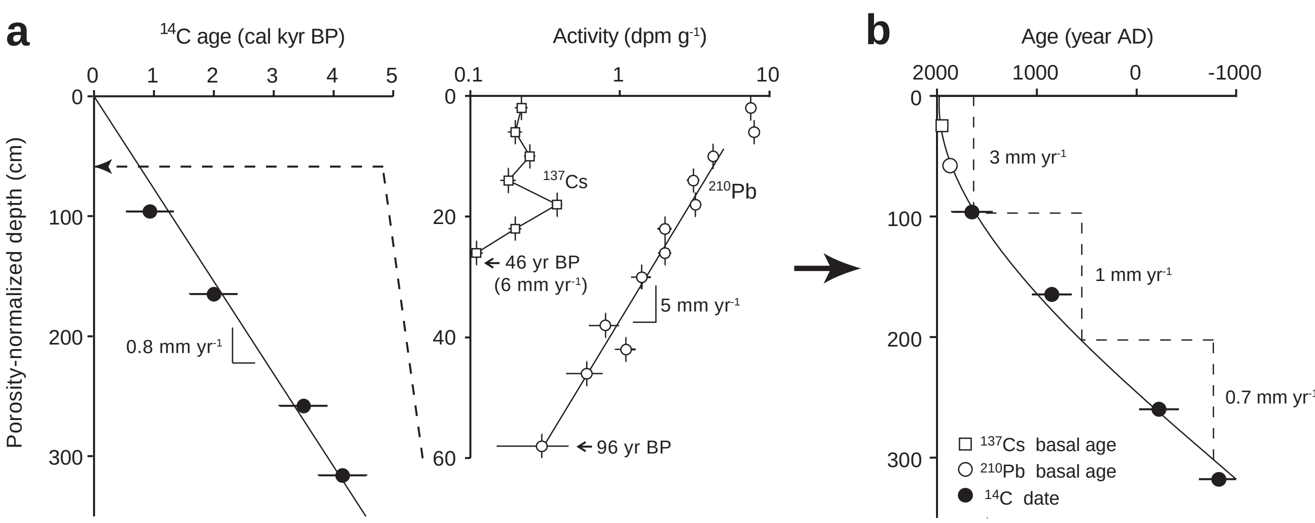

Fig. 9 Accumulation rates of Eel shelf sediments. (a) The two profiles show ^{14}C, ^{210}Pb and ^{137}Cs for the same site on the Eel shelf (95 m water-depth), and allow calculation of accumulation rates integrated over time-scales of ~3000 yr, ~100 yr and ~50 yr, respectively (data points have been adjusted vertically to a uniform bulk porosity). (b) Composite profile illustrating vertical changes in ages and accumulation rates within the seabed. The accumulation rates are greater for the uppermost sediment column, because it retains a record that is more complete than the underlying strata. (From Sommerfield *et al.*, this volume.)

special relevance here (Sommerfield *et al.*, this volume, pp. 157–212), ^{234}Th (half-life 24 days) and ^{7}Be (half-life 53 days) have primary sources, respectively, in ocean water (from decay of dissolved ^{238}U) and in terrestrial soils (from cosmogenic fallout). Lead-210 (half-life 22 yr) has several potential sources but, in ocean water, primarily comes from decay of ^{238}U-series radioisotopes. Lead-210 accumulation rates are commonly verified by profiles of ^{137}Cs (Fig. 9), a bomb-produced radioisotope. Caesium-137 was globally distributed by transport through the atmosphere and by subsequent fallout, and it first reached continental-margin sediments in ~1954. On the long end of these discussions, ^{14}C (half-life 5730 yr) ages are recorded in organic C (e.g. wood fragments) and inorganic $CaCO_3$ (e.g. shell fragments).

