Caen, July 28 - August 5, 2012

8[th]

International Conference on Tidal Environments

Tidalites 2012

Field-trip book

ASSOCIATION DES SÉDIMENTOLOGISTES FRANÇAIS

Booklet series of the field trips
organized in the frame of

The 8th International Conference on Tidal Environments

Tidalites 2012

31 july - August 2
Caen, France
(Chair: Bernadette Tessier, CNRS, UMR M2C)

Pre-conference field trips (July 28-30, 2012)

The Incised-Valleys of SW France: Marennes-Oléron Bay, Gironde Estuary and Arcachon Lagoon (3 days)
Leaders: Eric Chaumillon, Hugues Féniès

The Miocene Tidal Shelly Sands of Anjou-Touraine, France (2 days)
Leaders: Jean-Pierre André, Fabrice Redois, Cyril Gagnaison, Jean-Yves Reynaud

The Somme bay, NW France: a wave-dominated macrotidal estuary (2 days)
Leaders: Alain Trentesaux, Jose Margotta, Sophie Le Bot

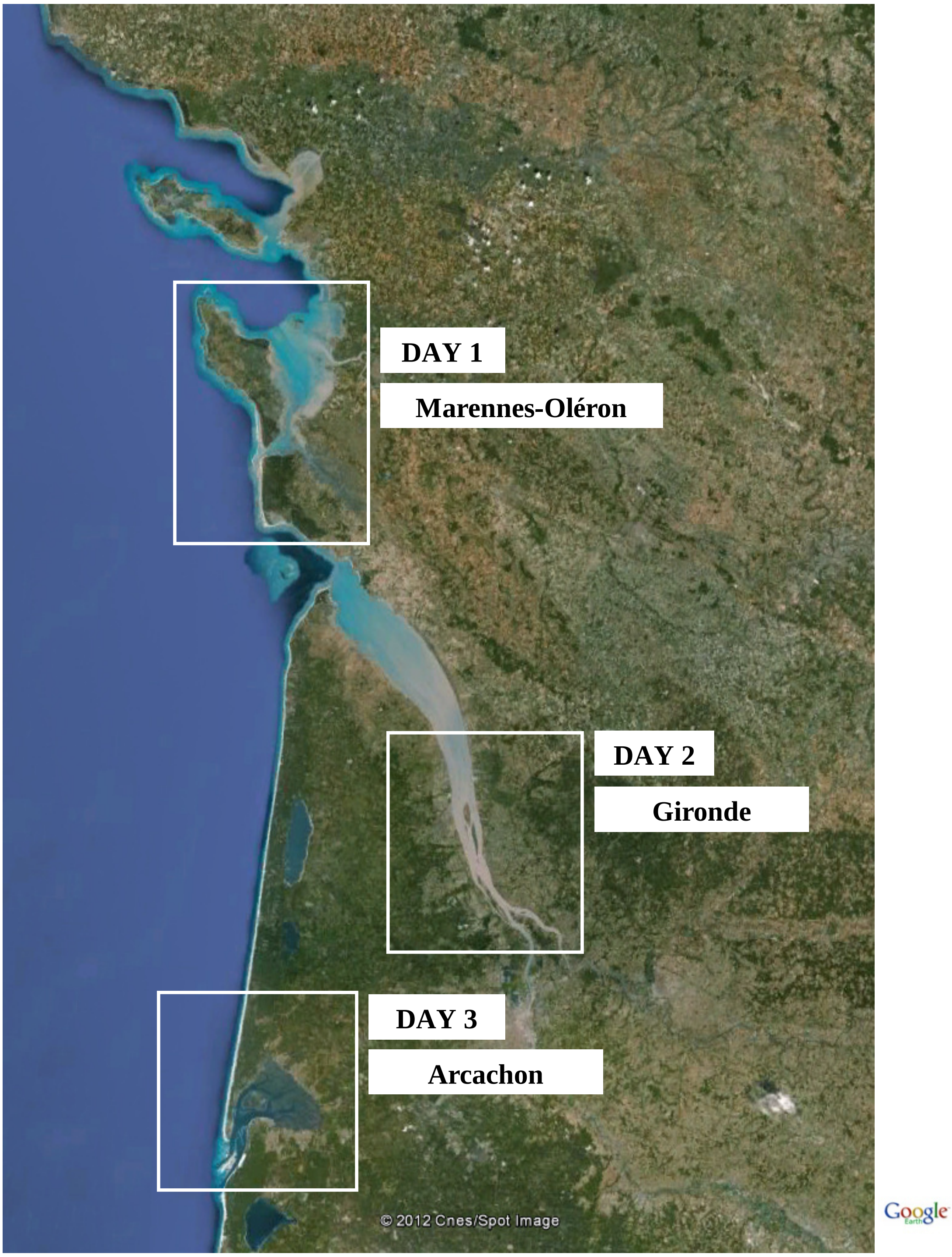
Post-conference field trip (August 3-5, 2012)

The Mt St Michel bay, NW France: Facies, sequences and evolution of a macrotidal embayment and estuarine environment (3 days)
Leaders: Bernadette Tessier, Chantal Bonnot-Courtois, Isabelle Billeaud, Pierre Weill, Bruno Caline, Lucille Furgerot

"The Incised-Valleys of SW France: Marennes-Oléron Bay, Gironde Estuary and Arcachon Lagoon"

(July 27-30, 2012)

Leaders: E. Chaumillon & H. Féniès ; eric.chaumillon@univ-lr.fr ; hugues.fenies@orange.fr



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MARENNES-OLERON BAY,GIRONDE AND ARCACHON FIELD TRIP AGENDA

On July 28 th. 2012, we will visit the Marennes-Oléron Bay (MOB), the first oyster farming area of Europe. This tidal bay is connected to the Atlantic Ocean by two entrances, a wide mouth with an elongated tidal sandbank and a narrow mouth with a tidal inlet. The Marennes-Oléron Bay belongs to the Pertuis Charentais incised-valleys system, showing a great variability in incised valley-fills (Chaumillon et al., 2008) with respect to the global model of Zaitlin (1994). Three stops in the MOB will illustrate: (1) the tide dominated northern mouth and the sediment dynamic of a mixed flat; (2) the mixed tide-and-wave dominated southern mouth and the long-term morphodynamic of a tidal inlet; (3) an example of fast regression at the location of the ancient Brouage Harbour. The last stop of this first day will be dedicated to **the Gironde Estuary Mouth** and the Bonne Anse Bay.

On July 29 th. 2012, we will visit the Gironde estuary: a worldwide reference model for tide-dominated incised valleys (Allen and Posamentier, 1993; 1994). The field trip will focus on the sandy tidal bars deposited in the Bay-Head delta. Detailed facies analysis of these deposits will be presented based on trenches and cores, as well as the internal architecture of these geobodies.
The sequence stratigraphic model of the Gironde incised valley (Lericolais, Berné and Féniès, 2001) will be presented based on high resolution seismic and the impact of tidal ravinement processes on the geometry of wave- and tide-dominated incised valleys will be discussed (Féniès, Lericolais and Posamentier, 2010).

On July 30 th. 2012, we will visit the Arcachon lagoon: a wave-dominated estuary. The field trip will focus on the sandy tidal channel deposits. Thanks to numerous trenches and cores the facies analysis of the tidal channel-fill will be presented, as well as the geometry of these deposits.
The sequence stratigraphic model of the Arcachon incised valley (Féniès and Lericolais, 2005) will be presented based on high resolution seismic. Comparison with the Gironde stratigraphic model will allow to identify specific processes controlling reservoir geometry in wave- and tide-dominated incised valleys (Féniès, Lericolais and Posamentier, 2010).

JULY 28 Th. 2012: FIELD SESSION IN THE MARENNES-OLERON BAY and GIRONDE ESTUARY MOUTH

1- STRATIGRAPHIC RECORDS AND VARIABILITY OF INCISED VALLEYS AND ESTUARIES ALONG FRENCH COASTS

This 3-days field trip gives the opportunity to visit and compare 3 estuarine environments belonging to modern incised-valleys of French Coasts. The coasts and inner shelves of France can be qualified in general as sediment-starved systems. This characteristic constitutes the common feature of all the study sites. This distinguishes these systems from many other published case studies, such as those along the South-East and Gulf coasts of the USA. A recent synthesis about the incised valleys along the French coasts (Chaumillon et al., 2010) have allowed comparing incised valleys within the same setting of tectonically stable and sediment starved margins, but showing contrasted conditions of hydrodynamics, sediment supply and bedrock control. At a stratigraphic level, sea-level variation is the main parameter controlling incised valley formation and sediment fill. The first-order controlling factor explaining the observed variations in valley fills is hydrodynamics. Three valley-fill categories are highlighted: tide-dominated, mixed tide-and-wave and wave-dominated, that match the classification based on hydrodynamics and morphology of present-day estuaries or lagoons (fig. 1). The second-order controlling factor explaining the observed variations in valley fills is the antecedent morphology of the bedrock, which in turn controls hydrodynamics and sediment supply. Finally, a promising result is the demonstration of the potential of incised valley fills to record high frequency environmental changes related to climate events and human activities.

The Marennes-Oléron Bay, Gironde and Arcachon Lagoon examples show an interesting variability in terms of morphology and stratigraphy, related to the wave-to-tide energy ratio and to the antecedent morphology of the bedrock. The Marennes-Oléron Bay and the Gironde Estuary are Mixed tide- and wave- meso and macrotidal estuaries (fig. 1) characterized by important bedrock control on the dynamic of their tidal channels and inlets (Bertin et al., 2004, Féniès et al., 2010). The Marennes-Oléron Bay is connected to small rivers whereas the Gironde Estuary is connected to large rivers are receive a large amount of fluvial-derived sediments. Sr and Nd isotope analysis of surface sediments (Parra et al., 1998) suggest that the greatest part (40 to 90%) of fine-grained sediment particles delivered to the Marennes-Oléron Bay is not provided by the small rivers flowing in the bay, but by the Gironde estuary, the mouth of which is located 60 km southward of the Marennes-Oléron Bay. The Arcachon Lagoon is a more wave-dominated estuary. This mesotidal semi-enclosed lagoon (fig. 1) is connected to a small river. Its sediment dynamic and stratigraphy is not controlled by the antecedent morphology of the bedrock.

Case study	Marennes-Oléron Bay	Gironde estuary	Arcachon Lagoon
Type of coast	Rocky	Rocky	Depositional
Tidal Regime	Macrotidal	Macrotidal	Mesotidal
Mean Spring tidal range (m)	5.10	4.35	3.95
Maximum tidal range (m) / Harbour	6.86 La Rochelle	6.09 Le Verdon	4.86 Arcachon
Wave Climate / Boy Name years	Yeu 2000-2005 *(Oléron 2001-2003)*	-	Cap Ferret 2001-2009
%Hs≤0,5 m	3 *(1)*	-	-
%Hs≤1 m	**18 *(7)***	**-**	**30**
Size of the River (mean water discharge)	Small	Large	Small
Present-day morphology of the estuary	Mixed Open (northern entrance) Semi-enclosed (southern entrance)	Mixed Open	Wave-dom Semi-enclosed
Main source of sediment supply	*Marine*	*Marine & Continental*	*Marine*

Table 1 – Depositional settings for the Marennes-Oléron Bay, Gironde Estuary and Arcachon Lagoon. The Valley segments designation follows Zaitlin et al. (1994). The tidal regime was defined after the classification of Davies [1964]. Maximum tidal range is given by the French hydrographic office (SHOM : http://www.shom.fr/fr_page/fr_act_oceano/RAM/RAM_P1.htm). Hs statistics are obtained from buoys deployed along the coasts of France (http://candhis.cetmef.developpement-durable.gouv.fr). For the Marennes-oléron Bay, two buoy records are indicated, because the nearest buoy (Oléron) has recorded wave climate only between 2001 and 2003. For the Gironde Estuary no buoy measurements are available, so the wave climate offshore the Gironde Estuary is estimated from the Yeu, Oléron and Cap Ferret buoy (see locations in Fig. 2). We feel the increasing wave height from Cap Ferret to Yeu and Oléron depends on buoy locations as our observations on the beaches generally show a contrasting increasing wave height southward. The present-day morphology of the estuaries was defined after the classification of Boyd et al. (1992).

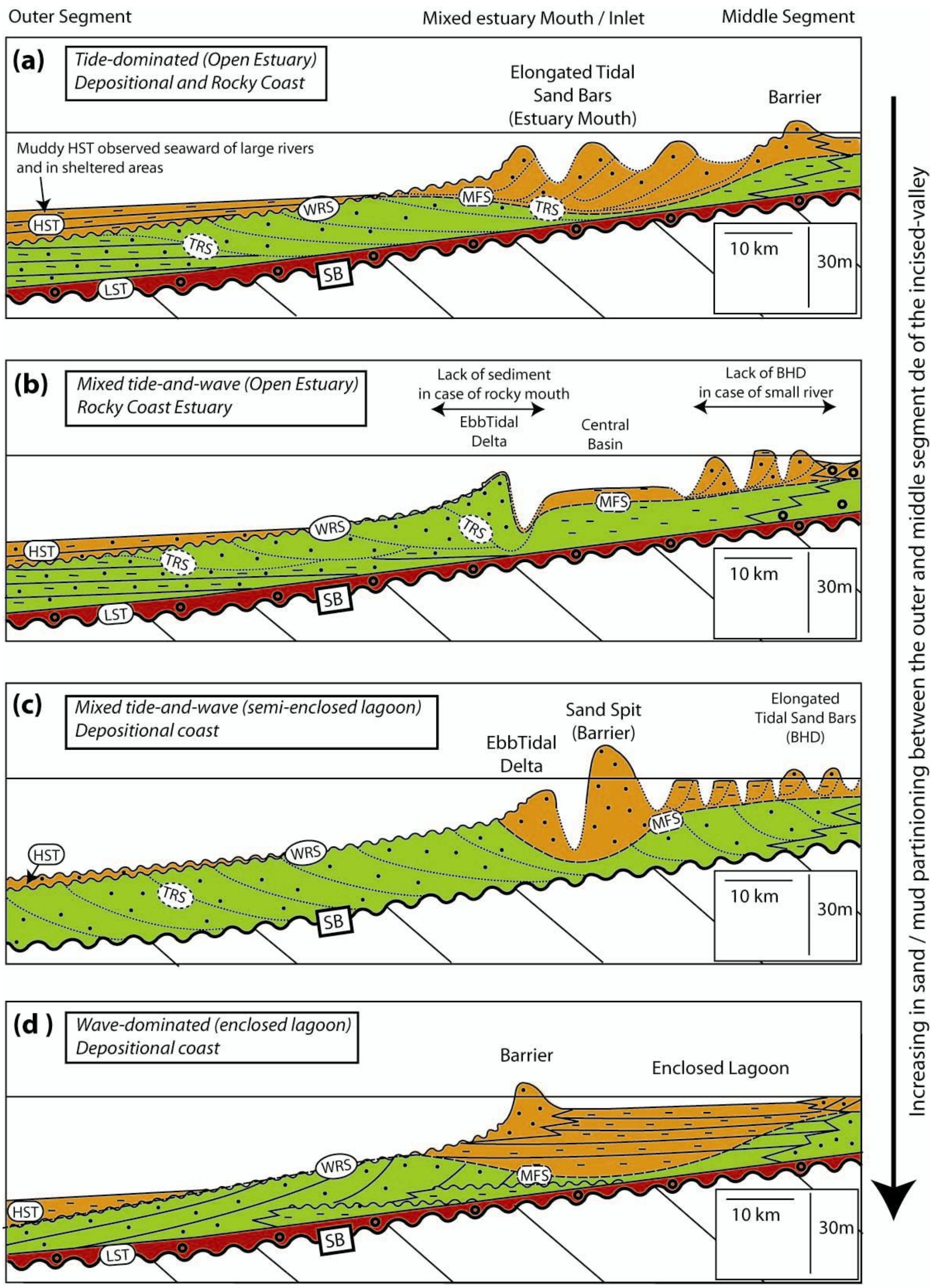


FIG. 1 – Schematic dip-line cross sections showing the variability of French valley fills related to their energetic and morphologic classification and to the presence / absence of rocky coasts bounding the present-day estuary (Chaumillon et al., 2010, BSGF n°181). (a) Tide dominated category associated with open estuaries and depositional or rocky coasts; (b) Mixed tide-and-wave category associated with open and rocky coast estuaries (Marennes-Oléron Bay and Gironde Estuary); (c) Mixed tide-and-wave category associated with semi-enclosed lagoon and depositional coast (Arcachon Lagoon); (d) wave-dominated category associated with enclosed lagoons and depositional coasts.

2- CHARACTERISTICS OF MARENNES-OLERON BAY

2.1- Morphology

The Marennes-Oléron Bay (MOB) is located in the Bay of Biscay about 20 km to the north of the Gironde estuary (fig. 2). The MOB is a 190 km2 wide semi-enclosed environment connected to the Atlantic Ocean through the Pertuis d'Antioche to the North, and the Maumusson Inlet to the South (fig 2). It is bounded seaward by Oléron Island and mainland whose NW-SE elongated rocky outcrops are fringed by extensive marshes (figs. 3 and 4). Intertidal flats represent about 60% of the total surface area of the bay and are cut by sandy tidal channels (fig. 3b). The Marennes-Oléron Bay can be subdivided into four geomorphologic domains : (1) the Longe de Boyard, a 8 km long elongated sandbank located between two tidal channels in the northern entrance (Chaumillon et al., 2002, 2008a; Bertin and Chaumillon, 2005a), (2) the northern MOB (NMOB, fig. 3b), the wider part of the bay (Bertin et al., 2005; Bertin and Chaumillon, 2005a) bounded to the south by the "Coureau d'Oléron Strait" (COS, fig. 3b), (3) the southern MOB (SMOB, fig. 3b) consisting of mixed sand-and-mud flats cut by tidal channels, and (4) the southern "Maumusson" tidal inlet (Bertin et al., 2004, 2005).

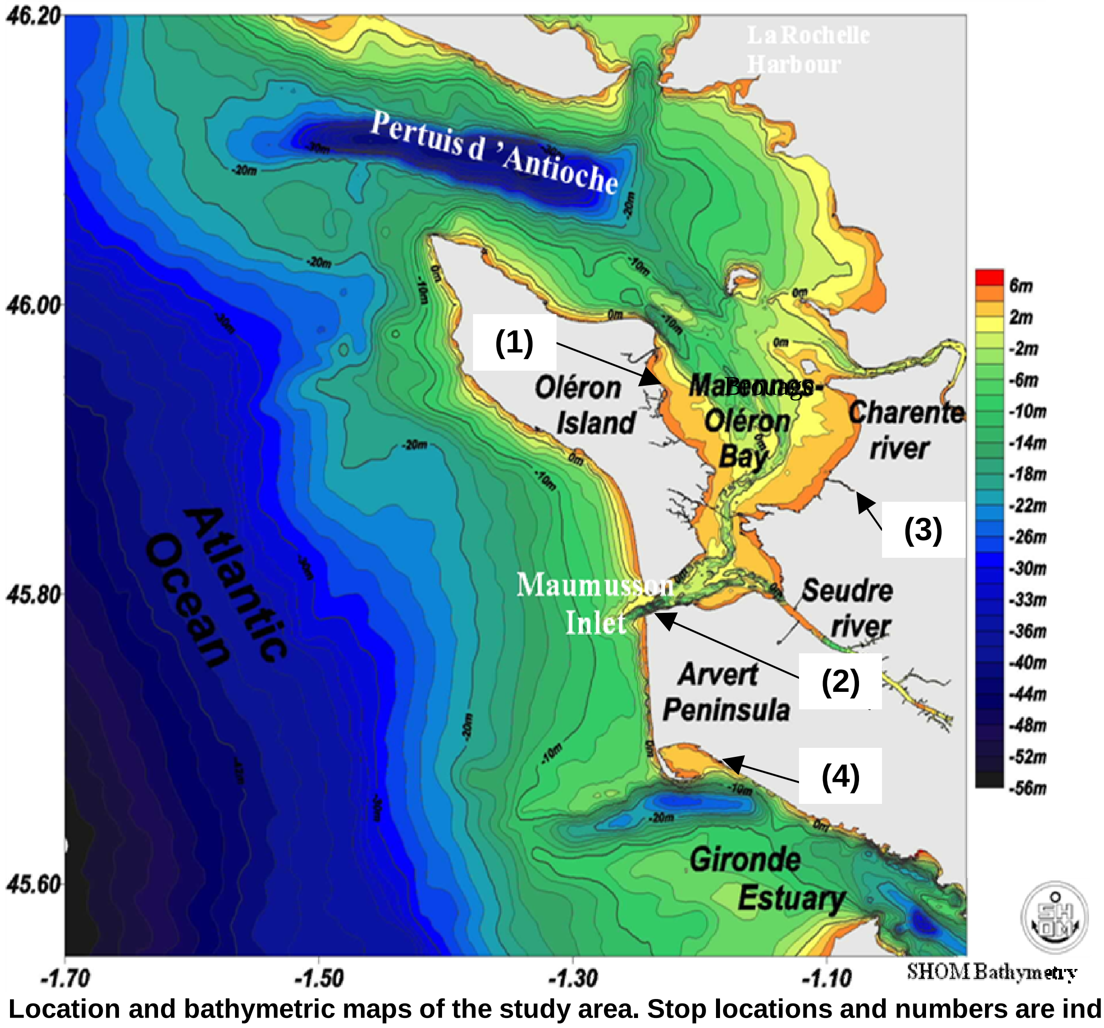


FIG. 2 – Location and bathymetric maps of the study area. Stop locations and numbers are indicated.

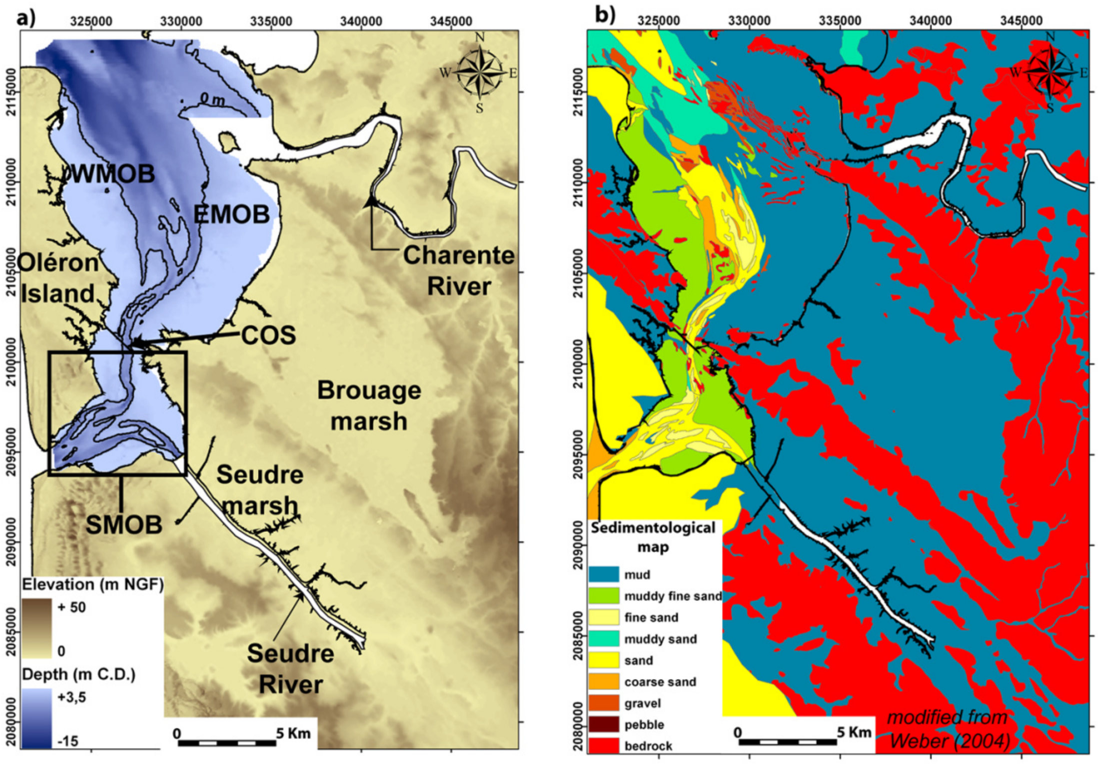


FIG. 3 – (a) Bathymetric (in reference to the marine chart-sounding datum at La Rochelle Harbor) and topographic (in reference to the Nivellement General de France) map of the Marennes-Oléron Bay and surrounding areas. WMOB, EMOB, SMOB and COS represent the western, the eastern, the southern Marennes-Oléron Bay, and the Coureau d'Oléron Strait, respectively. (b) Simplified sedimentological map of the Marennes-Oléron Bay and surrounding areas (modified from Weber, 2004). From Allard et al., 2010, BSGF n°181.

2.2- Hydrodynamic Setting

Tides affecting the Marennes-Oléron Bay are semi diurnal. MOB is a macrotidal coastal environment with tidal ranges ranging from less than 2 m during neap tides to more than 6 m during spring tides. Tides and the associated tidal currents are the dominant forcing in the sedimentary dynamic of MOB, which has already been investigated using numerical models. Tidal currents control the transport and redistribution of sediments, resulting in a flood-dominated residual sand flux within both the northern entrance and the northern part of the bay (fig. 5), an ebb-dominated residual sand flux in the southern Maumusson Inlet (fig. 5), and a pronounced onshore residual flux on the eastern mud flat of the Marennes-Oléron Bay (Bassoulet et al., 2000)

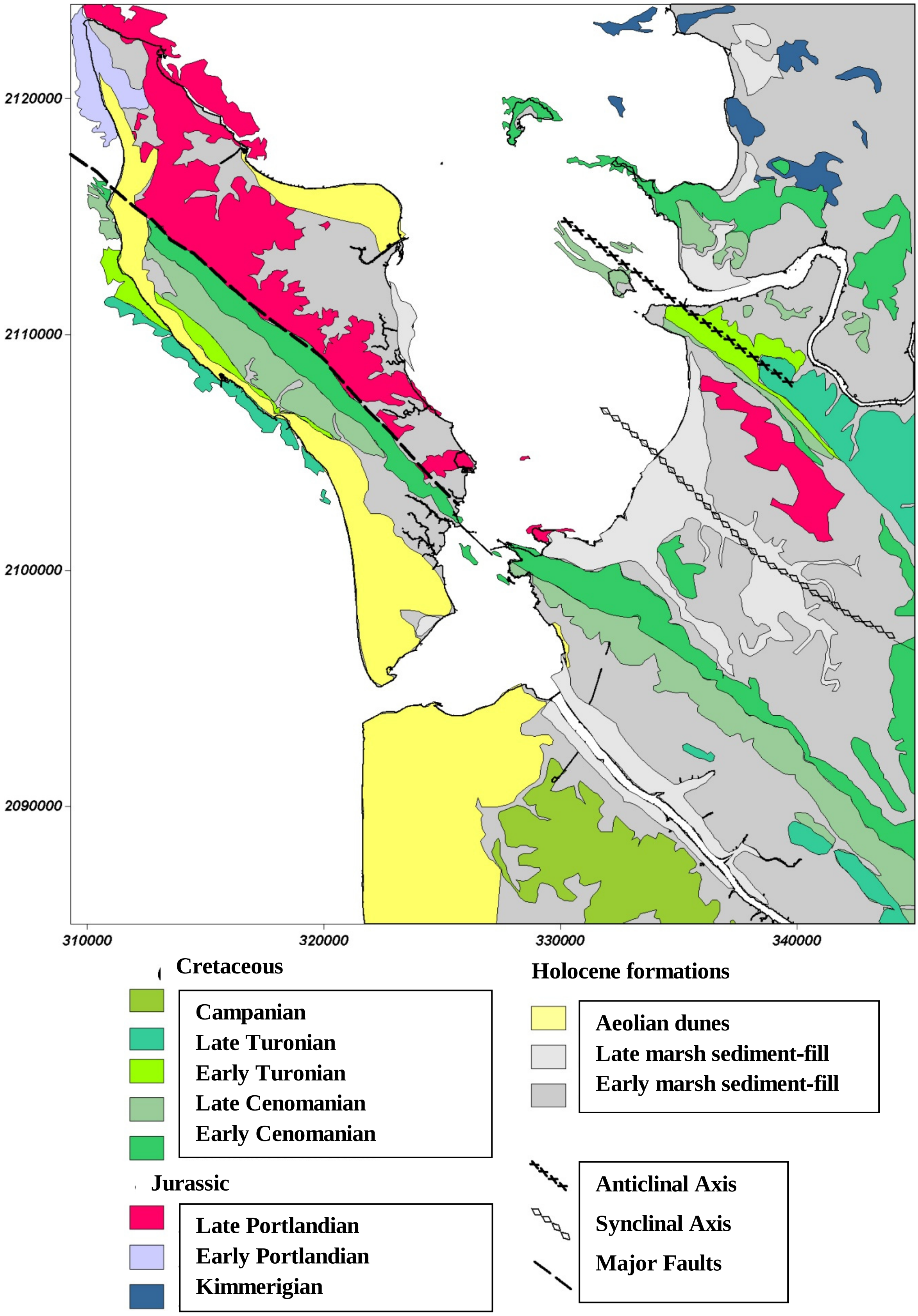


FIG. 4 – Simplified geological map of the Marennes-Oléron Bay (modified from Bourgueil and Moreau, 1974).

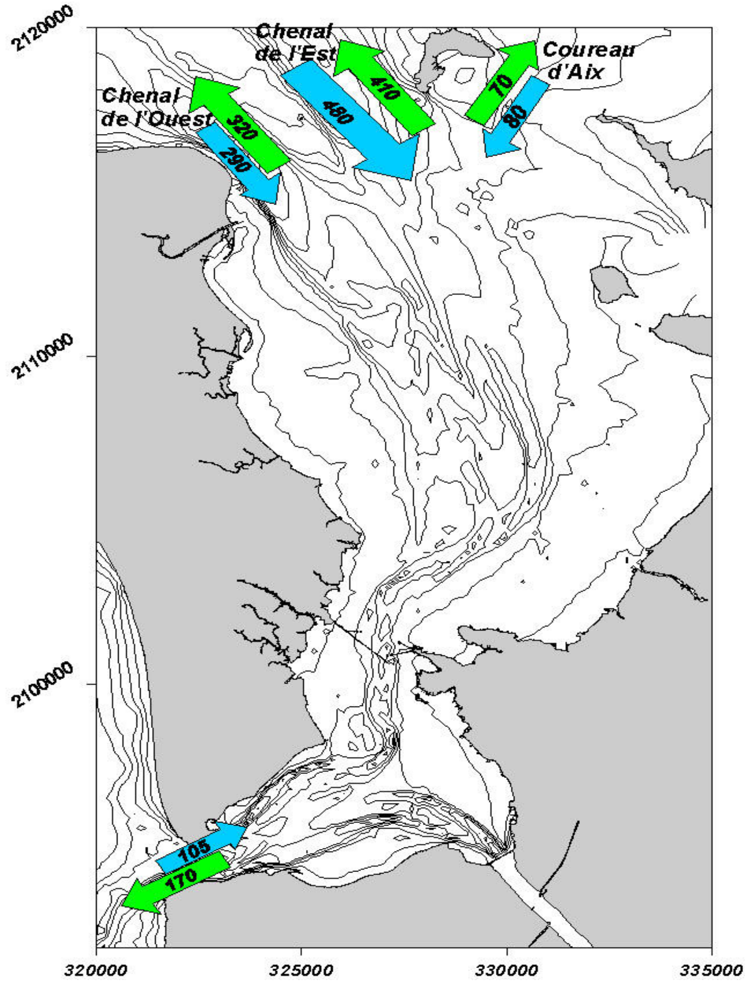


FIG. 5 – Tidal prisms at the inlets of the Marennes-Oléron Bay (*106m3), from tide modeling, Telemac2D on a bathymetry of 2001-2003 (Bertin et al., 2005, PhD Thesis).

Mean annual significant offshore wave height is about 1.5 m, but winter storms can episodically produce waves larger than 9 m (Bertin et al., 2008). Peak wave periods ranging from 8 to 12 s represent more than 60% of the annual wave climate while periods larger than 15 s occur more than 2% of the time. Wave directions are predominantly W to NW, which represent more than 60% of the annual wave climate and induce a southward net littoral drift along the southwestern coast of Oléron island (Bertin et al., 2008). Within the MOB, swells are strongly attenuated but their strong obliquity with respect to the shoreline induces a significant longshore transport, which results in the building of up to hectometer to kilometer long sand spits close to the two entrances of the bay (Bertin et al., 2005). Wind waves induce mud resuspension during the high-water slack period (Bassoulet et al., 2000).

Two small rivers flow into the MOB, the Charente and the Seudre rivers (fig. 2b and c). The volumes of freshwater introduced by these rivers are respectively 65 and 0.55 m3.s-1 (Tesson, 1973), which is more than two orders of magnitude less than those moved by the tide during each tidal cycle (Bertin et al., 2005). The mean annual solid discharge of the Charente River is about 41000 t (Gonzalez et al., 1991). Sediment discharge from the Seudre River was not measured but it is likely much lower than those of the Charente River.

2.3- Morphological Evolutions

A synthesis of historical data and maps (Pawlowski, 1998) was used by Bertin (2005) and Bertin and Chaumillon (2005) to show the evolution of the Marennes-Oléron Bay and Gironde Estuary Mouth (fig. 6 and 7). This synthesis allowed showing huge changes of the coastline since the last centuries and evidenced the communication between MOB and the Gironde estuary during the Middle Age (fig. 6).

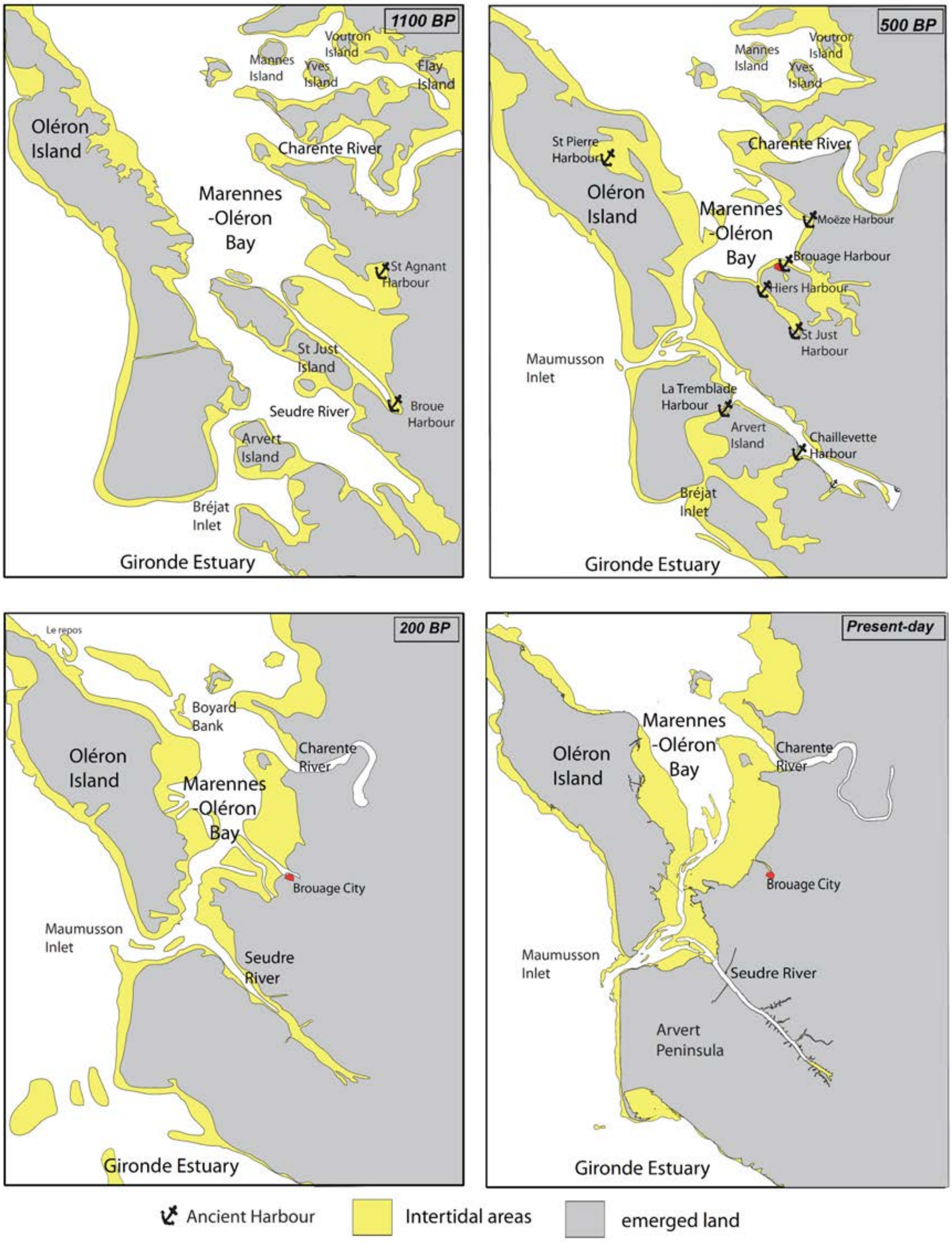


FIG. 6 – Historical evolution of the coastline in the vicinity of the Marennes-Oléron Bay (modified from Bertin, 2005, in Allard et al., 2010, BSGF n°181).

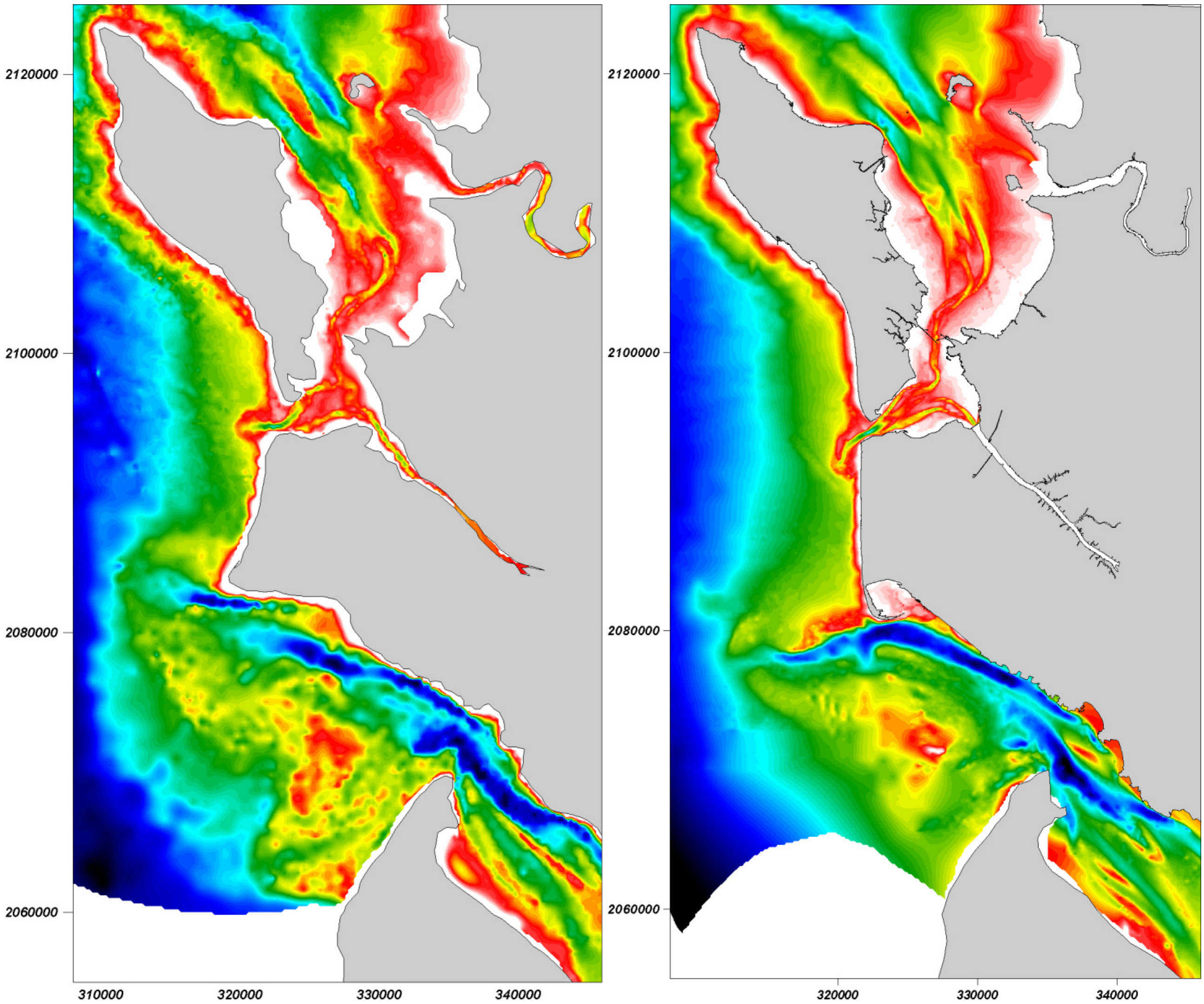


FIG. 7 – 1824–1825 and 1994–2003 bathymetric charts and coastlines of the Marennes-Oléron Bay and Gironde Estuary mouth, showing the some of the strongest morphological changes along the coastlines of France since the two last centuries: Southward migration 5 km (30 $m \cdot yr^{-1}$) of the main channel of the Gironde Estuary Mouth ; (2) Erosion (2,5 km corresponding to an average rate of 15 m·an−1) of the La Coubre headland (immediately northward of the Gironde Estuary Mouth); (3) Development of the 5 km-long La Coubre sandpit (30 $m \cdot yr^{-1}$) immediately northward of the Gironde Estuary Mouth; (4) Southwestward progradation (1,5 km corresponding to an average rate of 10 $m \cdot yr^{-1}$) of the southern part of Oléron Island from 1824 to 1960, leading to decrease of the Maumusson Inlet width from 2 to 1 km; (5) More than 120.10^{6} m^{3} of sediment accretion in the Marennes-Oléron Bay corresponding to an average sedimentation rate of 0,55 $cm \cdot yr^{-1}$.

Comparison of bathymetric maps produced in 1824 and 1995 has allowed computation of the sediment budget of the MOB (fig. 8, Bertin et al., 2005). Calculated sediment budget is clearly positive (106 x 10^{6} m3 added over the last 171 years). Sediment accretion occurred mainly in four types of areas: (1) close to the connections with the open ocean; (2) in the lower parts of intertidal flats; (3) in the wide tidal channel located in the northern part of the bay; and (4) in areas where land reclamation has occurred. Sediment erosion occurred mainly within channels of the southern part of the bay. The recent and rapid sediment infilling of the bay has been attributed to both natural processes and human activities (Bertin et al., 2005; Bertin and Chaumillon, 2006).

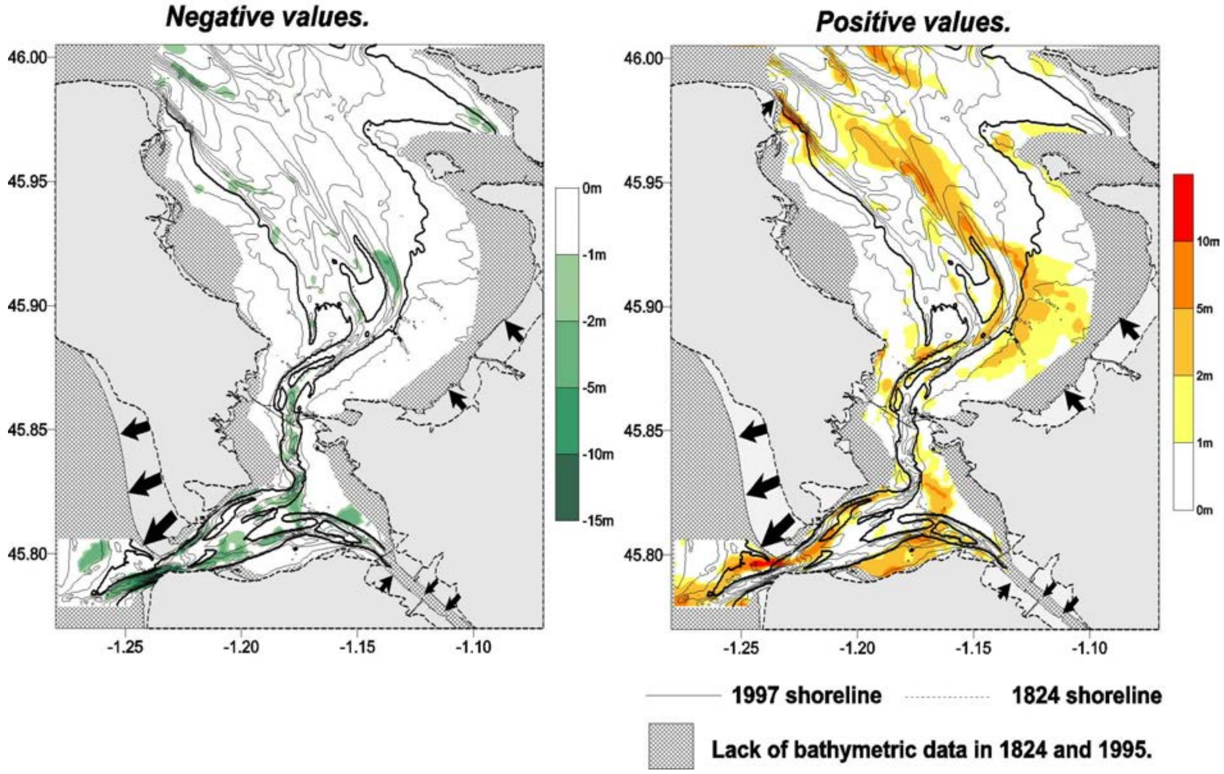


FIG. 8 – Bathymetric and differential maps of the Marennes-Oléron Bay. The bold line indicates the 0 m isobath. Since 1824, sediment filling of the tidal bay has led to a 20% decrease in its water volume, and a 35% reduction of the inlet throat section. Furthermore, the bay is subjected to a very high anthropic pressure, mainly related to oyster farming. Thus, both natural and human-related processes seem relevant to explain high sedimentation rates. Modified from Bertin et al., 2005, Continental Shelf Research n°25.

3- FIELD SESSION IN THE MARENNES-OLERON BAY AND IN THE GIRONDE ESTUARY MOUTH

STOP 1 (1h30): Bellevue sand spit & Northern entrance of the Marennes-Oléron Bay

Objectives :

- Hydrodynamics and sediment dynamic at the northern entrance of the Marennes-Oléron Bay (fig. 9).
- Seismic, bathymetric and core results description obtained on an elongated tide-dominated sandbank, the Longe de Boyard (figs. 10 and 11).
- Seismic and core results in the tidal flats and tidal channels (fig. 12).
- Sedimentary facies & bedforms observations on the Mixed sand-and-mud flat and on the Bellevue Sandspit (fig. 13).
- Development of a Sandspit and evidence of sand infill related to wave processes in the Marennes-Oléron Bay (fig. 14).

Multi time Scale Evolution of a wide estuary linear sandbank,The Longe de Boyard, Atlantic coast of France – *Chaumillon, E., Bertin, X., Falchetto, H., Allard, J., Weber, N., Walker, P., Pouvreau, N., and Woppelmann, G., 2008a. Marine Geology, 251, 209-223.*

The Longe de Boyard Sandbank, lies in a macrotidal estuary environment off the French Atlantic coast. Side scan sonar data combined with shipek grab samples and numerical modeling of waves and tides revealed its short-term dynamics. Historical (1824) and present-day (2000 and 2003) bathymetric data combined with numerical simulations of waves and tides and tide-related sand transport in 1824, and seismic profiling, were then used to demonstrate the long-term evolution of the sandbank and how this correlates with the short-term dynamics. The geological evolution (centuries to millennia) was finally deduced from seismic stratigraphy combined with an analysis of vibrocore samples.

Most of the long-term morphological changes in the ‘Longe de Boyard’ can be explained by the short-term dynamics involving sand transport convergence driven by both tides and waves. Seaward, the changes in the axial part of the bank correspond mainly to erosion and can be explained by wave and tide ravinment. Shoreward, sediment accretion is related to the convergence of tide-related sand transport during ebb and flood due to the dam-effect of the crest of the bank. The changes in the sandbank since 1824 can also be related to a decrease in the tidal channel cross section. The latter was the result of a 10% decrease in tidal currents and tidal prisms subsequent to the rapid sediment infill, and a related 20% reduction in the water volume of the estuary system where the sandbank lies. Seismic stratigraphy and cores showed that the modern sandbank consists of upper clinoforms made of fine sand built up over a core made of coarse sand and gravel related to high-energy environments. Hence, the Longe de Boyard is not only the result of sand convergence driven by both tides and waves but also integrates decreases in the tidal prism subsequent to sediment infilling of the surrounding estuaries on a century and millenia time-scale.

Correlation between VHR seismic profiles and cores evidences a major environmental change recorded in a macrotidal bay –

Billeaud, I., Chaumillon, E. & Weber, O., 2005. Geomarine letters, 25, 1-10.

New, very high-resolution (25 cm) seismic profiles have revealed the internal architecture of the infilling of a macrotidal bay, the Marennes-Oléron Bay, France. Within this geometry, a major seismic unconformity has been correlated with core data. This correlation provides evidence that the seismic unconformity corresponds to a sharp grain-size decrease. Both seismic and core data indicate that this change of grain size can be interpreted as a record of a recent (around 1,000 years B.P.) decrease in hydrodynamical energy with time and/or a larger supply of fine-grained material. This recent environmental change can be related to natural infilling of the Marennes Oléron Bay, and to tidal prism decrease, increasing human activities (e.g. land reclamation, deforestation, agricultural land use) and climate fluctuations during the late Holocene, such as the transition between the cold period of the Dark Ages (1,550–1,050 years B.P.) and the Medieval warm period (1,050–550 years B.P.).

Development of a Sandspit in the vicinity of the northern mouth of the Marennes-Oléron Bay –

The Bellevue Sandspit is an interesting example à wave-dominated sand body developed on a mixed mud-and-sand flat at the entrance of a tidal bay. Such sand body demonstrates that sediment transport related to wave processes can significantly contribute to the sediment-fill of a tidal bay. The Bellevue sand spit did not exist at the beginning of the 19th century. During the 20th century it rapidly developed with area gain rate ranging from less than 500 to almost 2000 m²/yr. At the first order spatial scale, this sand spit is a typical recurved spit. At a second order spatial scale, a cuspate morphology develops (fig. 14), indicating the key role symmetric or slightly asymmetric wave distribution and wind waves with high angle direction with respect to the coastline (Ashton et al, 2001).

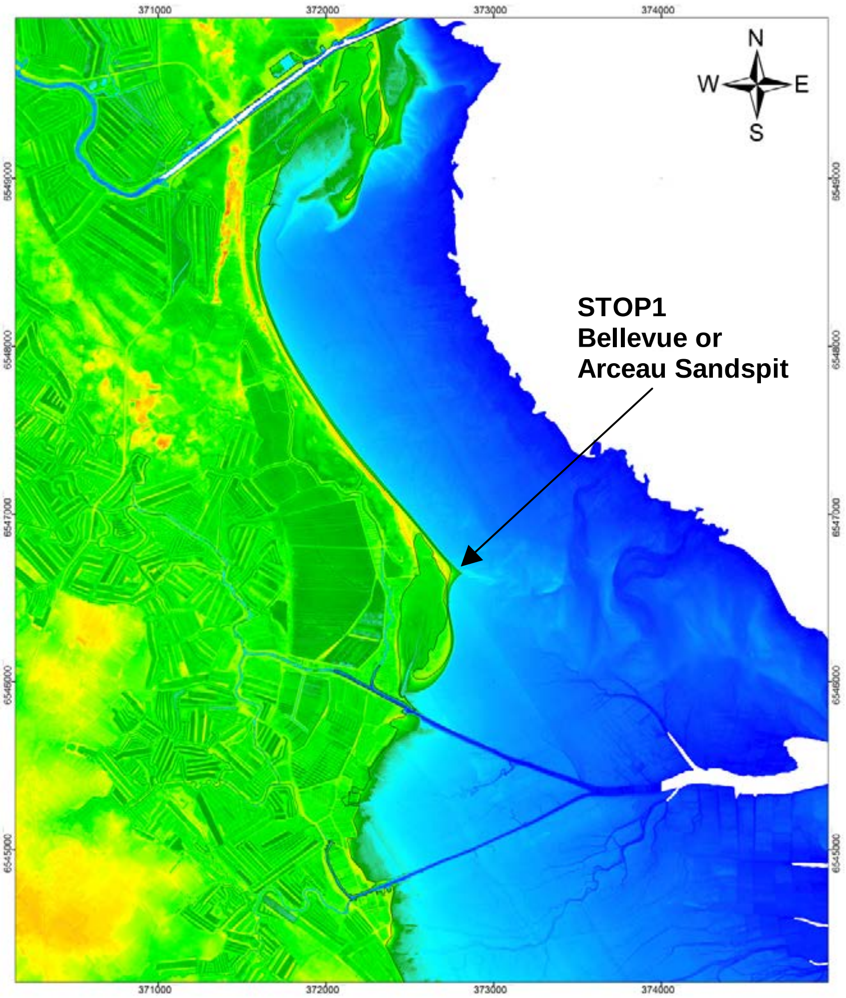


FIG. 9 – Topographic map (IGN Lidar 2010-2011) of the Bellevue Sandspit area.

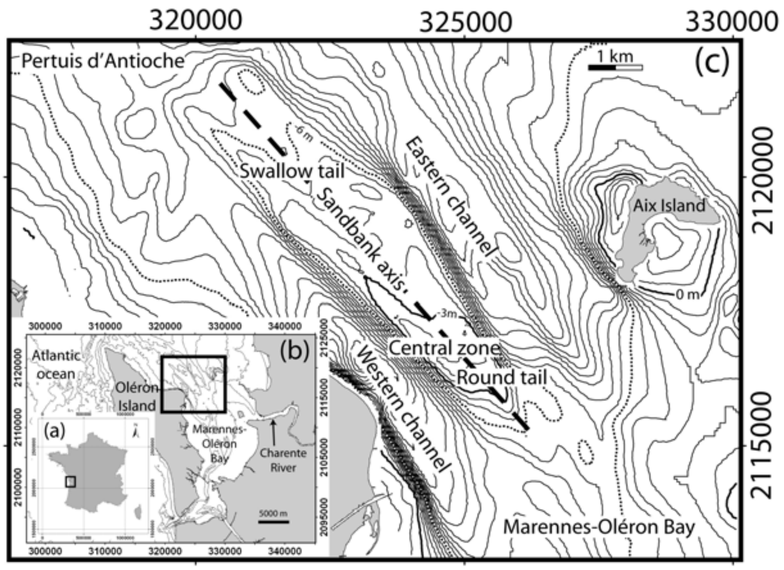


FIG. 10 – (a) Location of the Longe de Boyard Sandbank. (b) Simplified bathymetric map (5m contour intervals) showing the location of the Longe de Boyard Sandbank at the northern entrance of the Marennes-Oléron Bay. (c) Bathymetric map (year 2000, 1m contour intervals) showing the three main morphological zones of the Longe de Boyard Sandbank, the swallow tail northwestward (seaward), the central zone (approximately bounded by the −3 m isobath, in bold) and the round tail southeastward (shoreward). The reference hydrographic level (0 m isobath) is in bold. The sandbank contour is arbitrarily defined as the −6 m isobath (dotted line). From Chaumillon et al., 2008, Marine Geology n°251.

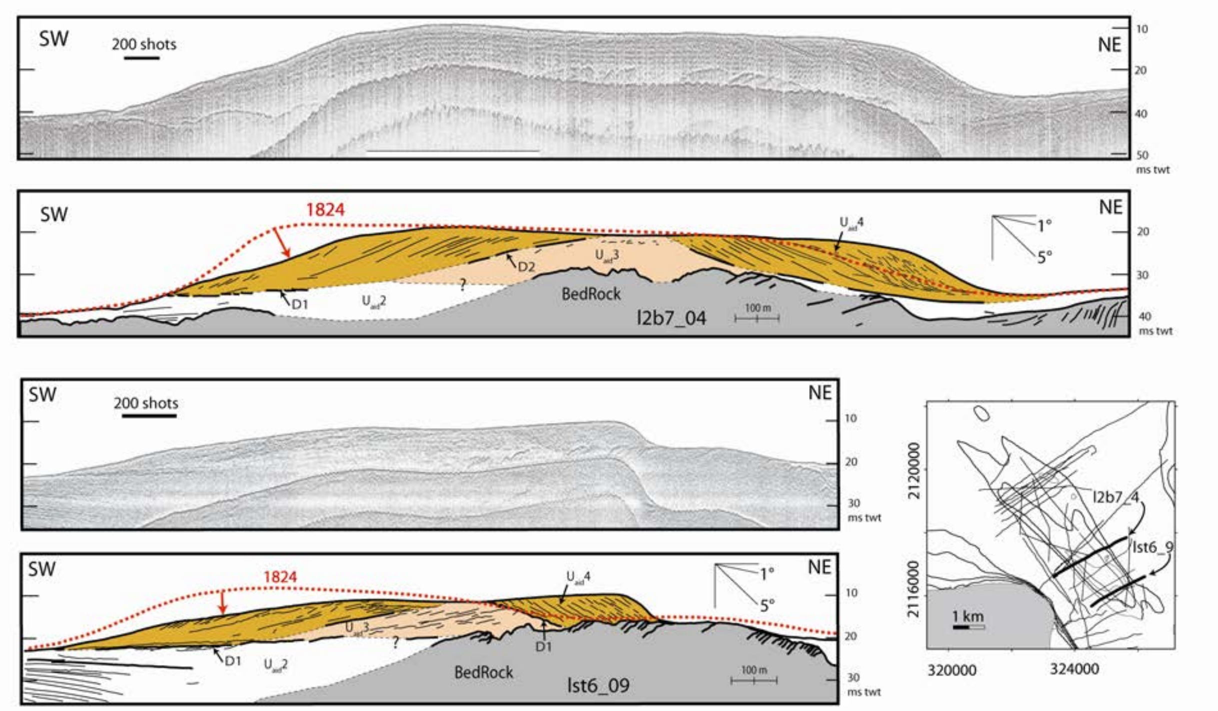


FIG. 11 – Two examples of Boomer seismic profiles (L2b7_04=2620 m and Ist6_09, length=1930 m) showing the internal geometry of the shoreward part of the Longe de Boyard Sandbank (round tail and central zone). The superimposition of the 1824 bathymetric profile shows the internal geometry of sediments emplaced during the past two centuries. The red dotted line corresponds to the 1824 bathymetric profiles and the red arrows underline areas of sediment loss. From Chaumillon et al., 2008, Marine Geology n°251.

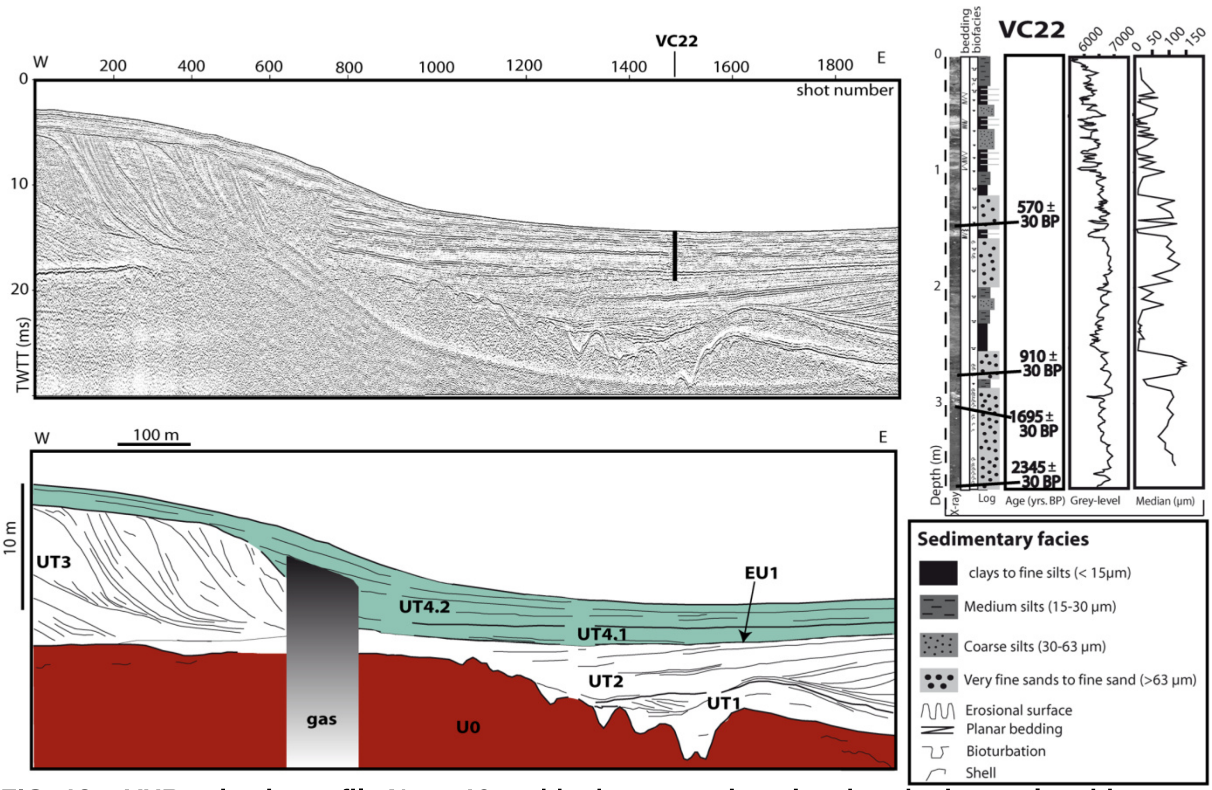


FIG. 12 – VHR seismic profile Noma10 and its interpretation showing the internal architecture of the sediment infilling within the western part of Marennes-Oléron Bay. Detailed description of vibracore VC22 include, from left to right: depth (m below seafloor: bsf), X-ray image, bedding and biofacies, grain size, 14C uncalibrated dates, grey-level curve and median grain-size in micrometers. From Allard et al., 2010, BSGF n°181.

