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Deltaic, mixed and turbidite sedimentation of ancient foreland basins

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Abstract

The marine fill of ancient foreland basins is primarily recorded by depositional systems consisting of facies and facies associations deposited by a variety of sediment gravity flows in shallow-marine, slope and basinal settings. Tectonism and climate were apparently the main factors controlling the sediment supply, accommodation and depositional style of these systems. In marginal deltaic systems, sedimentation is dominated by flood-generated hyperpycnal flows that build up impressive accumulations of graded sandstone beds in front of relatively small high-gradient fan-deltas and river deltas. During periods of tectonically forced lowstands of sealevel, these systems may commonly shift basinward to shelfal and slope regions. Instability along the edges of these lowstand deltas and sand-laden hyperpycnal flows generate immature and coarse-grained turbidite systems commonly confined within structural depressions and generally encased in distal delta-front and prodeltaic deposits. Because of the close vertical and lateral stratigraphic relations between deltaic and turbidite-like facies, these marginal systems are herein termed 'mixed depositional systems'. They are very common in the fill of foreland basins and represent the natural link between deltaic and basinal turbidite sedimentation.

Basinal turbidite systems form in deeper water elongate highly subsiding troughs (foredeeps) that developed in front of advancing thrust systems. The impressive volumes of sheet-sandstones that form the fill of these troughs suggest that basinal turbidite systems are likely to form following periods of dramatic tectonic uplift of adjacent orogenic wedges and related high-amplitude tectonically-forced sealevel lowstands. In such deep basinal settings, sediment flux to the sea is dramatically increased by newly formed sediment in fluvial drainage basins and the subaerial and submarine erosion of falling-sealevel deltaic deposits generated during the uplift. Turbidity currents are very likely to be mainly triggered by floods, via hyperpycnal flows and related sediment failures, but can fully develop only in large-scale erosional conduits after a phase of catastrophic acceleration and ensuing bulking produced by bed erosion. This process leads to deepening and widening of the conduits and the formation of large-volume highly efficient bipartite currents whose energy dissipation is substantially reduced by the narrow and elongate basin geometry. These currents can thus carry their sediment load over considerable distances down the basin axis.

Keywords: Foreland basin; Marginal deltaic and turbidite sedimentation; Basinal turbidite sedimentation

1. Introduction

The manner in which changes in the fluvial regime control the growth of basin-margin deltas with time and, directly or indirectly, affect the style of associated turbidite sedimentation in deeper waters has received little attention from sedimentologists and stratigraphers over the years. In fact, most classic 'fluvial', 'deltaic' and 'turbidite' sedimentological models have largely ignored each other, implicitly assuming the lack of close genetic relations between these kinds of sedimentation.

Stemming from the concept of hyperpycnal flows of Bates (1953) and later important research on some modern deltas (Wright, Yang, Bornhold, Keller, Prior, & Wiseman, 1986), a growing body of evidence has recently emphasized a close relationship between turbidity currents and rivers in flood in both modern (Piper & Normark, 2001) and ancient (Mutti, Tinterri, Remacha, Mavilla, Angella, & Fava, 1999) depositional settings. As a result, the structure of hyperpycnal flows exiting river mouths and the way in which these flows may evolve into turbidity currents have received considerable attention by several authors (Mulder & Syvitski, 1995). Quite surprisingly, however, no attempts have been made to assess the importance of hyperpycnal flows in both modern and ancient delta-front and prodeltaic sediments where these flows should be best recorded.

In this paper we describe and discuss ancient deltaic and turbidite depositional systems of foreland basins. The results of our studies highlight the fundamental importance of flood-dominated fluvio-deltaic systems in basin margin settings and the broad spectrum of facies and depositional

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styles that characterize these sediments from river mouths to slope regions. These flood-generated depositional systems form impressive sedimentary volumes that have been ignored or misinterpreted in previous literature. As intended in this paper, basinal turbidite sedimentation of foreland basins is restricted to elongate foredeeps which can only be reached by large-volume and highly efficient turbidity currents exiting large-scale submarine erosional conduits after experiencing a phase of catastrophic acceleration and bed erosion. These currents commonly deposit very thick accumulations of laterally extensive turbidite sandstones.

Most of the data and concepts that form the basis of this paper are derived from extensive field studies carried out in many exposed tectonically controlled basins over the years and particularly in the Eocene of the south-central Pyrenees, Spain, the Oligocene and Miocene of the Tertiary Piedmont Basin (TPB) and the northwestern Apennines, northern Italy. Some of the results of these studies have been discussed in previous papers (Mutti, 1992a; Mutti, Davoli, Tinterri, & Zavala, 1996; Mutti et al., 1999; Mutti, Tinterri, di Biase, Fava, Mavilla, Angella et al., 2000).

2. General depositional setting of foreland basins

Foreland basins form in front of active thrust systems in growing orogenic wedges where accommodation is essentially provided by the subsidence of an outer foreland plate under the load of an orogenic wedge (flexural subsidence), or through more complex crustal processes. The general structural and depositional setting of these basins has been amply discussed by several authors from a number of both ancient and modern basins (Allen & Homewood, 1986; Ricci Lucchi, 1986; De Celles & Giles, 1996).

An idealized transect (Fig. 1), oriented perpendicular to the main structural axes of an orogen and largely inspired from the northern Apennines and southern Pyrenees, shows that sedimentation of the broad foreland region takes place in three distinct and coeval basins including:

1. wedge-top basins, generally resting unconformably on the growing orogenic wedge, and filled in with alluvial, deltaic and mixed depositional systems;
2. a foredeep basin, i.e. an elongate and asymmetric trough developed adjacent to the thrust front and characteristically infilled with deep-water basinal turbidites; associated piggy-back basins form along the inner margin of the foredeep due to forward thrust propagation; and
3. an outer and shallower ramp developed on the passive foreland plate. Sedimentation associated with peripheral bulges developed in the outer regions of foreland basins is omitted from this discussion.

Although foreland basins may differ in terms of geodynamic setting, general basin configuration, rates of subsidence, style of structural deformation and deposition, and location of the main sediment source areas (Allen & Homewood, 1986; Covey, 1986; Ricci Lucchi, 1986), their overall evolution is essentially similar and involves three main stages (Fig. 2). The first stage records the inception of thrusting and flexural subsidence; the foredeep remains essentially underfilled and the passive margin of the foreland experiences a progressive drowning of its depositional profile. The second stage is recorded by turbidite sand deposition in the foredeep and the migration of the foredeep axis and sand depocenters due to forward thrust propagation. During the third and final stage, basinal turbidite sedimentation progressively ceases, being replaced by fluvio-deltaic and eventually alluvial sedimentation (Covey, 1986).

3. Deltaic, mixed and basinal turbidite systems of foreland basins: main facies associations and inferred processes

3.1. General

Terrigenous marine sedimentation of ancient foreland basins is primarily recorded by three types of depositional systems, which can be differentiated by their facies

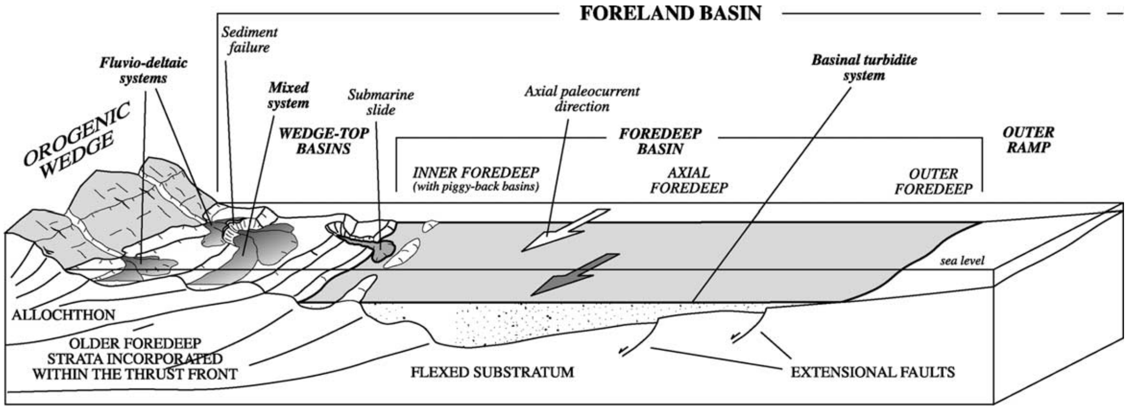

Fig. 1. Scheme showing the main structural and depositional elements of an alpine (mediterranean)-type foreland basin.

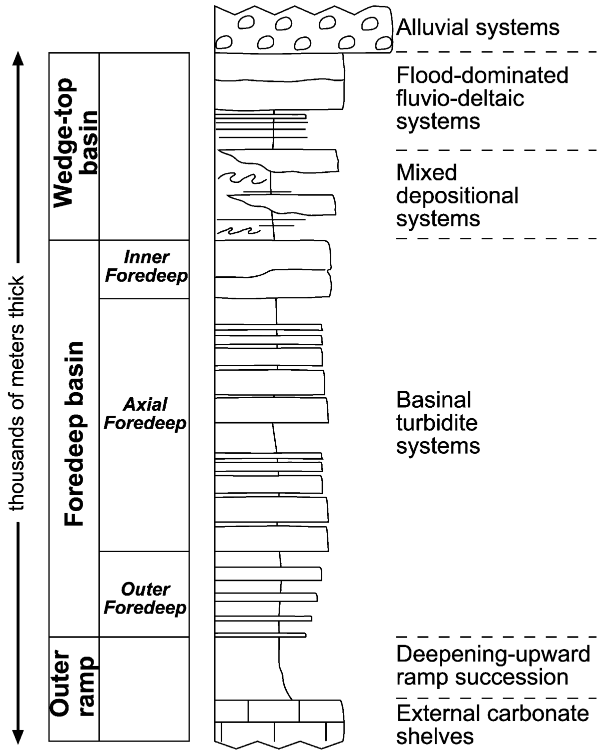

Fig. 2. Idealized vertical stacking pattern of a foredeep basin (mostly inspired by the Eocene of the south-central Pyrenees) from its inception to its final infill with alluvial deposits. The stacking pattern shows facies changes primarily produced by thrust propagation toward the outer margin of the basin (see text for a more extensive discussion).

associations and relative water depth (Fig. 1). These systems include (1) *flood-dominated deltaic systems*, extending from delta-front to slope and base-of-slope regions (for a brief account on related fluvial sedimentation see Mutti et al., 1996), (2) *mixed depositional systems* in which turbidite-like bodies deposited by poorly efficient gravity flows are associated with deltaic sediments both vertically and laterally, and (3) *basinal turbidite systems* and associated hemipelagic deposits, which are found as the infill of foredeeps. The main facies and facies associations of these systems are summarized in the following sections.

3.2. Flood-dominated deltaic systems

3.2.1. Introduction

Basin-margin shallow-marine and shelfal successions of many foreland basins contain thick and laterally extensive accumulations of parallel-sided graded sandstone beds commonly containing HCS (Fig. 3). The stratigraphic importance of these deposits was first recognized by Goldring and Bridges (1973), who termed them 'sublittoral sheet sandstones' and suggested various origins including storms, tsunamis, floods, tides, rips and turbidity currents. In subsequent literature, these sediments have been generally interpreted as storm-dominated shoreface and shelfal deposits mainly because of the abundance of HCS (Walker, 1984; Duke et al., 1991). More recent work suggests that the origin of these sediments is more complex than previously thought and that non-actualistic processes have to be envisaged to account for their stratigraphic importance and sedimentologic characteristics (Mutti et al., 1996; Myrow & Southard, 1996).

As pointed out by Myrow and Southard (1996) and Myrow et al. (2002), the large amounts of graded shelfal sandstones with HCS imply the suspension of similarly large amounts of sediment at the shoreline to generate density currents that can carry this sediment perpendicular to the shore for long distances. These authors suggest the possibility that the process might be associated with the 'oceanic floods' of Wheatcroft (2000), i.e. a process during which large quantities of fine-grained sediment are rapidly introduced to the sea by small rivers in flood and the river-sea system responds to the same storm event.

The genetic relations between these beds and rivers in flood have been documented by Mutti et al. (1996, 2000) through field mapping, detailed facies analysis and high-resolution stratigraphic correlations from a significant number of ancient depositional systems. In particular, vertical and lateral stratigraphic relations observed in these systems suggest that the vast majority of shelfal graded sandstone beds with HCS grade landward into flood-dominated fluvial systems with or without intervening estuarine zones showing evidence of reworking by wave action or tidal currents.

In such settings, fluvial floods generate sediment-water mixtures that enter seawaters as density-driven underflows, i.e. hyperpycnal flows in the sense of Bates (1953). Much of the sediments carried by these flows can escape river mouths and be transported farther seaward, thus increasing the sediment flux to shelfal regions. The shelfal graded sandstone beds with HCS deposited by hyperpycnal flows that could escape river mouth regions have thus been termed 'flood-generated delta front sandstone lobes' and are thought to record the sandy depositional zones of a broad spectrum of relatively small, coarse-grained and high-gradient fluvio-deltaic systems periodically dominated by catastrophic floods (Mutti et al., 2000).