By using radiochemical tools with different half-lives, a composite understanding can be obtained for continental-margin sedimentation over a range of time-scales. For example, the dichotomy between deposition and accumulation rates can be related to processes and patterns of sediment dispersal. The Yangtze River undergoes flooding during the quiescent summer months, and rapidly deposits much sediment on the continental shelf near its mouth. However, longer-term accumulation rates indicate that winter storms remove and transport > 50% of this sediment to distal portions of the dispersal system (McKee *et al.*, 1983; DeMaster *et al.*, 1985). In contrast, the Amazon River has peak discharge during intervals of seasonally intense tradewinds and waves, and most of its sediment discharge (> 50%) is immediately displaced along the dispersal system > 200 km from the river mouth, to shelf areas where it deposits and near where it ultimately accumulates (Kuehl *et al.*, 1986, 1996).

Eel margin sediment dispersal system

The Eel basin has experienced multiple decades of sustained wet, dry and variable conditions during the past 100 yr (Sommerfield *et al.*, this volume, pp. 157–212). El Niño–Southern Oscillation events can bring unusual precipitation, but the location of the basin between latitudinal weather bands precludes a clear repetitive signal (e.g. El Niño brought the driest year in 1977, and the wettest year in 1983). The second half of the 1900s was a period with increased logging in the Eel basin, and, together with enhanced precipitation (Sommerfield *et al.*, 2002), this land use significantly increased sediment yield (by 23–45%). Other forms of human interaction (e.g. damming) were minimal, so the increased sediment flux was transferred to the ocean.

In addition to the storm-related physical oceanographic processes near the Eel mouth that have been described above, regional circulation influences distal portions of the dispersal system. Seaward of the shelf break, the **California Current** flows southward (Hickey, 1979, 1998) and, on the shelf, the **Davidson Current** flows northward during the autumn and winter (Strub *et al.*, 1987). The local promontories (Cape Mendocino, Trinidad Head) can deflect these currents (Pullen & Allen, 2000), leading to the seaward transport of water and suspended sediment and to the development of eddies (Washburn *et al.*, 1993; Walsh & Nittrouer, 1999). Other morphological features on the Eel margin influence the fate of water and sediment, especially Eel Canyon, which forms a chasm across the southern boundary. More subtle across-margin ridges (**anticlines**) and depressions (**synclines**) are moving up and down at rates of millimetres per year (averaged over millennia; Orange, 1999).

Sediment deposition on the Eel margin is clearly demonstrated by the distribution patterns associated with the 1995 and 1997 flood events, which discharged ~24×10^6 t and ~29×10^6 t of sediment, respectively (Wheatcroft & Borgeld, 2000). Both events formed elliptical deposits on the middle shelf north of the Eel mouth (Figs 5 & 10), representing 20–30% and 15–30% of the mass discharged, respectively. The similarity of the two deposits suggests that the mechanisms of emplacement operated in a repetitive manner. The remainder of the sediment could deposit landward, northward, southward or seaward of these deposits. The inner-shelf sands contain some intermixed mud, and Humboldt Bay might receive some sediment through tidal exchange and estuarine circulation. The Davidson and California Currents could move some surface plumes of sediment beyond the confines of the Eel margin (e.g. Mertes & Warrick, 2001). However, the bulk of sediment is thought to be transported seaward of the Eel shelf by a combination of hyperpycnal flows, storm-induced fluid muds and intermediate nepheloid layers.

