Tidal and fluvial controls on the internal architecture and sedimentary facies of a lobate estuarine tidal bar (The Plassac Tidal Bar in the Gironde Estuary, France)

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Estuarine tidal bars emplaced in estuaries are complex sediment bodies composed of sand and mud provided by rivers. This study focuses on the Plassac Tidal Bar, which is the unique lobate-shape tidal bar of the Gironde Estuary, located at the very upstream extremity of the bay-head delta, because no detailed sedimentological studies have been made on such geomorphological category of tidal bar. Its stratigraphy was investigated using a dense grid of 65 very high resolution seismic profiles, ground-truthed by 6 to 9 meter-long cores and time-controlled by historical bathymetric surveys.

The base of the tidal bar consists of a tidal erosional surface. This surface has eroded muddy tidalites, corresponding to the maximum flooding surface of the Gironde Estuary. The bar is composed of two vertical sequences of facies, composed of thickening-up cross stratified fine- to medium-grained sand beds at the top and decimeter-thick mud-layers at the base. Those mud layers constitute the master bedding of the bar and are dipping in a channelward direction, perpendicularly to the tidal flows and to the sediment transport directions. One phase of lateral accretion of the Plassac Tidal Bar is correlated with variations in fluvial discharge of the last decades and suggests a climate control on the internal architecture of this sandbar. Low water periods lead to the upstream migration of the turbidity maximum and to the deposition and consolidation of the muddy master bedding planes. During the subsequent flood periods, the muddy master bedding planes are partially eroded and then buried by the fluvially-sourced sandy mini-flood lobes that merge with the tidal bar.

The internal architecture of the Plassac Tidal Bar is controlled by a lateral accretion process, its master bedding surfaces which record this process are dipping perpendicularly to the axis of the tidal flows. All the other tidal bars of the bay-head delta (around a dozen) exhibit an elongated morphology and prograde longitudinally by seaward frontal accretion occurring in the axis of the tidal flows. The master bedding surfaces which record this process are seaward dipping along the axis of the tidal flows. Those variations in internal accretion surfaces orientation of tidal bars are of considerable importance and must be taken into account for studies of ancient tidal deposits, paleoenvironment reconstructions and for heterogeneities characterization and quantification of reservoir properties in these types of sandstones.

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1. Introduction

Estuarine tidal bars emplaced in bay-head deltas of estuaries are mainly made of sand, but they also contain large quantities of mud, both of those sediments being provided by rivers. For that reason, estuarine tidal bars emplaced in bay-head deltas clearly differ from estuary mouth bars of truly tide-dominated estuaries, made of clean sand mostly coming from the sea. Tidal bars of the bay-head delta built a regressive wedge in the inner part of the estuary during highstand conditions. Thus, their sedimentary record is not controlled by sea level variations but more likely by: (1) tidal currents and related transport; (2) climate changes, through variations in fluvial discharge and sediment supply; (3) anthropogenic activities, like deforestation or dams leading to changes in fluvial sediment supply (Fénies and Tastet, 1998; Billy et al., 2012). Therefore tidal bars appear as key sediment bodies for the sediment record of environment changes that occurred during the last centuries or millenaries, like other highstand coastal sediment bodies including deltas (Ta et al., 2002; Yi et al., 2003, 2006), estuaries (Anderson et al., 2008), coastal barriers (Billeaud et al., 2009), lagoons (Freitas et al., 2003), lakes (Behling and da Costa, 2001) and wetlands (Goff and Chagué-Goff, 1999), sandspits (Allard et al., 2008) or mud flats (Poirier et al., 2011). Beyond this academic interest, tidal bars emplaced in bay-head deltas are of
considerable importance as they represent analogues for many oil reservoirs (Zaitlin et al., 1994; Hubbard et al., 2002), particularly those characterized by heterogeneities (Jackson et al., 2005; Ringrose et al., 2005). They also form hazards for shipping, they contain large quantities of sand for sand extraction and they are nursery for many fish species (Applied Marine Sciences, 2009).

The internal architecture of estuarine tidal bars is still not well known as seismic investigations remain a challenge in shallow water areas where strong tidal currents occur. Indeed high resolution seismic investigations on modern environments showing the detailed internal architecture of tidal bars and the hierarchy of bounding surfaces are scarce (Dalrymple and Zaitlin, 1994; Berné et al., 2002; Chaumillon et al., 2002; Dalrymple et al., 2003; Chaumillon et al., 2008). Most of the data available on tidal bars result from studies conducted on ancient environments (Mutti et al., 1985; Willis and Gabel, 2003; Olariu et al., 2012; Legler et al., 2013) or from cores obtained in modern environments (Feniés and Tastet, 1998). Moreover there is still an urgent need of case studies in present-day environment in order to improve facies models (Dalrymple, 2010).

This study aims at showing the internal architecture and sedimentary facies of an estuarine tidal bar from a dense grid of seismic profiles, ground truthed by vibro cores and time-scaled by bathymetric surveys, made almost every year since about one century. We focus on a lobate-shape tidal bar belonging to the Gironde Bay-head delta, the Plassac Tidal Bar, as no detailed sedimentological studies have been made on such geomorphological category of tidal bar. This study that is focused on the internal architecture and sedimentary facies of the Plassac Tidal Bar, based on seismic profiles and cores, complements the article of Billy et al. (2012) based on the bathymetry and the morphological evolutions of this tidal bar.