FIG. 13 – Internal stratigraphy of a swash bar (Bellevue Sandspit, 25 August 2011) indicative of a shore-parallel transport of sand toward the Marennes-Oléron Bay (to the left of the photo).

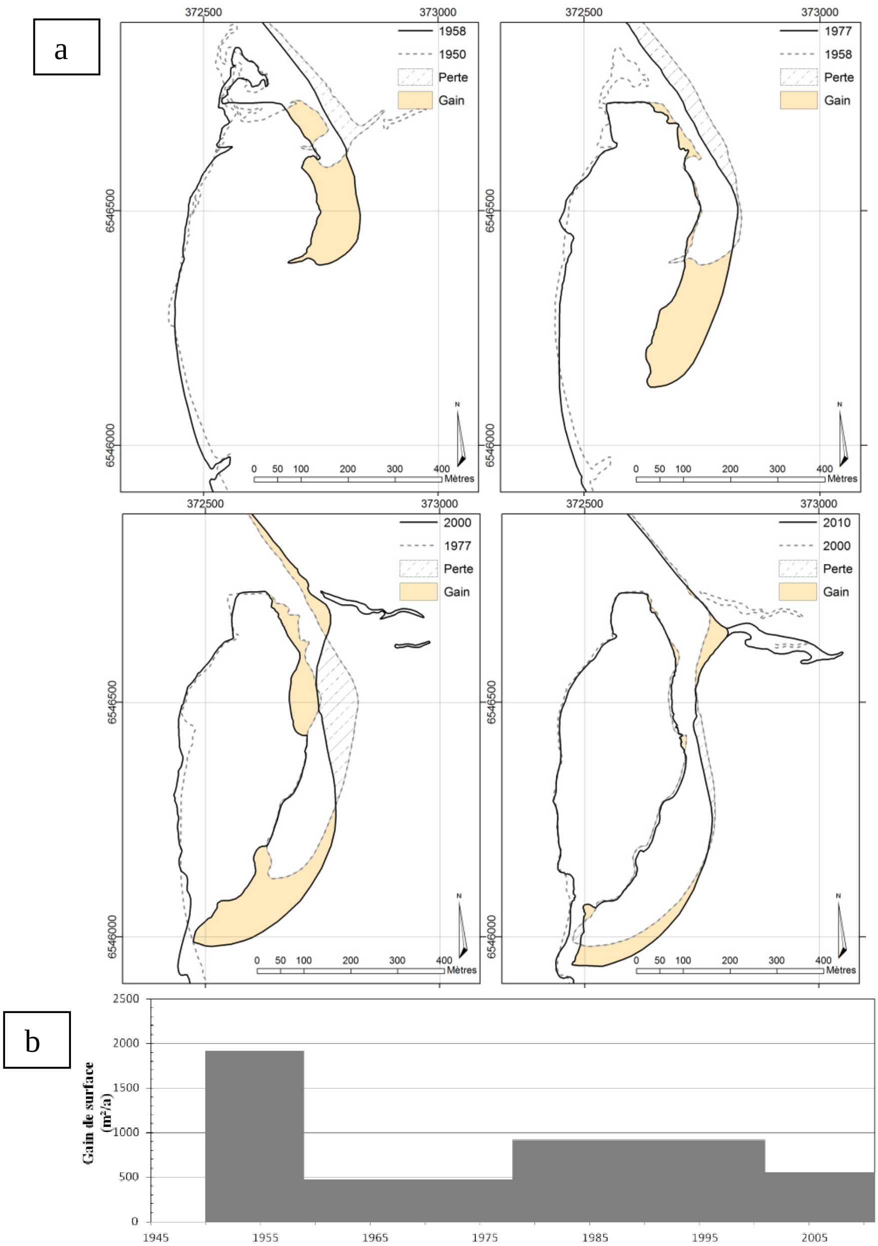


FIG. 14 – a) Morphological evolutions of the Bellevue sandspit (1950-2010). This recurved spit, locally shows a cuspate morphology (well expressed in 2010), probably related to a symmetric or slightly asymmetric wave distribution related to both oceanic and wind waves. b) Area gain rate ($m^2.yr^{-1}$) at the Bellevue sandspit (1950-2010).

STOP 2 (1h30) - Southern entrance of the Marennes-Oléron Bay: Maumusson Inlet

Objectives -

- Seismic & bathymetric results description Landscape observation (Figs. 15 and 16).
- Long-term morphodynamic of a tidal inlet related to the infilling of the tidal bay, implications for the adjacent channels (Figs. 17 and 18).
- Dynamic of bedrock-controlled tidal channels and inlet (fig. 19).
- Wave-driven sediment input in the tidal inlet (fig. 20).

Tidal inlet response to sediment infilling of the associated bay and possible implications of human activities: The Marennes-Oléron Bay and Maumusson Inlet, France –

Bertin X., Chaumillon E., Sottolichio, A. and Pedreros, A. 2005. Continental Shelf Research, Volume 25, Issue 9, 1115-1131.

Maumusson Inlet, located on the French Atlantic coast, connects the Atlantic Ocean with the Marennes-Oléron tidal bay. The tidal range (2 to 6 m) and wave climate (mean height 1.5 m) place this tidal inlet in the mixed energy, tide dominant, category of Hayes [(1979) Barrier island morphology. In: Leatherman, S.P. (Ed.), Barrier Island, Academic Press, New York, pp. 1-28]. An innovative method, combining high quality bathymetric data (nine accurate Digital Elevation Models since 1824) with a very high seismic resolution, demonstrates a major tidal inlet evolution from 1824 onwards and its dramatic acceleration since 1970. The chronology of those morphological changes suggests strong coupling between the location of the tidal channel and the behaviour of the adjacent shorelines. The recent shoaling and migration of the inlet channel can be attributed to a decrease in tidal prism due to the filling in sediment of Marennes-Oléron Bay. Seismic data give evidence that the inlet was located on a major incision of the bedrock. It can be inferred that the bedrock exerts control of channel location, this control varying in time as a function of channel depth. A conceptual model is proposed, including the inlet, its adjacent shorelines, the tidal bay and the time-varying bedrock control of main channel location. Such a model could be considered valid in similar cases along other coastlines, i.e. coastlines with a fine unconsolidated sediment sheet.

Morphological evolution and coupling with bedrock within a mixed energy tidal inlet: the Maumusson Inlet, Bay of Biscay, France –

Bertin, X, Chaumillon, E., Weber, N. and Tesson, M., 2004. Marine Geology 204, 187-202.

Tidal inlet characteristics are controlled by wave energy, tidal range, tidal prism, sediment supply and direction and rates of sand delivered to the inlet. This paper deals with the relations between inlet and lagoon evolutions, linked by the tidal prism. Our study is focused on the Maumusson Inlet and the Marennes-Oléron Bay (first oyster farming area in Europe), located on the western coast of France. The tidal range (2–6 m) and wave climate (mean height: 1.5 m) place this tidal inlet system in the mixed energy (tide, waves), tide-dominated category. The availability of high-resolution bathymetric data since 1824 permits to characterise and quantify accurately morphological changes of both the inlet and the tidal bay. Since 1824, sediment filling of the tidal bay has led to a 20% decrease in its water volume, and a 35% reduction of the inlet throat section. Furthermore, the bay is subjected to a very high anthropic pressure, mainly related to oyster farming. Thus, both natural and human-related processes seem relevant to explain high sedimentation rates. Current measurements, hydrodynamic modelling and cross-sectional area of the inlet throat are used in order to quantify tidal prism changes since 1824. Both flood and ebb tidal prism decreased by 35%. Decrease in the Marennes-Oléron Bay water volume is inferred to be responsible for a part of tidal prism decrease at the inlet. Tidal prisms decrease may also be explained by an increase in frictional resistance to tidal wave propagation, due to a general shoaling and oyster farms in the bay. A conceptual model is proposed, taking into account natural and human-related sedimentation processes, and explaining tidal inlet response to tidal bay evolutions.

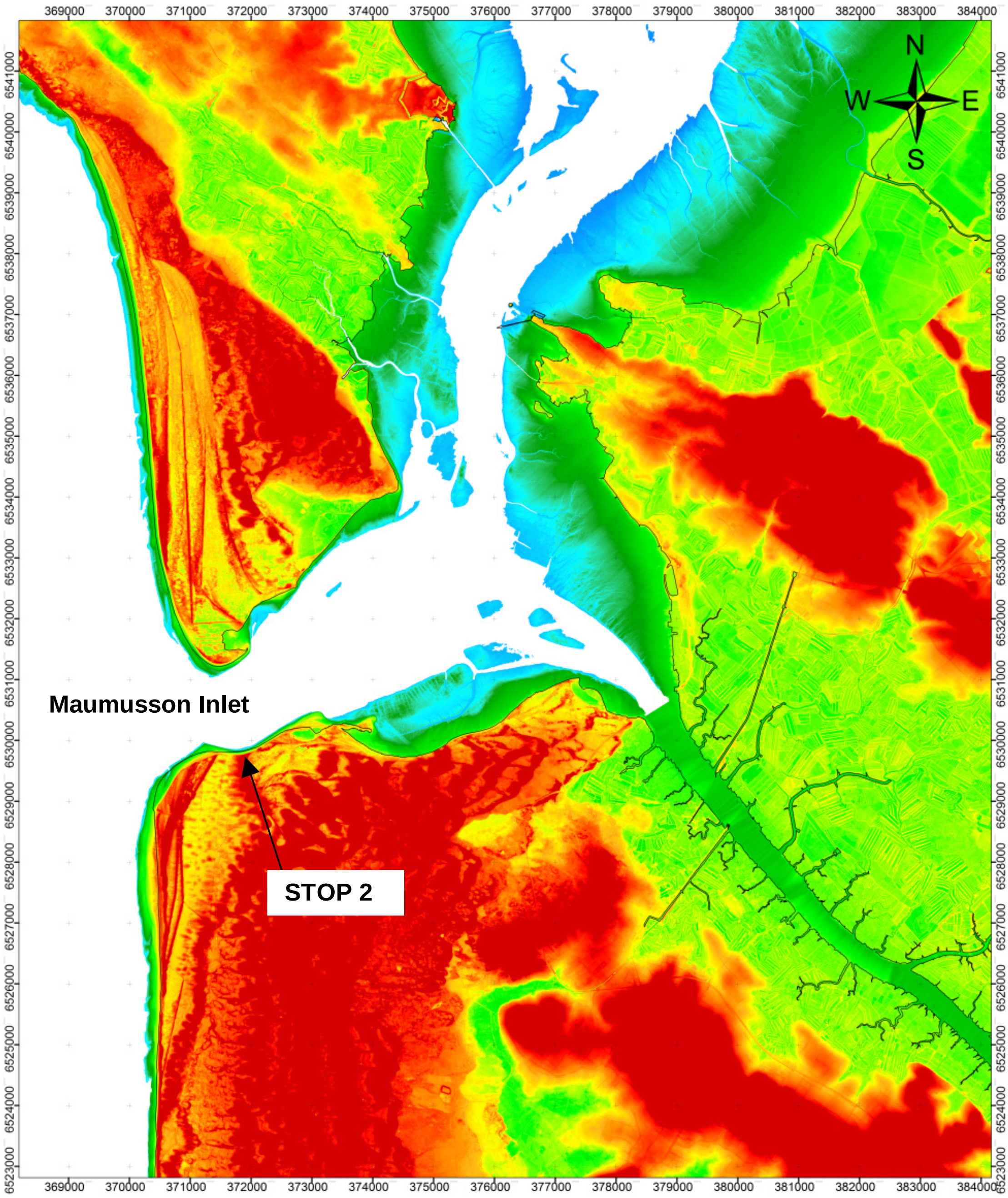


FIG. 15 – Topographic map (IGN Lidar 2010-2011) of the southern Marennes-Oléron Bay and Maumusson Inlet area.

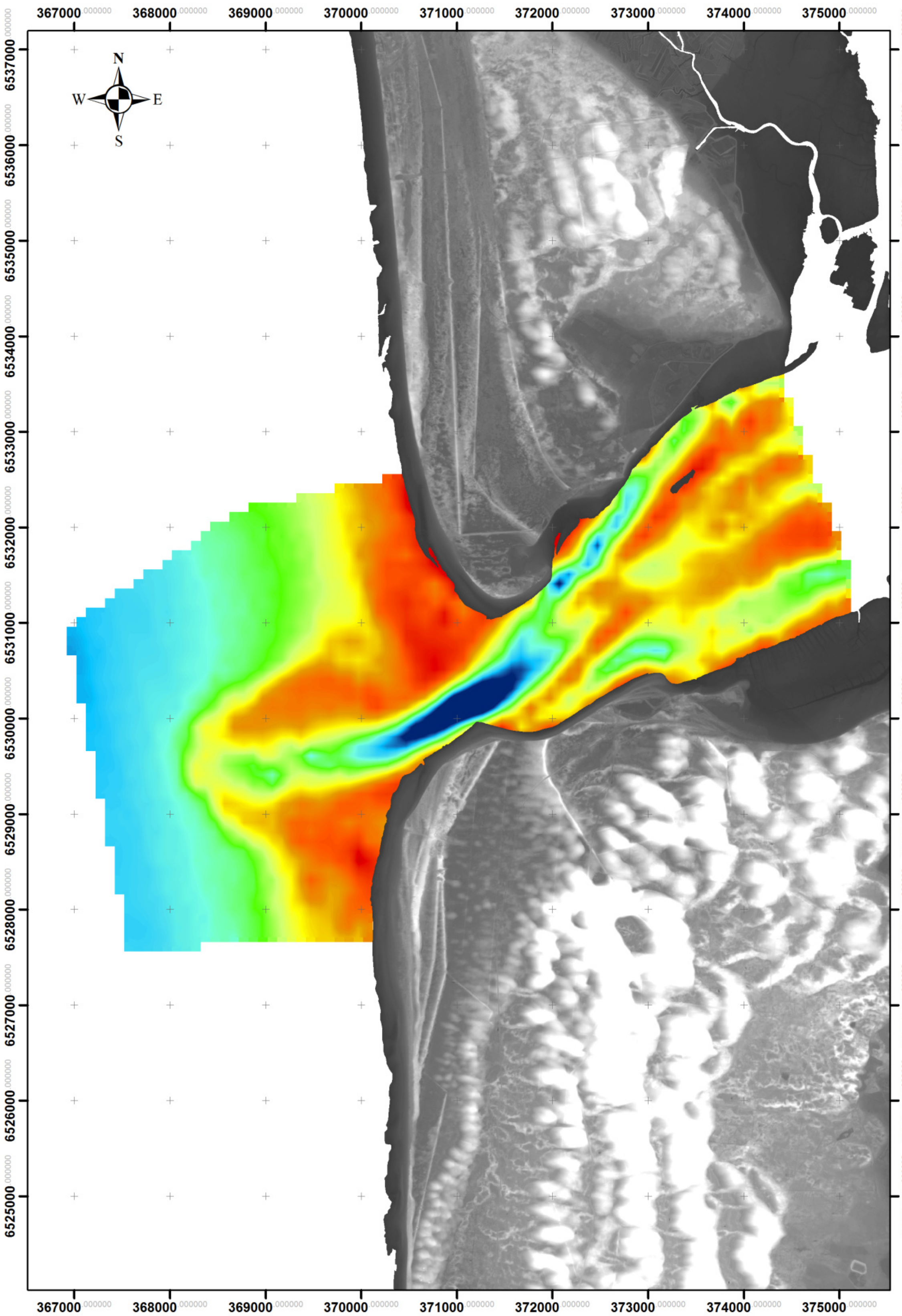


FIG. 16 – Topographic (IGN Lidar 2010- 2011) and bathymetric map (bathymetry of July 2011) of the Maumusson Inlet area.

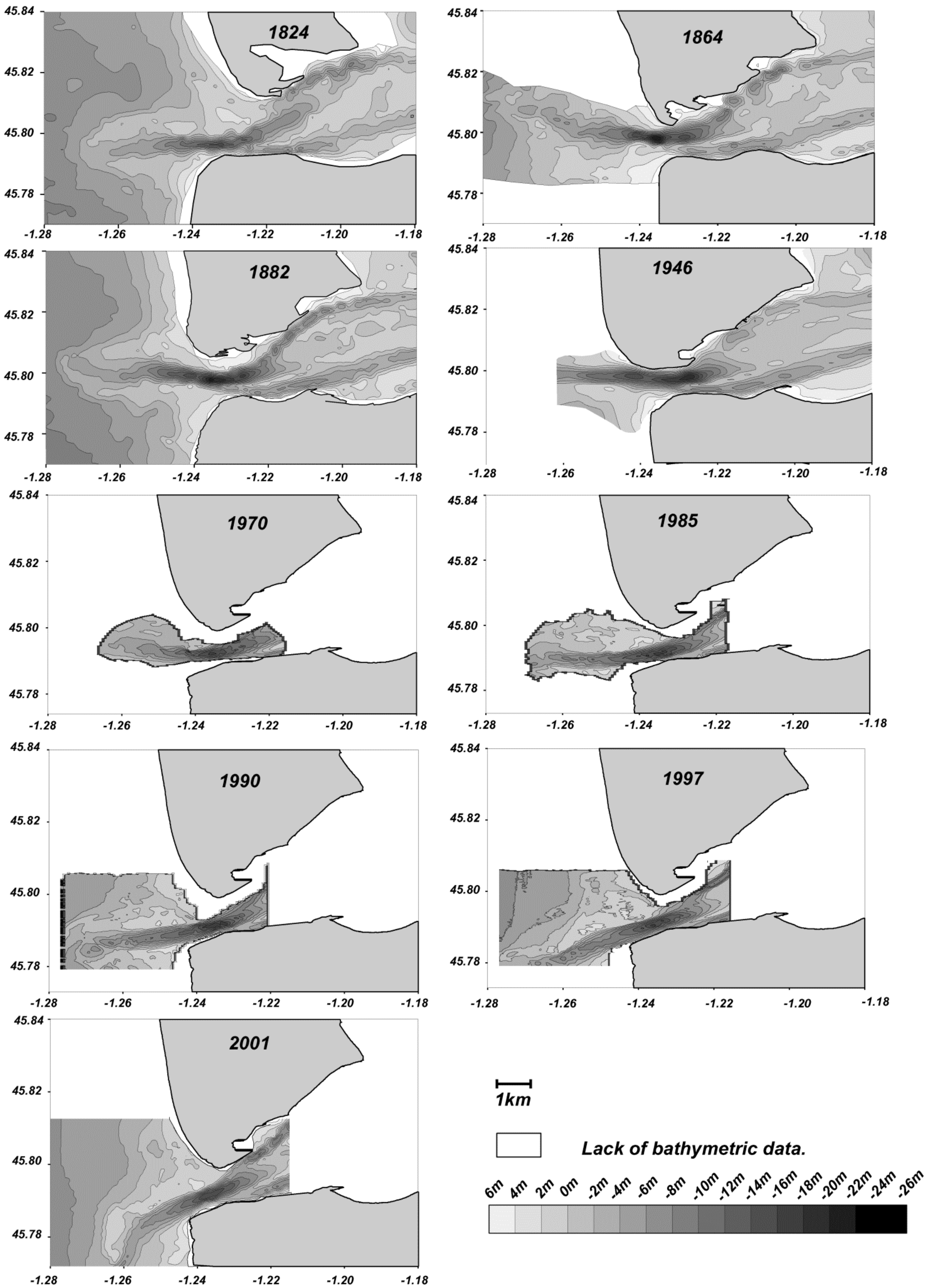


FIG. 17 – DEM displaying the evolution of Maumusson Inlet and its adjacent shorelines from 1824 to 2001 (bathymetric interval is 2 m). From Bertin et al., 2004, Marine Geology n°204.

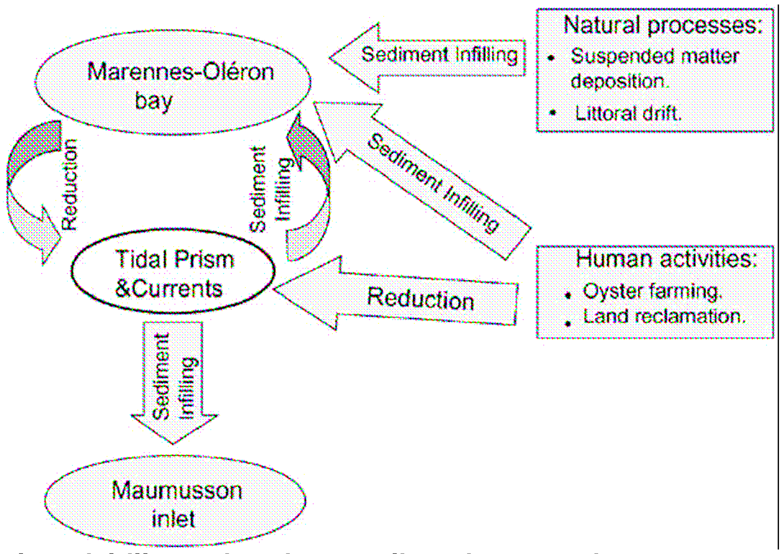


FIG. 18 – Conceptual model illustrating the couplings between the Marennes-Ole´ron Bay and the Maumusson Inlet, linked by the tidal prism (From Bertin et al., 2005, Cont Shelf Res, 25). The relation between the minimum cross-sectional area (A) and the tidal prism (Ω) (1) was firstly recognized by O'Brien (1931).

$$A = C \, . \, \Omega^{n}$$

where n is a dimensionless value ranging between 0.84 and 1.10 (Jarrett, 1976) and C is a scaling coefficient acquiring the dimensions necessary for dimensional balance in view of the selected exponent (Hugues, 2002).

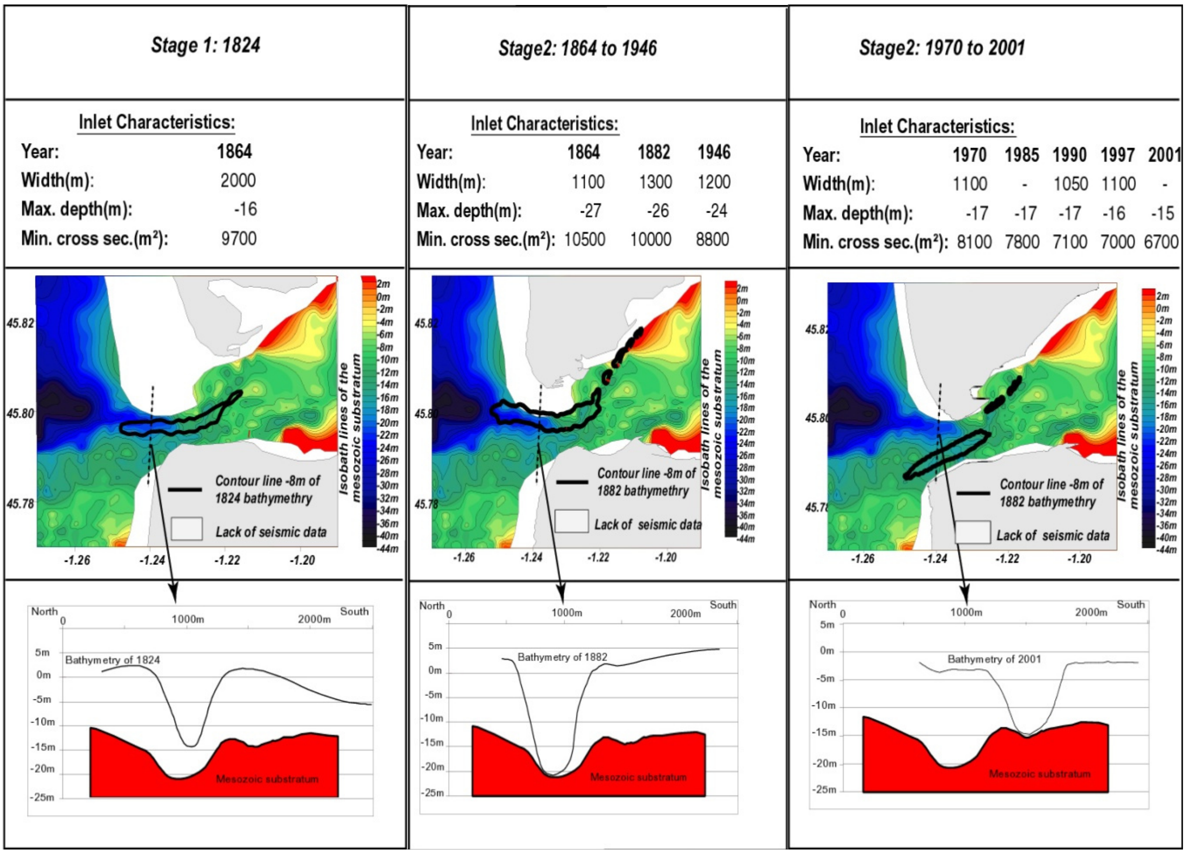


FIG. 19 – Superimposition of the main channel (bathymetric line 38 m) at different times on the isobath map of the Mesozoic substratum. The three stages proposed show that the bedrock control varies in time as a function of the channel depth. The isobath map of the Mesozoic substratum corresponds to a DEM obtained from our seismic data. From Bertin et al., 2004, Marine Geology n° 204.

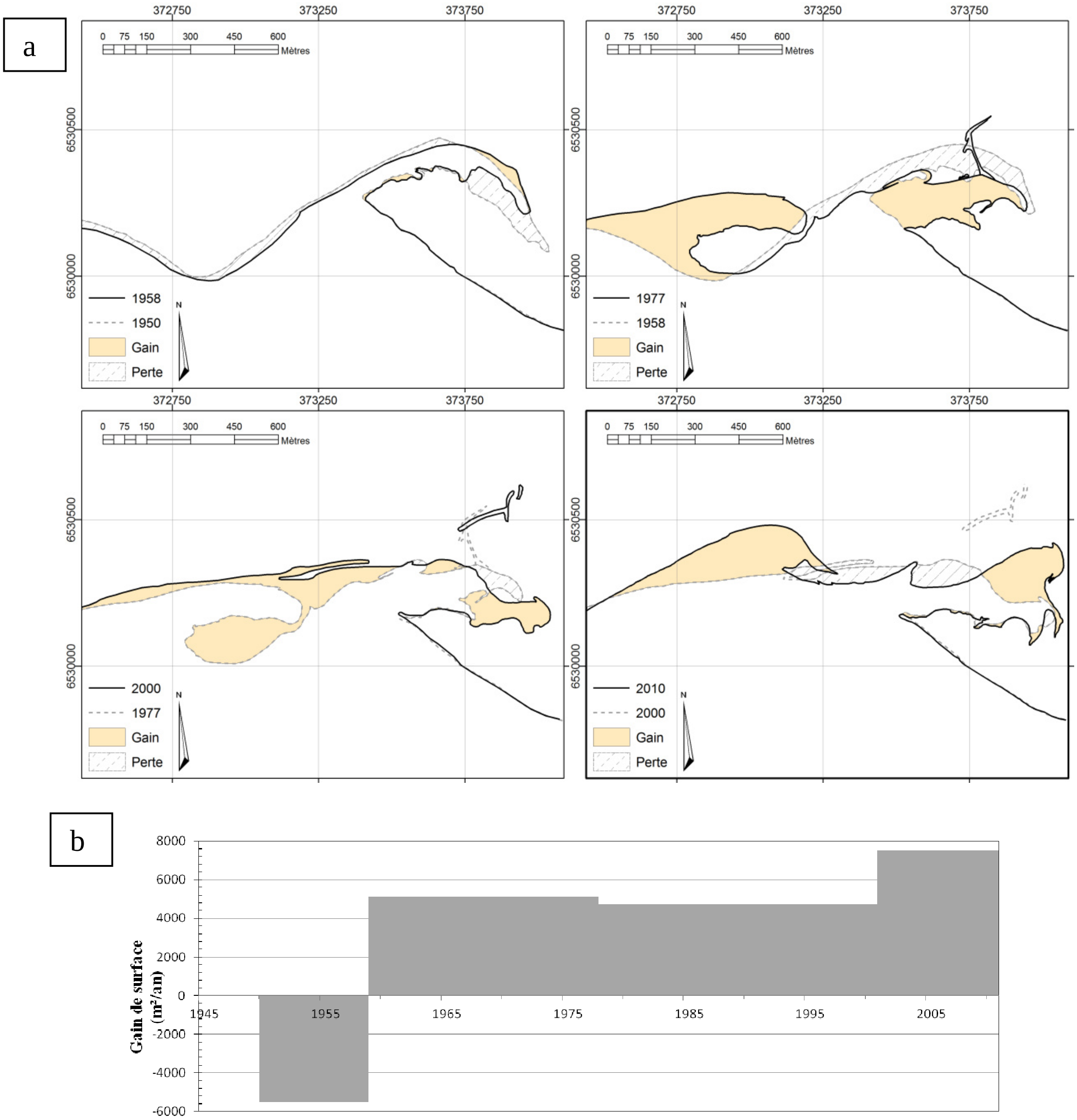


FIG. 20 – a) Morphological evolutions of the Galon d'Or sandspit (1950-2010) showing the wave-driven sediment input in the southern inlet of the Marennes-Oléron Bay. b) Area gain rate ($m^2.yr^{-1}$) at the Galon d'Or sandspit (1950-2010).

STOP 3 – Ancient Brouage Harbour: An example of fast regression (1h)

Objectives -

- Landscape and evidence of fast shoreward migration of the shoreline and HST regression (fig. 21).
- Historical review (figs. 22 and 23).
- Seismic and core results description obtained on a buried tidal channel (figs. 24 to 27).
- Sediment record of lower sea levels (fig. 25).
- Human and climate impact on the late HST (figs. 25 to 27).

Secular morphological evolution and Holocene stratigraphy of a macro tidal bay: the Marennes-Oléron Bay (SW France) –

Allard, L., Chaumillon, E., Bertin, X., Poirier, C., and Ganthy, F. 2010. In: E. Chaumillon, B. Tessier, and J.-Y. Reynaud, French Incised Valleys and estuaries, Bulletin de la Société géologique de France, numéro thématique, t. 181, n°2, 151-169.

The Marennes-Oléron Bay is characterised by a very high sedimentation rate and appears to be an ideal place to investigate the sedimentary record of the major environmental changes that occurred since the last several millennia. The sediment budget of the Marennes-Oléron Bay, between 1824 and 2003, is clearly positive. The flood-dominated northern Marennes-Oléron Bay displays sediment gain in both intertidal and subtidal areas whereas the ebb-dominated southern Marennes-Oléron Bay displays sediment gain restricted to the intertidal area and deepening of subtidal channels. In addition, human influences such as oyster farming may play a role in the sediment gain of the bay. The sediment-fill of the northern Marennes-Oléron Bay consists of five main phases: (1) lenticular units and flooded intertidal flats recording lower sea level periods before 7500 yr B.P.; (2) tidal channel-fills recording changes in tidal drainage pattern from 7500 to 5000 yr B.P.; (3) a subtidal unit which constitutes the main phase of sediment fill in the northern part of the bay from 5000 to 1500 yr B.P.; (4) a major channelized erosional surface related to huge coastline changes from 1500 to 1000 yr B.P.; and (5) a mud drape emplaced during the last millennia and potentially recording historical human impact (deforestation and land reclamation). The sediment fill of the southern Marennes-Oléron consists of sandbanks, mixed sand-and-mud flats and tidal channels, mainly emplaced under wave-and-tide processes since the last centuries. Despite its relatively thin (20 m at the maximum), recent and rapid sediment fill, the stratigraphic organization and morphological evolution of the Marennes-Oléron Bay is very complex and spatially variable. Like in many other estuaries, sediment fill of the Marennes-Oléron Bay was successively controlled by relative sea level changes, and then by sediment supply driven by hydrodynamic changes related to huge coastline migrations, and finally by human activities. Moreover, this kind of “rocky coast” estuary, where the sediment-fill is very thin and discontinuous, is characterised by a bedrock control at each phases of the sediment fill both in terms of preservation in topographic lows and in terms of control on hydrodynamics and related sediment input.

Siltation of river-influenced coastal environments: respective impact of late Holocene land use and high-frequency climate changes –

Poirier, C., Chaumillon, E., Arnaud, F., 2011. Marine Geology, 290, 51-62.

An 8.4 m long sediment core was recovered in the Pertuis Charentais (western France). An accurate 14C Bayesian age model was developed to perform a depth to time conversion of high-resolution sedimentological data obtained on the core. Forest loss and precipitation changes that occurred on the related catchments were reconstructed from data gathered in the literature. A sharp increase in winter rainfall (1700–1750 AD) intensified erosion of catchments previously sensitised by increased forest loss (1640–1710 AD),which resulted in a massive supply of silt to the coast from ca. 1775 AD onwards. A review of articles reporting similar mud layers is also proposed. Depending on their age, some major phases of mankind migration history could be traced back, which confirms the predominant influence of human activities on the siltation of river-influenced coastal environments. We therefore propose to group these mud layers into the “Anthropogenic System Tracts”, in reference to widely used sequence stratigraphy nomenclature, and we suggest that it might help to identify a Holocene to Anthropocene stratigraphic boundary in coastal areas.

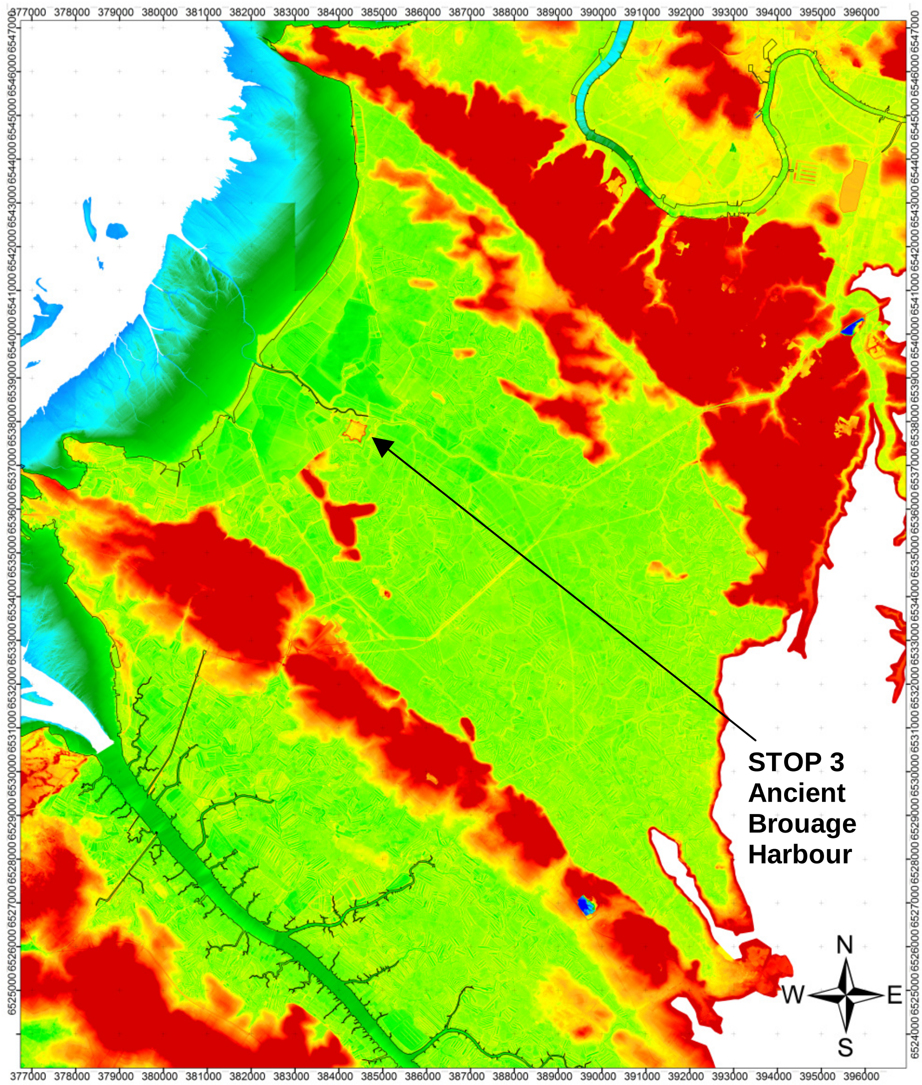


FIG. 21 – Topographic map (IGN Lidar 2010-2011) of the Brouage Marsh.

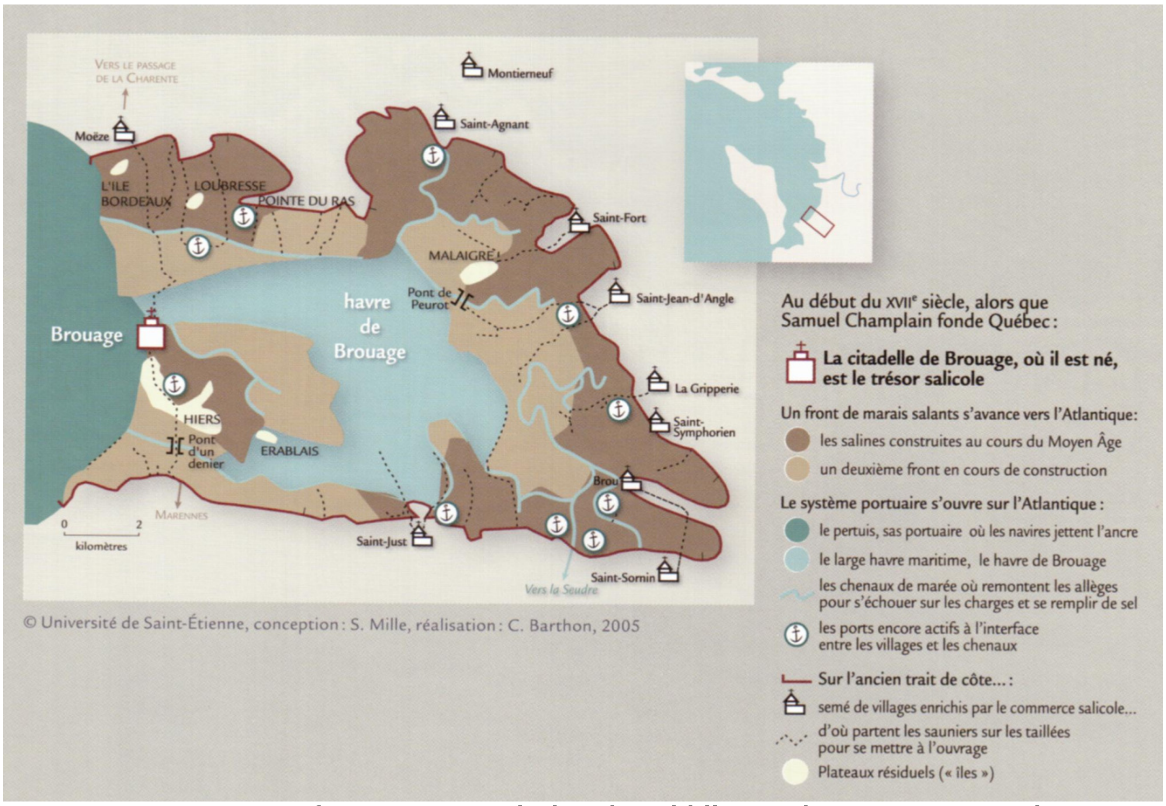


FIG. 22 – Map of Brouage area during the Middle Age (XIIe-XVIIe century).

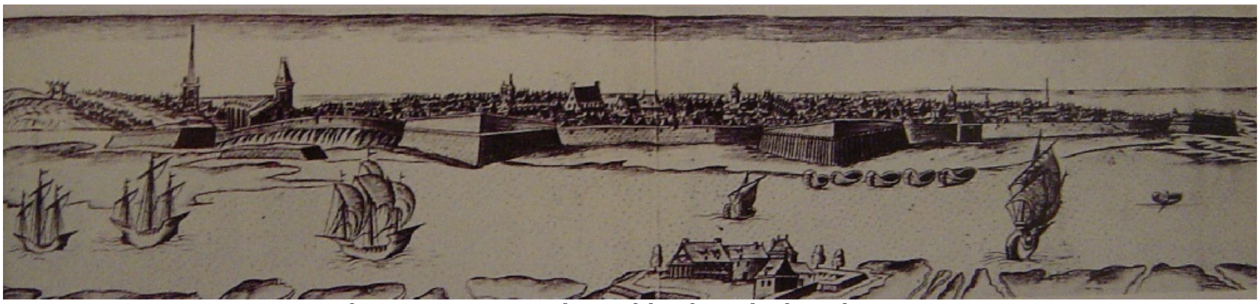


FIG. 23 – The Brouage Castle and harbor during the XVIIe century.

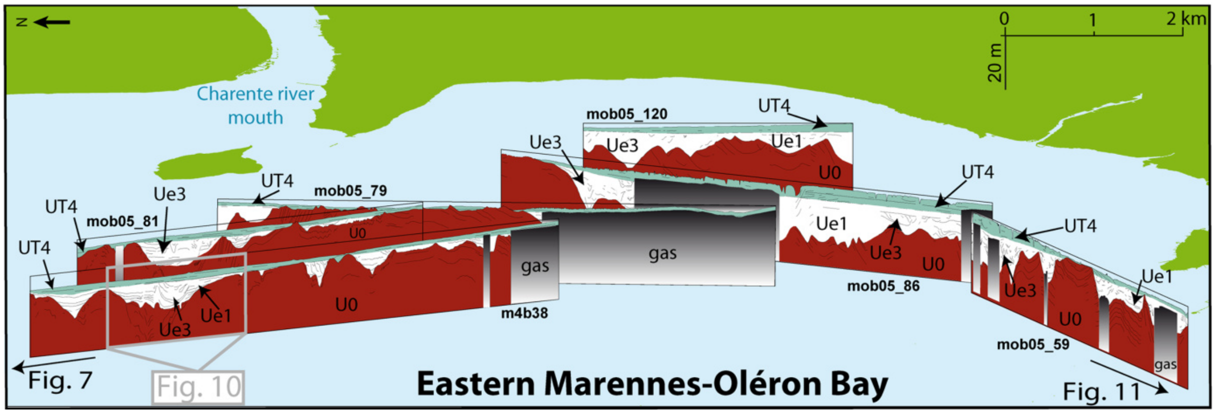


FIG. 24 – Fence diagram of seismic profiles M4b38, Mob05_59, Mob05_79, Mob05_81, Mob05_86 and Mob05_120, showing the internal architecture of the EMOB. Seismic unit U0 corresponds to the bedrock and seismic units Ue1.1, Ue1.2, Ue3 and UT4 correspond to the soft sedimentary cover. From Allard et al., 2010, BSGF n°181.

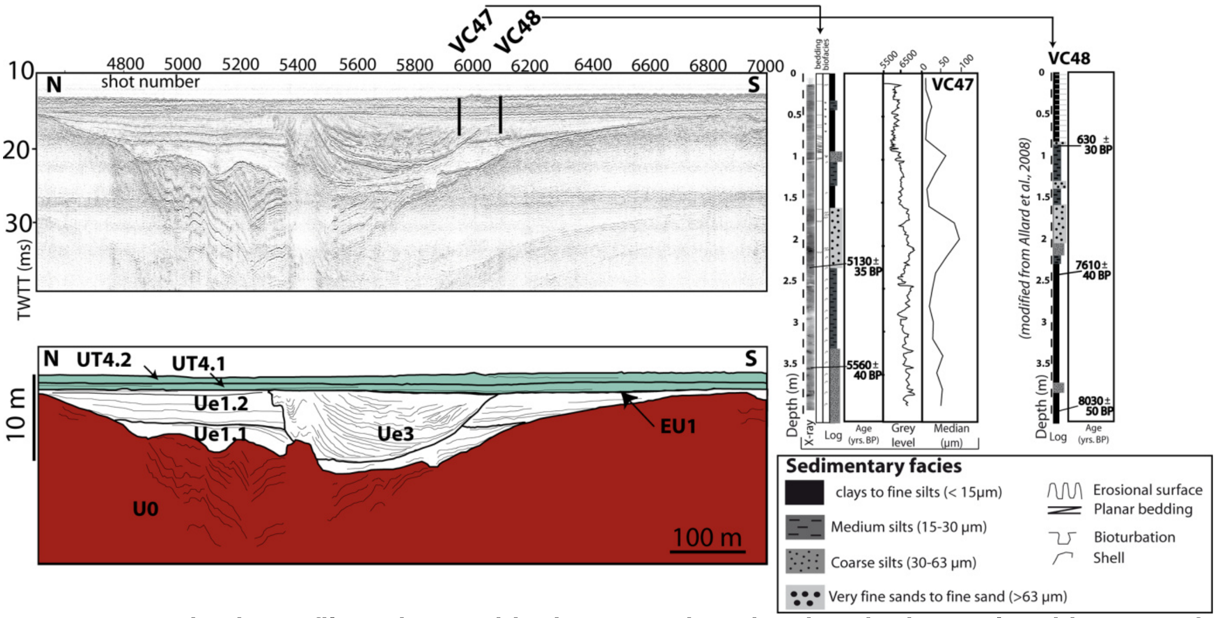


FIG. 25 – VHR seismic profile M4b38 and its interpretation showing the internal architecture of the sediment infilling within the EMOB. Detailed description of vibracores VC48 [modified from Allard et al., 2008] and VC47 includes from left to right: depth (m bsf], X-ray image, bedding and biofacies, grain size, 14C uncalibrated dates, grey-level curve and median grain-size in micrometers. From Allard et al., 2010, BSGF n°181.

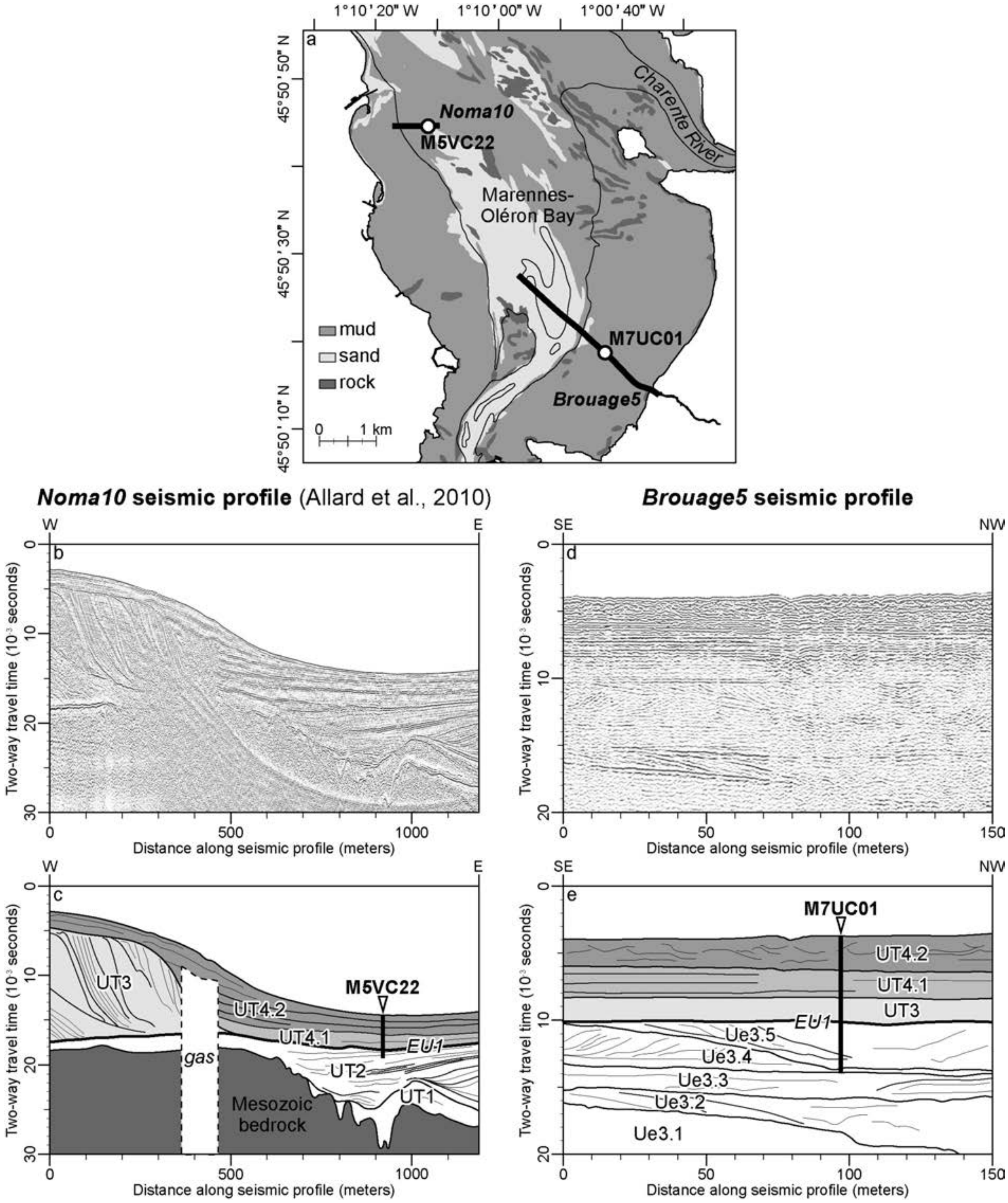


FIG. 26 – Holocene sediment fill of the Marennes-Oléron Bay. (a) Map of the northern Marennes-Oléron Bay, showing the location of the Noma10 and Brouage5 very high resolution seismic profiles and of the M5VC22 and M7UC01 cores. The thin dark line delineates the lowest waterline during spring tides. Sedimentological map is provided by the French hydrographic office SHOM. (b) Raw Noma10 seismic profile, sampled above the western subtidal Marennes-Oléron Bay mud drape, modified from Allard et al. (2010). (c) Interpreted Noma10 seismic profile, with the location of the M5VC22 core, modified from Allard et al. (2010). (d) Raw Brouage5 seismic profile, sampled above the eastern intertidal Marennes-Oléron Bay mud drape. (e) Interpreted Brouage5 seismic profile, with the location of the M7UC01 core. Note the different horizontal and vertical scales between the two profiles. Heavy black line delineates the EU1 erosional unconformity. From Poirier et al., 2011, Marine Geology, 290, 51-62.

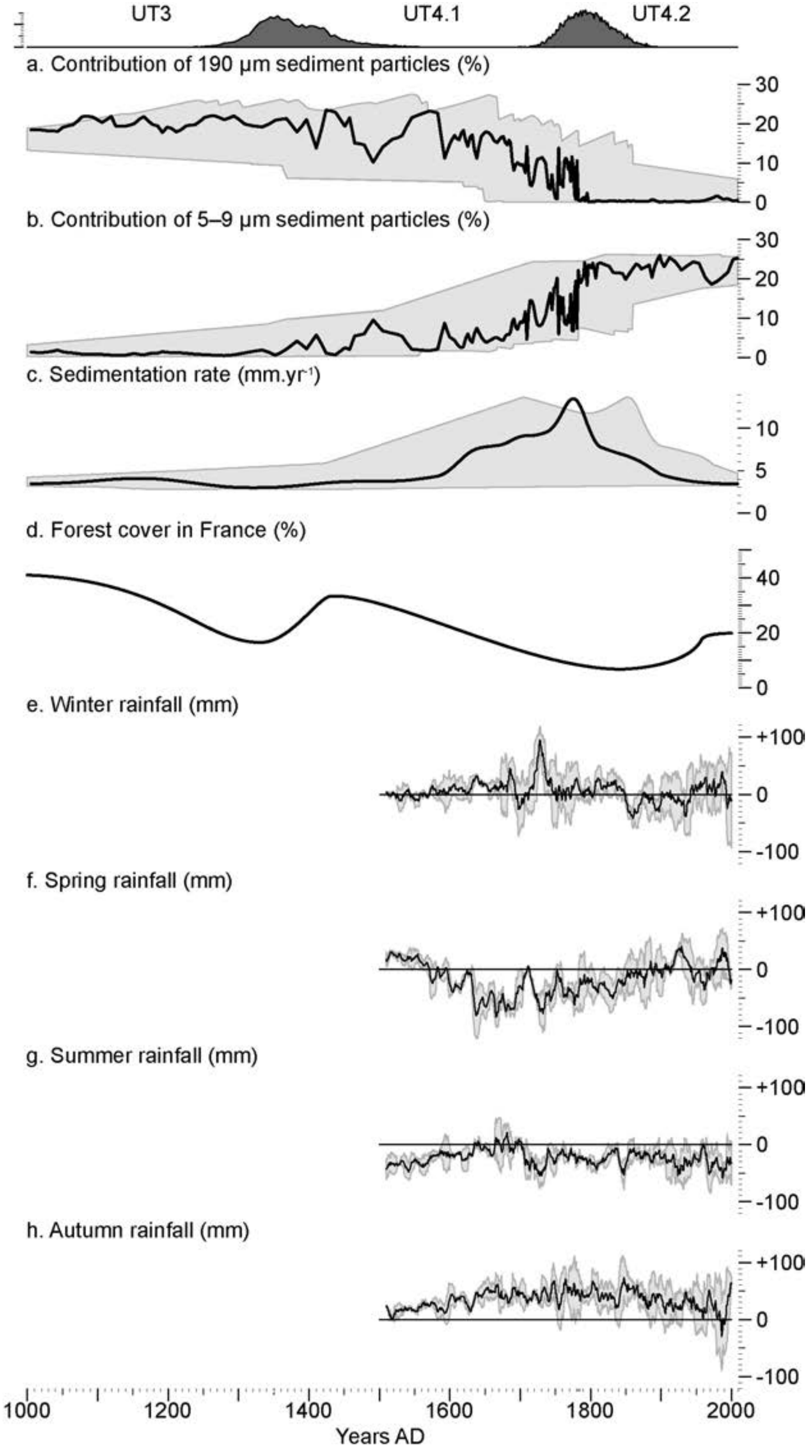


FIG. 27 – Temporal changes in (a) Contribution of 190 µm and (b) of 5 – 9 µm sediment particles in M7UC01 core (in %). (c) Sedimentation rate calculated from the Bchron age-depth model (in mm.yr-1). The grey shadings in a, b and c represent the 95% confidence interval of the Bchron age model after depth to time conversion (Fig. 4). (d) Forest gain and loss in France (in %.yr-1), calculated from Mather et al. (1999) forest cover data. Positive values correspond to forest gain (i.e. reforestation) and negative values to forest loss (i.e. deforestation). (e) Winter, (f) Spring, (g) Summer and (h) Autumn rainfall deviation to annual mean (in mm) over the Charente and Gironde catchments, extracted from Pauling et al. (2006) precipitation dataset. The dark line represents a 9-yr running mean and the grey shading represents the corresponding 9-yr running standard deviation. From Poirier et al., 2011, Marine Geology, 290, 51-62.

STOP 4 (1h30) - Bonne Anse Bay and the Gironde Estuary Mouth

Objectives -

- Bathymetric and coastline evolution description (figs. 28 to 32).
- Dynamic of the Gironde Estuary Mouth, a mixed tide-and-wave environment (figs. 29 to 32).
- Transgression and regression occur in the same area.
- Observation of sedimentary facies & bedforms on the sand flat and along tidal channels.

For this last stop, we will visit the Bonne Anse Bay, located at the mouth of the Gironde Estuary. This tidal bay developed following the rapid growth (Average rate growth of 50 m.yr-1) during the 20th century of a recurved sandspit (La Coubre Sandspit, Fig. 29). This 5 km-long sandpit record seaward and eastward long shore sediment transport. The sand supplied to this sandspit was partly originated from the La Coubre Headland which eroded with rates ranging from 15 to 20 m.yr-1 during the 19th and 20th centuries. This huge erosion leaded to the destruction of the La Coubre lighthouse in 1907. Within the Bonne Bay, extensive sand and mixed sand-and-mud flats develop. Many subaqueous dunes can be observed and usually demonstrate flood-dominated sand transport and the flats and ebb-dominated sand transport close to the tidal channels.

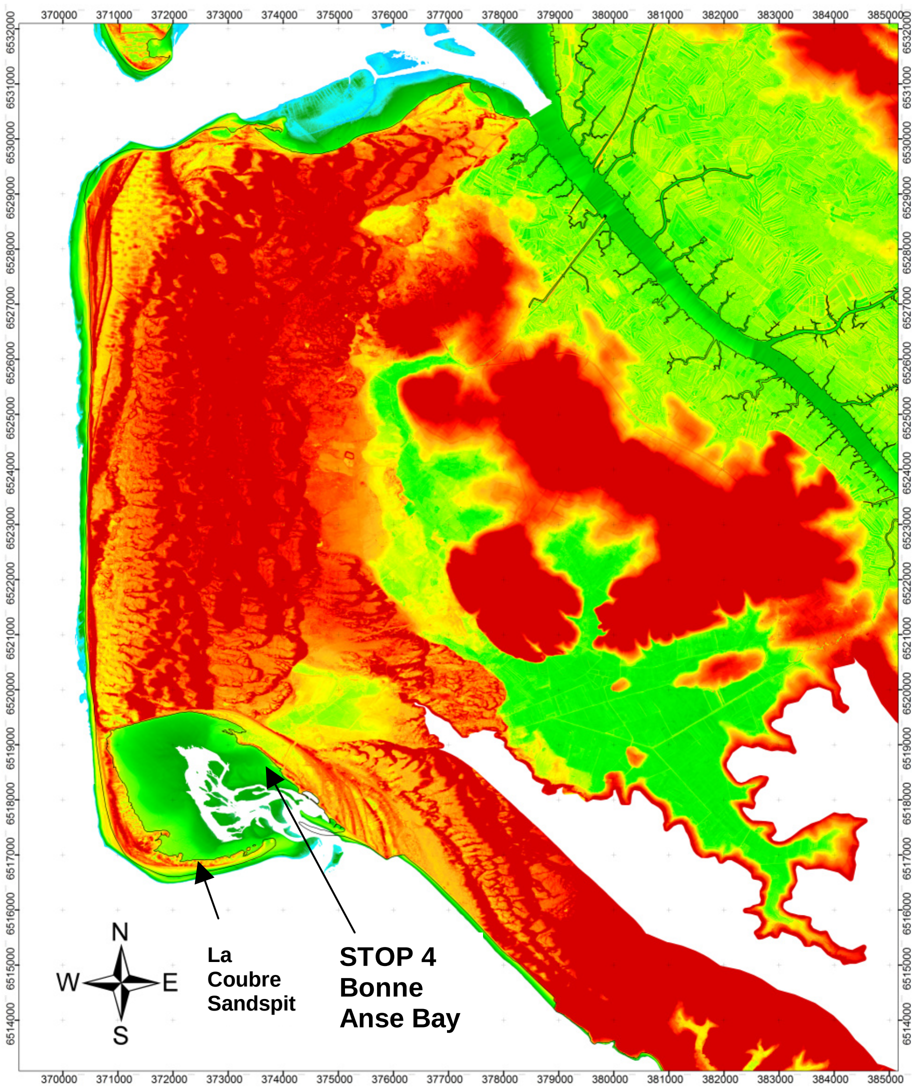


FIG. 28 – Topographic map (IGN Lidar 2010-2011) of the Arvert Peninsula.

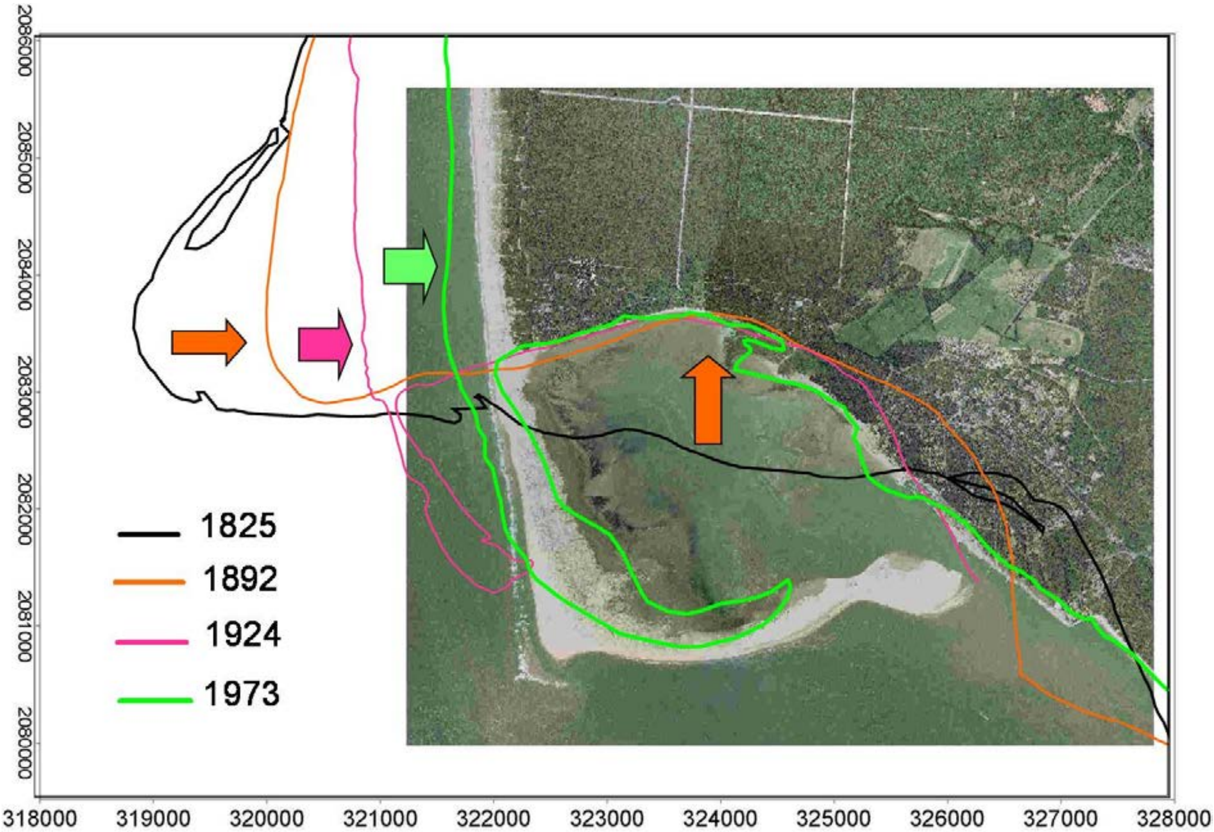


FIG. 29 – Evolution of the upfrift coast of the Gironde estuary mouth.