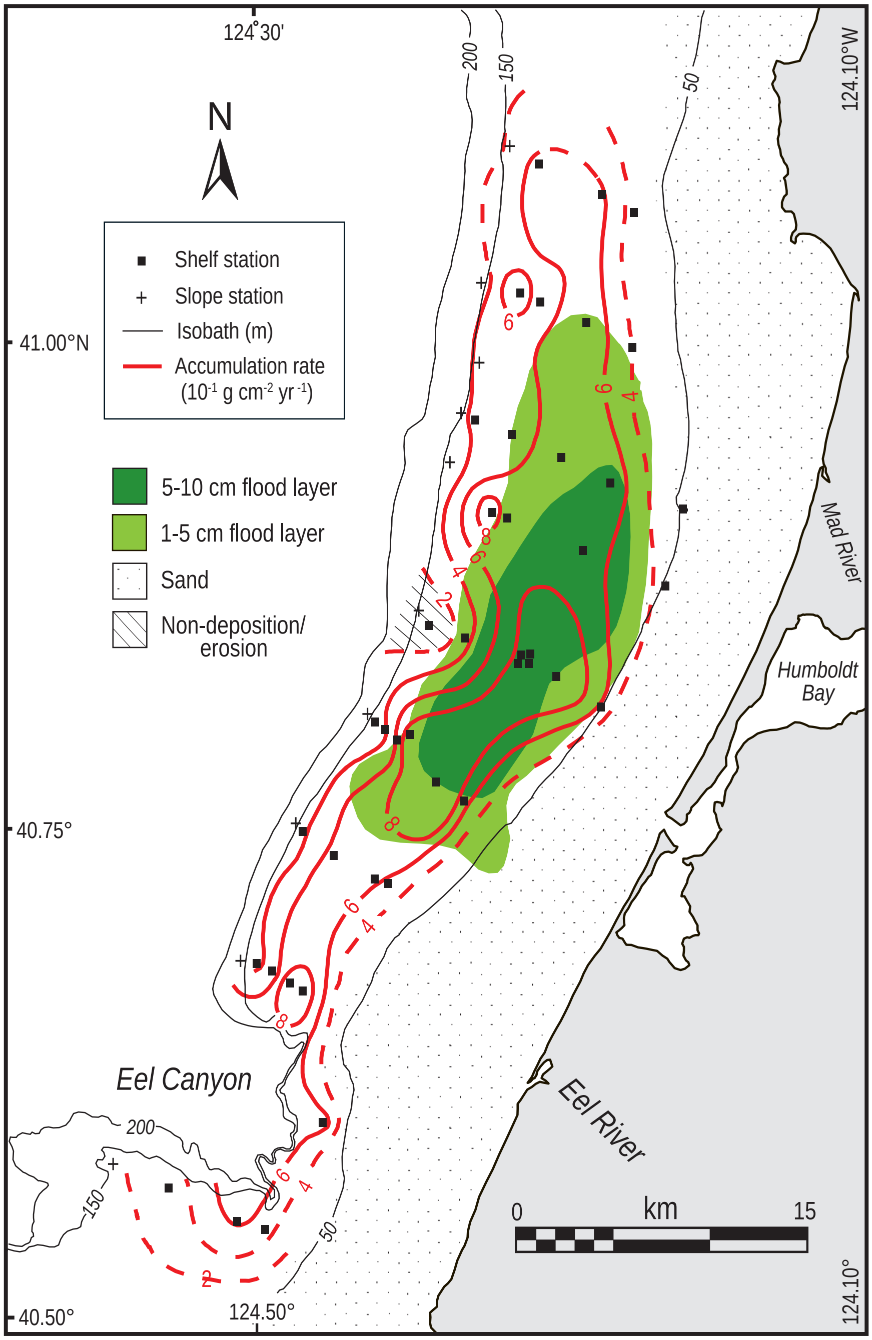

Fig. 10 Contour map of ^{210}Pb accumulation rates (red isopach lines) on the Eel shelf, superimposed on the thickness of the January 1995 flood deposit (green shaded areas). They coincide well, and indicate maximum values in ~50–70 m water depth and ~10–30 km north of the river mouth. Approximately 20–30% of sediment discharged by the 1995 flood remained on the shelf, and this fraction is similar to that retained over a 100-yr time-scale. (Modified from Sommerfield & Nittrouer, 1999.)

On time-scales of decades and centuries, the fate of silt and clay from the Eel River shows a similar pattern: ~10% is buried with inner-shelf sands (< 60-m water depth; Crockett & Nittrouer, 2004), ~20% accumulates on the middle and outer shelf (Fig. 10; Sommerfield & Nittrouer, 1999), and the remainder is exported to deeper water. Accumulation on the upper slope (150–800 m) accounts for ~20% of the Eel sediment discharge (Alexander & Simoneau, 1999) and Eel Canyon is the alternative pathway on the slope (Mullenbach & Nittrouer, 2000, 2006). These observations demonstrate that the Eel shelf traps less than a third of modern sediment discharge. They also highlight the importance of Eel Canyon for dispersing sediment seaward; as much as 50% of the river discharge could be moving into and through the canyon. Both the escape of sediment from the shelf and the large flux through Eel Canyon are occurring during the present highstand of sea level, and probably reflect sedimentation typical of narrow, tectonically active continental margins.

