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Hyperpycnal flows and hyperpycnites: Origin and distinctive characteristics

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Abstract: Growing evidence suggests that land generated sediment gravity flows are the most important source of clastic sediments into marine and lacustrine sedimentary basins. These sediments are mostly transferred from source areas during exceptional river discharges (river floods). During floods rivers discharge a sediment-water mixture having a bulk density that often exceeds that of the water in the receiving water body. Consequently, when these flows enter a marine or lacustrine basin they plunge and move basinward as a land-derived underflow or hyperpycnal flow. Depending on the grain-size of suspended materials, hyperpycnal flows can be muddy or sandy. Sandy hyperpycnal flows also can carry bedload resulting in sandy to gravel composite beds with sharp to gradual internal facies changes laterally associated with lofting rhythmites. Lofting occurs because flow density reversal due to the buoyant effect of freshwater when a waning turbulent flow loses part of the sandy load. On the contrary, muddy hyperpycnal flows are loaded by a turbulent suspension of silt and clay. Since the concentration of silt and clay don't decrease with flow velocity, muddy hyperpycnal flows will be not affected by lofting and the flow will remain attached to the sea bottom until its final deposition. The last characteristics commonly result in cm-thick graded shales disposed over an erosive base with dispersed plant debris and displaced marine microfossils. Deposits related to hyperpycnal flows are hyperpycnites. Although hyperpycnites display typical and diagnostic characteristics that allow a clear recognition, these deposits are often misinterpreted in the literature as sandy debrites, shoreface, estuarine or fluvial deposits. The correct identification and interpretation of hyperpycnites provides a new frontier for the understanding and prediction of conventional and unconventional reservoirs.

Key words: hyperpycnal flows; hyperpycnites; deep water sedimentary; turbulent flow

Introduction

Rivers are far the main responsible of transferring sediments from producing areas to the basin. According to Syvitski et al. (2003) present rivers transfer to the ocean about 25 GT/year of sediments, this is more than 90% of the total sediment influx from terrestrial sources. The river mouth was historically considered the zone where most terrigenous sediments accumulate, due to the drastic deceleration and loss of confinement of these stream flows when reaching the

coast. Nevertheless, growing evidences demonstrate that a lot of sediments can bypass this coastal area during river floods, allowing the transfer of a huge volume of sediments hundreds of kilometers basinward. This situation is possible because most rivers at least once a year (Mulder & Syvitski, 1995) discharge a mixture of water and sediments having a bulk density that exceeds that of the receiving water body (hyperpycnal flows since Bates 1953). As a consequence, fluvial discharges plunge in coastal areas and travel basinward as fluvial derived sediment

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gravity flows. The volume of sediments transferred in each flow can be enormous. Mulder et al. (2003) shown that during a single hyperpycnal discharge of the Var River that lasts 18 hours, the volume of sediments transferred to the marine was equivalent to that transported in 20 years under normal conditions. Consequently, only a limited fraction of the sediments is retained in the littoral areas since most of the sediment are transported to the shelf and deeper areas during periodical floods.

This change in paradigm has important consequences for the sedimentology and reservoir significance of shelfal and deep marine/lacustrine systems, since predicts the existence, and allows the explanation of the occurrence of sandstone deposits (channels and lobes) located far from the coast line not related to a sea/lake level fall.

The study of 230 actual fluvial systems revealed that most rivers (84%) produce periodical hyperpycnal discharges (Mulder & Chapron, 2011) in their as-

sociated marine basins. These data strongly suggest that hyperpycnal flow deposits (hyperpycnites) should be very common in recent and ancient successions. Nevertheless, the occurrence of recent and ancient hyperpycnites are poorly quoted in the geological literature. Probably, a large number of hyperpycnal deposits do exist in the geological record, and have been wrongly interpreted as estuarine, fluvial, deltaic (littoral delta front and delta plain), storm (tempestite), shoreface and sandy debrite deposits.

1 Hyperpycnal flows

Bates (1953) introduced a rational classification of deltas, considering the relationship between the density of the incoming flow (river discharge D_r) respect to that of the receiving water body (lake or sea, D_w). Basically, Bates recognized three categories, termed Hypopycnal flow, Homopycnal flow and Hyperpycnal flow (Fig. 1).

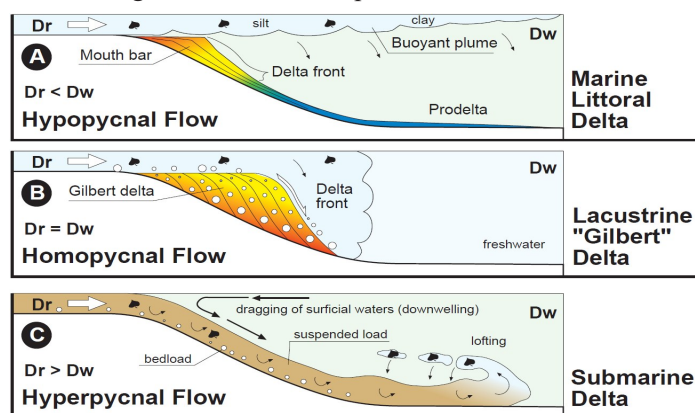


Figure 1: Classification of deltas according to Bates (1953), considering the relationship between the density of the incoming flow (river discharge or D_r) respect to that of the receiving water body (lake or sea, or D_w). A) Hypopycnal flow ($D_r < D_w$). B) Homopycnal flow ($D_r = D_w$). C) Hyperpycnal flow ($D_r > D_w$).

A Hypopycnal flow (Fig. 1 A) forms when the river discharge is less dense respect to the density of the reservoir water ($D_r < D_w$). In this situation the stream flow experienced a rapid deceleration and loss of confinement at the river mouth, with the consequent accumulation of the coarser grained fractions in coastal areas (forming mouth bars). Freshwater, suspended fine grained materials (silt/clay) and plant remains form a buoyant plume that can extend some distance from the coastline. The collapse of these materials compose a fine grained prodelta. The typical result of hypopycnal flows are marine and eventually lacustrine littoral deltas. The shape of these littoral deltas will be controlled by the existence of not of diffusion

processes at the coast (wave or tides), resulting in fluvial, wave or tidal dominated deltas (see the popular classification of Galloway, 1975).

Homopycnal flow (Fig. 1 B) corresponds to the situation when the incoming flow has a similar density respect to that of the receiving water body ($D_r = D_w$). In this particular case all the sediment fractions carried-up by the fluvial stream rapidly collapse at the river mouth. In rivers having coarse grained bedload this results in steep-gradient deltas (Gilbert-type deltas) dominated by debris falls (avalanches). The homopycnal condition is almost exclusive of sediment-free bedload dominated stream flows entering freshwater lakes.

A hyperpycnal flow (Fig. 1 C) occurs when the density of the incoming flow is higher than that of the water in the reservoir ($D_r > D_w$). In delta systems, this situation originated when a subaerial system (i.e. a river) discharges a mixture of water and sediment having a bulk density which is greater than that of the receiving water body. In the case of the marine environment, a concentration of 35-45 kg/m³ of suspended sediment is required in the fluvial discharge to overcome the density contrast with sea water (Mulder and Syvitski, 1995). On freshwater lakes, hyperpycnal flows are very common, since only 1 kg/m³ of suspended sediment is required in the incoming flow to went hyperpycnal (Tab. 1). When this situation occurs, the fluvial discharge sinks below the sea water forming a hyperpycnal flow (Bates, 1953) which can travel considerable distances carrying large vol-

umes of sediment directly supplied from a river in flood. A hyperpycnal flow can be a sediment gravity flow, but not all sediment gravity flows are hyperpycnal flows. A subaqueous sediment gravity flow can only be considered as hyperpycnal (from the Greek *ὑπέρ* (hyper) meaning "over", *pycnal* = density, from Greek: *πυκνός* (puknos) meaning "dense") if it was originated in the continent. This is because the hyperpycnal condition is achieved at the coast, according to the density contrast between incoming and receiving fluids. Consequently, sediment gravity flows generated within the basin, as the case of mass-transport complexes, intrabasinal turbidites, tempestites, cascades, and turbulent flows derived from convective instability (Parsons et al., 2001) should not be considered true hyperpycnal flows.

RIVER DISCHARGE		CONCENTRATION		AVERAGE DENSITY/Kg • m ⁻³	SUSPENDED AEDIMENT/ Kg • m ⁻³
		WEIGHT/%	VOLUME/%		
Critical concentration required to produce a hyperpycnal flow	Sea water	>3.4-4.7	>1.3-1.7	>1022-1027	>35-45
	Fresh water	>0.1	>0.04	>1000.4	>1

Table 1: Critical sediment concentration in fluvial inflows required to produce a hyperpycnal flow in marine and lacustrine basins. After Mulder and Syvitski (1995). Density differences of sea water are related to latitudinal variations.

Hyperpycnal flows originated from a river discharge are characterized by a turbulent suspension with interstitial freshwater. When this relatively denser fluvial discharge plunges in coastal areas, the resulting downward flow (hyperpycnal flow) induces a downwelling in surface reservoir waters, progressively avoiding the formation of buoyant (hypopycnal) plumes (Fig. 1 C). Consequently, during a hyperpycnal discharge, freshwater and other relatively light materials originally transported within the fluvial discharge, like plant debris, leaves, trunks and charcoal, are forced to sink and to travel basinward within the hyperpycnal discharge.

Depending on the duration of the relatively dense subaerial discharge, a hyperpycnal flow can be episodic or sustained (long lived). An episodic hyperpycnal flow usually last few hours and commonly develops in fan-delta settings with steep gradients and small catchment areas. Since these hyperpycnal dis-

charges are relatively highly concentrated and short lived, these deposits commonly have a limited distribution in the associated basin. In high gradient delta slopes they can also trigger intrabasinal sediment gravity flows. Their deposits could be very variable depending on the original density of the incoming flow (cohesive debris flow, hyperconcentrated flow, turbulent flow) and their possible evolution into intrabasinal turbidites. According to their higher relevance, this paper will be focused only in sustained (long lived) hyperpycnal flows and their deposits.

2 Sustained hyperpycnal flows

Hyperpycnal discharges associated to medium to large-sized rivers can last for weeks or even months depending on climate and the size and shape of the associated fluvial drainage area. These last characteristics results in turbidity currents that will be very different compared to those associated to intrabasinal

sediment gravity flows. Fig.2 summarizes the main characteristics of long lived (sustained) hyperpycnal flows. These characteristics include:

An origin related to a direct fluvial discharge, which is often characterized by long lived flows with fluctuating changes in velocity and concentration.

Common occurrence of associated bedload pro-

cesses (carrying terrestrial and basinal eroded components) with shear provided by the passing-by long lived hyperpycnal flow.

A turbulent flow having a light interstitial fluid (freshwater) together with other light components in suspension, like plant remnants, leaves, trunks, etc.

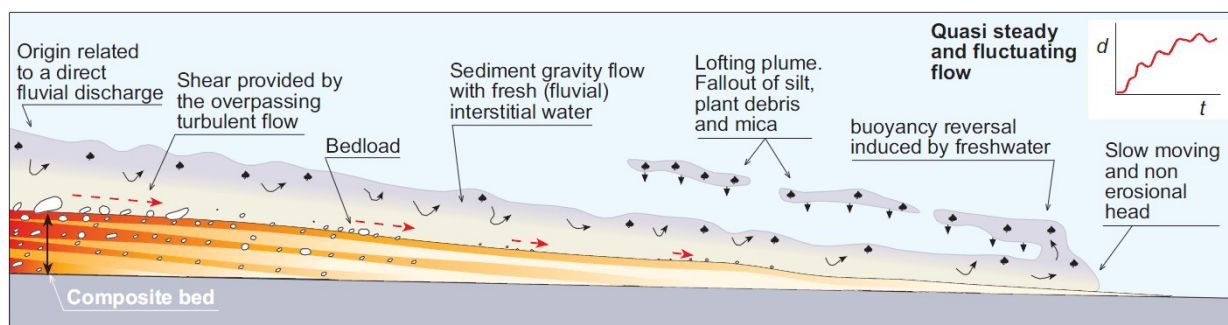


Figure 2: Main characteristics of long-lived hyperpycnal flows and their typical deposits (From Zavala et al., 2008 a; Zavala et al., 2011). The complexity of these flows results in the accumulation of composite beds (Zavala et al., 2007).

Sustained hyperpycnal flows consist of three distinctive parts: the plunge region, the main body, and the leading head (Kassem & Imran, 2001). In contrary to intrabasinal turbidity currents, sustained hyperpycnal flows have a slow moving and non-erosional leading-head (Zavala et al., 2011; Zavala & Arcuri 2016) and main erosion and deposition occurs in the flow body instead of the flow head (De Rooij & Dalziel, 2001; Peakall et al., 2001; Zavala et al., 2006 a; Zavala et al., 2011). Field observations of recent and ancient hyperpycnal deposits indicate that some fine hyperpycnites are composed of an inversely graded (waxing flow) basal unit, followed in transition by a normally graded (waning flow) unit (Mulder & Alexander, 2001; Mulder et al., 2003; Zavala et al., 2006 a). The inversely graded interval is accumulated at the flow head, and commonly starts with climbing ripples (Zavala et al., 2006 a) suggesting a flow speed smaller than 0.2 m/s for the advancing leading head. Nakajima (2006) described hyperpycnites in the central Japan Sea 700 km away from the Toyama Bay. This author estimates a flow speed of 0.3 m/s and a hyperpycnal discharge lasting for about 3 or 4 weeks to achieve the indicated distance.