2. Study area

2.1. Physiography and hydrology of the Gironde Estuary

The Gironde Estuary is located in south-western France (Fig. 1A). This 75 km-long estuary is the largest one in Europe. It displays a typical funnel shape; it is 12 km-wide at the maximum close to its mouth and 2 km-large upstream at the confluence of the Garonne and Dordogne Rivers (Fig. 1B). It is a hypersynchronous estuary with tidal amplitude ranging from 1.5 to 5 m at the mouth to 3 to 6 m at Bordeaux (Fig. 1B). The tidal currents are characterized by a landward-increasing time–velocity asymmetry, which results in longer and weaker ebb currents and shorter and stronger flood currents (Salomon and Allen, 1983; Allen, 1991). Two rivers flow within the Gironde Estuary: the Dordogne and the Garonne Rivers, respectively corresponding to 1/3 and 2/3 of the total fresh water discharge. Periods of high river discharge occur during the winter floods (average flow rate above 2500 m³/s and up to 5000 m³/s) and periods of low river discharge occur during the summer months (average flow rate less than 200 m³/s). During low river discharges, tidal-current reversal can occur up to 160 km from the estuary mouth. Within the estuary, wind waves are relatively small, typically less than 1 m, due to the small fetch.

2.2. Sedimentation pattern of the Gironde Estuary

A large volume of fluvial sand and mud is supplied to the estuary by the Garonne and Dordogne Rivers (Castaing, 1981). All the fluvially sourced sand is deposited within the estuary (Allen, 1972) and 75% of the fluvially sourced suspended silt and clay is deposited within the estuary. Sediment within the estuary form a distinctive longitudinal facies pattern (Allen, 1991) characterized by a tripartite morphological zonation successively comprising, (1) sand and mud point bars in the upper estuary channels and elongated tidal sand bars in the bay-head delta, (2) estuarine mud in the central basin and (3) coarse sand and gravels in the estuary mouth, that are sourced from the adjacent oceanic coast by wave erosion and littoral drift. From the hydrodynamic parameters, its morphology and sedimentation pattern, the Gironde Estuary is a typical mixed tide- and wave-dominated estuary (Dalrymple et al., 1992).

2.3. Stratigraphy of the Gironde Estuary

The stratigraphy of the Gironde Estuary was described by many authors (Allen et al., 1970; Feral, 1970; Castaing, 1981; Castaing and Allen, 1981; Allen, 1991). A sequence stratigraphic model was developed by Allen and Posamentier (1993, 1994) and Lericolais et al. (2001). The valley-fill is a fifth order sequence which is divided in three distinct systems tracts, including: a thin lowstand several systems tracts made of fluvial sands and gravels; a thick transgressive systems tract made of estuarine sands and muds, including estuarine point bars, tidal flats and marshes, and estuary mouth well-sorted medium and coarse grained-sands seaward; a regressive hightstand several systems tracts, located in the modern estuarine funnel, made of estuarine sands and muds and constituting the estuarine bay-head delta. The landward boundary of the bay-head delta (BHD) is located at the confluence of the Garonne and Dordogne Rivers (75 km from the estuary mouth; Allen and Posamentier, 1993). It has prograded since the end of the last Holocene eustatic rise and is still prograding down the estuary. It is sourced by the present discharge from the rivers (Allen, 1991).

2.4. Geomorphology, hydrodynamics and sedimentology of the tidal bars of the BHD

Tidal bars belonging to the bay-head delta are developed in the estuary funnel located 45 to 75 km upstream of the mouth (Fig. 1B). They consist of fluvial sands brought to the estuary by alluvial floods and reworked by tidal currents. When the sand builds up to the intertidal zone, the bars are capped by supratidal mud which is then colonized by vegetation and form permanent islands. The regression of the bay-head delta is demonstrated by the island area increase as the total area of the islands mapped in 1950 was four times larger than in 1750 (Migniot, 1971).

Two kinds of tidal bar morphologies are identified in the bay-head delta which infill the upper estuary funnel: elongated (Feniés and Tastet, 1998) and lobate (Billy et al., 2012). The distribution of those morphologies, elongated downstream and lobate upstream, may be compared to the longitudinal variations in the morphology of tidal bars in the tide-dominated delta of the Fly River (Dalrymple et al., 2003), where U-shaped tidal bars (also called lobate) occur in distributary channels and elongate bars occur in the distributary-mouth-bar area. In the Gironde Estuary, lobate and elongated tidal bars show morphological evolutions that can be related to variations in fluvial discharge (Feniés and Tastet, 1998; Billy et al., 2012).

The Trompeloup Tidal Bar displayed an elongated morphology and was around 6 km long and 600 to 200 m wide in 1980. The detailed morphological evolution of the bar (Feniés and Tastet, 1998) has been reconstructed over the last century thanks to bathymetric surveys, it showed an alternation of phases of longitudinal bar progradation along the axis of the tidal currents (bar front migrates seaward 8 km during 24 years − 1901/1925) and phases of bar abandonment (bar is eroded and breached by swash channel in 1968). The Plassac Tidal Bar was 4.6 km long and 1.4 km wide in 2010, and it displayed a lobate morphology, with two elongated ebb spits partially separated by a dead end flood channel which terminates onto an ebb shield (Fig. 1C, D; Billy et al., 2012). Current measurements conducted on the Plassac Tidal Bar have shown peak current velocities larger than 1 m s⁻¹ during both spring tide flood and ebb (Allen, 1972; Feniés et al., 1999). Residual currents are strongly asymmetric, they are flood-oriented on the tidal bar, during both neap and spring tides under conditions of low water fluvial discharge, and they are ebb-oriented, in the tidal channel adjacent to the tidal bar (Allen, 1972). Radioactive
tracer experiments were conducted on the Plassac Tidal Bar under conditions of low water fluvial discharge. Those experiments have demonstrated the net sediment transport is upstream-oriented with a migration of the tracer cloud centroid of about 14 m after 8 days. The detailed morphological evolution of the bar has also been reconstructed over the last century thanks to bathymetric surveys; it showed an alternation of phases lateral bar accretion and phases of bar abandonment. The lateral accretion of the ebb spits is caused by the incorporation of downstream migrating mini-flood lobes on the outer branches of the spits and occurs during the periods of higher fluvial discharge, whereas the partial ebb shield breaching occurs during periods of lower fluvial discharge (Billy et al., 2012).