The accumulated strata reveal interesting sedimentary trends along the dispersal system. Grain size decreases progressively with distance from the Eel mouth, both northward and seaward. Fining continues across the slope, but includes an anomalously coarse zone below the shelf break (250–350 m water depth; Alexander & Simoneau, 1999), possibly due to winnowing by shoaling internal waves (Cacchione *et al.*, 2002). Organic carbon shows a progressive increase in the marine component relative to the terrestrial component with distance from the Eel River (Blair *et al.*, 2003). Temporal changes are also observed for the past ~4000 yr (Sommerfield *et al.*, this volume, pp. 157–212). The accumulating sediment has progressively become finer (less sand, more silt) as the dispersal system has evolved since sea-level rise slowed. For the past 200 yr the upward fining trend has been accelerated by human impacts on land use. The magnitude and frequency of flood events also have increased, imposing the sedimentary characteristics of those events: high sediment flux, increased clay content, much terrestrial carbon. The changes have been particularly acute during the past 50 yr (Sommerfield *et al.*, 2002), and reflect the combined effects of land use (clear cutting, road building) and climatic increases in precipitation intensity.

There is a distinct similarity of the shelf patterns in flood deposition and accumulation rates over decades/centuries (see Figs 5 & 10). This is due to the correlation of river discharge and energetic oceanic conditions. Most sediment is immediately transported to a stable location for accumulation (i.e. where it will not be eroded by strong boundary shear stresses), rather than being temporarily deposited and subsequently transported. In this regard, the Eel margin more closely approximates the conditions of shelf deposition/accumulation near the mouth of the Amazon River than those near the Yangtze River. However, the accumulation pattern over decades/centuries also matches well with thicknesses of late Holocene strata (see below), which are related to tectonic features on the shelf (Orange, 1999; Burger *et al.*, 2002; Spinelli & Field, 2003). Millenial accumulation rates are < 1 mm yr^{-1} over anticlines, and reach 6 mm yr^{-1} in synclines. The similarity of accumulation patterns suggests that tectonic activity on the margin impacts sedimentation on scales as short as decades (as detailed in Sommerfield *et al.*, this volume, pp. 157–212). Likely candidates for the operative mechanisms are gravity flows, which are common on the Eel margin and respond to subtle gradients of the seabed.

SEABED FAILURE

The processes and products of seabed failure on continental margins have been addressed in this volume by Lee *et al.* (pp. 213–274) and Syvitski *et al.* (pp. 459–529).

General considerations

The dispersal of sediment to sites of accumulation is a continuing process; new sediment buries old sediment, causing consolidation and development of strength to resist subsequent shear forces. However, in some cases, forces exerted on the seabed are stronger than the strength developed, and the seabed fails. The resulting **mass movement** is driven by body forces (i.e. gravity) rather than by fluid stresses exerted on the seabed surface. In this way, mass movement differs from sediment erosion and transport. Some famous failures have occurred in the past 100 yr, including the 1929 Grand Banks (Heezen & Ewing, 1952), 1964 Alaska (Coulter & Migliaccio, 1966; Lemke, 1967) and 1998 Papua New Guinea (Tappin *et al.*, 1999; Geist, 2000) landslides; all were triggered by earthquakes and all initiated tsunamis. Failures can be triggered by other processes, including large waves associated with storms, such as Hurricane Camille in 1969 (Sterling & Strohbeck, 1973; Bea *et al.*, 1983) and more recent hurricanes in the Gulf of Mexico. Large landslides have also occurred in the geological past leaving scars and deposits as evidence, such as the Storrega landslides (Bryn *et al.*, 2003) during the Pleistocene and Holocene (most recently ~8200 yr ago). These removed a large piece of the Norwegian continental margin (~3000 km^3) and displaced it over a region stretching ~800 km.

The largest landslides on Earth are found in the ocean.