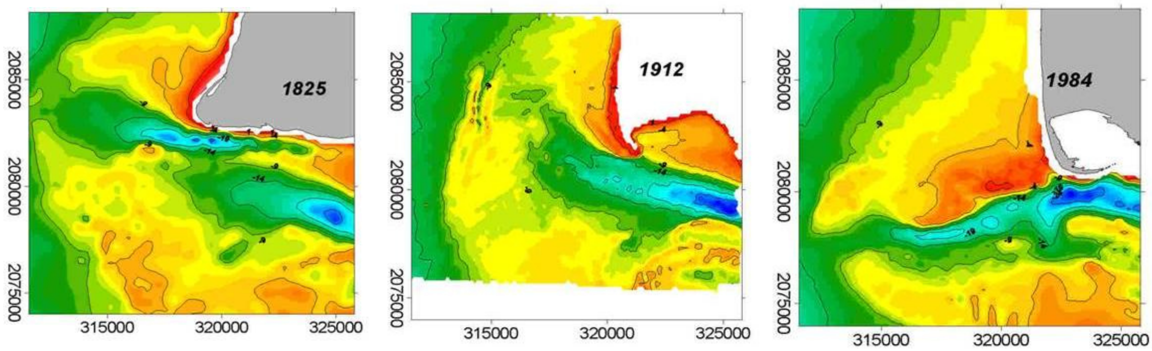


FIG. 30 – Bathymetric evolution of the Gironde estuary mouth.

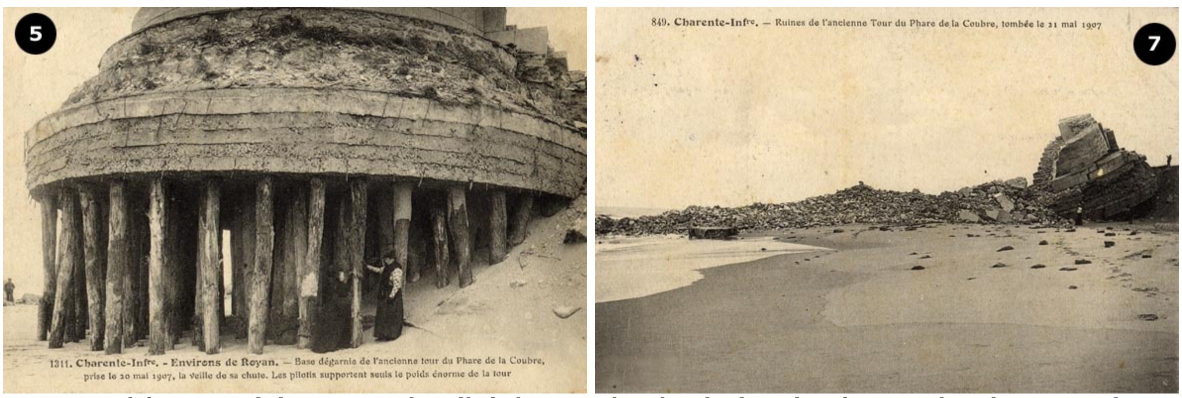


FIG. 21 – (5) Base of the La Coubre lighthouse the day before its destruction (20 May of 1907). (7) La Coubre lighthouse after its destruction.

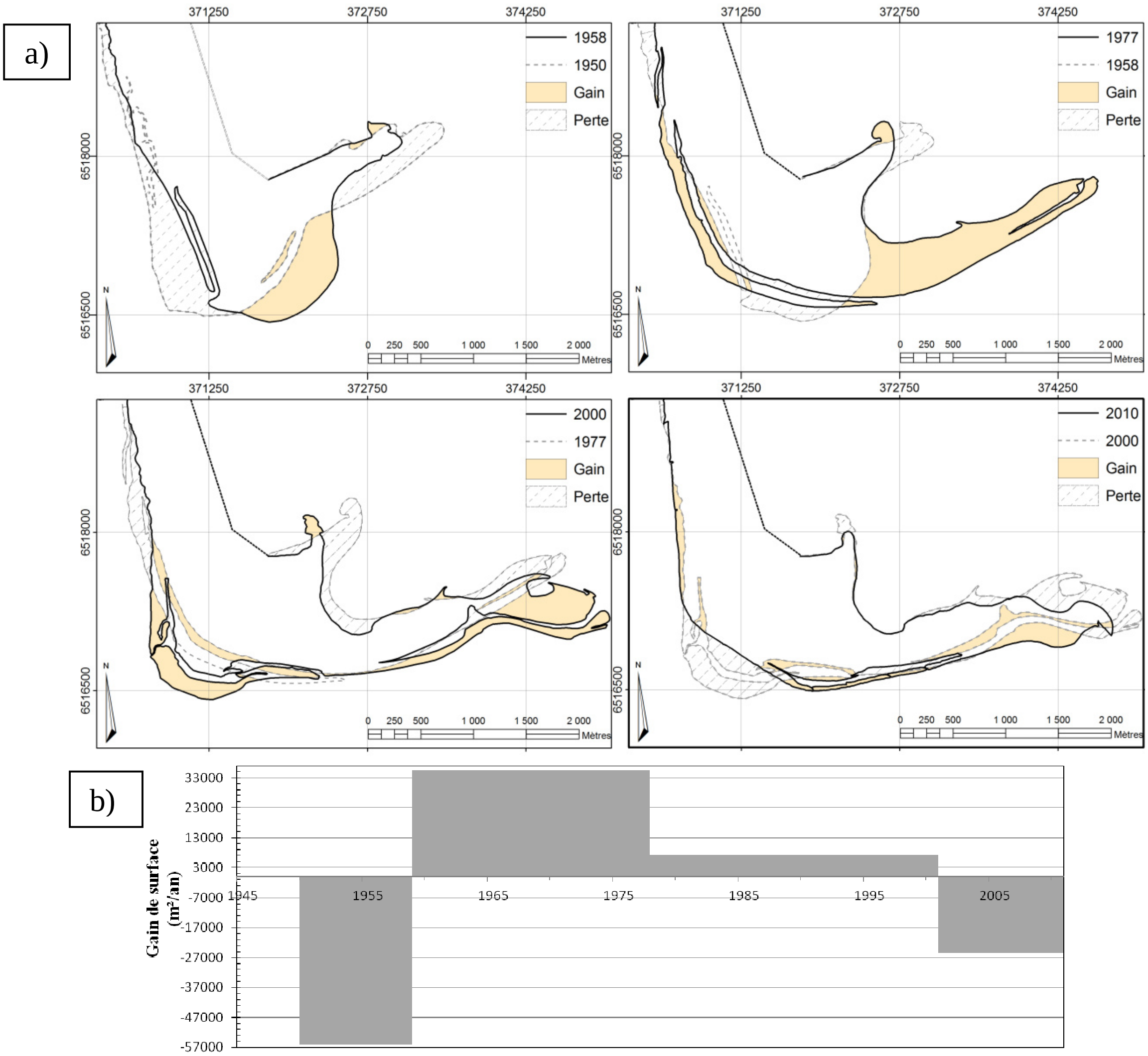


FIG. 32 – a) Morphological evolutions of the La Coubre sandspit (1950-2010).
b) Area gain rate ($m^2.yr^{-1}$) at the La Coubre sandspit (1950-2010).

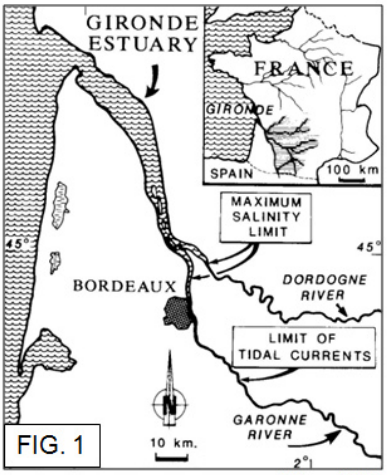
JULY 29 Th. 2012: FIELD SESSION IN THE GIRONDE ESTUARY

1- CHARACTERISTICS OF GIRONDE INCISED VALLEY

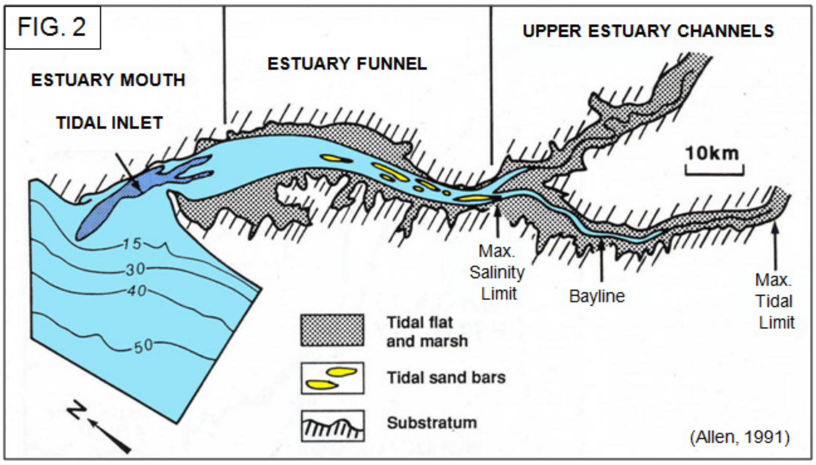
Morphology, hydrology and sedimentary patterns

Morphology

The Gironde Estuary forms an elongate coastal indentation at the confluence of the Dordogne and Garonne Rivers (Fig. 1). The estuary tapers landward from a maximum width of 18 km near its mouth, to less than 3 km at the confluence of the two rivers 80 km upstream. The drainage basin of the Dordogne and Garonne Rivers covers approximately 75,000 km² and includes the Massif Central and the Pyrenees mountains.
The morphology of Gironde estuary is typical of wave- and tide-dominated estuaries and is characterised by a tripartite morphological zonation (Allen, 1991) comprising : 1) meandering upper-estuary channels with tide-dominated sand and mud estuarine point bars, 2) a mid-estuary funnel-shaped channel containing linear tidal sand bars and estuarine mud, and 3) an estuary mouth with a deeply-scoured (35 m) tidal inlet terminating seaward and landward in sandy tidal-delta shoals (Fig. 2).



(Allen, 1991)



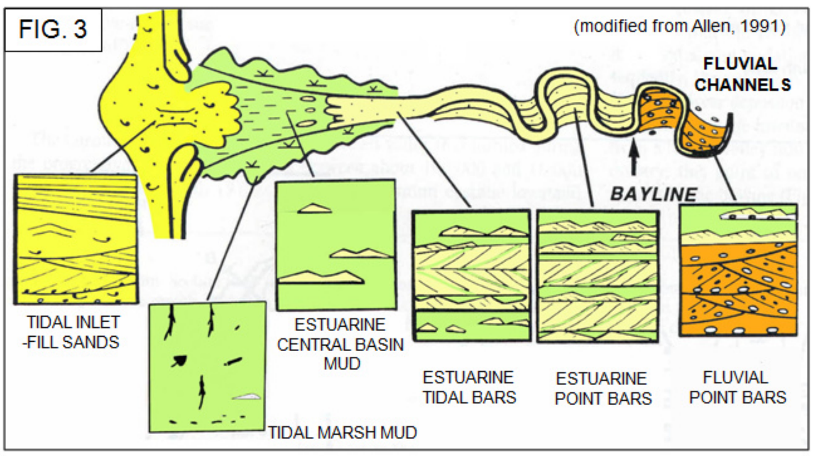


Fig.1 :
Drainage basin of the Gironde estuary

Fig.2 :
Morphology and depositional environments of the Gironde estuary

Fig.3 :
Present-day facies distribution along the Gironde incised valley

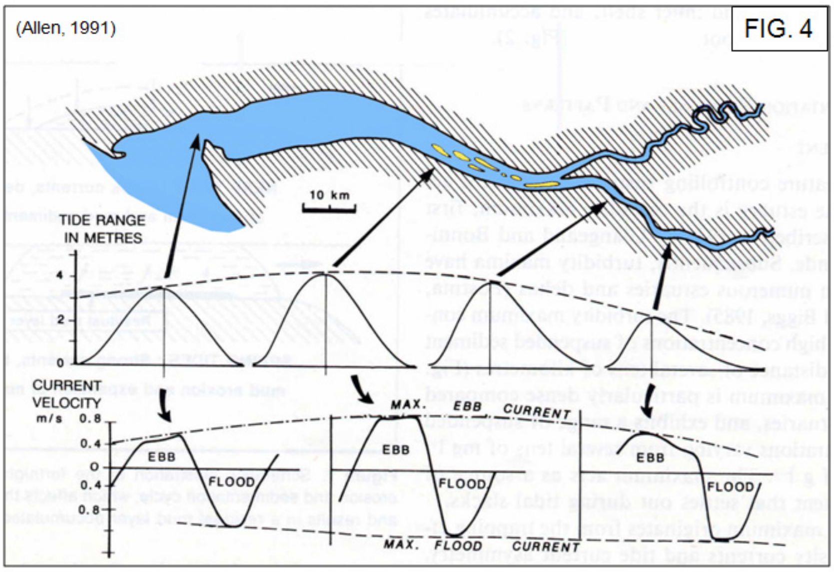
Hydrology

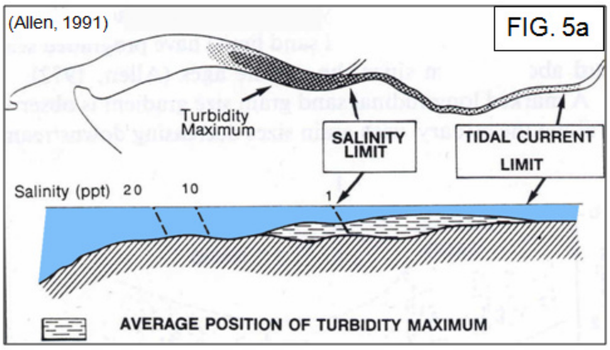
Hydrological studies of the Gironde (Allen, 1972; Castaing and Allen, 1981) have shown that tides in the Gironde are semi-diurnal, with amplitudes ranging from 2.5 m during neap tides to more than 5 m during spring tides (Fig. 4). During low river discharge, tidal current reversal occurs up to 130 km from the mouth, but during river floods tidal current reversal occurs only as far as 100 – 115 km upstream . The salinity intrusion within the estuary varies seasonally, extending landward up to 80-100 km during low fluvial discharge, and 40-50 km during high fluvial discharge. Consequently, the upstream limit of tidal currents is always a considerable distance landward of the limit of salt water (Figs. 5a, 5b).

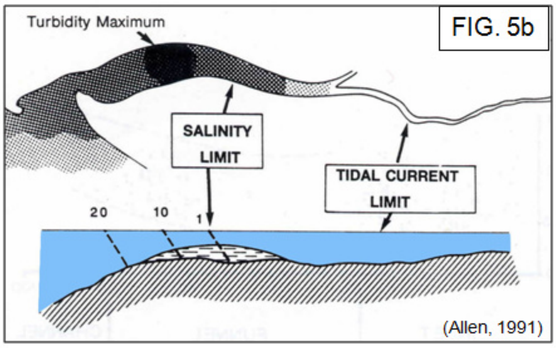
Fig.4 :
Landward evolution of tides and tidal currents within estuary funnel and estuarine channels (average tide and low water discharge)

Fig. 5a:
Location of the turbidity maximum and salinity intrusion during the low river discharge period (summer, fall)

Fig. 5b:
Location of the turbidity maximum and salinity intrusion during the high river discharge period (winter, spring)







Sedimentary patterns

A large volume of fluvial sand and mud is supplied to the estuary by the Garonne and Dordogne rivers**.** At present however, the entire fluvial sand supply is deposited within the estuary, and no fluvial sand reaches the estuary mouth or adjacent oceanic coastline (Allen, 1972). Studies by Castaing and Allen (1981) indicate that approximately 75% of the fluvial suspended silt and clay is deposited within the estuary while the remainder is evacuated out to the shelf where it is accumulating seaward of the 50 m isobath.

Sediments within the estuary form a distinctive longitudinal facies pattern (Allen, 1991), with coarse sand and gravel point bars in the rivers landward of the bayline, tide-dominated sand and mud point bars in the upper estuary channels and elongate tidal sand bars and estuarine mud in the mid-estuary funnel (Fig. 3). The estuary mouth contains coarse sand and gravel that is supplied by wave erosion of the adjacent oceanic coast and tidal current reworking of Pleistocene fluvial terraces in the estuary mouth. These coarse sediments form a flood tidal delta and tidal inlet fill system, and are in contrast with the muddy estuarine deposits further landward within the estuary.

Two facies transitions observed in the present estuary are of particular importance within the incised valley fill (Fig. 3): (1) the fluvial gravel to tidal estuarine sand and mud transition at the landward limit of the estuary, and (2) the mud to sand transition at the estuary mouth (Allen, 1991).

The fluvial-tidal facies transition at the landward limit of the estuary occurs in the vicinity of the bayline, i.e., c. 115 km landward of the estuary mouth. Although tidal currents during low river discharge can extend landward of the bayline, this limit marks the zone where fluvial flood currents are sufficiently damped out by tides so that tidal mud deposited during periods of low river discharge can be preserved within the channel sands.

The mud to sand transition at the estuary mouth marks the point of convergence of the seaward prograding estuarine muds and the transgressive landward-migrating estuary mouth sands.

Sequence stratigraphy of the Gironde estuary: a tide-dominated incised valley

The sequence stratigraphic model of the Gironde incised valley (Fig. 6) is based on onshore data (Allen and Posamentier; 1993, 1994) and offshore data (Lericolais, Berné and Féniès, 2001) and is characterised by 3 key-points:

1- The valley-fill contains a single sequence which corresponds to a fifth-order cycle (Vail et al., 1991) and is composed of three systems tract (LST, TST, HST).

The basal Sequence Boundary (SB) was initiated by the step-wise Würm glacio-eustatic fall between 120,000 and 18,000 BP and most of the valley-fill was deposited between 18,000 years BP and the present.

The Lowstand (LST) is constituted of a one-channel thick deposit of fluvial gravel and coarse sand. It is poorly developed in the proximal part of the incised valley and pinches out in its distal part, no fluvial aggradation has been observed. In the landward part of the valley, it is bounded by a sequence boundary (SB) generated by fluvial erosion during the Lowstand time. In the seaward part of the valley, this sequence boundary (SB) is eroded by the Tidal Ravinement Surface (TRS; Allen, 1991) described below.

The Transgressive Systems Tract (TST) constitutes the bulk of the incised valley-fill. It is constituted by estuarine tidal sand and mud sediments and comprises 2 facies associations: 1- heterolithic estuarine point bars and tidal flats, located in the inner estuary, 2- tidal inlet channel-fill coarse-grained sands, located in the estuary mouth. These estuarine sediments onlap onto the fluvial deposits of the Lowstand systems tract.

The basal surface of the Transgressive systems tract could be a non-erosive flooding surface (FS), or an highly-erosive Tidal ravinement surface (TRS):

In the seaward part of the incised valley located on the shelf, the TST strata are composed of landward-prograding estuarine clinoforms bounded at their base by a non-erosive flooding surface. These TST strata are not bounded by a basal tidal ravinement surface. This is attributed to the fact that, when the small and Lowstand anastomosed fluvial channels were flooded during the transgression, the oceanic tidal range was not amplified by a funnelling effect within the incised valley, and consequently the tidal-ravinement process did not occurred. Oppositely, in the landward part of the incised valley, the TST strata are exclusively composed of landward-migrating tidal inlet deposits, bounded at their base by a highly-erosive tidal ravinement surface. Such deposits were generated when the deeper and larger Lowstand channels have been flooded during the transgression. Here, the oceanic tidal range was locally amplified by the spit-constricted estuary mouth and the tidal-ravinement process started. The landward retreat of the estuary mouth within the incised-valley thalweg has partially eroded the underlying Lowstand strata. The morphology of the Lowstand fluvial channels seems to have controlled the sedimentary processes and therefore the internal architecture of the Transgressive Systems Track (TST) deposits.

The upper surface of the Transgressive systems tract (TST) could be an erosive Wave ravinement surface (WRS) or a non-erosive Maximum flooding surface (WRS):

- The seaward part of the TST is capped by another erosive surface: the *wave-ravinement surface* (WRS), which develops in response to the passage of the shoreline across the estuarine deposits. Wave erosional processes associated with the retreat of the shoreface result in removal of several meters of sediment (Swift et al., 1972).

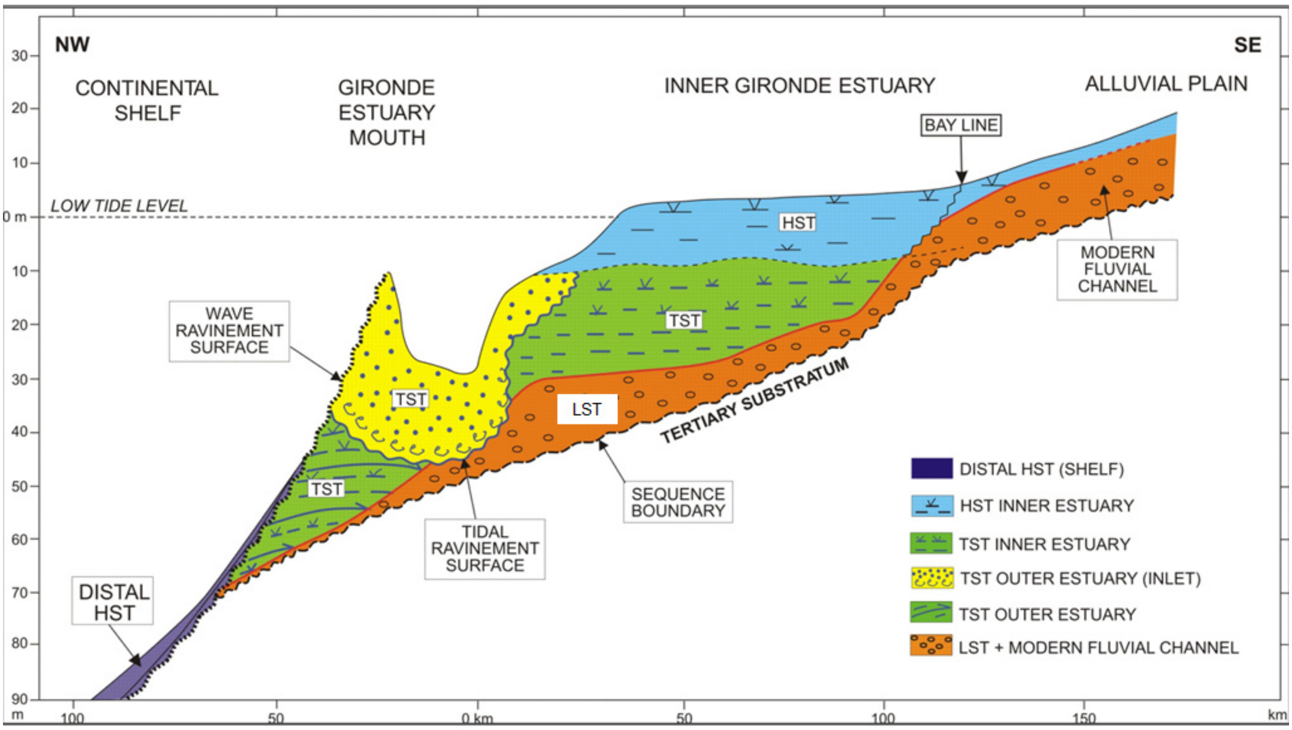
- The landward part of the TST is capped by the Maximum Flooding Surface (MFS), which is a mud layer that was deposited in the estuary funnel before it was filled by the Highstand Systems Tract (HST).

The Highstand Systems Track (HST) strata are constituted of a seaward-marching estuarine bay-head delta, located in the inner part of the estuary funnel, which progrades seaward and downlaps onto the underlying Maximum Flooding Surface. The characteristic facies association is composed of heterolithic tidal bars and muddy tidal channel fills. Eventually, the bay-head delta sediments will infill the entire estuary funnel and prograde offshore onto the shelf generating a wave- and tide-dominated delta.

Internal architecture of a wave- and tide-dominated incised valley-fill : the Gironde IVF FIG. 6

* **Three key-surfaces of erosion are observed within the valley fill :**
1- Fluvial Sequence Boundary (SB), 2- Tidal Ravinement Surface (TRS), 3- Wave Ravinement Surface (WRS),

* **Most of the Incised valley is filled by the Transgressive Systems Tract (TST), the Lowstand and the Highstand Systems Tracts are present (LST, HST)**



(Lericolais, Berné, Féniès, 2001)

2- The Gironde incised valley tapers on the shelf, 50 kms seaward of the estuary mouth (Lericolais, Berné and Féniès, 2001).

The fluvial incision (SB) located at the base of the Lowstand Systems Tract (LST) decreases gradually seaward and is eventually truncated by the Wave Ravinement Surface (WRS) around - 70 m below the present-day mean sea level (Figs. 7a, 7b). The decreasing depth of incision of the sequence boundary (SB) is interpreted to be due to a change of morphology of the fluvial Lowstand channels : large and deep sinuous channels located upstream change downstream to smaller and shallower anastomosed channels.

Similar observations have been made for the English Channel paleoriver (Lericolais et al., 1996), which supports the idea that rivers do not always generate continuous cross-shelf incised-valleys during the phase of sea-level drop. When sea-level drops too rapidly for fluvial sedimentation to keep pace with it, Lowstand deposition occurs as prograding anastomosed fluvial channels which eventually died out in the proximal part of the shelf. Similar observations have also been reported by Van Wagoner (1995) on the Cretaceous strata of the Bookcliff formation (USA). The result of such a process is that the Lowstand incised-valley deposits are disconnected from the Lowstand deposits located seaward on the shelf break.

3- The Gironde incised valley has a sand ribbon-shaped narrow and elongated geometry, due to a specific process occurring at the estuary mouth: the "Anchored tidal ravinement process" (Féniès, Lericolais and Posamentier, 2010).

The inlet of the estuary, bounded by the tidal ravinement surface, has not been shifted laterally from the North to the South by the littoral drift, because the valley is incised in a very resistive Eocene carbonate bedrock (named: "Plateau de Cordouan"). The inlet is anchored in the bedrock (Fig. 7a).

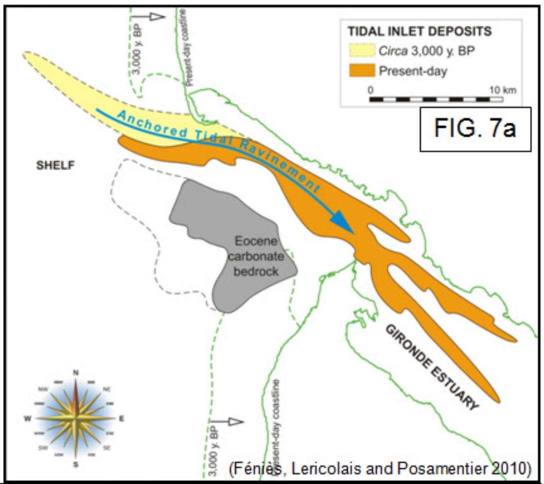
Consequently the reservoir geometry of the valley-fill has a sand-ribbon shape (Fig. 7b) : length = 180 km, width = 10 km, thickness = 40 m.

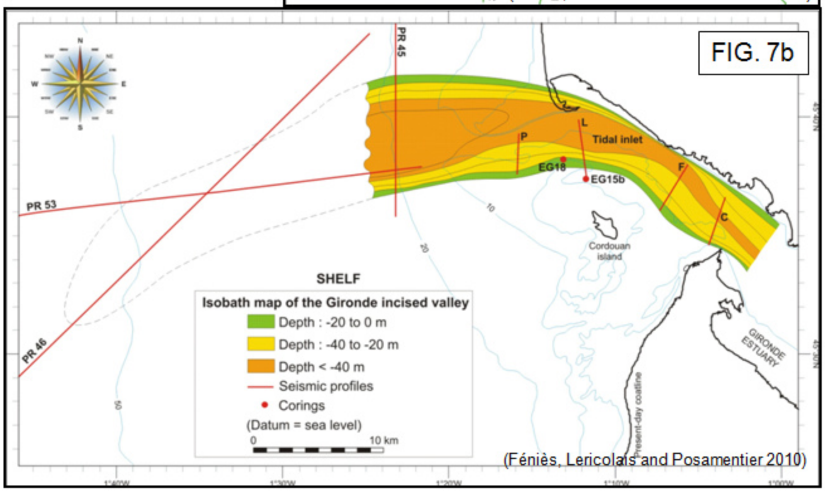
IMPACT OF TIDAL RAVINEMENT PROCESSES
ON WAVE- & TIDE-DOMINATED INCISED VALLEY GEOMETRY
(Féniès, Lericolais and Posamentier, 2010)

EXAMPLE OF THE GIRONDE INCISED VALLEY

PROCESS = Anchored tidal ravinement (Fig. 7a)
The inlet of the Gironde valley is anchored within the resistive Eocene carbonate bedrock and cannot be shifted to the South by the littoral drift

PRODUCT = Narrow Ribbon-shaped incised valley geometry (Fig. 7b):
Valley dimensions: length = 180 km, width = 10 km, thickness = 40 m





2- FIELD SESSION IN THE GIRONDE ESTUARY BAY-HEAD DELTA

Tidal bar complex of the estuary funnel, field stop : Plassac tidal bar

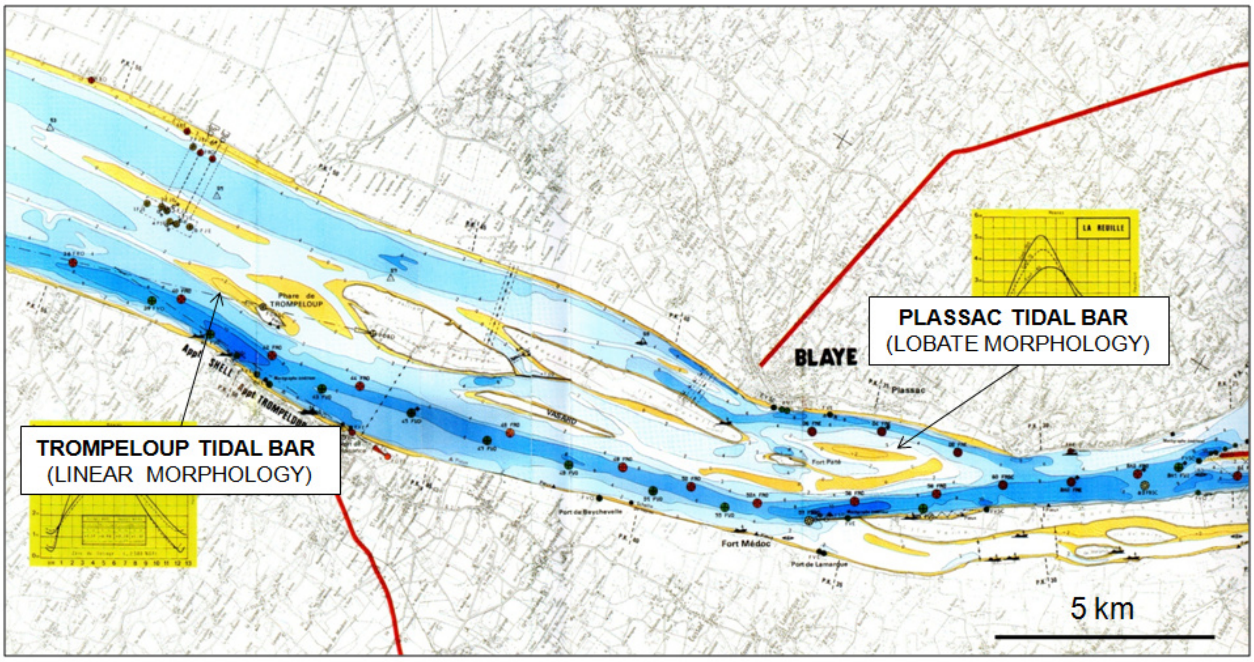
Location and historical evolution

Tidal bars are well developed in the landward part of the estuary funnel (Fig. 8), in a 30 km long bay-head delta (Allen, 1991), located from the junction of the 2 rivers (PK 25) to the mid-estuary funnel (PK 55). These consist of fluvial sands brought to the estuary by alluvial floods and reworked by tidal currents. These bars quickly prograde seaward in the estuary funnel, in a water depth of around – 10 m. When the sand builds up into the upper part of the intertidal zone, the bars are capped by mud, which is then colonised by vegetation to form permanent islands.

FIG. 8

Gironde estuary bay-head delta, exhibiting:

1- Tidal channels, 2- Tidal bars (lobate and linear morphologies), 3- Supratidal islands (tidal bars capped by marshes)



Morphology

Two key-examples illustrate the different tidal bar morphologies in the bay-head delta.

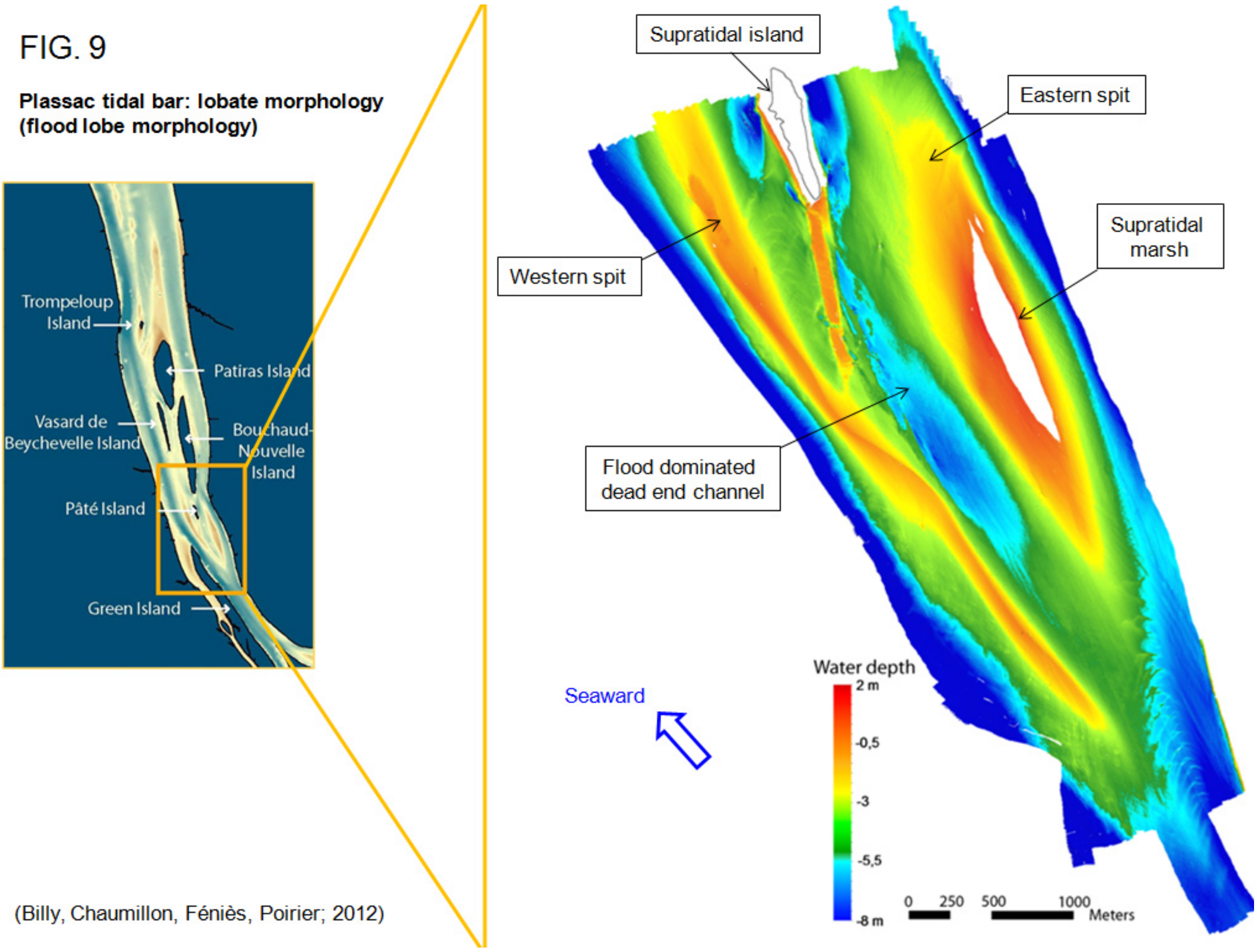
1- The Plassac sand bar, which will be visited, is a good example of a lobate morphology (Fig. 9). It is 4.6 km long, 1.4 km wide and culminates 2.5 m above the lowest water level of the tide. It is capped by a small supratidal island. This lobate morphology is stable though time, showing no radical morphological evolution through last century (Billy, Chaumillon, Féniès, Poirier, 2012).

2- The Trompeloup sand bar is a good example of the linear morphology (Fig. 10). It is 8 km long, 400 m wide and culminates 3 m above the lowest water level of the tide, a small supratidal island covers its upstream part (Féniès and Tastet, 1998). The linear morphology is characterised by major morphological changes: during active phases of seaward progradation the northern extremity of the bar was migrating seaward at a rate of 400 m/year.

In the bay-head delta of the Gironde estuary, only one tidal bar has the lobate morphology (Plassac), all the other bars have the linear morphology (e.g.: Trompeloup).

FIG. 9

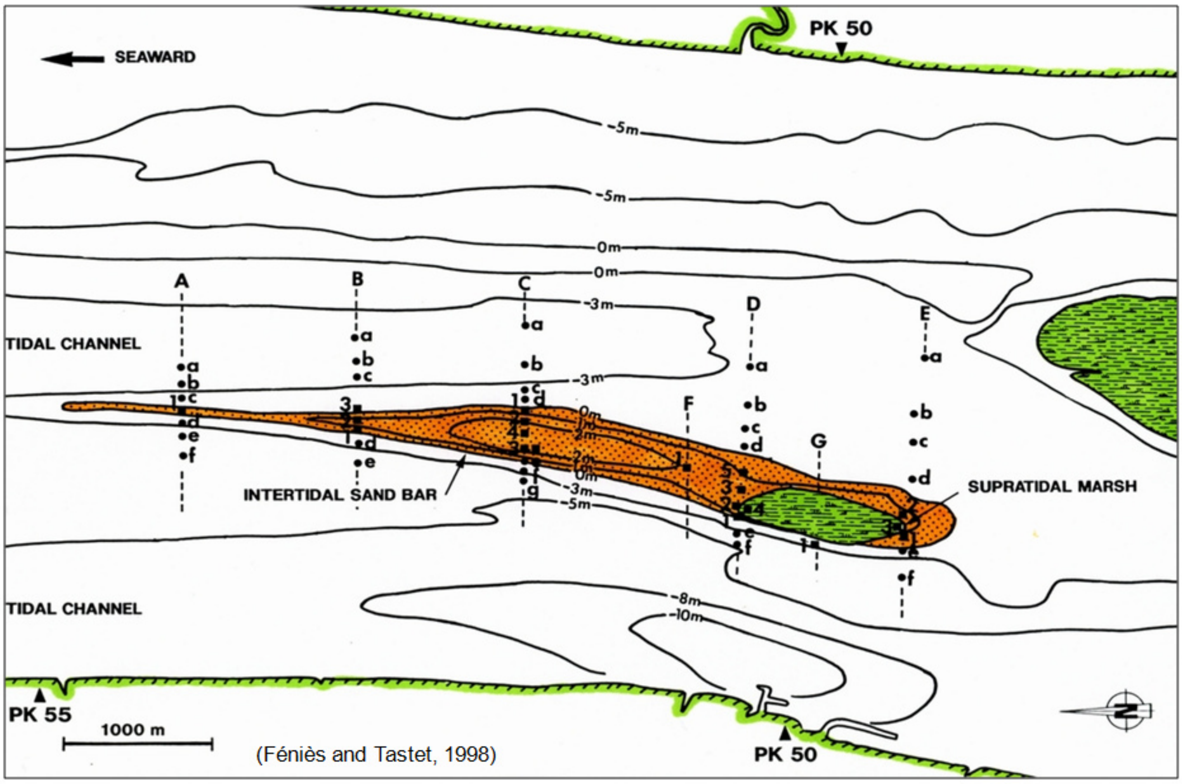
Plassac tidal bar: lobate morphology (flood lobe morphology)



(Billy, Chaumillon, Féniès, Poirier; 2012)

Trompeloup tidal bar: linear morphology

FIG. 10



Hydrology

The estuary is a macrotidal environment (Allen, 1991), in the bay-head delta the average tidal amplitude is 4 m, and varies between 2,7m (neap tides) and 5,3m (spring tides).

Current velocity only varies with tide amplitude, the estuary is essentially tide-dominated, fluvial currents play a negligible role. Tidal currents show a pronounced time-velocity asymmetry typical of estuarine environments. The dominant current could be the ebb or the flood depending upon the local morphology of the tidal bar / channel.

On the surface of the Plassac tidal bar in the intertidal zone, during the spring tides, both tidal current, ebb and flood, last equally the same time (around 3h), there is no time asymmetry, but there is an important velocity asymmetry: ebb reaches 80 cm/s whereas flood 52 cm/s (Féniès, de Resseguier and Tastet, 1999).

In the subtidal channels bounding the tidal bars, there is an important lateral tidal current segregation : the ebb and the flood flow along different paths, the ebb current predominates in the deep tidal channels located on both sides of the bar and the flood current predominates in the dead-end channel located inner part of the lobe (Fig.9).

Tidal bar sedimentary facies

The Plassac and Trompeloup tidal bars are covered by a network of 2D and 3D dunes generated by the tidal currents. During each phase of the dominant tidal current (ebb or flood) the crest of the dunes migrates down current and a package of sand stratifications is deposited. This sand package is named the tidal bundle, individual tidal bundles have a sigmoidal shape (Mutti et al., 1985). The individual bundle thickness vary from 20 cm during the neap tides to 110 cm during the spring tides (Féniès and Tastet, 1998).
During each phase of the subordinate tidal current, a reactivation surface erodes a part of the tidal bundle generated by the dominant tidal current and a bed of subordinate current ripples is formed. During the slack water periods the velocity of the tidal currents is nill, mud particles in suspension settle down and a mud drape is deposited over these bedforms on the surface of the bar. In the intertidal zone, the high-tide slack-water mud drape is deposited when the bar is submerged and drapes the entire surface of the bar. The low-tide slack-water mud drape is deposited in the pools of turbid water located in the troughs of the megaripples. The preservation of both high-tide and low-tide slack-water mud drapes generates an intertidal mud drape couplet that is a diagnostic feature of the tidal deposits (Féniès et al., 1999).

Facies sequence of a linear tidal bar (Trompeloup tidal bar)

51 cores, 4 to 6 m long (Fig. 10), have been cut in the subtidal, intertidal and supratidal parts of the Trompeloup tidal bar (Féniès and Tastet, 1998). They indicate that the facies sequence formed by the seaward progradation of the tidal bar is around 10 m thick (Fig. 11). Within this facies sequence, the thickness of sand bodies is up to 6 m.

In this facies sequence, two upward-coarsening sand bodies up to 3 to 5 m thick are isolated by a shaly layer 1 to 2 m thick made of sand-shale tidal alternations. Each sand body exhibits tidal cross stratified sands, i.e. consisting of bundles separated by clay drapes and reactivation surfaces. In most cases this cross stratification is unidirectional, however bidirectional structures can also be observed i.e. juxtaposed ebb/flood bundles. Each sand body is made up of well-sorted, medium-grained sand (0.3 mm). The muddy unit which separates the sandy units is made of mud and ripple laminated sand beds which exhibits the classical tidal bedding.

These two sand bodies were deposited during 2 distinct phases of bar progradation and are separated by a phase of bar abandonment (Figs. 11, 12). Analysis of detailed bathymetric maps since 1940 has enable to reconstruct the formation of the tidal bar vertical facies sequence.
1- The basal upward-coarsening sand body was an active seaward-prograding tidal sand bar between 1940 to 1960.
2- From 1960 to 1972, the bar was abandoned and the muddy layer was deposited over the tidal sand bar. During this phase of abandonment, sand accumulated further upstream in a small-size flood lobe. The mud deposits became bioturbated and no major sand influx occurred until 1975.

3- From 1975 to 1985, another sand bar rapidly prograded seaward over the muddy layer and again formed an elongate sand ridge morphology. This progradation generated the upper sand body.

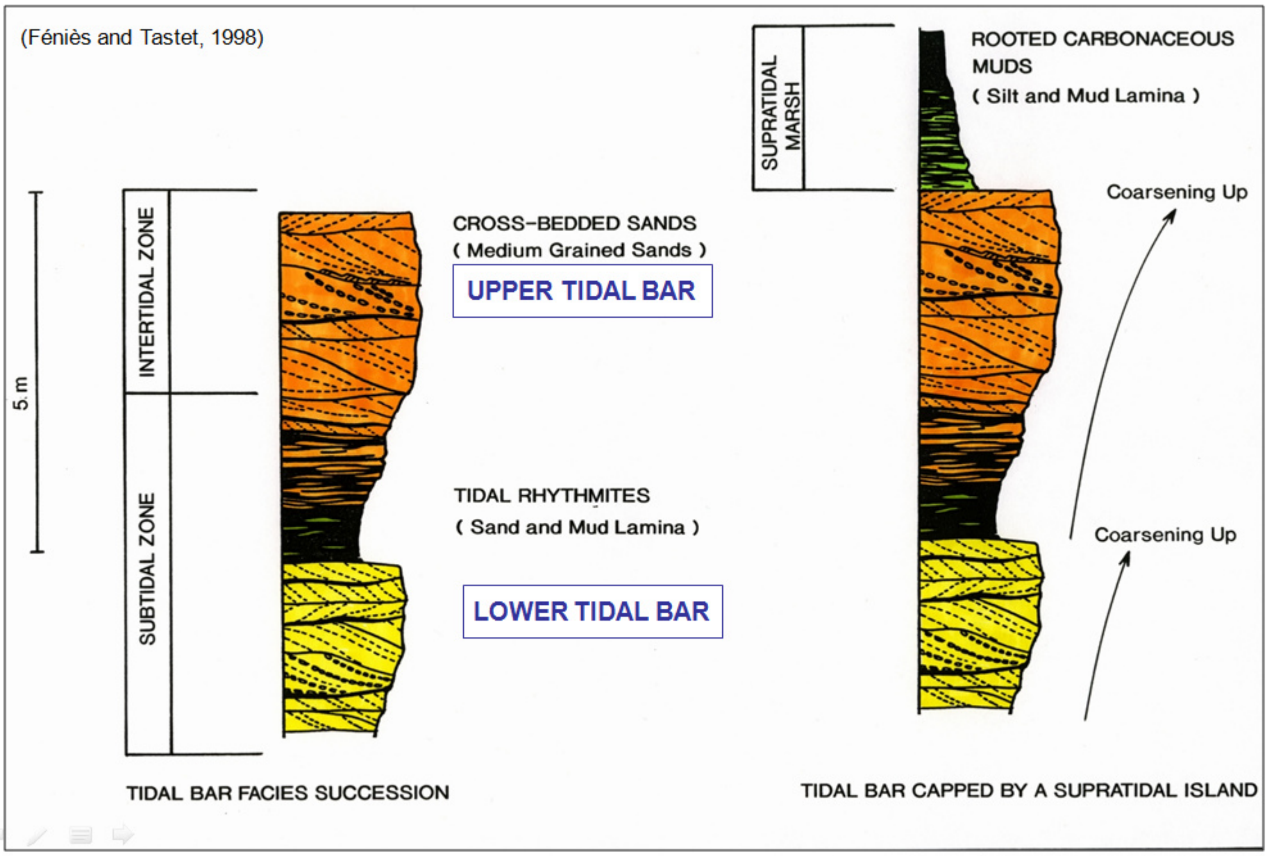
The 2 upward-coarsening sand bodies are related to a specific bar morphology, i.e. elongated sand ridges that quickly prograde seaward during periods of strong fluvial sand influx. The shaly layer characterise a period of bar abandonment, which occurred during a period of weak fluvial sand influx.

These processes of bar progradation and abandonment occur in a time frame of 10 to 20 years. The tidal bars are made of fluvial sands introduced by the fluvial floods within the estuary. The periods of active seaward progradation of the tidal bars may be climatically controlled, i.e., the periods of strong fluvial sand influx within the estuary may occur during the years of heavy rain fall and the periods of bar abandonment, caused by a weak fluvial sand influx, may occur during years of moderate to poor rain fall.

When the sandy tidal bar aggrades to the top of the intertidal zone, it is capped by mud flat and marsh which form a supratidal island in the estuary. These islands are comprised of 2 to 3 m of silt and shale alternations covered by salt marsh vegetation. Root patterns are very well developed in these clayey sediments. These supratidal islands represent the ultimate stable stage of tidal bar progradation.

FIG. 11

Vertical sequence of facies generated by the migration of a tidal bar (based on Trompeloup tidal bar)



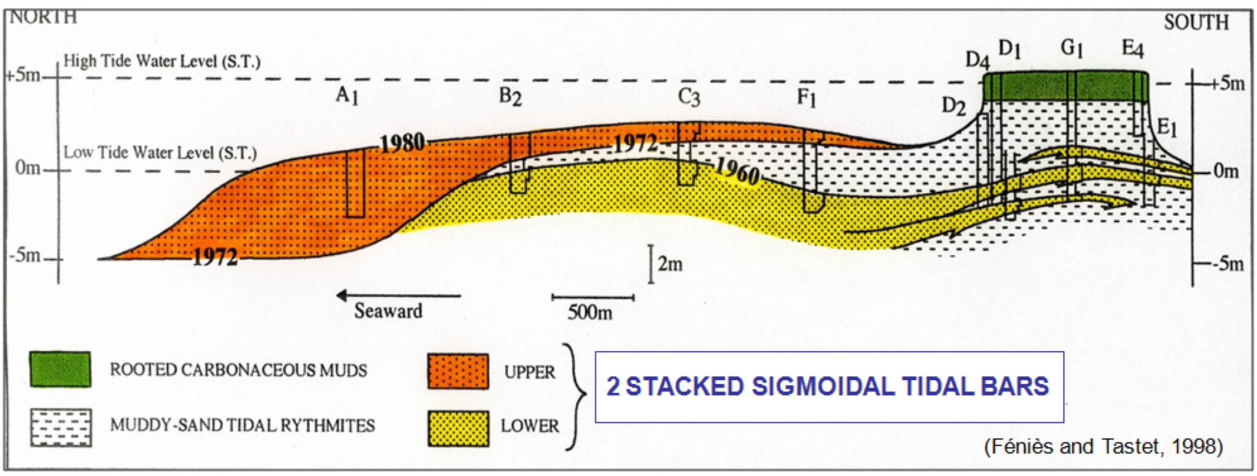
Reservoir architecture of a linear tidal bar (Trompeloup)

Based on numerous cores and high resolution seismic data, a number of cross-sections have been established (Fig. 12) through the preceding tidal bar system illustrating its internal architecture (Féniès and Tastet, 1998).

As described above, the tidal bar is made of 2 sand bodies partially separated by a shaly-layer. Each of these sand bodies was deposited by the seaward progradation of an elongated linear sand bar. These elongated bar form large sigmoïdal sand ridges up to 8 km long, 400 m wide, and 2 to 6 m thick. The interconnectivity of these sand bodies may be restricted by the shaly layer which might act as a permeability baffle.

FIG. 12

Longitudinal cross section trough the Trompeloup tidal bar exhibiting the juxtaposition of 2 sigmoidal tidal bars



JULY 30 Th. 2012: FIELD SESSIONS IN THE ARCACHON LAGOON

1- CHARACTERISTICS OF ARCACHON/ LEYRE INCISED VALLEY

The Arcachon lagoon sediments are infilling the upper part of the incised valley-fill of the Leyre estuary. The incised valley of the Leyre was cut during the last period sea level fall (from 120,000 y. BP, to 18, 000 y. BP) and filled during the last period of sea level rise (from 18,000 y. BP, to present day). The present-day lagoonal morphology was progressively acquired since 3,000 years BP, when the rate of the eustatic rise slowed down and the sea level was practically stable at its present day level (Féniès, 1984; Faugères, Cuignon, Féniès, Gayet, 1986; Féniès, Lericolais and Posamentier, 2010).

Arcachon lagoon : morphology, hydrology and sedimentary processes

Morphology

The Arcachon Lagoon is located on the oceanic coast of the Bay of Biscay to the south west of Bordeaux (45°North ; 1° West). It is a large triangular lag oon, 15 Km on a side, 160 Km² in area, protected from the erosive action of ocean waves by a 20 Km long littoral barrier spit named : "le Cap Ferret" (Fig. 13). Behind the barrier, a wide area of sandy and muddy tidal flats and marshes have developed, cut by an extensive network of tidal channels. These tidal channels form a seaward contributive network which merges into a deep (25 m) tidal inlet at the mouth of the lagoon. At low tide, intertidal flats cover about 75 % of the surface of the lagoon and tidal channels about 25 %. During high tide, only a small marsh remains exposed in the central part of the lagoon : "l'Ile aux Oiseaux" (covering 1 Km²).

Tidal flats exhibit the classical facies pattern described by Reineck and Singh (1980)

The tidal flats are predominantly muddy in the inner part of the lagoon and are entirely covered by marine grass: "Zostera Noltii". Sandy tidal flats predominate in the outer lagoon where they are only sparsely covered by "Zostera Noltii".

Tidal channels exhibit 2 types of morphological patterns:

In the inner lagoon, channels are highly meandering, narrow (50 to 120 m) and deep (3 to 10 m). In the convex part of the meanders, sandy side bars accumulate up to 200 m long and 60 m wide.

In the outer lagoon, tidal channels converge and coalesce to and form slightly less sinuous, wider (up to 1.5 Km) and deeper (up to 15 m) channels. Large scale side bars (2 Km long, 300 m wide) are deposited in the convex part of the meanders.

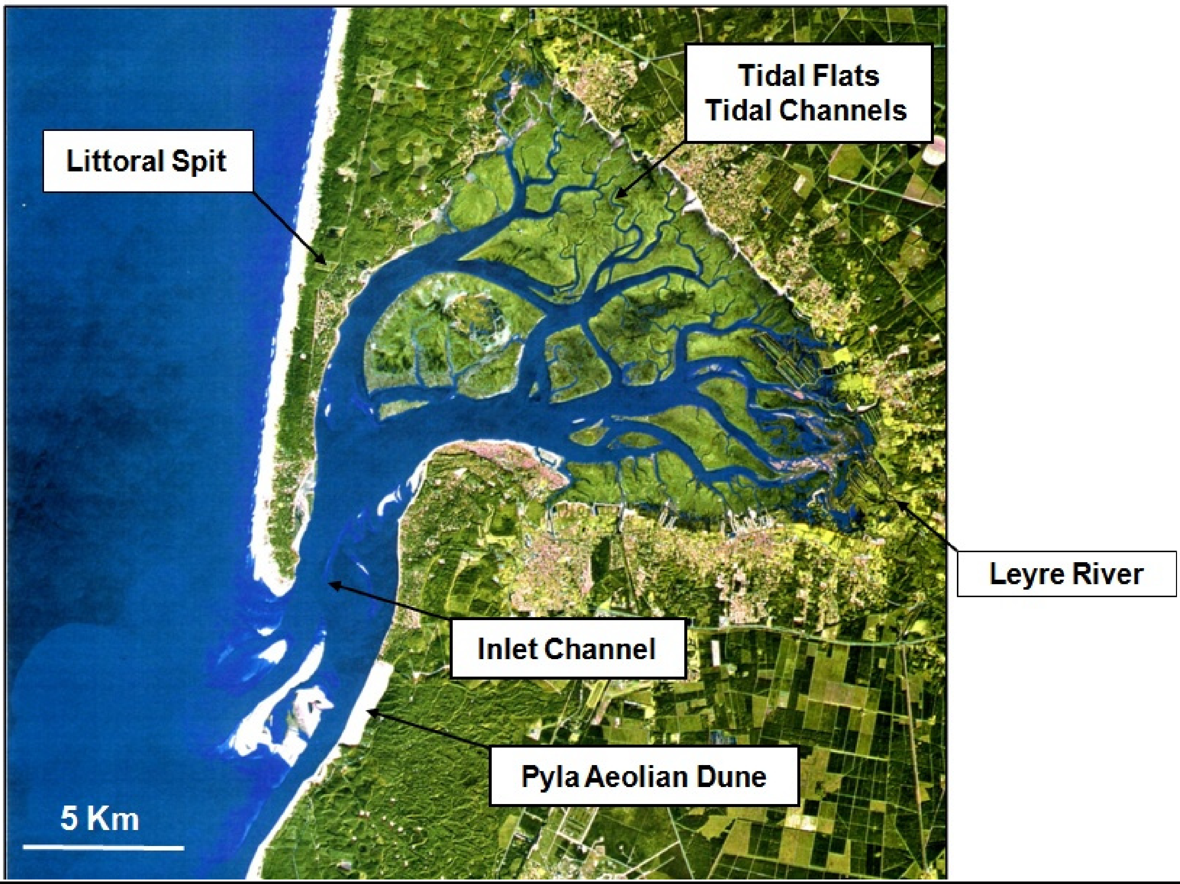
Seaward, at the mouth of the larger tidal channels, lobate tidal sand ridges (Caston, 1972) form alternating ebb- and flood-dominant lobes up to 600 m wide.

In the area adjacent to the tidal inlet, flood-dominant lobes amalgamate and form a flood tidal delta complex. The tidal inlet is 12 km long, 3 km wide and - 25 m deep. A large-size ebb tidal delta is present at its seaward extremity and progrades onto the shoreface in a water depth up to - 20 m.

Morphology of the Arcachon lagoon at low tide:

a dense tidal channel network incise a muddy tidal flat covered by marine grass

FIG. 13



Hydrology and Sedimentary Processes

Tidal amplitudes in the lagoon range from 2.1 m (neap tides) to 4.9 m (spring tides) which places it in the mesotidal range. Tidal currents are the dominant sediment transport mechanism in the tidal channels. Wave action is restricted to the very outer part of the inlet and fluvial currents only affect the mouth of the Leyre river in the SE corner of the lagoon (Féniès, 1984).

Tidal channels mainly comprise sand which is transported by tidal currents and a variety of tidal sedimentary structures have been observed in these channels (Féniès, 1984 ; Féniès and Faugères, 1998). Two scales of tidal cycles are responsible for the transport of the sand : the semi-diurnal ebb-flood cycle (12 hours 40 mins) and the fortnightly neap-spring cycle (14.7 days).

During the semi-diurnal cycle (12 h 40’), ebb and flood currents last 6 hours. The tidal slack-water periods (high-tide and low-tide slack-water) last around 20 mins each.
In the subtidal part of the channels, ebb is the dominant tidal current in terms of velocity. There is no time asymmetry, as ebb and flow last equally 6 hours.
In the intertidal part of the channels, ebb and flood maximum velocities are similar and both currents last equally the same time.

During the neap-spring cycle (14.7 days), the amplitude of tides and corresponding current velocities varies from a minimum during the neap tides, to a maximum during the spring tides.
In the subtidal area of a channel, the maximum current velocity is always above the threshold for dune migration.
In the intertidal area of a channel, e.g. the summit of the side bars, during 4 days of a fortnightly neap-spring cycle, tidal current velocity is less than the threshold for dune migration and only ripples are generated; during 10 days however, tidal current velocity exceeds this threshold and dunes are formed on the side bars.

Sequence stratigraphy of the incised valley of the Leyre estuary (Arcachon lagoon): a wave-dominated incised valley

The Leyre valley was discovered 10 kilometres West to the Arcachon lagoon on the continental shelf (Féniès and Lericolais, 2005) - Fig. 14-. This incised valley appears to be similar to the Gironde incised valley located on the continental shelf around 100 kilometres North.
Both incised valleys were formed within the same depositional wave- and tide-dominated environment and present the same stratigraphic framework, they are characterized by 4 key-points:

(1)-The Leyre and the Gironde river valleys have been cut during the last sea-level drop that took place during the Upper Pleistocene from 120,000 years B.P. to 18,000 years B.P.

During the following sea-level rise, from 18,000 years B.P. to 6,000 years B.P., both fluvial valleys were flooded and transformed into estuaries. When the sea-level reached a near still-stand, around 6,000 years B.P., the Gironde and the Leyre estuaries were partially filled.
The beginning of the formation of the present day Arcachon lagoon is dated around 3,000 years B.P. (Féniès, 1984). Since this period the littoral spit (“Cap Ferret”) has protected the inner lagoon from the erosive wave influence and the Arcachon lagoon has progressively acquired its present-day morphology (Féniès, 1984; Cuignon, 1984; Faugères, Cuignon, Féniès, Gayet, 1986). The North-South progradation of the "Cap Ferret" littoral spit caused a shift of the axis of the tidal channel network, from Northwest-Southeast to Northeast-Southwest direction.

(2)-The Leyre and the Gironde valleys have been formed in a wave- and tide-dominated environment and are characterized by a specific strata architecture.

The sequence stratigraphy analysis of the Leyre valley-fill (Figs.14) reveals that it is a fifth order depositional sequence (Posamentier and Vail, 1988) characterised by four original key points :

(a)- A complete erosion of the Lowstand systems track and of the fluvial sequence boundary by a tidal ravinement surface (Fig. 14).