The internal architecture of the elongated tidal bar category is illustrated by the work conducted on the Trompeloup Tidal Bar located at the seaward extremity of the Gironde Bay-head delta (Féniès and Tastet, 1998). The internal architecture of this bar was revealed using a dense coring grid (more than 50 vibro cores) and a few seismic lines (Féniès, 1981). It is composed of large-scale longitudinal accreting sigmoidal sand bodies deposited during years of strong fluvial discharge, partially isolated from one another by thick mud blankets deposited during years of weak fluvial discharge. The vertical succession formed by the seaward progradation of the tidal bar records these alternating phases of growth and abandonment. The internal architecture of this tidal bar is representative of the elongated tidal bars deposited in the bay-head delta of wave- and tide-dominated estuaries. The elongated tidal bars represent the vast majority of the tidal bars deposited in the Gironde Bay-head delta (12 elongated bars, out of the 13 bars of the bay-head delta).

This paper focuses on the internal architecture of a lobate tidal bar, so far unpublished: the Plassac Tidal Bar. It is based on new very high resolution seismic profiles ground-truthed by 8 cores. This tidal bar is the only lobate one deposited in the Gironde Bay-head delta and it is located at the very landward extremity of the bay-head delta, close to the junction of the two estuarine channels: Garonne and Dordogne (Fig. 1B).

3. Methods

The internal architecture of the Plassac Tidal Bar was investigated from 65 very high resolution seismic profiles combined with 8 vibro cores. Historical bathymetric profiles where used to provide an age control on the stratigraphic units. Interferometric multibeam bathymetry was used to evidence bedforms located at the surface of the tidal bar. The timing of deposition of the main sedimentological units and discontinuities were compared to the Gironde’s fluvial discharge.

3.1. Very high resolution seismic profiling

During the GiRaFS cruise (August - September 2008), about 100 km of new high resolution seismic profiles (65 profiles, Fig. 1D) were shot on the Plassac Tidal Bar. The distances between profiles ranged between 500 and 100 m and positioning of seismic profiles was obtained using a GPS system with its antenna directly mounted on the seismic profiler. The seismic profiler was the IKB Seistec, a boomer plate associated with a line-in-cone short streamer (Simpkin and Davis, 1993). This seismic profiler dedicated to very shallow waters allowed us to record seismic profiles on intertidal areas with water depth of less than 2 m. The band pass frequency was 1 to 10 kHz giving a vertical resolution of about 20 cm. A 50 J power supply was selected. The shot interval was 250 ms, corresponding to distance between adjacent traces of about 25 cm for a ship speed of about 2 knots.
3.2. Vibro core sampling and laboratory analysis

Core sampling was carried out with a portable vibro corer (De Rességuier, 1983) during 3 legs conducted in 2009 and 2010. Three 9 m long cores (65b, 44a and 40b) and five 6 m long cores (65a, 65c, 44a, 16a, 48a) were obtained along selected seismic profiles (Fig. 1D). During the coring process, cores were oriented seaward, in order to distinguish between ebb and flood dune stratifications after core opening.

A medical Computed Assisted Tomography scanner (CAT scanner) was used to obtain very-high resolution images in longitudinal and transversal directions of the cores before opening. Those images were used to identify the bedding, the bedforms and the grain size changes. Cores were opened with respect to the seaward orientation, directly photographed and an accurate sedimentological description was carried out. Grain size analysis was performed on three cores (65a, 65b and 65c) using sieving in order to calibrate CAT scanner images. Correlations between core results and seismic profiles have been achieved using a P-wave velocity of 1800 m s⁻¹ for sand. Compaction of sediment during coring is difficult to quantify, indeed correlation between seismic profiles and cores was made only for first-order reflectors.

3.3. Historical bathymetry

Historical bathymetries of the Plassac Tidal Bar were extracted from the database of the Bordeaux Harbor (Grand Port Maritime de Bordeaux, GPMB). Until 1958, bathymetric data were obtained with lead line and positioned by triangulation. Since 1958, bathymetric data were obtained with echo sounders. For both methods, bathymetric profiles spacing was 200 m and their location was constant from one year to the other. The maximum vertical error margin was estimated to 0.5 m for lead line data and 0.2 m for echo sounders data (Billy et al., 2012). The reference level of sounding reduction is the marine chart contouring at Fort-Médoc (Fig. 1B), which is —2.296 m with reference to the 0 m NGF (Nivellement Général de la France), corresponding to the mean sea level in Marseille. For the 1905 to 2008 period, 29 maps have been scanned, georeferenced, digitized and processed using ArcGis 9.3. From each original map, DEM were generated using nearest-neighbor algorithm and a square elementary cell of 60 m to compute changes at and marsh deposits. Then, these two vertical sequences of facies will be projected on the seismic lines to reveal the overall internal architecture of the Plassac Tidal Bar.

4. Results

The results section includes: (1) the description of the vertical sequences of facies and internal architecture of the Plassac Tidal Bar and its substratum; (2) the detailed internal architecture of the eastern spit, where the progradation phases of a lateral accretion package can be reconstructed through time; (3) the analysis of the hydrology of the Gironde Estuary (1950–2008), which partly controls the internal architecture of the tidal bar.

4.1. Vertical sequences of facies and internal architecture of the Plassac Tidal Bar

Core results and field observations showing the substratum of the tidal bar and the two vertical sedimentary sequences of facies of the Plassac Tidal Bar will be described first. The basal sequence is generated by the progradation of the Plassac Tidal Bar onto the muddy bottom of the estuarine channel in the bay-head delta. The upper sequence is generated by the lateral accretion of the tidal bar (to the west or to the east). Both the basal and the upper sequences are locally capped by mud flat and marsh deposits. Then, these two vertical sequences of facies will be projected on the seismic lines to reveal the overall internal architecture of the Plassac Tidal Bar.