The scenario outlined above is very similar to that envisaged by Milliman and Syvitski (1992) in their highly perceptive discussion of active margin sedimentation, i.e. settings where sediment flux to the sea is enhanced by high-elevation source areas close to the shoreline, lack of extensive alluvial and coastal plains, and the periodic flooding of small and medium-sized 'mountainous' rivers characterized by relatively short and high-gradient transfer zones. Mulder and Syvitski (1995) have introduced the term 'dirty rivers' to describe a limited number of modern rivers with low average discharge and small drainage basins, which are able to trigger weather-induced underflows during one or more periods of the year. However, even in the case of the Yellow river, the only modern large river that generates semi-permanent underflows at its mouth

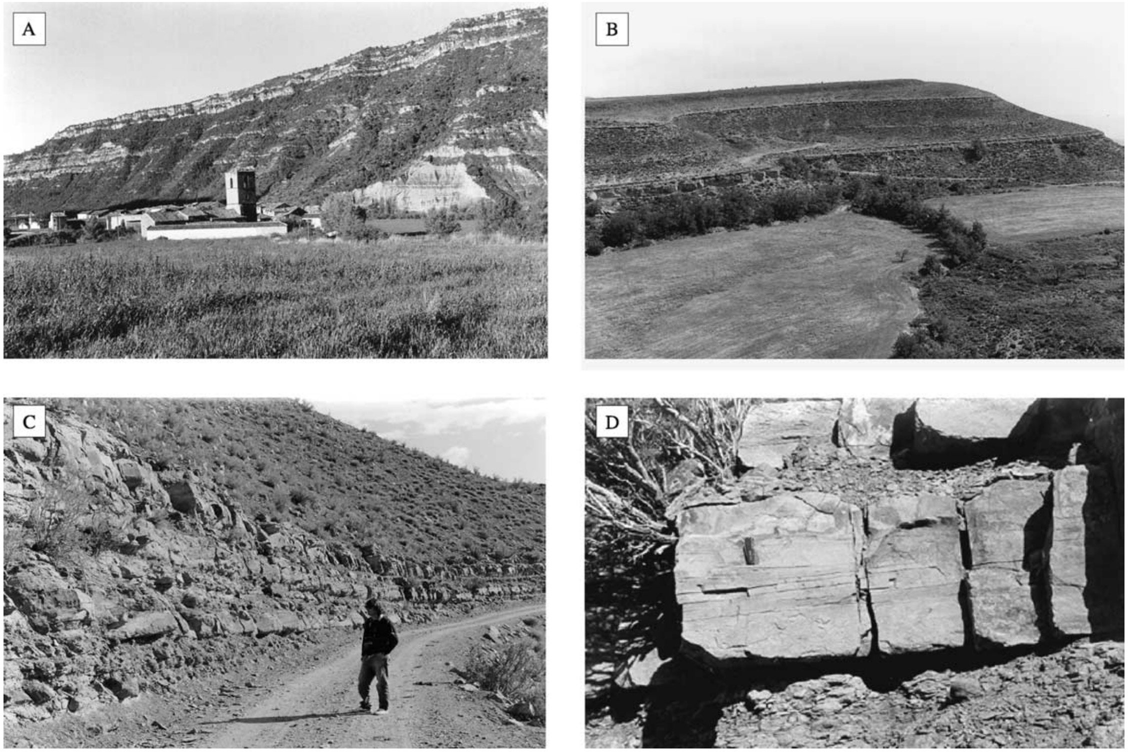

Fig. 3. (A) Flood-generated delta-front sandstone lobes of the Eocene Santa Liestra Group fan-delta system, south-central Pyrenees. Note the tabular geometry of sandstone packets. These sandstone lobes form a succession with a thickness of about 400 m grading northward (right in the photograph) into an equally thick succession of conglomerates and pebbly sandstones, which were deposited by catastrophic gravelly flows. (B) Sheet-like flood-generated fine-grained sandstone lobes interbedded with highly-bioturbated and fossiliferous shelfal sandy mudstones of a river-delta system. Lower Eocene Figols Group, south-central Pyrenees, near Tremp, Spain. (C) Typical aspect of an individual sandstone lobe, i.e. a metre-thick packet of parallel-sided graded sandstone beds with HCS, encased in finer-grained and highly bioturbated fossiliferous finer-grained strata. Jurassic Bardas Blancas Formation, Neuquen basin, Argentina. (D) Close-up of a sharp-based graded sandstone bed with HCS. Jurassic Bardas Blancas Formation, Neuquen basin, Argentina, (Knife for scale).

(Van Gelder et al., 1994), there is no evidence of substantial sand deposition in the delta front region.

Clearly, most ancient flood-dominated fluvio-deltaic systems can only be viewed in terms of catastrophic processes that were able to transport large amounts of gravel, sand and mud to delta-front and shelfal regions—a setting that is probably difficult to perceive from what we know from the Recent (Mutti et al., 1996).

The abundance of HCS in the upper part of the graded sandstone beds (see Figs. 3D and 5A) indicates deposition from bipartite hyperpycnal flows in which the more diluted upper part contains an oscillatory component (De Celles & Cavazza, 1992). Apparently these beds are very similar to those described by Myrow and Southard (1996) with the term 'wave-modified turbidite'.

The origin of the oscillatory component in the hyperpycnal flow can be related to different processes. For example, it can be associated with internal waves that develop along density interfaces (Wright et al., 1988; see also Nemec (1995)). The oscillatory component can be added by enhanced wind and wave energy in coastal waters due to storm events (see concept of 'oceanic floods' of Wheatcroft (2000)). We also speculate that in shallow and tectonically confined basins, large-volume and high-momentum hyperpycnal flows can generate 'sloshing' of sea water, thus strongly enhancing the oscillatory component of the flow (Mutti, 1992b; Mutti et al., 1996).

3.2.2. Deltaic deposits

Flood-generated delta-front sandstone lobes form thick and laterally extensive accumulations showing cyclic alternations between metre-thick, sheet-like sandstone bodies and muddier facies (Fig. 3). Spectacular examples of this kind of sedimentation can be observed in the upper Cretaceous and the Tertiary of the south-central Pyrenees, Spain, the Pleistocene of the southern Apennines, the Oligocene and Miocene of the Tertiary Piedmont Basin, Italy and the Jurassic and Cretaceous of the Neuquen basin, Argentina (Mutti et al., 1996).

Flood-generated delta-front sandstone lobes constitute a depositional element common to a broad spectrum of relatively small intergradational and coarse-grained fluvio-deltaic systems ranging between two end members, fan-deltas and river-deltas. In fan-delta systems, individual lobes form roughly tabular units extending from alluvial conglomerates to shelfal siltstone and mudstone. Coarse-grained facies are thus observed also in open marine environments. Conversely, in river-delta systems, lobes are generally finer grained since the coarse-grained sediment is trapped at river mouths. Facies tracts and inferred processes

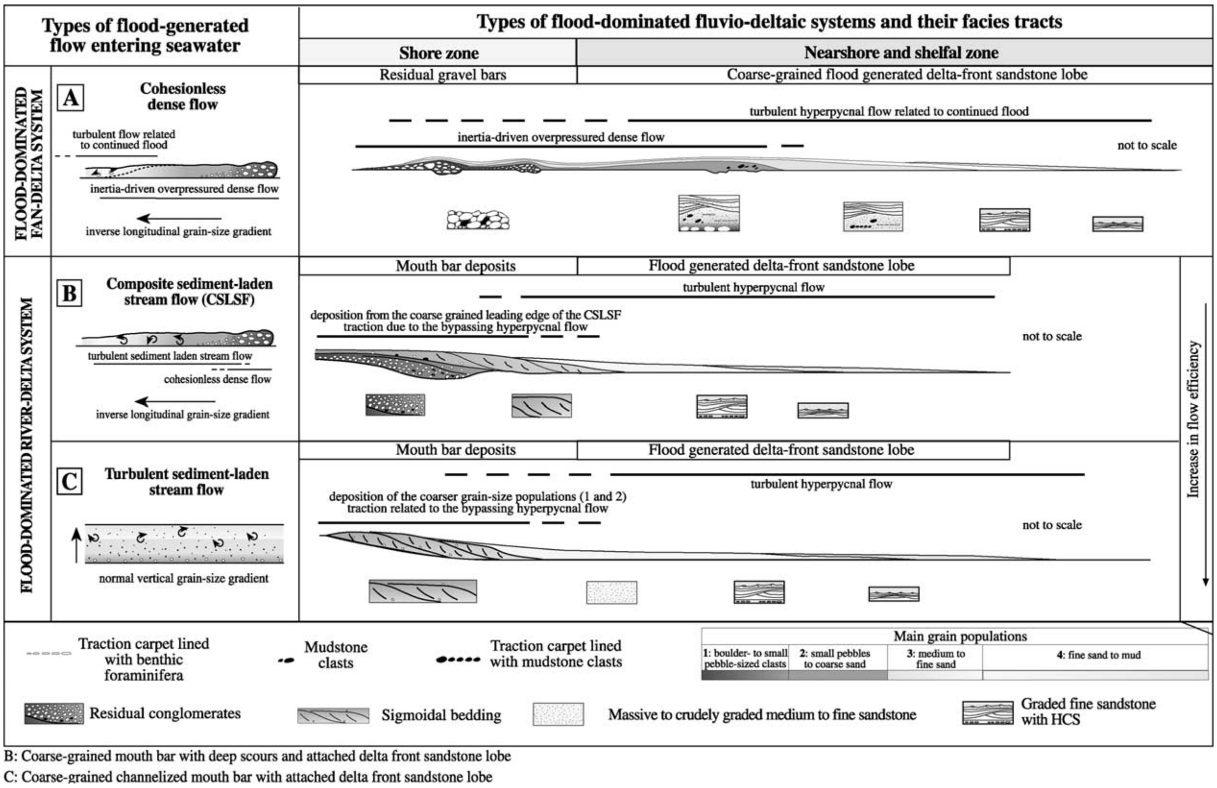

Fig. 4. Idealized diagrams showing facies and inferred processes of flood-dominated fan-delta and river-delta systems. The term dense flow is used herein in the sense of Norem et al. (1990). The term is thus essentially synonymous of 'debris flow', 'flowslide' or 'granular flow' as used by Morhig and Marr (2003), Shanmugam (2000) and Mutti et al. (1999). See sections on mixed and basinal turbidite systems for a more extensive discussion.

of these systems, as well as the terminology adopted in this paper, are summarized in Fig. 4.

In fan-delta systems, flood-generated dense flows (see Fig. 4 for a definition) are accelerated along the steep and confined reaches of incised streams and enter seawaters as catastrophic and relatively unconfined sediment–water mixtures. In such inertia-driven dense flows, moving under conditions of excess pore pressure, coarser grain populations tend to collect at the front of the flow giving way to a horizontally negative grain-size gradient (Major & Iverson, 1999; Sohn, Chul, & Kim, 1999; Mutti et al., 2000). Consequently, the gravelly portions of these flows will outdistance the sandy ones during downslope motion.

In the marine environment these dense flows generate sustained bipartite currents (hyperpycnal flows) in which an initially faster-moving dense flow probably moves, as suggested by Sohn et al. (1999), through a series of progressively finer grained surges (Fig. 4). Each surge is forced to deposit when the loss of excess pore pressure along its leading edge causes frictional freezing of the flow. As indicated by abundant rip-up mudstone clasts and shell debris (Fig. 5), substantial flow bulking (in the sense of Smith & Lowe, 1991, p. 64) must occur through bed erosion at the edges of each surge.

Basinward, sedimentation is dominated by the upper and more dilute part of each hyperpycnal flow, i.e. a trailing turbulent flow which is continuously fed from behind as long as the flood continues. The flow thickens in the shorezone due to mixing with seawater (Wright, 1977; McLeod, Carey, & Sparks, 1999) and must undergo additional thickening farther seaward through turbulence generated at the leading edge of the preceding dense flow. The flow must also increase its sediment concentration through (1) fall-out from the overlying suspension (cloud collapse in the sense of McLeod et al., 1999), (2) sediment escaping from the underlying granular flow due to the loss of pore-fluid pressure (Mutti et al., 1999), (3) sediment erosion from and turbulent mixing at the head of the basal flow (see Mohrig and Marr, 2003) and (4) substantial bed erosion taking place at the head of the inertia-driven basal flow (see later section on basinal turbidites). As a result, the lower part of the flow becomes a relatively high-density turbulent flow that will eventually bypass the frozen, coarser grained deposit of the basal dense flow (Fig. 6) and move farther basinward carrying its suspended load to more distal delta-front regions.