As a consequence of the low interaction of the flow head with ambient waters, hyperpycnal flows

can transport freshwater and plant debris (if any) for long distances basinward (Fig. 2) as long as the river discharge is maintained (Prior et al., 1987; Zavala et al., 2011). This fact is supported by an increasing number of direct measurements and observations that revealed that the original freshwater supplied by a sustained river discharge and its associated content of terrestrial light materials can travel long distances without substantial mixing with ambient waters. As an example, Johnson et al. (2001; see also Farnsworth, 2000) reported over 12 yr, the direct association between extreme turbidity activity in the Monterrey Submarine Canyon and largest flood events in the nearby Salinas River (California). These hyperpycnal flows were studied at different research stations, located between 5.2 and 22 km far from the river mouth, and water depths ranging between 230 and 1170 meters. The studied hyperpycnal flows are characterized by fresher and warmer water that extended to depths below 1000 m, carrying substantial amount of terrestrial organic carbon (Baudin et al., 2010). A large turbidity current in the Zaire submarine valley at 4000 meters and 330 km seaward of the Congo River mouth was registered by Khripounoff et al. (2003), composed of warm and turbulent flows with velocities exceeding the 121.4 cm/s. This discharge lasted

for about 10 days, and transported basinward large amounts of fine (0.15 to 0.2 mm) quartz sand and large plants debris (wood, leaves, and roots). These hyperpycnal flows were probably related to flooding episodes in the Congo River (Savoye et al., 2009) since they correlate with submarine cable breaks across the canyon (Heezen et al., 1964). More recently Kao et al. (2010) observed at 180 km off southwestern Taiwan an anomalously warm and low salinity turbid water current at 3000 – 3700 m depths immediately after Typhoon Morakot in 2009, thus proven the ability of freshwater to travel long distances during hyperpycnal events. The direct input of freshwaters (meteoric- waters) in deep- water turbidites via hyperpycnal flows has also being proposed by Mansurbeg et al. (2006) for the formation of kaolinite owing to the dissolution of detrital silicates in the Shetland – Faroes Basin on the British continental shelf. Recently Gwiazda et al. (2015) documented the presence of the pesticide DDT in recent sediments from the Monterey Submarine Canyon located in water depths deeper than 3000 meters. These chemicals were transported up to 250 km basinward from the present canyon head during hyperpycnal discharges of the Salinas and Pajaro Rivers. Since DDT was only applied on land in California between 1944 and 1972, the presence of this pesticide suggests not only the activity of the Monterey Fan during the present Highstand, but also the ability of hyperpycnal flows to transport extrabasinal elements (like DDT and freshwater) for long distances with limited mixing with ambient waters.

Main key points to understand why freshwater and plant debris can achieve such a long distances probably reside in a) the slow moving head, compared with the velocity along the main body, which prevent the development of strong turbulence vortices and substantial mixing with basin waters; b) lofting processes at flow head and flow laterals (especially in marine settings), which avoid the reincorporation of the lifted-up water and sediments; and c) the continuous fluvial discharge, which supply the energy to maintain turbulence and flow velocity for long distances.

3 Hyperpycnal flow deposits (hyperpycnites)

A hyperpycnite (Mulder et al., 2003) is the deposit of a hyperpycnal flow. Hyperpycnites are basically extrabasinal turbidites having distinctive and at present poorly known facies characteristics (Zavala and Arcuri 2016). Its origin is closely related to a direct fluvial discharge (Mutti et al., 1996) and results in facies types and depositional features that often resemble those considered as typical of fluvial environments (bedload, meandering etc.), but commonly associated with clear marine and lacustrine indicators. The transfer and accumulation of a huge volume of continental sediments during a single long-lived flood frequently results in confusing facies types from a conventional point of view. The resulting clastic sedimentary bodies can be biostratigraphically sterile or can display a wide range of water depth indicators. Very thick hyperpycnites often lack or show rare and very specific trace fossils favoring the confusion of these strata with estuarine deposits. As for every sediment gravity flow, the location and thickness of coarse grained hyperpycnal sedimentary bodies are very sensitive to the contemporaneous subaqueous topography, often resulting in the accumulation of very thick laminated or structureless sandstone bodies of difficult interpretation using non- hyperpycnal facies models.

This complex scenario of multiple interaction between intra and extrabasinal elements results in depositional bodies of difficult interpretation according to conventional facies models. Commonly sedimentologists feel lost when dealing with these deposits, and are forced to assign conventional depositional environments supported on weak and inconsistent evidence (i.e the interpretation of thick fine grained massive sandstones as estuarine or shoreface deposits).

Key points to unravel the characteristics and complexity of hyperpycnal flow deposits reside in the understanding the main processes that govern the movement of sustained hyperpycnal flows. Sustained hyperpycnal flows are fully turbulent flows, with a relatively bulk density provided by denser suspended

particles (basically fine-grained sand, silt and clay). The relative proportions of these elements in the turbulent suspension could be highly variable, and will result in the accumulation very different facies associations. Coarse grained materials could not be supported by turbulence at the velocities commonly achieved by sustained hyperpycnal flows, but can be dragged at the base as bedload by shear forces provided by the passing-by flow. Finally, in the case of marine basins, the buoyant effect of the freshwater transported within the hyperpycnal discharge can induce the density reversal (lofting) of the flow when part of the suspended load is lost by deposition. These three elements will result in typical facies categories that will give important information about: 1) position of the studied section respect to the whole system (proximal and lateral indexes), 2) system dimensions and expected facies changes (reservoir prediction) and 3) salinity of the basin.

The characteristics above discussed result in the accumulation of three main facies families related to the three main elements that govern the movement of almost all sustained hyperpycnal discharges in marine settings: bedload, suspended load and lofting (Fig. 3). These facies categories are here termed as B (bedload related sedimentary facies), S (suspended-load related sedimentary facies) and L (lofting related sedimentary facies).

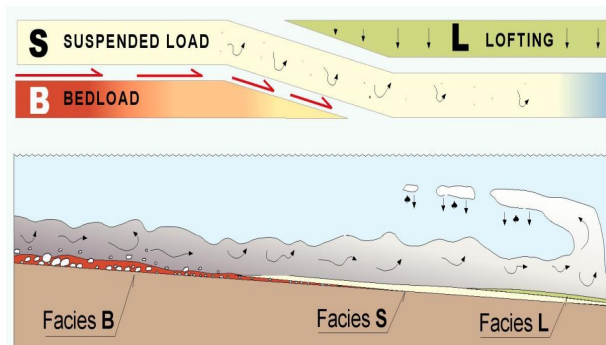


Figure 3: Main sedimentary processes and related facies families originated during hyperpycnal discharges with associated bedload in marine basins. (From Zavala et al., 2006 b; Zavala, 2008; Zavala et al., 2011).

Each facies family is in turn composed of individual facies types differentiated according to some distinctive internal characteristics. The proposed fa-

cies tract is shown in Fig. 4, and their constituting facies will be discussed in detail along the following pages. All the hyperpycnal flow-related sedimentary facies recognized in this paper resulted from changes in flow competence and flow capacity as the hyperpycnal flow wanes during its travel basinward (long distance) and towards the flow laterals (short distance). Competence refers to the large size clast that a flow can carry, whereas capacity concerns to the total volume of carried sediment. These facies types display distinguishing and characteristic depositional features that often easily allow to diagnose the hyperpycnal origin of the deposits. In the following section, main typical characteristics of facies related to bedload, suspended load and lofting processes will be discussed.

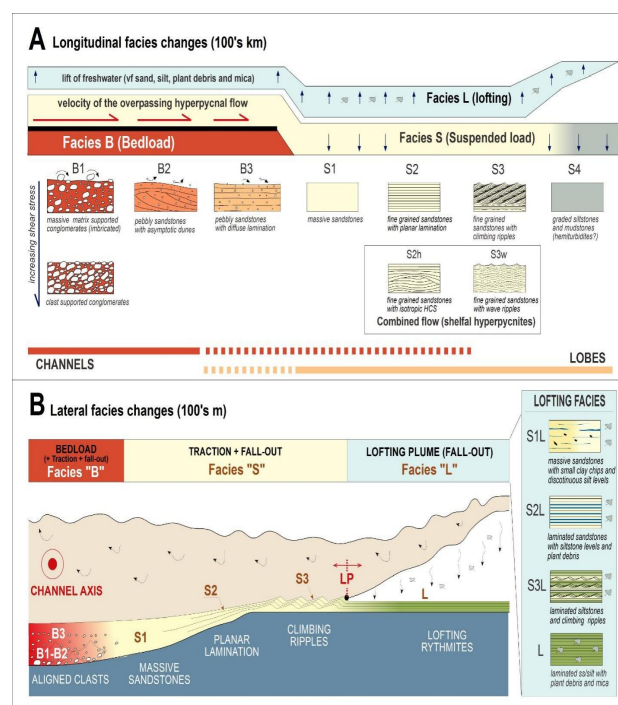


Figure 4: Genetic facies tract for the analysis of sustained hyperpycnites with associated bed load. A) Facies association along the depositional system. B) Lateral facies changes towards flow margins. Modified after Zavala & Arcuri, 2016.

3. 1 Facies related to bedload processes (facies B)

Type B (bedload related) facies family includes a broad spectrum of coarse grained deposits accumulated by the joint occurrence of (1) shear/drag forces exerted by the overpassing long-lived turbulent (hy-

perpycnal) flow over coarse grained materials carried as bedload, and (2) partial fallout of fine-grained suspended load trapped at the rising flow bottom. As a consequence, the texture of hyperpycnal bedload deposits are essentially bimodal, reflecting the joint accumulation of sediments related to these two dominant processes (bedload and suspended load). The accumulation of facies B type deposits is associated to a change in flow competence, since the flow progressively loses its ability of transporting large clasts as saltation, sliding and rolling. Since rolling is influenced by clast shape, well rounded clasts often can travel longer distances independently of the individual clast size. A direct consequence of rolling is imbrication, since during rolling non spherical clasts tend to rest in a stable position (Fig. 5). Imbrication suggests that large clasts were free to move and rotate at the base of an overpassing non-cohesive turbulent flow. Consequently, the recognition of clast imbrication in these deposits is very important because provides a diagnostic criterion for the recognition of fluid (non-cohesive) flows. Large clasts are probably transported by sliding and rolling whereas the finest materials composing the matrix would correspond to suspended load materials trapped at the low velocity and relatively high concentrated basal zone (Manville and White, 2003). The occurrence of large clasts “floating” in a sandy matrix is therefore a consequence of a combination of basal grain-to-grain interaction and hindered settling grain supporting mechanisms, and should not be wrongly interpreted as an evidence of flow cohesion (i.e. sandy debris flow or sandy debrites of Shanmugam & Moiola, 1995, Shanmugam 2008). Sandy debris flow deposits were interpreted as accumulated from sediment gravity flows characterized by a plastic rheology and a laminar state, where deposition occurs “in masse” through cohesive freezing (Shanmugam 2016). This last proposed mechanism cannot adequately explain clast imbrication and the well sorted fine grained sandy matrix of some “sandy debrites” (see pag. 30, 50, 52 and also Fig. 19 of Shanmugam 2016). According to these evidences, is here interpreted that the “sandy debrites” recognized in the Triassic of the

Ordos Basin in China (Li et al., 2011; Zou et al., 2012) probably correspond to bedload facies developed at the base of sustained turbulent (hyperpycnal) flows.

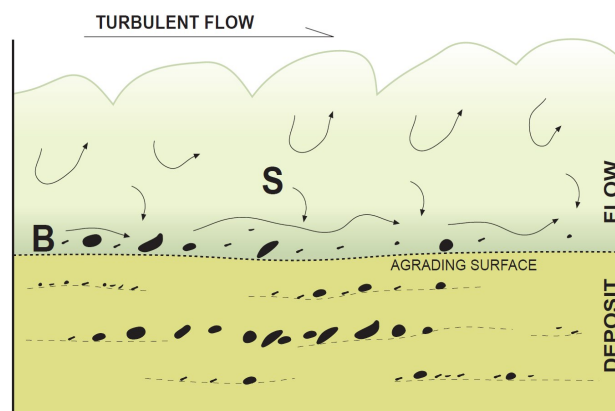


Figure 5: The occurrence of bedload at the base of sustained turbulent flows often results in bimodal deposits, composed of aligned and imbricated large clasts floating in a matrix of fine to medium grained massive sandstones.

Coarse grained particles transported as bedload can be extrabasinal (mostly composed of rounded and well-rounded pebbles and gravels), intra or extrabasinal (clay chips and clay clasts), intrabasinal (fragments of mollusks and other lacustrine or marine body fossils), or a mixture thereof. The existence of extrabasinal components in B facies suggests that bedload was probably inherited from the original subaerial fluvial discharge.

Three main categories of bedload facies are recognized (Zavala et al., 2011), termed B1, B2 and B3 (Fig. 4). In the case of sandy successions with the exclusive occurrence of clay clasts, these categories are named B1 s, B2 s and B3 s (Fig. 4).

3.1.1 Facies B1

Facies B1 is composed of massive and crudely stratified conglomerates with abundant medium to fine-grained sandstone matrix. Largest clasts appear “floating” in the sandy matrix or imbricated following diffuse subhorizontal alignments (facies B1, Figs. 6 A and 6 B). The overall texture is matrix supported, although some varieties of clast supported conglomerates are recognized (facies B1 c, Fig. 6 C). Depending on the energy and duration of the associated fluvial discharge, the deposits can show normal, inverse or internal complex vertical changes in

grain size. Field relationships suggest that a vertical evolution between facies B1 c and B1 (Fig. 6 C) could be related to a progressive decrease of drag forces exerted by the overpassing turbulent flow. In

systems that lack coarse grained clastics, B1 facies could be entirely composed of matrix or clast supported clay clasts (facies B1 s, Fig. 6 D).