4.1.1. Substratum of the tidal bar

The Plassac Tidal Bar migrated on a substratum that was sampled by the deepest cores (see: cores 65b, 40b, 48a, Fig. 2). This substratum is composed of 2 facies:

- The basal facies (Fs1) is a coarse to medium-grained sand layer (Figs. 3 and 4). This sand layer contains gravels and exhibits amalgamated dune sets with abundant clay drapes and mud pebbles (710 to 780 cm on core 65b). The thickness of this sand layer is unknown, because it is located at the base of the cores and was not entirely cored.
- The uppermost facies (Fs2) is a 30 cm thick, at the minimum, muddy layer (Figs. 3 and 4), which exhibits a rhythmic layering where groups of sandy ripples alternate with groups of lenses in a cyclic pattern, suggesting a neap-spring cycle layering (651 to 710 cm on core 65b, Fig. 4).

4.1.2. Basal vertical sequence of facies, generated by the progradation of the Plassac Tidal Bar onto the bottom of the estuarine channel (based on cores observation)

The basal vertical sequence of facies, generated during the progradation of the Plassac Tidal Bar onto the muddy bottom of the estuarine channel (upper most facies of the substratum), is a few meter thick (127 to 556 cm on core 65a, 118 to 651 cm on core 65b, 264 to 552 cm on core 65c, Figs. 3 and 4). From base
(estuarine subtidal channel floor) to top, it includes two different facies:

- Bottomsets of the basal vertical sequence (Fs3) are deposited in the subtidal zone and are around one meter thick (462 to 556 cm on core 65a, 530 to 651 cm on core 65b, 472 to 552 cm on core 65c, Figs. 3 and 4). They are made of decimeter-thick sandy beds (1 to 3 dm), isolated from one to another by decimeter-thick muddy beds (0.5 to 1 dm). Sandy beds are composed of medium-grained sand ripples and small-size dune sets with abundant clay drapes and mud pebbles. Muddy beds are made of amalgamated millimeter-thick slack water drapes intercalated with millimeters-thick sand ripples and limns.

- The core of basal vertical sequence (Fs4) is deposited in the subtidal and intertidal zones and is 2 to 4 m thick (127 to 462 cm on core 65a, 118 to 530 cm on core 65b, 264 to 472 cm on core 65c, Figs. 3 and 4). It corresponds to a homogenous sandy unit (devoid of the decimeter-thick muddy beds described in the bottomsets), composed of medium-grained amalgamated dune sets with abundant clay drapes and mud pebbles. This facies has already been described in details by Féniès et al. (1999).

This basal vertical sequence shows a thickening upward of the sand beds (Figs. 3 and 4) from the bottomsets of the bar (alternation of decimeters-thick sand beds and decimeter-thick shale beds), deposited in the subtidal zone, to the top of the sand bar (sand body composed of amalgamated sand dune sets), deposited in the intertidal zone.

4.1.3. Upper vertical sequence of facies, generated by the lateral accretion of the Plassac Tidal Bar (based on cores and field observation)

Upper vertical sequence of facies, generated by one phase of lateral accretion is up to 2.64 m thick (see: cores 65a and 65c; Figs. 3 and 4). The lateral accretion process of the Plassac Tidal Bar is related to the incorporation of a seaward migrating mini-flood lobe onto the spit of the tidal bar (Fig. 5, see also the detailed morphological evolution of the Plassac Tidal Bar described by Billy et al., 2012).

From base to top (0–127 cm on core 65a, 0–118 cm on core 65b and 0–264 cm on core 65c, Figs. 3 and 4), this lateral accretion package includes two different facies:

- Basal layer of the lateral accretion package (Fs5) is a decimeter-thick layer, composed of two distinct sub-facies. Sub-facies Fs5a is composed of a mud matrix exhibiting regular alternations of millimeter-thick beds of very fine sand linsens and centimeter-thick beds of fine sand ripples, with plant debris and roots. It was observed in cores (78–127 cm on core 65a and 57–118 cm on core 65b, Figs. 3 and 4) and on the field (where Fs5a crops out at the surface of the tidal bar and gently dips to the west, Fig. 4). Sub-facies Fs5b is a dense lag of clay pebbles, clay drapes and organic matter debris (238–264 cm on core 65c, Figs. 3 and 4).

Both sub-facies Fs5a and b outcrop at the surface of the tidal bar and exhibit gently dipping westward or eastward dipping strata.

- Upper layer of the lateral accretion package (Fs6) is a couple of meter-thick thickening upward succession (0–78 cm on core 65a, 0–57 cm on core 65b, Fig. 3 and 0–228 cm on core 65c, Fig. 4) of medium sand beds (sand beds thickness range: a few centimeters at the base, to a few decimeters at the top). Sand beds are composed of juxtaposed sand dune sets, which are more and more amalgamated upward and which constitute an homogenous sandy unit, very similar to the facies of the core of the basal vertical sequence (Fs5c) described above.

4.1.4. Top of the Plassac Tidal Bar: mud flat and supratidal marsh deposits

In the upper intertidal zone, where sandy bar is capped by a meter-thick mud flat (Fs7, Fig. 4), there is an abrupt decrease of sand bed thickness and grain size. Sand bed thickness is reduced to the ripple size (a few millimeters to a centimeter) and decreases upward to the limns size (a few millimeters). Grain-size decreases from fine-grained sands to very-fine grained sands and silts.

Mud flat deposited in the upper part of the intertidal zone exhibits the classic tidal bedding with the regular alternations fine-grained sand ripples and limns deposited in a muddy matrix (Fs7, Fig. 4). Skolithos ichnofacies is observed here, with millimeter-size branching vertical burrows (Fig. 4).

To the top of the mud flat, a 2 m thick marsh is deposited in the supratidal zone (Fs8, Fig. 4). The muddy matrix of supratidal marsh shows a dense network of reed roots (Fig. 4).

4.1.5. Internal architecture of the Plassac Tidal Bar (based on cores and seismic lines)

Two main reflector categories were distinguished on the basis of their seismic character and lateral extension (Figs. 2 and 3). The first-order reflector is named R1 and is observed on distances longer than 5 km (longer than the Plassac Tidal Bar). It is generated at the interface located between the bottomsets of the tidal bar (Fs3) and the substratum of the bar (Fs1 and 2). The second-order reflectors are observed on distances of hundred of meters; they are named R2 and correspond to the master bedding within the tidal bar and outline the lateral accretion packages.