River delta systems are dominated by two intergradational types of flow, including (1) sediment-laden stream flows, and (2) composite sediment-laden stream flows (Fig. 4). Sediment-laden stream flows are common in relatively mature and low-gradient river systems and can be viewed as long-lived density-stratified turbulent flows in which sand and mud are essentially transported as suspended load. Composite sediment-laden streamflows

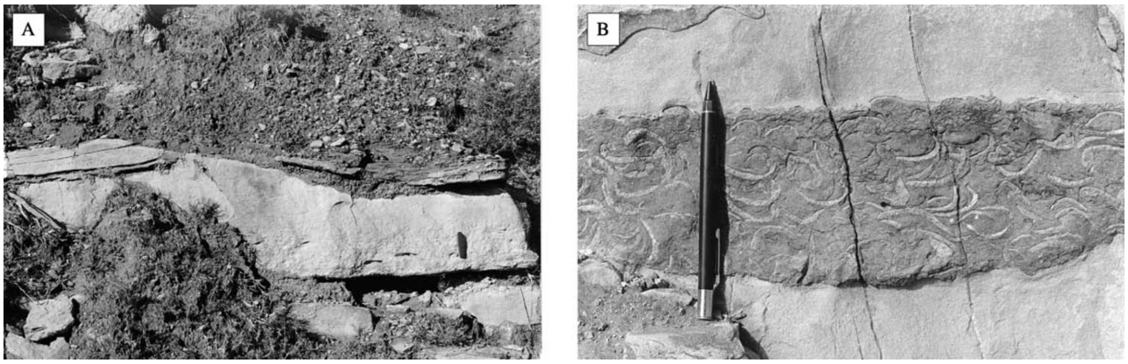

Fig. 5. (A) Example of a graded sandstone bed deposited by a hyperpycnal flow in the delta-front region of a flood-dominated fan-delta system. Note the basal coarse-grained division with abundant rip-up mudstone clasts and larger foraminifera sharply overlain by HCS. Mudstone clasts and fossil debris indicate that the basal division was deposited by a flow characterized by substantial bed erosion at its head. Knife for scale. Lower Eocene Santa Liestra Group, south-central Pyrenees, Spain. (B) Skeletal-rich (mostly valves of the pelecypod Trigonia) basal division of a flood-generated sandstone bed in the Jurassic Bardas Blancas Formation, Neuquen basin, Argentina, indicating bed erosion at the head of a hyperpycnal flow.

are short-lived flows, characterized by a marked inverse longitudinal grain-size segregation, which are common in relatively small, high-gradient and gravel-rich systems (Fig. 4b).

Sediment-laden stream flows and composite sediment-laden stream flows produce substantially different types of deposits, particularly at river mouths. As a consequence, mouth-bar deposits exhibit a great variability in terms of geometry and facies types, essentially recording locally prevailing conditions that may range from erosion and sediment bypass to deposition of the entire sediment load of fluvial outflows. Facies distribution patterns are thus mainly controlled by how much sand is trapped at river mouths and how much sand can escape this region through turbulent hyperpycnal flows, which can move farther basinward and deposit their sand load as delta-front lobes. This ratio can be considered in terms of efficiency of hyperpycnal flows exiting river mouths, i.e. the ability of these flows to carry their sediment load basinward, primarily controlled by their momentum, sediment concentration, discharge and duration (see below).

In this paper we restrict our discussion to systems dominated by relatively long-lived sediment-laden streamflows with high sediment discharge that build up systems with well-developed clinoforms and very distinctive mouth-bar and lobe elements. An excellent example of this kind of delta system is that of the Eocene Roda Sandstone in the south-central Pyrenees (Tinterri, 1999, with references therein).

Flow efficiency and water depth at river mouths exert a fundamental control on the local depositional setting. Because of their density and momentum, sediment-laden streamflows must enter seawater as inertia-dominated outflows, thus forming either axial or plane turbulent jets, depending on the depth of seawater at and seaward of the river mouths (Wright, 1977). Assuming a constant depth, the depositional setting is mainly controlled by flow efficiency (Fig. 7).

Flood-dominated mouth-bar deposits of this type of delta are characterized by facies types ranging from poorly sorted massive or crudely laminated very coarse-grained sandstone and pebbly sandstone to better sorted and finer grained sandstone exhibiting various types of internal stratification. Massive sandstone facies can be interpreted as the deposit of flows which underwent expansion at river mouths followed by the sudden gravitational collapse of their coarse-grained

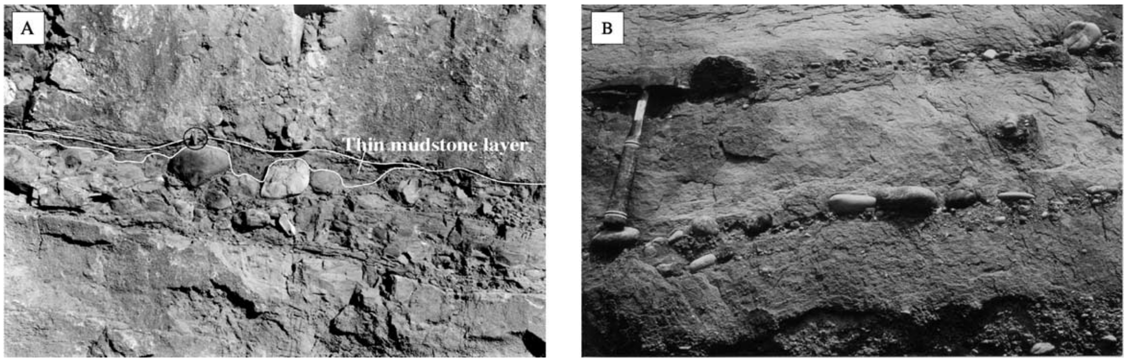

Fig. 6. (A) Conglomerates deposited by frictional freezing of gravelly dense flows in a flood-dominated fan-delta system. Note a pebble alignment resulting from extensive reworking and winnowing by bypassing turbulent flows capped by a thin mudstone layer. Encircled knife for scale. Lower Eocene Santa Liestra Group, south-central Pyrenees, Spain. (B) Pebble alignment produced by a bypassing hyperpycnal flow. During the falling stage of the flood, the pebbles are covered by a division of medium-grained sandstone with faintly developed laminae. Lower Molare Unit, Tertiary Piedmont Basin, Italy.

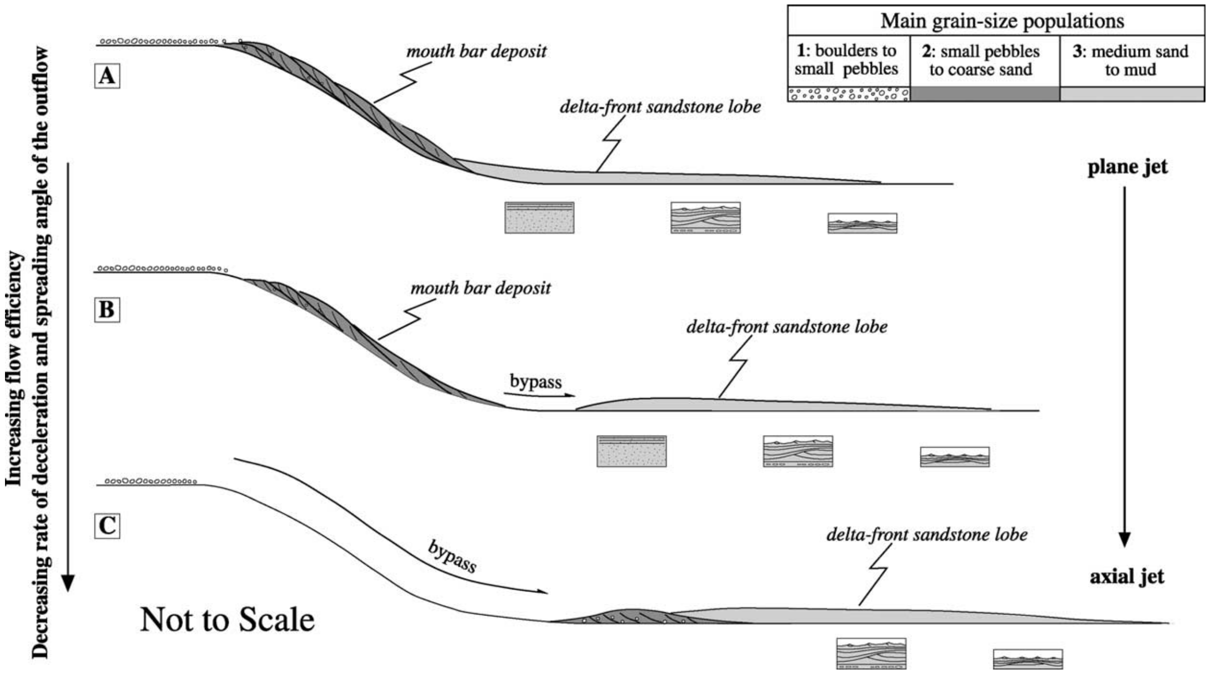

Fig. 7. Relationships between mouth-bar deposits characterized by downstream accreting sigmoidal bars and tabular delta-front sandstone lobes for increasing efficiency of river outflows, mainly controlled by flow momentum and duration, sediment concentration and lateral flow spreading, under constant depth conditions (see text for more details).

sediment load in the absence of substantial traction. In the case of higher outflow efficiency, coarse sediment collapsed at river mouths is transported as tractive bed load by sustained turbulent flows exiting river mouths as hyperpycnal flows. This process sets up the formation of tractional bedforms that are primarily controlled by water depth and the velocity and steadiness of the bypassing hyperpycnal flows.

Flood-generated sigmoidal cross bedding (Mutti et al., 1996) is the most characteristic expression of this process (Fig. 8). This type of bedding consists of cosets of sigmoidally shaped cross-stratified units that characteristically thin and flatten in a downcurrent direction. The upper boundary of each coset is a sharp erosional surface produced by the bypassing of the turbulent flow. In very shallow waters, cosets of sigmoidal units stack vertically, being bounded by essentially horizontal or slightly convex-upward surfaces; for increasing water depth, with a progressively more pronounced relief of clinoforms, these cosets form very distinctive downstream accreting units bounded by convex-upward surfaces conforming to the clinoform profile.

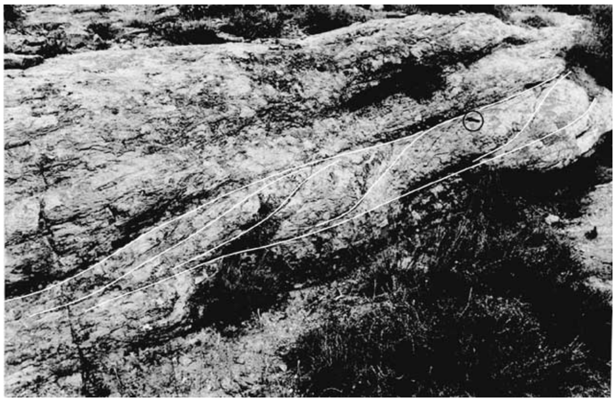

Fig. 8. Internally cross-stratified sigmoidal units constituting a downstream accreting deposit in a flood-dominated river-mouth bar. These sediments are a tractive bed-load deposit formed by bypassing turbulent flows exiting the river mouth. Note that individual sigmoidal units thin and flatten in a downcurrent direction (from right to left) as a result of progressive waning-flood conditions (see text for more details). Encircled knife for scale. Lower Eocene Figols Group, south-central Pyrenees.

As shown in Fig. 7, clinoform relief determines the water depth at which bypassing hyperpycnal flows start to deposit their sediment load in the lobe region. This depth is highly variable, ranging from a few meters where deltas prograde in very shallow marine environments to 100–300 m, where deltas prograde across deeper shelfal regions. Despite this difference, processes and depositional architectures remain the same, indicating trapping of coarser-grained sediment in shallow waters and bypassing of finer-grained sediment carried in suspension by turbulent hyperpycnal flows. These can reach clinoform toes and move greater distances seaward to deeper waters, depending on flow efficiency and local submarine topography (see also Pirmez, Pratson, & Steckler, 1998).

3.2.3. Prodeltaic mudstone wedges

Basinward of river mouths and proximal delta-front regions where the bulk of the sand is deposited, flood-dominated delta systems grade into prodeltaic mudstones with interbedded fine-grained and thinly-bedded sandstones

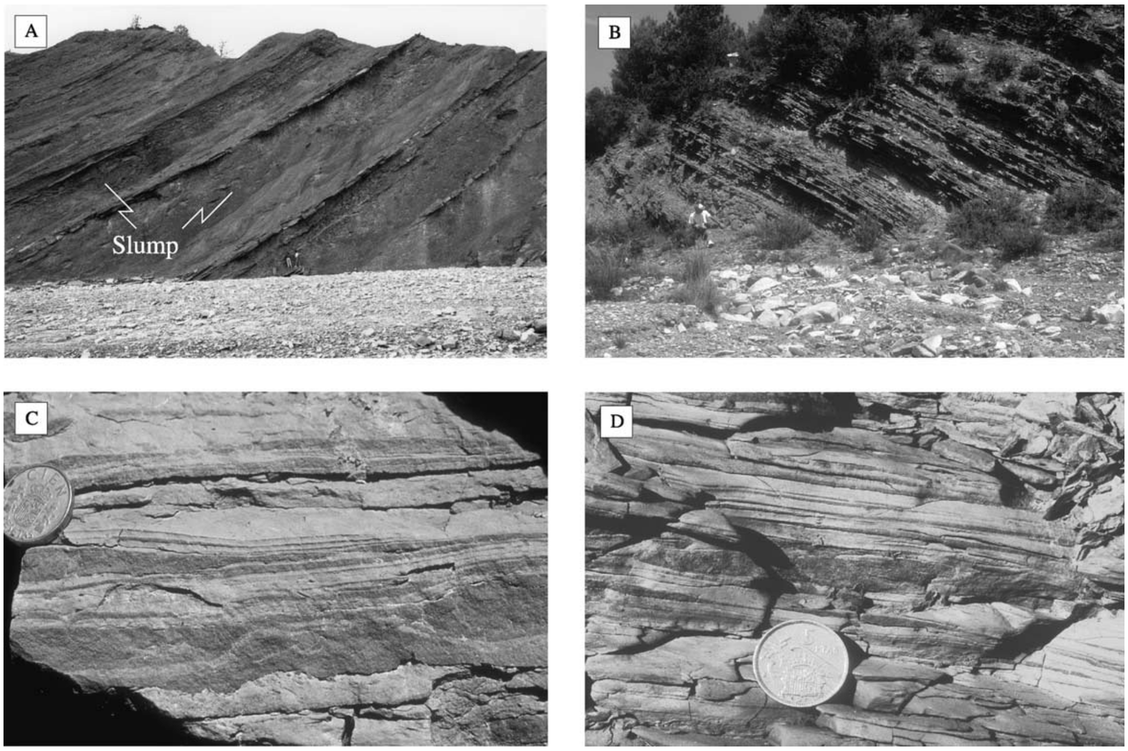

Fig. 9. Examples of stacking patterns and facies characteristics of prodeltaic wedges. (A) Shows the cyclic stacking pattern of mudstone and thin-bedded sandstone facies. Arrows indicate chaotic intercalations produced by slumping and sediment creeping. (B) Depict cyclic alternations of mudstone-dominated facies and thin-bedded sandstones forming metre-thick packets with good lateral continuity. (C) Shows the highly complex internal organization of current laminae suggesting unsteady flow conditions. (D) Depicts details of very thin sandstone/mudstone couplets deposited by deltaic plumes and/or dilute hyperpycnal flows. Lower Eocene Castissent Group, south-central Pyrenees.