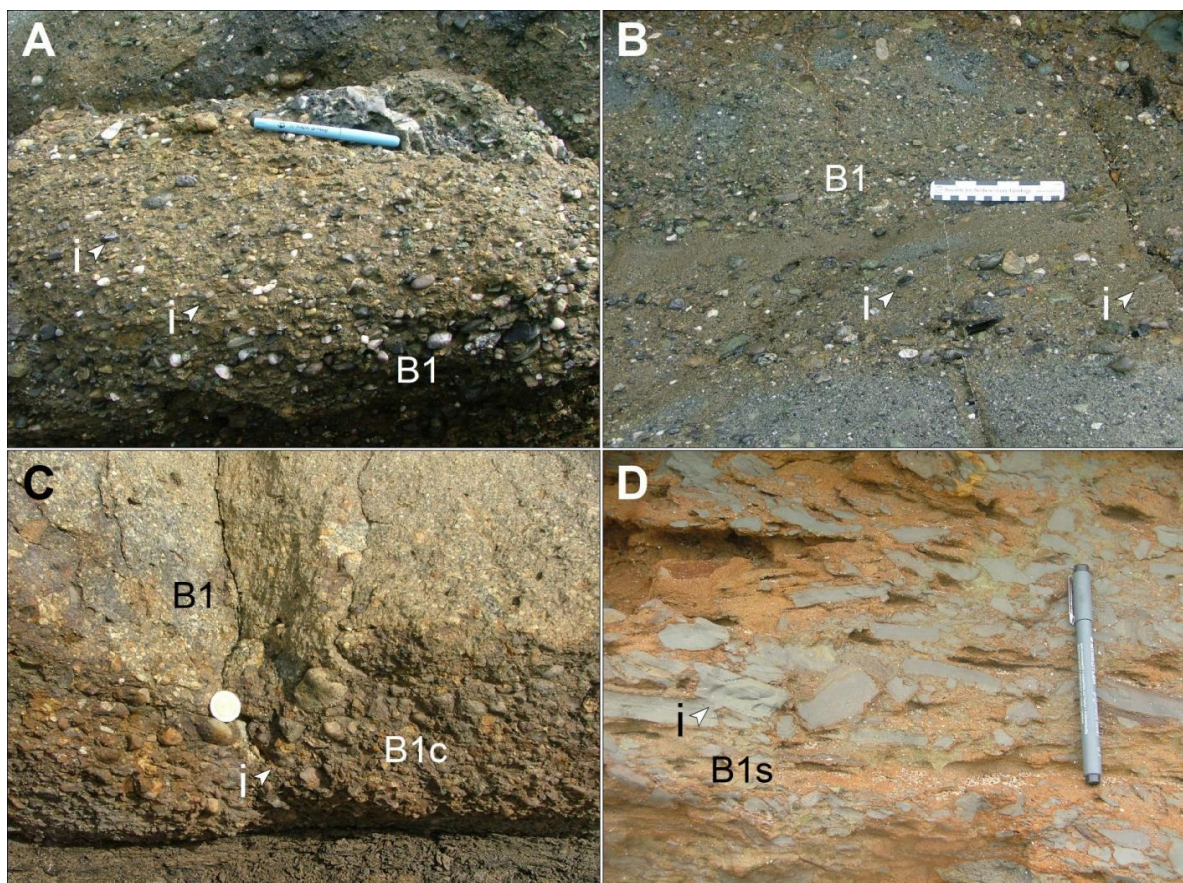


Figure 6: Field examples of B1 facies. A & B) Example of facies B1 in Eocene deep water deposits from the Pampatar Formation (Margarita Island, Venezuela). Note the fine grained matrix and clast imbrication (i). C) Clast supported conglomerates (facies B1 c) in shelfal turbidites from the Jurassic Los Molles Formation, Neuquén Basin, Argentina. D) Conglomerates composed of clay clasts (facies B1 s) floating in a fine grained sandy matrix. Jurassic Los Molles Formation, Neuquén Basin, Argentina.

3.1.2 Facies B2

Facies B2 is composed of sandstones, pebbly sandstones, and fine-grained conglomerates with low angle asymptotic cross stratification. This type of cross bedding forms at the base of a sustained turbulent flow with sand-size suspended load, and shows several distinguishing characteristics respect to that formed at the base of sediment-free streamflows (or open channels), the last very common in subaerial (fluvial) environments. Streamflow dunes (or grain-flow dunes, Fig. 7 A) are characterized by high angle foresets with flow separation. Flow separation provokes a zone of hydraulic “shadow” immediately after the brink point (Fig. 7 A) with no surficial cur-

rents. This hydraulic shadow forces coarse grained materials transported as bedload along the stoss side to accumulate, resulting in the generation of repeated gravitational avalanches at the lee side. Gravitational avalanches are inertia flows that accumulates longitudinally graded laminas (Fig. 8 A). In consequence, cross bedded sets originated from sediment-free flows are characterized by an overall fining upward trend (Fig. 7 E) with sharp and high angle basal boundaries (often erosional). Grainfall dunes (Fig. 7 B) on the contrary, are formed by flow expansion. Flow expansion at the foreset provokes a gradual decrease in flow velocity with the consequent collapse of suspended materials (fine grained sandstones) along the

lee side. This sediment fallout along the dune front results in asymptotic cross bedding (Figs. 8 A-D) often accompanied with abundant plant remains (and small clay chips) at the lower foreset (Figs. 8 B & 8 D). Asymptotic cross beddings typically compose thickening and coarsening upward successions, (Figs. 7 B, 8 B & 8 C) resulting from the dominance of suspended load at the lower foreset and bedload in the upper foreset (Fig. 7 B).

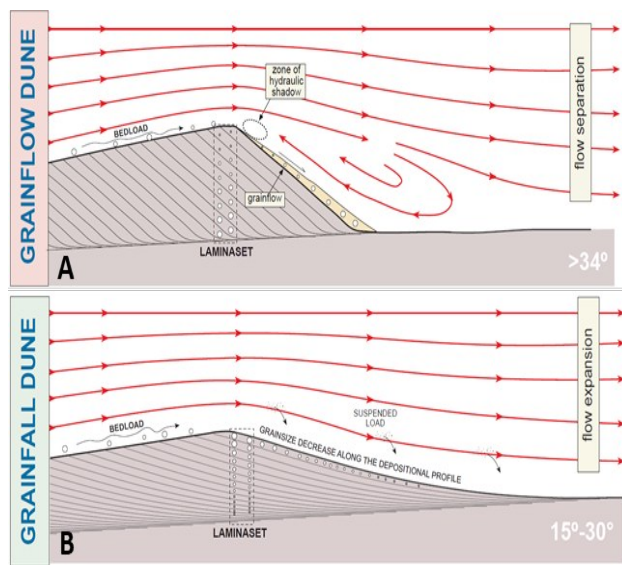


Figure 7: Main differences between cross bedding generated at the base of a streamflow (A, grainflow dunes) respect to that originated at the base of a sustained turbulent flow with abundant sand-sized suspended materials (B, grainfall dunes).

Large clasts in facies B2 often appear floating in a medium to coarse grained sandstone matrix (Fig. 8 C) along foreset laminae that in general does not exceed 20 degrees in inclination. Depending on the rate of sediment fallout from the overpassing turbulent flow, bedset bounding surfaces can be erosional (like an anisotropic hummocky cross stratification) or transitional, forming climbing dunes (in the sense of Mutti et al., 1996).

Field observations suggest a close association of this facies with channel fill deposits. In gravel-free systems, facies B2 is almost entirely composed of medium to coarse grained sandstones (Figs. 8 B & 8 D) often displaying abundant clay clasts and plant remains in the lower foreset. This sandstone variety of facies B2 is termed facies B2 s (Fig. 5). If clay

clasts are numerous and of small size, this structure can easily be confused with a tidal bundle. The key for a correct interpretation resides in differentiating the small clay clasts from a true mud couplet. Like facies B2, facies B2 s relate to the migration of straight or sinuous bedforms at the base of a long lived turbulent flow carrying high suspended load.

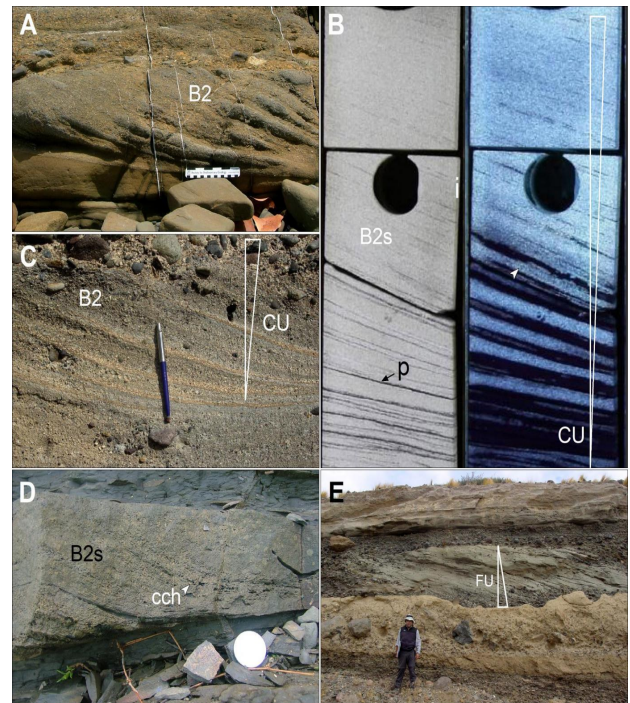


Figure 8: Field and core examples of facies B2. A & D) Low angle asymptotic cross bedding in deep water deposits the Pampatar Formation (Margarita Island, Venezuela). Note in D the abundant clay chips draping the foresets. B) Core example (daylight and UV) of asymptotic cross bedding in shallow marine channels from the Lower Cretaceous of the West Siberian Basin (Russia). Note the coarsening upward grain size and the abundant plant remains in the lower foreset. C) Asymptotic cross bedding in Pleistocene coarse grained lacustrine deposits of the Huarenchenque Formation, Neuquén Basin, Argentina. Note the coarsening upward trend. E) Contrasting example showing a cross bedding related to a subaerial streamflow (fluvial deposits). Note the abrupt basal boundary and the fining upward trend. Huarenchenque Formation, Neuquén Basin, Argentina.

3.1.3 Facies B3

Facies B3 is characterized by medium-grained to pebbly sandstones with diffuse horizontal to subhorizontal stratification and levels of small aligned pebbles (Figs. 9 A & 9 B). Pebbles appear dispersed or aligned often showing imbrication. From a textural viewpoint, this facies is characterized by a strongly

bimodal grain-size distribution, with large clasts “floating” on a sandy matrix. These characteristics reflect simultaneous deposition of small and large particles by two distinct mechanisms: (i) accretion of the sandy matrix from a turbulent suspension as the flow loses capacity, and (ii) progressive emplacement of pebbles as the flow loses the competence to roll them.

The sandy variety of this facies (facies B3 s) is characterized by aligned clay clasts (Fig. 9 C, 9 D & 9 E) commonly associated with plant fragments. Facies B3 composes tabular to lenticular bodies often filling erosive depressions. It is interpreted that these

facies accumulated by the combined effect of bedload and the gravitational segregation of sandstone materials transported by turbulence in the overpassing hyperpycnal turbulent flow. In shallow water settings, B3 facies may display low angle diverging and truncated laminae which closely resemble gravelly hummocky cross stratification (facies B3 h in Fig. 4). Mutti et al. (1994) suggested that subaerially-derived gravity flows may develop an internal oscillatory component during their downslope travelling through the setting in motion of standing bodies of water in the shallower parts of receiving water bodies.

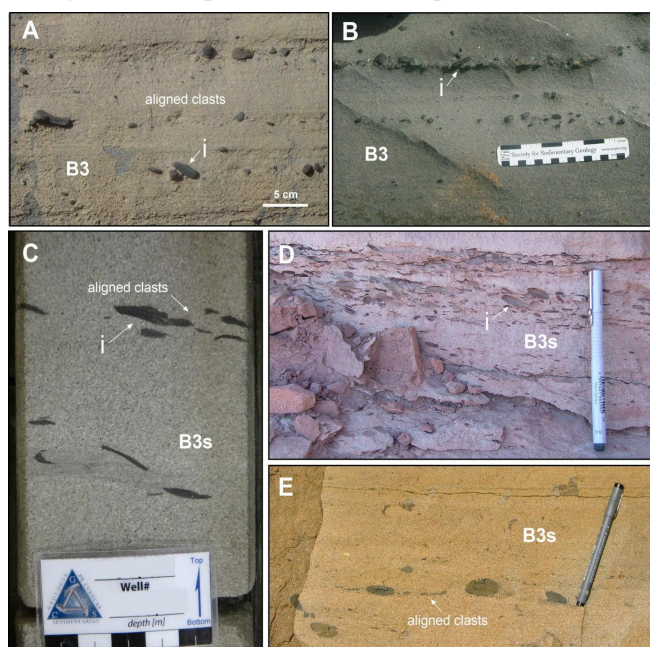


Figure 9: Field and core examples of facies B3. According to the presence of “floating” clasts, this facies has been wrongly interpreted as related to sandy debris flows (sandy debrites). A) Pebbly sandstones with aligned pebbles showing imbrication. The last suggests the joint occurrence of bedload and suspended load in aggrading beds.

Pleistocene lacustrine deposits of the Huarenchenque Formation, Neuquen Basin, Argentina. B) Pebbly sandstones with imbricated pebbles. Eocene deep water deposits of the Pampatar Formation (Margarita Island, Venezuela). C) Core example of massive sandstones with aligned clay clasts (facies B3 s). Lower Cretaceous of the West Siberian Basin (Russia). D) B3 s facies with imbricated clay clasts. Lower Cretaceous lacustrine deposits of the Rayoso Formation, Neuquén Basin, Argentina. E) Sandstones with aligned clay clasts. Lower Jurassic outer shelf deposits of the Los Molles Formation, Neuquén Basin, Argentina. i: Imbrication.