Underwater **landslides** move sediment with a range of speed and internal deformation of the deposit (Varnes, 1958). Subclasses of movement include **creep,** when the movement is slow, and **slumps,** when sediment blocks rotate along a curved failure surface. **Liquefaction** occurs when loosely packed particles temporarily lose contact with each other, and the weight of the deposit becomes supported by pore fluids. All styles of failure can lead to disintegration of the sediment deposits, and development of gravity flows (e.g. debris flows, turbidity currents).

Failures and landslides are prevalent in environments of the continental margin where thick deposits of soft sediment accumulate. Fjords can receive large amounts of rock flour (with limited cohesion) that rapidly accumulate on steep gradients (some > 5°). Fjord sediments are commonly organic rich and produce methane gas. Subsequent earthquakes or even very low tides can initiate failure (Syvitski & Farrow, 1983; Prior *et al.*, 1986). Deltas are also loci of rapid accumulation, and despite gentle gradients (usually < 2°) can fail in response to earthquakes or storms (Coleman *et al.*, 1980; Field *et al.*, 1982). Continental slopes are extensive and steep (> 4°) regions with a propensity for failure, which is accentuated during lowstands of sea level when fluvial and glacial sediment discharge occurs directly at the top of the slope. Gas and gas hydrates, which commonly form on continental slopes, can be responsible for failures (Field & Barber, 1993), especially with sea-level fall that reduces hydrostatic pressure on the seabed and causes dissociation of hydrates (Kayen & Lee, 1991). Submarine canyons are regions of preferential sediment accumulation, and failures near their heads can lead to gravity flows that supply sediment to submarine fans at the bases of the canyons (Hampton, 1972; Booth *et al.*, 1993). Especially important are failures triggered by earthquakes on active margins that cause turbidity currents to transport much sediment long distances (e.g. Goldfinger *et al.*, 2003). During the present highstand of sea level, continental slopes are generally below the depth of surface-wave influence, but the heads of submarine canyons are in shallower water and can be impacted by energetic waves (Puig *et al.*, 2004). From observations in a range of sedimentary environments, the factors recognized to influence failures are sediment accumulation rates, bathymetric gradients, seismicity, storm waves and gas.

Failures and landslides occur when and where driving stresses exceed shear resistance. Bathymetry is important because it defines the gravity-induced stresses. Earthquakes cause cyclic accelerations in addition to gravity (Lee & Edwards, 1986). Similarly, large storm waves produce alternating pressures that create stresses superimposed on those from gravity (Henkel, 1970). In opposition to the applied stresses is the **shear strength** of the seabed, which is defined as the limit of stress before failure. The shear strength of sediment increases as it is buried by subsequent accumulation and as the seabed consolidates. The **factor of safety** for the seabed is the shear strength divided by the shear stress. In addition to large stresses, the factor of safety can be reduced by a loss of shear strength. A common mechanism is the development of excess pore pressures, due to (i) the inability to remove pore fluids during consolidation (e.g. under high accumulation rates; Coleman & Garrison, 1977), (ii) the development of gas bubbles (e.g. from the decay of organic matter or the dissociation of hydrates; Kayen & Lee, 1991) and (iii) the infusion of additional water (e.g. by groundwater seepage). Earthquakes and storm waves apply stresses cyclically, which destroys particle fabric (i.e. grain-to-grain contact), causes liquefaction (Seed, 1968), and increases pore pressures. Human activity can cause failure as well, commonly from construction at or near the shoreline that destabilizes the seabed. Sometimes, the resulting landslides even stimulate tsunamis, e.g. during 1979 in Nice, France (Seed *et al.*, 1988) and during 1994 in Skagway, Alaska (Rabinovich *et al.*, 1999).

Whether by increased stresses, reduced strength or both, marine sediments can fail. After failure, they create landslide deposits or disintegrate into fluid flows (Hampton *et al.*, 1996), depending on the bulk density (i.e. porosity) of the sediment (Poulos *et al.*, 1985; Lee *et al.*, 1991). A critical threshold separating these two fates (i.e. slide deposit from fluid flow) can be defined for each sediment type. If seabed conditions have densities below this threshold (**contractive sediment**), then excess pore pressures will develop after failure, and the sediment will flow. Densities above this threshold (**dilatant sediment**)