(Fig. 9a and b). Depending on the local basin configuration, these fine-grained prodelta deposits can form on shelves or extend farther seaward beyond the local shelfedge to form thick slope wedges commonly characterized by pervasive sediment creep and slump features (Fig. 9a). These slope wedges thin basinward where they may interfinger with the upslope terminations of basinal sand-rich turbidite systems (Mutti, 1992a).

Prodeltaic mudstone wedges can be viewed as the zone of terminal depositional of deltaic systems recorded by a variety of facies types mainly deposited by hypopycnal flows (buoyant plumes), interflows and low-density hyperpycnal flows. Mainly on the basis of experimental work and numerical modeling, these processes have been amply discussed by several authors (e.g. Sparks, et al., 1993).

In ancient prodeltaic mudstone wedges of foreland basins, massive or finely-laminated mudstones record the background hemipelagic sedimentation from hypopycnal plumes during periods of predominantly normal river regimes. Alternating mudstone and thin-bedded siltstone and fine-grained sandstone form decimetre- to metre-thick packets interbedded with hemipelagic mudstones showing remarkably well-developed cyclic stacking patterns (Fig. 9a and b). Based on sandstone/mudstone ratio, sandstone bed thickness, and depositional structures within individual sandstone beds, these alternations comprise distinct subfacies.

Mud-rich subfacies (Fig. 9c and d) contain very distinctive millimetre- to centimetre-thick, graded siltstone/mudstone couplets. Siltstone commonly occurs as mm-thick layers and streaks which are either structureless or exhibit very thin horizontal or very low-angle cross laminae. These sediments are here interpreted as the product of sediment fallout in the absence of substantial traction. Silt-rich buoyant plumes and especially lofting (in the sense of Sparks et al., 1993) of dilute hyperpycnal flows seem the most plausible process for this kind of sedimentation.

Sand-rich subfacies (Fig. 9b) are laterally extensive packets of centimetre-thick sandstone and siltstone beds each overlain by a mudstone division. The internal structures exhibited by the sandy divisions of these beds indicate deposition mainly from traction-plus-fallout processes associated with highly unsteady turbulent flows. Most beds have a relative high sand/mud ratio, suggesting that large amounts of mud kept moving basinward carried by low-density hyperpycnal flows.

A correct understanding of the origin of thin-bedded and fine-grained sandstone facies, which are a volumetrically important component of the fill of foreland basins, appears to be crucial to basin analysis. These sediments can actually form in a variety of settings such as prodeltaic wedges, channel-levee complexes, and basin plains. Their correct interpretation should be based on careful facies analysis framed within stratigraphic correlation patterns established at a basinwide scale.

3.3. Mixed depositional systems

3.3.1. Introduction

The term ‘mixed systems’ is introduced herein to define relatively small and generally sand-rich depositional systems sharing several characteristics with basinal turbidites (see below), but differing from these by showing a more immature facies development (cf. ‘poorly-efficient turbidite systems’ of Mutti (1979)) and, most importantly, for their close vertical and lateral stratigraphic association with deltaic deposits. These systems can thus be viewed as immature, marginal and poorly efficient turbidite-like systems formed seaward of, but adjacent to feeder delta complexes. Consequently, these systems formed at shallower depth than that of basinal turbidites (see below). The importance of these systems was clearly perceived in earlier work by Chan and Dott (1983) and Heller and Dickinson (1985) with their model of delta-fed turbidite systems associated with submarine delta ramps. The term ‘delta-fed turbidites’ is not retained in this paper in order to avoid confusion since most basinal turbidites are also ultimately fed by deltaic systems and, most importantly, are characterized by facies, processes and depositional settings which are substantially different from those of mixed systems (see later).

Mixed systems constitute sand-rich facies associations (Figs. 10 and 11) confined within structural depressions generated by faulting and folding and are an important component of the fill of many ancient basin fills, particularly the Tertiary Piedmont Basin, northwestern Italy, the Ainsa basin of the south-central Pyrenees, Spain, the Neuquen basin, Argentina, and the uppermost stratigraphic part of the Marnoso-arenacea Formation in the northern Apennines, Italy (Mutti, Ricci Lucchi, & Roveri, 2002).

In foreland basins, mixed depositional systems are probably related to local tectonic uplift, which produces relative sealevel variations of probably low to moderate amplitude. These tectonically forced sealevel lowstands along basin margins cause steepening of the depositional profile and ensuing erosion and resedimentation of falling-stage and lowstand deltas through sediment failures and hyperpycnal flows into adjacent structural depressions. Once an equilibrium profile is re-established along basin margins and accommodation is resumed landward of the tectonic hinge-line, these delta-fed turbidite-like systems are overlain by a transgressive systems tract recorded by flood-dominated deltaic sandstones and prodeltaic mudstones.

Depending on many local factors (e.g. type of feeder deltaic system, origin, volume and textural composition of

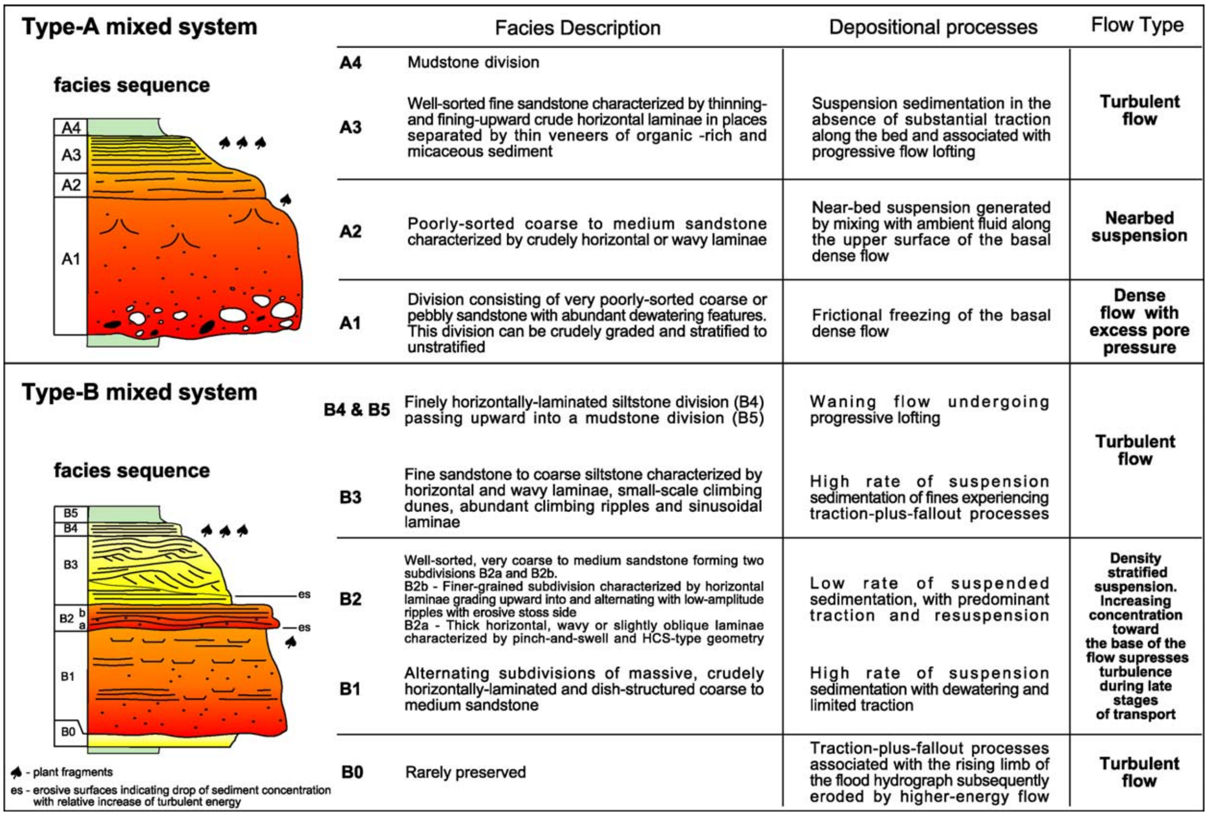

Fig. 10. Idealized facies sequences describing deposits associated with marginal mixed turbidite systems. Bed scale is in the order of a metre or so. See text for explanation.

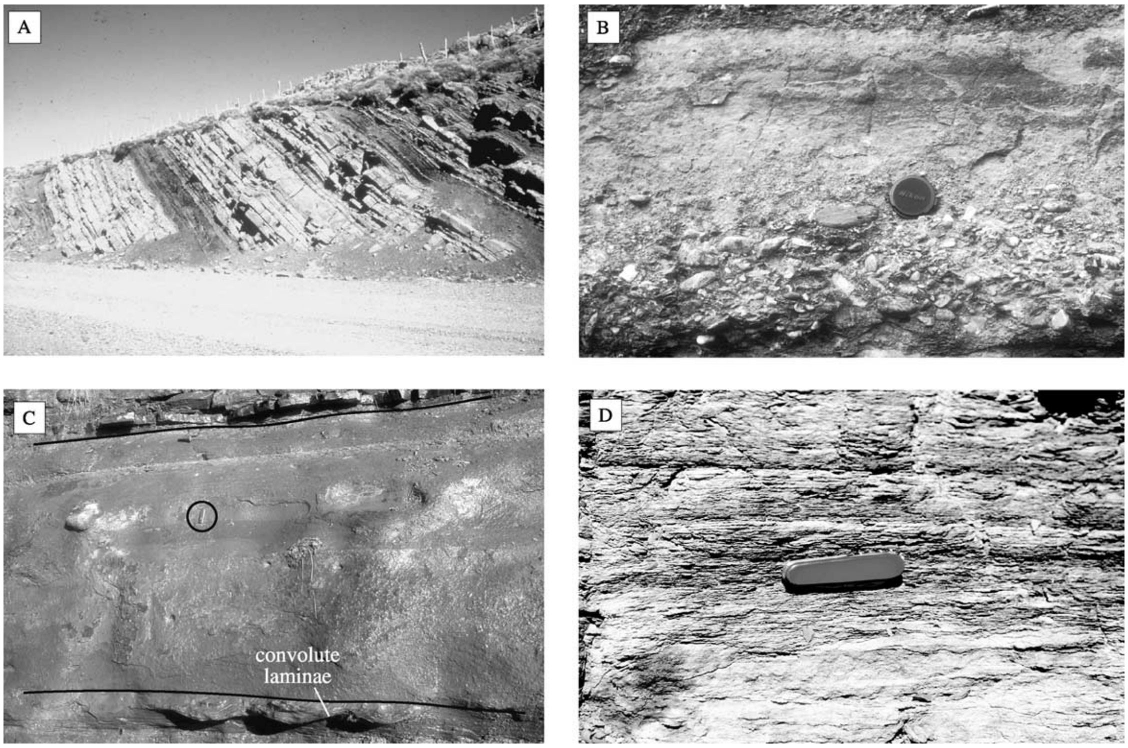

Fig. 11. Examples of bedding patterns and facies characteristics of type-A mixed systems. (A) Stacking pattern of a mixed turbidite system composed of alternating metre-thick sandstone lobes and finer-grained facies. The latter are locally fossiliferous (Jurassic Los Molles system, Neuquen basin, Argentina). (B) Graded pebbly sandstone bed. The lower conglomeratic division is interpreted as the deposit of a gravelly dense flow (A1 division). The upper sandstone division, showing faint horizontal laminae, is thought to represent the deposit of a near-bed suspension (A2 division), (see text for more details). Lower Miocene Noceto system, Tertiary Piedmont Basin, Italy. (C) Thick sandstone bed sharply resting on a finer-grained bed with convolute laminae. The thick sandstone bed is normally graded and primarily consists of horizontal laminae that thin and fine upward and are eventually capped by a thin mudstone division. Beds of this type are here interpreted as the deposit of relatively dense near-bed suspension (see text for details) (Lower Miocene Noceto system, Tertiary Piedmont Basin, Italy). (D) Bed entirely composed of fine sandstone characterized by horizontal laminae passing upward into a mudstone division without intervening ripple laminae. (Jurassic Los Molles system, Neuquen basin, Argentina) (see text for more details).

sediment gravity flows, basin size and physiography), mixed systems differ greatly from each other in terms of size, geometry, internal architectural styles, and facies and facies associations. These systems can thus exhibit all the gradational types from short-lived, small, coarse-grained and extremely poorly organized deposits derived from sediment failures and trapped in adjacent small fault-bounded depressions to relatively longer lived and larger systems deposited by more mature flows and characterized by channel and lobe elements.