3. 2 Facies related to the collapse of suspended load (facies S)

The S facies family is mostly fine grained (fine grained sandstone & siltstone). It is composed of sediments transported as suspended load within a sustained turbulent flow and accumulated by a gradual gravitational collapse as the flow wanes and loses flow capacity. These deposits often form thick and internally complex beds that can be massive or characterized by traction plus fallout sedimentary structures.

3.2.1 Facies S1

Facies S1 is one of the most common facies within the facies tract of hyperpycnal systems. This facies composed tabular fine to medium grained massive sandstone beds (Fig. 10). These sandstones usually integrate monotonous and very thick successions in-

ternally showing subtle and gradual grain-size variations (Fig. 10 D). Small floating clay chips are common and could appear dispersed within the sandstone body or grouped towards the top of the bed (Fig. 10 B), close to the boundary with facies related to more diluted flows (facies S2). Carbonaceous remains, charcoal (Fig. 10 F) and woody fragments are also common within massive sands, often displaying entire leafs with exceptional preservation (Fig. 10E). Leaves in massive sands can be very abundant and have been proposed as the main source of hydrocarbons in the Kutei Basin, Indonesia (Saller et al., 2006). S1 facies mostly lack any kind of bioturbation. Nevertheless, some massive, very thick intervals show isolated Ophiomorpha (Fig. 10 C) and Thalassinoides, which can be related to crustaceans

(doomed pioneers) , bulked and transported by the turbulent flow from shallower areas (Grimm and Follmi, 1994).

The origin of this facies would be related to the progressive aggradation from the bottom by long lived flows (Fig. 11) having high suspended load (Sanders, 1965; Kneller and Branney, 1995; Camacho et al., 2002). This progressive aggradation has been proposed as a mechanism that inhibits the formation of primary sedimentary structures at the basal layer. Massive deposits could therefore be related to the

absence of a sharp boundary between the moving flow and the deposit, but rather a zone of aggrading transition characterized by a high sediment concentration associated with water escape. Experimental studies (Banerjee, 1977; Arnott and Hand, 1989; Sumner et al., 2008) indicate that this facies originates from a turbulent flow with fallout rates greater than 0.44 mm/s. For smaller fallout rates at similar flow velocities, the result will be laminated sands similar to facies S2.

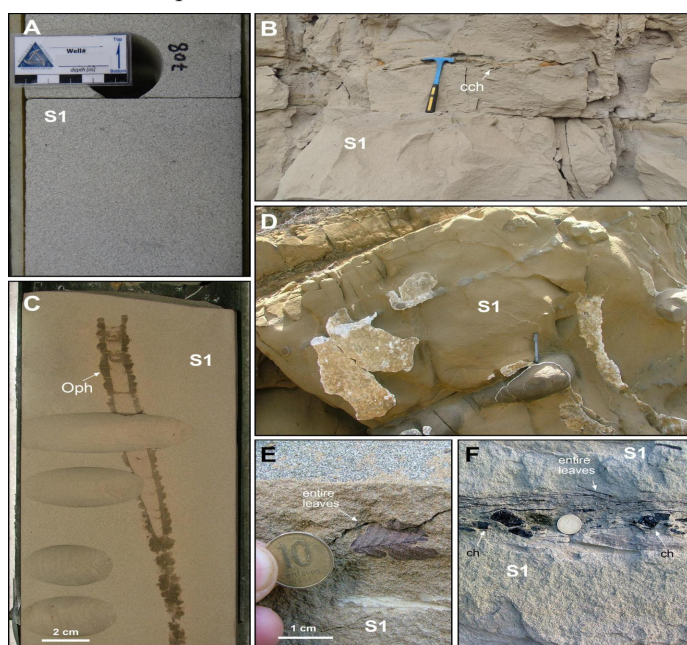


Figure 10: Field and core examples of facies S1. A) Massive fine grained sandstones. YK field, Cretaceous of Siberian Basin, Russia. **B)** Massive sandstones with clay chips (cch) on top. Lacustrine sandstone lobes in the Triassic Ordos Basin, China. **C)** Massive fine grained sandstones with *Ophiomorpha nodosa* traces as doomed pioneers. Miocene sandstone lobes in the Columbus Basin, Trinidad & Tobago. **D)** Tabular deep water sandstone lobes in the Eocene Pampatar Formation (Margarita Island, Venezuela). **E)** Entire leaves in deep water massive sandstones. Eocene of Tierra del Fuego, Argentina. **F)** Massive sandstones with entire leaves and charcoal (ch) clasts. Deep water deposits in the Eocene of Tierra del Fuego, Argentina.

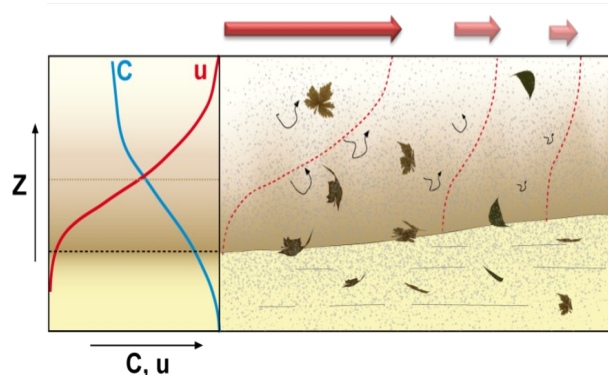


Figure 11: Diagram showing the accumulation of massive sandstone deposits with plant debris at the base of a long lived and waning turbulent flow. C= flow concentration; u= flow velocity. From Zavala et al., 2012. Modified after Kneller & Branney, 1995.

The high rates of fallout of sand-size materials results in the trapping of plant debris and other light components (leaves) transported within the turbulent

flow in the interstices of sandstone grains (Fig. 11). Also this rapid fallout results in a loose packing responsible of an original very high primary porosity of massive sandstones. In general S1 facies is characterized by a well sorted fabric (Fig. 12), since the maximum available grain size in a turbulent suspension is limited by flow velocity. Nevertheless, smaller grain size particles (silt and clay) are commonly found partially filling the interstices. These silts and clay are part of suspended materials trapped together during the collapse of fine grained sand at the lower boundary. Fig.12 shows an example of massive fine grained sandstones (microphoto and laser grain size) from the Pliocene of Trinidad. Note the relative well sorting of these fine to very fine grained sandstones, usually showing abundant associated plant debris.

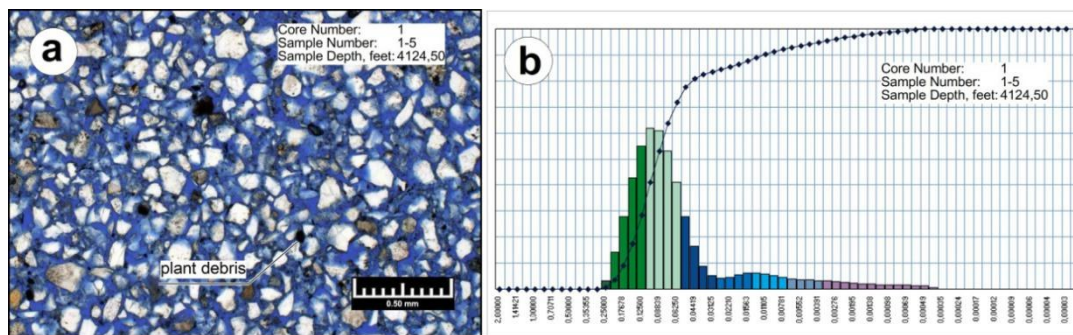


Figure 12. Microphotography (a) and laser grain size analysis (b) of massive sandstones of Pliocene turbidites in Trinidad. In “a”, black components are plant debris. This deposit is mainly composed of poorly consolidated fine to very fine grained sands.

In structurally controlled depocenters, the individual thickness of single massive sandstone bodies can be dramatically increased (Amy et al., 2007), with individual beds that in some cases exceeds the

45 meters in thickness (Arcuri and Zavala, 2007; 2008). This phenomenon is a consequence of the trapping of a fluid and subcritical sediment gravity flow in topographically confined depocenters (Fig. 13).

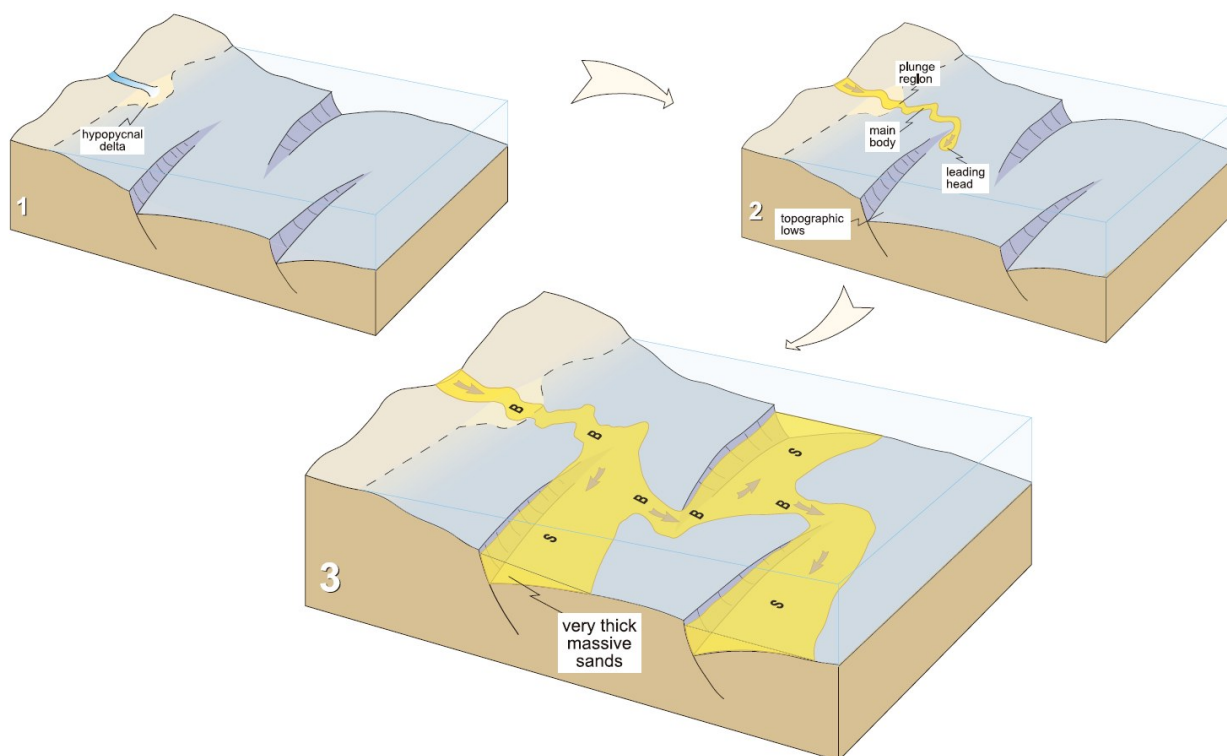


Figure 13. Accumulation of very thick massive sandstone bodies in structurally confined depocenters. The hyperpycnal flow is guided by the slow moving leading head trying to reach the lower topographic positions. If the flow is “trapped” in confined depocenters, it is forced to decelerate and accumulate forming lens-shaped thick sandstone accumulations. After Zavala et al., 2011.

3.2.2 Facies S2

Facies S2 is composed of tabular, fine-grained sandstone bodies showing subhorizontal parallel lamination (facies S2, Fig. 14) or internally low angle, diverging and truncating (hummocky-like) lamination (facies S2h, Figs. 5 and 14 D) disposed over a

sharp or transitional boundary. Individual laminae are millimeter thick and in some cases are bounded by thin levels showing abundant mica, carbonaceous material (Figs. 14 C, 14 F & 14 G) and even charcoal (Fig. 14 E). Facies S2 appears in close association with massive (S2) and climbing ripple (S3) struc-

tures, and often shows parting and other current line-

tion structures like those marked by heavy minerals.