The first-order reflector (R1) is observed at the base of the tidal bar deposits. It is a high-amplitude, low frequency, discontinuous reflector, most often sub horizontal and displaying a step-like morphology (Figs. 2 and 3). It crops out in the tidal channels located along the spits (Fig. 6). The R1 reflector is correlated with the facies of the substratum of the tidal bar. The high impedance contrast of the R1 reflector is probably related to the sharp grain size changes between the mud layer at the base of tidal bar (Fs2) and the coarse-grained estuarine channel deposits (Fs1), located immediately below this mud layer and the overlying sandy beds of the bottomsets of the bar (Fs3). High impedance contrast of R1 may also relate to compaction of older sediments at the base of the tidal bar. Step-like morphology of R1 reflector identified on seismic profiles and its stratified morphology identified on acoustic imagery where R1 outcrops (Fig. 6) is in accordance with the cohesive nature of the Fs2 facies of the substrate, which is a mud layer eroded by tidal currents.

Second-order reflectors correspond to the master bedding of the tidal bar and outline the sigmoidal shape of the lateral accretion packages. They are labeled R2e (R2e1, R2e2, R2e3) and R2w for the eastern and the western ebb spits, respectively (Figs. 2, 3, 7 and 8).

Within the eastern spit, 3 inclined reflectors are observed from west to east (Figs. 3 and 7). R2e1 is a low frequency, strong amplitude, relatively continuous, gently westward dipping (1.5° dip-angle) reflector. R2e2 is a low frequency, strong amplitude, eastward dipping reflector (3° dip-angle) with a very good continuity. Downlap of R2e1 and R2e2 onto basal R1 reflector is not observed because those reflectors end at about 4 ms twt below the tidal bar top. R2e3 is a high frequency, low amplitude, discontinuous eastward dipping reflector (4° dip angle).

R2e1 was sampled by cores 65a (78 to 127 cm) and 65 b (57 to 118 cm) and is correlated with the basal layer of a lateral accretion package (sub-facies Fs5a, a mud matrix with regular alternations of millimeter-thick beds of very fine sand linsens and centimeter-thick beds of fine sand ripples, exhibiting plant debris and roots). R2e2 was sampled by 65c core (238 to 264 cm) and is also correlated with the basal layer of a lateral accretion package (sub-facies Fs5b, a dense lag...
of clay pebbles, clay drapes and organic matter debris). R2e3 was not cored.

The dipping reflectors (R2e1, R2e2 and R2e3) can be correlated, from one seismic profile to another along the axis of the eastern spit (Fig. 7), which allows the reconstruction large-scale dipping planes gently inclined to the East, in a direction normal to the tidal currents.

Above these reflectors, meters-thick sub transparent seismic units lie and show a sigmoidal shape (dipping to the East or to the West, seismic line 65, Figs. 3 and 7), which are correlated with the thicker upper sandy layers of the lateral accretion packages (Fs6).

Consequently, master bedding of the Plassac Tidal Bar is composed of the horizontal juxtaposition of sigmoidal-shaped lateral accretion...
Fig. 4. Detailed description of the three cores (65a, b and c) showing sedimentary facies names, grain size results and bedforms. For each sedimentary facies sampled by the cores, a scanner image and photography of the core are shown (without horizontal exaggeration). Outcrop photographs of sedimentary facies Fs5, Fs7 (upper muddy intertidal flat) and Fs8 (supratidal marsh) show additional observations. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
packages, made of meters-thick sandy units (upper layers of the lateral accretion packages, Fs6), isolated from one another by decimeter-thick muddy units (lower layers of the lateral accretion packages, Fs5).

Within the western spit (seismic profiles 44, 48 and 56, Figs. 2 and 8), another category of reflector (R2w) is observed. They are strong amplitude and good continuity reflectors displaying erosive concave-up channel shapes. In some areas (seismic profile 44, Fig. 8), the concave-up channel shape is filled by high frequency, small to middle amplitude seismic reflectors, which are dipping to the east (5–10° dip-angle). R2w reflector was sampled by 44b core and corresponds to a gravely layer (few millimeters to 5 cm grain size interval, Fig. 8). Those erosive channel-shaped reflectors separate two sandy sequences (see seismic line 44 and cores 44a and b, Fig. 8). In cores 44a (Fig. 2) and 44b (Fig. 8), both the lower and upper sandy sequences exhibit the classic tidal bar facies/thickening upward sand bed trend, observed in the lateral accretion packages described in cores 65a, 65c (Fs 6, Figs. 3 and 4). Consequently, the lower and upper sandy sequences can be interpreted as two lateral accretion packages, showing evidence of lateral accretion to the west (upper unit) and to the east (lower unit). These two lateral accretion packages are separated by an erosive channel-shaped surface (reflector R2w correlated with a gravely layer), which partially truncates the lower one. The western spit had a complex story, made of phases of lateral accretion and erosion.

4.2. Internal architecture of the eastern spit: a detailed analysis of the emplacement of the sigmoidal-shaped lateral accretion packages through time

Superimposition of bathymetric profiles on seismic data shows the internal geometry of sediments emplaced before and after those time-lines (Figs. 2, 3 and 7). From those time-lines, it appears that a part of the western spit was emplaced before 1905 and that the entire eastern spit was emplaced from 1950 onwards (Fig. 7). Given that fluvial discharge data in the Gironde Estuary are available since 1905, we focus on the eastern spit because it is composed of seismic units that can be easily compared with bathymetric profiles and fluvial discharge data.

Bathymetric profiles show a rapid lateral accretion of the tidal bar through time, with a rate of about 200 m to the east in 16 years (between 1993 and 2009, year of the seismic profile record). Lateral accretion process is not continuous through time: it is punctuated by very fast phases of lateral accretion, where upper sandy layer of the lateral accretion packages (Fs6) are deposited over the muddy basal layers (Fs5) within a few years. Bathymetric changes between 1991 and 1993 (Fig. 7), are a clear illustration of the geometry of the fast-deposited lateral accretion package: within 2 years an elongated layer of sediments (until 3 m thick, 200–400 m wide and at least 5 km long) was deposited along the flank of the eastern spit of the tidal bar.