Mixed systems may form at variable water depths, thus showing facies types that are intergradational between delta-front sandstone lobes and progressively deeper-water deposits lacking evidence of shallow-marine conditions. In relatively shallow waters, these sediments may contain HCS and abundant skeletal debris and be interbedded with highly bioturbated and fossiliferous finer-grained shelfal deposits. For increasing water depth and sediment gravity flow efficiency, mixed systems tend to develop a more basinal turbidite-like character, are interbedded with prodeltaic mudstones and may contain chaotic intercalations produced by slumping and sediment creeping. In some cases, these systems exhibit facies and facies associations that become very difficult to distinguish from those of basinal turbidites if these sediments are not carefully examined in terms of component facies and framed within their stratigraphic and structural setting.

An extensive review of the great variety and complexity of facies types and processes of mixed systems is beyond the purposes of this paper (see Mutti et al. in preparation). For this reason, we will herein briefly consider only the two types of systems that probably constitute the end-members of a broad spectrum of intergradational systems. These two types of systems are referred to herein as type-A and type-B systems and briefly discussed below.

3.3.2. Type-A systems

The basic facies types observed in type-A systems are summarized in Fig. 10 showing the complete vertical sequence of depositional divisions of an ideal bed. The sequence exhibits a basal A1 division consisting of a crudely graded and internally unstratified or crudely stratified subdivisions commonly consisting of very poorly sorted coarse-grained sandstone or pebbly sandstone with dewatering features formed during and shortly after deposition (Fig. 11b and c). This basal division is overlain

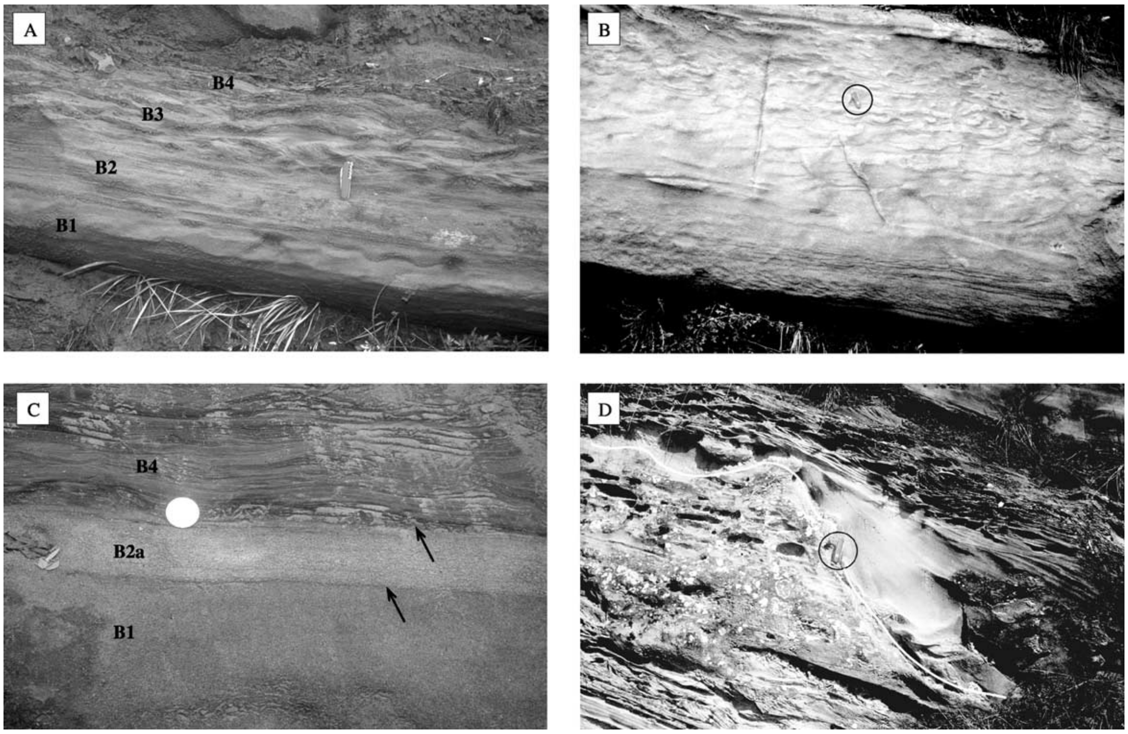

Fig. 12. Some typical facies characteristics of type-B mixed systems (see text and Fig. 10 for a more extensive discussion). (A) Sandstone bed showing the complete vertical sequence of depositional divisions deposited by a sand-laden turbulent hyperpycnal flow. (B) Dish-structured and crudely horizontally laminated medium to coarse sandstone division (B1 division of Fig. 10). Note vertical Ophiomorpha burrow. Encircled knife for scale. (C) Bed showing the sharp contacts bounding below and above the lighter and more resistant B2a division. The lower contact is thought to be the product of winnowing and resuspension of the sediment of the underlying B1 division; the upper contact, which is also marked by a distinct break in grain size, is interpreted as a bypass surface. The overlying B4 and B5 divisions record the waning flow stage of the hyperpycnal flow. (D) Deep intrabed scour produced by a highly turbulent flow that cuts into the underlying B1 and B2 divisions. Note small granules and sparse mudstone clasts aligned parallel to the steep scour wall. The scour was first infilled with subtly inclined thick laminae followed upward by finer-grained horizontal and climbing-ripple laminae during waning-flow stage. All photographs are from the Miocene Marnoso-arenacea Formation, Northern Apennines, Italy.

by a relatively thin division of crudely horizontally- or wavy-stratified and poorly sorted coarse- to medium-grained sandstone (A2). The latter grades into a division of better sorted and finer grained sandstone characterized by thinning- and fining-upward horizontal laminae (A3) which may be separated by thin veneers of organic-rich and micaceous sediment (Fig. 11d). A3 Division grades into a thin mudstone division (A4) without intervening current ripples.

The interpretation of this sequence of depositional divisions is shown in Fig. 10 and suggests deposition from a tripartite flow consisting of (a) a basal and faster moving inertia-driven dense flow with excess pore pressure, (b) a near-bed suspension, and (c) an upper and dilute turbulent flow. A1 Division represents deposition from the basal flow due to frictional freezing (Fig. 11b and c). A2 Division records deposition from the near-bed suspension, i.e. a relatively dense flow formed by turbulent mixing with ambient fluid at the leading edge of the basal dense flow during its motion. Near-bed suspensions, herein thought to represent a fundamental and very common process in sediment gravity flows (see later section on basinal turbidites), can be considered as a transitional flow stage during which the particle support mechanisms of inertia-driven dense flows (high sediment concentration and excess pore pressure) are progressively replaced by turbulence developed in the mixing zone at the head of the dense flow. Smaller particles can be incorporated as fully suspended load in the upper turbulent flow. Conversely, coarser particles are too large to be fully suspended and thus move as suspended or intermittently suspended load near the bed, being probably supported within the flow also by grain-to-grain collision due to the relatively high sediment concentration. The sediment of near-bed suspension is characteristically composed of essentially aggradational horizontal laminae showing a marked thinning- and fining-upward trend.

The development of A3 division depends on how much fine sediment can be incorporated as fully suspended load within the upper turbulent flow. This is controlled by the amount of turbulent energy developed in the mixing zone and the amount of fines which are contained within the frontal part of the dense flow undergoing turbulent mixing and elutriated from the dense flow. A3 Division can be thus either suppressed or developed through thinning- and fining-upward horizontal laminae commonly containing very abundant plant fragments. The general lack of ripple laminae suggests that laminae of A3 division mainly record suspension sedimentation in the absence of substantial traction along the bed due to progressive flow lofting.

We suggest that this kind of sequence of depositional divisions essentially records deposition from parental dense flows containing small proportions of fines and which were not sufficiently accelerated during their downslope motion to entrain fine-grained sediment through bed erosion. As a result, these dense flows cannot generate affiliate turbulent flows (in the sense of Mohrig and Marr, (2003)) of sufficient efficiency to produce extensive development of fine-grained facies characterized by traction-plus-fallout divisions.

3.3.3. Type-B systems

Type-B mixed systems are characterized by facies types predominantly deposited by sustained sediment-laden turbulent flows generated by hyperpycnal flows exiting river mouths. Where completely developed, beds deposited by these flows include five main depositional divisions herein termed, in ascending stratigraphic order, B1, B2, B3, B4, and B5 (Figs. 10 and 12a).

B1 division consists of alternating massive, crudely horizontally stratified, and dish-structured sub-divisions of very coarse to medium sandstone (Fig. 12b). Rapid deposition from an overlying suspension is suggested by the abundance of dewatering features that can form before, during or immediately after deposition of the overlying divisions. The B1 division commonly consists of relatively poorly sorted deposits and the upper part may contain organic-rich intervals, suggesting the ascent of plant fragments associated with dewatering. The B2 division sharply rests on the underlying division and consists of well-sorted coarse to medium sandstone forming two distinct sub-divisions. The lower B2a subdivision consists of relatively thick horizontal, wavy or low-angle cross laminae that are characterized by a pinch-and-swell geometry, similar in some respects to those observed in shallower water HCS (Fig. 12c). The origin of this sub-division is thought to be associated with reduced sediment fallout and extensive traction with resuspension and winnowing of sediment from the underlying division. Higher in the B2 division (B2b), medium and coarse sandstone contains horizontal laminae grading upward into and alternating with small-amplitude ripples with pronounced stoss-side erosion. The suggested origin of this sub-division is due to a predominantly tractional process occurring along the bed during relatively low rates of sediment fallout from an overlying turbulent flow. The B3 division is composed of very fine sandstone and coarse siltstone exhibiting thin horizontal laminae and particularly small-scale climbing dunes and abundant climbing ripples and associated sinusoidal laminae. The B4 division consists of thinly horizontally laminated siltstone and organic-rich mudstones passing upward into a mudstone division (B5). B3 and B4 divisions suggest increasing sediment fallout and decreased traction with time associated with the final waning stage of a turbulent flow carrying only fine-grained sediment as suspended load.

We interpret these beds as the deposits of sand-laden hyperpycnal flows that experienced an overall waning process through time punctuated by changes in depositional characteristics primarily due to changes in flow discharge and sediment concentration (Fig. 10). The hyperpycnal origin of these flows is well documented by the preservation of finer-grained rising-limb flood deposits at the base of some B1 divisions (B0 division, Fig. 10). A careful examination of the depositional divisions and their bounding surfaces within each flood unit permits the recognition of two very important stages related to changes in sediment concentration with time. The first surface is that separating B1 from B2 divisions; the surface is commonly erosive suggesting increased turbulent energy as a result of lowered sediment concentration (Fig. 12c). The second deeply erosive surface can be found within the B3 division separating the basal horizontally and wavy laminae from the overlying small-scale climbing dunes and climbing ripples. This surface records a dramatically decreased sediment concentration corresponding to the waning limb of a flood, which still retains sufficient free turbulent energy to deeply erode the underlying divisions and resuspend part of their sediment (Figs. 10 and 12d). The occurrence of many bypass surfaces and the complex internal structure of current-ripple divisions indicate that highly unsteady flow conditions dominate the waning-stage of most flows during deposition of type-B mixed systems.

3.4. Basinal turbidite systems

Impressively thick and laterally extensive sedimentary prisms of basinal turbidite sandstones crop out in many foreland basins. These prisms, formed in front of advancing thrust systems (Fig. 1), constitute an integral part of the external folded foreland of most orogenic belts as, for instance, in the Alps, the Apennines, the Carpathians, the Hellenides, and the south-central Pyrenees. The concept of turbidites (Migliorini, 1943; Kuenen & Migliorini, 1950), the Bouma sequence (Bouma, 1962), and the early turbidite facies and fan models of Mutti and Ricci Lucchi (1972), as well as many other important concepts on turbidite sedimentation, originated from outcrop studies carried out in these synorogenic strata. As indicated by angular unconformities along basin margins (Mutti, Seguret, & Sgavetti, 1988) and the impressive volumes of sand and mud infilling these turbidite basins, sediment flux to the sea must have been dramatically increased by the tectonic uplift of the source areas and therefore by tectonically forced sealevel lowstands of considerable amplitude.

The turbidite fill of foredeep basins, where the chief paleocurrent direction is generally oriented parallel to the basin axis, typically consists of three main elements (Mutti & Normark, 1991; Mutti, 1992a; Mutti et al., 1999). These include, (1) large-scale submarine erosional features, with relief up to 500 m and length up to 10–15 km that acted as conduits for turbidity currents, (2) sandstone lobes formed at the exit of these conduits, and (3) finer-grained basin-plain deposits formed in the distal and ponded sector of the basin.

This setting highlights two fundamental aspects of foredeep turbidite sedimentation. Firstly, turbidity currents must have been highly erosive when moving along their conduits and have thus entrained considerable amounts of fines through bed erosion. Secondly, large-volume and highly-efficient sand-laden currents were required to achieve the great runout distances (recorded by the lateral continuity of individual sandstone beds) and deposit the impressive accumulations of tabular sandstone lobes and associated basin-plain deposits.