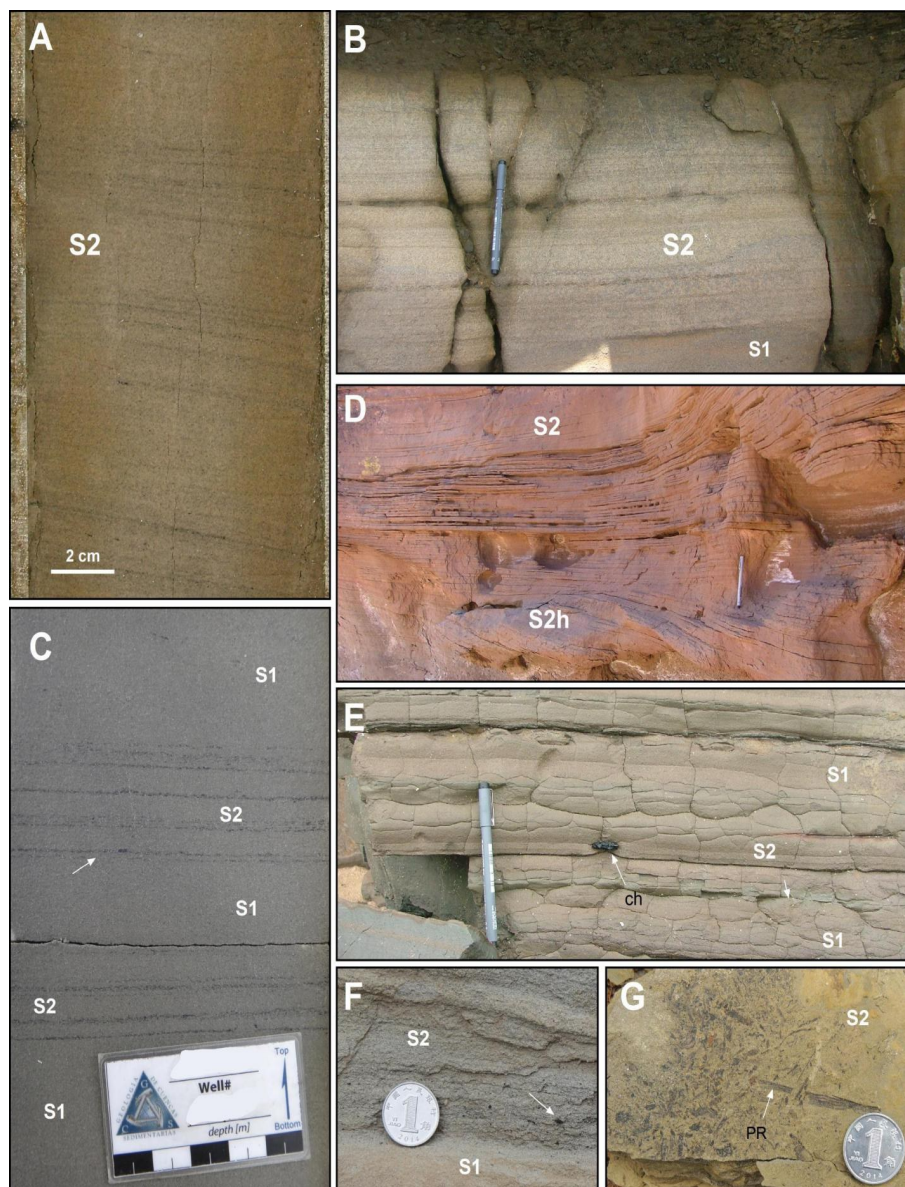


Figure 14: Core and field selected examples of facies S2. A) Laminated fine-grained sandstones in shelfal sandstone lobes. Upper Miocene of Columbus Basin, Trinidad & Tobago. **B)** Laminated sandstones (S2) on top of massive beds (S1). Eocene Pampatar Formation, Margarita Island, Venezuela. **D)** Laminated (S2) and HCS like structures (S2 h) in Lower Cretaceous lacustrine lobe deposits of the Rayoso Formation, Neuquén Basin, Argentina. **C)** Laminated sandstones (S2) alternating with massive sandstones, suggesting cyclic changes in the rate of fallout. Black lines are plant remains (arrows). Shelfal sandstone lobes of the Lower Cretaceous Centenario Formation, Neuquén Basin, Argentina. **E)** Cyclic alternance of massive and laminated fine grained sandstones. Note the charcoal clast (arrow). Shelfal sandstone lobe deposits. Mayaro Formation, Pliocene of Trinidad & Tobago. **F)** Massive (S1) and laminated fine grained sandstones in lobe deposits of the Triassic Ordos basin, China. **G)** Plat view of laminated sandstones showing abundant plant remains. Triassic Ordos Basin, China.

Laminated sandstones are common in turbidite beds (Bouma 1962). Nevertheless, the interpretation of the origin of laminated fine grained sandstones has been controversial since similar features were generated in flume experiments with unidirectional stream flows at upper flow regime (Simons et al., 1965). However, Sanders (1965) noted that parallel lamination related to turbulent flows often grades laterally into climbing ripples, suggesting a common origin by traction plus fallout processes. This last conclusion is also consistent with the experiments conducted by Arnott and Hand (1989) and Sumner et al. (2008), and it is also supported by the common association of facies S2 with massive intervals (facies S1, Figs. 14 C

& 14 E) often constituting very thick rhythmical successions. In hyperpycnal flows, laminated fine-grained sandstones represents a transitional facies between massive sandstones (facies S1) and sandstones with climbing ripples (facies S3). The transition from massive (S1) to laminated (S2) facies occurs at similar flow velocities because a decrease in the rate of sediment fallout (below 0.44 mm/s, Sumner et al., 2008). On the other hand, the passage between laminated (S2) and climbing rippled sandstones (S3) responds to a decrease in flow velocity (Sanders, 1965) with a consequent increase in the fallout rate (Fig. 15).

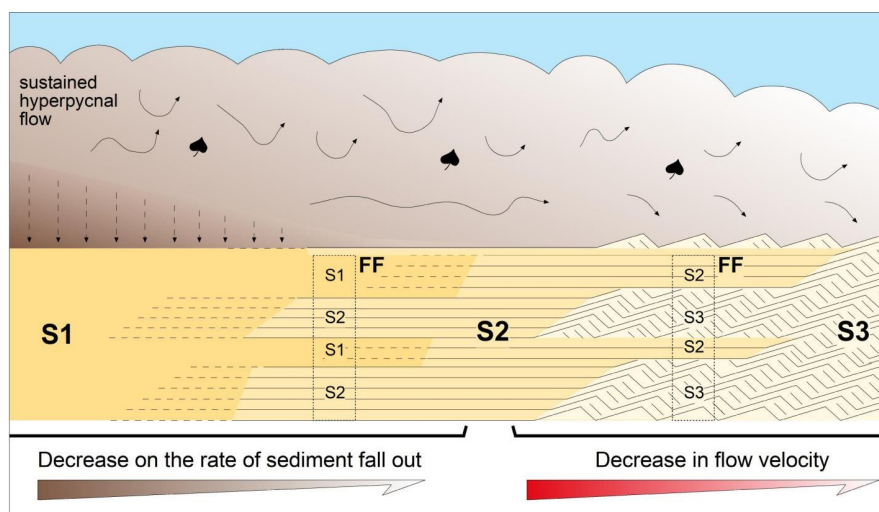


Figure 15: The origin of massive (S1), laminated (S2) and climbing rippled (S3) sandstones as related to changes in the rate of sediment fallout and decrease in flow velocity of sustained turbulent suspensions. The repeated and cyclic variation of these conditions with time (flow fluctuations, FF) result in the alternance of massive-laminated or laminated-rippled beds.

In consequence, in long lived and fluctuating hyperpycnal flows (FF, Fig. 15), the cyclic alternance between S1-S2 (Figs. 14 C & 14 E) is produced by cyclic changes in the rate of fall out, while the alternance between S2 and S3 is related to cyclic changes in flow velocity (Fig. 15). The alternance between massive and rippled sandstones is not hydraulically possible.

In flows having a lateral anisotropy on velocity and rate of sediment fallout, the unbalanced grown of single laminae can result in low angle and laterally diverging beds bounded by internal and discontinuous erosional surfaces (facies S2 h, Figs. 4 A and 14 D). These forms are similar to isotropic hummocky cross stratification related to combined flow (Harms et al., 1975, 1982; Southard, 1991; Mutti et al., 1994, Morsilli & Pomar, 2012), but can also occur in lacustrine systems (Fig. 14 D).

3.2.3 Facies S3

Facies S3 is composed of tabular to irregular fine-grained sandstone bodies with climbing ripples (Fig. 16). Individual sets are up to 5 cm thick with asymptotic lower foresets, commonly showing mica and small plant remains (Figs. 16 B & 16 D). If the angle of climb is maintained for long periods, the bedset can show “pseudo cross bedding” (Fig. 16 A).

Flume experiments demonstrate that current ripples are stable bedforms developed at the base of

streamflows moving at low velocities (0.1 – 0.4 m/s) over sandy substrates (Guy et al., 1966). The existence of ripples not always implies deposition, since ripples can occur in erosional, transfer or depositional context. A critical point to understand the deposition from ripples is the angle of climb (see the discussion in Harms, 1975; 1982). The angle of climb is directly proportional to the rate of fallout. In the case of climbing ripples, the high angle of climb (critical or supercritical climbing ripples, Hunter 1977) implies that sediments accumulated in the ripple foreset not derived from erosion at the stoss side, but are supplied from an overpassing and waning turbulent flow. Consequently, climbing ripples are considered a diagnostic sedimentary structure indicative of turbulent flows with high suspended sandy load (Jopling and Walker, 1968; Ashley et al., 1982; Mulder and Alexander, 2001; Sumner et al., 2008; Jobe et al., 2012). Facies S3 often grades horizontally and vertically with facies S2 (Sanders, 1965; Zavala et al., 2006 b), thus evidencing a common origin for both facies, controlled by fluctuations (FF Fig. 16 B) in the velocity of the overpassing turbulent flow. In shallow water environments affected by combined flows, ripples can show aggrading wave structures (Facies S3 w, Fig. 16 E) suggesting sediment fallout associated with unidirectional to oscillatory flows.

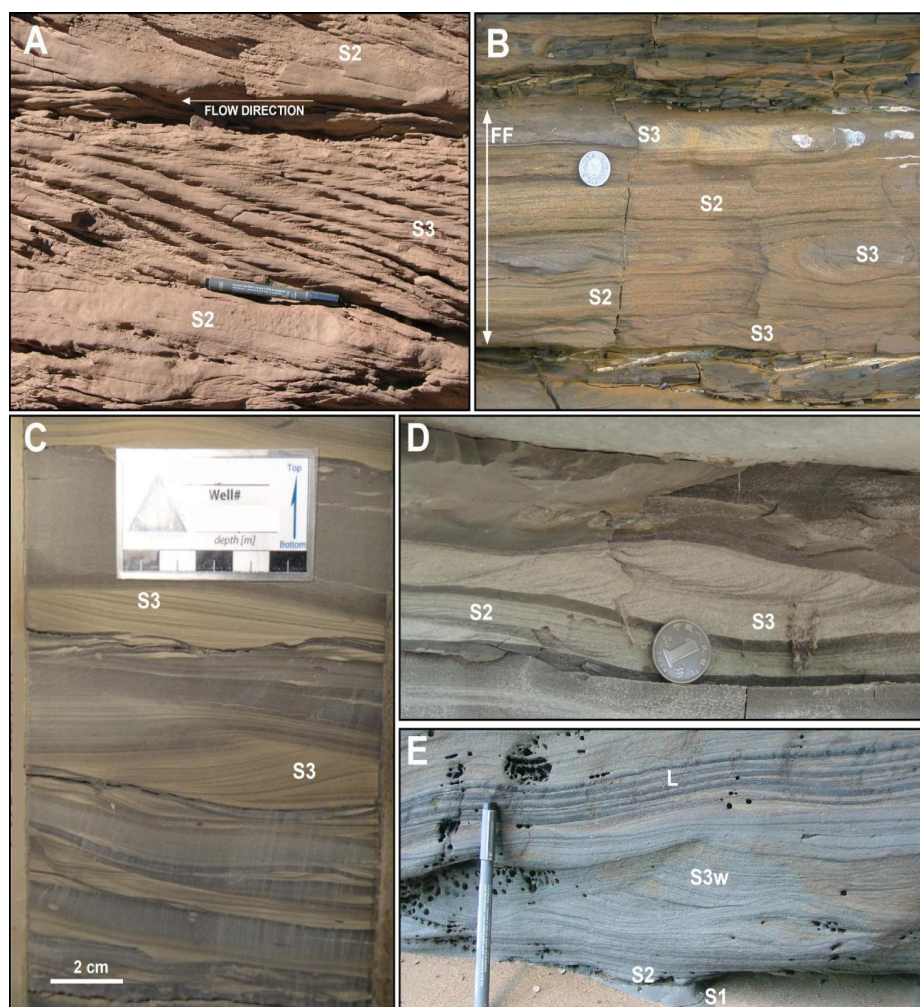


Figure 16: Selected field and core examples of facies S3. A) Fine-grained sandstones with climbing ripples (facies S3) interbedded with laminated sandstones. Note the high angle of climbing that provokes a “pseudo cross bedding”. Lower Cretaceous Rayoso Formation, Neuquén Basin Argentina. B) Cyclic alternance between rippled (S3) and laminated (S2) sandstones, suggesting velocity fluctuations (FF) in the parent flow. Eocene Pampatar Formation, Margarita Island, Venezuela. C) Climbing ripples grading upward into fallout structures. Lower Cretaceous Vikulovskaya Formation, Siberian Basin, Russia. D) Climbing ripples. Black lines are plant debris. Triassic lacustrine deposits in the Ordos Basin, China. Wave modified climbing ripples (S3 w) with abundant plant remains. Upper Pliocene Morne L’Enfer Formation, Trinidad & Tobago.

3.2.4 Facies S4

Facies S4 is characterized by massive to laminated graded siltstones and mudstones, disposed over sharp or erosional basal boundaries. It is composed of the finest materials transported by the hyperpycnal flow, which accumulated by normal settling when the flow completely stopped. Consequently, facies S4 is useful to identify boundaries between different hyperpycnal events. The differentiation of facies S4 from prodelta / shelfal mudstones may be difficult, and sometimes requires micropaleontological studies, since facies S4 often contain a mixture of continental/shallow water species. In coarse grained

(sandy-gravelly) systems this facies is typical of non-marine hyperpycnal flow deposits, because in marine settings fine grained materials are commonly lofted-up by the buoyancy provided by the lighter interstitial fluid (freshwater) and accumulate as facies L (Fig. 17 A & 17 B). S4 facies is also important in muddy hyperpycnal flows in marine settings. Since these fine grained hyperpycnal flows are mostly loaded by a mixture of silt and mud, the loading provided by the sediment concentration does not directly depend on the flow velocity, and flow reversion (lofting) does not occur (Fig. 17 C).