During the last sea-level drop the river Leyre cut its valley into Pliocene sediments, along the structural axis of the Leyre fault. When the valley was flooded by the Holocene transgression, an intense tidal ravinement process (Allen, 1991) took place within the Leyre estuary and caused the complete erosion of the lowstand systems track and of the fluvial-generated sequence boundary. Consequently, the basal surface bounding the Leyre incised valley is a composite sequence boundary, of fluvial origin, that has been reworked by a tidal ravinement surface.

Under the Arcachon lagoon, this tidal ravinement surface has a funnel-shape morphology (Fig. 15b) deepening and widening towards the West (Féniès and Lericolais, 2005) . Below the spit, this surface is 20 km wide and -30 m to -35 m deep below the sea level. The lateral boundaries of the incised valley do not extend beyond the present day northern and southern coasts of the Arcachon lagoon.

On the continental shelf, ten kilometres West to the Arcachon lagoon, the tidal ravinement surface bounding the Leyre incised-valley shows a deep incision exhibiting a flat U-shaped morphology (Féniès and Lericolais, 2005). The depth of this incision is around - 60 m below sea level. The slope gradient of the sequence boundary, from the "Cap Ferret" littoral spit to the seismic line offshore is : 0.23 %. It is very similar to the slope gradient of the sequence boundary of the Gironde incised valley.

(b)- A thick Transgressive systems track which infill most of the valley (Fig. 14).

The Transgressive systems tract is composed of coarse-grained sands and gravels with marine shells debris. This sedimentary facies is typical of tidal channel deposits (Féniès, 1984; Féniès et Faugères, 1998). The seismic facies of this Transgressive systems track is characterised by a set of high frequency reflections, showing high impedance contrasts. These reflections are 5 m to 15 m high, dipping for the most part of them towards the West with an angle of around 4° and downlapping on the basal sequence boundar y. These reflections are interpreted to be the bedding generated by the lateral migration of inlet tidal channels, when these channels are shifted towards the South by the littoral drift.

The thickness of the Transgressive systems track deposits varies from 30 m to 40 m below the Cap Ferret littoral spit. Ten kilometres to the West of the littoral spit on the continental shelf, their thickness decreases to 20 m to 25 m because the upper part of the Transgressive systems tract is eroded by the wave ravinement surface (Swift et al., 1972).

(c) An important wave ravinement surface, which eroded the upper part of the Transgressive systems track (Fig. 14).

The wave ravinement surface erodes the upper 10 m of the Transgressive systems track. The interfluves of the incised valley (supratidal and intertidal deposits) are entirely eroded and only the deepest subtidal channels of the inlet are preserved.

(d)- A thinner and muddier Highstand systems tract (Fig. 14), composed of: (a) the tidal flats of the present day Arcachon lagoon, (b) the Bay-head delta of the Leyre estuary, located onshore in the South East corner of the Arcachon lagoon (Cuignon, 1984).

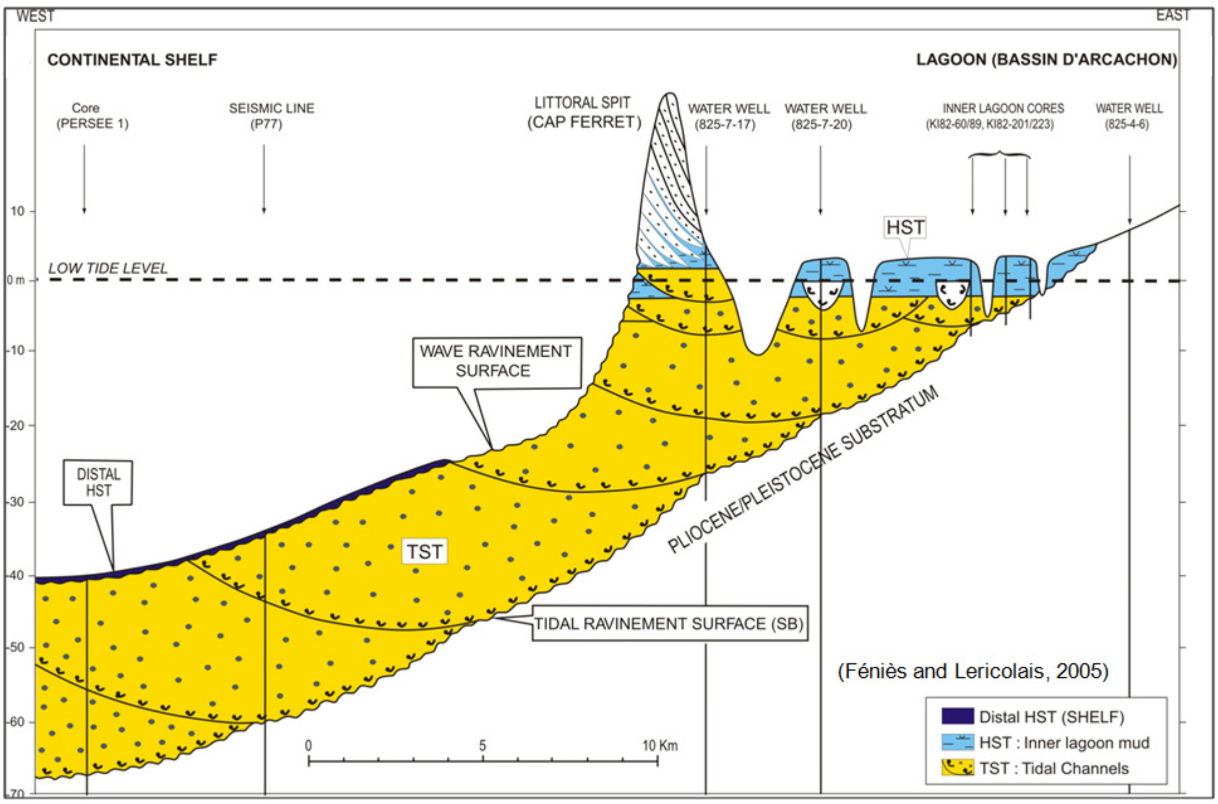
Consequently, the description of the internal architecture of the Leyre incised valley is very similar to the stratigraphic model of the Gironde incised valley. The main difference is the absence of the Lowstand systems track which was not preserved within the Leyre valley-fill, because the depth of fluvial incision generated during the sea-level fall is shallower than the depth of the tidal ravinement surface generated during the sea-level rise.

Internal architecture of a wave- and tide-dominated incised valley-fill : the Leyre IVF

FIG. 14

*** Two key-surfaces of erosion are observed within the valley fill :**
1- Tidal Ravinement Surface (TRS / SB), 2- Wave Ravinement Surface (WRS)

*** The Incised valley is entirely filled by the Transgressive Systems Tract (TST)**



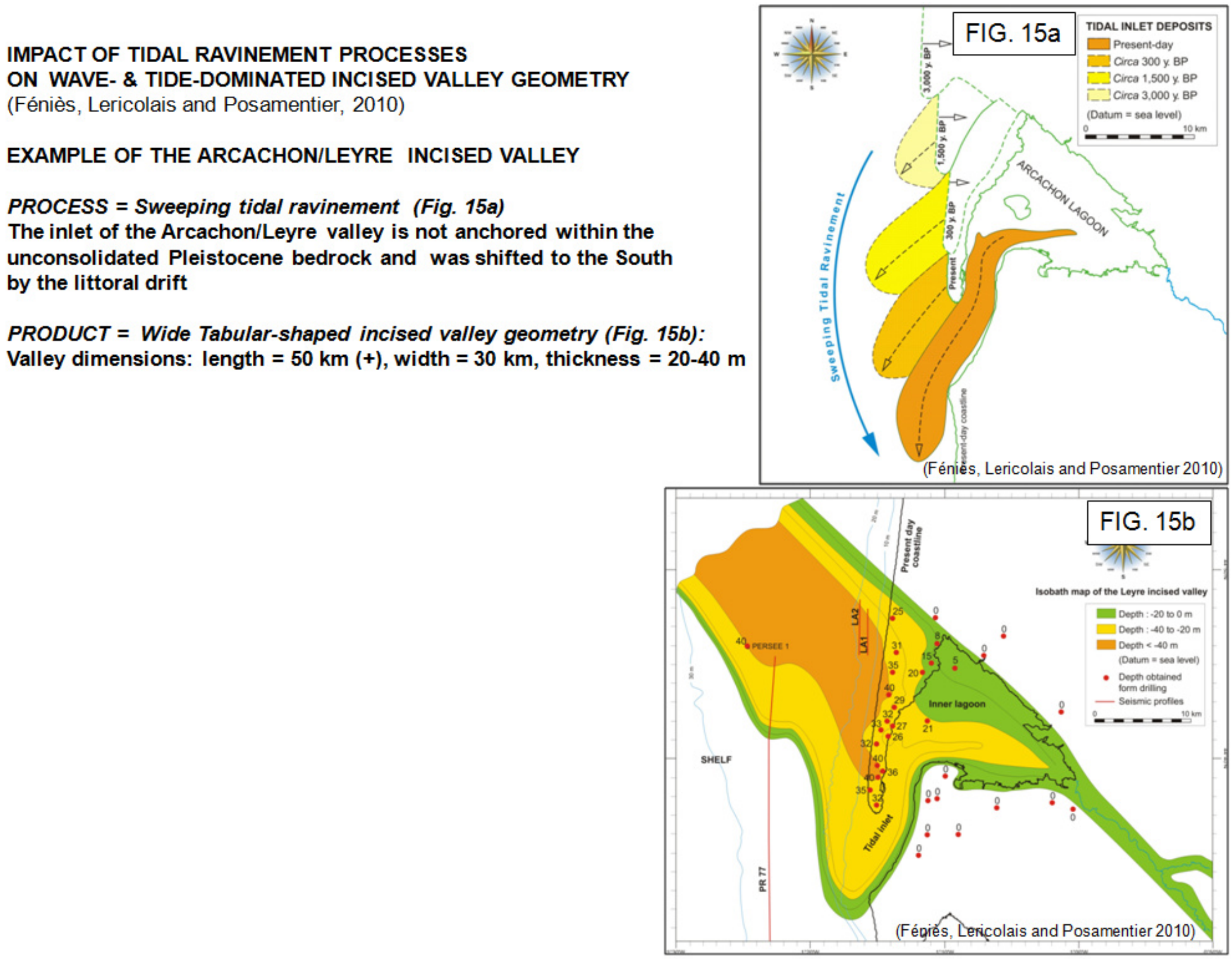
(3)-The Leyre incised valley has a tabular-shaped reservoir geometry (Figs. 15a, 15b), due to a specific process occurring at the estuary mouth: the "Sweeping tidal ravinement process" (Féniès, Lericolais and Posamentier, 2010).

The inlet of the estuary, bounded by the tidal ravinement surface, has been shifted laterally from the North to the South by the littoral drift over a distance of 30 km, because the valley is incised in non-consolidated

Pleistocene sediments (Fig. 15a). Oppositely to the Gironde estuary, the Leyre inlet is not anchored in the bedrock.

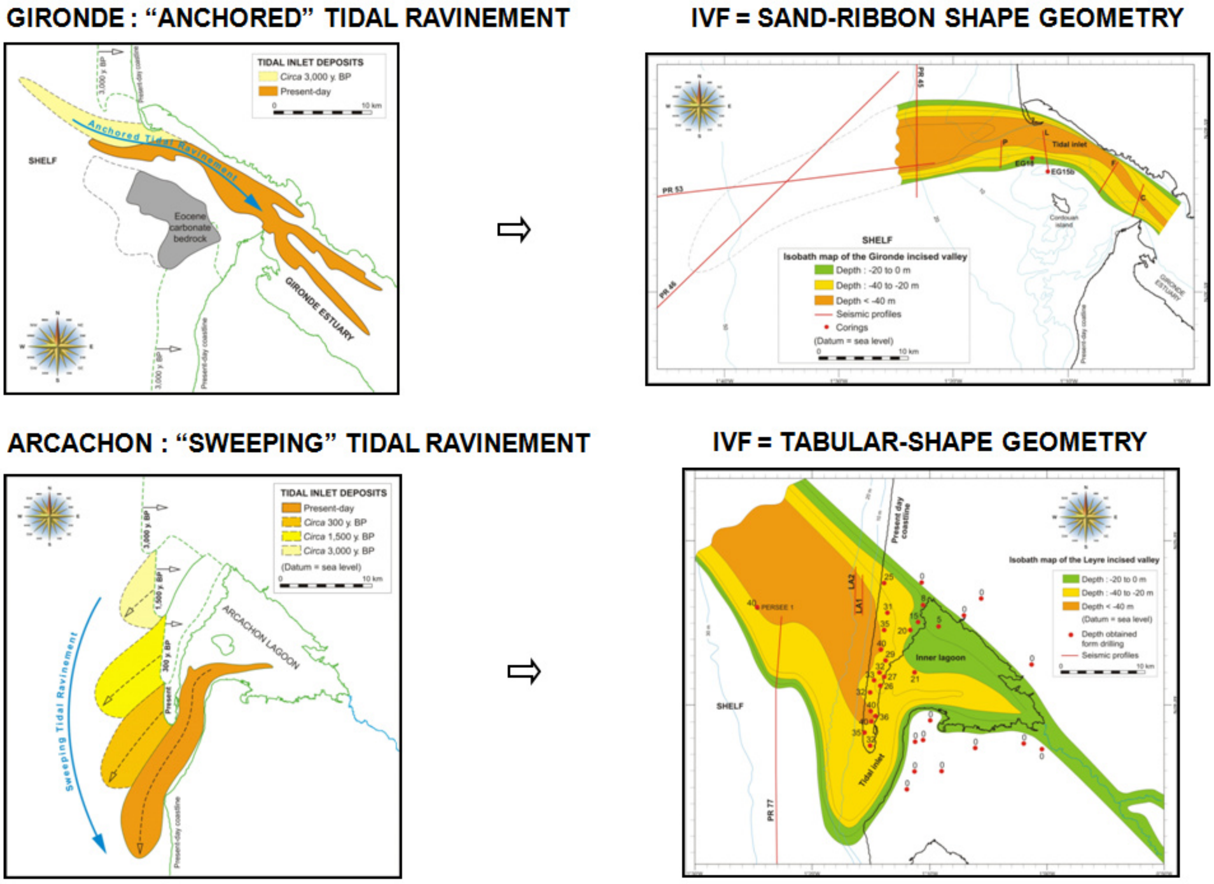
Consequently the reservoir geometry of the valley-fill has a wide tabular shape (Fig. 15b) :

length = 50 km (+), width = 30 km, thickness = 20m-40 m. The Leyre incised valley is 3 times wider than the Gironde incised valley.



(4)-The category of tidal ravinement process directly controls the reservoir geometry of wave- and tide-dominated incised valleys (Fig. 16): an Anchored tidal ravinement generates narrow ribbon-shaped reservoir geometry (Gironde IVF), whereas a Sweeping tidal ravinement generates wide tabular-shaped reservoir geometry (Féniès, Lericolais and Posamentier, 2010).

Comparison of the impact of tidal ravinement processes on wave- & tide-dominated incised valley geometry FIG. 16
(Féniès , Lericolais and Posamentier, 2010)



2- FIELD SESSIONS IN THE ARCACHON LAGOON

Tidal channels and Tidal inlet of the Arcachon lagoon

The meandering tidal channels of the inner Arcachon lagoon (Féniès and Faugères, 1998) will be examined during the first stop and the tidal inlet channel located at the seaward extremity of the lagoon will be examined during the second stop.

Sedimentary facies (Fig. 17)

The tidal channel sand is well-sorted, clean and medium-grained in the core of the side bar. In the thalweg of the channel (up to - 10 m deep), a thick channel lag overlies the erosive base of the channel. This lag is mainly constituted of pelecypods shells (a dozen of different species are present here) with a few mud pebbles, in a coarse sand matrix.

In the subtidal zone of the channel the sedimentary structures are ebb-dominated sinuous-crested dunes which cover the outer lobe of the side bars. The average cross-bed set thickness is 10-30 cm. The bedding is made of superimposed sets of unidirectional ebb-oriented trough cross beds, truncated by numerous reactivation surfaces.

In the intertidal zone of the channel, the side bar surface is covered by a network of 2D and 3D dunes which migrate during the tidal cycles and form sigmoidal bundles.
On the outer lobe of the bar, ebb is the dominant tidal current, the subsequent bedding is unidirectional, ebb-oriented. At the bar crest, ebb and flood currents reach very similar velocities, ebb and flood dunes actively migrate and generate bidirectional ebb and flood sigmoidal bundles bounded by reactivation surfaces.

A notable feature of these tidal channel sediments is the quasi total absence of the mud drapes and flasers. The only fine-grained sediments that are deposited during the slack water periods are silts and organic matter particles.

FIG. 17

Meandering tidal channels filled by sandy side bars (C8 to C2)

and

Adjacent muddy intertidal flats

(Inner Arcachon lagoon)

Tidal channel-fill sequence (Fig. 18)

More than 90 cores (4 to 7 m long) were taken in the intertidal and subtidal zones of the tidal channels, these have enabled a fairly detailed description of the facies sequences formed by a tidal channel-fill (Féniès and Faugères, 1998).

As illustrated in Fig.18, the tidal channel-fill succession comprises well-sorted, medium-grained sands which form a cylindrical to fining-upward profile. The coarser grain sizes (> 0.5 mm) are only present as a lag at the base of the channel-fill. The bulk of the channel fill is constituted by well-sorted, medium (0.3 mm) sand.

The erosive base of a channel-fill section is overlain by a coarse lag deposit up to 1 m thick, mainly comprised of pelecypod shell debris, a few mud pebbles and coarse sands.

A characteristic suite of sedimentary structures is also observed within the channel-fill succession. In the subtidal part of the channel-fill, unidirectional ebb-oriented, bundles truncated by flood reactivation surfaces predominate. At the top of the channel-fill, i.e. in the intertidal zone, ebb- and flood-oriented cross strata predominate. A characteristic feature of the tidal cross-stratification is the sigmoidal geometry of the bundles formed during the semi-diurnal tidal cycles and the abundance of reactivation surfaces formed by the reversal of the tidal currents.

When a channel is abandoned, these intertidal channel-top sands are capped by the sand/mud alternations of the intertidal flat. These tidal flat deposits exhibit all the "classical" tidal facies, e.g. : flaser bedding, wavy bedding, lenticular bedding, roots, worm burrows.

The total thickness of a tidal channel-fill succession ranges from 5 m to 15 m, comprising up to 10 m of subtidal fill and about 5 m of intertidal fill (tidal range).

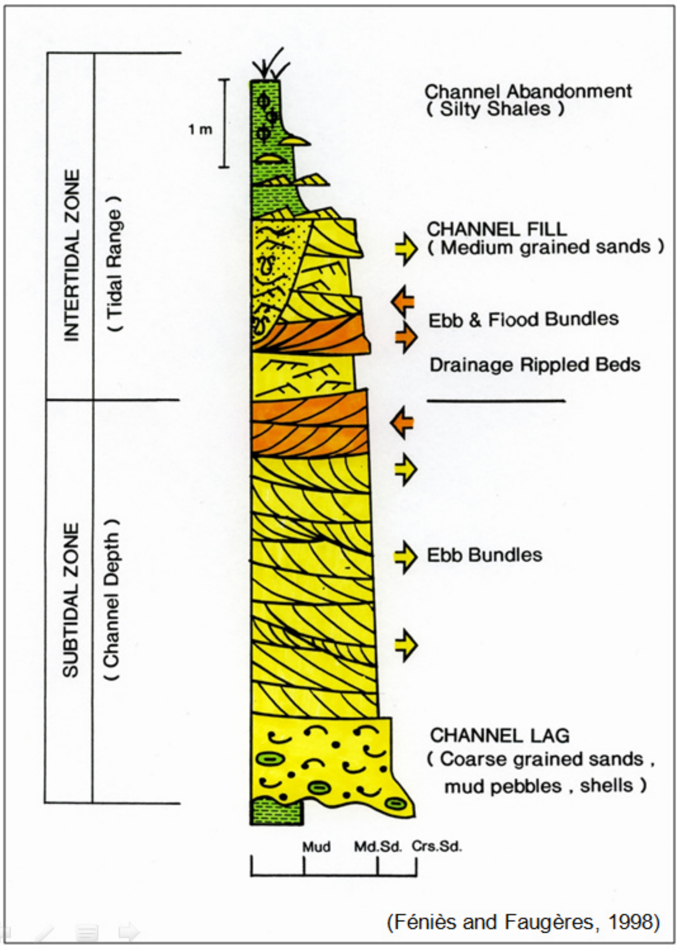


FIG. 18

Vertical sequence of facies generated by a tidal channel-fill

Geometry of the channel-fill deposits (Fig. 19)

The geometry of the deposits of a meandering tidal channel-fill sand has been described in the inner Arcachon lagoon. Thanks to numerous cores, an isopach map of the channel-fill sands have been established (Féniès and Faugères, 1998).

This map shows that the meandering morphology of the channel governs the pattern of the sand accumulation along the axis of the channel. Side bars are located in the convex part of the meanders. Each side bar forms an elliptical sand body with a thickness varying between 1 to 6 m ; a width between 40 to 60 m ; and a longitudinal extension between 150 to 200 m. These elliptical sand bodies coalesce longitudinally along the channel axis and form a sinuous sand ribbon. The geometry of this ribbon comprised a series of elliptical-shaped sand units joined by thinner sand ribbons. The overall dimensions of the channel fill are : 2 to 6 m thick, 150 to 200 m wide and a few km long.

The primary characteristic of this channel-fill sand ribbon is its limited lateral extent. The meandering morphology of the channel suggested a widespread meander belt. However, in spite of the high channel sinuosity, there is no evidence of any lateral channel migration. The lack of channel migration as shown by coring has also been documented by bathymetric maps since the 18 th century (Fig. 20) and aerial photographs since 1934. This stability of the tidal channels is a characteristic of all the Arcachon tidal channels that are not fed by river input.

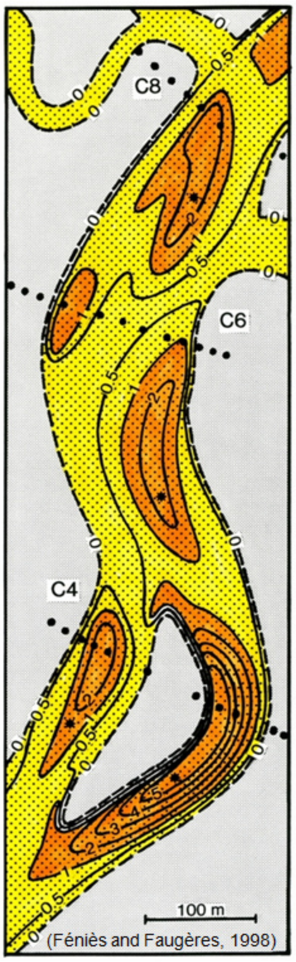


FIG. 19

Geometry of the tidal channel-fill deposits, note that :

- **the sand isopach has a narrow ribon shape,**
- **the side bar deposits generate elliptical sand bodies which coalesce all along the longitudinal channel axis.**

Two mechanisms could explain this lack of migration (Féniès and Faugères, 1998).

1- The adjacent tidal flats are made of 3 to 5 m of silty-muddy sediments that are covered by the marine grass : zostera. This rooted sediment is very cohesive and could prevents erosion and lateral migration of the channel.

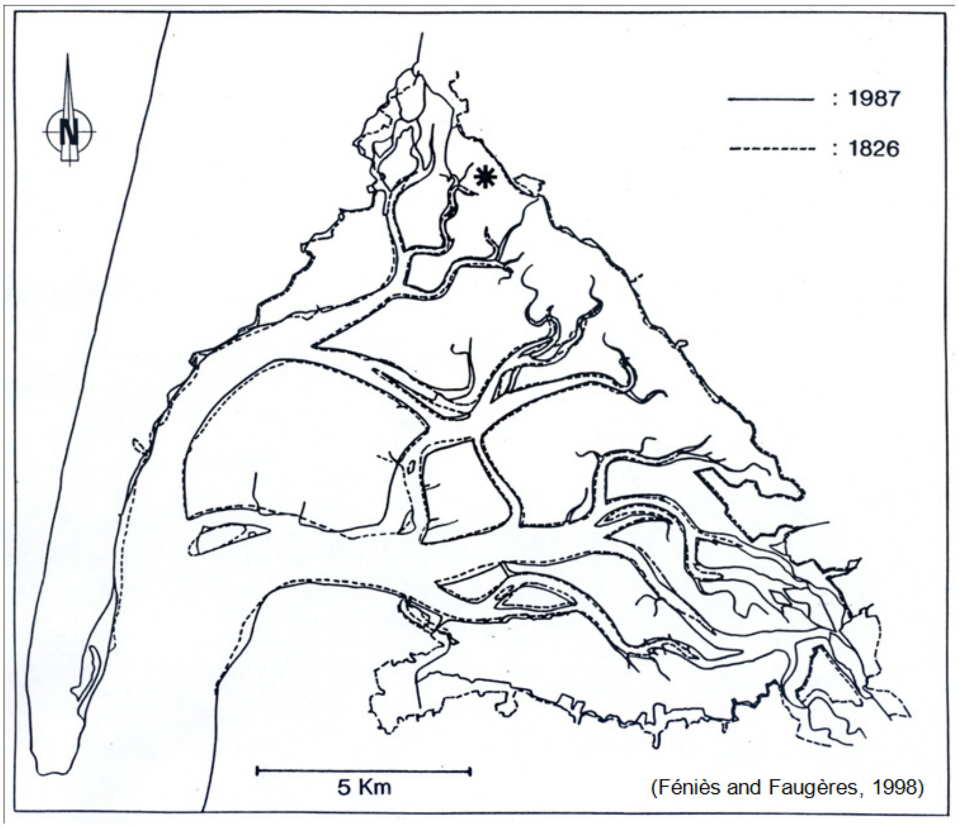
2- The sand that fills the channel is sourced by tidal current erosion of underlying beach. These sands have been progressively incorporated into the channel during its growth and are now deposited in the meanders as point bars. The morphology of these bars is in equilibrium with the channel hydrology and represent a stable stage in the channel development.

In contrast, the tidal channels that are fed by the river Leyre in the south-east corner of the lagoon, migrate laterally. Here, fluvial sands eroded by the river are introduced during the alluvial floods into the tidal channels. These channels quickly migrate laterally to reach a morphology that is in equilibrium with their new hydrological stage.

Overall, this lack of lateral channel migration reduces the potential reservoir volume of the tidal channels in the lagoon. However, this limited lateral extent is compensated by the longitudinal continuity of the channel network. Because the tidal channels converge seaward up to the tidal inlet (Fig. 13), their overall reservoir geometry will present a high degree of interconnectivity.

Superposition of one of the oldest reliable map (dated 1826) and a present day map (dated 1987), note that most of the tidal channels have not migrated laterally during this 160 years period

FIG. 20



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Tidalites 2012

8th International Conference on tidal Environments (Caen, France)

Preconference field trip (July 29-30, 2012)

The Miocene Tidal Shelly Sands of Anjou-Touraine, France

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[1] Université d'Angers, [2] Institut Lasalle-Beauvais,

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FOREWORD

This fieldguide contains basic informations about the location, description and interpretation of selected sites and outcrops visited during a two-day field trip in the Faluns d'Anjou-Touraine. The english translation of the french word «Falun» is «shelly sand» ; As in french the term «Falun» (shelly sands) has a lithologic and stratigraphic value, we use hereafter the term Shelly Sands (with capitals) as an equivalent.

The purpose of this trip is to show tide-influenced or tide-dominated bioclastic deposits infilling a shallow-marine Miocene embayment that mainly occupied the lower Loire Valley during the Middle to Late Miocene : the Shelly Sands Sea.

The tidal signature of these deposits is the reason why this field excursion is scheduled at the event of the 8th International Conference on Tidal Environments. However, although their current-dominated origin was recognised for long due to the various crossbedded facies, the tidal interpretation was generally not invoked.

By contrast, the paleogeography and paleoecology of the Shelly Sands Sea has been studied extensively for over one century, due to the outstanding variety and preservation of the shelly faunas. This paleontological heritage is put forward in regional Museums, one of which is to be visited during this trip.

The Shelly Sands Sea was shallow and flat and the area was not deformed since emersion of the area. As a consequence, natural outcrops are very rare. Fortunately, because the Shelly Sands form good building stones, many old quarries display excellent exposures, allowing to study the 3D architecture of the deposits. Some of these quarries were subterranean, they can be visited and account for the touristic attractiveness of this region.

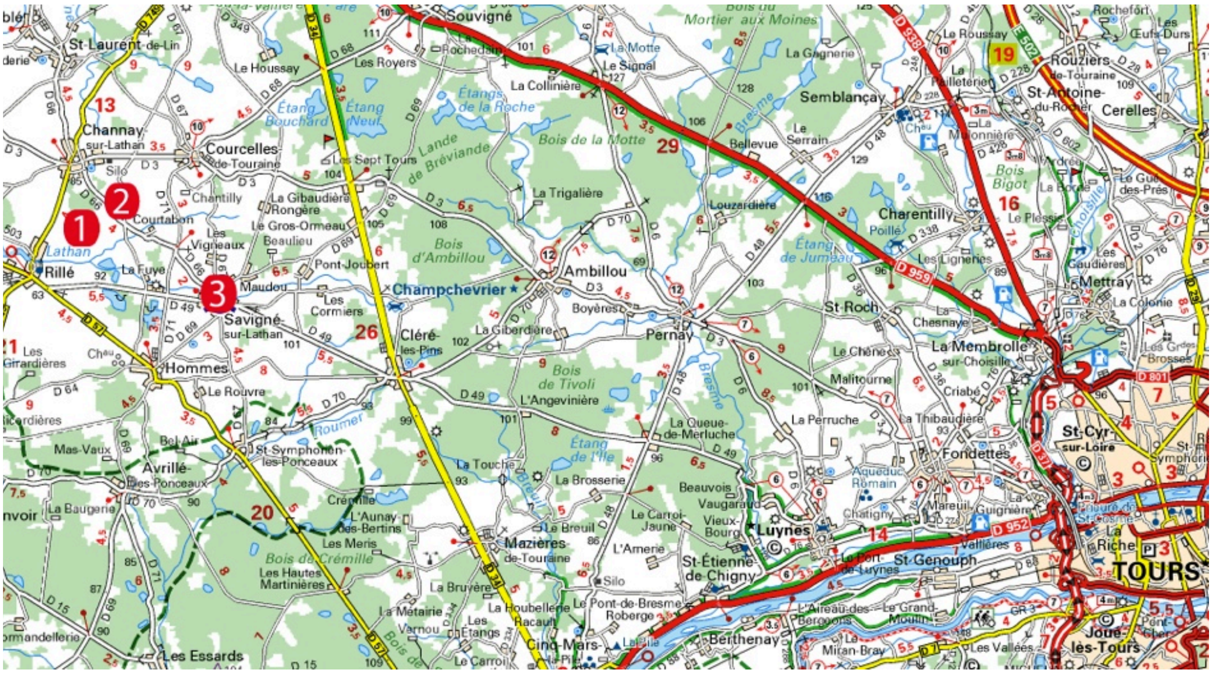
Finally, the Anjou and Touraine are famous historical places of the Loire Valley, reknown for their castles and good wines. During these two days, we shall save time to appreciate these important detours of geology...

SCHEDULE AND ITINERARY

Detailed inset map : page 5

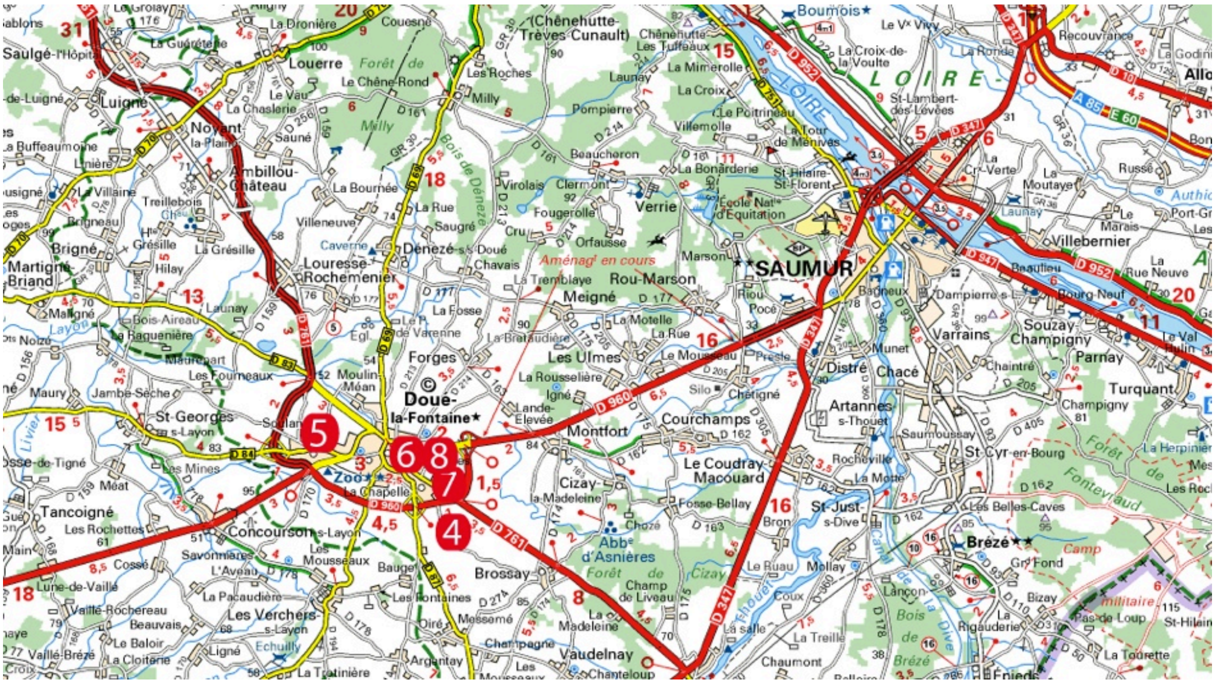
Day 1

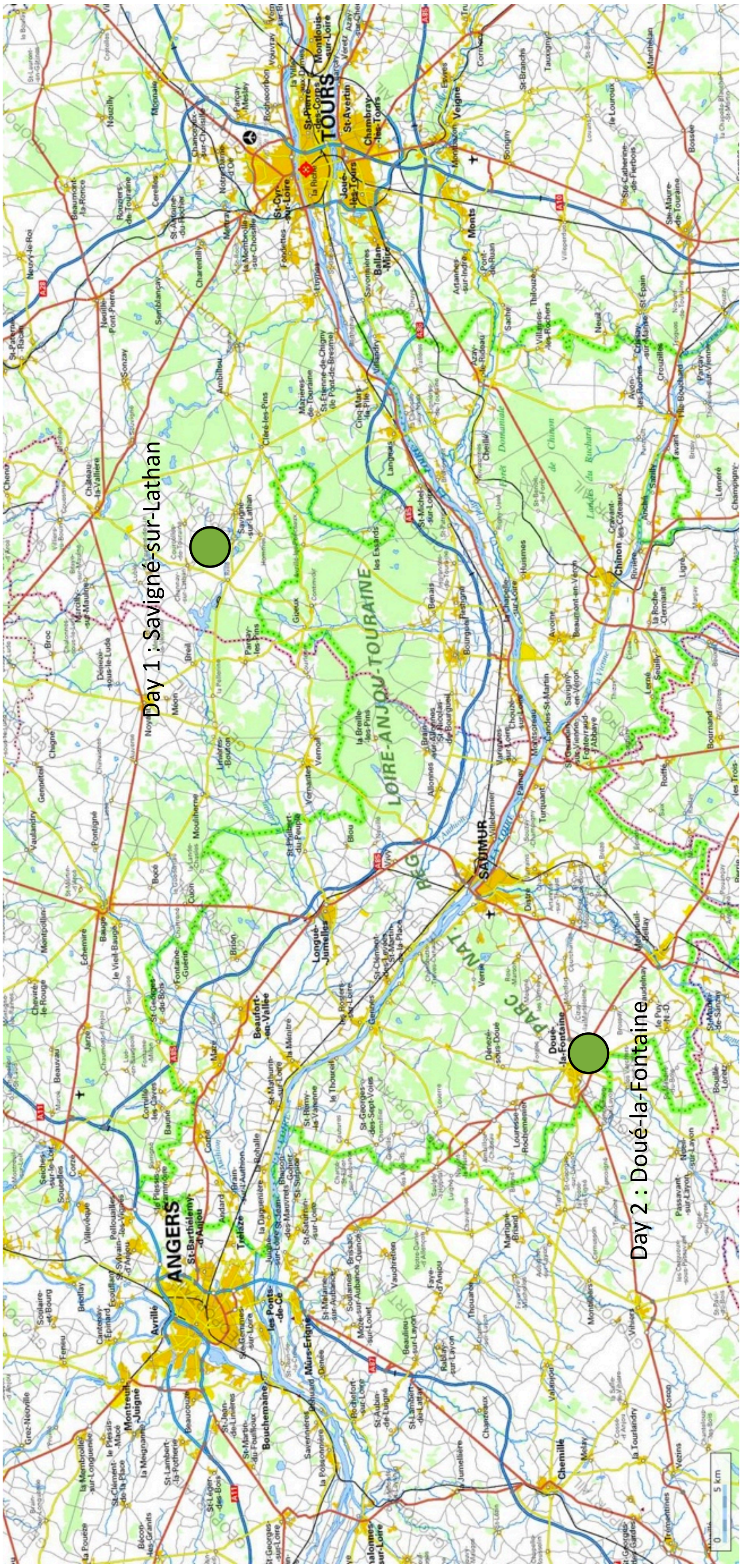
7h30-10h30	Bus transfer from Caen to Savigné-sur-Lathan
10h30-12h00	**1. La Morfassière** - the Savignean Shelly Sands
12h15-13h30	French lunch: Le bouff'tard (Hommes)
13h50-15h15	**2. The Savignean Museum**
15h30-17h30	**3. The Quarry-Museum** - architecture of the Savignean
17h30-18h30	Bus transfer from Savigné to Doué-la-Fontaine
18h30	Arrival at the hotel de La Saulaie
20h00	Diner at Doué-la-Fontaine



Day 2

8h30-10h00	**4. AFPA quarries** - age and general setting of the Shelly Sands
10h00-10h45	**5. Soulanger** - overview of the tidal facies
11h00-12h00	**6. Les Arènes** - large tidal dunes
12h00-13h00	Pic-nic in the Arènes
13h15-14h15	**7. Clos-Melon quarry** - tidal deltas
14h30-17h00	**8. Les Perrières** - 3D analysis of a tidal dune complex
	Visit of the subterranean quarry with Mr Aubineau. Wine testing.
20h30	Arrival in Caen




Day 1 : Savigné-sur-Lathan
Day 2 : Doué-la-Fontaine
TOURS
ANGERS
SAUMUR
LOIRE-ANJOU-TOURAINE
Chinon
Doué-la-Fontaine
Savigné-sur-Lathan
5 km

INTRODUCTION TO OFFSHORE TIDAL DEPOSITS

J.-Y. Reynaud

In this introduction, we wish to summarize some ideas about offshore tidal depositional systems that can be useful in the discussions which the visit of the Shelly Sands outcrops may raise. For a more detailed review, the reader will refer to Reynaud & Dalrymple (2012).

Until recent years, tidal deposits in epeiric seas have been paid less attention than their counterparts in estuarine or coastal settings. One of the reasons is that the basic criteria of recognition of tidal dynamics in the sedimentary facies (the bidirectionnal currents and the presence of mud drapes) are generally lacking. This is due to the fact that the offshore tidal currents are commonly rotary and therefore without a slack water stage. Another reason is that it is difficult to study the hydrosedimentary dynamics of the modern tidal swept sea floors.

The basic knowledge dates back to the 60s, and was indebt to the images provided by side-scan sonar surveys and very high resolution seismics of the shelf seas surrounding Great Britain. It was compiled in the pionneering book of Stride (1982). This book also showed the importance of tidal dynamics as regarding to the production and taphonomy of shelly faunas - the same as those recorded in the Shelly Sands. Due to the large size of the offshore areas hosting tidal systems, the sediment bodies are generally of kilometric lateral extent. On the sandy shelves of the English Channel or the Southern North Sea, they consist of a bedform tract ranging from gravel lags with sand ribbons to large dune fields forming tidal sand sheets and locally sand ridges over 30 m thick and 10 km long (Fig. 1).

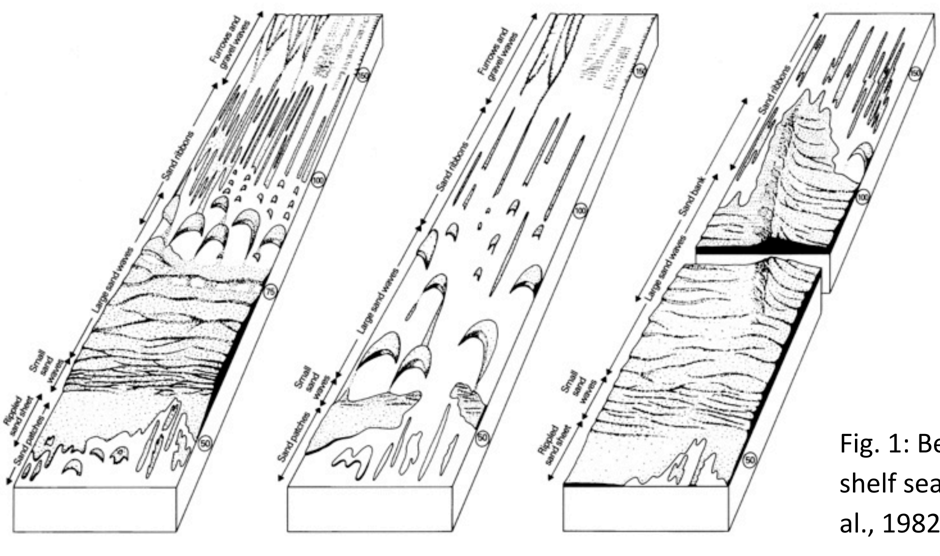


Fig. 1: Bedform tracts of tidal shelf seas (after Belderson et al., 1982).

One of the most important achievements of the work of Stride and collaborators was to relate this bedform tract to tidal-transport pathways, sourced in erosional bedload-parting areas and ending in depositional bedload convergence areas (Fig. 2). Years later, numerical modelings demonstrated that the tidal transport pathways were the consequence of the modification of the tide propagating through these complex shelves, embayments and narrow seaways.

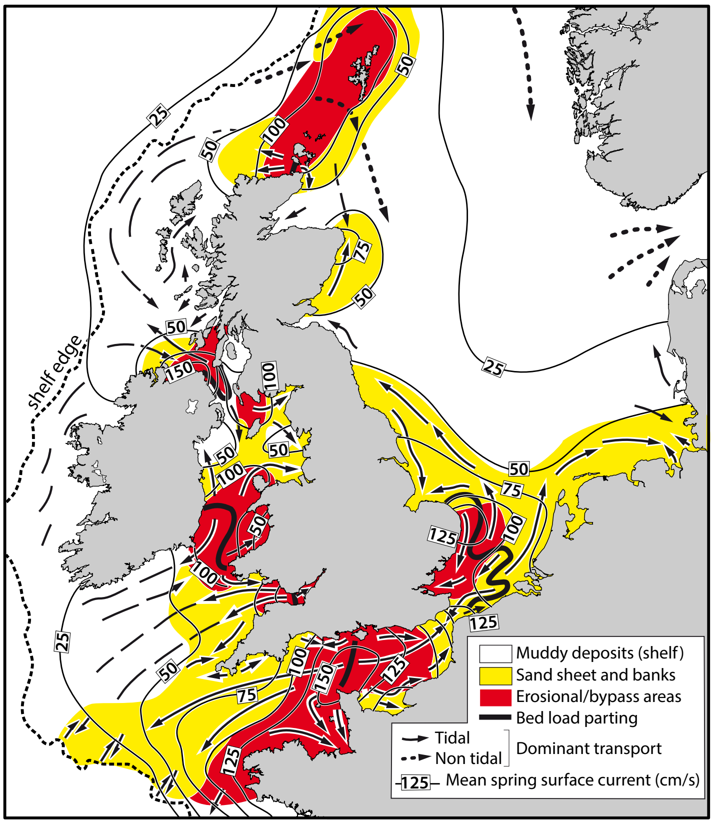


Fig. 2: Sediment cover and tidal-transport pathways of the shallow seas surrounding Great Britain (after Howarth, 1982 and Johnson et al., 1982).

In the English Channel or the Southern North Sea, only the tidal ridges form significant accumu-lations that may be preserved in the stratigraphic record. They are formed by the lateral accretion of sand on the lee side of the dominant tidal current. The sand is transported as fields of tidal dunes (which can be large and compound) so that the basic architecture of the tidal ridges is an oblique cross-bedding at several orders. The master-beds dip of generally less than 10°, which is the lee slope of the active ridges. The residual bedload transport is generally oposite on the two flanks of the ridge, which is a morphody-namic response of the flow over the ridge (Fig. 3).

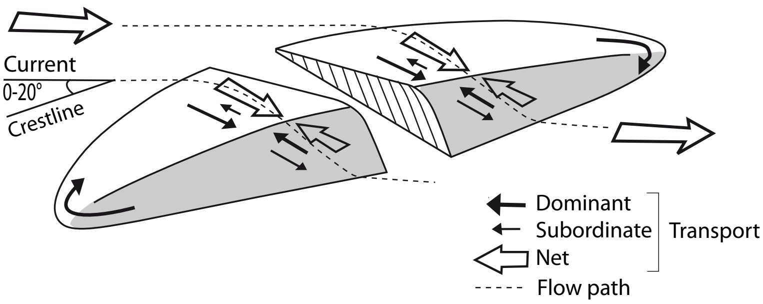


Fig. 3: Conceptual model of flow over an offshore tidal ridge and resulting internal architecture. From Reynaud and Dalrymple (2012)

Numerical models express the ridge growth as the result of friction- and Coriolis-driven vorticity forces generated above an initial bump of the sea bed (Fig. 4). They also show that the ridge size, spacing and shape are scaled to water depth.

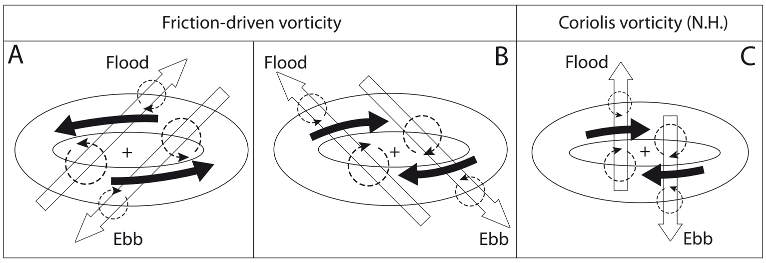


Fig. 4: Vorticity forces accounting for the Huthnance (1982) stability model of offshore tidal sand ridges. From Reynaud & Dalrymple (2012).

Most of the offshore tidal sands that are observed in modern shelf seas are reworked from underlying sequences, namely incised valley fills, in favor of the last post-glacial sea-level rise. This interpretation is the basis of the general stratigraphic model of tidal deposits (Fig. 5).

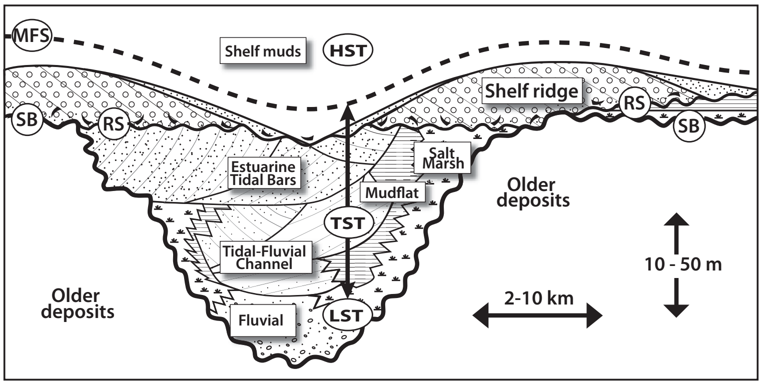
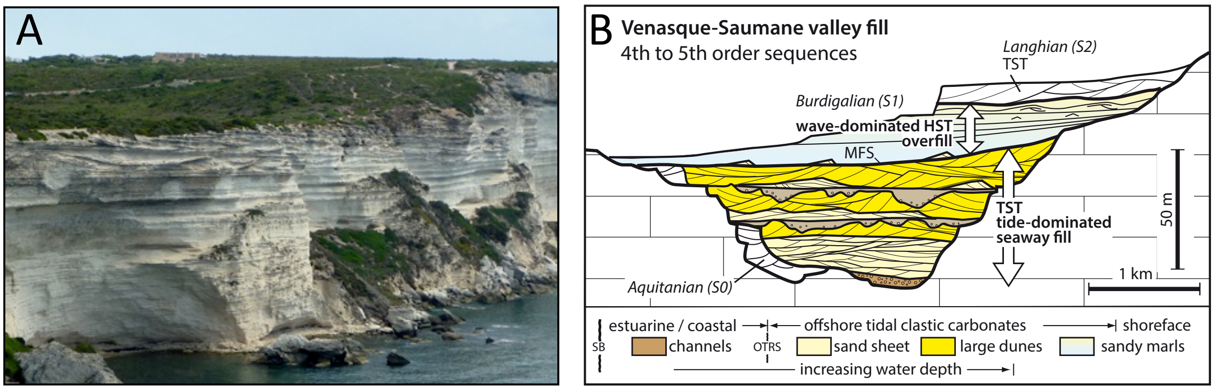


Fig. 5: Model of depositional succession on transgressive tidal shelves, after Dalrymple (2010). From Reynaud & Dalrymple (2012).

It is likely that offshore tidal deposits mostly form during transgressions. On straight shelves, the main reason is that the increase in width of the shelf favours the tidal resonance (an increase in amplitude of the tide that happens where the tidal wave reflected at the coast is in phase with the incoming semi-diurnal tide). Transgressions favor the formation of open-mouthed embayments, where the tide is funneled and therefore accentuated in a landward direction. Also, a rise in sea-level increases the size of semi-enclosed epicontinental seas and therefore their tidal prism (the volume of water exchanged with the ocean during a half tide). Finally, transgressions favor the development of straits and seaways, where tidal flows can be accelerated by constriction or at the junction between two tidal basins.

There are many examples of offshore tidal deposits in the rock record that do not conform, however, to the schematic model of a standard transgressive sequence on tidal shelves. Other deposits than isolated tidal ridges can be preserved in areas where both accommodation and sediment supply are high. Considering french Miocene examples, by contrast to the general tidal ridge model, offshore tidal deposits are more commonly composed of agrading subtidal sand sheets infilling incised valleys or forming subtidal deltas at the outlet of seaways (Fig. 6). This is also the case of the deposits preserved in the Shelly Sands of Anjou and Touraine.



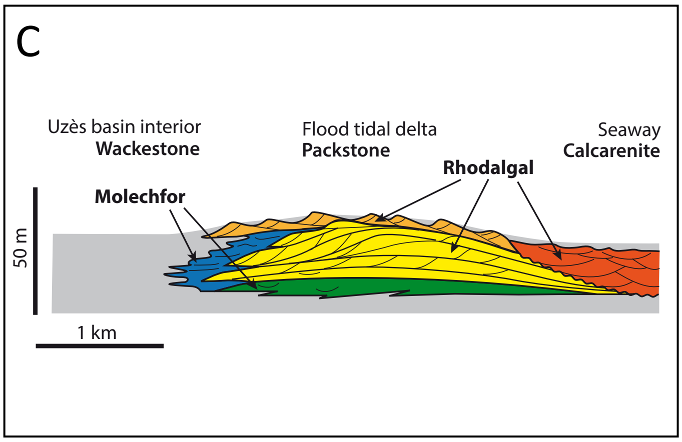


Fig. 6: A: The 80m-high cliff of Bonifacio (Corsica) shows an agrading tidal sand sheet topped by giant 3D dunes. See André et al. (2010). B: Subtidal dunes and channels infilling a Burdigalian incised valley. After Besson et al. (2005). C: Subtidal delta at the entrance of the Miocene Uzès Basin. After Reynaud et al. (2006). In those examples the tidal deposits are composed of bioclastic carbonates.

SOME ELEMENTS ON THE SHELLY SANDS SEA

J.-P. André & C. Gagnaison

The Shelly Sands Sea was a semi-enclosed epicontinental sea that extended over 100,000 km^2. It can be compared to the English Channel, in size and depth, and was probably subject to similar tides. It was connected to the Atlantic Ocean to the W and by a narrow seaway to the Western Channel to the N (Fig. 7).

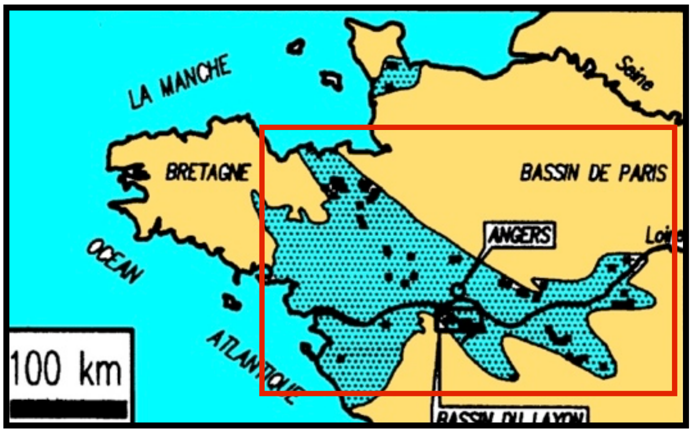


Fig. 7: Maximum extent of the Faluns Sea in the Late Miocene. The black dots indicate the most important fossil sites. The red dots correspond to the places to be visited during the field trip. Red square corresponds to the area displayed on the geological map of Fig. 8.

The seaward connections of the Shelly Sands Sea followed structural lineaments of the metamorphic basement of the Paleozoic Armorican Massif (i.e. Brittany; Fig. 8). To the south, it corresponds roughly to the lower reaches of the Loire valley. The northern seaway followed an older suture between the Armorican block and the basement of Normandy (Cadomian and Brioverian), reactivated during the Oligocene. To the east, the Shelly Sands Sea ended into a large embayment located between Angers and Blois. There, its substratum is composed of Mesozoic rocks (Jurassic carbonate platforms, Cenomanian gravelly sands and marls, and finally Turonian chalks, named «tuffeau», which provide most of the ancient building stones of the area), and Paleogene fluvial to lacustrine deposits forming the final stages of infilling of the Paris Basin.

During the Middle to Late Miocene, a transgression came from the west, and resulted in the deposition of bioclastic sands, in a tidally influenced hydrodynamic regime. These Shelly Sands are famous for their mixture of marine and continental faunas. It is a thin succession, about ten meters thick sometimes reaching around 25 m in Brittany (Durand, 1960), and in some places 20 m in Anjou (less in Touraine). The outcrops are widely scattered due to post-Miocene erosion and the original patchy nature of the deposition. However, they could be used by Lecointre (1947) to draw the assumed shorelines of the Miocene Sea (Fig. 8).

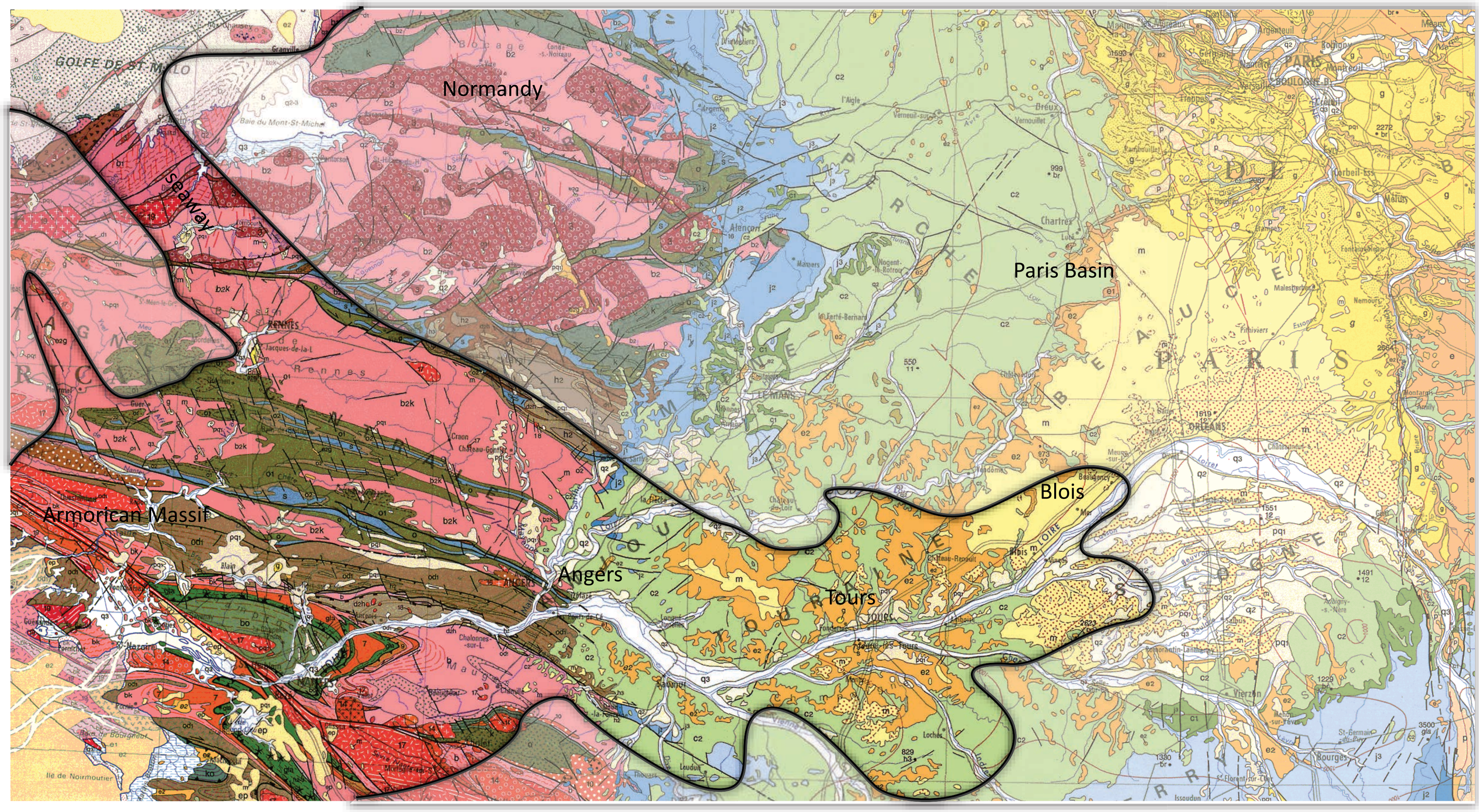


Fig. 8: Geological map of France at 1:1,000,000 scale, with superimposed contours of the Faluns Sea. The Neogene rocks (light yellow) correspond to the Middle and Upper Miocene Faluns d'Anjou and Touraine, except in the easternmost part of that sea, where they correspond to Early Miocene lacustrine limestones.

Uncertainties remain as regarding the duration of this transgression, which roughly corresponds to a 3rd order sea-level half cycle (sensu Vail et al., 1977). However, only higher order sequences may be preserved in the various study sites, the sum of which would encompass the overall transgressive period of the Shelly Sands Sea (Fig. 9).

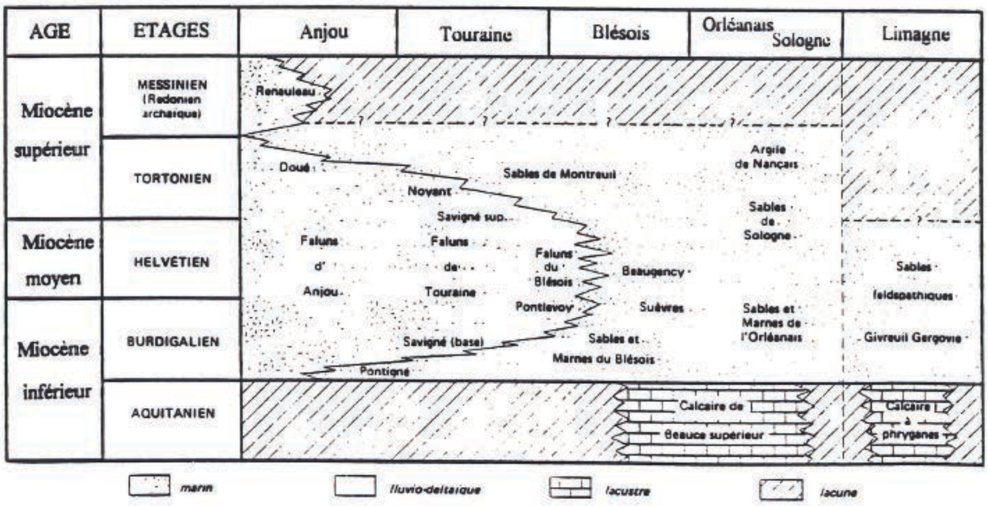


Fig. 9: Chronostratigraphic chart of the Miocene deposits of the Loire Basin (Cavelier et al., 1980).

Lithological components of the Shelly Sands consist mainly of bioclastic shelly sands, siliciclastic particles mixed with 30 to 90% vol. carbonate skeletal grains. The sediment is often crumbly, weakly cemented, and with only hardened zones which show micritic coating. Sands and gravels constitute the siliciclastic fraction (quartz grains, lithoclasts, feldspars and heavy minerals). This detrital material was reworked from a coastal fluvial system (Cavelier, 1980), and also from the transgressive ravinement of the Lower Cenomanian and Senonian sandstones, which were easily mobilized by marine coastal erosion during the Miocene transgression (Biagi, 1993). As a consequence, the well-rounded shape of the siliciclastic grains contrasts with the weaker wearing degree of the Miocene bioclasts. Silts and clays are poorly represented, restricted in some places to the matrix of either the basal conglomerate (in Anjou) or peculiar wackestone facies of the Shelly Sands (in Touraine).