Fig. 13 shows the main types of facies observed along the axis of a foredeep basin depicting the transfer and the depositional zones of an idealized turbidity current. Although turbidity currents can experience local important velocity variations dictated by submarine topography and these variations may cause erosion, bypass or deposition resulting in important local facies changes (Kneller & McCaffrey, 1999), it is clear that these local variations can only be appreciated if compared with an ideal along-axis evolution of facies and processes that should be taken as a standard reference.

As indicated in Fig. 13, facies types show an overall fining in a downcurrent direction and can be subdivided into four main groups, each defined by a distinctive grain population. The grade classes include: (A) boulder- to small-sized clasts, (B) small pebbles to coarse sand, (C) medium to fine sand, and (D) fine sand to mud. The four grain-size populations coincide with those used by Lowe (1982), Mutti (1992a) and Mutti et al. (1999) in their facies classification schemes.

The facies tract of Fig. 13 is here interpreted as the deposit of a bipartite turbidity current in which a dense basal flow—mainly impelled by inertia forces under conditions of excess pore pressure—was initially moving faster than an overlying and more dilute turbulent flow generated by mixing at the head of the dense flow (Sanders, 1965; Ravenne & Beghin, 1983; Norem, Locat, & Schieldrop, 1990; Mutti et al., 1999; Mohrig & Marr, 2003). When the dense flow decelerates due to progressive dilution and mixing with ambient water combined with the elutriation of the trailing edge of the flow, the turbulent flow bypasses the dense flow and moves basinward over a much greater distance (Norem et al., 1990). This general flow evolution is shown in the scheme of Fig. 14, which is inspired by the work of Ravenne and Beghin (1983) and Norem et al. (1990). Our scheme further subdivides dense flows into gravelly and sandy flows and shows that, upon their freezing, gravelly flows are bypassed by sandy dense flows which are characterized by a considerably greater runout distance largely controlled by their ability to generate and maintain excess pore pressure (Norem et al., 1990; Mutti et al., 1999; Tinterri, Drago, Consonni, Davoli, & Mutti, 2003).

The term ‘dense flow’ (see also previous sections) is used herein for the sake of simplicity to denote highly concentrated mixtures of sediment and water moving close to the bed. These mixtures are referred to in the literature with a variety of terms such as ‘debris flow’, ‘sandy debris flow’, ‘high-density turbidity current’, ‘hyperconcentrated flow’, ‘granular flow’ and ‘flowslide’. All these terms essentially denote dense flows with viscoplastic behavior (non-Newtonian flows), which eventually transform into turbulent flows during their downslope motion. Both types of flow are herein considered to be an integral part of a turbidity current in the sense of Kuenen (1965), in Sanders, (1965) (see Mutti et al., 1999 for an extensive discussion), though other authors (Shanmugam, 2000; Mohrig & Marr, 2003) suggest that the term ‘turbidity current’ should be restricted only to the turbulent flow. Following Kuenen’s definition bears considerable advantages not only for a more stable terminology, but, and most importantly, because it permits us to treat as genetically linked facies a broad spectrum of lithologies ranging from conglomerates to mudstones (Mutti, 1992a; Mutti et al., 1999). The four grain populations of Figs. 13 and 14 tend to be transported and deposited by turbidity currents as naturally distinct entities, thus forming similarly distinct facies groups. The first two populations move within a dense flow; the third population initially moves within a dense flow, but can be incorporated as suspended load into the overlying turbulent flow; the fourth population is the typical suspended load of fully turbulent flows.

The deposits of gravelly dense flows include a variety of facies types whose characteristics are primarily controlled by the original textural composition of the parental flow and the amount and type of sediment incorporated through bed erosion during flow motion. These deposits are essentially disorganized mixtures of cobbles, pebbles, and coarse sand floating in a sandy mudstone matrix (the classic pebbly mudstone facies, commonly referred to as a debris-flow deposit or debrite and herein termed F2 facies, see Fig. 15a). Bed-erosion produced at the head of gravelly flows is documented by the common abundance of rip-up mudstone clasts, ranging in size from mm-scale chips to m-scale blocks, incorporated in the deposit (Fig. 15a). Equally common are clast-supported conglomerates (F3 facies of Fig. 15b) which are herein interpreted as a record of frictional freezing at the leading edges of gravelly flows. These clast-supported conglomerates can form amalgamated and extensively scoured units, with the local development of crudely developed downstream accreting bars, or are found as isolated units characterized by a lenticular convex-upward geometry suggesting a lobate planform of the original flow. These lenses typically contain an inner core made up of a mudstone-clast breccia (Fig. 16) recording bed erosion at the head of the gravelly flow.

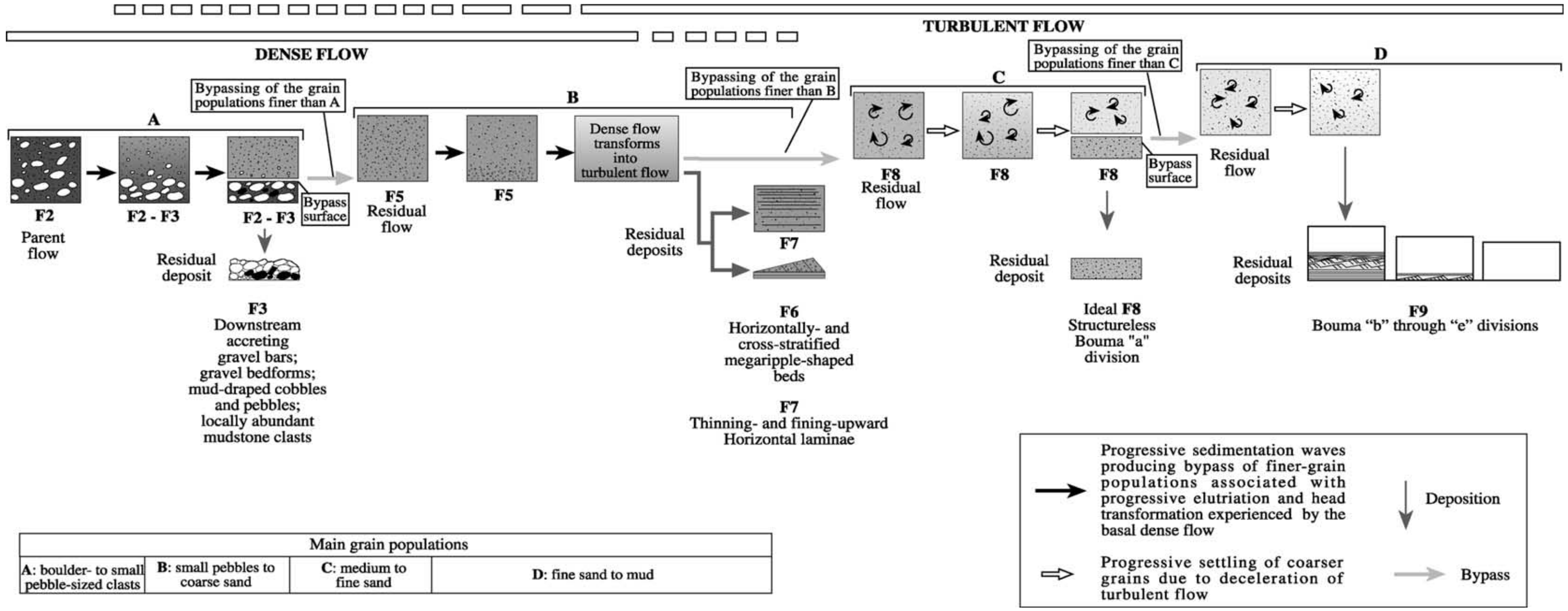

Fig. 13. Facies and inferred processes associated with an ideal bipartite turbidity current flowing along an elongate and flat axial zone of a foredeep basin (slightly modified from Mutti et al. (1999)).

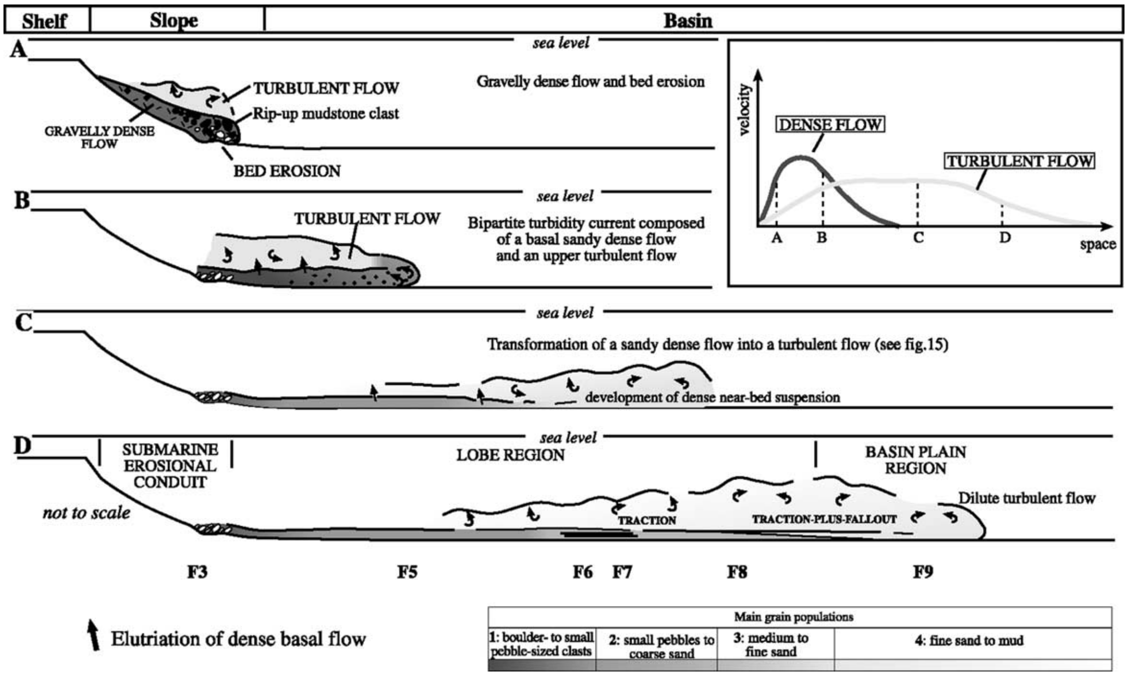

Fig. 14. Main erosional and depositional processes associated with the downslope evolution of a turbidity current. The current evolves from an inertia-driven gravelly dense flow moving under conditions of excess-pore pressure to a quasi-steady turbulent flow. Note in the insert in the upper right of the figure the different velocity-space relations of the inertia-driven dense and turbulent flows (slightly modified from Mutti et al. (1999) to which the reader is referred for more details).

The abundance of out-sized mudstone clasts indicates that gravelly dense flows must become highly erosive when accelerating along submarine conduits, being thus responsible for their deepening and widening together with sediment failure from the conduit walls. Mudstone clasts derived from these processes are progressively disaggregated within moving gravelly flows and substantial amounts of fine sediment are thus incorporated as suspended load within the upper turbulent flow.

The deposits of sandy dense flows bypassing the zone of deposition of the preceding gravelly flows are characterized by poorly sorted massive or graded divisions (F5 of Fig. 15c and e; see Mutti (1992a) for typical examples) forming thick and laterally extensive units that can be traced for several kilometers in a downcurrent direction. F5 deposits contain abundant mudstone clasts, suggesting substantial bed erosion also at the head of these flows (Mutti, 1992a), and display a variety of syn- and post-depositional dewatering features indicating that transportation and deposition took place under conditions of excess pore pressure (see above).

Facies relationships indicate that the final transformation of a sandy dense flow in a turbulent flow takes place in two different and probably intergradational ways. Facies and inferred processes associated with this transformation are shown in Fig. 17. In the first and most common type of transformation (upper part of Fig. 17), the massive distal deposit of a sandy flow (F5) is overlain by and pass laterally into crudely horizontally-laminated divisions made up of medium and coarse sandstone (F7 of Figs. 15c,d, and 17); within each bed, these laminae are distinctly fining and thinning upward and are sharply capped, through a break in grain size, by a thin ripple-laminated division deposited by the dilute tail of the turbidity current in its final waning stage. In the second case, the massive F5 division is either erosively capped or entirely replaced by relatively well-sorted coarse-sandstone divisions whose geometry can vary from highly discontinuous lenses, bounded by erosional surfaces and containing out-size mudstone clasts, to more laterally continuous divisions exhibiting plane bed and megaripple stratification (F6 of Figs. 15e,f, and 17; see details in Mutti (1977) and Mutti and Normark (1987), their Fig. 15). In this case, these coarse-grained divisions are generally sharply capped, through a bypass surface, by finer-divisions deposited by the dilute tail of the turbidity current.

F7 divisions are herein interpreted as the deposit of a near-bed suspension (see previous sections on mixed depositional systems) generated by progressive turbulent mixing at the head of a sandy dense flow with relatively low rates of deceleration. Turbulent energy developed through mixing is not sufficient to fully suspend the entire sediment load carried by the head of the dense flow. As a result, only the smaller particles can be fully suspended and transferred to the upper and bypassing fully mixed turbulent flow and thus carried farther downcurrent. Conversely, coarser particles form a relatively thin and density-stratified near-bed suspension in front of and above the transforming head of the dense flow. Deposition from these near-bed suspensions is apparently controlled by the settling of progressively finer-grained particles with limited traction.