Superimposition of bathymetric profiles on seismic profiles (Fig. 7), allows the timing of deposition of the basal layers of the lateral accretion package to be estimated (R2e2 and R2e1 reflectors are correlated with the basal muddy layers of the sigmoidal-shaped lateral accretion package Fs5):

- Eastward dipping reflector R2e2 was deposited between 1991 and 1993 (Fig. 7),
- westward dipping reflector R2e1 and the eastward dipping reflector R2e3 were deposited around 1993 (Fig. 7).

Distance between R2e2 and R2e3 reflectors (around 60 m) is the width of the lateral accretion package deposited in less than 2 years (1991–1993).

Therefore timing of the internal architecture of the eastern spit of the Plassac Tidal Bar can be precisely reconstructed through time. Deposition of this spit started around 1950. During around 54 years (1950–2009), the 400 m wide spit (east–west dimension)
was constructed by a lateral accretion process: juxtaposition of sigmoidal-shaped lateral accretion packages constituted of few meter-thick sandy units (the upper layers of the lateral accretion packages Fs6, around 60 m wide) are isolated from one another by decimeter-thick muddy units (the basal layers of the lateral accretion packages Fs5).

Mechanism of the lateral accretion process described above was reconstructed by Billy et al. (2012) on the Plassac Tidal Bar, with the comparison of bathymetric maps of the bar (1905–2010): the lateral accretion packages are generated by downstream migration of fluviolysourced mini-fluvio lobes, which are incorporated into the spits of the tidal bar (Fig. 5, Billy et al., 2012).

4.3. Analysis of the hydrology of the Gironde Estuary (1950–2010), which partly control the internal architecture of the tidal bar: number of floods and low waters of the Gironde Estuary (1950–2010)

Given a fluvial influence on the morphological evolution of the Plassac Tidal Bar was evidenced (Billy et al., 2012) and given the timing for seismic unit deposition can be inferred from bathymetric data between 1950 and 2010, this section is focused on fluvial discharge evolution during this period of time (numbers of days of rivers flood and low water stages per month and per year, Fig. 9). A more complete analysis of fluvial discharge evolution in the Gironde Estuary is available in Billy et al. (2012). Number of days of river flood exceeds 15 during 20 years between 1950 and 2010. Most of floods occur from December to May. Nevertheless the year 1992 is atypical, because during this year, floods occur from July to December. Number of days of low water stage exceeds 28 during 20 years between 1950 and 2010. Most of the low water stages occur from July to September. Nevertheless the period from 1988 to 1991 is exceptionally dry, and as during those three years, number of days of low waters was higher than 100 per year. Moreover during the years 1989 and 1990, many low water stages occurred in January and February as those months are often characterized by high water discharges.

Computation of the numbers of days of flood and low water stages of the Gironde Estuary per month and per year from 1950 to 2008, shows that the period from 1989 to 1992 is unique as it includes the three successive driest years (1989 to 1991) followed by the only wet year where the floods occur in summer and early autumn (1992).

5. Discussion

5.1. Correlation between acoustic impedance and grain-size changes

Correlation between first- and second-order reflectors (R1 and R2) with core samples shows that the highest contrasts of acoustic impedance reveal major grain size changes. Indeed, major acoustic impedance changes, R1, R2e1 and R2e2, are correlated with decimeter-thick mud-dominated layers interstratified between sand-dominated units. In addition, reflector R2w corresponds to a gravelly layer interstratified between sandy units. Those correlations between seismic reflectors and core samples support the works of Alexander et al. (1986) and Billeaud et al. (2005) that showed grain size changes are responsible for major reflectors of similar very high resolution seismic profiles.

5.2. Sequence stratigraphy interpretation and internal architecture of the Plassac Tidal Bar

Given the recent bathymetric evolutions of the Plassac Tidal Bar (Billy et al., 2012), vertical sequences of facies of this tidal bar belong to the Highstand Systems Tract (HST), as described by Allen and Posamentier (1993) and Féniès and Tastet (1998). The muddy layer (Fs2), at the base of the tidal bar, records an estuarine environment that occurred before the deposition of regressive tidal sand bars of the bay-head delta. Thus this muddy layer is interpreted to be the Maximum Flooding Surface (MFS), which was deposited around 4000 years B.P. during the period of the maximum flooding of the estuary, before estuarine accommodation space started to be filled by fluvial sediment supply (Allen and Posamentier, 1993). Below the muddy layer (Fs2), the coarse-grained sand layer of the substratum of the tidal bar is interpreted to be the facies of estuarine channels of the Transgressive Systems Track (TST), as described by Allen and Posamentier (1993).

Overall, individual sand units of the Plassac Tidal Bar (basal and upper sequences of the eastern spit and western spit) consist of thickening-upward sand beds lying most of the time on mud-rich strata. This stratal pattern is very similar to the vertical sequence of facies generated by the seaward progradation of the Trompeloup Tidal Bar, located 20 km seaward in the Gironde Bay-head delta (Féniès and Tastet, 1998). This stratal pattern is explained because fluid muds are well developed in the channel bottom of the Gironde Estuary. An upward grain size decrease of the sand is observed at the top of the eastern spit (Fs6 to Fs7 and Fs8). This upward grain size decrease is explained by upward decrease of both water depth and current speed, from the thalweg toward the bar crest, as already proposed by Dalrymple et al. (2003).

Thus the Plassac Tidal Bar, like the Trompeloup Tidal Bar, is a composite sedimentary body made up of meters-thick sandy units isolated from one another by mud layers. These mud layers, which are evidenced on seismic as strong contrasts of acoustic impedance, represent the master bedding of the tidal bars, in between those master bedding planes, the meter-thick sandy units are clinoforms which exhibit a sigmoidal morphology.