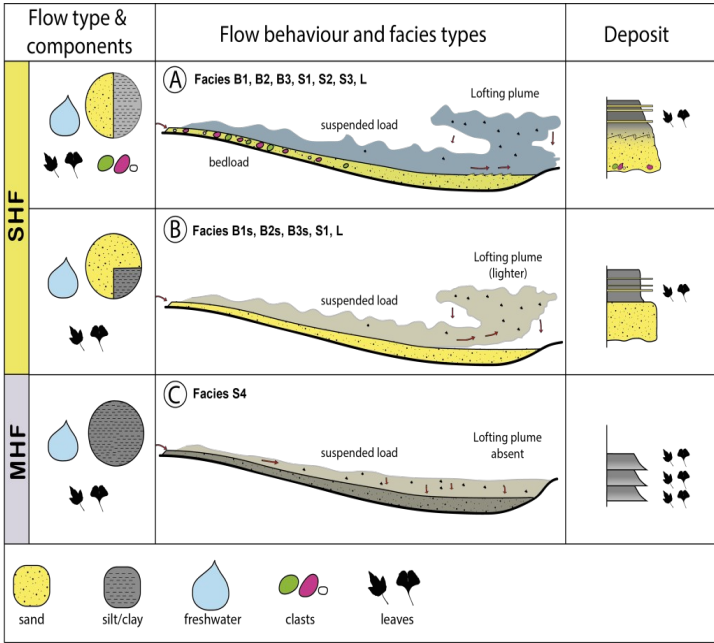
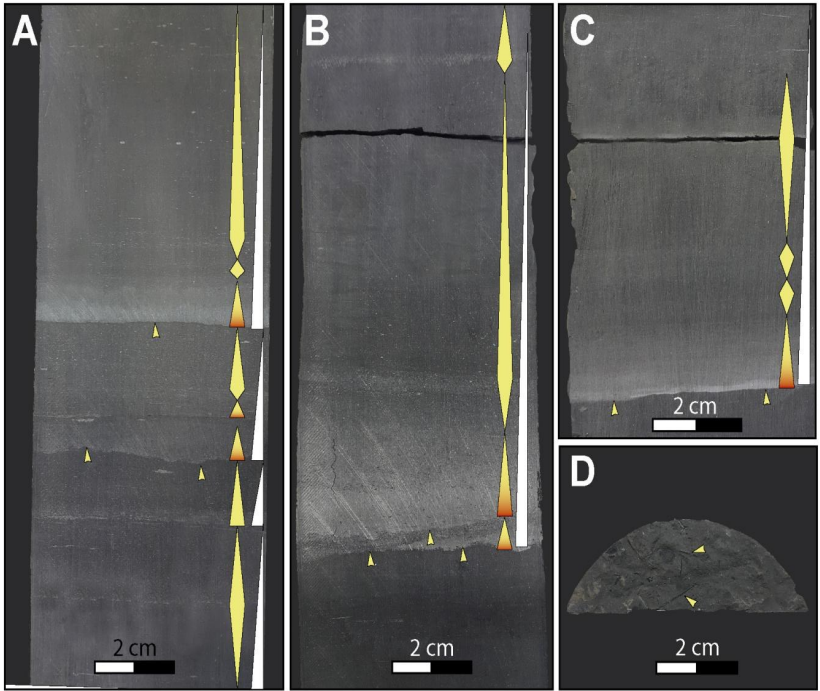


Figure 17: Composition of hyperpycnal flows and their resulting deposits in marine settings. A and B) Sandy hyperpycnal flow with (A) or without (B) associated bedload. C) Muddy hyperpycnal flow. According to the abundance of silt/clay fractions in the parent flow in A, the flow lofts at very low velocities allowing the development of plane bed and ripples on top of the bed. In B, the low content of silt/clay results in lofting at higher velocities respect to those required for develop parallel lamination and ripples. In consequence the resulting bed will be composed of massive sandstones (S1) overlaid by lofting deposits. Note also that density reversal (lofting) doesn't occur in muddy hyperpycnal flows (C) since the flow bulk density don't decrease with flow velocity. Modified after Zavala & Arcuri, 2016.

The final deposit of muddy hyperpycnal flows will be centimeter to meter thick graded mudstone beds, often disposed over sharp or erosional basal boundaries (Fig. 18). Internally, these deposits can show a mixture of intrabasinal and extrabasinal fossil remnants (Fig. 18 D). In contrary to lofting deposits (where plant debris are disposed along plane beds) , plant remains are commonly dispersed within the mudstone beds (Fig. 19). The existence of erosional basal boundaries, flames, and cyclic changes in grain size suggests that muddy hyperpycnal flows are highly dynamic currents (Fig. 19). These evidences

are against the traditional model of “normal fallout” previously assumed as the main depositional mechanism for the accumulation of mud in offshore areas. Muddy hyperpycnal flows are fluid flows capable of eroding the muddy soupy substrate. The internal cyclic changes in the grain size suggests also the existence of long living “pulsating” flows. These flows are an efficient mechanism for eroding, transferring and accumulate intrabasinal an extrabasinal organic matter in the inner basin. The rapid burial contributes in preserving the organic matter from oxidation at not totally anoxic bottoms.

Figure 18: Selected core examples of S4 facies in cores of the Upper Jurassic – Lower Cretaceous Vaca Muerta Formation, Neuquén Basin, Argentina. A-C) Stacked beds related to muddy hyperpycnal flows. Note the basal erosional contact and the overall normal grading of the deposit. Internally, these deposits commonly show subtle grain-size changes that suggests pulsating flows. D) Detail of plant remains on plant view (arrows) suggesting an extrabasinal origin of these beds.



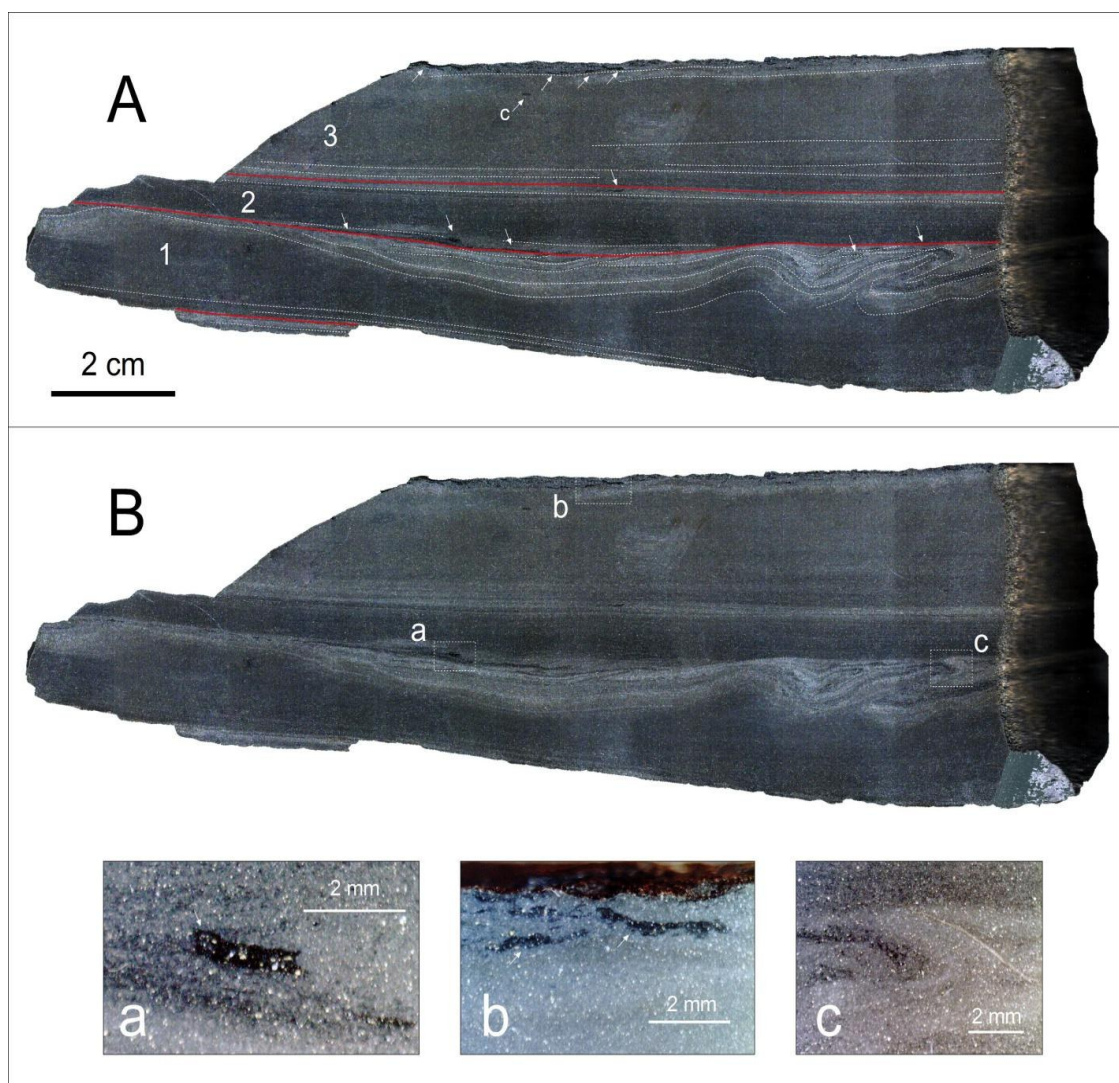


Figure 19: Example of deposits related to muddy hyperpycnal flows (Facies S4). A) Line tracing of the three graded beds shown in B. Note the erosional surface at the base of bed 2, and the flames at the top of bed 1, suggesting an erosional fluid muddy flow moving from left to right. Flames in 1 also indicate that the flow was moving over a “soupy” substrate. Plant remains are abundant. B) General view showing the location of photomicrographs a, b and c, with examples of small and dispersed plant debris. Lower Jurassic Los Molles Formation, Neuquén Basin, Argentina. After Zavala & Arcuri, 2016.

3.3 Facies related to flow lofting (facies L)

Facies L is characterized by widely extended thin (millimeter thick) couplets of siltstones and very fine grained sandstones, often showing abundant plant debris and micas on bounding surfaces (Fig. 20). These couplets display abundant, small scale load cast structures, commonly associated with syneresis cracks and siderite nodules. Trace fossils are scarce and mostly limited to some forms of *Palaeophycus*. These laminated levels can form bedsets up to 0.5 meters thick (termed lofting rhythmites, Zavala et al., 2006 c; 2008 b) alternating with shelfal or hemipelagic shales, but more commonly associated

with massive sandstones (facies S1, Figs. 20 C, 20 D).

Individual levels display a variable thickness from few millimeters up to one centimeter, and are separated by thin layers with abundant plant debris (Figs. 20 B and 20 E). Plant debris provides a definitive evidence of a direct connection between the fluvial system and the marine/lacustrine related basin (Petter and Steel, 2005; Lamb, 2008; Zavala et al., 2012). The absence of tractive structures in these sandstones suggests an accumulation by normal settling from a suspension cloud elevated over the depositional surface.

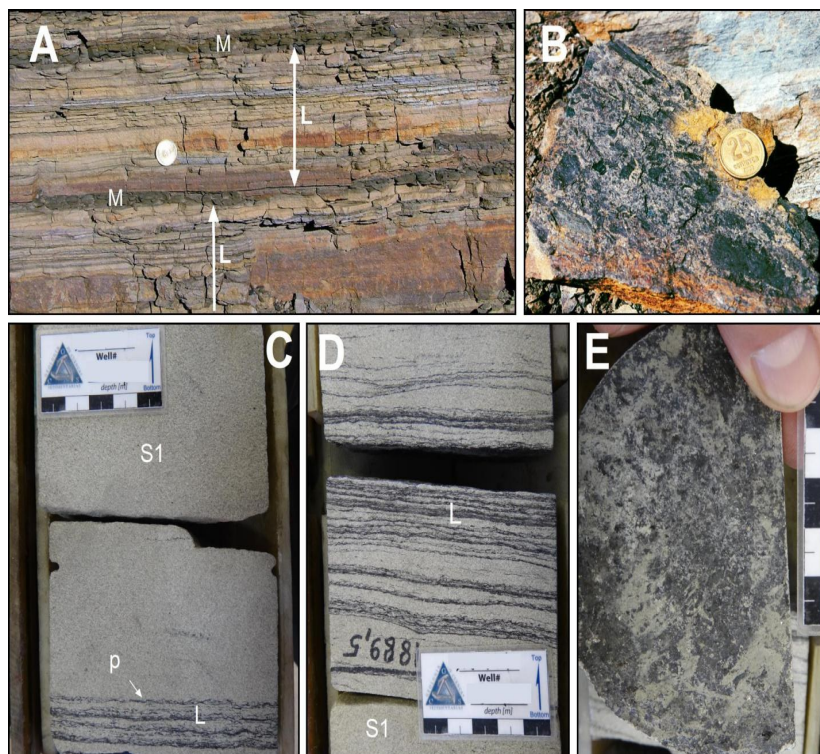


Figure 20: Field and core examples of facies L (lofting rhythmites). A) Deposits of lofting rhythmites (L) interbedded with shelfal mudstones (M). Inner shelf deposits of the Lower Jurassic Los Molles Formation, Neuquén Basin, Argentina. B) Plant view of lofting deposits shown in A. Note the abundant plant remains. C & D) Lofting deposits (Facies L) interbedded with massive sandstones (Facies S1). Black lines are plant remains. Cretaceous of the Siberian Basin, Russia. E) Plant view of D, with abundant mica and plant remains.