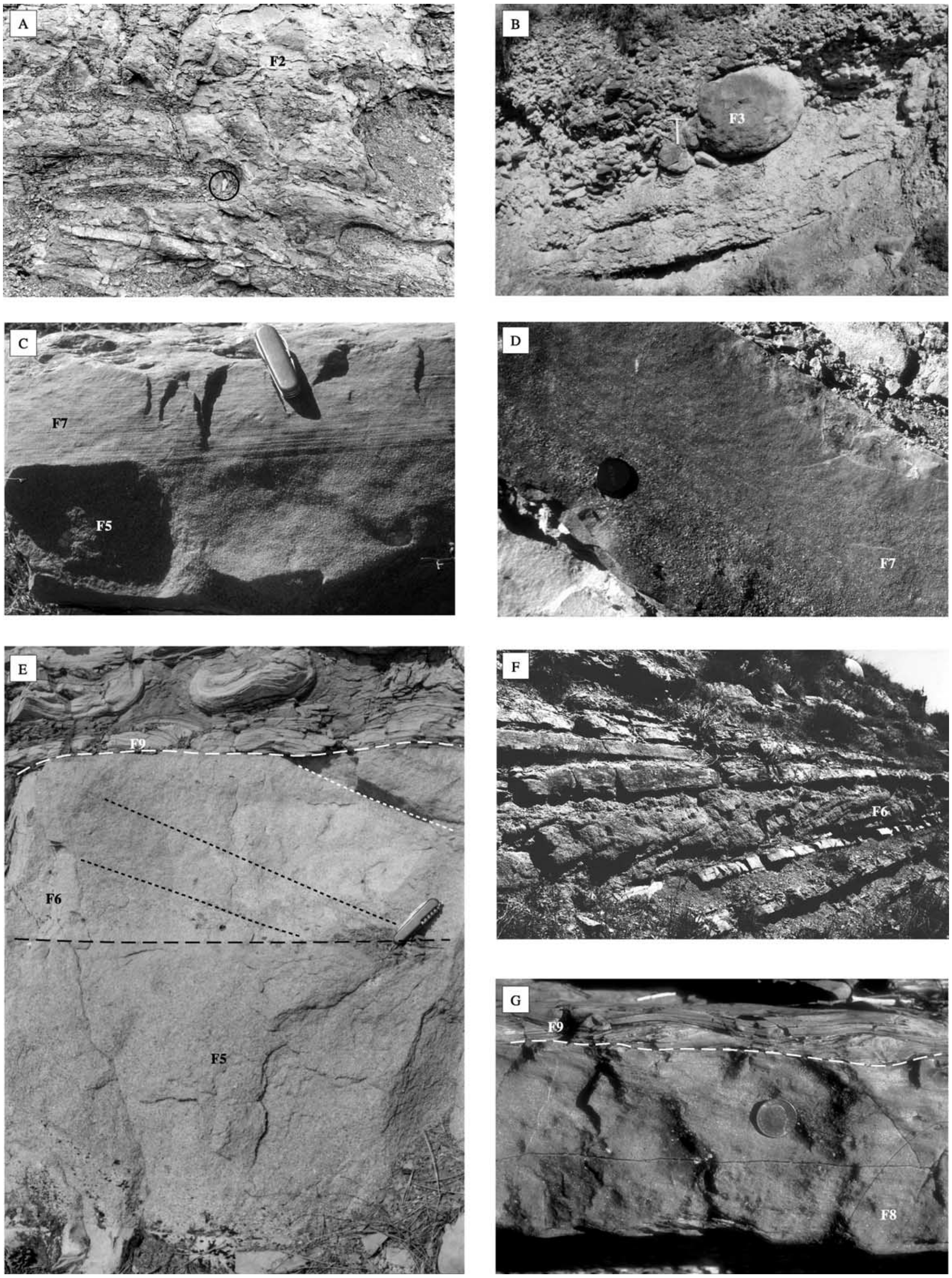

Fig. 15. Some basic facies of basinal turbidite systems (see text and Figs. 13 and 14 for an extensive discussion). (A) Highly disorganized deposit (F2 facies) frozen during early stages of downslope motion. The bed consists of a sandy mudstone with abundant granule- and small pebble-sized particles containing abundant large blocks of mudstone and thin-bedded sandstone. These blocks, most of which are highly contorted due to soft deformation, float within a matrix which is thought to represent very closely the original parental flow. These deposits document the highly erosive character of inertia-driven gravelly flows within submarine conduits. Encircled knife for scale. (B) Clast-supported conglomerates (F3 facies) recording the frictional freezing of the elutriated heads of gravelly flows within submarine conduits. (C) Horizontally-laminated sandstone division (F7 facies) abruptly overlying a massive and poorly sorted F5 division. The laminated division is thought to result from the resuspension of the finer-grained sediment of the underlying division during the development of a dense, near-bed turbulent suspension that progressively replaces the basal dense flow in a downcurrent direction due to mixing with ambient fluid and increasing turbulent energy. (D) F7 facies constituting the main division of an overall graded turbidite sandstone bed. Note that, in this case, the original F5

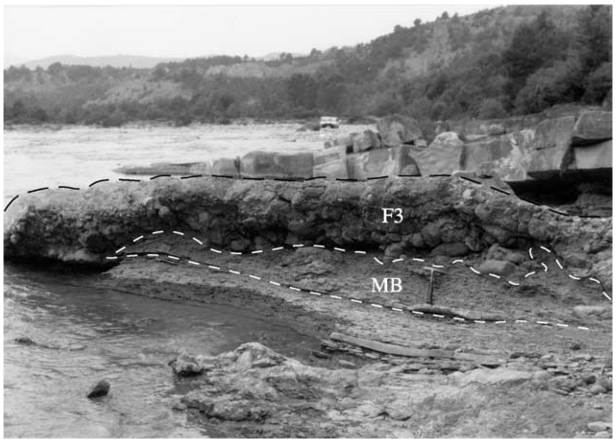

Fig. 16. The bed shown in this photograph consists of a basal lenticular division made up of a mudstone-clast breccia (MB) with a coarse-sandstone matrix which is overlain by a clast-supported division consisting of pebbles, cobbles and some boulders (F3 division). The F3 deposit is sharply overlain by fine-grained current-ripple division deposited by the dilute tail of the turbidity current. The MB division is here thought to represent the product of extensive bed erosion at the head of a gravelly flow. Both processes were recorded by frictional freezing of the gravelly flow. See text for more details. Eocene Hecho Group, south-central Pyrenees, Spain.

F6 deposits are here interpreted as the result of sudden deceleration of the dense flow, probably forced by subtle depositional topography and favoured by the characteristics of the dense flow (textural composition, degree of liquefaction, etc.), followed by flow expansion and full turbulent mixing (see hydraulic jump of Mutti (1977), and Mutti and Normark (1987)). Turbulent energy developed by the process produces extensive bed erosion and can fully suspend much of the sediment carried by the head of the dense flow except for the coarsest particles. The latter keep moving as bedload at the base of the bypassing turbulent flow forming distinctive tractive bedforms recording progressive waning flow conditions expressed by the vertical succession of planebed, megaripple and ripple bedforms. When fully developed, megaripples have a typical 3D geometry, height of about 17 cm, and wavelength up to 3 m, suggesting reworking from sustained and large volume turbulent flows. The fine-grained deposits of the dilute tail of the turbulent flow usually cap these coarse-grained sediments through a marked grain-size break indicating a phase of sediment bypass.

Basinward from the zone of transformation of dense flows into turbulent flows, sedimentation in ancient turbidite foreland basin appears to be dominated by facies types essentially deposited by turbulent flows. The volume of sand and mud deposited in these distal basin regions depends on the efficiency of turbulent flows, which is primary controlled by the amount of fines these flows can incorporate as suspended load through bed erosion, elutriation of the dense flow, and mixing at the head of the dense flow during its downslope motion (Figs. 14 and 17).

The deposits of turbulent flows are those which best conform the model of the Bouma sequence (Bouma, 1962). The Bouma *a* division is considered here as a massive and subtly graded deposit made of medium to fine sand (F8 of Figs. 13 and 15g), thus differing from the coarser grained facies of group B. Its interpretation is that suggested by Middleton and Hampton (1973), i.e. the result of high rates of sediment fallout from an overlying suspension, preventing the formation of tractive features and causing liquefaction (‘quick bed’) because of excess pore pressure (see also Kneller and Branney (1995)). The classic *b* through *d* divisions of the Bouma sequence are invariably made up of fine sand and coarse silt showing structures associated with a well-developed traction-plus-fallout process recording the depletive and waning stages of a turbulent flow. In distal ponded basin-plain regions, thick and relatively dilute turbidity currents become ‘contained turbidity currents’ (in the sense of Pickering and Hiscott (1985)), thus experiencing deflections, reflections and ponding recorded by very complex beds deposited by unsteady and nonuniform flows (Pickering & Hiscott, 1985; Remacha & Fernández, 2003). In these ponded settings, most turbidite beds are characterized by very thick mudstone divisions (Mutti & Ricci Lucchi, 1972; Remacha, Fernández, Maestro, Oms, & Estrada, 1998; Remacha & Fernández, 2003) suggesting that a large amount of suspended mud is the necessary condition for turbidity currents to become highly efficient flows that can travel considerable distances along the axes of foredeep basins (Mutti et al., 1999).

3.5. Some remarks on basinal turbidite systems

Based on modern settings, it is difficult to perceive the way in which the impressive volumes of foredeep turbidites could accumulate. Many individual sandstone beds and sandstone lobes can be traced along the axis of these basins over distances in excess of 100–200 km with volumes on the order of a few to 10’s km^3, respectively (Ricci Lucchi & Valmori, 1980). This requires that comparably large volumes of sand and mud must were available in the source areas. On the other hand, evidence discussed in above

deposit has been completely replaced by a horizontally-stratified division characterized by thinning- and fining-upward laminae deposited by a near-bed suspension. (E) F5 division sharply and erosively overlain by a cross-stratified coarse-grained sandstone division (F6). The F6 facies is sharply capped, through a very distinct break in grain size recording a phase of sediment bypass, by the fine-grained deposits laid down by the dilute tail of the turbidity current (F9). (F) Typical example of an F6 division characterized by large-scale cross stratification in coarse-grained sandstone (G) Example of an F8 (Bouma *a* division) made up of unstratified subtly graded medium to fine sandstone sharply overlain by a current-ripple Bouma *c* division (F9). Photographs A, B, C, D are from the Eocene turbidite Hecho Group, south-central Pyrenees, Spain. Photograph E is from the Eocene and Oligocene Annot Sandstone, Maritime Alps, France; photograph F is from the Oligocene Reitano flysch, Southern Italy; photograph G is from the lower Miocene Cervarola Sandstone, northern Apennines, Italy.

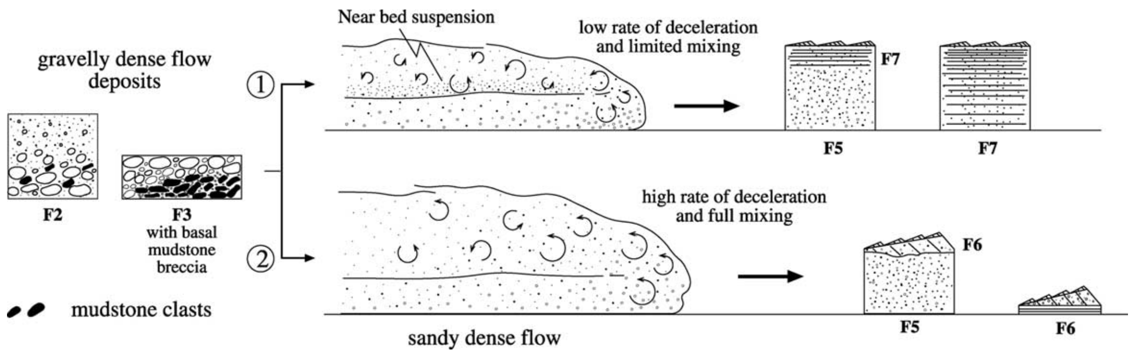

Fig. 17. Facies types observed at the final transformation of a dense sandy flow into a turbulent flow (see text for an extensive discussion).

sections indicates that fluvio-deltaic systems developed along the margins of foredeep basins are relatively small, an inherent characteristic of active margin settings. Therefore, none of these systems, if taken in isolation, could produce hyperpycnal flows of sufficient sediment discharge to generate turbidity currents of the volume required for foredeep turbidite sedimentation.