A main difference between the two tidal bars is related to the thickness of mud layers, which are decimeter-thick in the Plassac Tidal Bar and meter-thick in the Trompeloup Tidal Bar (Féniès and Tastet, 1998). These different thicknesses may be explained by the downstream location of the Trompeloup Tidal Bar closer to the turbidity maximum of the Gironde Estuary (the Trompeloup Tidal Bar is located at the front of the bayhead delta).

Another major difference between the two tidal bars is related to the master bedding orientation, which dips in a direction normal to the tidal currents axis in the Plassac Tidal Bar (towards the east and towards the west) and which dips in the axis of the tidal currents in the case of the Trompeloup Tidal Bar (towards the North, i.e.: seaward, Féniès and Tastet, 1998).

Detailed studies of the morphological evolution of the two tidal bars through the last century evidence that:

- The Plassac Tidal Bar is laterally accreting (Figs. 3 and 4 in Billy et al., 2012), this lateral accretion process generates a master bedding which dips in a direction normal to the axis of the tidal currents;
- The Trompeloup Tidal Bar is longitudinally accreting (Fig. 3 in Féniès and Tastet, 1998), this longitudinal seaward accretion process generates a master bedding which dips in a direction parallel to the axis of the tidal currents.

5.3. Processes controlling the internal architecture of the tidal bars

Internal accretion surface orientations are commonly used as a criterion to differentiate tidal bars from tidal dunes (Olaru et al., 2012). Tidal dunes migrate forward in the same direction as the current and they have internal accretion surfaces that are oriented essentially normal to the currents. Oppositely, tidal bars have a strong and characteristic component of lateral migration and they have internal accretion surfaces that are frequently oriented parallel to the currents (Olaru et al., 2012). Unlike this basic distinction, the internal architecture of lobate (Plassac) and elongated (Trompeloup) tidal bars of the Gironde Estuary demonstrates that their internal accretion surfaces are both parallel and normal to the tidal currents axis. In the case of the lobate Plassac Tidal Bar, dune stratifications dip in a direction that is the axis of the tidal currents: to the North (ebb current) or to the South (flood
current); see: Féniès et al. (1999); Billy et al. (2012) and are oriented perpendicularly to the dip of the master bedding planes (dip to the West or to the East). Meter-thick sandy sigmoidal clinoforms located in between the muddy master bedding planes dip in the same direction than the master bedding. Similar orientation of the clinoforms was already observed in other examples of tidal bars or sandbanks emplaced in estuary mouth (Chaumillon et al., 2008), or in deltas (Dalrymple et al., 2003). Such clinoform orientation was proposed to record the lateral migration of sandbanks and tidal bars, which has been widely documented (Houbolt, 1968; Harris, 1988; Berné et al., 1994; Dalrymple and Zaitlin, 1994; Dalrymple and Rhodes, 1995; Dalrymple et al., 2003; Chaumillon et al., 2008; Olariu et al., 2012). In the case of the Plassac Tidal Bar, the process of lateral accretion is generated by the merging of mini-flood lobes with the ebb spits following periods of high water discharges (Fig. 5 and Billy et al., 2012). Thus lateral migration and accretion processes may be generalized to all lobate-shape tidal bars and sandbanks. Bertin and Chaumillon (2005) carried out a detailed hydrodynamic study on a similar lobate-shape sandbank and showed that the sandbank crest was slightly oblique to predominant current and induce a gradient of the free surface elevation, inducing a crest ward deviation of the tidal current and related residual transport. Obliquity of tidal bar crests in estuaries with respect to predominant

![Fig. 6. A) High resolution digital elevation model (1.3 m grid) of the western spit in bathymetry in 2010 showing the stratified morphology of the erosional surface R1 outcropping on the eastern flank of the western spit. Reference level of sounding reduction is the marine chart sounding datum at Fort-Médoc. B) Very high resolution seismic profile GiRaFS_48 showing the step-like morphology of the erosional surface revealed by the R1 reflector and its outcropping on the eastern flank of the western spit. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)](image-url)
currents may be an important issue to explore for the lateral migration and accretion processes.

In the case of the elongated Trompeloup Tidal Bar (Féniès and Tastet, 1998), dune stratifications, observed on cores, also dip in a direction that is the axis of the tidal currents: to the North (ebb current) or to the South (flood current). But, oppositely to the Plassac lobate Tidal Bar, the master bedding plane (meter-thick mud layer) dips towards the North in the same direction that the ebb tidal current. The meter-thick sandy sigmoidal clinoform also dips towards the North, in the same direction (see Fig. 13 in Féniès and Tastet, 1998). This dip is generated by the fast process of longitudinal seaward accretion of the tidal bar towards the North, in a direction that is slightly parallel to the axis of the tidal currents (the crest of the Trompeloup Tidal Bar prograded towards the North, by 5 km in 10 years; see Fig. 4 in Féniès and Tastet, 1998).

Although lateral accretion is widely documented in tidal bars (Houbolt, 1968; Harris, 1988; Berné et al., 1994; Dalrymple and Zaitlin, 1994; Dalrymple and Rhodes, 1995; Dalrymple et al., 2003; Chaumillon et al., 2008; Olariu et al., 2012; this study), it is not the unique accretion process of sandbars as longitudinal accretion has been documented in the Gironde Estuary (Féniès and Tastet, 1998) and at the front of tide-dominated deltas (Kuehl et al., 1997; Willis et al., 1999; Bhattacharya and Willis, 2001; Lambiase et al., 2003). The work realized by Legler et al. (2013) on Eocene Tidal Bars deposited in a tide dominated delta in Egypt also supports the fact that longitudinal accretion is a key-process which controls the tidal bar architecture. Therefore internal accretion surfaces parallel to the flow cannot be used as a unique criterion to characterize tidal bars. Tidal bars, like tidal dunes, may have their internal accretion surfaces normal to the flow. This is of considerable importance for studies in ancient tidal deposits, paleoenvironment reconstructions and for heterogeneities characterization and quantifying reservoir properties in these types of sandstones (Jackson et al., 2005; Ringrose et al., 2005).