Lofting rhythmites accumulate from a pulsating lofting plume, which is a typical feature of hyperpycnal inflows in marine environments. A hyperpycnal flow is basically a heterogeneous mixture of components having different individual densities. According to their lower density respect to marine waters (averaged surface density 1.025 g/cm^3), freshwater (1.0 g/cm^3), charcoal (0.208 g/cm^3) and plant debris ($0.09\text{--}0.55 \text{ g/cm}^3$) are considered buoyant components (Fig 22). On the other hand, sand ($2.2\text{--}2.8 \text{ g/cm}^3$), silt ($2.4\text{--}2.8 \text{ g/cm}^3$) and clay ($2.7\text{--}2.8 \text{ g/cm}^3$) are load components. Under certain conditions of velocity, these components are mixed in different proportions within a turbulent flow providing an average bulk density. If this bulk density is above the threshold of sea water density, the flow will be attached to the sea bottom. On the contrary, if the flow loses part of the load components (i.e. sand - silt) by deposition, the flow will lift from the substrate through buoyancy reversal (Sparks et al., 1993; Kneller & Buckee, 2000; Mutti et al., 2003; Hesse et al., 2004; Hesse and Khodabakhsh, 2006), forming lofting plumes charged with fine-grained sediments, plant debris and micas

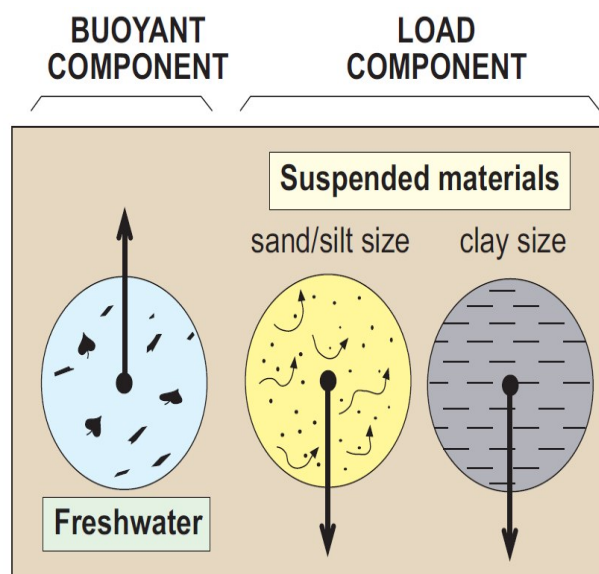


Figure 21 Buoyant and load components as the basic constituents of a turbulent hyperpycnal flow. After Zavala et al., 2011.

Lofting rhythmites mostly accumulate at flow margins (Zavala et al., 2011, Fig. 4 B) and show a more extended distribution compared with that of the main hyperpycnal flow. Facies L constitutes a diagnostic and characteristic element of hyperpycnal sedimentation in marine settings since it suggests the existence of a less dense interstitial fluid (freshwater) with respect to that of the environment (marine wa-

ter), derived from direct fluvial discharge (Zavala et al., 2006 c; 2012). Vertical and lateral relationships between L and S (S1, S2 and S3) facies are not sharp, and result in transitional categories (Fig. 4 B) termed S1/L (massive sandstones with discontinuous

siltstone levels (Fig. 22) , S2/L (laminated sandstones with abundant plant debris and micas) and S3/L (siltstone levels interbedded with sandstones with small and low angle climbing ripples).



Figure 22: Outcrop showing the lateral relationship between massive sandstones (facies S1) and lofting (facies L). Transitional levels include massive sandstones with levels of plant debris (facies S1/L). Lower Jurassic Los Molles Formation, Neuquen Basin, Argentina.

Analysis of thin sections (Fig. 23) of lofting rhythmites provides clear evidences about the origin of this feature apart of giving some diagnostic criteria for its recognition (Zavala et al., 2008 b; 2012). According to its accumulation from normal settling, the deposition of fine-grained materials from lofting

plumes is very selective. The free fallout of fine grained clastic materials from suspension clouds is governed by the Stokes' law. The genetic analysis of the thin section of Fig. 23 allows to tract the origin of lofting rhythmites.

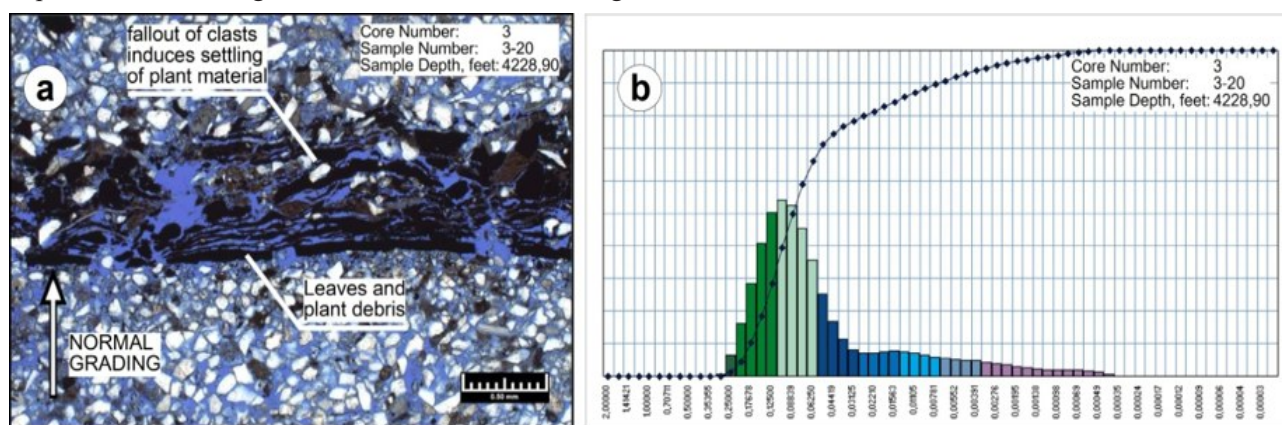


Figure 23: Thin section (a) and grain-size analysis (b) of lofting deposits in the Pliocene of the Columbus Basin, Trinidad & Tobago. Fallout of sand-silt materials from suspension clouds are normally graded, and are bounded by thin levels of plant debris, trapped by the next wave of sand fallout. After Zavala et al., 2012.

Fig. 24 shows a step-by-step accumulation of lofting rhythmites, according to the interpretation of the microphotograph shown in Fig. 23. In (A) a heterogeneous lofting cloud is introduced by a hyperpycnal wave, in a lateral position respect to the main flow. In (B) the free fallout of different grain-sized clastic materials results in a normally graded interval with silt and oriented micas on top. According to their

low density, leaves and plant materials remain in suspension. In (C) , a new suspension cloud is introduced by a new wave during the same hyperpycnal discharge. The free fallout of largest sand grains forces the deposition of leaves and plant fragments. As a consequence, a thin level of carbonaceous materials develops in between massive sand-silt levels.

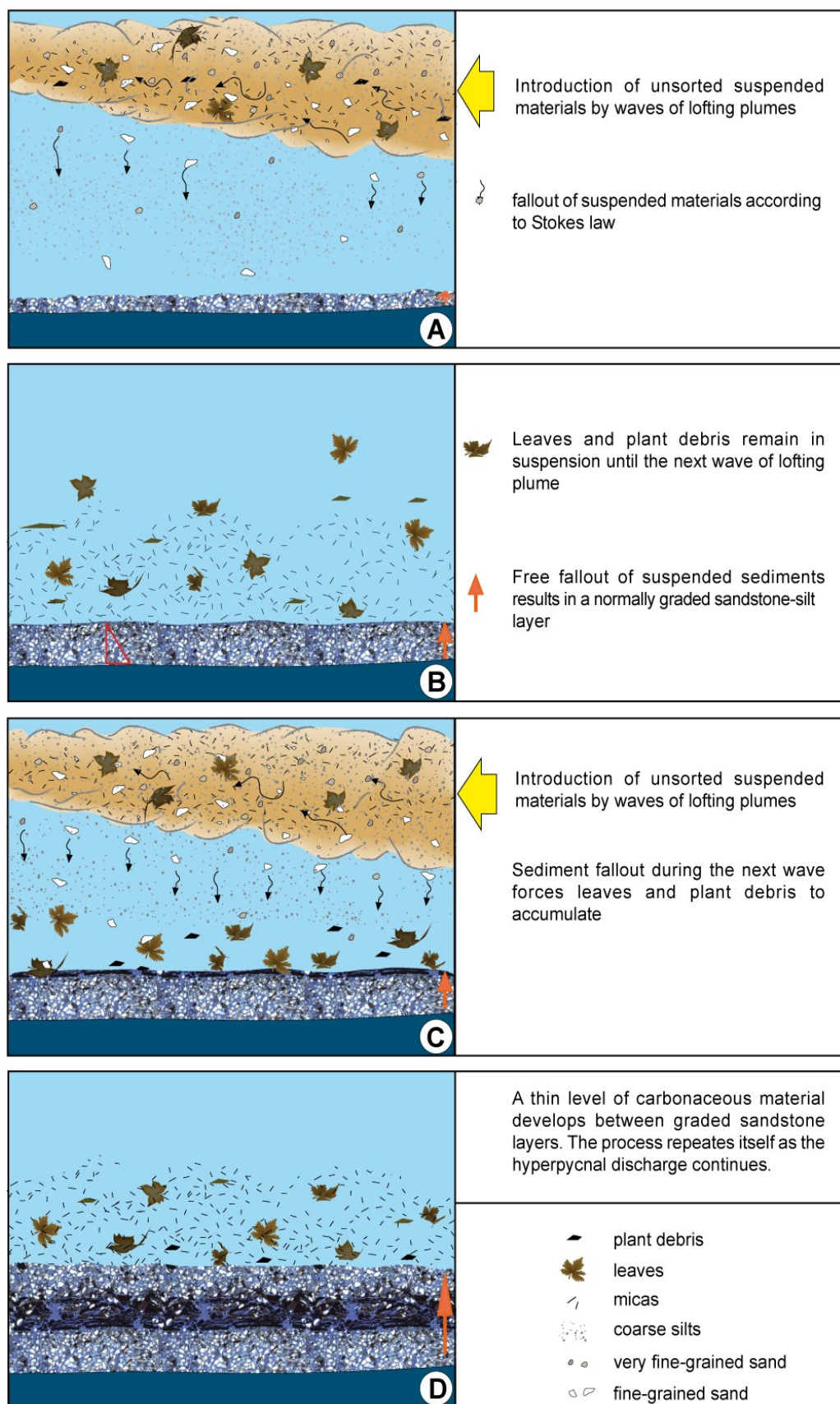


Figure 24: Lofting rhythmites are the result of the repeated aggradation of fine grained materials from suspension clouds related to the buoyant inversion of hyperpycnal flows at flow margin areas. Explanations in the figure. Modified from Zavala et al., 2008 b; 2012.

4 Depositional elements: channel fill, lobes and levee deposits

In contrary to intrabasinal (surge like) turbidites, sustained hyperpycnal discharges result in very dynamic jet flows (in the sense of Hoyal et al., 2003) capable of travel for long distances on gentle or flat basin slopes. The energy that allows such long travel is provided by the continuous pumping from the river flood. Consequently, the distance achieved by the hyperpycnal discharges will directly depend on the duration of the associated river discharge and the topography of the related basin. According to the common accumulation of climbing ripples at the leading head of sandy hyperpycnal flows, it is assumed an average velocity of advance of the hyperpycnal flow of 20 cm/s (climbing ripples develop at flow velocities ranging from 15 to 25 cm/s, Ashley et al., 1982). At this velocity, a sustained river discharge of about two months will be required to achieve a distance of 1,000 km along sea bottom. In contrary to that of the leading head, the velocity along the flow body can be

high, according to the evidences provided by sedimentary structures and internal erosional surfaces. The understanding of the basinward evolution of a hyperpycnal flow is crucial to understand the final depositional geometry of channels fills and lobes. The initial flow path will be traced by the advance of the leading head, which will move basinward trying to reach the lower topographic positions of the basin landscape. In this context, the existence of structurally controlled closed topographic lows will result in a flow “trapping” and the forced accumulation of very thick and discontinuous massive sandstone bodies (due to the associated high rates of fallout). In the case of more regular (gentle or flat) landscape, the acceleration of these jet flows along the flow body result in erosion and channel incision. The conceptual step by step evolution of hyperpycnal channel fill sandstone bodies comprise three stages: (1) initial deposition, (2) erosion & bypass, and (3) channel fill (Fig. 25).

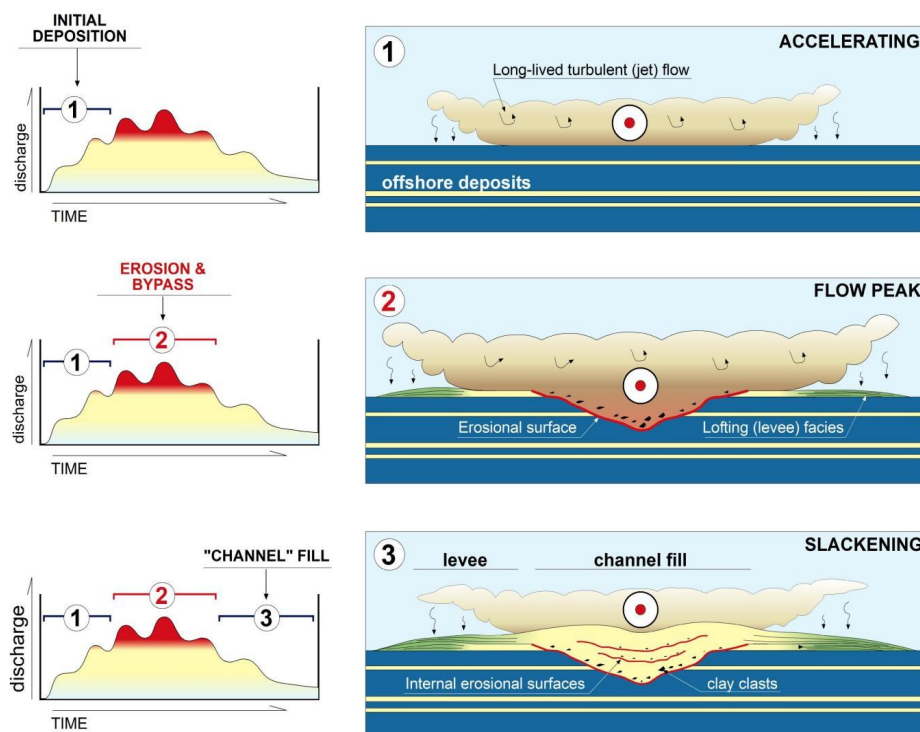


Figure 25. Depositional schema showing the step by step origin of hyperpycnal channel fill deposits. In 1 the arrival of the flow leading head accumulated traction plus fallout deposits. In 2, high velocities at flow axis produces a flute-like basal erosional surface. In 3, during flow slackening, the hyperpycnal flow infills the previous erosional relief. Modified from Zavala et al., 2008 b.