Well-known examples of turbidity currents of extraordinary volume (associated with catastrophic flows triggered by equally catastrophic events) include those related to the Grand Banks earthquake (see summary in Hughes Clarke, Shor, Piper, & Mayer, 1990) and the late Pleistocene Missoula megafloods (Zuffa, Normark, Serra, & Brunner, 2000; Piper & Normark, 2001). It seems difficult to advocate similar events to account for the many thousands of large-volume turbidity currents that reached ancient foredeep basins, though earthquakes and megafloods certainly played a role in such tectonically active settings (Labaume, Mutti, & Seguret, 1987). In particular, we argue that catastrophes of this magnitude are very unlikely to occur as high-frequency cyclic events to account for the stacking patterns exhibited by the typical metre-thick alternations of sandstone lobes and finer-grained facies that characterize the axial fill of foredeep basins (Mutti et al., 1999). We therefore tentatively suggest that the origin of large-volume turbidity currents of foredeep basins and their

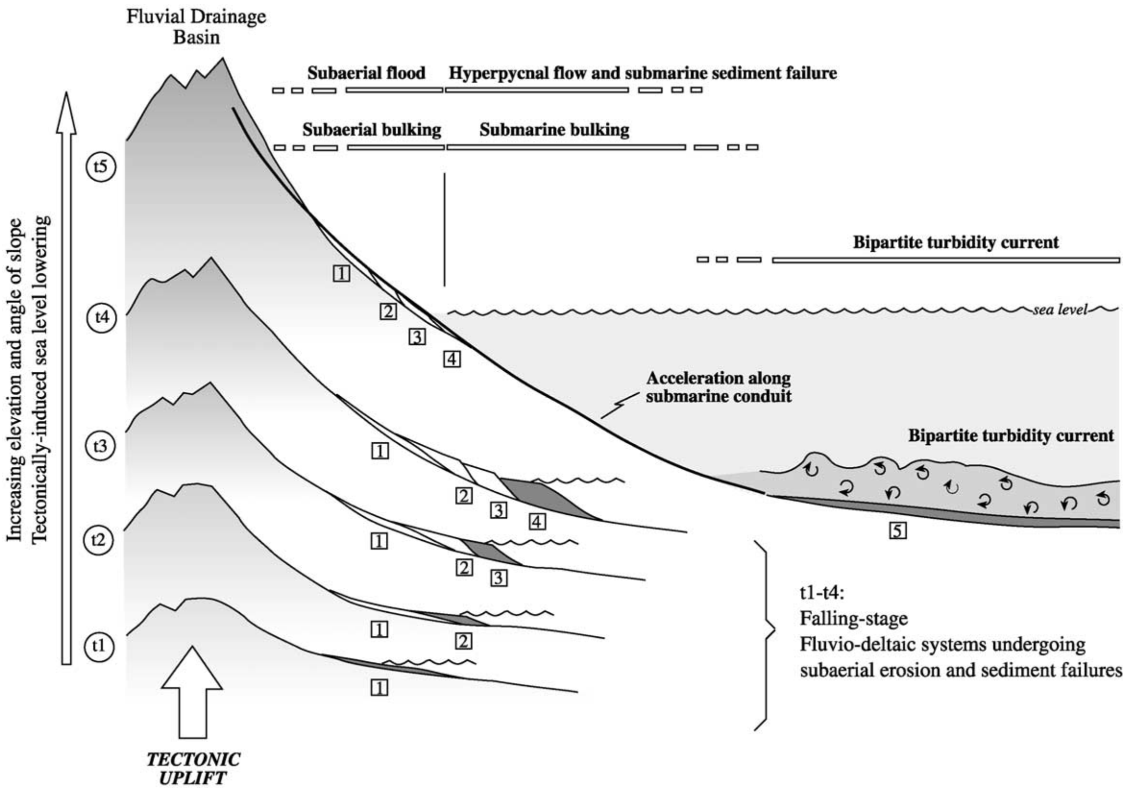

Fig. 18. Sketch showing the conditions required to form large-volume and highly-efficient turbidity currents in foredeep basins. Tectonic uplift controls sediment availability and generates a progressively steeper depositional profile. Climate-controlled floods trigger subaerial gravity flows that progressively transform into turbidity currents.

cyclic occurrence can be explained by an interaction of many factors, which are typical of these basins and difficult to perceive from the Recent.

Although a physical link with marginal flood-dominated fluvio-deltaic sedimentation and related hyperpycnal flows cannot be established, multiple lines of evidence suggest that sandy basinal turbidites originated from catastrophic floods and sediment failures during relative falling- and lowstand-stages of sealevel forced by the dramatic uplift of basin margins (Fig. 18). Tectonic uplift progressively increases the sediment yield of small rivers through the elevation of drainage basins and their merging into progressively larger rivers. In eustasy-controlled lowstands, this would result in the formation of progressively larger fluvio-deltaic systems where the role of floods would be considerably reduced (Mulder & Syvitski, 1996). In tectonically forced lowstands, we argue that basin margin physiography may conversely enhance the role of floods due to the steepened gradient, the increased elevation of drainage basins, and the bulking of flood-generated flows through the erosion of older alluvial and deltaic deposits, which formed during earlier relative falling-stages of sealevel. If these flows enter seawaters as high-momentum jet flows, bypassing and eroding narrow and high-gradient shelf regions, they may travel down and be further accelerated along steep slopes, thus generating submarine conduits. The process must trigger extensive sediment failures in shelfal and slope regions and subsequently increase the volume and sediment concentration of these flows. Catastrophic bed erosion and acceleration of these flows along submarine conduits generate the bipartite and highly efficient turbidity currents discussed in the preceding pages. In conclusion, these currents can apparently form only if substantial amounts of fines are added to the parental flow through submarine slides and bed erosion.

4. Discussion and conclusions

4.1. Fluvio-deltaic systems

Fluvio-deltaic systems of foreland basins are dominated by flood-generated flows in their alluvial, nearshore, shelfal and slope elements. Although constituting impressive sedimentary volumes in exposed orogenic belts, these systems and their component facies and facies associations are essentially ignored in current sedimentological models. Hyperpycnal flows, formed in fan-delta and river-delta systems during catastrophic flood events, are able to carry sand and gravel over considerable distances (up to tens of km) across shelfal regions—a character of hyperpycnal flows that is difficult to perceive from modern settings where these flows are essentially viewed as suspensions of mud and fine sand. The typical deposits of these flows are flood-generated delta-front sandstone lobes a category of deposits which is still unfortunately mistaken, in most recent literature, either for storm-dominated shoreface and shelfal deposits due to the common occurrence of HCS (Back et al., 2001) or for basin-floor turbidites, where they develop in relatively deeper shelfal regions at the toe of prograding deltaic clinoforms (Plink-Bjorklund, Mellere, & Steel, 2001).

4.2. Mixed depositional systems

Mixed depositional systems are made up of turbidite-like facies and facies associations formed at relatively shallow depths at the seaward edges of flood-dominated deltaic systems. The term 'delta-fed turbidites' is not retained in this paper in order to avoid confusion and consequently, these sediments can be viewed as marginal turbidites associated with basin margin deltas. Although careful analysis of facies and stratigraphic relationships usually permit the differentation of these marginal systems from basinal turbidites, this differentiation may in some cases be difficult, particularly when based on limited exposures or solely on core analysis. For this reason, most of these systems have been commonly described and interpreted as basinal turbidites in previous literature (Mutti, 1979, 1992a). This clearly shows that there has been a general tendency among sedimentologists and stratigraphers over the years to lump into the broad category of turbidites all those sandstones deposited by density currents, lacking obvious evidence of shallow-marine processes, and interbedded with mudstone facies. Parallel-sided graded sandstone beds, sole markings, and Bouma-type depositional divisions within individual beds are thus criteria that, if taken out of context, are insufficient to establish the processes, environment and water depth of the depositional system under consideration. The problem may also have obvious and important implications in hydrocarbon exploration and exploitation regarding the prediction of facies distribution patterns and reservoir quality.

4.3. Basinal turbidites

Mainly because of the greatly increased economic importance of hydrocarbon-bearing turbidite sands in many offshore basins worldwide (e.g. Gulf of Mexico, west Africa and Brazilian offshore), there has been a renewed strong interest in this kind of sedimentation in recent years with emphasis on divergent continental margin basins. Based on the great advances in marine geology and hydrocarbon exploration techniques, this has resulted in a large number of publications focused in particular on slope channels and intra-slope basins (see summaries in Prather, Booth, Steffens, and Craig (1998); Prather (2003); Pirmez, Beaubouef, Friedmann and Mohrig, 2000; Pirmez and Imran, (2003); Posamentier, (2003)).

The new insight gained from these studies has inevitably added further problems to an understanding of deep marine sedimentation. Clearly, turbidite sedimentation of divergent continental margins differs dramatically from that recorded by ancient foredeep basins. If we maintain the term 'turbidites' for the classic basinal sheet-like deposits of the latter, then it may become confusing to use the same term for the channelized and intra-slope basin-fill sandstones of continental margins. Sheet-like sand-rich basinal turbidites along divergent continental margins fed by submarine canyons are probably deposited only in relatively outer regions of deep-sea fans (see summary in Piper and Normark (2001)), i.e. in ultra-deep offshore regions that remain largely unknown in terms of their depositional architecture and facies types, at least based on the information available in the public domain.

The extent to which deep basinal turbidite sedimentation of ancient foreland basins can be used to improve our understanding of the same kind of sedimentation in continental and particularly divergent margin basins, and vice versa, is a basic issue well beyond the purposes of this paper. A few aspects of these problems are briefly discussed below.

Slope systems of divergent margins usually lie seaward of lowstand deltas and are commonly associated with significant tectonically induced topography caused by salt or mud diapirs and growth faults. For this reason, among others, we argue against the real 'turbidite' character of these systems and tentatively suggest that they may be rather and more significantly interpreted as mixed depositional systems built up by gravity flows produced by sediment failures and hyperpycnal flows, which were forced to deposit their sediment load in topographic depressions or low-gradient slope segments before acquiring sufficient acceleration to become highly efficient bipartite turbidity currents.

These systems commonly exhibit channel systems forming spectacular meandering belts with associated levees and a variety of lateral and frontal splays (Mayall & Stewart, 2000; Kolla, Bourges, Urruty, & Safa, 2001; Posamentier, 2003). Although with some minor differences, these channelized systems strongly resemble those of typical fluvial systems (Pirmez et al., 2000; Pirmez & Imran, 2003) and this similarity probably reflects an actual fluvial-like character of these submarine systems.

These channel and overbank deposits could be the product of relatively long-duration hyperpycnal flows, essentially loaded with mud and fine sand, exiting river mouths during stages of delta-recession forced by sealevel rise (transgressive systems tract) and culminating with the deposition of condensed sections (Booth, DuVernay III, Pfeiffer & Styzen, 2000). These relatively dilute hyperpycnal flows would behave as submarine 'rivers', i.e. fluidal turbulent flows with enough fine-grained sediment concentration to form density currents. These flows should thus be strongly affected by the local slope profile, being accelerated along steep segments, where they become erosive, and forming depositional meandering belts where the gradient decreases. Most of the sand deposited in these meandering belts, in places exhibiting spectacular laterally accreted bars, could be partly the 'bedload' of these submarine 'rivers', resulting from the erosion and resuspension of underlying coarser-grained sediment deposited by dense sandy flows.

Depending on their magnitude (duration and sediment concentration), these relatively dilute hyperpycnal flows may form relatively small channel-levee complexes in slope regions fed by relatively small rivers (Beaubouef & Friedman, 2000; Booth et al., 2000), or extend to basinal regions (deep-sea fans), forming large channel-levee complexes (Pirmez et al., 2000; Piper & Normark, 2001) within which individual channels may have lengths up to several hundreds of kilometers. These large channel-levee complexes can be interpreted as the deposit of long-lasting (weeks or months) and dominantly mud-laden hyperpycnal flows exiting the mouths of large rivers and directly funneled down adjacent canyons where their flow is contained within canyons and fan-valleys with associated levees (Piper & Normark, 2001; Mulder, Savoye, Piper, & Syvitski, 1998). These flows may thus reach deep-water fans where they will eventually undergo lateral spreading or maintain sufficient energy to move farther basinward across low-relief fan surfaces forming channels with sinuous to straight courses, extending considerable distances from related fan-valleys. Channels of this kind have never been documented from foredeep turbidites of orogenic belts probably because of the dramatically different style of their associated fluvio-deltaic sedimentation and basin-margin physiography.

In summary, both small (shallower) and large (deeper) channel-levee complexes of continental margins can in some way be compared to the prodeltaic wedges of relative small fluvio-deltaic systems of foreland basins discussed in earlier sections. These wedges, however, do not connect to large canyons and are built up by considerably smaller-volume, mud-laden flows experiencing lateral spreading over short distance from river mouths, followed by deposition in shelfal and slope regions.

4.4. Tectonism and climate

The cyclicity developed at different hierarchical orders within foreland basin successions is apparently controlled by two main factors, tectonism and cyclic climate changes (for an extensive discussion see Milliman and Syvitski (1992), Mutti et al. (1996, 1999, 2000) and Dewey and Pitman (1998)). Tectonism increases sediment availability necessary for generating the huge volumes of sand and mud required for the infilling of adjacent basins. Alternating periods of tectonic uplift and relaxation (subsidence) result in cycles of relative sealevel variations and related unconformity-bounded units of basinwide extent (Mutti et al., 1996, 1999).

Higher frequency depositional sequences, spectacularly punctuating the longer-term and larger-scale tectonically controlled ones are herein thought to result mainly from climatic and small-scale eustatic variations generated by orbitally forced cyclicity within the Milankovitch range. Climate, in particular, seems the basic factor controlling the frequency and magnitude of floods and therefore the origin of hyperpycnal flows that dramatically increase sediment flux to the sea. As noted earlier, most turbidity currents of tectonically active basins seem to be directly or indirectly related to hyperpycnal flows, thus recording periods of time during which sediment flux to the sea attains its maximum. A growing body of evidence seems therefore to suggest that the final depositional zone of most ancient fluvial systems lies in deep waters, far away from river mouths, and is recorded by basinal turbidite sandstones. A better understanding of these sandstone thus largely depends on our knowledge of the depositional history of their marginal fluvio-deltaic systems.

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