5.4. Fluvial discharge control on the internal architecture of the Plassac Tidal Bar

Superimposition of historical bathymetric measurements on seismic data (Fig. 7), shows that, within the eastern spit, the main reflector correlated with the mud pebble layer (R2e2 correlated with Fs5b), was emplaced between 1991 and 1993. The reflector R2e1 is not considered in this section as it is emplaced in the western part of the ebb spit where accurate dating based on bathymetric profiles cannot be achieved as there was an alternation of erosion and accretion phases. Oppositely, sediment accretion occurred continuously in response to the lateral eastward migration of the eastern spit in
the eastern part of the ebb spit (Fig. 7 and Billy et al., 2012). Between 1991
and 1993, the spit eastward lateral accretion was very important, as
revealed by the largest width documented for this sandbar between
1905 and 2008 (Billy et al., 2012) and by the about 3-m-thick sediment
gain and the 10^6 m^3 eastern spit volume gain (Fig. 7). It has been
already shown that lateral accretion of the ebb spits is caused by the
incorporation of downstream migrating mini-flood lobes on the outer
branches of the spits and occurs during the periods of higher fluvial
 discharge (Billy et al., 2012). Moreover the fluvial discharge variations
have also a control on mud deposition in the Gironde Estuary, as the
turbidity maximum migrates upstream during low waters and down-
stream during floods (Doxaran et al., 2009) and thicker clay drapes
are deposited on the Plassac Tidal Bar every summer in response to
upstream migration of the turbidity maximum during low waters.

Given fluvial discharge variations have a control on the Plassac Tidal
Bar morphological evolutions and migration of the areas of mud
deposition, Gironde discharges were compared to the internal architec-
ture of this part of the eastern spit. The period between 1991 and 1993
was unique at the scale of the last decades (1950–2010) in terms of
fluvial discharge. Indeed, this period included the last year of the
three successive driest years (1989 to 1991) and the year 1992 during
which the floods occurred exceptionally in summer and autumn.

Given the timing of deposition of the mud layer (Fs5b) and overlying
sand units (Fs6) coincides with years characterized by exceptional
low waters followed by summer and autumn floods, a three-step de-
positional model is proposed to explain the deposition of the upper
sequence of the eastern spit. From 1989 to 1991, the large numbers
of low waters lead to upstream migration of the turbidity maximum
in the vicinity of the bay-head delta for a long period of time and to
deposition and consolidation of an abnormally thick mud layer in
this part of the Gironde Estuary and on the Plassac Tidal Bar. In
1992, the high number of floods that occurred in summer and au-
tumn leads to the erosion of this mud layer (R2w), as attested by
the presence of angular mud pebbles (F5b) and to the rapid burying
by massive sand transport, as attested by the thick sand unit depos-

6. Conclusion

The Plassac Tidal Bar is the unique lobate-shape tidal bar of the bay-
head delta of the Gironde Estuary, all the other tidal bars are of an
elongated shape (e.g.: Trompeloup Tidal Bar; Fenié and Tastet, 1998).
Its stratigraphy was investigated using a dense grid of 65 very high
seismic profiles ground truthed by 6 to 9 m long cores.

The base of the tidal bar consists of a tidal erosional surface. This sur-
face has eroded muddy tidalites, corresponding to the maximum
flooded surface of the Gironde Estuary. Below the main flooding sur-
face, coarse sands, deposited in estuarine channels, belong to the Trans-
gressive Systems Track.

The bar is composed of two vertical sequences of facies, composed of
thickening-up cross stratified fine- to medium-grained sand beds
at the top and decimeter-thick mud-layers at the base. Those mud
layers constitute the master bedding of the bar and are dipping in
a channelward direction. The dip of the master bedding planes is
perpendicular to the tidal flows and to the sediment transport direc-
tions. The master bedding planes underline phases of lateral accre-
tion generated by the merging of sandy mini-flood lobes with the
ebb spits following periods of high water discharges (mini-flood
lobes are fluvially-sourced).

One phase of lateral accretion of the Plassac Tidal Bar, which gener-
ates a thickening upward sequence of sand beds, can be correlated with
variations in fluvial discharge of the last decades and suggests a climate
control on the internal architecture of this sandbar. Periods character-
ized by a large number of low waters lead to the upstream migration
of the turbidity maximum and to the deposition and the consolidation
of the muddy master bedding planes. During subsequent flood periods,
muddy master bedding planes are partially eroded and then buried by
the sandy mini-flood lobes which merge with ebb spits.

Lateral accretion appears to be the dominant process recorded in
the master bedding of lobate-shape tidal bars (Plassac Tidal Bar and
Longe de Boyard Tidal Bar; Chaumillon et al., 2008) and of other exam-
ple s of sandbar or sandbank (Houbolt, 1968; Harris, 1988; Berné et al.,
1994; Dalrymple and Zaitlin, 1994; Dalrymple and Rhodes, 1995).

Fig. 8. A) Very high resolution seismic profile GiRaFS_44 showing the internal architecture of the western spit of the Plassac Tidal Bar (see location map on Fig. 2). Simplified core results sampled along this profile are superimposed. Mud is represented in black and mixed sand-and-mud and sand-dominated strata are respectively represented in gray and white. B) Scanner images obtained from the core 44b showing a gravelly layer correlated with R2w reflector.
Nevertheless, lateral accretion is not the unique accretion process of sandbars as longitudinal accretion was documented in elongated tidal bars placed in estuaries (Fenié and Tastet, 1998) or at the front of tide-dominated deltas (Kuehl et al., 1997; Willis et al., 1999; Bhattacharya and Willis, 2001; Lambiase et al., 2003; Legler et al., 2013). Such variations in accretion processes are of considerable importance for reservoir property characterization and for studies in ancient tidal sandstones and paleo-geography reconstructions.

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