The carbonate fraction is composed of medium to coarse skeletal grains linked as grainstone texture. Fossils are dominated by bryozoans that belong to all zoarial forms (Stach, 1936; Nelson, 1988; Moissette, 1989). The main bryozoan fragments are constituted by debris from vinculariform colonies and numerous articles of cellariiforms. Red algae are common, namely rhodolith pavements, which can be associated with bryozoan colonies ("bryoliths"). Fragments of bivalve shells such as pectinids, chlamyds and oysters are also abundant. Echinoids, brachiopods, barnacles, and some gastropods (often found as internal moulds) are present. Finally, the Shelly Sands contain phosphatized bones and teeth. These include mainly shark, ray and teleostean teeth, and various chelonian, crocodilian and terrestrial mammal remains.

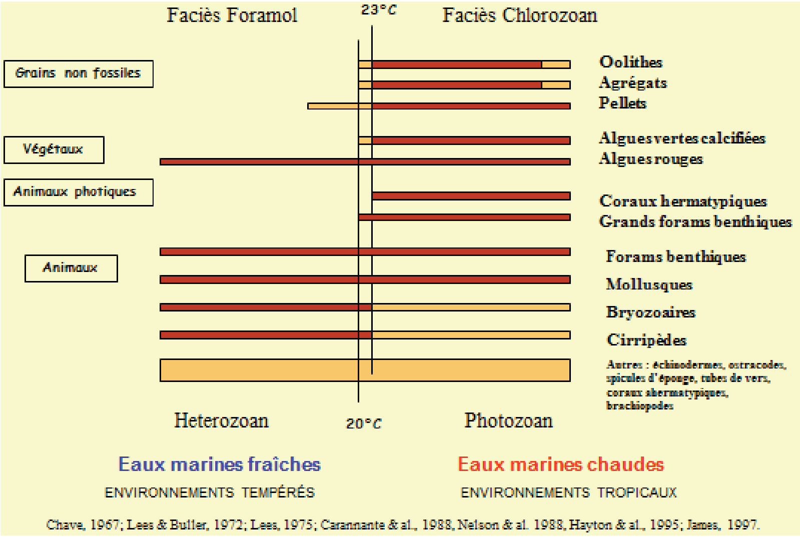


Fig. 10: Biota variation with temperature in marine carbonate ecosystems.

This bioclastic carbonate facies dominated by bryozoans and mollusks indicate a *bryomol facies* (Nelson, 1988; Nelson *et al*., 1988) which corresponds to temperate marine water sedi-mentation (Fig. 10). Neither ooids, nor *Halimeda* or biohermal scleractinians are reported. Furthermore, the siliciclastic part is comparable to those described in ancient and modern formations of New Zealand and Australia (James, 1997). The mineralogy of the rocks is dominated by calcite in contrast to the aragonitic component of warm marine deposits, and dolomite has never been found in this Anjou region. In support of these characteristics, $\partial^{18}O$ studies (Lécuyer *et al*., 1996) indicate an average temperature of 20°C. According to James (1997), these deposits are placed at the upper limit of the marine temperate model, such as the *Heterozoan association*.

Savigné-sur-Lathan : the Shelly Sands of Touraine
C. Gagnaison

Introduction

In the adventures of the giant Gargantua, François Rabelais (Fig. 11) described the Faluns are "dépattures" of Gargantua, who one had gone to the Atlantic Ocean. Once back in Touraine, the beach sand sticked to his boots began to dry and fall out little by little as «depattures», giving birth to the Faluns...

F. Rabelais (1483-1553)

Gargantua (1532)

Fig. 11: Two monuments of the french literature of the XVIth century.

The Savigné-sur-Lathan basin

The basin of Savigné-sur-Lathan is known since the XVIIIth century for the intense quarrying of Miocene marine shelly sands. Since then, several studies were carried out about the Shelly Sands genesis (Dujardin, 1837) as well as the fossils of invertebrates (Bardot, 1980) and marine and continental vertebrates (Ginsburg & Mornand, 1986 ; Ginsburg, 1989).The basin is gently folded following a NNW-SSE direction (Charrier *et al.*, 1977). The Miocene deposits are best preserved along the Esvres syncline (Fig. 12), running through the villages of Noyant-sous-le-Lude and Savigné-sur-Lathan. The Miocene deposits overlay a Late Oligocene lacustrine limestone (Lecointre, 1947). The Miocene deposits are channeled in the Esvres syncline, which thus already existed at that time (Temey, 1996 ; Bouchet, 2009).

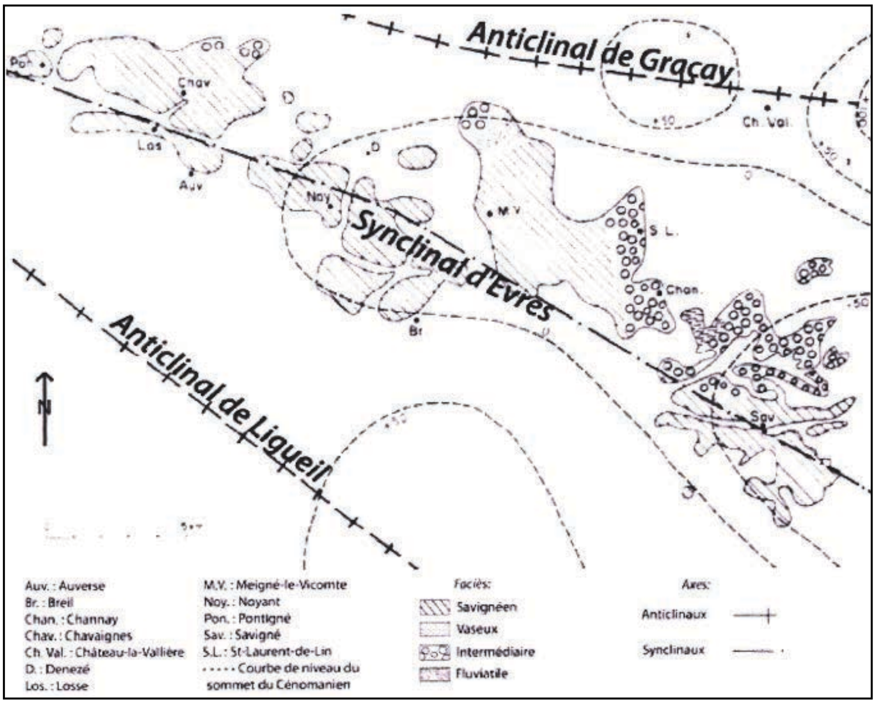


Fig. 12: Structural map of the Savigné-sur-Lathan basin (Ginsburg & Janvier, 1970).

This folded paleotopography was followed as soon as the Burdigalian by the paleo-Loire River system, remnants of which are preserved at the base of the Miocene series in the basin. From Middle to Late Miocene, the area underwent three successive transgressions. At that time, the Shelly Sands Sea entirely flooded the lower reaches of the former fluvial basin (Ginsburg, 1989 and 2001 ; Bouchet, 2009). The stratigraphy of the resulting deposits can be reconstructed from several quarries (Fig. 13).

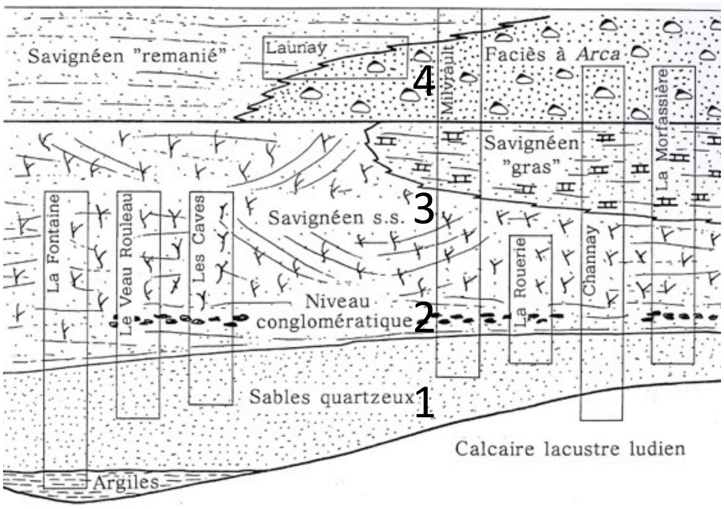


Fig. 13: Schematic stratigraphy of the Miocene deposits of the Shelly Sands Sea in Touraine. The Shelly Sands correspond to the «Savignéen» facies : a bryozoan-dominated cross-bedded calcarenite.

Four layers are exposed in the Shelly Sands quarries of Savigné-sur-Lathan/Noyant-sous-le-Lude (Figs. 13 & 14): (1) Burdigalian fluviatile sands, (2) Langhian calcarenite comprising *Parascutella faujasi* and *Amphiope bioculata*, (3) Langhian to Serravalian clayey shelly sands comprising bryozoans such as *Cellepora* sp. and *Retepora* sp., and (4) a platy calcarenite with *Anadara turonica* (Lévêque, 1989 ; Temey, 1996).

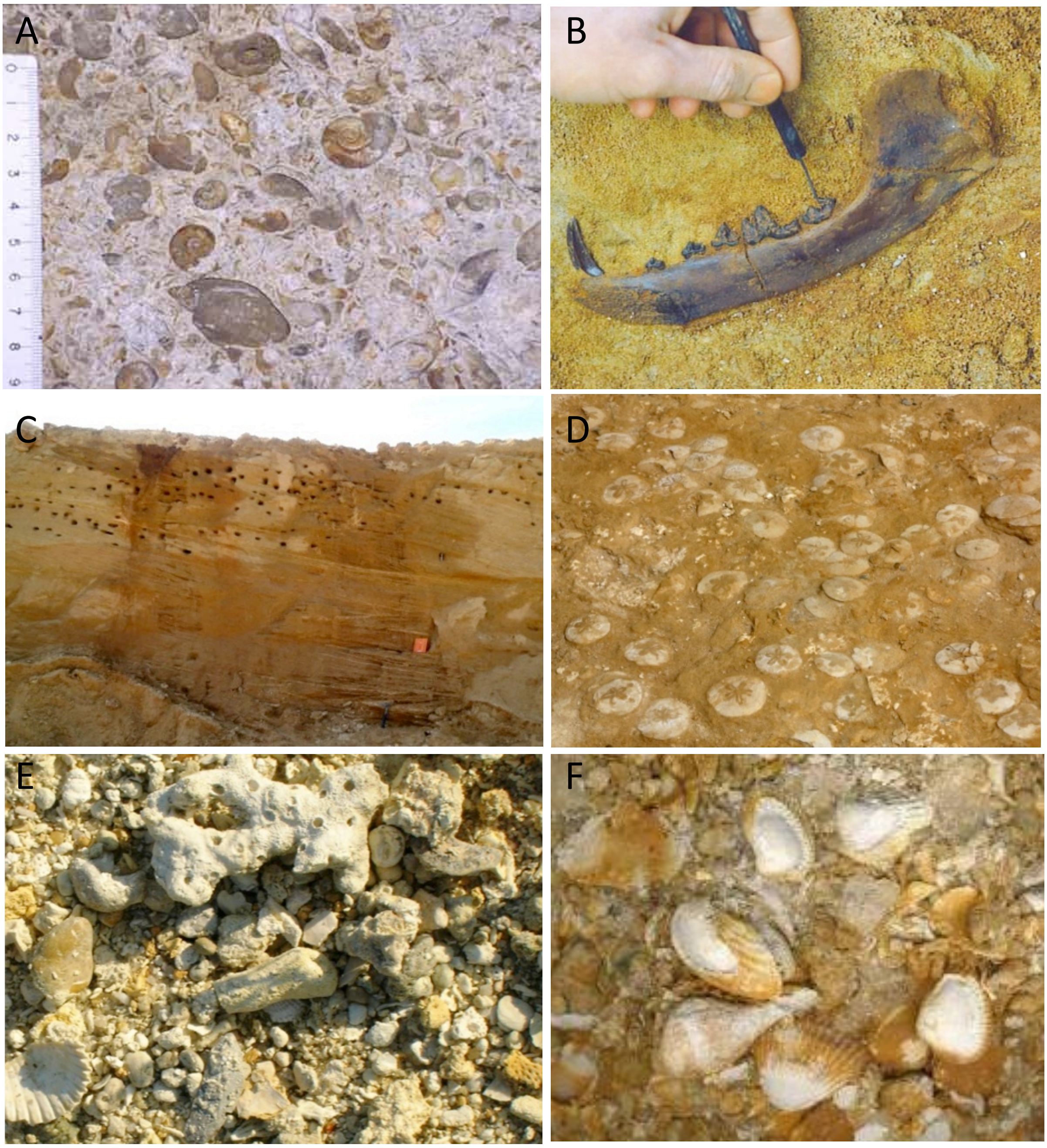


Fig. 14: Deposits of the Miocene of Touraine. A: Late Oligocene lacustrine limestone; B: Orleanian (Burdigalian) fluviatile sands with land fauna (*Cynelos bohemicus*); C: Tidal bundles in the Langhian calcarenite (Les Bournais quarry's); D: Langhian calcarenite also including *Parascutella faujasi* and *Amphiope bioculata* (Les Bournais quarry's, 2011); E: The clayey shelly sands comprising bryozoans such as *Cellepora* sp. (Langhian to Serravalian); F: *Anadara turonica* platy calcarenite (Tortonian).

Stop 1.- **The Savignéen limestone: The Shelly Sands with celleporiforms bryozoans (La Morfassière)**

This old quarry offers good exposures of the classic Shelly Sands of the Savigné-sur-Lathan/ Channay-sur-Lathan basin. Sedimentary structures are dominated by large-scale cross-bedding, which are interpreted as tidal bedforms. The Shelly Sands exhibit numerous bioclasts and well preserved celleporiforms bryozoans.

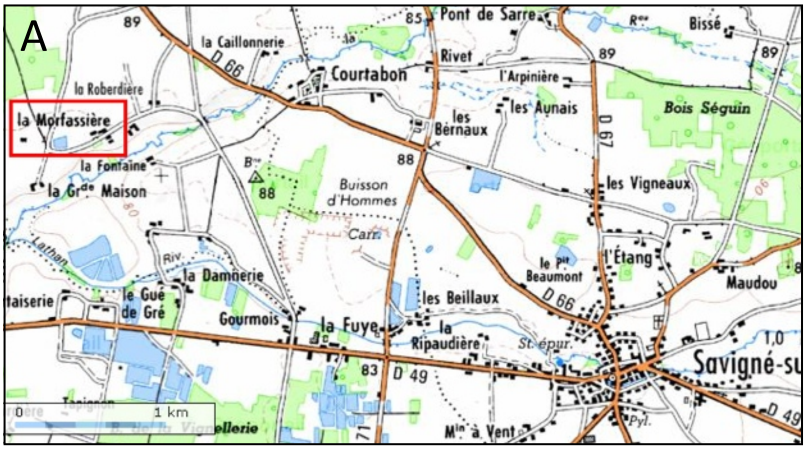
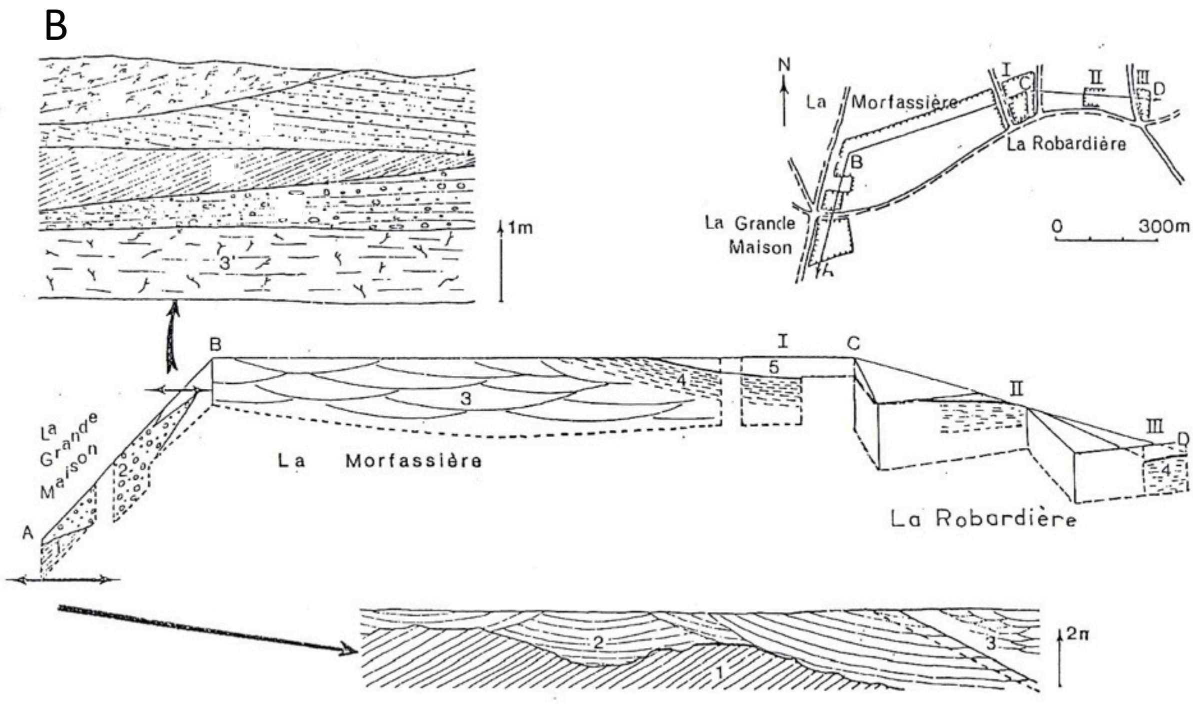
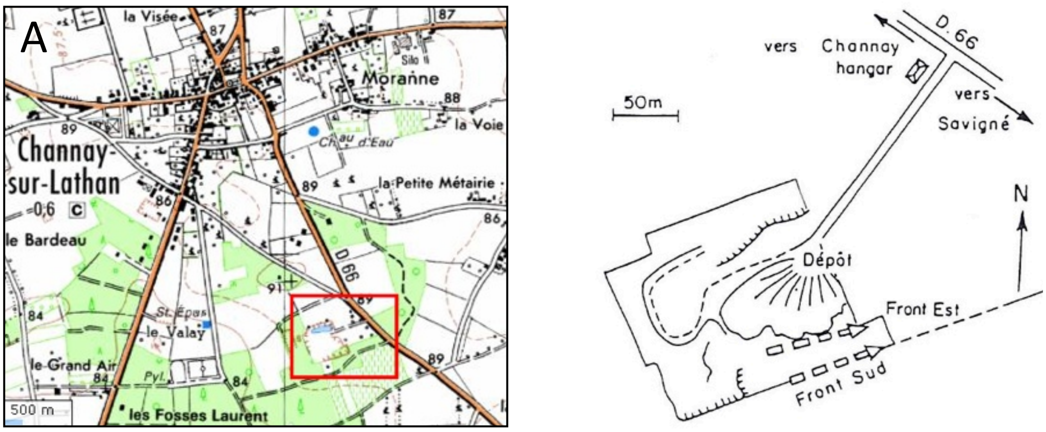


Fig. 15: A: Location map. B: Sketch of the quarry. 1: fine-grained sandstone. 2: coarse-grained sandstone. 3: Shelly Sands. 4: Muddy Shelly Sands. 5: *Arca*-rich platy calcarenite. Note the large-scale oblique and trough cross-beddings.



Stop 2.- **The Quarry-Museum : the Oligocene-Miocene boundary, the Miocene calcarenites and the Shelly Sands**

The Quarry-Museum presents the contact between the Oligocene lacustrine limestone and the overlying Miocene marine deposits (Fig. 16). All the Miocene facies of the Savigné-sur-Lathan/Channay-sur-Lathan basin are exposed in the quarry. Sedimentary structures of various types are visible. The Langhian calcarenite and the Tortonian *Anadara turonica* platy calcarenite exhibit crossbeds that interpreted as produced by the migration of tidal dunes.



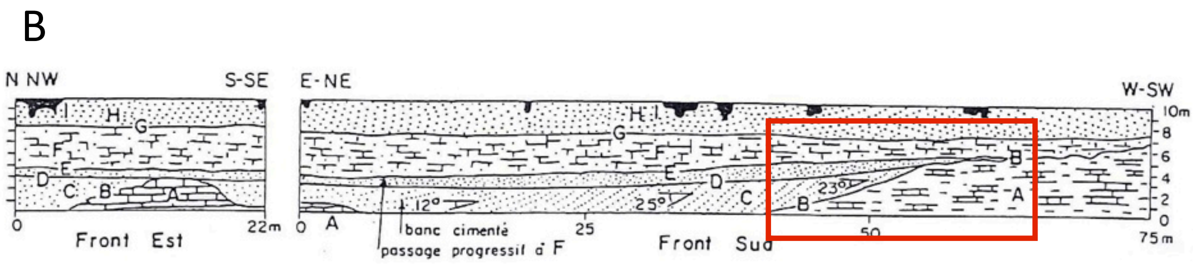


Fig. 16: A: Location map of the Quarry-Museum. B: Architecture of the Miocene deposits. a: Oligocene lacustrine carbonate. b: erosion surface with borings. c: sandy, oblique bedded deposits (channel fills). d: transgressive ravinement surface. e: nodule-bearing transgressive lag. f: Muddy Shelly Sands. g: ravinement surface. h: *Andanara turonica*- rich coarse-grained calcarenite. This latter deposit exhibits a large-scale trough cross-bedding. C: Close-up on the outcrop (red square in B).

Stop 3.- **The Savignean Museum**

The «Savignean» is the bryozoan-rich facies of the Shelly Sands, named after the village of Savigné-sur-Lathan. The Savignean Museum (Fig. 17) presents the local geology of the Miocene and an important paleontological collection of the related marine and terrestrial faunas (Fig. 18 & 19). Interestingly, it fills the gap between the geological heritage and the folk traditions of Touraine.

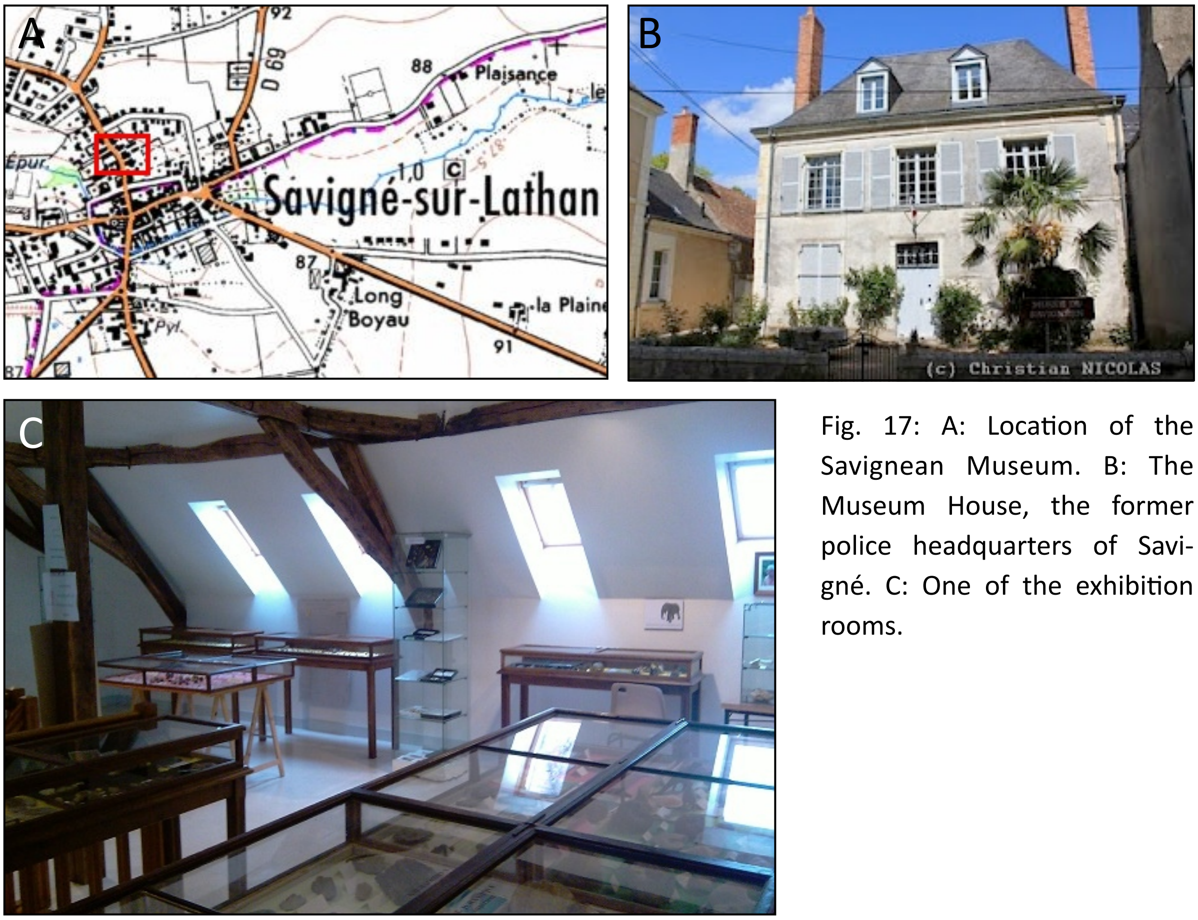


Fig. 17: A: Location of the Savignean Museum. B: The Museum House, the former police headquarters of Savigné. C: One of the exhibition rooms.

The terrestrial vertebrate fauna reworked in the Shelly Sands (Fig. 18) indicate a warm paleoclimate (arboreaceous savannah), while the marine faunas of the Bryomol indicate a rather cooler sea temperature. Similar relationships can be observed today at the coastal borders of West Africa, such as Mauritania.

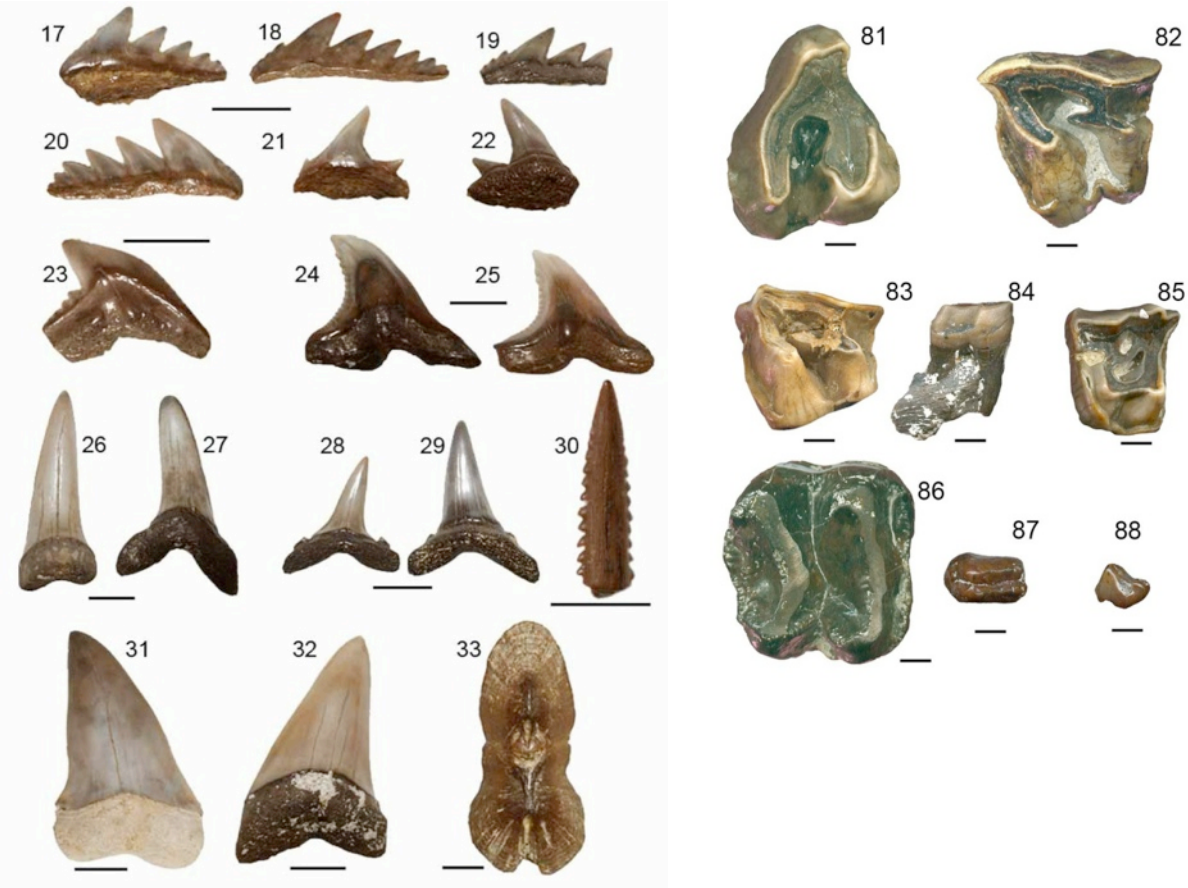


Fig. 18: The HARTMANN Collection (Miocene fossils) of the Savignean Museum (Pouit, 2009 and Gagnaison & Hartmann, 2009). Marine fishs.- 17 to 22: *Notorynchus primigenius.* 23: *Galeocerdo aduncus.* 24: *Hemipristis serra*. 25: *Hemipristis serra*. 26 & 27: *Isurus desori*. 28 & 29: *Carcharias* sp. 30: Tail spine of Myliobatiforms. 31: *Lamnidae* sp. 32: *Cosmopolitodus hastalis*. 33. Suldered rings of *Dasyatis* sp. Terrestrial vertebrates.- 81: *Brachypotherium brachypus*. 82: *Prosantorhinus douvillei*. 83: *Prosantorhinus germanicus*. 84: *Protaceratherium minutum*. 85: *Plesiaceratherium platyodon*. 86: *Deinotherium bavaricum*. 87: *Zygolophodon turicensis*. 88: *Deinotherium bavaricum*. Scale bars : 1 cm.

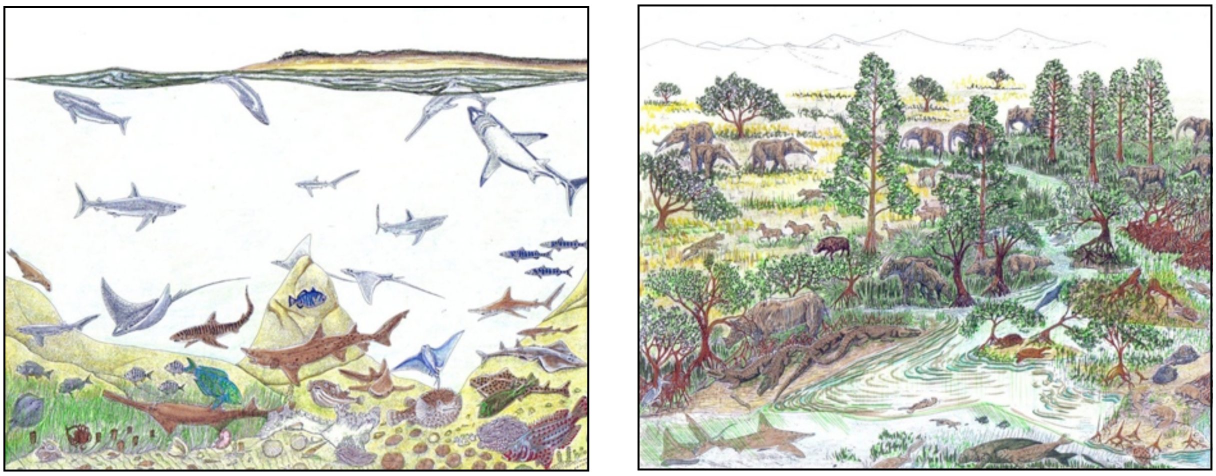


Fig. 19: Artist view of the paleoenvironements of the Savignean (http://www.museedusavigneen.com/)

Doué-la-Fontaine : the Shelly Sands of Anjou

J.-P. André

Introduction

As Touraine, Anjou is marked by the Loire Valley that contributes much of its history and economy. Anjou is also known for its mild climate ("La douceur angevine" from Joachim Du Bellay) allowing market gardening and wine production. Also, Anjou is associated to building stones (schists, slates and limestones). This the reason why there is a dense network of quarries, the oldest of which date back from the Middle Age. Another consequence is the amount of troglodytic houses and cellars (Fig. 20).

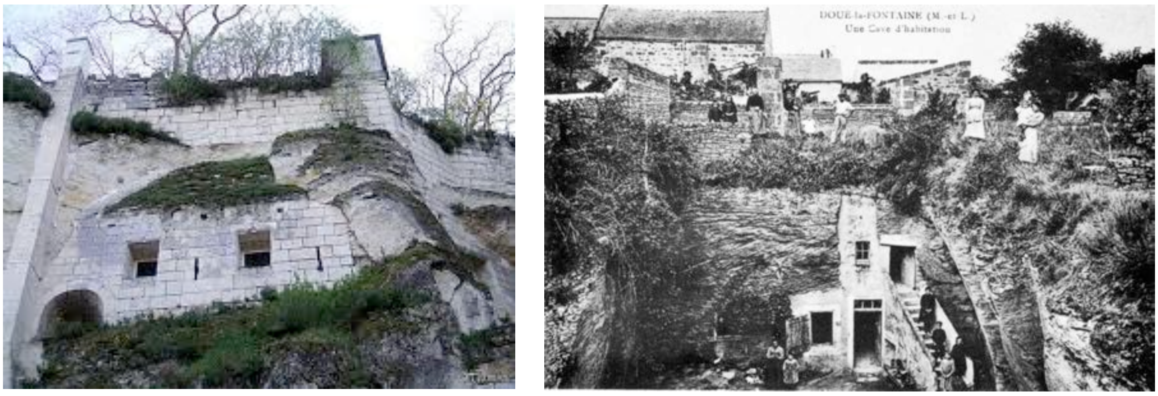


Fig. 20: Troglodytic housing in the Turonian chalk in Saumur city (A) and in the Miocene Shelly Sands in Doué-la-Fontaine (B).

Age of the Anjou Shelly Sands

The discovery of vertebrate fossils as Hipparion (Fig. 21 and B) in the base levels of the faluns of Doué-la-Fontaine allowed to relate the formation of the Valesian, MN9 zone, corresponding to the Tortonian (Upper Miocene; Ginsburg & al., 1979). Isotopic measurements carried out from phosphatic biogenic fragments collected throughout the basin, from Blois to Britanny, yielded ages ranging from 10 to 11 My (Barrat & al., 2000). The lithology, facies and architecture of the Anjou Shelly Sands are described in André et al. (2003).

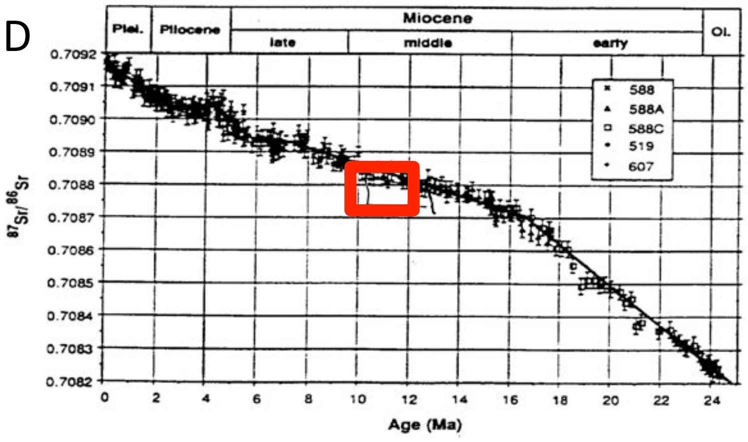


Fig. 21: A: Artist view of Hipparion. B: Tooth of Hipparion. C: Stratigraphic stages and absolute ages of the Shelly Sands in Anjou. D: Strontium isotope curve and range (red square) of the Shelly Sands. after Hodell et al., (1991).

Bryozoans are the main components of the Shelly Sands of Anjou, mostly composed of encrusting and rigid erect zoarial forms (Fig. 22). Zoarial forms give indications on the depositional environment. They generally indicate a low to moderate sedimentation rate. They may grow on a rocky substrate but also encrust or be fixed on particles of a loose substrate during weak energy periods (Fig. 23). Encrusting zoarial forms correspond to membraniporiforms, which mostly encrust shells of dead animals, and celleporiforms, which may encrust all kinds of particles, namely algae thalli. Fixed living forms comprise the rigid erect colonies (platy adeoniforms and eschariforms, reteporiforms and delicate branching vinculariforms), flexible erect colonies (catenicelliforms and cellariiforms, often preserved as very small debris), and free living colonies (lunulitiforms). Fixed zoarial forms may have several stages of development, between hydrodynamic periods of seabed reworking by currents.

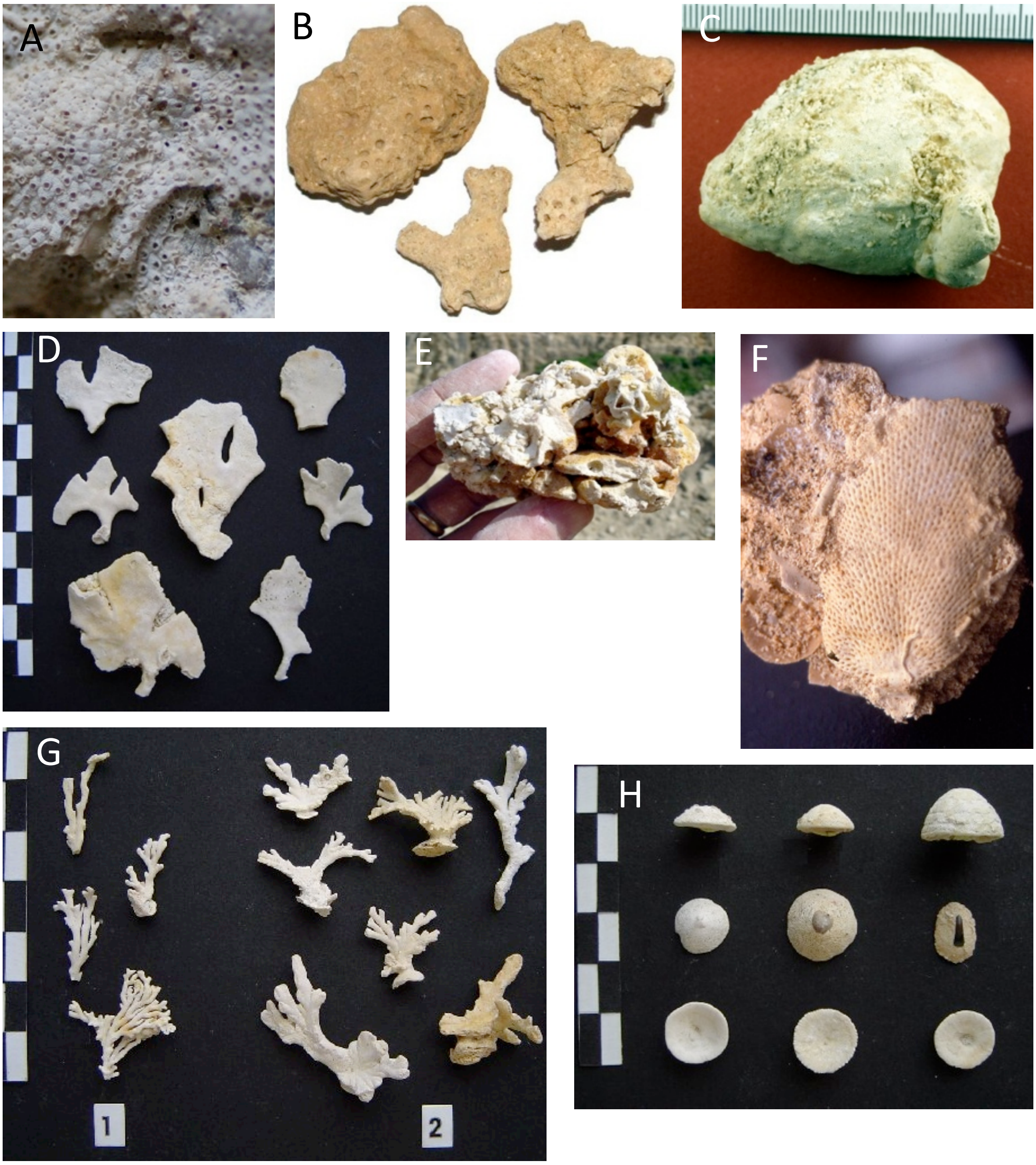


Fig. 22: Encrusting colonies of Bryozoans. MEMBRANIPORIFORMS- A: Modern Bryozoan colony (Rhodes Island, Greece). Photo P. Moissette. CELLEPORIFORMS- B: Celleporaria Palmata. C: Celleporiform colony encrusting an algae thallus. ADEONIFORMS- D: Several species, coll. D. Millet. ESCHARIFORMS- E: Mixed encrusting and erect colonies (bindstone like). RETEPORIFORMS- F: Large colony (6 cm). VINCULARIFORMS- G: 1: *Hornera radians,* 2: *Hornera frondiculata*; coll. D. Millet. LUNULITIFORMS- H: Several species, coll. D. Millet.

PALEOECOLOGY OF ECTOPROCTS (BRYOZOANS)

TABLE *I*—AN ESTIMATE OF THE ASSOCIATION OF ASPECTS OF GROWTH OF ECTOPROCTS WITH ENVIRONMENTAL FACTORS

Growth type	Substratum			Current (cm./sec.) 20 100			Rate of sedimentation (cm./10^3 yrs.) 10 100 1000			
	Hard	Flexible	Particulate	Low	Moderate	High	Low	Moderate	High	Very high
Erect, rigid										
Adeoniform	XX			X	XX		XX			
Eschariform	XX		X		XX		XX			
Reteporiform	XX			X	XX		XX			
Vinculariiform	XX			XX			XX			
Erect, flexible										
Catenicelliform	X	XX	X		X	XX		XX	X	
Cellariform	XX	X	X	X	XX	X	X	X	XX	
Flustriform	XX	XX	XX	XX	X	X	XX	X	X	
Encrusting										
Celleporiform	X	XX			XX	X	XX			
Conescharelliform		XX			XX		not applicable			
Membraniporiform A	XX	X			XX	X	XX			
Membraniporiform B	X	XX			XX		XX			
Petraliiform	XX	X	X		XX		XX			
Pseudovinculariform		XX			XX		XX			
Setoselliniform			XX	XX	X		XX			
Free-living										
Lunulitiform			XX		XX	X	X	XX	XX	

X, occasional association; XX, frequent association. The quantitative limits on any group are mostly poorly known; they represent an attempt to be more specific than the vague terms "high", "moderate", or "low". The degree of association is based on data accumulated from the continental shelf off New England and from other reports of this type of analysis by Brown (1952), Canu & Bassler (1920), Cheetham (1963), Harmer (1926), Labracherie & Prud'homme (1967), Lagaaij & Gautier (1965), and Stach (1936, 1937).

SCHOPF, 1969

Fig. 23: Relationship between substrates and zoorial forms of bryozoans, from Schopf (1969).

Stop 4.- **AFPA quarries : paleo-onlap of the basin**

This stop aims to show the paleo-onlap of the southern border of the Shelly Sands Sea onto its basement, as exposed in an old quarry. The basement is constituted here by the Jurassic limestones. The transgressive surface is widely bored by bivalves. The base of the Shelly Sands contains several reworked debris : flints from the Jurassic, pebbles and gravels of Paleozoic (glauconized quartz and quartzite gravels) several fossils from Jurassic (*Ammonites*) and Cretaceous (oysters as *Pycnodonta*). Some relicts of paleokarst are assumed.

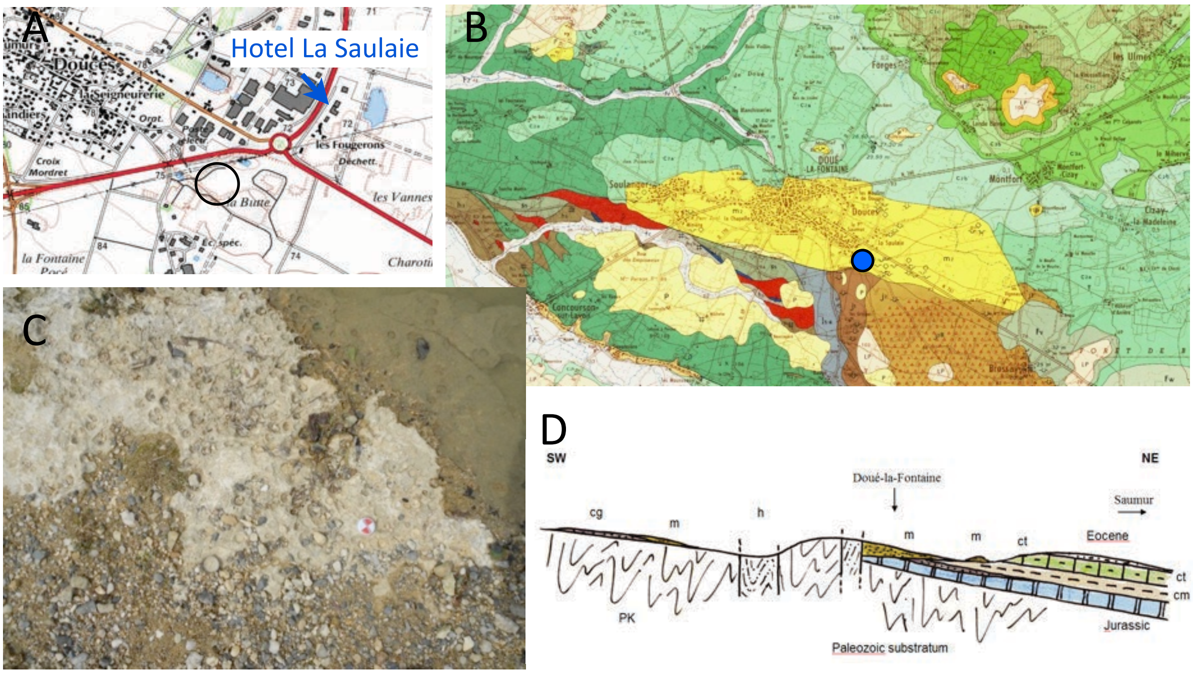
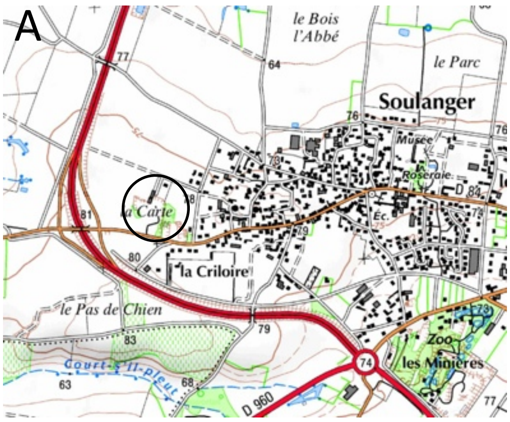


Fig. 24: A: Location map. B: Geological map at 1:50,000. Yellow: Miocene. Green: Cretaceous. Grey and brown: Jurassic. Other tones: Paleozoic and Precambrian. C: Bored surface of onlap of the Miocene deposits of the Shelly Sands above Jurassic limestones. Cap for scale! D: Schematic section of the Doué-la-Fontaine area. The fault which separates the Miocene from the substratum formed a paleocliff during sedimentation of the Shelly Sands. Pk: Precambrian. h: Carboniferous. cg: sandy Cretaceous. cm: marly Cretaceous. ct: chalky Cretaceous.

Stop 5.- **Soulanger : very large offshore dunes**

This outcrop shows large-scale, low angle cut-and-fill structures preserved in an old quarry (Fig. 25). By contrast to the architecture of channel filling structures (Fig. 26A), we can see that the internal packages of strata are commonly isopach (which is exceptionnal in channels, where there is a strong change in energy from the deep to the bank, and therefore a variation in depositional rate). As a consequence, we interpret these cut-and-fill structures as very large 3D dunes over which flow variations are less constrained. The transport direction is globally toward the WNW.



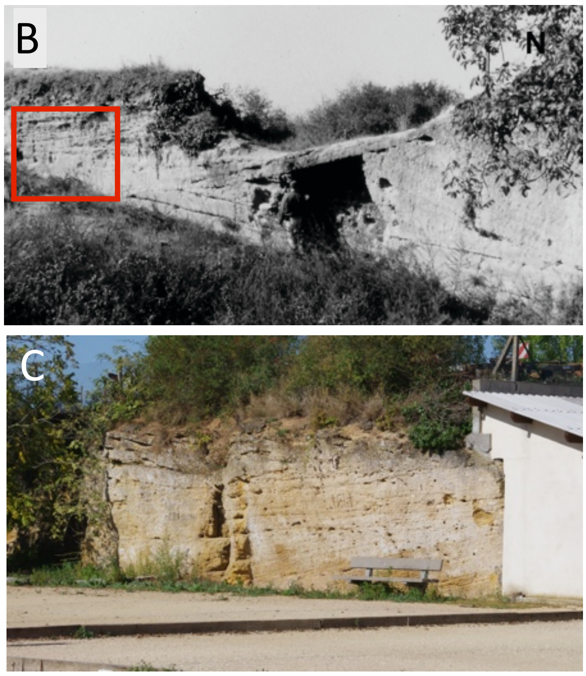


Fig. 25: A: Location map. B: The quarry wall as it was 20 years ago. R. Biagi for scale. C: The quarry is now backfilled, and only the upper part of the wall can be observed at places, between new buildings.

Strong currents are necessary to create these very-large dunes (Fig. 26B). In addition, their formation requires a water depth equal to at least five times the dune height. The bathymetry can thus be evaluated here to about twenty meters.

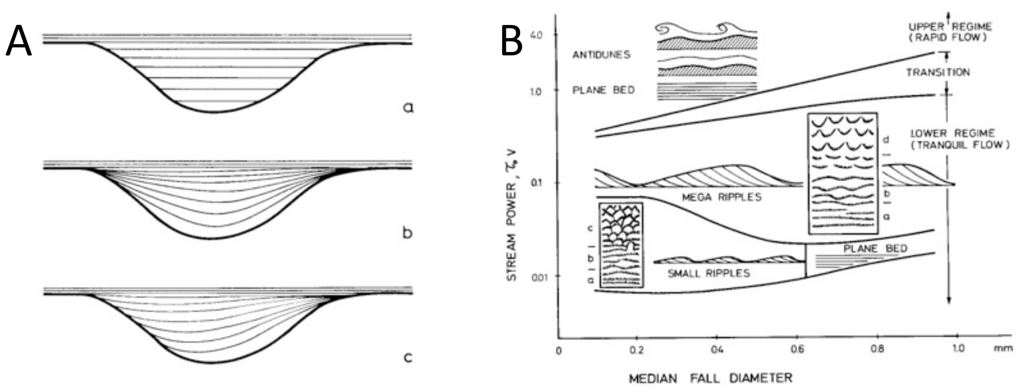


Fig. 26: A: Three different geometries of channel fills. Except for a passive infilling (a), strata are anisopachous (b and c). B: Stability diagram of bedfroms as a function of stream power and grain-size. From Allen (1968) in Reineck & Singh (1973).

Moving eastward at some hundreds of meters, a bryozoan-rich facies can be observed (Fig. 27). Some adeoniform or reteporiform colonies are exceptionally big and well developed. A careful observation shows that the development of some reteporiform colonies can «keep up» with sedimentation, as some laminae of the dune foresets onlap their branched bodies. Knowing that such colonies can grow up of some centimeters per month, a very high sedimentation rate is deduced.

Fig. 27: Bryozoan colonies on the lee side of a dune at Soulanger. They fix the sediment and are able to keep up with sedimentation. Picture frame is about 10 cm wide.

Stop 6.- **Les Arènes of Doué-la-Fontaine : architecture of the very large dunes**

The Arènes of Doué-la-Fontaine are ancient quarries used to extract building stones because the Shelly Sands here are relatively well cemented. Nowadays the site hosts various events as the Feast of the rose or banquet associations.

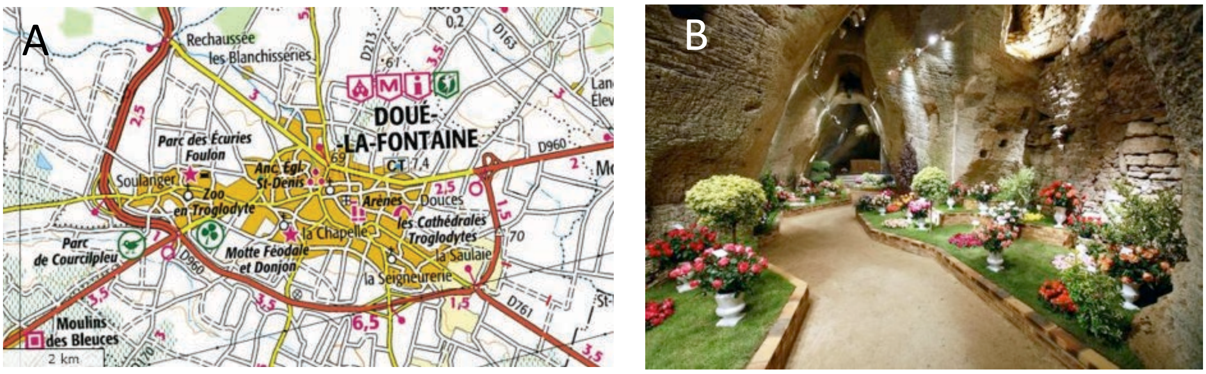


Fig. 28: A: Touristic map of Doué-la-Fontaine showing that the most important places are related to geology (subterranean quarries and troglodytic houses). B: The Rose feast at the Arènes in july.

Various structures can be observed. The most spectacular is a large 2D dune with foresets dipping to the W (Fig. 29). At the base of the lee side laminae, it is possible to distinguish some small structures which laminae have a reversal dip : they are interpreted as backflow ripples (Fig. 30).

Fig. 29: Architecture of a very large dune showing oblique planar bedding (foresets) overlying horizontal trough cross-bedding (bottomsets). The dominant accretion direction in the trough cross-beds is opposite to the slope dip direction of the large foresets.

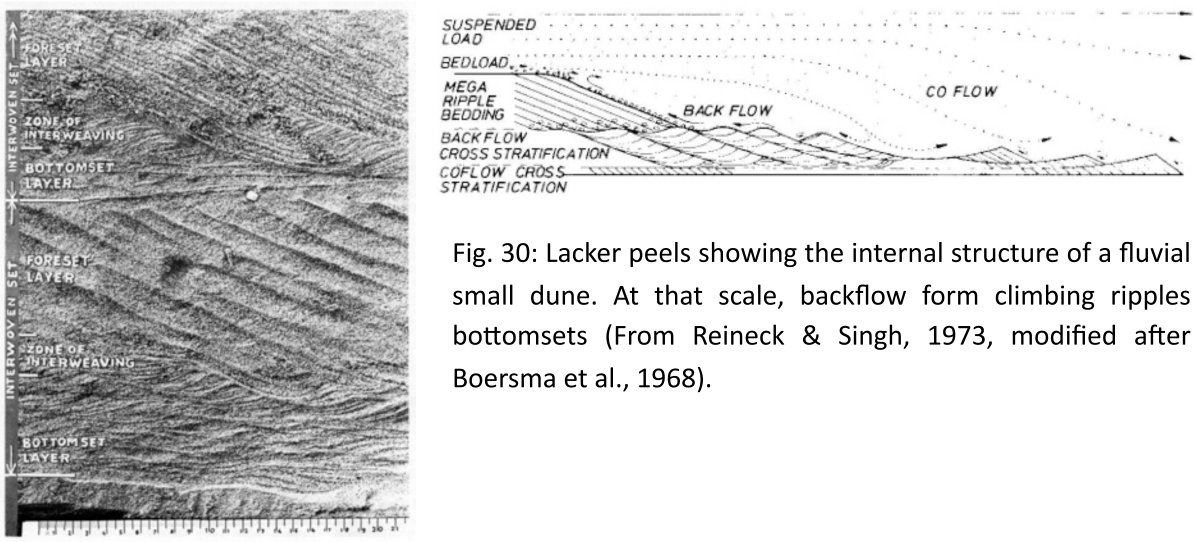


Fig. 30: Lacker peels showing the internal structure of a fluvial small dune. At that scale, backflow form climbing ripples bottomsets (From Reineck & Singh, 1973, modified after Boersma et al., 1968).

This architecture is typical of that of a dune formed under unidirectional flows. However, up-climbing ripples are also locally observed superimposed to the foresets of the large dunes in the Arènes. This implies the reversal of the current shaping them. Therefore, these dunes are interpreted as tidal dunes.

Fig. 31: Up-climbing ripples within the foresets of one large dune (black arrow). The white arrow points to a slumped bed, which suggests that avalanching also proceeded on the lee face of the dune.

Fig. 32: Entrance of Charlemagne cellar.

At the entrance of the Charlemagne cellar (Fig. 32), metric 3D dunes systems and also more flat structures with metric ondulations can be related to antidunes or hummocky cross stratification. It could be result of the surimposition of a swell control on the main tidal control.

Stop 7.- **Clos Melon : subtidal delta**

To the East of Doué-la-Fontaine, we observe a large quarry from the road (Fig. 33B). The lithoclastic content in this facies is higher in proportion than in the previous stops. The deposit is organized at first order in large-scale oblique tabular beds, with 25° slope dip to the West. This stratification resemble to the foresets that constitue a Gilbert delta. The visible thickness of the system is at least 15 meters. As a consequence, the water depth would have exceeded this value.

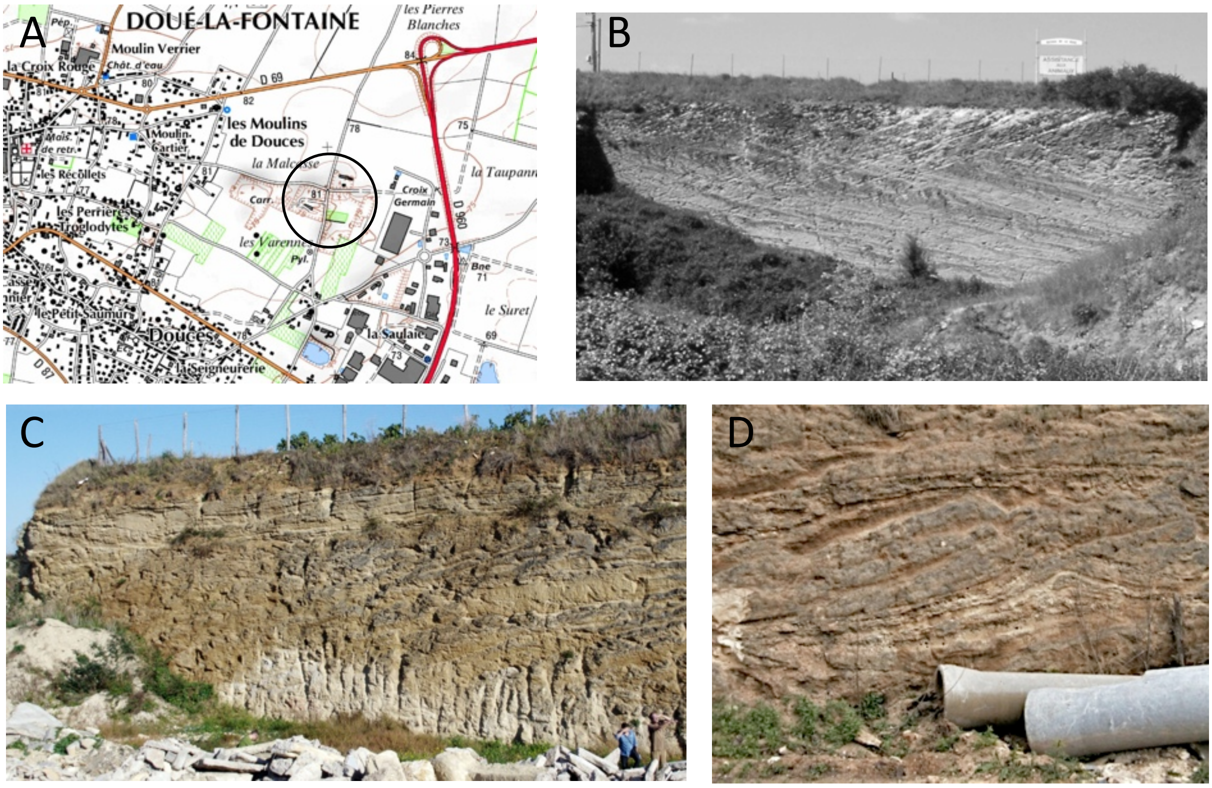
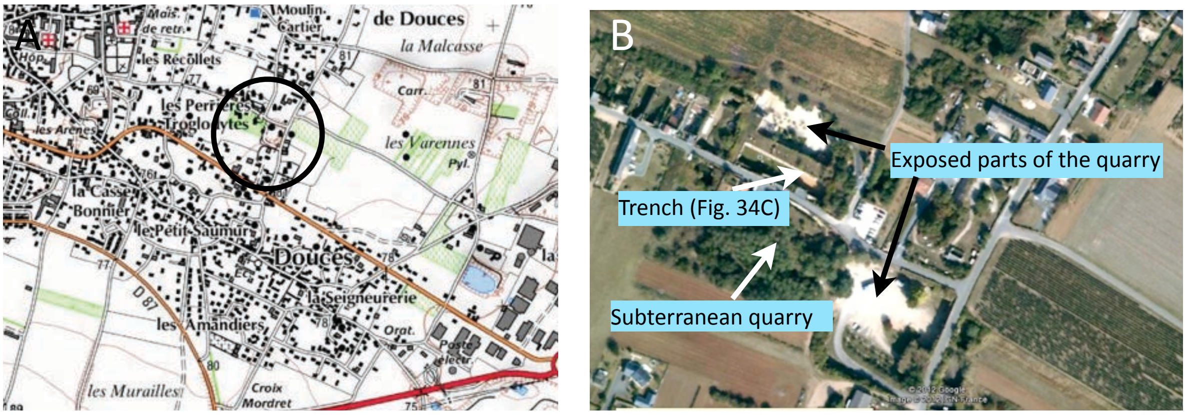


Fig. 33: A: Location map. B: Large-scale high angle oblique tabular bedding (be aware not to mistake for the bedding the lower angle scratch features made by quarrymen tools). C: At the top of the succession, note the large-scale hummocky beds. D: Close-up on soft deformation structures with the oblique masterbedding.