The initial deposition stages are characterized by traction plus fallout deposits (S3-S2 facies) and associated levee (L facies) accumulated during the flow acceleration immediately after the arriving of the flow

leading head (1 in Fig. 25). The progressive increase in the flow velocity at the flow axis results in strong turbulence at the jet region. Once the erosional threshold is exceeded (2 in Fig. 25), the erosion of the sub-

strate provokes a linear flute- like erosional scour (Hoyal et al., 2003) with associated levee deposits (L facies) at both sides. This second stage is characterized also by an important sediment bypass, allowing the basinward transfer of a huge volume of sediments. The interplay of erosion, temporary deposition and bypass in this stage results in bedload (B1,

B2 and B3 facies) and suspended load (S1 facies) deposits at the channel bottom, often associated with lateral accretion (in case of meandering channels). The Lateral accretion associated to hyperpycnal channel fills is characterized by an aggrading component (Fig. 26).

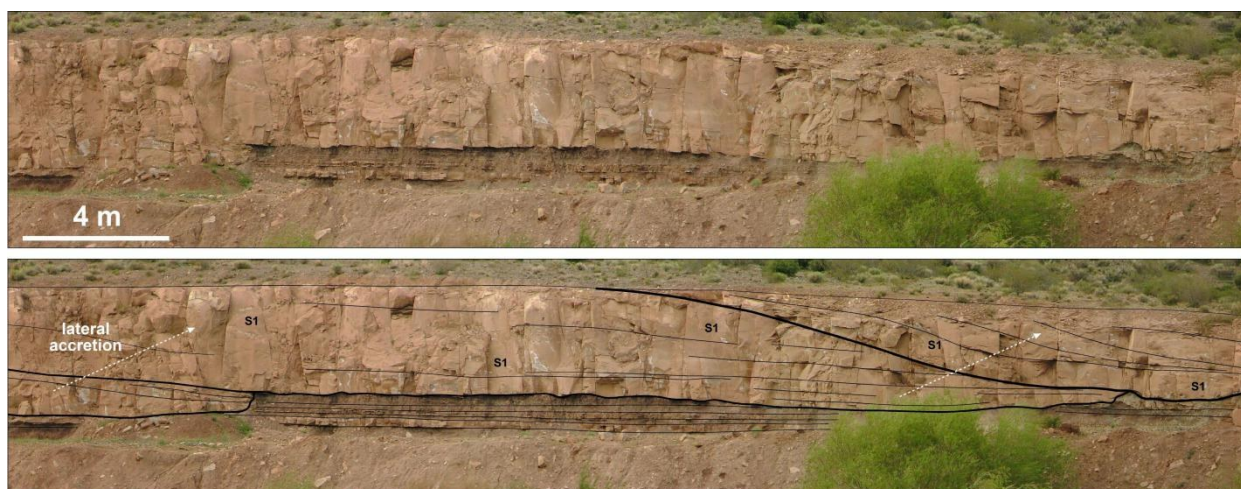


Figure 26. Field example of lacustrine hyperpycnal channel fill with lateral accretion. Note that in contrary to lateral accretion of subaerial fluvial channels, this kind of lateral accretion has an aggrading component. Flow direction is perpendicular to the outcrop. Lower Cretaceous Rayoso Formation, Neuquen Basin, Argentina.

The final channel infill occurs during stage 3, which is characterized by an overall fluctuating waning flow. The associated deposit comprises sand size materials collapsed from the sustained turbulent flow (S1, S2 and S3 facies) with associated levee deposits at flow margins (L facies). In contrary to fluvial channel fills, the evolution and infill of hyperpycnal channels result in positive topographic elements on the sea bottom. This sedimentary induced relief will control the location of the future hyperpycnal channels, resulting in compensation cycles in the sense of Mutti & Sonnino (1981). Compensation cycles provides a very complex scenario for detailed correlation of sandstone bodies.

The transition from channels to lobes occurs when the sustained hyperpycnal flow loses the capacity of eroding the substrate. In topographically (fault) controlled depocenters, the flow could be forced to wane and accumulate according to fill and spill processes. The “fill and spill” process is relatively well known from the study of deposition related to dilute sediment-gravity flows (turbidites) on topographical-

ly irregular basins (or minibasins) located on the continental slope of the Gulf of Mexico and the Tertiary Alpine basins (Winker, 1996; Sinclair & Tomasso, 2002; Toniolo et al., 2006). This process involved at least four stages: (1) Flow ponding, where incoming flows are totally trapped, depositing thick, sheet-like sand – mud couplets. (2) Flow stripping, where the finer, more dilute portion of the flow is able to escape over the confining topography to be deposited elsewhere. (3) Flow bypass, when flows traversing over the filled basin and transfer the sediments into a deeper position. (4) Backfilling of the basin during the final stage, usually with meandering channel – levee complexes with low sand/mud ratios (Sinclair & Tomasso, 2002). Fig. 27 depicts the process of fill and spill in the offshore of the Columbus Basin, in Trinidad & Tobago. During the initial infill of the fault controlled depressions (Fig. 27 A) the hyperpycnal discharge flows along the axis of the depocenter (1 in Fig. 27 A), and the associated deposits compose thick intervals of massive and laminated sandstones (Fig. 28). The infill of the former depocenter

forces the flow to bypass this area and accumulate in a deeper area (Fig. 27 B). In consequence, the final

accumulation in the first depocenter will be dominated by fine grained lofting deposits (2 in Fig. 27 B)

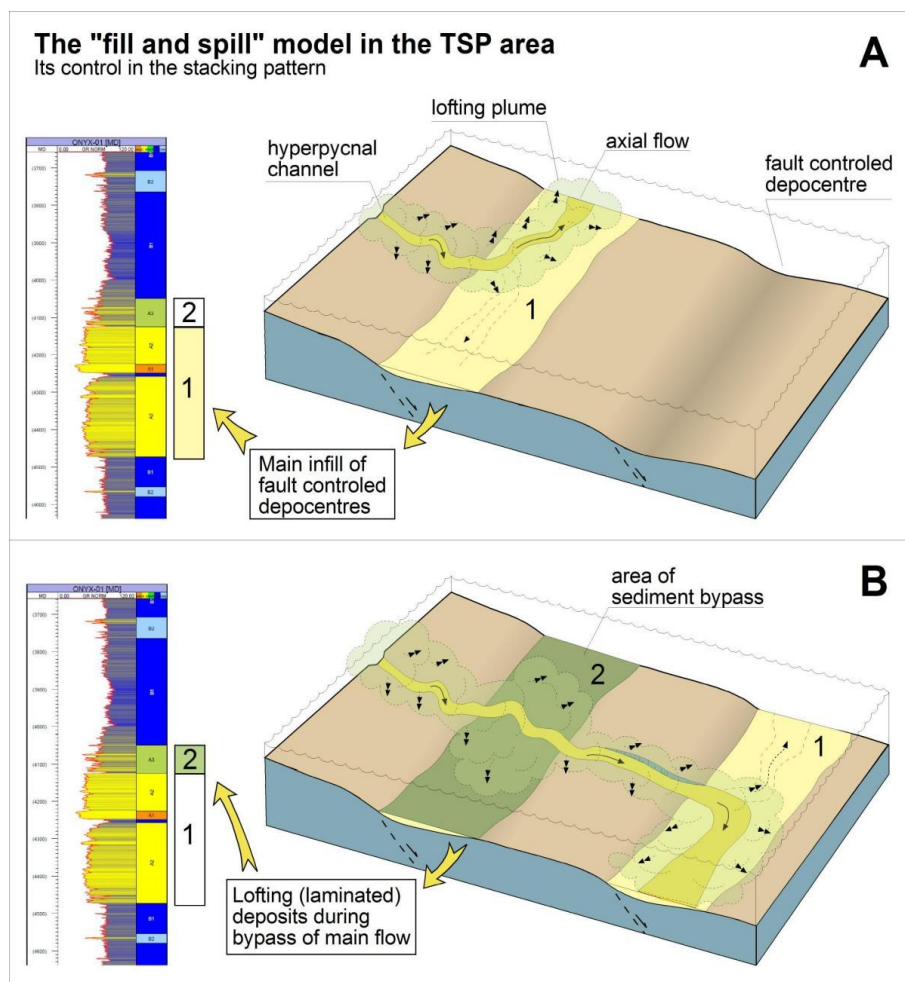


Figure 27. 3 D model showing the progressive infill of fault controlled depocenters and its control in the stacking pattern of the associated reservoirs. "A" depicts the initial and main infill, resulting from the accumulation of thick sandstone packages (1 in the well log) by axial hyperpycnal flows. Once the first depocenter is almost filled ("B"), the flow bypass towards a deeper depocenter, resulting in the accumulation of lofting deposits (2 in the well log) at the top of the previously filled depocenter.



Figure 28. Thick massive sandstone bodies in the Mayaro Formation (Pliocene), Columbus Basin, Trinidad & Tobago. Thick massive sandstones result from the accumulation in structurally controlled basin floor depressions.

In more topographically regular basins, channels usually extend for 10's to 100's of kilometers until the transported sediments spread out as terminal

lobes. Fig. 29 provides an exceptional example for discussing the origin and distribution of hyperpycnal channels, and their relationship with littoral deltas. This marvelous seismic time slice correspond to Late Cretaceous shallow to deep water lacustrine deposits of the Nengjiang Formation in Songliao Basin, China. Littoral deltas clearly develop in the northern area, from where hyperpycnal flows are triggered during exceptional river discharges (river flood). Initially these channels are straight, but become meandering at the lower slope. Integration of cores, well logs, and seismic allows to estimate an averaged channel width of 500 meters (maximum 900 meters). Channels extend up to 80 km from littoral delta areas, and show a sinuosity ranges from 1.35 up to 1.71) average 1.54. The channel end is characterized by more laterally extended lobe deposits, typically covering an area from 4 to 10 km². It is interesting to

note that in hyperpycnal systems main sandstone accumulation not necessarily occur at lobe area. Chan-

nel fill deposits constitutes also important clastic reservoirs.

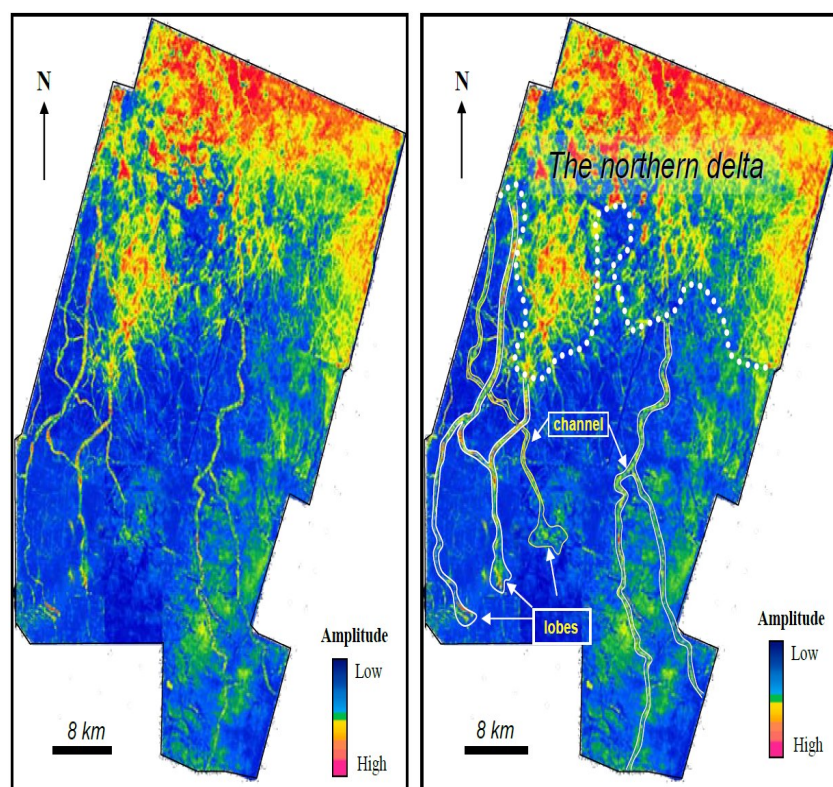


Figure 29. Time slices of Late Cretaceous shallow to deep water lacustrine deposits of the Nengjiang Formation in Songliao Basin, China. Littoral delta deposits clearly develop in the northern area, from where hyperpycnal flows are triggered during periods of high river discharge (flood). Note that initially hyperpycnal channels go straight, and become meandering at the lower slope. Note that sandstone accumulation occurs in the final lobes and along the channels. After Shuxin et al. (2017).

5 Discussion and conclusions

Growing evidences suggests that hyperpycnal flows are, and have been, a very common and important mechanism for transferring huge volume of sediments from producing areas to marine and lacustrine depositional basins. Nevertheless, their related deposits are still poorly recognized in the literature. Our ignorance of present clastic depositional environments (especially subaqueous ones) resulted in an overemphasis of “geomorphological environments” having little or no preservation potential in the geological record. As an example, a number of littoral geomorphologic features (like strand plains, estuaries, tidal flat, barrier island, etc) have been created during the Holocene by the rapid marine flooding (more than 120 m of sea level rise) over a mature continental landscape. These temporary geomorphologic features have been then “upgraded” to the category of “depositional system”, and their sedimentary models have been applied to explain the origin of very thick successions of clastic deposits. Consequently, it is highly probable that a number of lacustrine and marine hy-

perpycnites were wrongly described and interpreted in the literature as fluvial, shoreface, estuarine, or strand plain deposits. The correct interpretation of the significance of these deposits will require further studies.

This paper provides a synthesis of the main characteristics of hyperpycnal flows and their deposits. Nevertheless, our understanding of these deposits is still very limited, and new studies and effort will be required in the next years to achieve a complete knowledge of the characteristics and significance of hyperpycnal systems.

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