On the other side of the road (Fig. 33C and D), the outcrop delivers abundant remains of bryozoans, dominant bivalve debris and locally shark's teeth, which relate to a marine environment. Therefore, it is unlikely that the Glibert delta would be fluvial in origin. Also, the topsets are lacking. Instead, the upper part of the section shows sub-horizontal deposits with small inset channels or possible large scale HCS. These features are interpreted as a wave-influenced nearshore deposit following the formation of the submarine delta. They would indicate a sea-level fall. The material eroded by the waves would have been carried westward to the Arènes (stop 6) and Perrières (stop 8), which are deeper areas of the basin, where it would have been reworked as very large dunes.

Stop 8.- **Les Perrières : old subterranean quarries - the way to get 3D**

The Perrières Centre is a field of subterraneous quarries that was partially exhumed by a later excavation (Fig. 34). Preserved caves are used as troglodytic housing. This allows a 3D assessment of the stratigraphic architecture of the deposit.



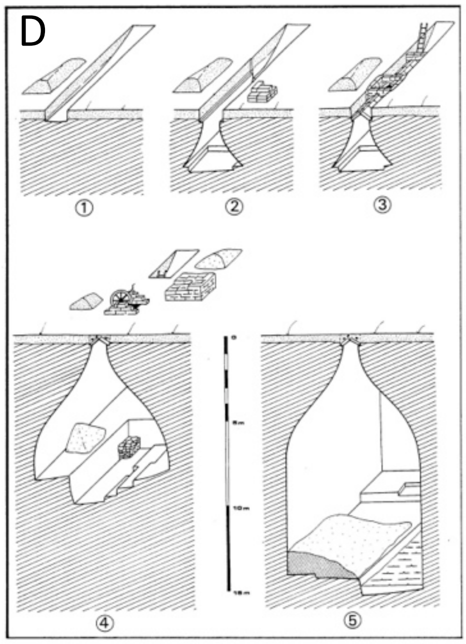


Fig. 34: A: Location map. B: Aerial photo of the site. C: Trench with the entrances of the quarries. D: How they did dig it...

The facies is almost the same bryozoan-rich calcarenite as in the Arènes site. The deposit is formed by the aggradation of small to very large tidal dunes (Fig. 35A). The tidal cycles are mostly recorded as foreset bundles in medium-sized dunes (Fig. 35B). A tidal origin of all the other large dunes observed in other places is inferred. Following the model of Allen (Fig. 35C), the architecture of the dunes suggest a high asymmetry of the tidal currents (strong dominant and weak subordinate currents). This general model is used as a basis to interpret the large to very-large compound dunes at Les Perrières. A strong tidal asymmetry, leading to a- or b-type dunes (Fig. 35C), is common in shallow tidal seaways where the tidal wave is deformed by complex bottom interaction.

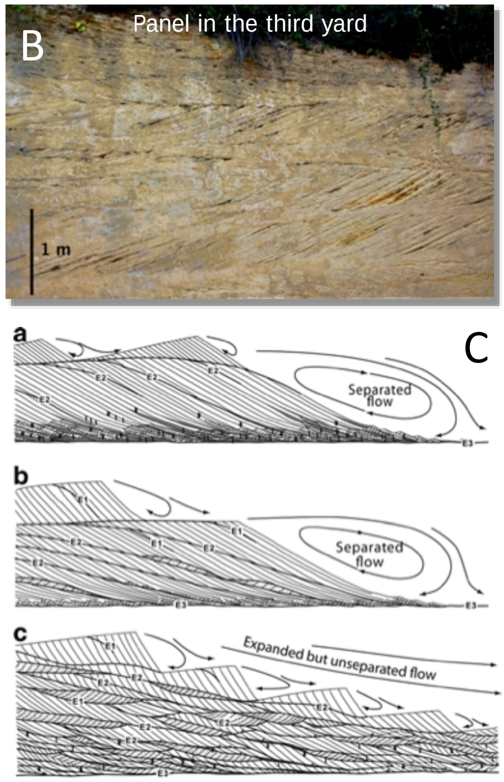


Fig. 35: A: Planar to arcutate, high angle cross-beds interpreted as the foresets of a very large dune. B: Rythmic bundling in the foresets of smaller-scale dunes. C: Variation of the internal structure of subtidal dunes in relationship with the asymmetry of tidal currents (after Allen, 1980, modified by Dalrymple, 2006).

Thanks to the numerous walls, the stratigraphic architecture of the deposit can be studied in 3D. At least 4 units are recognized, with distinct bedding patterns and an overall upward increase in amount and grain-size of lithoclasts and decrease in bryozoans (Fig. 36).

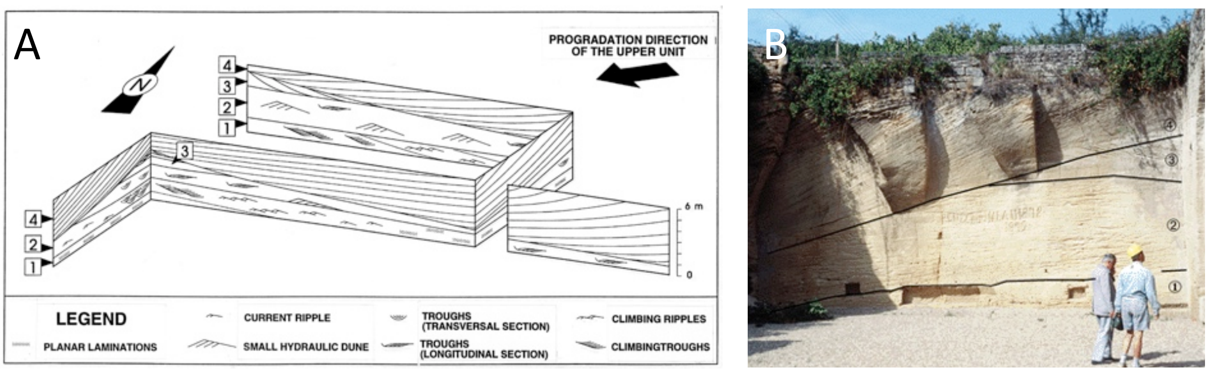


Fig. 36: A: Architecture of the Perrières deposit, from Biaggi (1993). Note the increase in size of the formsets generating the cross-beds from unit 1 to unit 4. B: detail of the erosion at the top of unit 3.

Units 1 to 3 are composed of cross-beds left by migrating dunes the size of which overall increases upward (Fig. 36). This is interpreted as an overall relative sea-level rise. The tidal dominance reverses between unit 2 and unit 3. The upper boundary of unit 3 corresponds an erosion surface which is refilled by unit 4 (Fig. 36B). Locally, unit 4 displays a progradational pattern above this surface. The coarser and more siliciclastic content, as well as the strong decrease in abundance of bryozoans in unit 4, suggests a more proximal setting, and therefore a sea-level fall between unit 3 and unit 4.

Conclusion

The Doué-la-Fontaine area offers several key outcrops which show the depositional setting evolution and the sedimentary dynamics of the Shelly Sands close to their southern onlap. An attempt of correlation of these sites is proposed (Fig. 37), owing to the fact that very little deformation occured after the Miocene and that the outcrops are relatively close to eachother.

This builds out the image of a coastal to shallow-marine domain undergoing a transgressive-regressive cycle of deposition. The tidal system benefitted from a bryozoan-dominated carbonate factory. This allowed the offshore tidal dynamics to be well recorded in this basin. More work would be required to decipher on the coastal complex paleogeographies and morphodynamic evolution of the deposit (subtidal deltas, tidal ridges?), based on paleocurrents and residual transport trends.

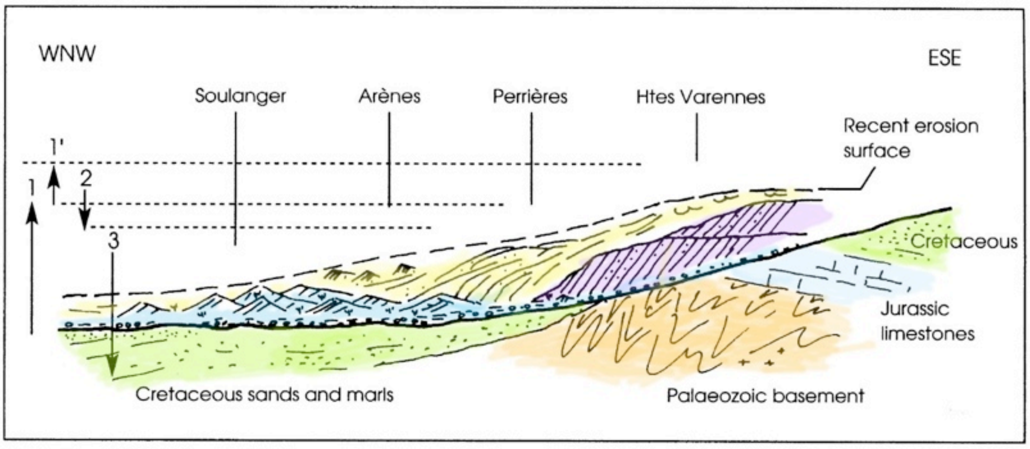


Fig. 37: Synthetic sketch of the stratigraphy of the Shelly Sands in the area of Doué-la-Fontaine.

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8th International Conference on tidal Environments
Caen, France, July 31 – August 2

Pre-conference field trip

The Bay of Somme
A wave dominated, macrotidal estuary

(July 29-30, 2012)

Leaders

Alain Trentesaux[1], José Margotta[1], Sophie LeBot[2] & Guillaume Villemagne[3]

With the sponsor of Oscar Savreux Quarry.

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General programme

Sunday, 29th July 2012
The Southern Bay. Following the gravels on their way North.

Tidal range: 5.65 m High tide at 08h37, 8.20 m Low tide at 15h53, 2.55 m

Stop 1: Ault
Introduction to the bay of Somme. Geological context.
Panorama on the chalk cliffs of Normandy, source of flint gravels.
Planning issues concerning the cliff retreat.

Stop 2: l'Amer du Sud. Cayeux-sur-Mer
Zone of gravel-spit high vulnerability.

Stop 3: Brighton-les-Bains.
Gravel spit evolution.

Stop 4: Pointe du Hourdel.
The modern end spit.
Changes in sedimentation rhythms
First view on the inner bay.

Stop 5: The Cap Hornu
The paleocliff and its foot sediments.
Sediments of the inner bay

Monday, 30th July 2012
The inner bay. Where the mud sticks to boots.

Tidal range: 6.30 m High tide at 09h55, 8.50 m Low tide at 17h00, 2.20 m

Stop 6: The seamen's chapel
Panorama of the bay at high tide from a Tertiary High.

Stop 7: Gravel quarries
Pleistocene evolution of the bay

Stop 8: Le Crotoy
How to slow down the sedimentation in the estuary?
The Northern end of the Bay

Stop 9: Le Crotoy -> Saint-Valery
Journey along the salt marshes
Use of the steam train to join the next stop

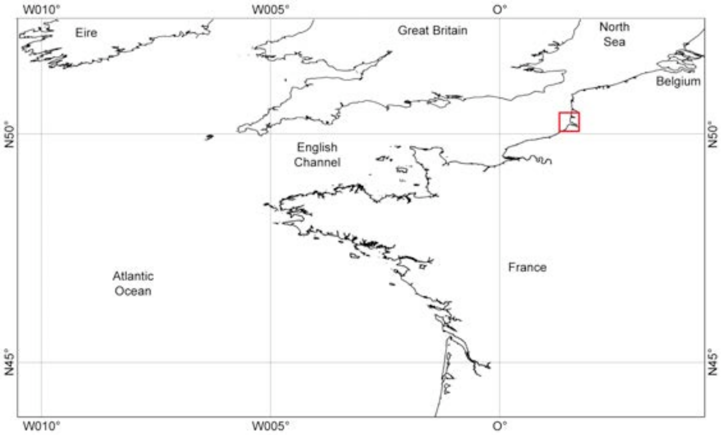
Stop 10: Saint-Valery
Inner-Bay sedimentation

Alain Trentesaux, Sophie LeBot and José Margotta have prepared this volume.

Abstract

The Somme Bay is a nice example of a macrotidal estuary developing characteristic and original features. The main objective of this field trip is to examine the modern sedimentation that occurs in the Bay of Somme as a function of the exposure to different hydrodynamic agents, such as waves, tides or river flows. As most of the European estuaries, this area is rapidly evolving, following a typical estuarine infill pattern (Chaumillon et al., 2010). In the context of eventual climatic changes and ecological realization, some environmental and planning issues will be also discussed.
The second objective consists in the presentation of the Holocene evolution of the Bay in relation with sea-level variations.
Two other field-trip guides can complete this booklet: Dupont et al., 1993, and Ducrotoy, 2004. The French Association of Sedimentologists (ASF) publishes both.

Stops



Road map displaying the proposed stops. IGN documents (www.geoportail.fr)
Red square: figure 1, next page.

1- Introduction

The Somme Bay is located in the North of France, in Picardy. It forms a vast embayment open to the NW in the Southern part of the coastal plain of Picardy (Figs. 1 and 2). The inner part of the estuary is continuously silted up and displays a mobile and fragile coast. Landscapes and sedimentary objects are very diverse: rapid changes of the estuarine channel courses, gravel bars to the South, sand beaches and dunes to the North. They are in front of a low-lying zone under the risk of flooding. Land reclamation has reinforced the natural accretion processes and led to deep modifications of environmental uses (e.g. fisheries or navigation).

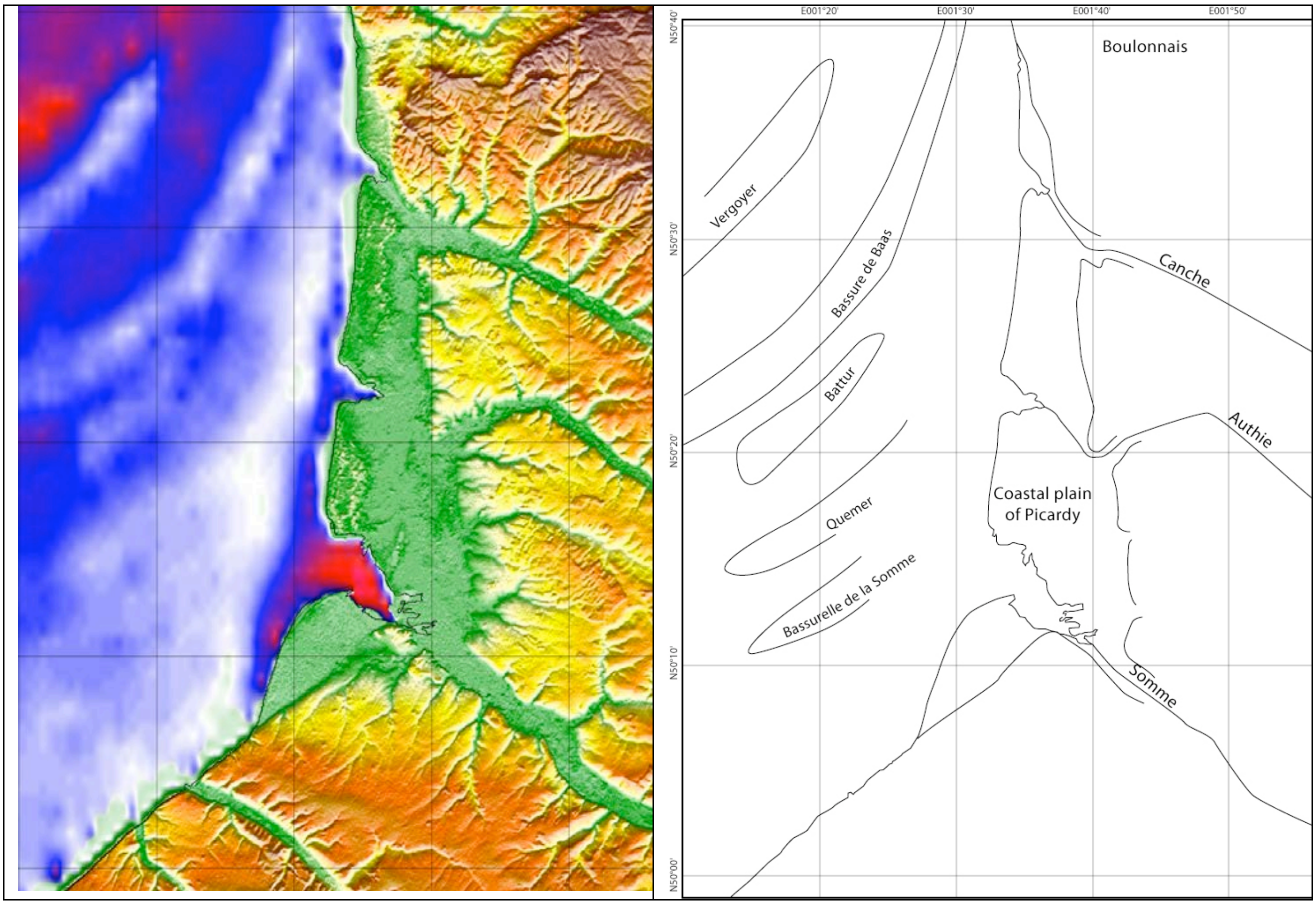


Figure 1. Digital Terrain Model of the coastal plain of Picardy. Distinction between the coastal plain in green and the relief from the Artois Chalk plateau is easily visible. Three main streams dissect the plain: the Somme, the Authie, and the Canche, respectively from South to North. The offshore zone displays some large 'tidal' sand banks. Topography from SRTM V4 (NASA, 2008) 90 m resolution. Bathymetry from ETOPO1 (NOAA, 2009) 1800 m resolution. Projection: Mercator, WGS84. Treatment: F. Graveleau, Géosystèmes, CNRS-Lille 1.

1.1- Geological framework

On a broad sense, the Bay of Somme is located in the Paris Basin North-western end. The Paris basin is intracratonic with a Mesozoic to Cenozoic infill history. The geological map displays three groups of terrains based on their age (Fig. 2). The substratum covering the entire zone consists in Cretaceous chalk. Outcrops display a wide range of chalks from Lower Turonian to Lower Santonian. The estuary lies on a large faulted syncline (van Vliet et al., 2000, Augris et al., 2004) at the origin of a large open bay that is filled with mostly marine sediments since the Cretaceous (Loarer, 1986).

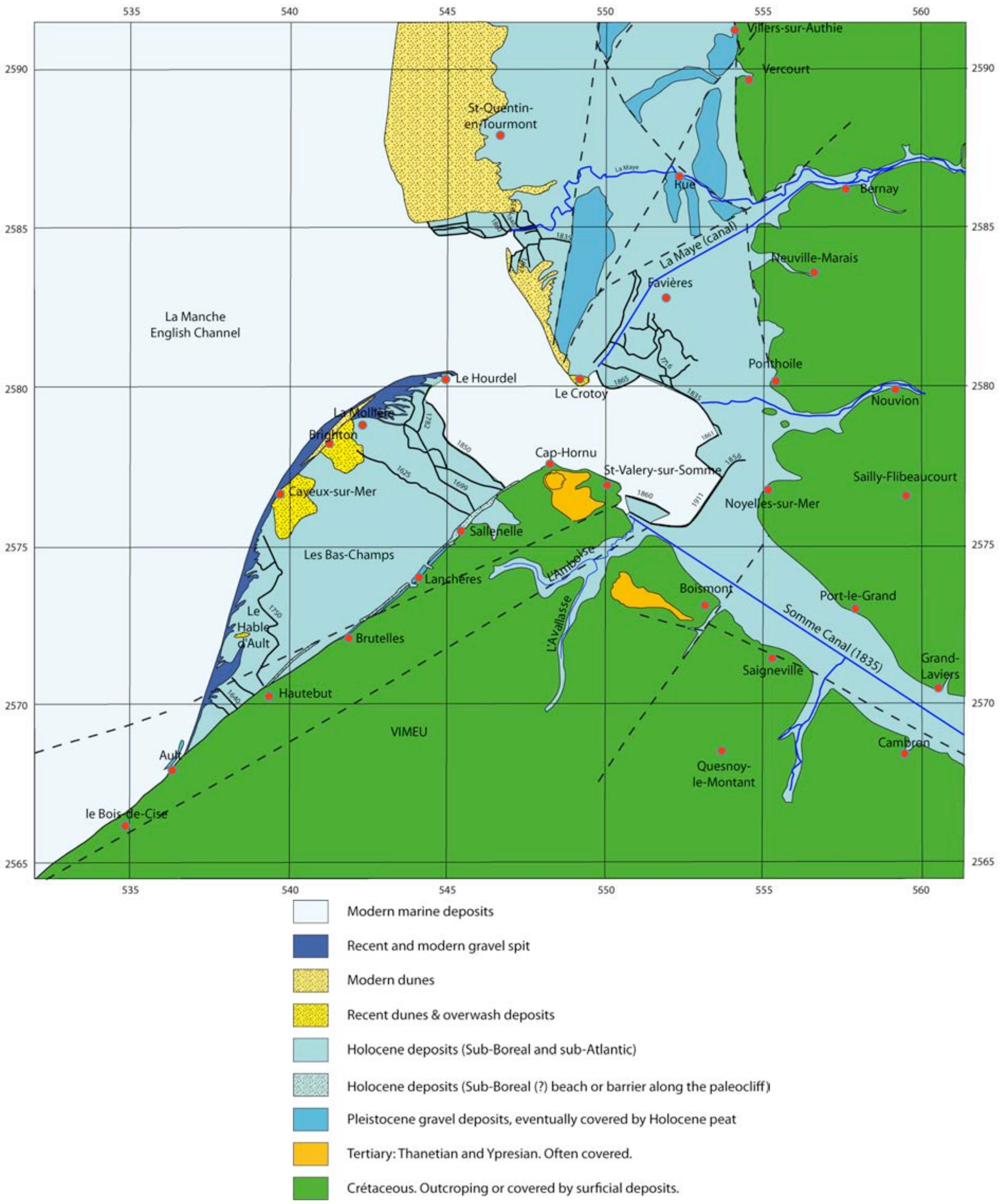


Figure 2. Geological sketch based on the two published geological map. (BRGM, 1981; 1985). Dashed lines refer to presumed faults or hypothetical Quaternary fault or flexure. Some dates have been reported along ancient seawalls.

Mesozoic rocks outcrop along the coast from Normandy (Le Havre) and constitute some well-know cliffs (Fig. 3). On a lithological point of view, except for Lower and Upper Turonian where the chalk is argillaceous, the chalk is pretty pure and displays a white colour. Flint layers are present throughout the series but their abundance is variable. The Cliff coast of Normandy ends at the southern extremity of the Bay of Somme. It continues as a paleocliff (ravines) separating the Artois plateau from the Pleistocene coastal plain.

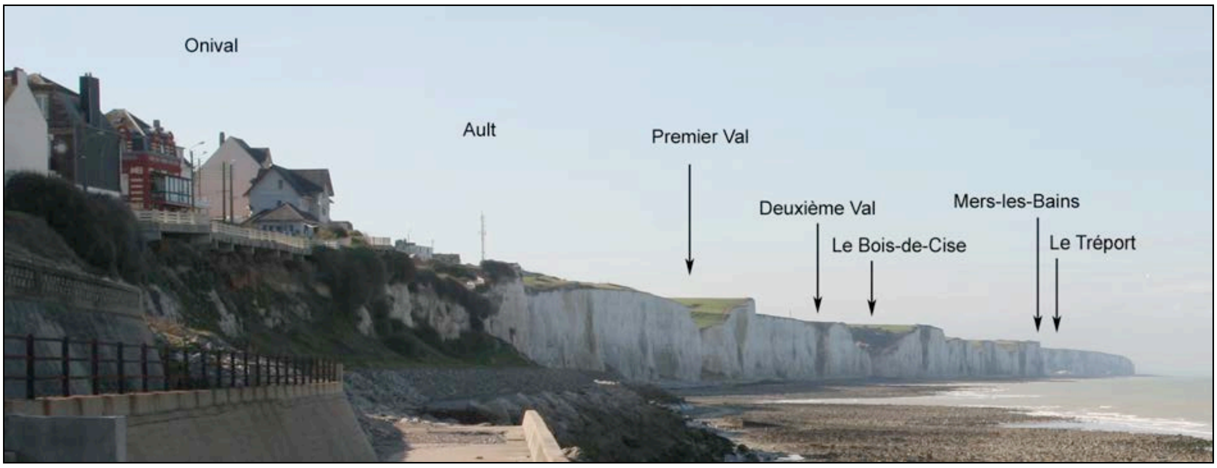


Figure 3. View from the beach of Onival toward the SW and displaying the active cliffs. On the shore appears the rocky platform. At the top of the cliffs, dry valleys appear. They are locally called 'val' or 'valeuse'.

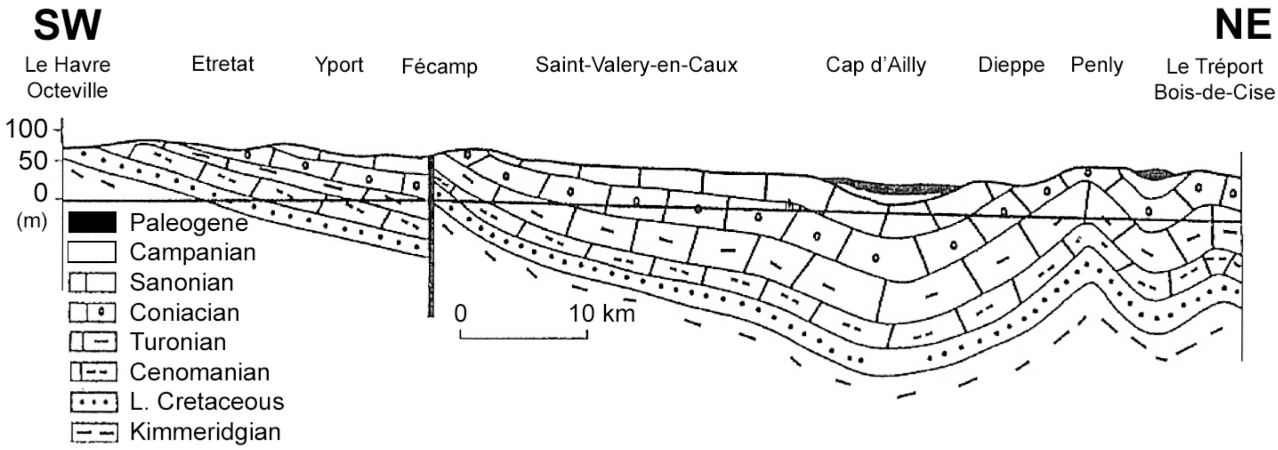


Figure 4. Simplified geologic section South of the Bay of Somme (Augris et al., 2004).

The second unit consists in Tertiary rocks well developed offshore. Inland it appears as witness buttes in the area of Saint-Valery. Around this locality, Thanetian sands and Lower Yresian sands, muds, silts and coquinas outcrop. These formations, characteristic of the Paris Basin, are mapped and described in Quesnel (1997) and Laignel et al. (2002).

The third unit is composed of Pleistocene deposits. These are of continental origin on the Plateau (e.g. loess; Lautridou, 1995), fluvial in the continental valleys, and mostly marine on the coastal zone. Ante-Holocene gravel bar formation and marine deposits will be further discussed in part 4.

1.2- Holocene context

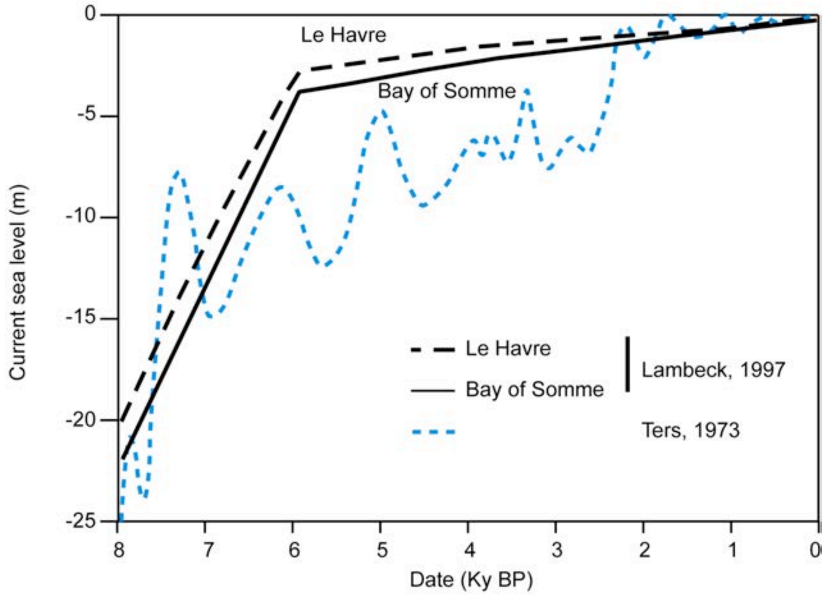


Figure 5. Sea-level evolution for the last 8 ka (Ters, 1973 and Lambeck, 1997)

Due to changes in sea-level (Fig. 5), the English Channel was strongly modified between glacial ages and interglacial stages, leaving a large fluvial plain between France and Great Britain. Different hypothesis have been proposed to explain how and when the opening of the Dover Strait occurred at each changes (Gupta et al., 2007, Destombes et al., 1975, Auffret et al., 1980, Lericolais et al., 2003.). One certainty is that a vast fluvial system was present at the place of the English Channel (See section 6). During low sea-level stages, the Somme was one among the many tributaries feeding this network.

1.3- Physical, modern framework

1.3.1- Tidal conditions

The Somme Bay is located in the Eastern English Channel, where the tidal regime is macrotidal and semi-diurnal. The tidal range reaches 9-10 m in spring conditions. It is the second location for tidal range value; after the Mont-Saint-Michel Bay (See this volume), along the Eastern English Channel (Fig. 6) Tidal values vary therefore strongly between 4.52 and 5.58 m (IGN 69, Tab. 1).

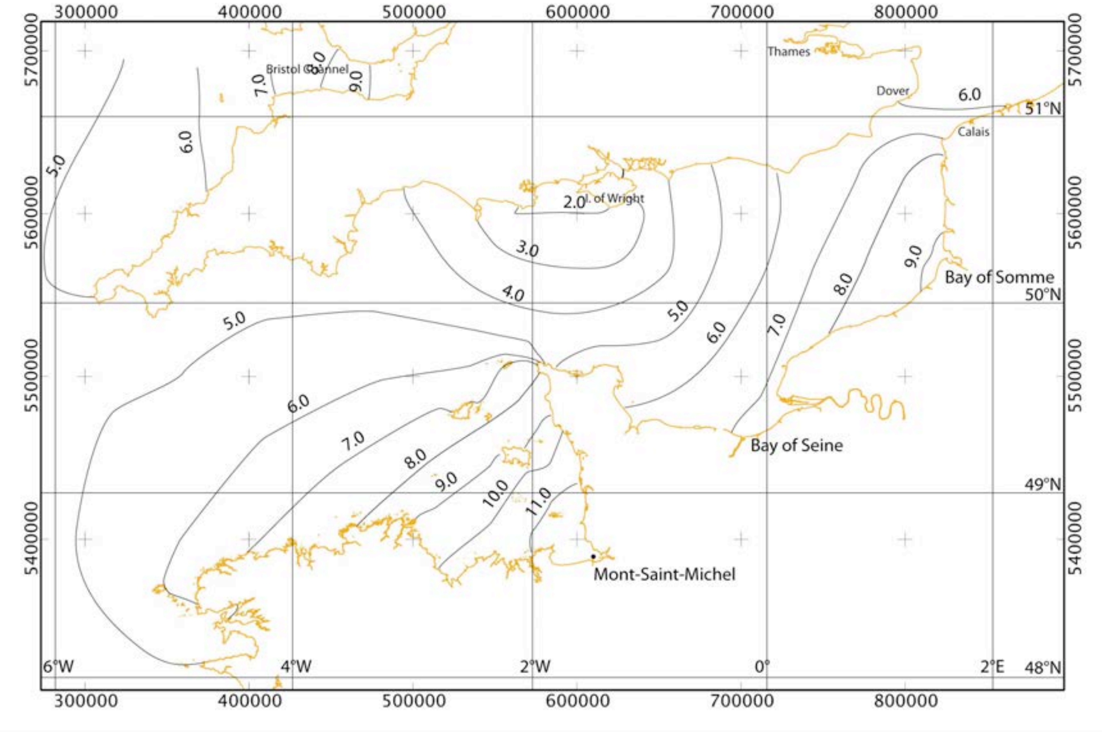


Figure 6. Mean spring tidal range along the English Channel. Data from Telemac model (EFDF-DRD, in SHOM, 2000).

Table 1. Characteristic tidal-levels at Cayeux (EPSHOM, 2001). *Coeff. (for coefficient) refers to a French tidal factor varying between 20 and 120.

	Coeff.*	Height (Nautical charts)	Height (IGN69)
Highest Astronomical Tide	-	10.55	5.58
Mean High Water Spring	95	9.85	4.88
Mean High Water Neap	45	8.00	3.03
Mean Water-Level	-	5.49	0.52
Mean Low Water Neap	45	2.95	-2.02
Mean Low Water Spring	95	1.20	-3.77
Lowest Astronomical Tide	-	0.45	-4.52

The tidal regime is flood-dominated (Fig. 7), although the flood effect tends to decrease due to the migration of the gravel spit toward the North that progressively closes the bay. The maximum current velocity measured offshore is lower than 1 $m.s^{-1}$ (Fig. 8). Tidal currents only increase when entering the bay (Fig. 9). The tidal cycle is strongly unbalanced: flood phase displays the maximal velocities, but only lasts 2 hours, whereas ebb phase lasts 5h45. Most of the bay is totally emerged during 4h30, and only the water flowing from the different rivers fill the inner channels of Le-Crotoy and Saint-Valery.

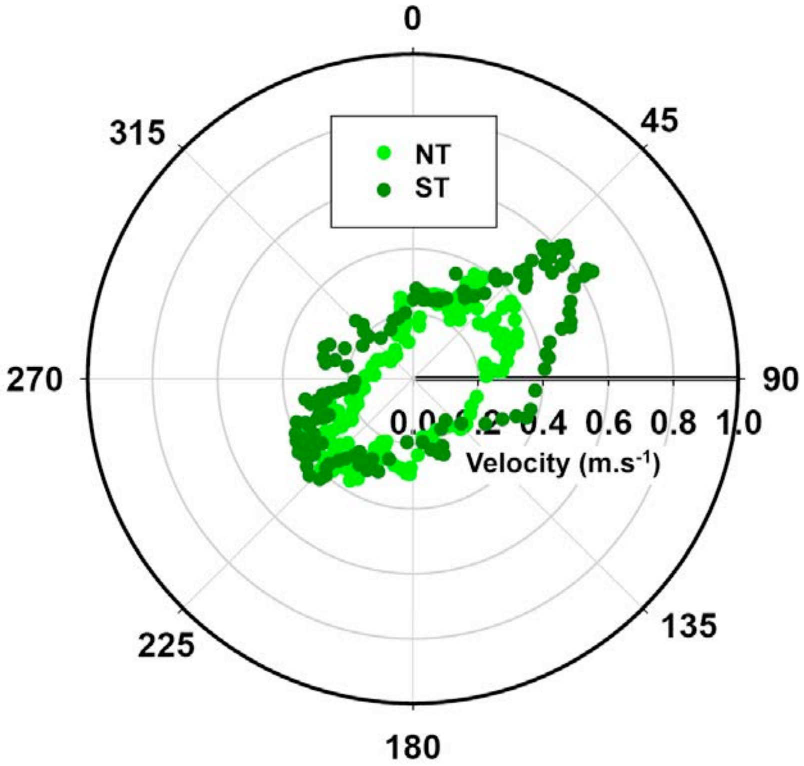


Figure 7. Averaged current rose measured between 2 m above the sea-floor and the surface offshore the Somme Bay at N50°09', E-001°17,5' (Ferret, 2011). NT refers to neap tides, ST to spring tides.

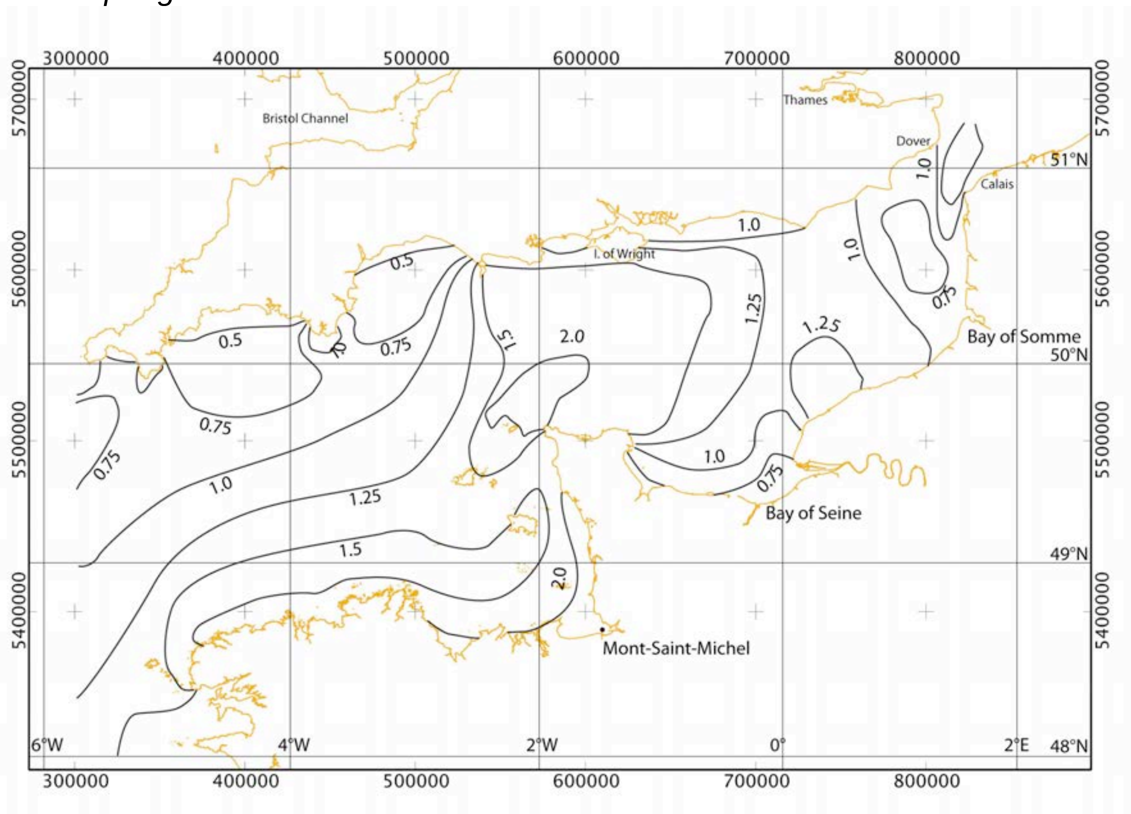


Figure 8. Maximum tidal current velocity in m.s^{-1} (from Larsonneur et al., 1982)

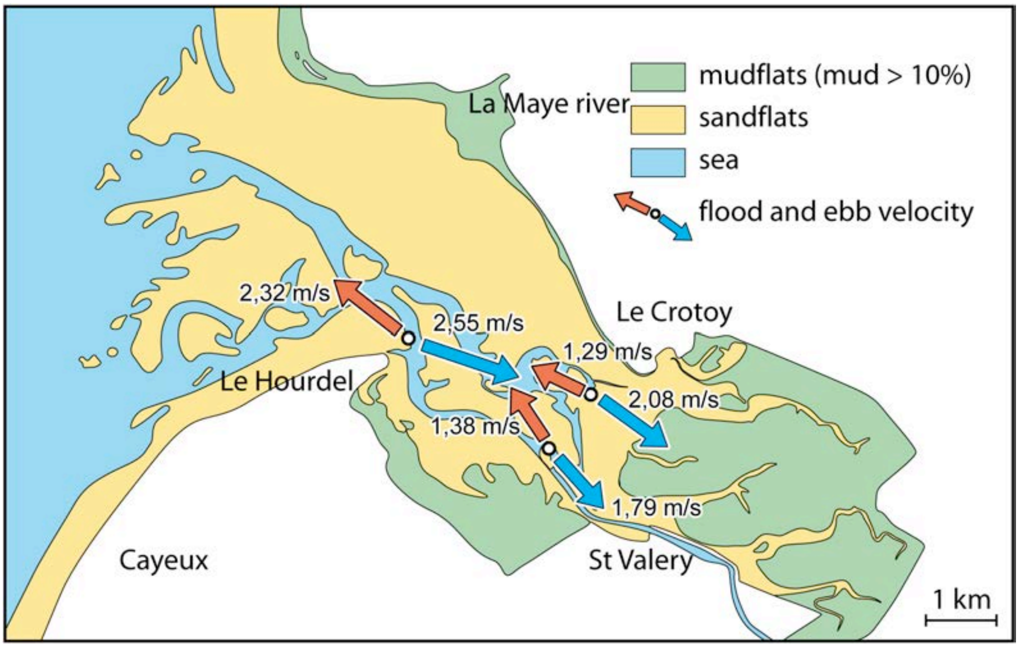


Figure 9. Flood and ebb currents in the Bay of Somme at low tide. Courtesy from François Baudin redrawn from diverse sources.

Fluvial discharge of the Somme River is low (5-60 $m^3.s^{-1}$, Dupont et al., 1993) with an annual mean around 32 $m^3.s^{-1}$ and does not counterbalance the flood dominance. The Somme is canalised between Saint-Valery and Abbeville, 15 km upstream and a sluice gate connect the river with the bay. Other small rivers only add 2.4 $m^3.s^{-1}$ to the estuary.

A tidal bore is sometimes observed (Fig. 10). Ancient documents seem to indicate that it was strong enough to help some vessels on their way to Abbeville in the XVIIth century (Cloquier, 2012), but nowadays, it is seldom observed and only helps canoe users in returning to Saint-Valery-sur-Somme from the Hourdel. Its position and occurrence strongly depends on tide conditions, but also on the movement of sand banks.

Figure 10. The tidal bore in front of the Jeanne d'Arc Quay. 10th September 2006. Picture: R. Grosléziat. The city of Le-Crotoy is visible in the distance.

1.3.2- Wind conditions

Local winds, generating wave agitation, dominantly Blow from west (50%, Fig. 11).

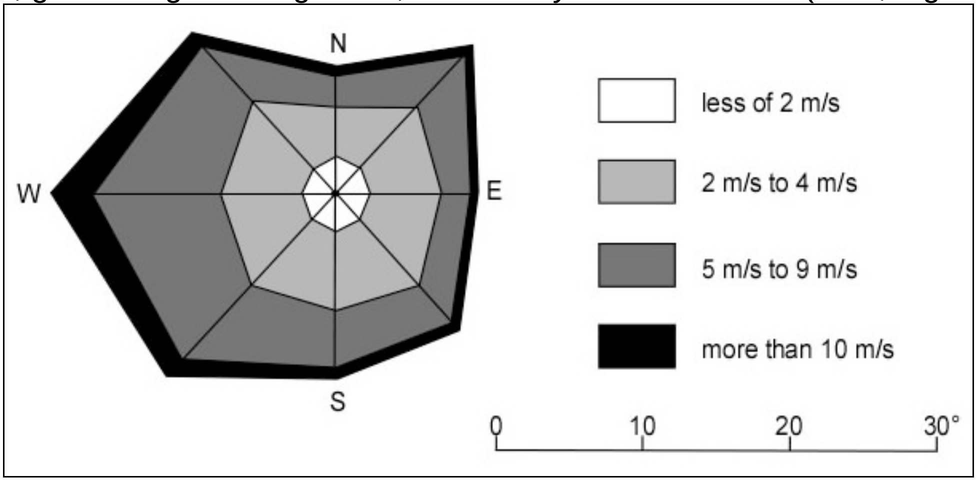


Figure 11. Wind rose measured at Abbeville (Clique et Lepetit, 1986).

1.3.3- Wave conditions

The Somme bay is concerned by a macrotidal regime, but waves act significantly due to high-energy wave conditions (Fig. 12). They first induce a strong littoral drift leading to the development of a gravel barrier at its mouth, on the southern part (Anthony and Héquette, 2007, Marion et al., 2009).

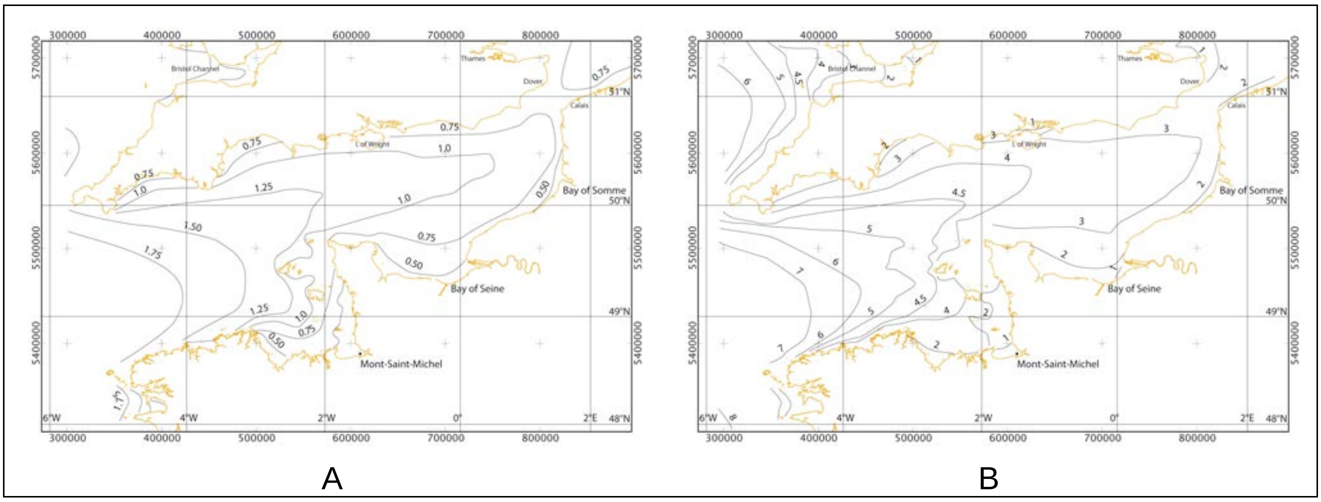


Figure 12. (Left) Mean of significant wave height and (Right) 99th wave-height percentile (Data from SHOM and Ifremer numerical models)

The local regime is characterized by low amplitude (< 1m) and short period (3-6 s) waves, mostly coming from the western sector (Fig. 13).

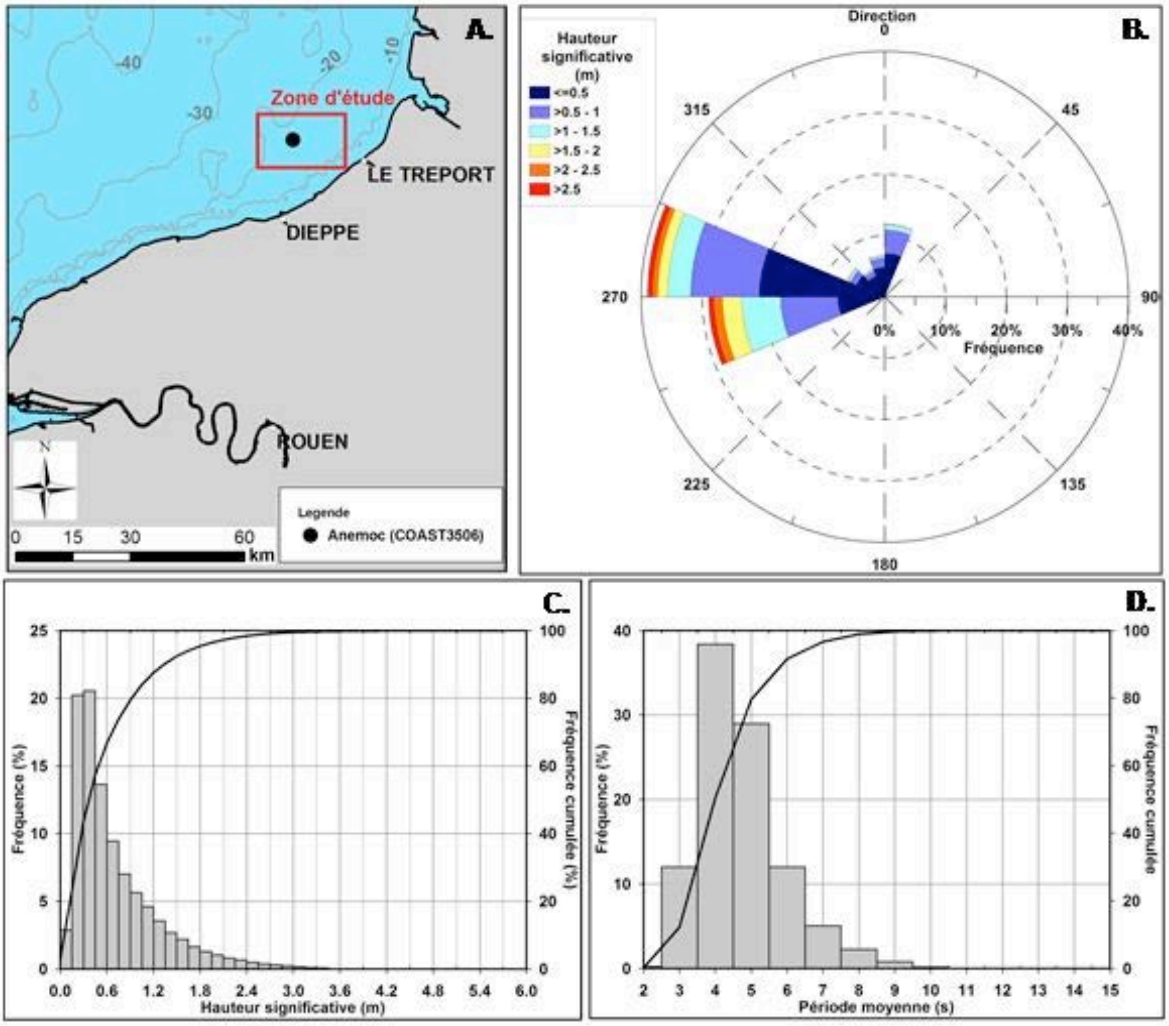


Figure 13: Wave regime offshore Dieppe on the period 1979-2002 (ANEMOC data, point Coast-3506). A. Location of the point Coast-3506 corresponding to the node of the analysed coastal grid; B. Wave frequency as a function of its provenance (hourly measurements); C. Histogram of the mean significant wave height; D. Histogram of the mean wave period (Ferret, 2011).

Table 2. Wave characteristics offshore the coast of Normandy at Paluel and Penly.

	Paluel	**Penly**
H_{max}:	80 cm	60 cm
$H_{1/3}$:	40 cm	35 cm
H_{mean}:	30 cm	25 cm
T_{max}:	8-9 s	6-7 s
T_{mean}:	6-7 s	5-6 s
Annual wave height:	5.6 m	4.1 m
Decennial wave height:	7.6 m	5.6 m
Centennial wave height:	9.6 m	7.3 m

Paluel and Penly are 44 and 32 km southward along the coast. The measurement points were located in about 15 and 10 m water depth (marine charts), respectively.

2 – The modern gravel spit

The characteristic that makes the Somme Bay different from the other Picardy estuaries is the presence of a long spit made of gravels while other are made of sands. This gravel spit started to form 2 500 years B.P (Dupont, 1981). In this section we'll discuss the source of these gravels, the spit dynamics, and some management problems linked to the spit. Stops 2, 3, and 4, along the 15.5 km-long spit will follow the gravel course due to residual wave action from their source along the Normandy cliff to the Northern end.

2.1- The source

The Somme estuary corresponds to the Northern end of the cliffs of Normandy (Fig. 14).

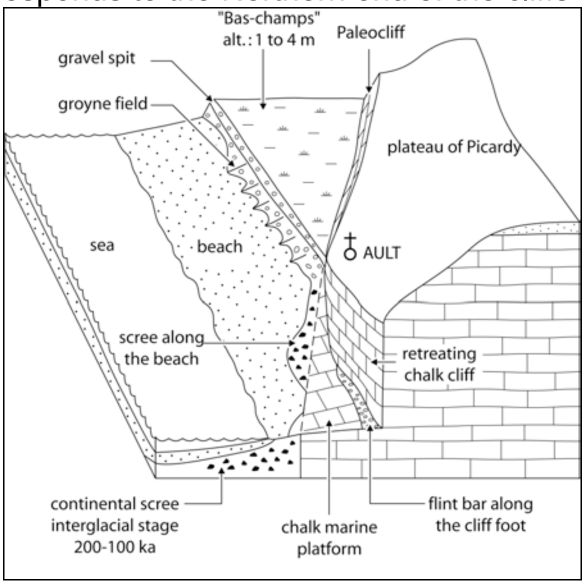


Figure 14. Sketch of the zone close to Ault where the retreating chalk cliff evolves in a paleocliff due to a longshore gravel spit. This spit isolates some lowlands, often reclaimed, locally called "bas-champs" that are under the level of spring high tides.

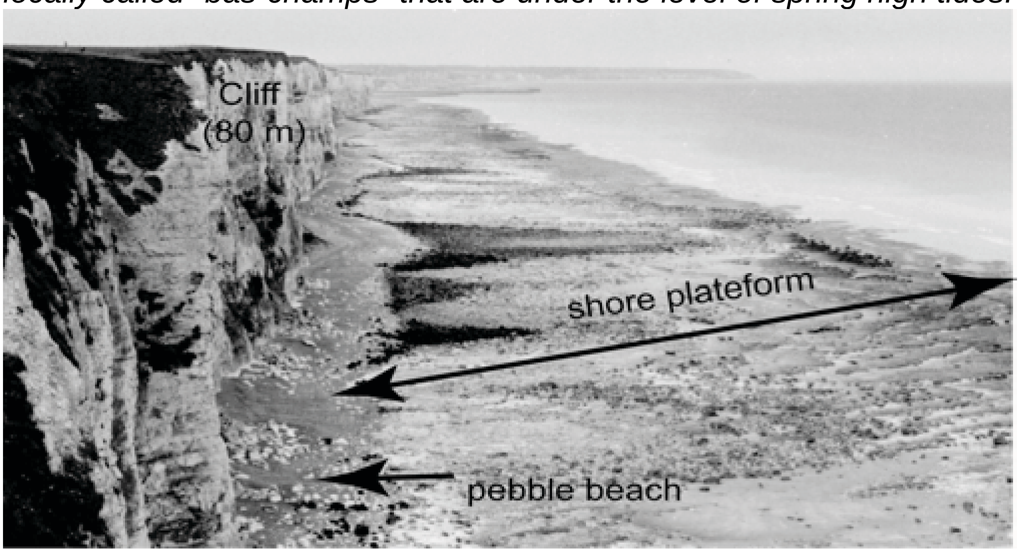


Figure 15. The cliff, shore platform, and pebble beach system from Haute-Normandy (Costa et al., 2002).

The Normandy cliffs are made of Cretaceous chalks characterized by their more or less high content of flint layers, especially high in Coniacian (8-13.5%), Santonian (10-16%), and Campanian (13.5-14.5%) layers (Laignel, 2003). Flints constitutes the source of the gravels: cliff retreat by marine action leads to landslides of cliff faces (Fig. 15). Blocks of chalk are quickly destroyed by the see in a couple of weeks (or months) but flints resists and are shaped in rounded gravels. Along this littoral, they are made of 98% of silica. On average, flint layers represent about 1 to 2% of the total rock volume. The cliff erosion feeds the gravel transport. Around Ault-Onival, the cliffs are made of Upper Turonian to Lower Santonian chalks (Fig. 16).

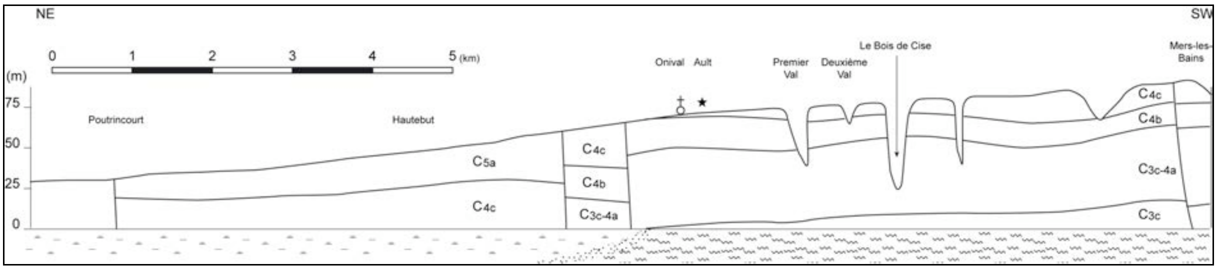


Figure 16. Synthetic geological cross-section along the nick-points located at the top of the cliff or paleocliff. Data drawn from the Geological map (BRGM, 1985). No indication of superficial deposits. The topographic profile does not take into account the profile along the coast, but the location of the highest nick-point. From Onival, Northward, the cliff is located inland and consists in a paleocliff. Vertical exaggeration: x20.
C3c: White or yellowish chalk with isolated or layered flints. Upper Turonian. C3c-4a: Chalk rich in flints. Latest Turonian and Lower Coniacian. C4b: White or yellowish, locally silicified chalk with rare flints. Middle Coniacian. C4c: White chalk with few flint layers. Upper Coniacian. C5a: White chalk with rare and small flints. Lower Santonian.

2.2- The gravel budget

The evolution of the gravel spit strongly depends on equilibrium between (i) gravel delivery from the chalk cliffs along the littoral of Seine-Maritime, from Antifer (Le Havre) to Ault, and (ii) the Northward gravel movement along the shore. If one of these two components is modified, the subtle equilibrium is broken and the gravel spit can grow or thin. This was the case along the Holocene, but has been strongly affected by human occupation along the shore, and often on the shore itself.

2.2.1- Incomes - Cliff-retreat rates

Mean cliff retreat is in the order of 0.21 m.yr-1 between Antifer and Ault on the 1966-1995 period (Costa, 2000; Costa et al., 2001). However, there is an important spatial variability of the cliff retreat rates (Fig. 17). The cliff evolution rhythm is strongly controlled by the lithology of the chalk (Costa et al., 2002) and also by the presence of obstacles such as harbour jetties, large landslides…

Close to the gravel spit, between Le Tréport and Ault, the shore retreat has been evaluated to 18 cm.year^{-1} on average. This retreat releases 2000 m^3 of gravel each year (Tab. 3).

Table 3. Annual gravel production along different sections of the Normandy coast (volume data from LCHF, 1986).

Zone	Length of coast (Km)	Annual production (m^3)	Annual production (m^3.km-1)
Antifer - Fécamp	38	6 100	160
Fécamp - Dieppe	55	10 400	190
Dieppe – Le Tréport	25	1 100	44
Le Tréport - Ault	7	2 000	285

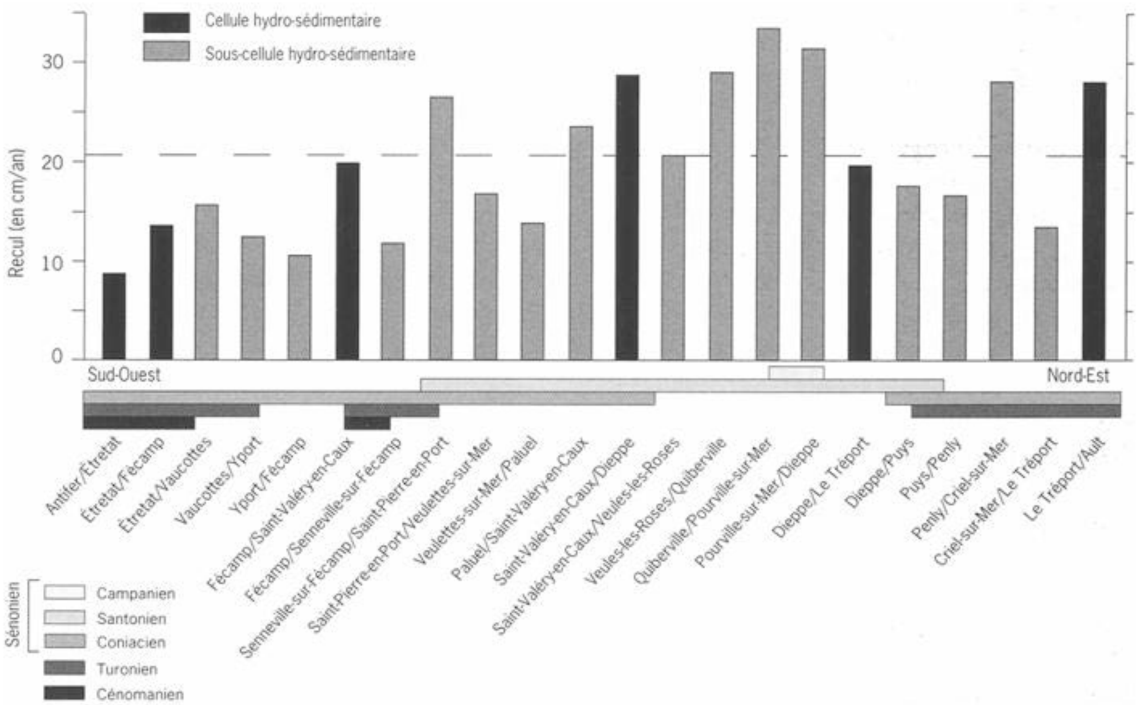


Figure 17. Cliff retreat rates by hydro-sedimentary cells (black) and sub-cells (grey). Horizontal bars inform on the stratigraphy of the chalk cliffs (Costa et al., 2000).

2.2.2- Outcome – Gravel extractions

From ages, gravel has been extracted along the shore to be used for house constructing or as road stones. These extractions remove gravels from the littoral budget and could have significant impact on the shore evolution. From the XXIst Century, this work decreased, and is now only located in strongly accreting zones. Gravel extraction on the beach itself is now restricted to manual collection for specific industrial use of the most rounded gravels (Fig. 18).

A B

Figure 18. Gravel exploitation. A. Gravel collection in the early days of the XXth century (in Bastide et al., 2010). B. Hand-made flint-stone gravel collection for industrial use. Picture: June 2010. Ancient cane baskets handled between the knees are still in use.

2.2.3- Gravel budget evolution along the shore

Gravel transit is directed eastward along the littoral of Seine-Maritime and Somme, from Antifer to Le Hourdel (Fig. 19)). Gravel budget has been estimated at the feet of the chalk cliffs and at the river mouths (LCHF, 1972; LCHF-BRGM, 1987). Except the low intake from retreating cliffs and official shingle extractions, these results allow establishing the main evolution pattern of the gravel budget along the coast (Costa, 1997). These studies suggest a generalized decrease of shingle beaches between Fécamp and Le Tréport. Several harbours and nuclear plant jetties especially disrupt the gravel transit and keep gravels on the western upstream-drift side of the jetties (Fig. 19 and 20).

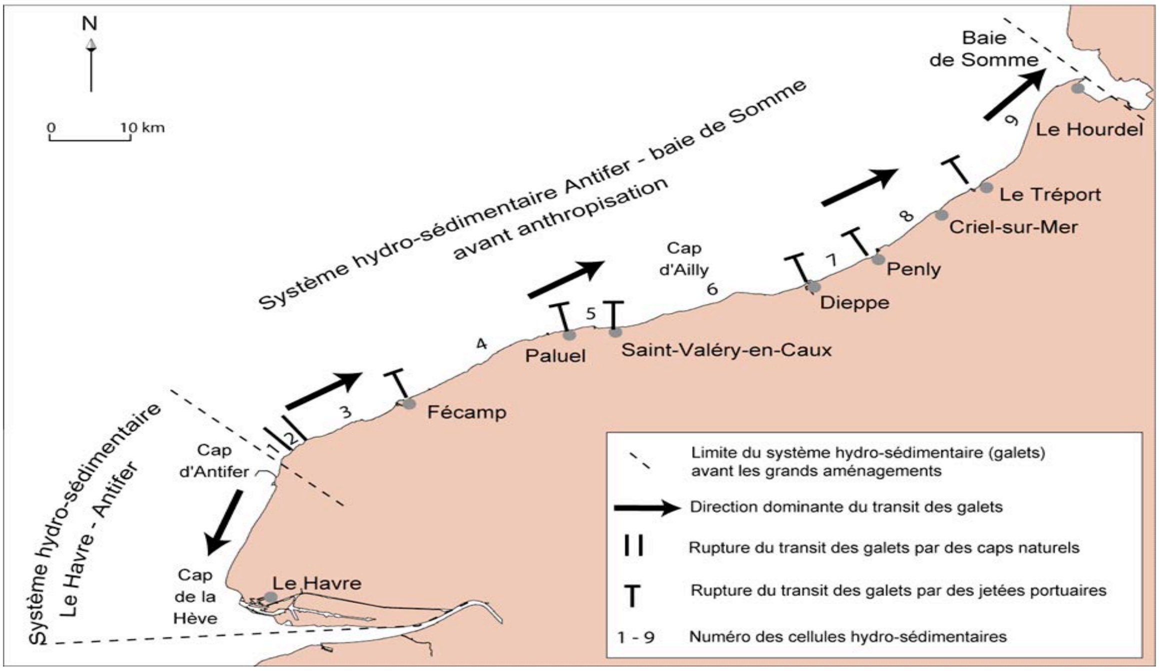


Figure 19. Direction of the resulting transit of gravels and sedimentary cells along the littoral of Seine-Maritime (Augris et al., 2004). Arrows: dominant direction of the gravel transit. Rupture in gravel transit by: (i) natural capes (double line), or (ii) harbour jetties ("T" symbol). 1-9: number of the hydro-sedimentary cells.

Figure 20. Volume of gravel along the shore between Mesnil-Val and Le Tréport (Costa et al., 2000).

2.3- Gravel spit migration

The gravel spit is characterised by various longitudinal dynamics, function of gradients in the littoral drift, typical of "open" spits (Orford et *al*., 2002; Anthony, 2009). Several studies have contributed to characterize and quantify the secular to decennial dynamics of the gravel spit (Briquet, 1930; Dallery, 1955; Costa, 1997; Dolique, 1998; Costa et al., 2000; Bastide, 2011; Fig. 21). Two spots are favourable to illustrate how the sand spit is moving north. The Northern end offers a scenic point on the Bay. The other point, between Ault and the end, is located at Brighton-les-Bains. At this locality the active gravel spit is bordered by a series of ancient ridges.

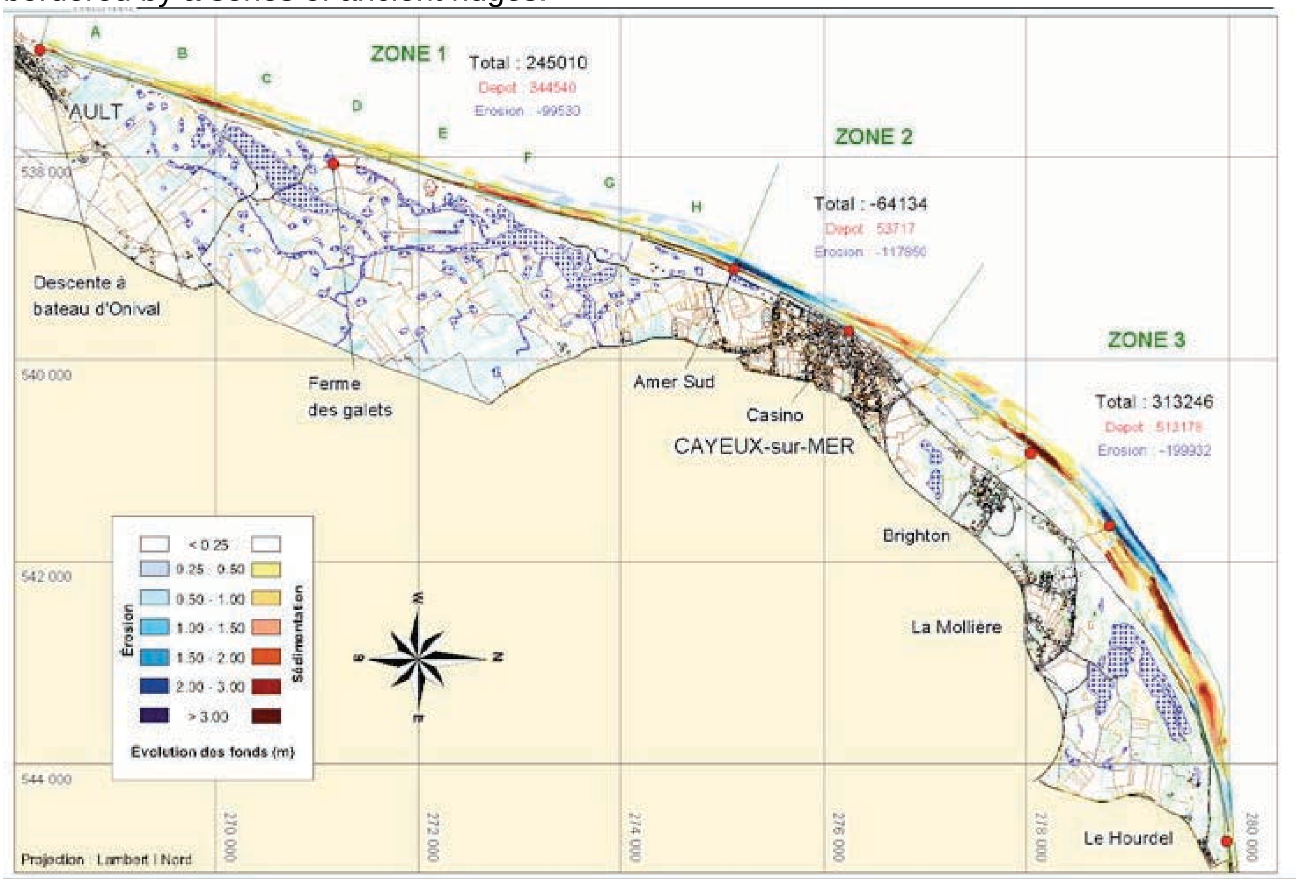


Figure 21. Volumetric balance of the gravel spit for the 05/1994 – 10/2001 period (Bastide, 2011).

2.3.1- Le Hourdel. The final spit

The end of the spit is rapidly evolving. The city of Le Hourdel moved to the coast during the XVIIIth century as some fishing vessels started to stop there, maybe due to increasing silting of the inner bay. Thanks to this occupation, many maps are available and have been compared (Fig. 22 and 23). They allow following quite accurately the evolution of the final spit. Nowadays, the spit migration is evaluated to 2 m.year^{-1}, and the gravel budget reaching the extremity of the spit is estimated to 4.000 m^3.year^{-1} (Bastide, 2011).

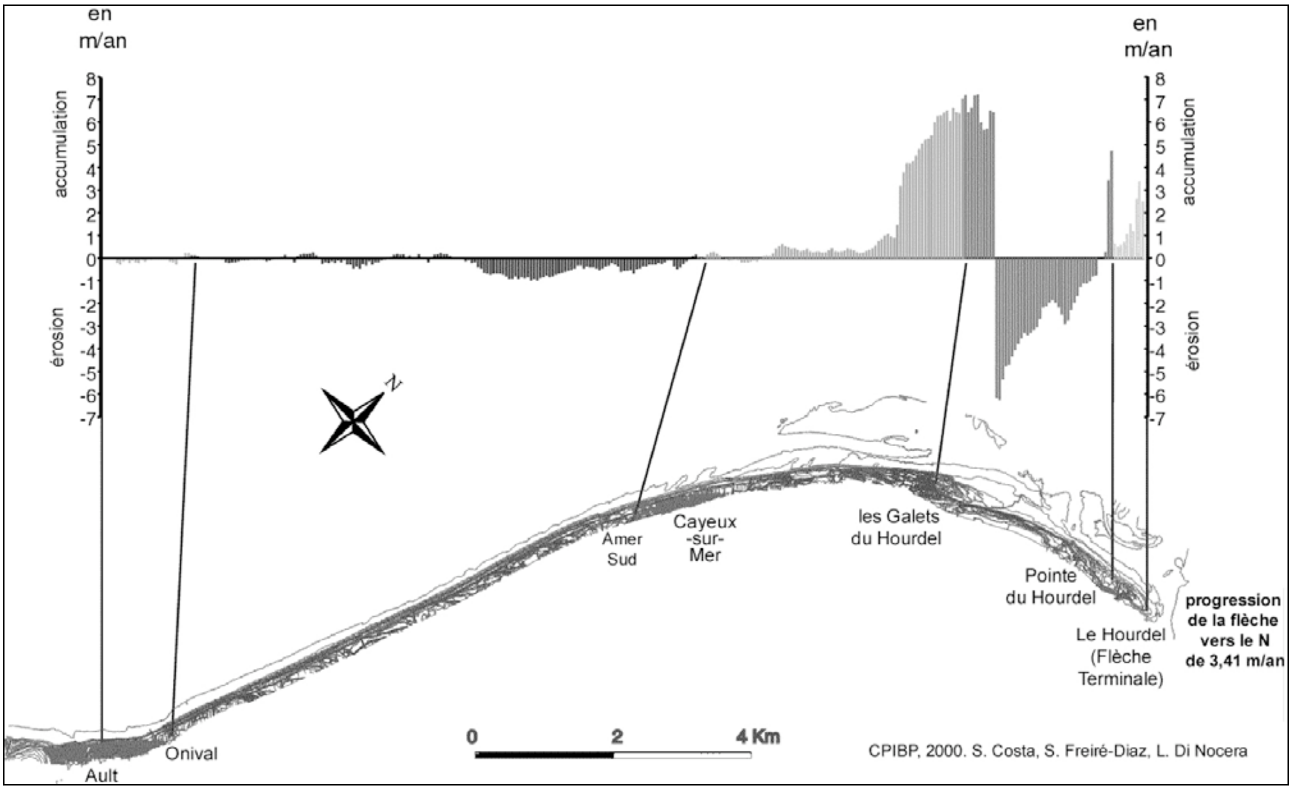


Figure 22. Coastline dynamics on the 1961-65/1999 period, between Ault and the Hourdel every 50 m (Costa et al., 2000).

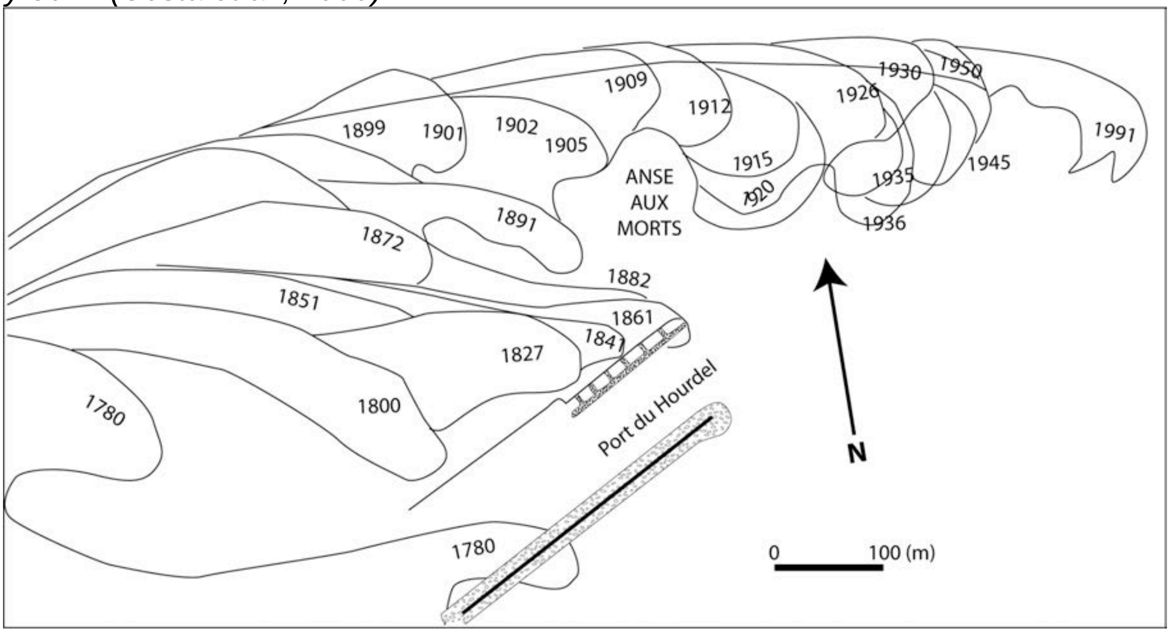


Figure 23. Two hundred years of evolution of the final spit. Work from diverse sources (Dallery, 1955 and Dolique, 1991) and partly based on photogrammetric data.

2.3.2- Brighton-les-Bains

Close to the lighthouse, a small track is perpendicular to the shore and crosses first a dune, then a series of gravel ridges. At that place, the movement of the gravel bar is exceptionally high, with advance rates reaching 120 m.year^{-1} (Fig. 24; SOGREAH, 2005).

Different geomorphologic features allow understanding the evolution of the gravel spit in a context of high gravel income: hooks, overwash deposits, silting in the runnel, dune installation, plant evolution… A cross-section illustrates how this place evolved recently (Fig. 25).

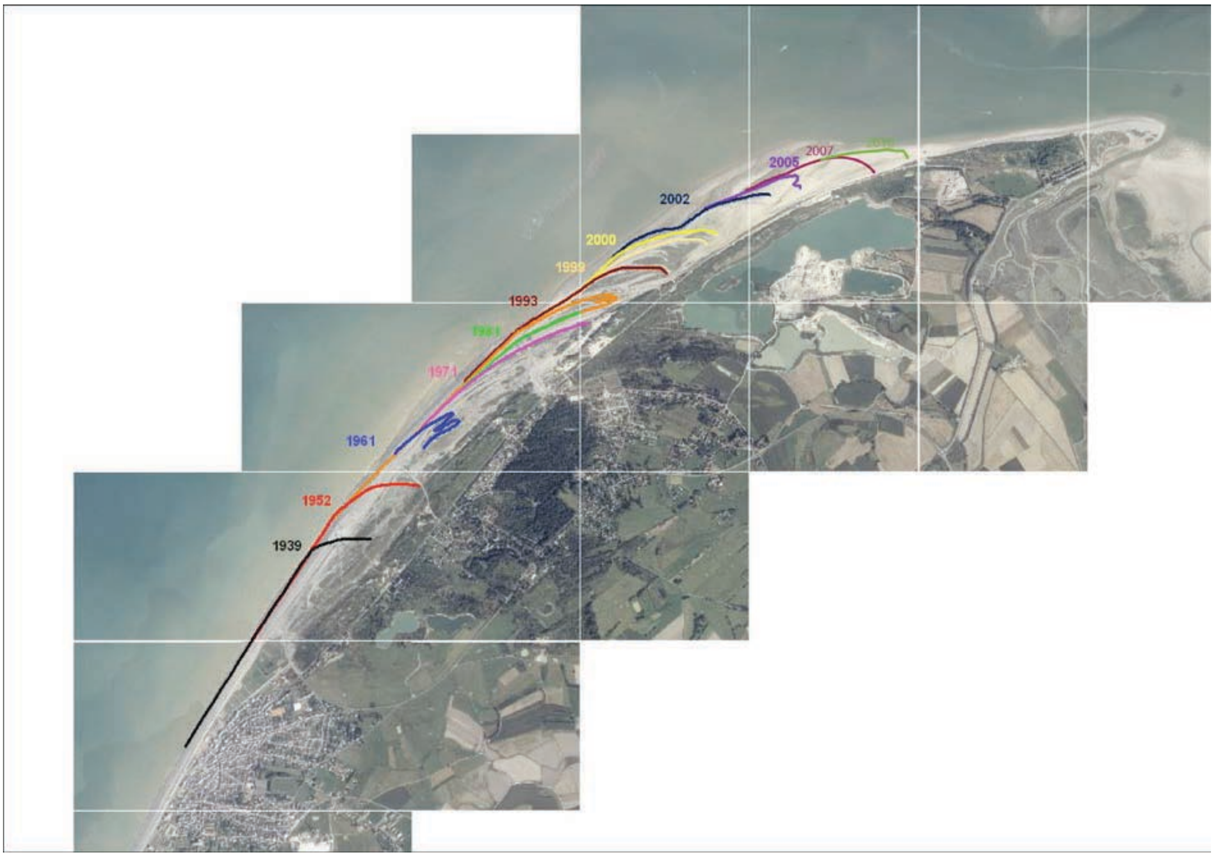


Figure 24. Evolution of the gravel spit North of Cayeux-sur-Mer between La Mollière and Le Hourdel between 1939 and 2005 (SOGREAH, 2005, Bastide, 2011). Aerial photograph © ORTHOLITTORALE 2000

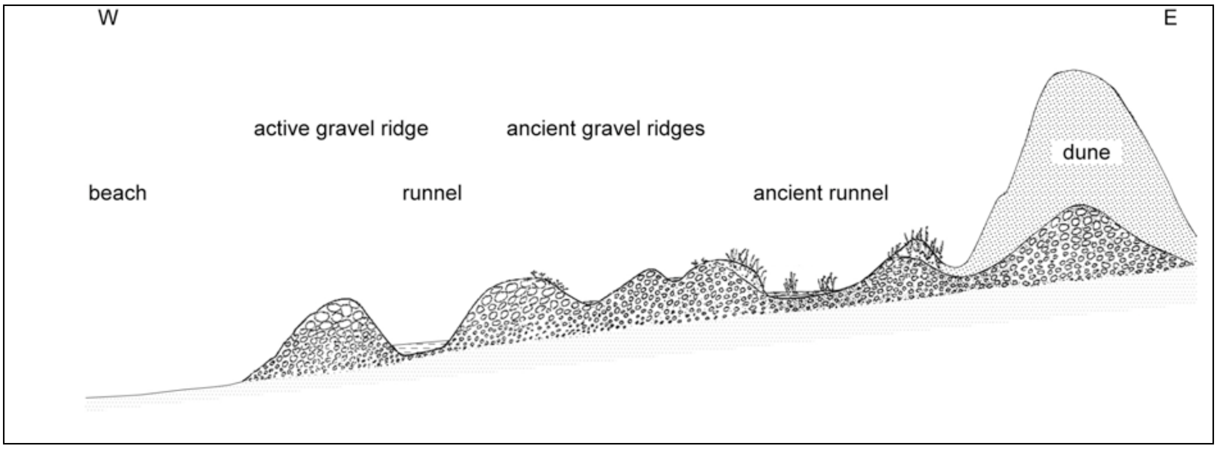


Figure 25. Cross-section around the Brighton lighthouse. (From Wiber, 1980)

2.3.3- Gravel migration rates

Even if the final position of the spit allows evaluating a NE-prograding velocity, the movement of individual gravel is still poorly known. This is however important for management reasons, especially in low-budget areas such as South from Cayeux-sur-Mer. Different experiments were conducted along the shore to measure the current and wave fields and to follow individual gravels (Fig. 26).

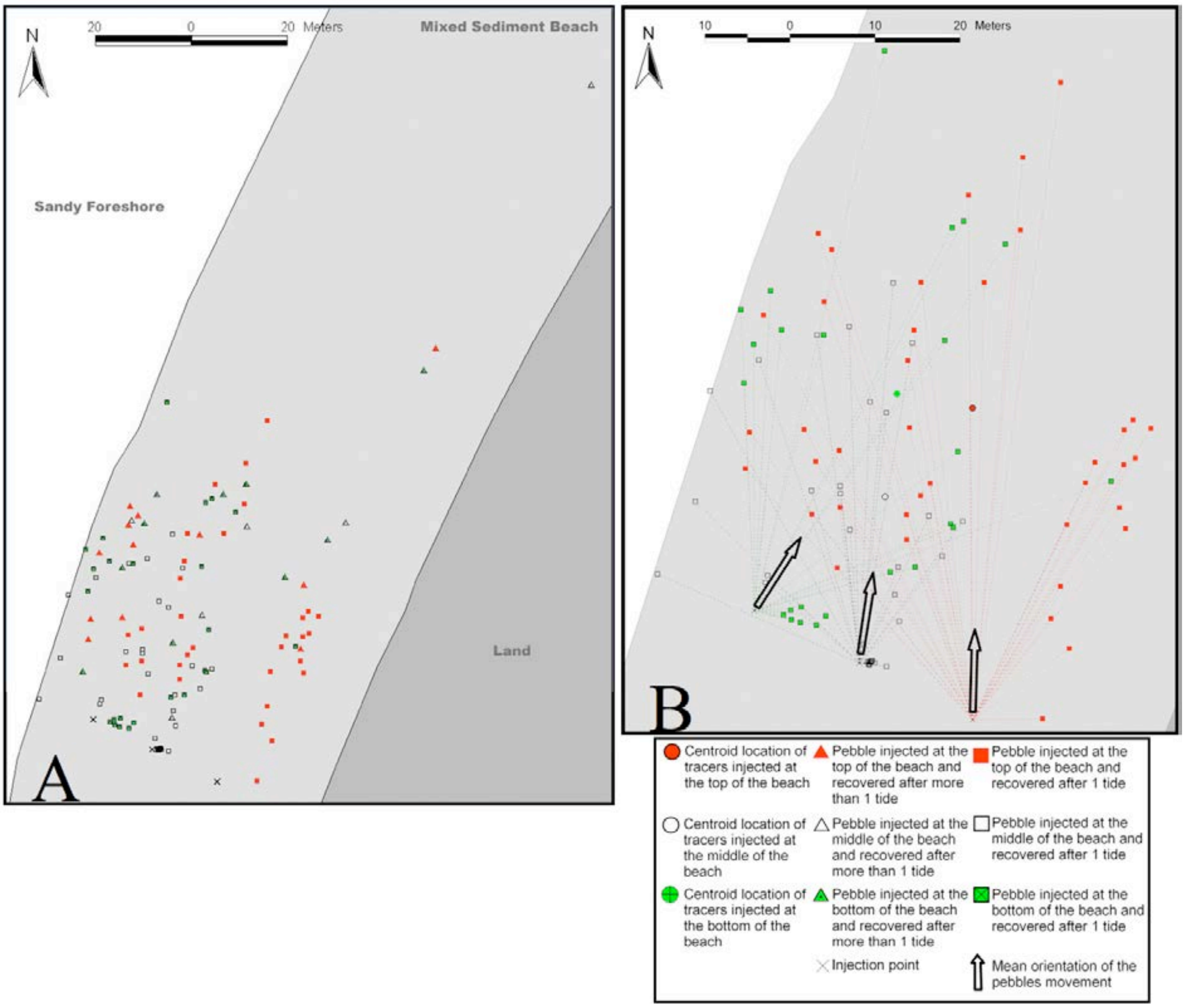


Figure 26. Movement of tracer pebbles deployed on the 8th of November 2005 close to the city of Cayeux-sur-Mer. A. Scattering observed over the whole survey period. B. Scattering observed after one tide (J. Curoy et al., U. Sussex, personal communication).

For this experiments it can be noticed that the movement of pebbles depends on their location at different levels of the bar. Material coming from the upper part tends to migrate seaward and further downdrift than the other parts of the beach. This is interpreted as linked to a continuous swash action from the high tide to the low tide (Curoy et al., U. Sussex, Pers. Comm.). Pebbles from the lower beach migrate upward, although alongshore transportation dominates. It is suggested that these movements are explained by the combined effect of groundwater flow, swash flow and breaking-wave action. Pebbles from the middle part of the beach move hanks to the same processes, but tend to migrate down the beach. For all situations, the alongshore drift is significantly greater than cross-shore movements.

2.4- Human impacts related to the gravel spit evolution

The gravel spit constitutes a natural protection against submersion for the polders of the 'Bas-Champs'. Part of its recent evolution seems to be strongly linked to human impacts along the shore both at the source of the gravel bar, and along its way North. Around Onival, a resort station developed in the early 1900's (Fig. 27). A casino was built on a natural riprap at the foot of the cliff (Fig. 28). This cliff was then eroded following the general cliff retreat characterizing the Normandy coast and the casino disappeared.

A B

Figure 27. Two postcards from the early 1900's. A. The beach of Onival. The church and some houses are still visible. B. The cliff at Ault. The casino 'on the beach' and most of the front-row houses have been destroyed.

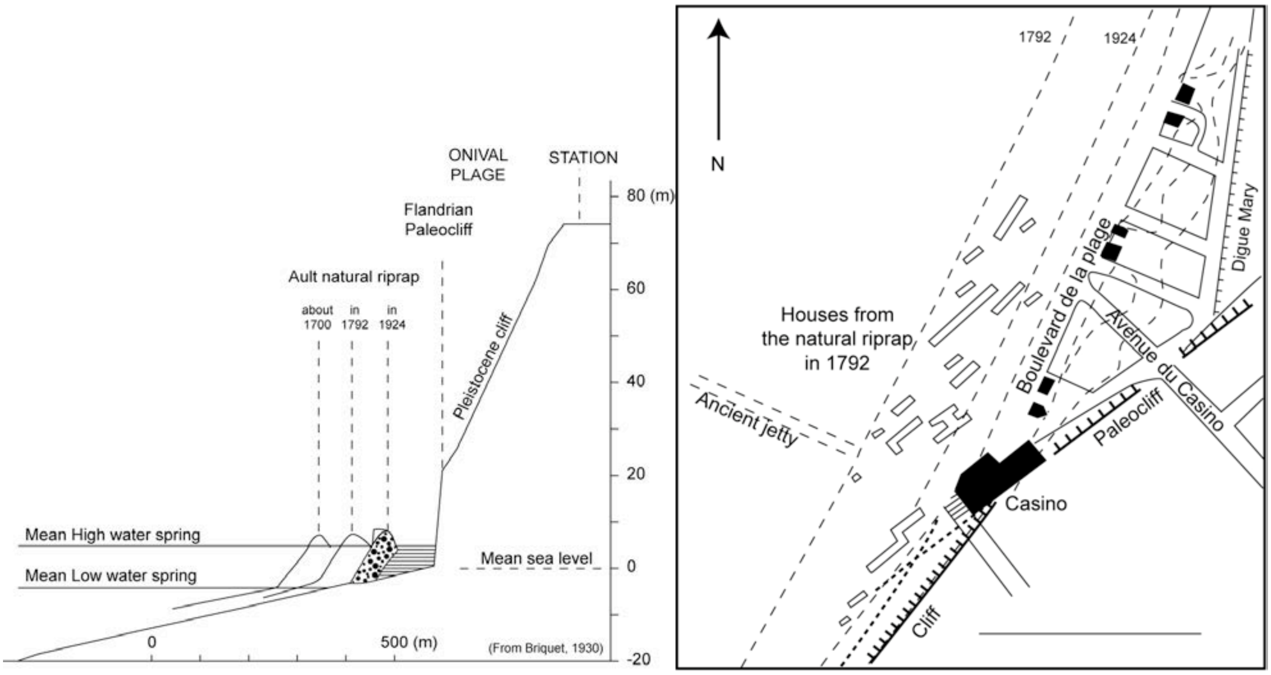


Figure 28. In the 30's, there was still a platform at the foot of the Flandreau cliff at the foot of which a casino was present. A gravel bar acting like a natural riprap protected this platform. (Drawing adapted from Briquet, 1930). The drawing indicates a series of houses on this platform and indicates they were still present in 1792. Some papers indicate that they were destroyed during a severe storm in 1579 or 1583. There is also an ancient map from 1667 displaying windmills and houses.

From 1835, different series of groynes were installed at the foot of the cliff, but also along the spit to limit the gravel drift and help protecting the shore (See e.g. Dallery, 1955, and later works).

At some places, such as Onival, heavy solutions were decided and built between 1981 and 1986 (DREAL, 2011) to protect the cliff from sea undermining, but also aerial erosion due to continental-water seepage (Fig. 29).

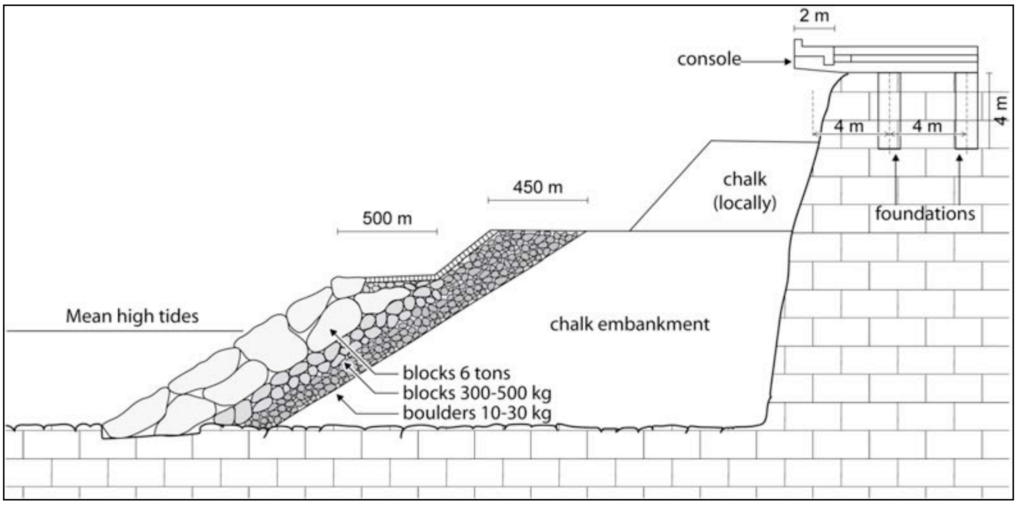


Figure 29. Last coastal defence built in the 70's and regularly maintained at Onival. DDE document.

The decrease of the gravel budget, observed since 2 centuries, has lead to the erosion of the spit in some places (e.g. Cayeux) (Costa, 1997; Costa et al., 2000). The gravel spit has been strongly comforted by groynes and spikes at the beginning of the 80's between Ault and Le Cayeux, in order to limit spit erosion and protect the area of the Bas-Champs from marine submersion. In strong relation with human constructions, the gravel spit can be divided in 3 sectors displaying a contrasted dynamics, from its proximal area, close to the gravel source to its distal area at the entrance of the Somme Bay: (1) the sector between Ault-Onival and the South of Cayeux-sur-Mer, with 83 spikes, built to trap the gravels, in response to an important erosion (0.1 to 1.8 m.yr^{-1} on the 1800-1991 period; Dolique 1998), (2) the sector of Cayeux-sur-Mer, in erosion, where frequent nourishments are conducted to avoid spit rupture and make possible an artificial gravel transit, and (3) the sector of Brighton-Le Hourdel, in progradation (150 m between 1780 and 1930; Briquet, 1930), where gravel extractions are still active (e.g. 'Silmer' and 'Delarue' concessions).

3- Intertidal sedimentation

In the bay of Somme, the sedimentation is driven by sediment budget, strongly linked to sea level rise and, much later, human influence. Modern intertidal sedimentation can only be seen on a quite limited area compared to the ancient size of the Bay of Somme. This is due to long-term continuous infill of the estuary as for many estuaries along the English Channel (Tessier et al., 2011). Land reclamations (embankments, polders) reinforce the natural accretion process (Bastide, 2011). The infilling then leads to important modifications of environment uses (e.g. fisheries, navigation).

3.1- Modern sedimentation

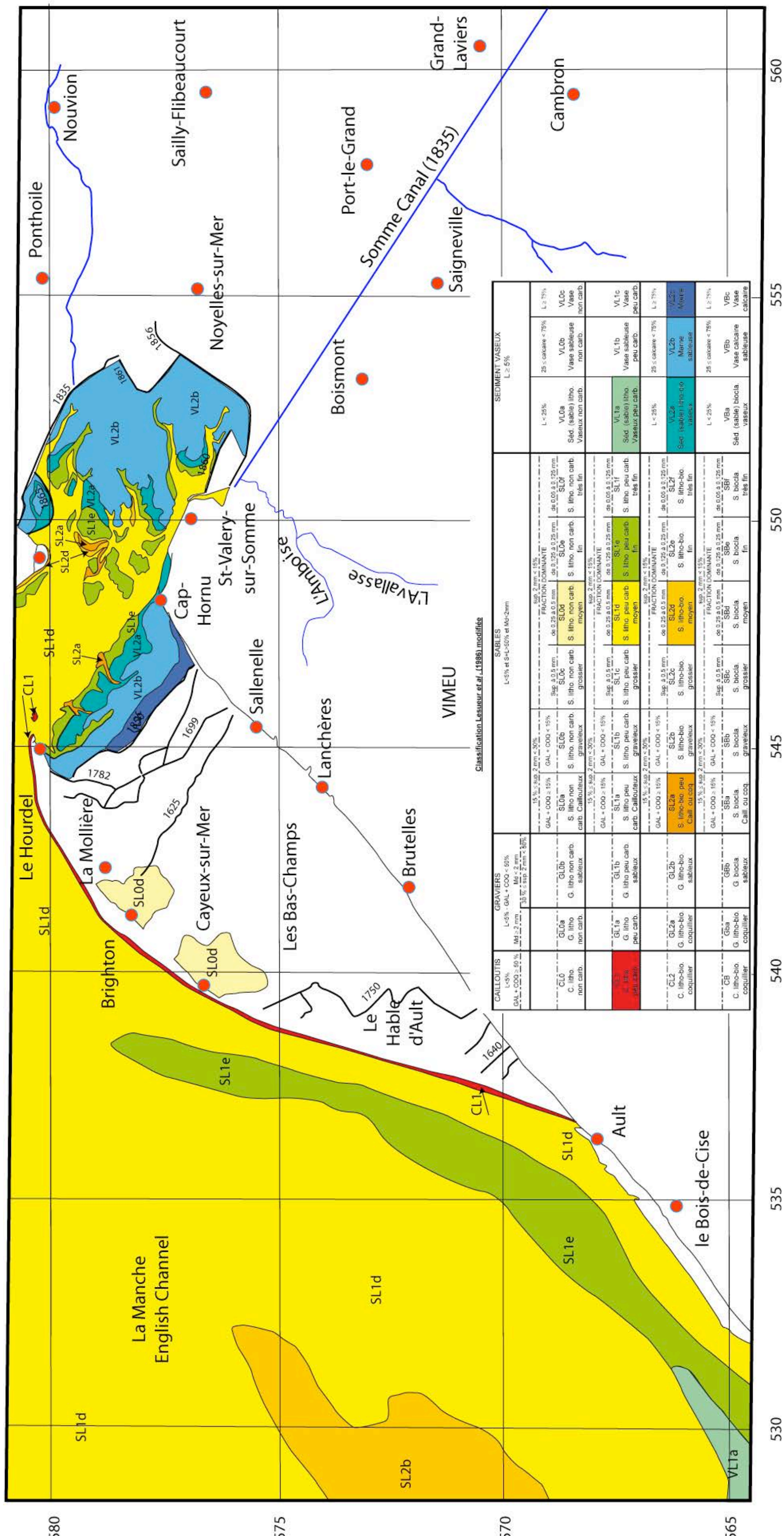
Recent sedimentation has been intensively studied in the 80's (e.g. Dupont and Homeril, 1980), but then the attention decreased replaced by management-oriented studies due to the high vulnerability of the coast (see sections 5).
The Somme estuary is macrotidal, but wave-dominated due to high energy wave conditions, the resulting strong littoral drift leading to the development of the gravel spit

(see part 2). The Somme Bay (Fig. 30) is almost exclusively filled with sands of marine origin (Dupont, 1981) and bioclasts from endemic benthic production (Desprez et al., 1998). Sedimentation rate in the Somme estuary is about 700 000 $m^3.yr^{-1}$ corresponding to a mean seabed elevation between 1.3 and 1.8 $cm.yr^{-1}$ (Verger, 2005; Bastide, 2011). These rates are similar to those recorded in the Authie estuary tens km further North (0.71-1.6 $cm.yr^{-1}$; Marion, 2007).

Dupont summarized the influence of tides and waves on the sedimentation in the Bay (1981). Tidal currents are responsible for the formation of an ebb delta protecting the inner estuary (Figs 31 and 32). Fine sands penetrate the estuary as sandy layers associated to flood-channelized currents. Flood dominance favours the infill pattern and the construction of sandy bodies at high topographic levels. Waves contribute to the protection of the internal estuarine domains by edifying littoral spits and swash bars at high altitudes. In internal estuarine domains, decantation of suspended sediment operates on the rapidly prograding salt marshes (schorres; Fig. 33), but also on the mixed flats (slikke) in sheltered areas (e.g. Mollières d'Aval to the South). In the estuary, a grain size fining trend is observed from open-sea sites to sheltered sites, due to various exposure degrees to tide and wave action. This is shown in the sedimentation in *Spartina* and *Halimione* communities observed on the low marsh (between slikke and schorre) flooded by mean tides and on the mid-marsh flooded by spring tides for 3 different sites in the Bay (Le Bot et al., 2012).

At Le Hourdel, intertidal modern sedimentation can be seen on both inner and outer estuary sides (Fig. 33). On the seaward side, small to medium dunes (Fig. 34) cover the intertidal areas, while south-eastward, fine sedimentation occurs. The total energy rapidly decreases and is at the origin of most of the fine sedimentation.

Figure 30. Surface sediments based on their grain-size and $CaCO_3$ content. Data from BRGM, 1985. Classification according to Larsonneur et al., 1978. From left to right, the mean gain-size decreases. From top to bottom, the $CaCO_3$ content increases. CL1 (red): lithoclastic gravels. SL2a (dark orange): litho- and bio-clastic sand bearing between 15 and 30% of gravels. SL2d (orange): litho- and bio-clastic medium sand. SL1d (dark yellow): lithoclastic medium sand poor in carbonate. SL0d (pale yellow): lithoclastic medium sand free of carbonate. SL1e (yellowish green): lithoclastic fine sand poor in carbonate. VL1a (green): muddy lithoclastic sand poor in carbonate. VL2a (blue green): muddy litho- and bio-clastic sand. VL2b (blue): sandy marl. VL2b: marl.


La Manche
English Channel
Le Hourdel
La Mollière
Cayeux-sur-Mer
Brighton
Les Bas-Champs
Le Hable d'Ault
Ault
le Bois-de-Cise
Brutelles
Lanchères
Sallenelle
Cap-Hornu
St-Valery-sur-Somme
VIMEU
L'Amboise
L'Avallasse
Ponthoile
Nouvion
Sailly-Flibeaucourt
Noyelles-sur-Mer
Port-le-Grand
Somme Canal (1835)
Saigneville
Boismont
Cambron
Grand-Laviers
SL1d
SL1e
SL2b
SL0d
SL2a
VL2a
VL2b
VL1a
CL1
Classification Lesueur et al. (1986) modifiée
CAILLOUTIS
GRAVIERS
SABLES
SEDIMENT VASEUX
2580
2575
2570
2565
530
535
540
545
550
555
560

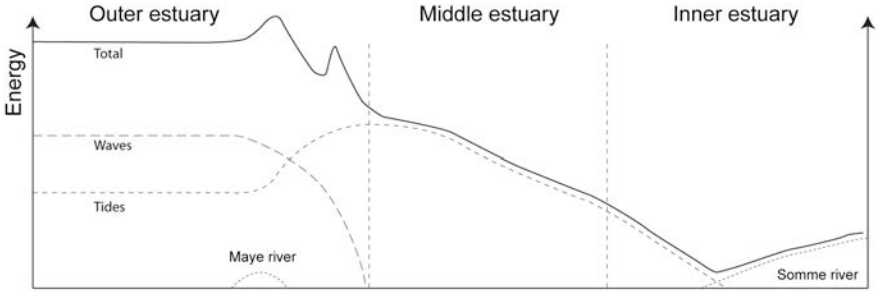


Figure 31. Energy diagram in the Bay of Somme. Partly from Dolique, 1998. This figure can be compared with fig. 32.

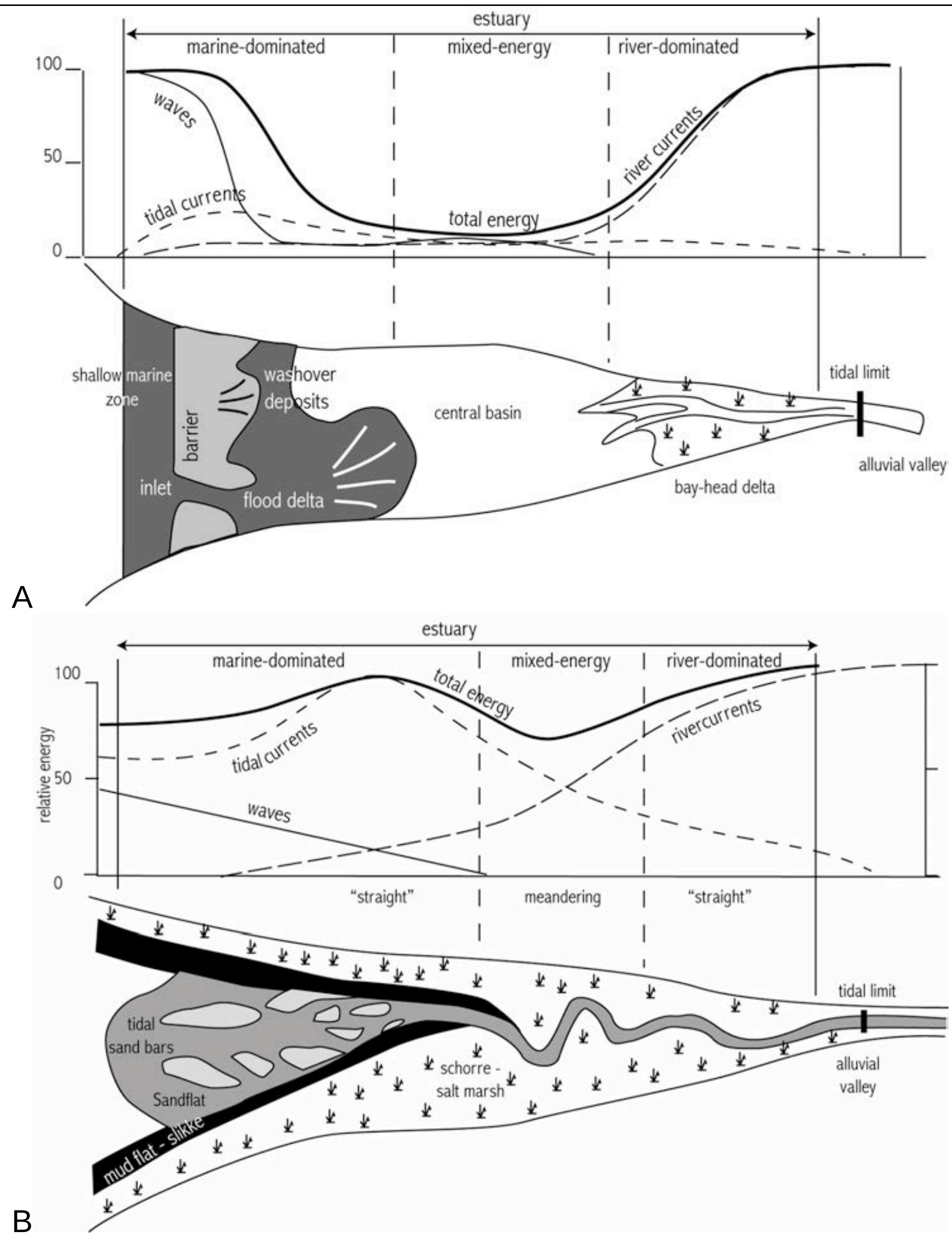


Figure 32. Distribution of energy types and morphological components of an idealised A) wave-dominated, and B) tide-dominated estuary (From Dalrymple et al., 1992).

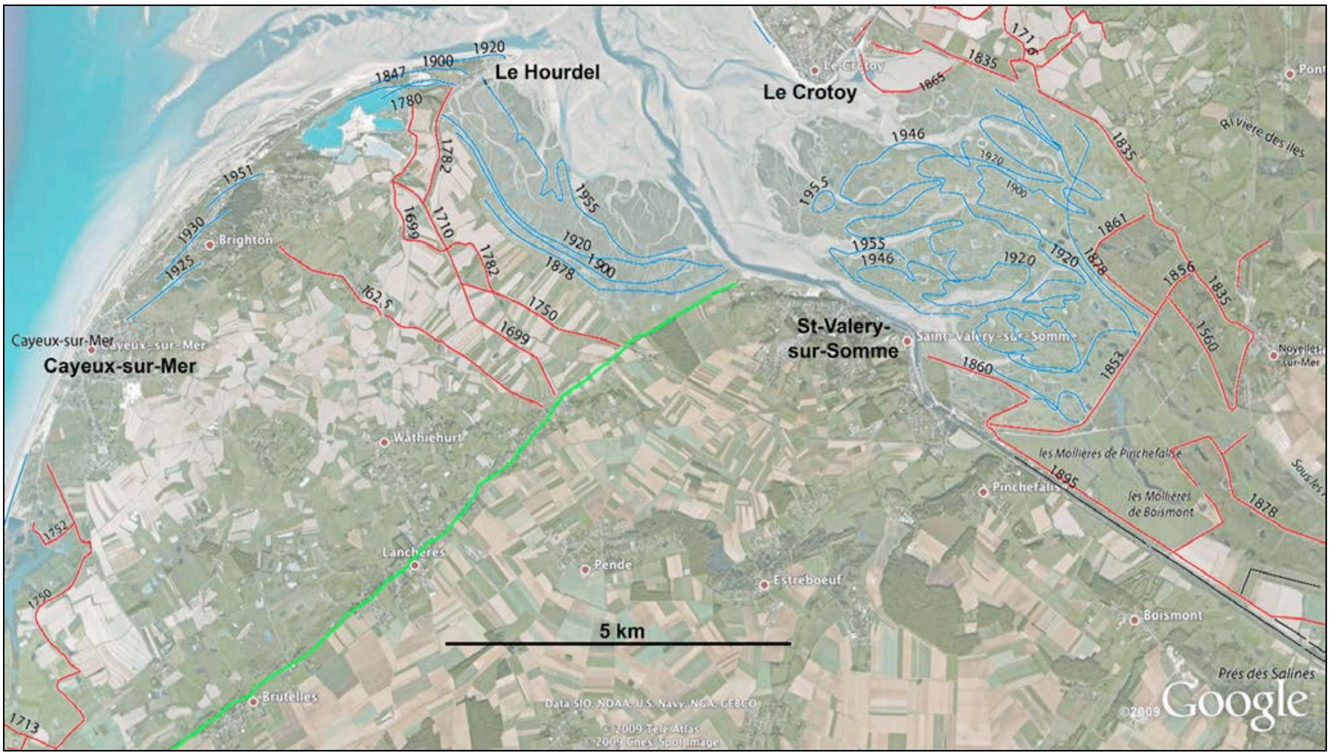


Figure 33. Central part of the Bay of Somme. In red, dated dykes. In blue, vegetated area evolution. In green, Paleocliff. In black, Abbeville-St-Valery-sur-Somme canal. Base map: Google Earth. Image date: Dec. 2007.

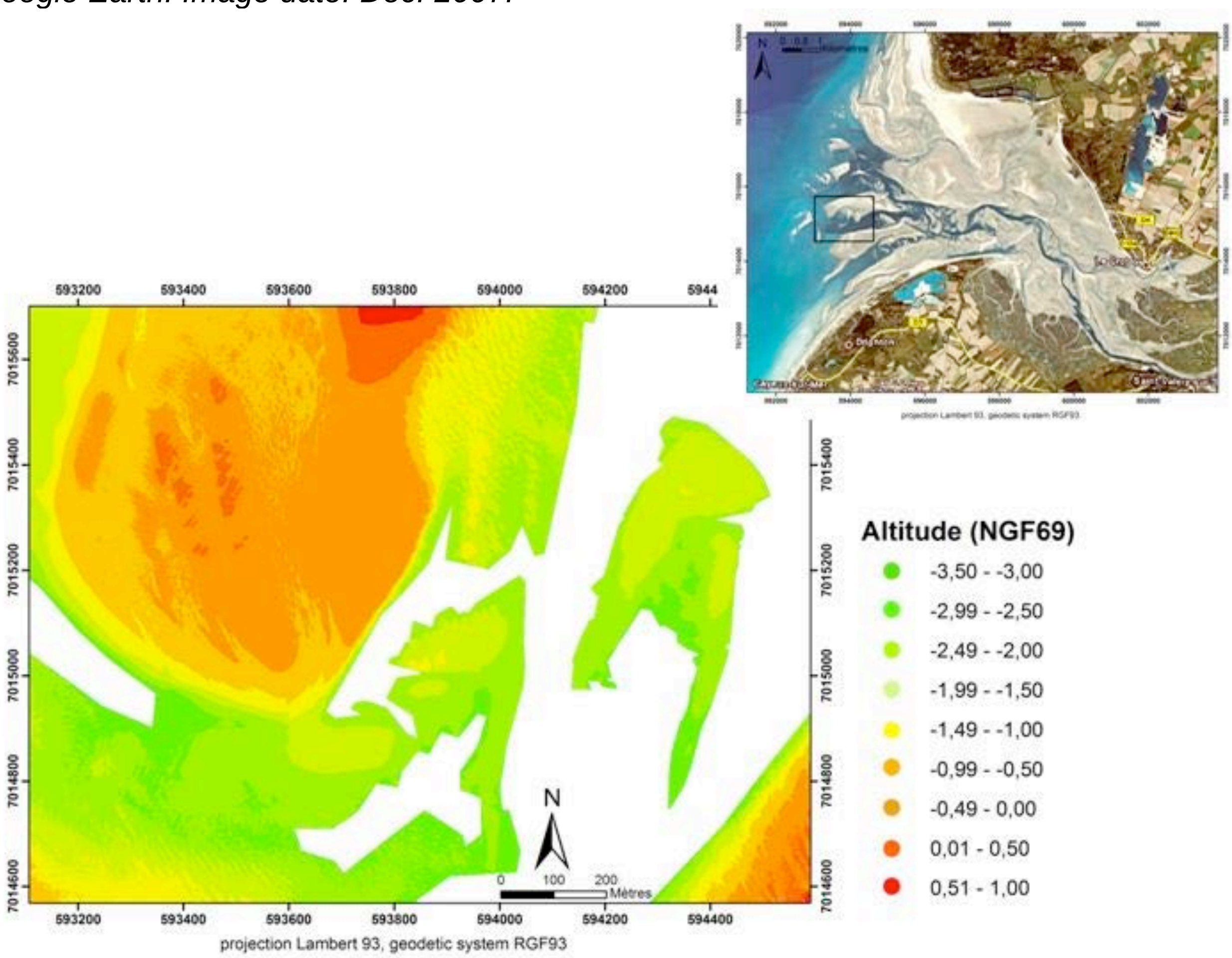


Figure 34. Topography of the channel area, in the external part of the Bay. DTM from LIDAR data (data source: operational team CLAREC, CNRS-UCBN, 28 September 2011). Up, right picture from Google Maps.

3-2. Long-term evolution

A series of coring were done in the Bas-champs allowing reconstructing the shore movements from 7 500 BP (Fig. 35).

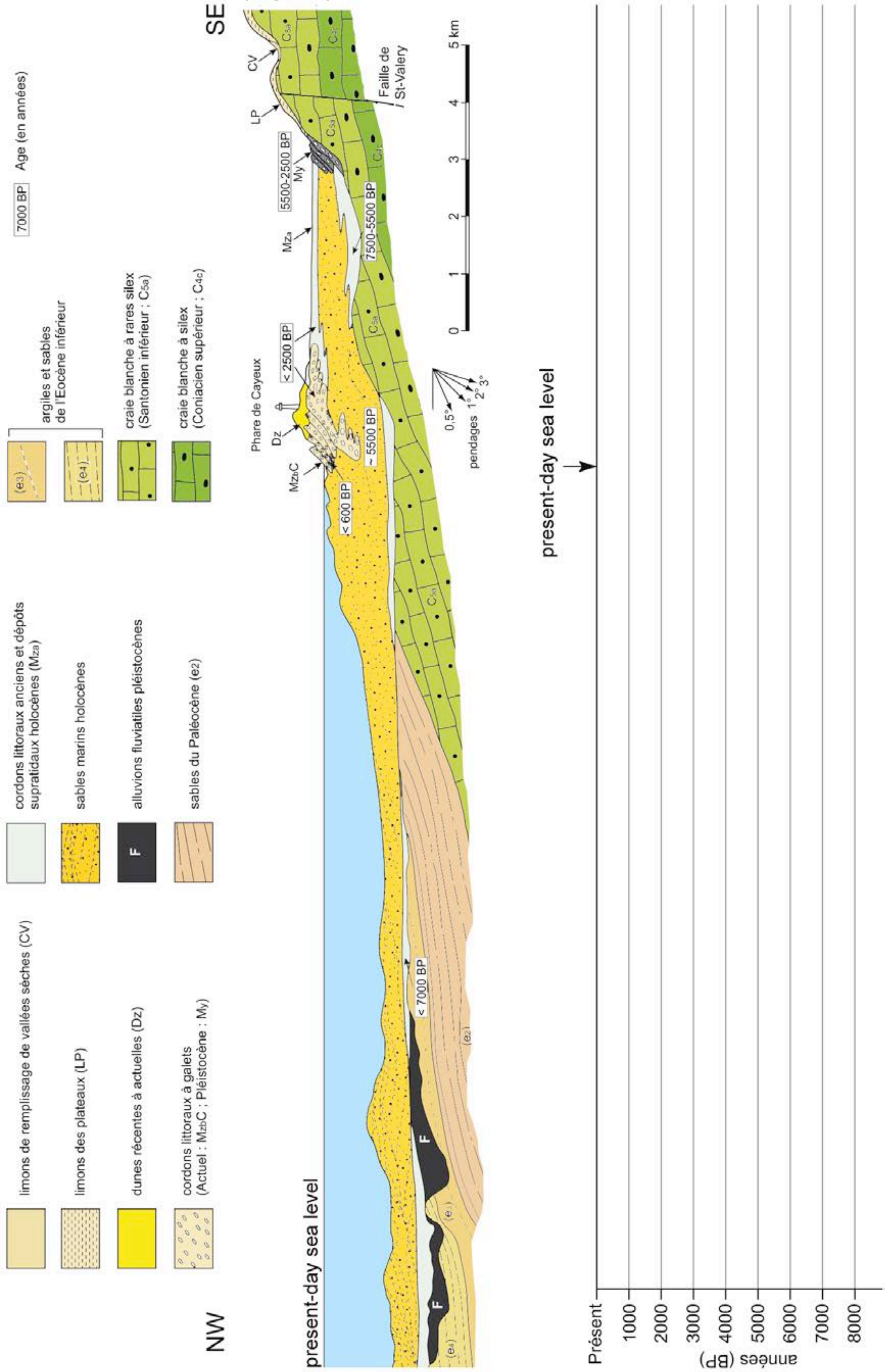


Figure 35. Cross section of the Bas-champs by the Cayeux lighthouse (BRGM, 1985).

3.3- Man-influenced coastal evolution

From the Middle Ages, for agriculture needs, man influenced the evolution of the coast in building embankments on the upper intertidal zone. This was done all along the Picardy coastal plain (Fig. 36), deeply modifying the landscape, closing the smallest estuaries. Works have been attested from the XIIth century in the bay of Authie, while in the bay of Somme the earliest embankments are from the mid XVIIth century. In the left bank of bay of Somme, thanks to continuing silting up, embankments were done in the southernmost part of the bay, and on its northern part on both side of a line linking Cayeux-sur-Mer to the paleocliff (Fig. 37). To the South, the objective was to reclaim the Hable d'Ault, a former estuary that regularly opened through ages. To the North, the embankments followed the gravel-spit evolution in its way North. On the right bank of the Somme embankments also occurred between St Valery-sur-Mer and Le Crotoy, but also around the Maye river (Fig. 36).

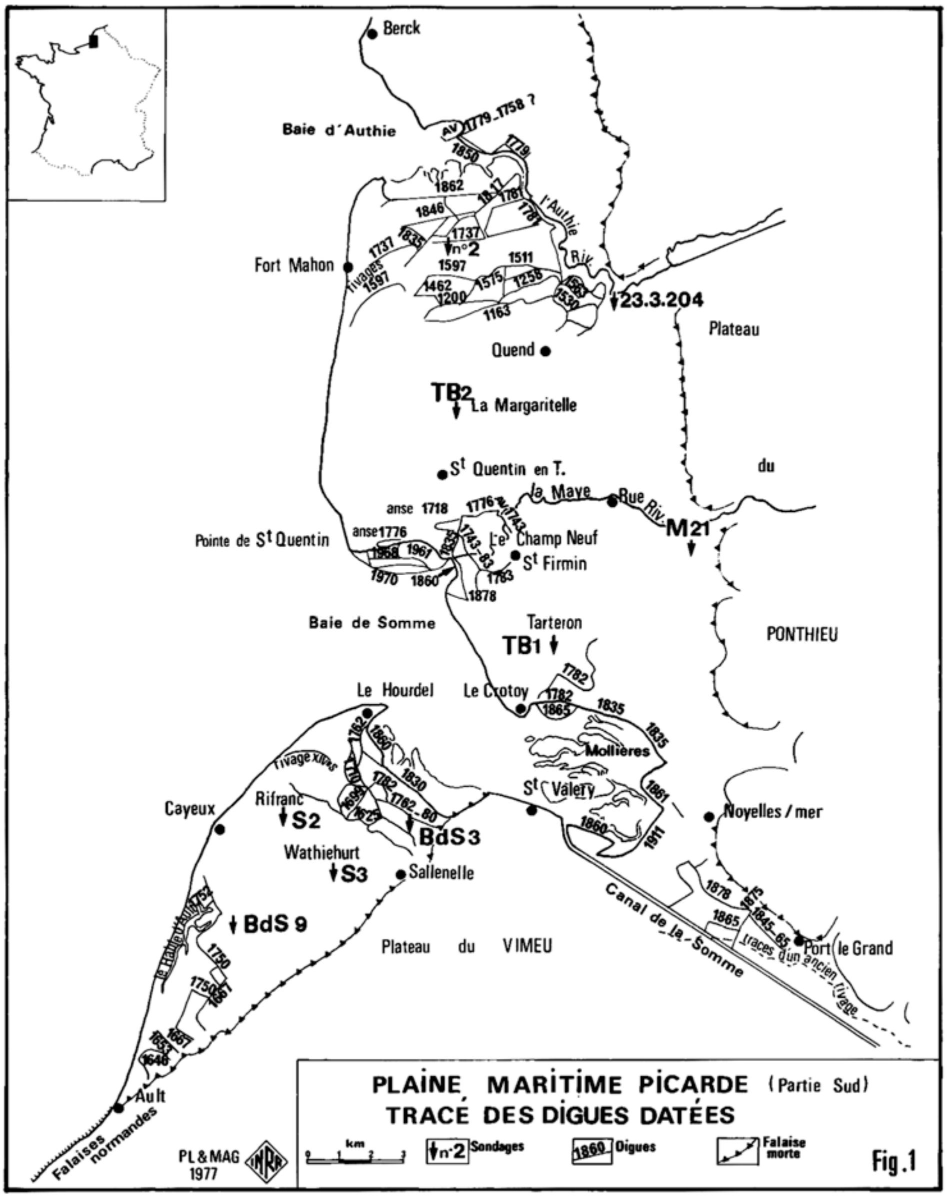


Figure 36. Embankment evolution in the Southern Picardy coastal plain (Lefevre, 1977).

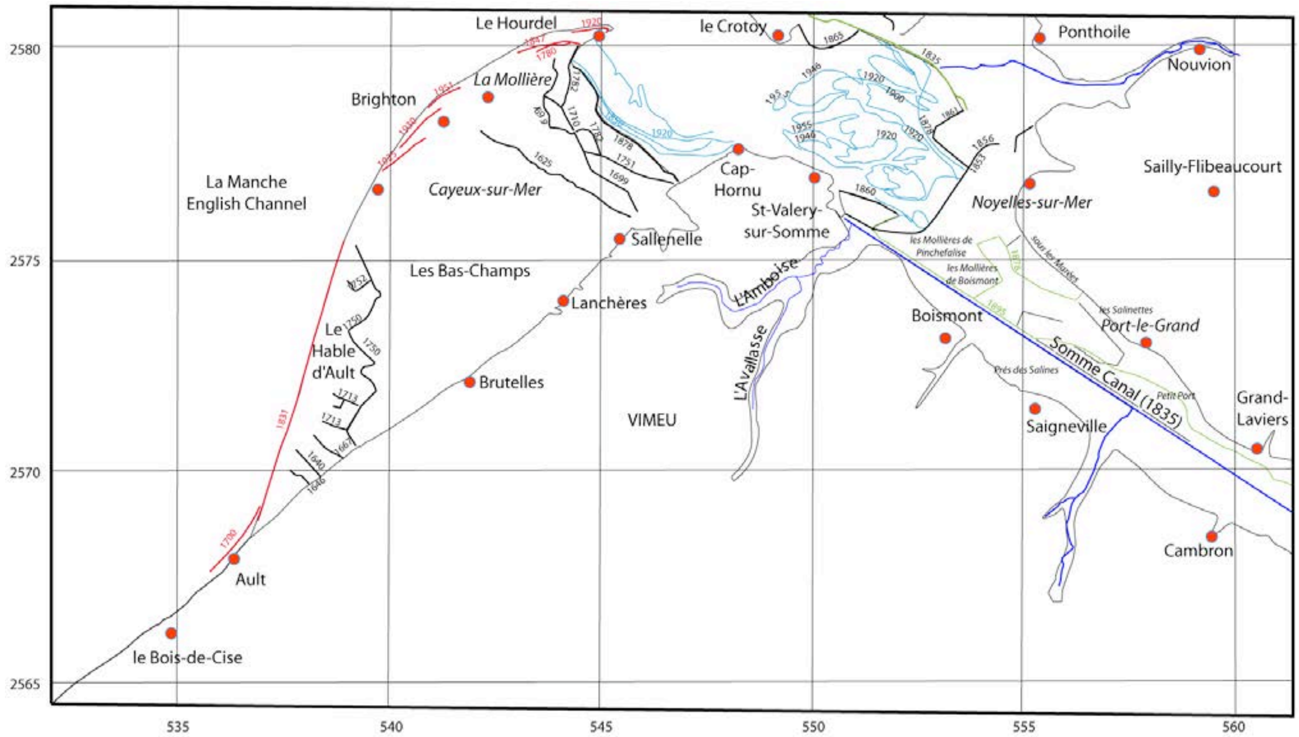


Figure 37. Significant lines indicative of shore evolution. In red, gravel spit location. In blue, shorre/slikke limit with dates, in black, main embankments. In italics, some villages or locations have been distinguished to illustrate a former maritime influence, e.g. Port-le-Grand (large harbour), les-Salinettes or Prés-des-Salines that refer to salt occurrences.

4– Ante-Holocene gravel spit evidences

Fossil gravel spits outcrop on the Northern side of the Somme. They correspond to a series of small hills clearly visible in the topography. These relative elevations have often been chosen to establish some cities from Gallo-Roman time (Gosselet, 1906). The city of Rue was an important trade city in the Middle Ages. The sea was still along its walls in the XIIth century. From the XVth Century largest vessel were not able to reach the harbour that disappeared during the XVIIth Century.

The gravel relief were especially visible prior to the gravel extraction (Fig. 38) as on modern maps these 10-metres high hills have been transformed either on quarries or in recreation lakes after exploitation. Cities of Quend or Rue are still on these remaining hills.

Figure 38. Topographical map drawn from the IGN 1:80 000 map. Recognition of dated element such as sea walls and groins indicate that the map could have been surveyed between the XIXth and XXth century. Geological information has been overlain from the geological map (BRGM, 1981). In blue, My: Rue Formation (Gravels) in beige: hillslope and underlying terrains. It combines CLP: hillslope loams, LPS: sandy-clayish red loams bearing flint pebbles, and C4C: White Chalk with flint of Upper Coniacian age not detailed on this figure.

The gravel spit rests on a white chalk (craie) marine abrasion platform (Fig. 39). Some reliefs seem to characterize the surface of the platform. Diverse interpretations have been proposed to explain the changes in altitude and in orientation of these gravel spits. The presence of faults or flexure is mostly accepted, but some erosive processes could be evocated.

The maximum altitude is at +13 m NGF with a water level at about +4 m NGF. The base of the gravel spit close to -8m NGF.

The sediment is composed of quartz sand and flint gravels. Cretaceous fossils are seldom observed as gravels. Exotic rocks such as diorites, pink granites, metamorphic rocks or sedimentary rocks can be observed. Up to 4 m wide blocks of Tertiary sandstones can be found in the lowermost intervals. It is thought that these last elements were transported on ice rafts during ice breaking up. Other exotic rocks are similar to some observed along the coast of Brittany (further South) or Boulonnais (to the North) (Petit, 1959). The gravel deposits are free of shells (Briquet, 1930), probably due to decalcification processes.

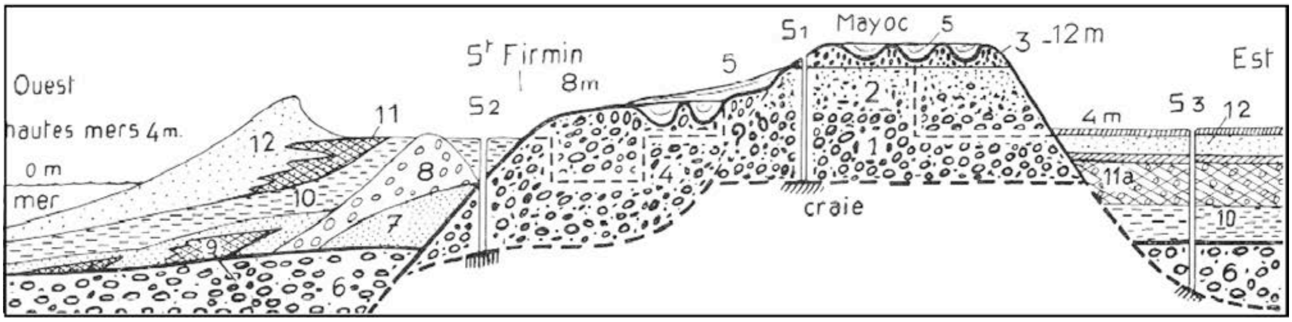


Figure 39. Cross section in the Northern end of the Le-Crotoy gravel bar. This interpretation proposes a multiphase setting for the bank and separates the group in two (St Firmin and Mayoc) individual bars (Agache et al., 1963).

Pebbles are usually organised in horizontal or slightly (2 to 5°) SE-ward dipping beds (Fig. 40-A). Stratification is made of an alternation or more or less sandy intervals. The uppermost part of the spits consists in finer sediment and often present some evidences of cryoturbation such a loem-filled ice wedges (Fig. 40-B).

A B

Figure 40. Typical views of the fossil gravel bar at Oscar Savreux quarry. A. Layering in the Pleistocene gravel spit. B. Loem-filled ice wedge at the top of the gravel formation. Total thickness of this picture is about 2 m.

Further west, some miles offshore Dieppe, ancient gravel spit (16 km long, 4 km wide) is also observed (Claveleau, 2007). It is made of 3 sediment bodies; each one composed of several prograding units and is interpreted as being of Pleistocene age. Its geometry is similar to what is observed in the quarries.

5– Environmental considerations

The low-lying area back from the gravel spit is fragile. Despite many attempts to consolidate it by gravel imports or groyne constructions from late XIXth century, some places are often subject to breaches that lead to land inundations (Fig. 41). On this picture, four successive submersions are distinguished.

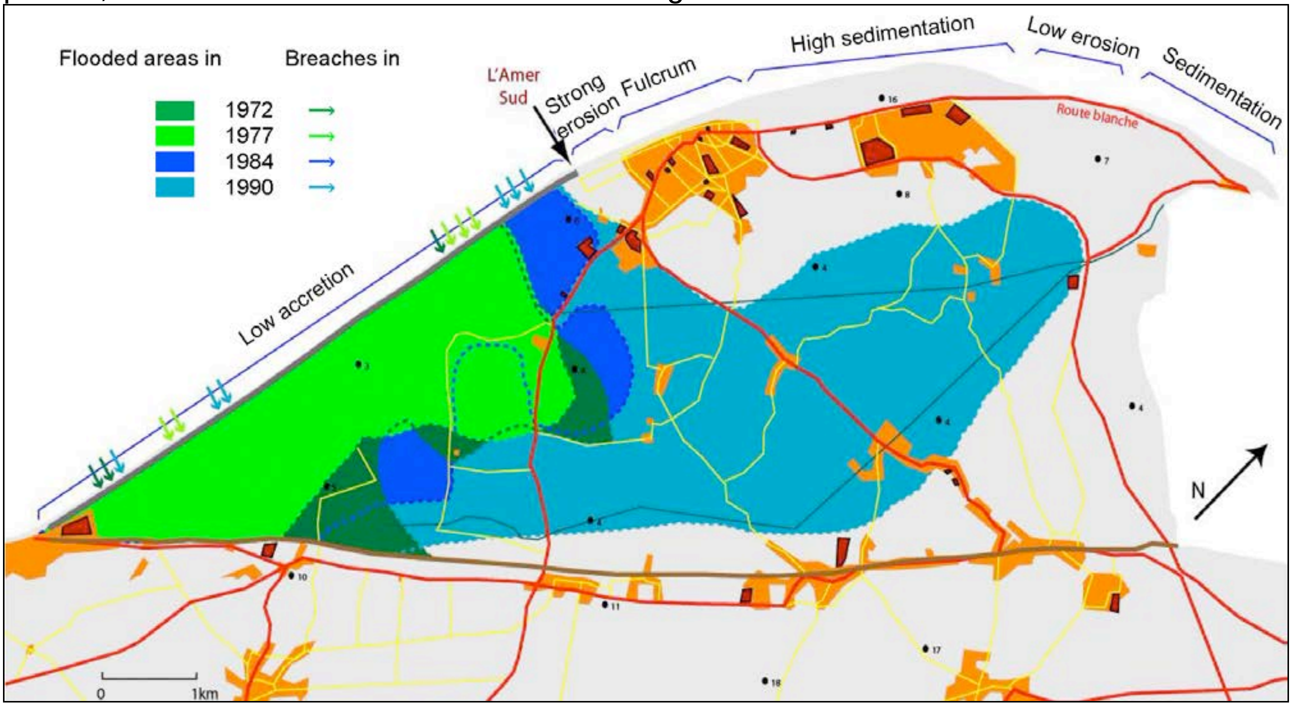
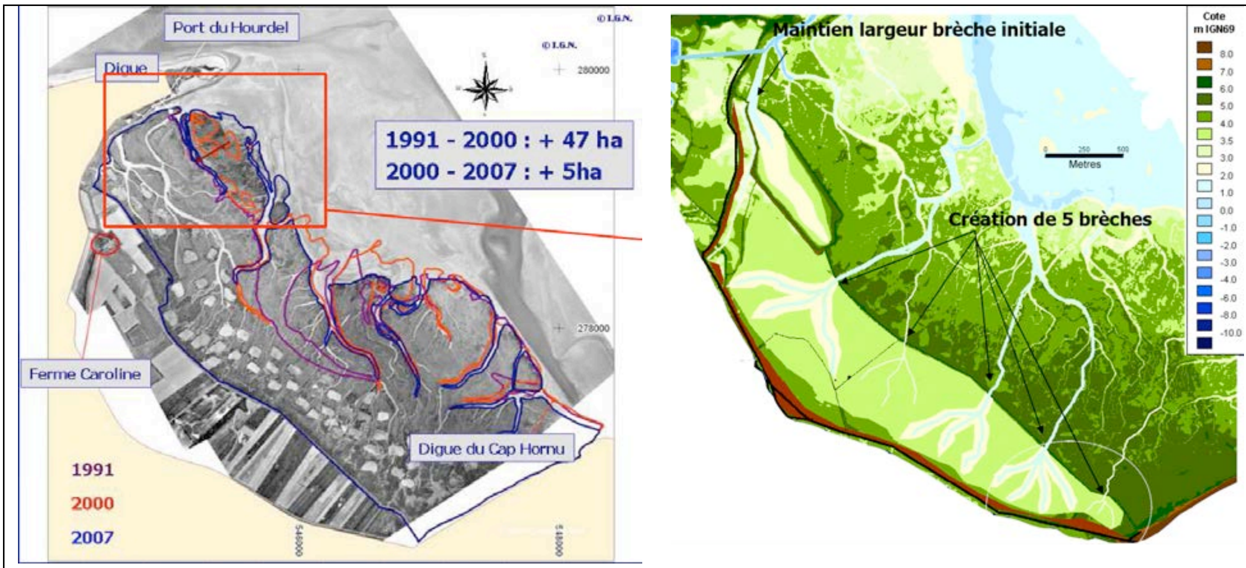


Figure 41. The Bas-champs of Picardy. Areas that were submerged by floods in 1972, 1977, 1984, and 1990 (Carel, 2009). The data used by Carel are from Costa, 1997, Dolique, 2004, and Creocean, 2005.

Many issues deal with the gravel budget along the spit. To help protecting the shore a lorry noria feed the gravel spit at its weakest positions, between groynes and at the Amer-Sud. Gravels are taken either from the final spit or from local quarries.

The other problem that the Somme bay has to face with is the continuous silting up. It leads to strong reduction of the intertidal zone (Fig. 42-A) and reduces the possibilities for boats to join their harbour. Biggest one moved to further deepest harbour out of the Bay, while being replaced by smaller units and recreation vessels. One possibility to reduce sedimentation and leave an open access to harbour is to re-open ancient reclaimed areas locally called polders as in the Netherlands. This is the case close to the final spit in the Ferme-de-la-Caroline polder (Fig. 42-A). In opening this area, and removing a huge amount of sediment, one can expect, on a very long-term basis, to recreate a more open area and maintain the access to Le Hourdel harbour (Fig. 42-B). In this configuration, dykes would be consolidated to protect the population.
Such a scenario could be difficult to accept for a population that continuously increased its acquisition of intertidal areas but is encouraged by most local authorities. A second project of reclaimed-area reopening is planned to the South at Le Hable d'Ault, as continuous gravel-spit feeding is too expensive compared to the local possibilities. This gravel spit was definitely closed in 1752 by a gravel dyke, the Digue-du-grand-barrement, but is often submerged and need to be repaired. A politic of controlled of the coastline could be more economic than a politic of strict maintenance, but is sometimes facing local lobbies.



A B

Figure 42. A. Evolution of the intertidal zone close to the harbour of Le Hourdel between 1991 and 2007. Lines correspond to the altitude of 4.1 m. they mark the limit between the vegetated schorre (salt marshes) and the non-vegetated slikke (mixed-flats). B. Digital terrain model on a very long-term basis after opening the Caroline farm polder to the North, and the possibility to open 5 new breaches.

During this excursion, we'll visit an area on the other side of the bay, close to Le Crotoy. A 50 ha basin has been opened to the sea to flush the sediment that is silting up in the harbour. Despite a certain efficiency to maintain water in the harbour very high sedimentation rate occurs in the basin that need to be often dredged. On average, 30 m^3 are deposited at each tide and more than one million m^3 were dredged between 1973 and 1982. In 1993, 800 000 extra m^3 were dredged again (Dupont et al., 1993).

6– Marine geology

Offshore the Somme Bay, the coarse seabed of the Eastern English Channel is covered with a thick cover made of a sand and gravels mixture. Its construction results from the Holocene transgression (Auffret *et al.*, 1980; Dewez, 1988). This wedge is moulded by a series of tidal sand-banks, generally covered with dunes. Tidal sandbanks are parallel to the coast, although some tend to connect the shallow area when the orientation of the coastline suddenly changes in the surroundings of Ault and the gravel spit (Figs. 43 and 45). Very large dunes display heights between 4 and 10.5 m and wavelengths between 250 and 1800 m (Ferret et al., 2010; Ferret, 2011; Fig. 44). Dune migration rate is not very high, varying between 0.8 ± 0.25 $m.yr^{-1}$and 6.6 ± 0.7 $m.yr^{-1}$. Sediment transport and dune residual movements are toward the East, in the direction of the dominant flood, but small waves may reverse sediment transport direction to the West and slow down (and even reverse) dune migration. Sediments of the sedimentary wedge of Picardy partly supply the Somme Bay, even though fluxes and budgets are not yet known.
The area is also characterized by a well-developed paleovalley system (Fig. 44) that drained main rivers and the now coastal rivers. A complete network was first drawn by Auffret et al. (1980) and was refined in its shape (Auffret & Alduc, 1982), in its stratigraphy (e.g. Lericolais et al., 2003) or in the history of its origin (Gupta et al., 2007).
Recently shot seismic profiles offer the possibility to describe the architecture of coastal tidal sand banks, but also to reconstruct the paleochannel network in an area that is close to the coast, where seismic profiling is less efficient due to rapid income of the first multiple (Fig. 46, 47, and 48; Trentesaux et al., 2011). Careful analyses of the 200m-spaced

profiles allow defining a series of surfaces mostly dipping offshore and sometimes incised by ancient small or larger streams. The Somme paleovalley is well expressed in the South of the study area, while, in the North a strange meandering channel occurs (Fig. 49). This valley is not in front of the Authie River at its present-day location, but could correspond to an ancient stream combining waters from the Authie and further-south smaller coastal rivers.

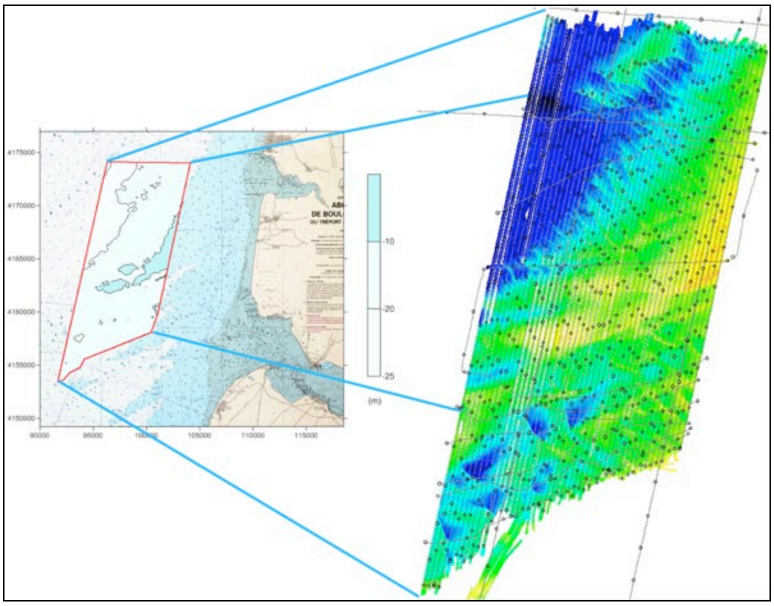


Figure 43. Bathymetry offshore the Somme Bay. The end of a sandbank, the Quemer, is clearly visible in yellow. Its asymmetry, the steepest flank facing the coast, is also visible. Dunes of different heights cover the bank and surrounding areas. Courtesy from Laure Simplet, 2010, Ifremer.

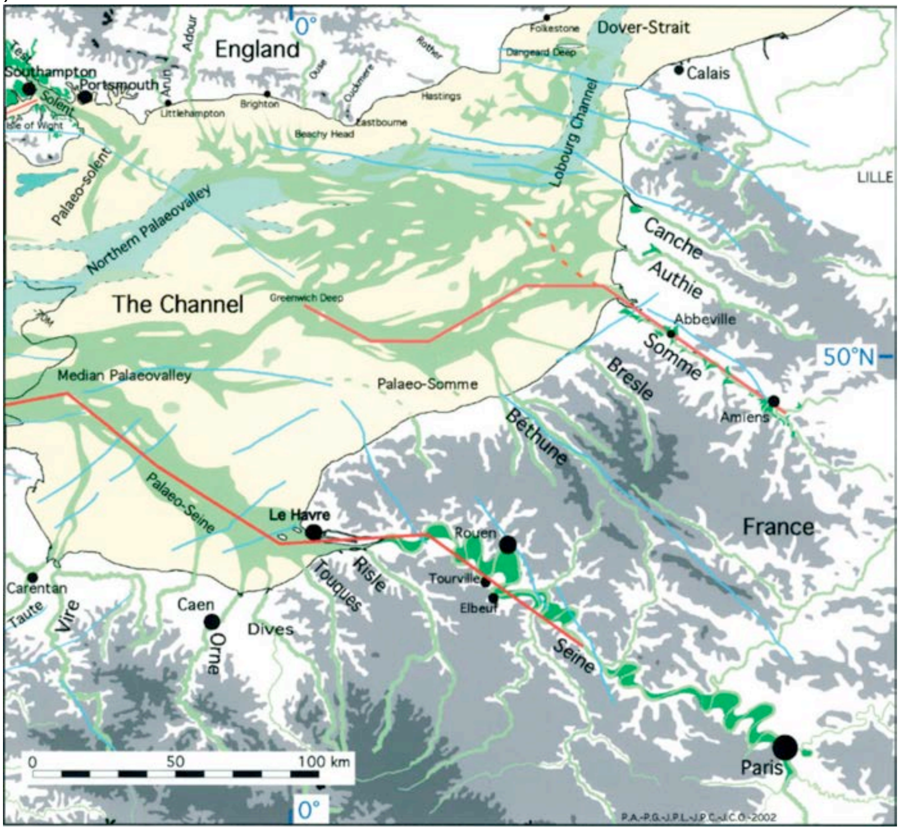


Figure 44. Map displaying the valley and offshore paleovalley system in the Eastern English Channel (Antoine et al., 2007).

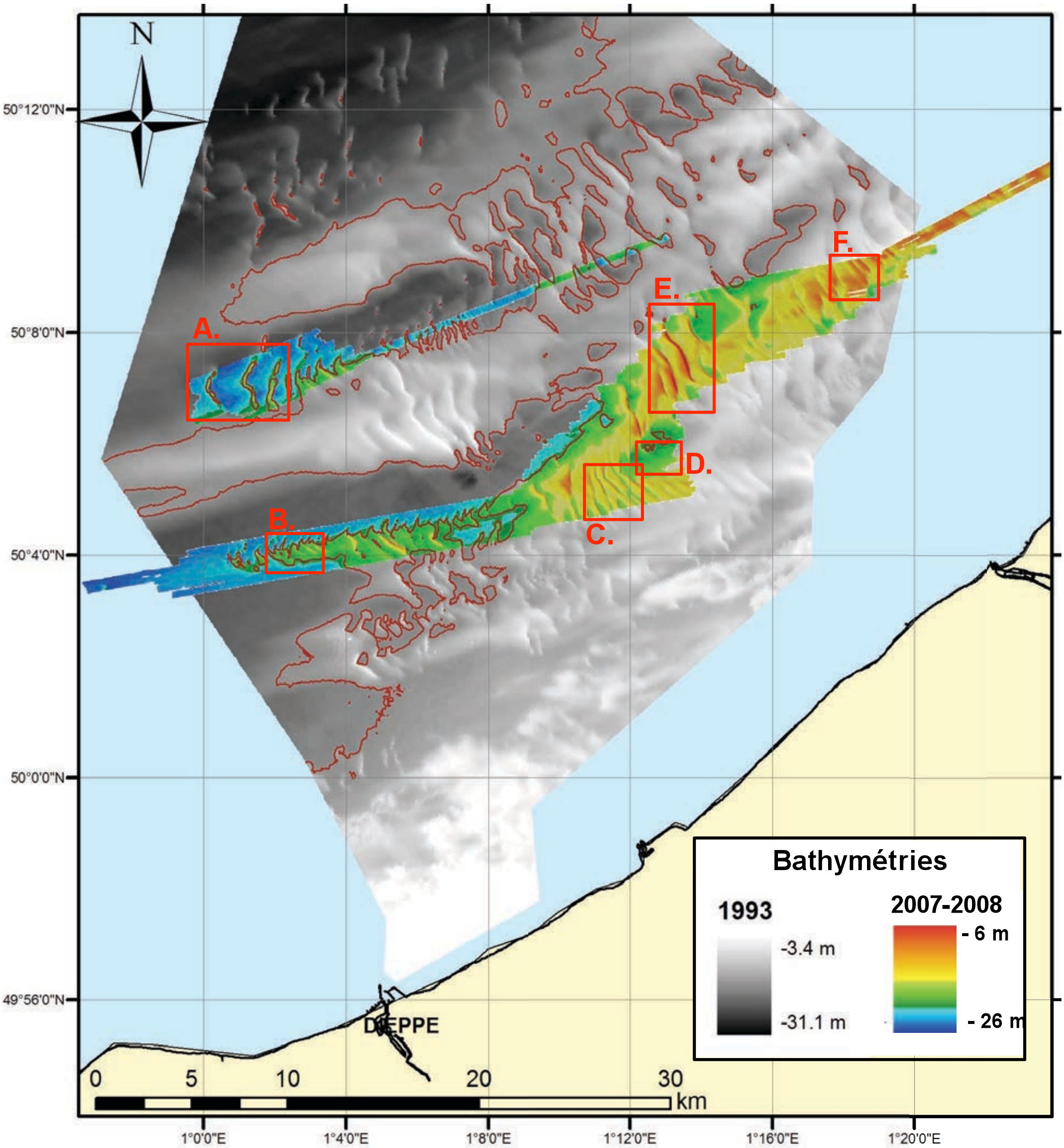
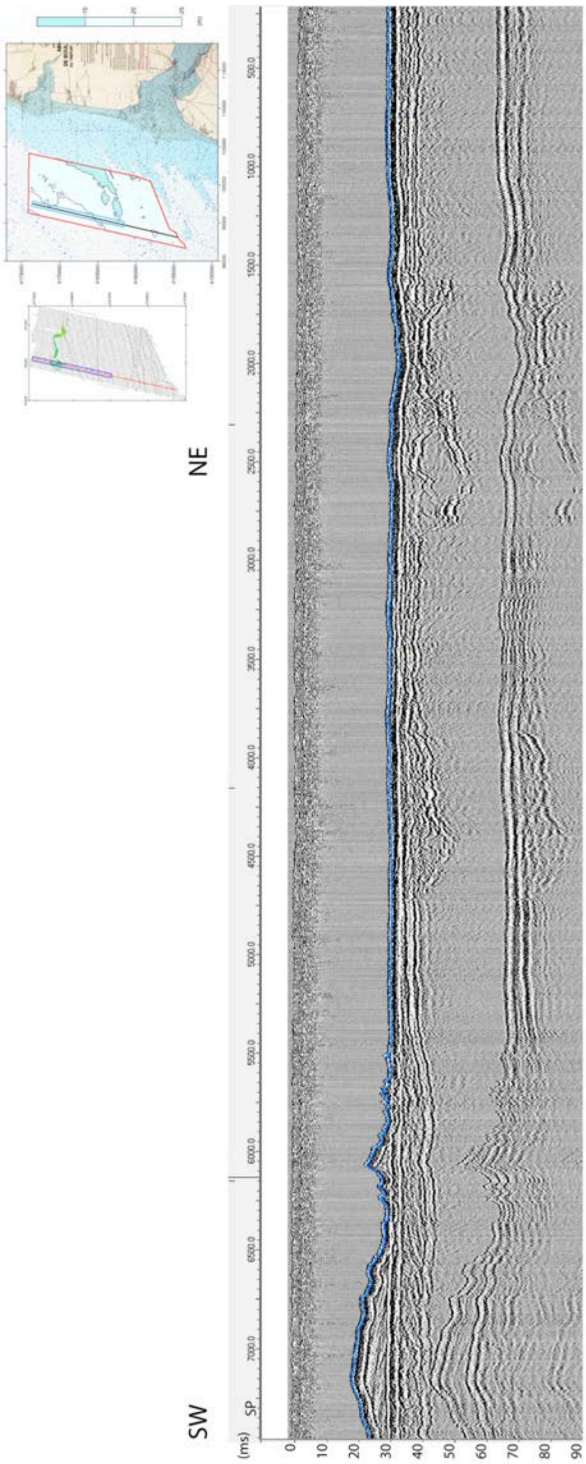
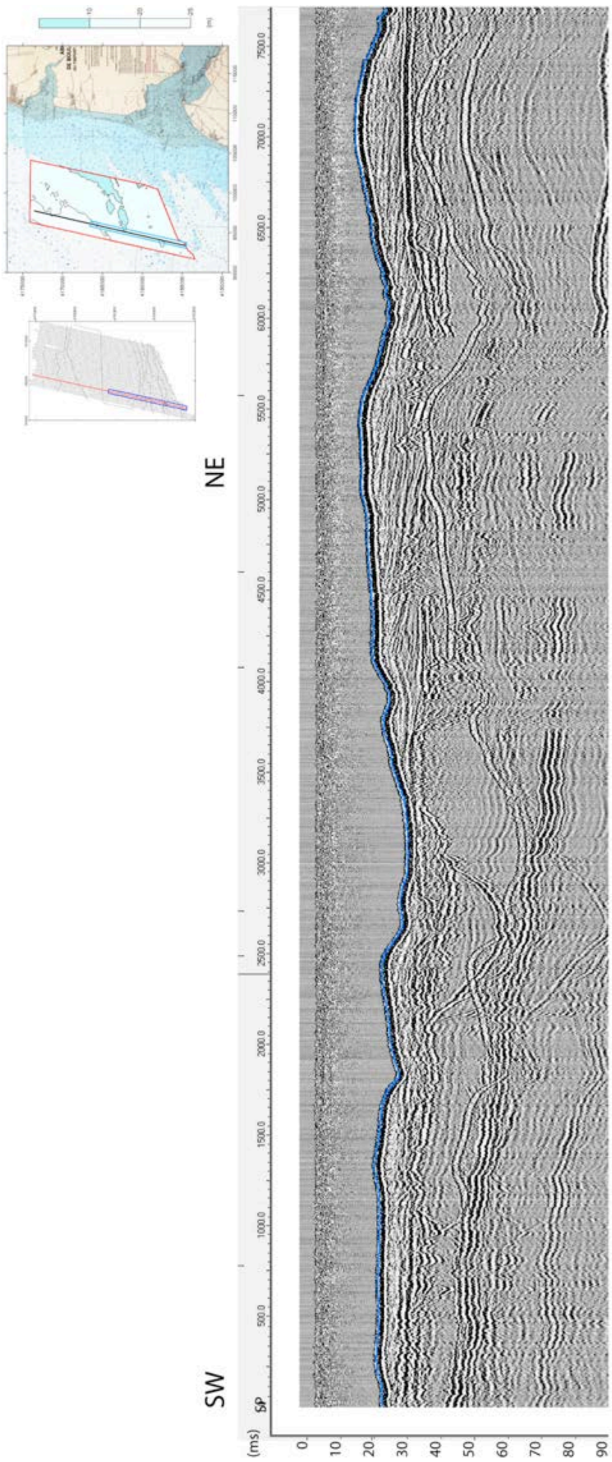


Figure 45. Bathymetry offshore Dieppe and Le Tréport (Northern harbour; Ferret, 2011). Single and multibeam data acquired respectively in 1993 (data source: SHOM) and 2007-2008. The red line corresponds to -20 m and underlines the tidal sandbank boundaries.



A



B

Figures 46 and 47. High-resolution sparker seismic profiles shot offshore the Somme estuary. Depths have been corrected from tide-related sea level. A. Profile 124 displays a series of three small incisions that are not in front or the Somme nor the Authie at their present-day locations. B. Profile 123 displays a deep incision related to the Somme valley. The South-dipping reflector corresponds to the Top Cretaceous. Gas, probably of biogenic origin, seems to be trapped in the sandbanks on the east side of the profile.

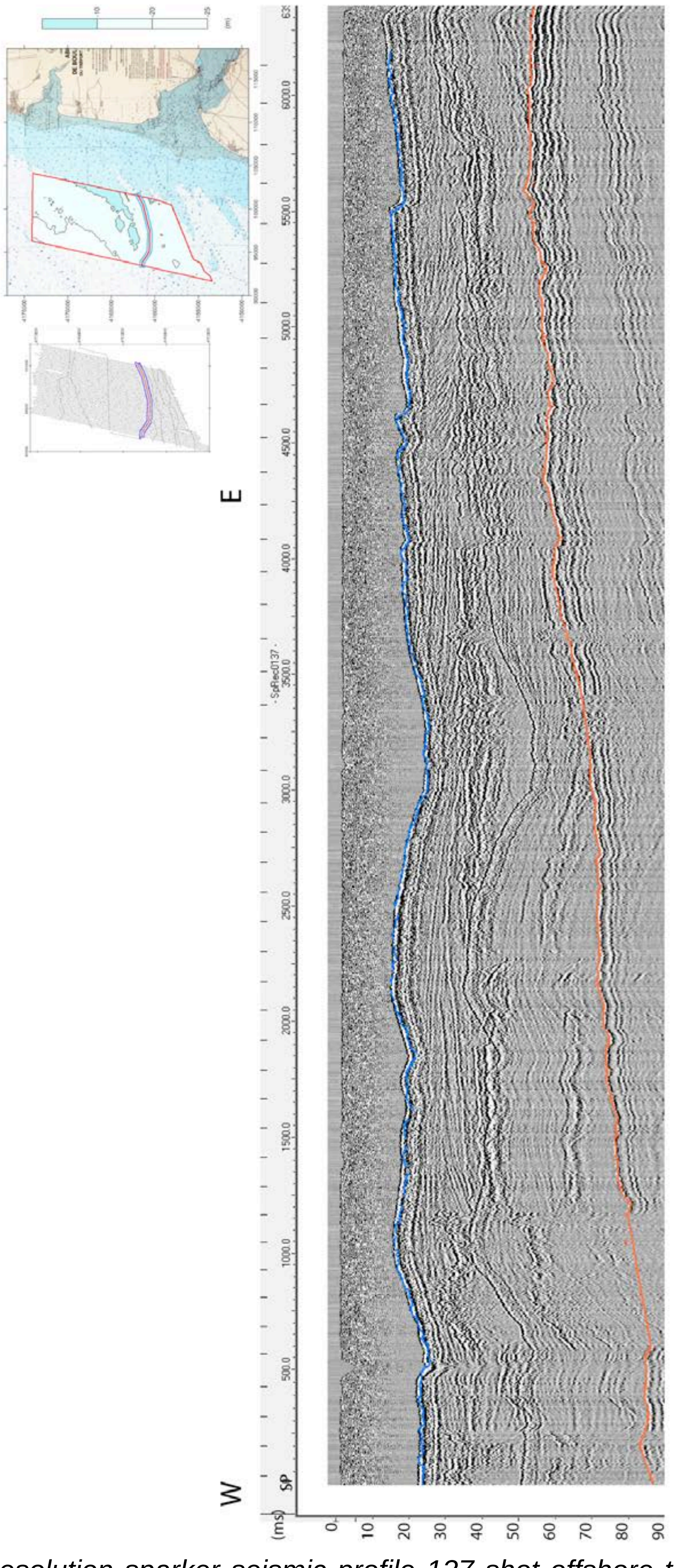


Figure 48. High-resolution sparker seismic profile 137 shot offshore the Somme estuary. The top Cretaceous surface in orange is dipping toward the centre of the Dieppe Basin. The uppermost part of the profile shows inclined bedding related to the upper Quaternary structure of the offshore banks.

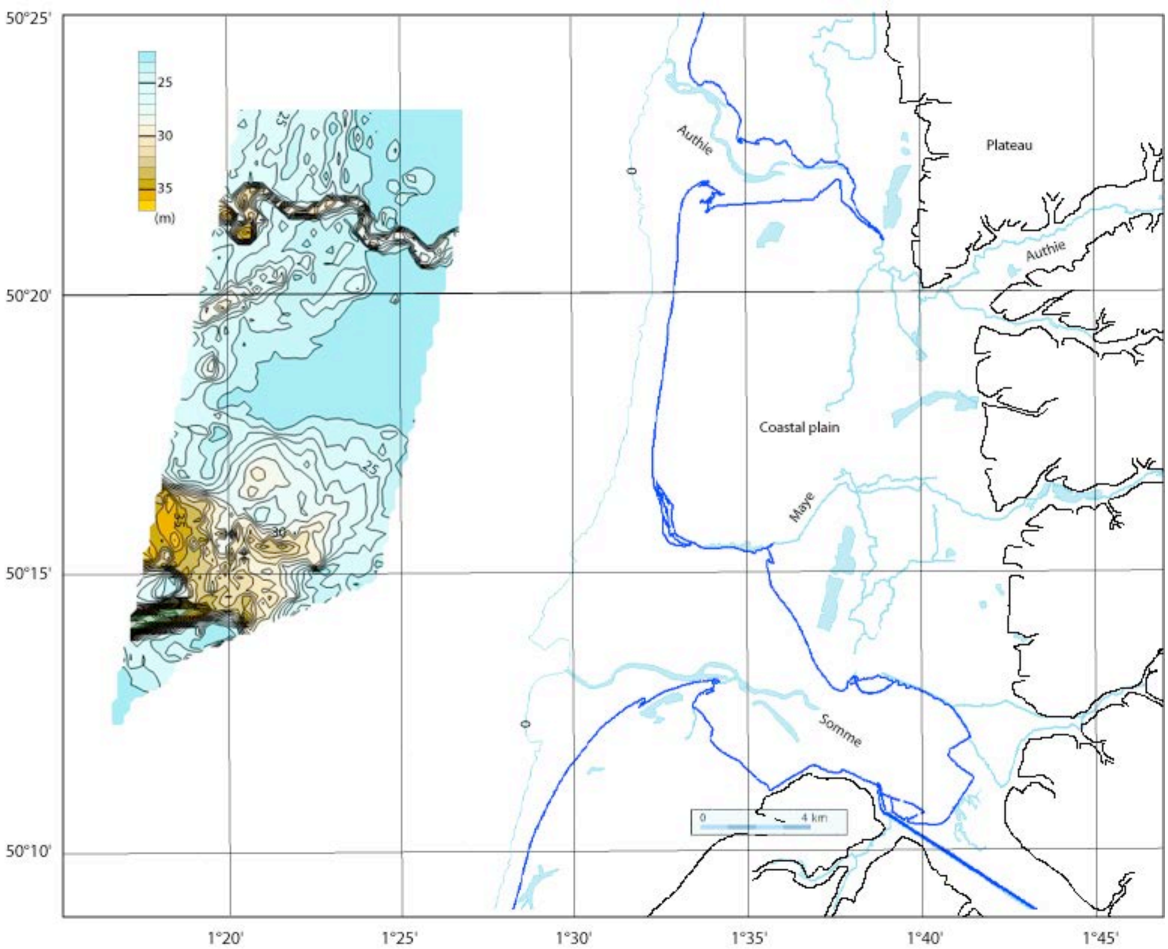


Figure 49. Bathymetry of the top-Tertiary surface offshore the Somme River (Lassue, 2010).

Acknowledgements

Thanks to Professor François BAUDIN who provided some digitized pictures from literature. It saved us some time.

Thanks to the Oscar Savreux Quarries, especially Olivier LECOEUR, Director, for their financial support.

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Seals at low tide on a mixed flat. A colony of a few hundreds seals are present in the bay and its vicinity throughout the year. Notice some linear traces ending where the seals are resting. Picture: Les Editions Gaud, Syndicat Mixte Baie de Somme – Grand Littoral Picard.

Tidalites 2012 Conference
Caen, France, July 31 – August 2

Post-conference Field Trip

The Mont-Saint-Michel bay (NW France)

Facies, sequences and evolution of a macrotidal embayment and estuarine environment

August 3 - 5, 2012

Leaders

Bernadette Tessier[1], Chantal Bonnot-Courtois[2], Isabelle Billeaud[3], Pierre Weill[4], Bruno Caline[3], Lucille Furgerot[1]

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[4] MINES ParisTech, Fontainebleau

Photo P. Gigot

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Field guide prepared by

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The book "The Bay of Mont-Saint-Michel and the Rance Estuary" by Bonnot-Courtois *et al.* (2002)* is offered to the participants as a complementary document to this field guide. Reference to the text or figures included in this book is indicated as follows: e.g. *Text § 4.3.3, p. 183* or *Fig. 1-2-2, p. 33*. Note that a full list of bibliographic references is also available in the book.

*Bonnot-Courtois, C., Caline, B., L'Homer, A. & Le Vot, M. (ed.). – La baie du Mont-Saint-Michel et l'estuaire de la Rance. Environnements sédimentaires, aménagements et évolution récente. [*The bay of Mont-Saint-Michel and the Rance estuary. Recent development and evolution of depositional environments*]. – Bull. Centre Rech. Elf Explor. Prod., Mém. 26, 256 pp. 158 fig., 1 pl., août 2002, TotalFinalElf – CNRS- EPHE

We are thankful to Total for providing this Memoir 26 as well as a backpack to each participant. We also thank the "Syndicat Mixte Baie du Mont-Saint-Michel" for providing detailed documentation on the restoring operation of the Mont Saint Michel's maritime character.

General programme

Day 1 – The western embayment

Friday, August 3, 2012			
High tide	Low tide	High tide	Tidal range
9h00	16h00	21h15	12 m

Departure from Caen: 9h00			
Stop 1	10h30 – 12h00	Mont Dol: General presentation of the Bay and of the field trip	
Stop 2	12h30 – 14h30	Pointe du Grouin: rocky entrance of the Bay	picnic
Stop 3	15h00 – 17h30	Hirel - Vildé area: mixed flat and shelly banks	
Return to the Hotel (La Croix d'Or, Avranches) at 18h00			Diner at 20h00

Day 2 – The embayment/estuary transition and tide-dominated inner estuary

Saturday, August 4, 2012			
High tide	Low tide	High tide	Tidal range
9h30	16h50	21h50	11,5 m

Departure from the Hotel: 07h30			
Stop 4	08h00 – 10h30	The Mont-Saint-Michel (Tourism, Environmental management operation)	
Stop 5	11h00 – 14h00	Chapelle Ste Anne: sandflat and litho-bioclastic banks	picnic
Stop 6	14h30 – 17h30	Gué de l'Epine – Pontaubault: Inner estuary	
Return to the Hotel (La Croix d'Or, Avranches) at 18h00			Diner at 20h00

Day 3 – The northeastern wave-dominated coastline

Sunday, August 5, 2012			
High tide	Low tide	High tide	Tidal range
10h10	17h20	22h25	11 m

Departure from the Hotel at 07h45 (with all the luggage)			
Stop 7	08h30 – 09h30	Grouin du Sud: Tidal bore passage	
Stop 8	10h00 – 12h00	Dragey: wave-dominated sandy barrier and back barrier system.	
Stop 9	12h30 – 14h30	Saint-Jean-Le-Thomas: Holocene evolution	picnic
Stop 10	15h00 – 17h00	The Champeaux Hermelle reef (polychaete bioherm)	
Departure for Caen around 17h30 (arrival to Caen around 19h)			

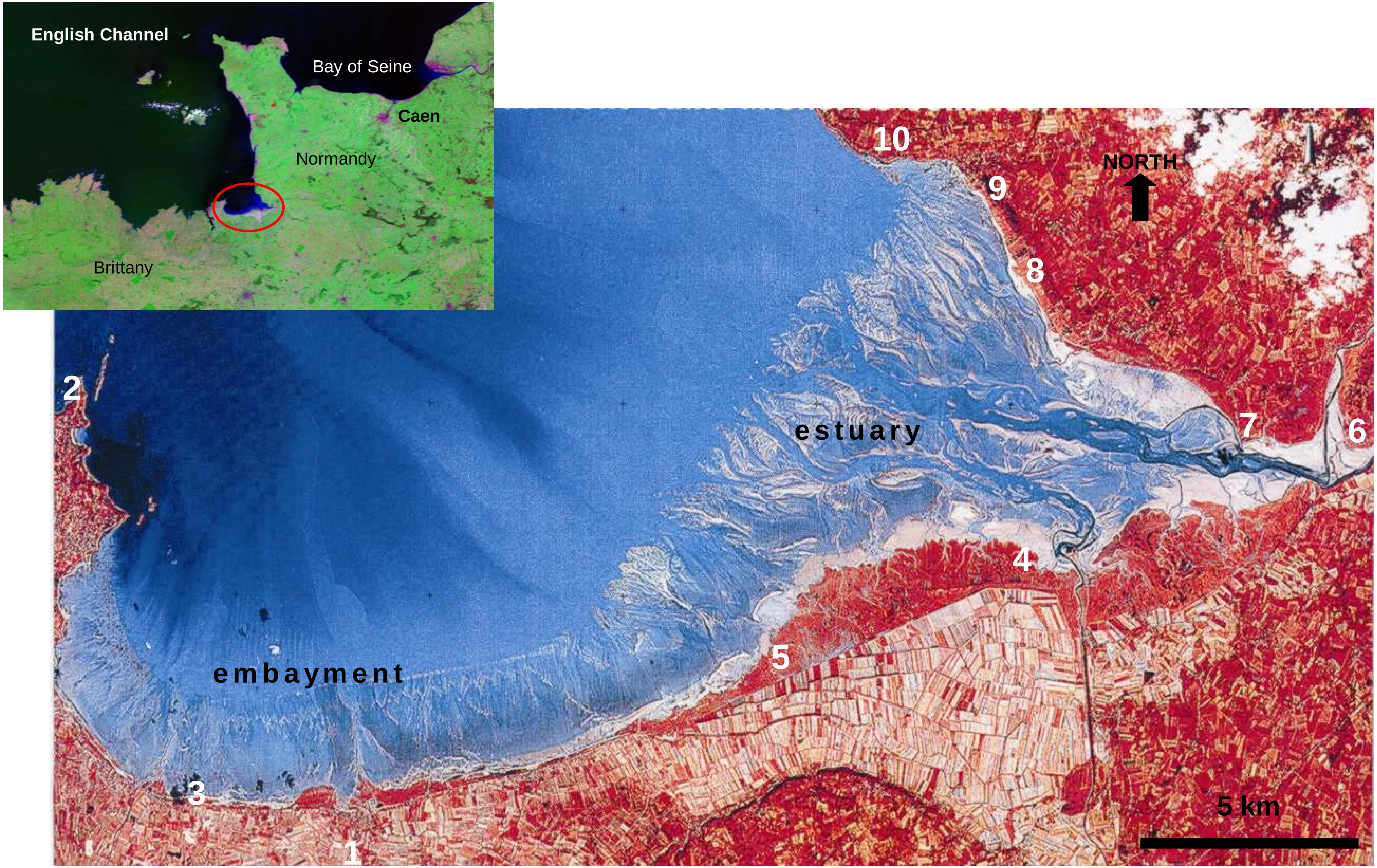


Figure 1. Location of the different stops during the field trip (SPOT Image, February 28, 1999; Processing by J. Le Rhun & C. Bonnot-Courtois, UMR CNRS Prodige, EPHE)
1: Mont Dol; 2: Cancale-Pointe du Grouin; 3: Vildé-Hirel; 4: Mont-Saint-Michel; 5: Chapelle Ste Anne;
6: Gué de l'Epine –Pontaubault; 7: Grouin du Sud; 8: Dragey; 9: St Jean le Thomas; 10: Champeaux

Day 1

The western embayment

Objectives

General presentation of the Mont-Saint-Michel bay (hydrodynamics, geological context, morphosedimentary organization, sediment distribution) and of the field trip. Focus on: - The wave-generated shell banks developed on the upper tidal flat (facies, geometry, morphogical evolution and migration rate); - The Holocene infilling (Dol marshes, embayment, paleogeographic reconstruction, Seismic and core data).

Successive stops

Stop 1	10h30 – 12h00	Mont Dol: General presentation of the Bay and of the field trip
Stop 2	12h30 – 14h30	Pointe du Grouin: rocky entrance of the Bay, hydrodynamics
Stop 3	15h00 – 17h30	Hirel - Vildé area: mixed flat and shelly banks

Stop 1 – The Mont Dol

General presentation of the Mont-Saint-Michel Bay

As the Mont-Saint-Michel and Tombelaine, the Mont Dol is one of the Cadomian leucogranite intrusions that rise above the bay. But contrary to the Mont-Saint-Michel and Tombelaine, which are still situated in a marine environment, the Mont Dol is located in the middle of the Dol Marshes (Stop 1 – Fig. 1).
From the Mont Dol a nice panoramic view on the flat landscape of the Dol marshes is offered. The geological context (Fig. 2), the general morphosedimentary characteristics and the Holocene evolution of the Bay (*Text § 4., 153-196*) are discussed from this point.

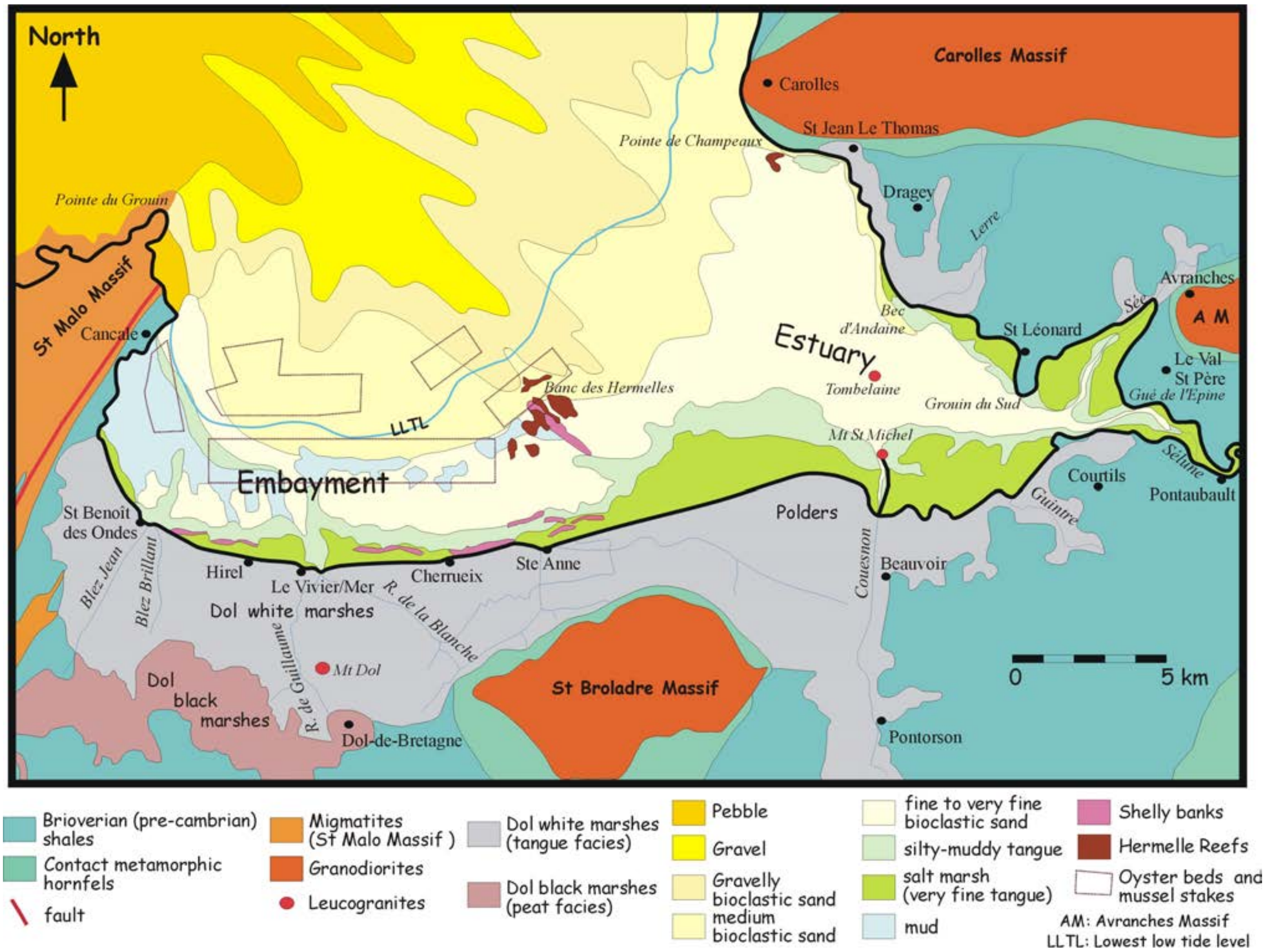


Figure 2. Simplified geological and sedimentological map of the Mont-Saint-Michel bay (after Larsonneur and coll., 1989 ; L'Homer *et al.*, 1999)

The Mont-Saint-Michel Bay (MSMB) belongs to the geological frame of the Armorican Massif (L'Homer et al., 1999). Its substrate is made of pre-cambrian sedimentary rocks (Brioverian turbiditic shales) and igneous rocks (Cadomian granites) (Fig. 2).
As most coastal landscapes around the English Channel, the morphology of the MSMB is a heritage of the plio-pleistocene glacio-eustatic fluctuations. During sea-level drops and lowstands, rivers incised the substrate and shaped the depression that were filled during subsequent rise and highstand. Because the regional subsidence is negligible, only the last post-glacial transgression is recorded into the infilling of the MSMB, the previous sea-level

fall having reworked almost all older marine sediments (L'homer et al., 1999). The last post-glacial sea-level rise was very rapid (about 6 mm/year, Lambeck, 1997; L'homer et al., 2002) and marine flooding already reached the most internal zones of the MSMB around 8000 yr B.P.(Fig. 3). Around 7000 - 6000 yr B.P. the transgression slowed down significantly (3 mm to progressively 1 mm / year, L'Homer et al., 2006, Fig. 4), allowing coastal wedge construction in general, and a rapid infilling of estuaries and embayments such as the MSMB (Fig. 5). Since that time of highstand sea-level or at least of very slow transgression, the different coastal environments composing the present-day landscape of the MSMB developed and evolved, each under the influence of specific geomorphogical, hydrodynamical and sediment supply conditions.

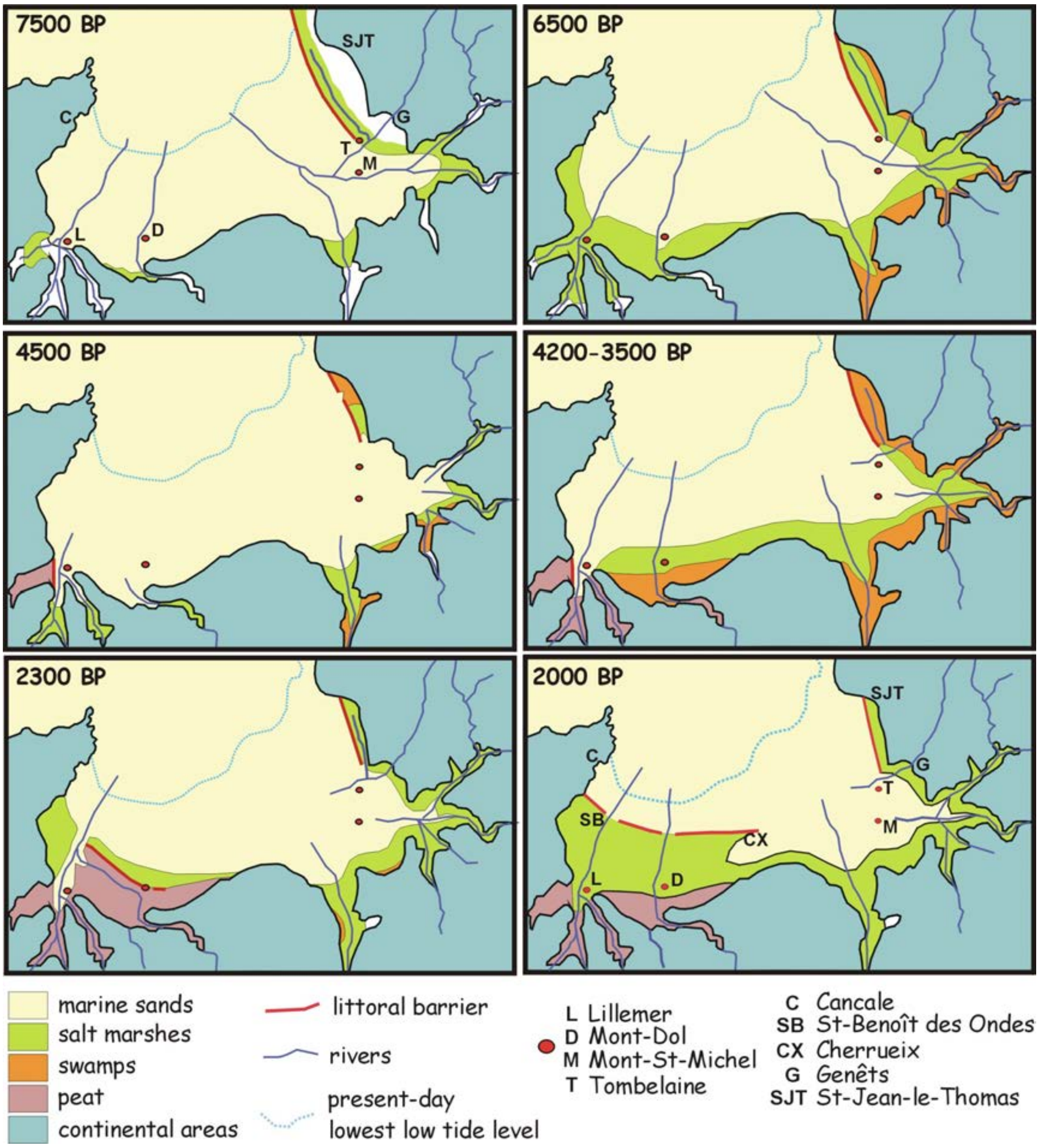


Figure 3. Paleogeographical reconstruction of the Mont-Saint-Michel bay
during the main stages of its Holocene evolution
(after Morzadec-Kerfourn, 1974; 1975, in Bonnot *et al.*, 2002 and in Larsonneur, 1989)

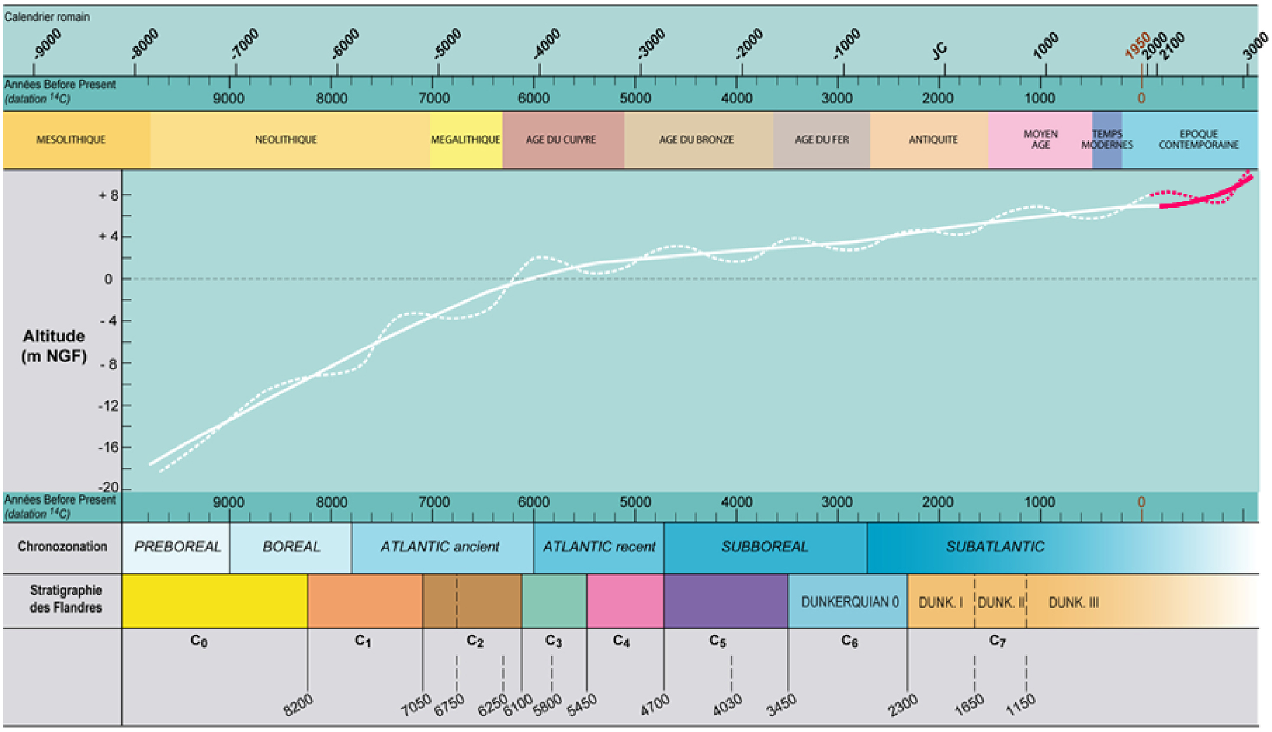


Figure 4. Reconstruction of the Holocene relative sea-level fluctuations in the Mont-Saint-Michel bay. Before 6500 y. BP, the rate is of about 10mm/y and then decreased to 3 to 1 mm/y (After L'Homer et al. 2006).

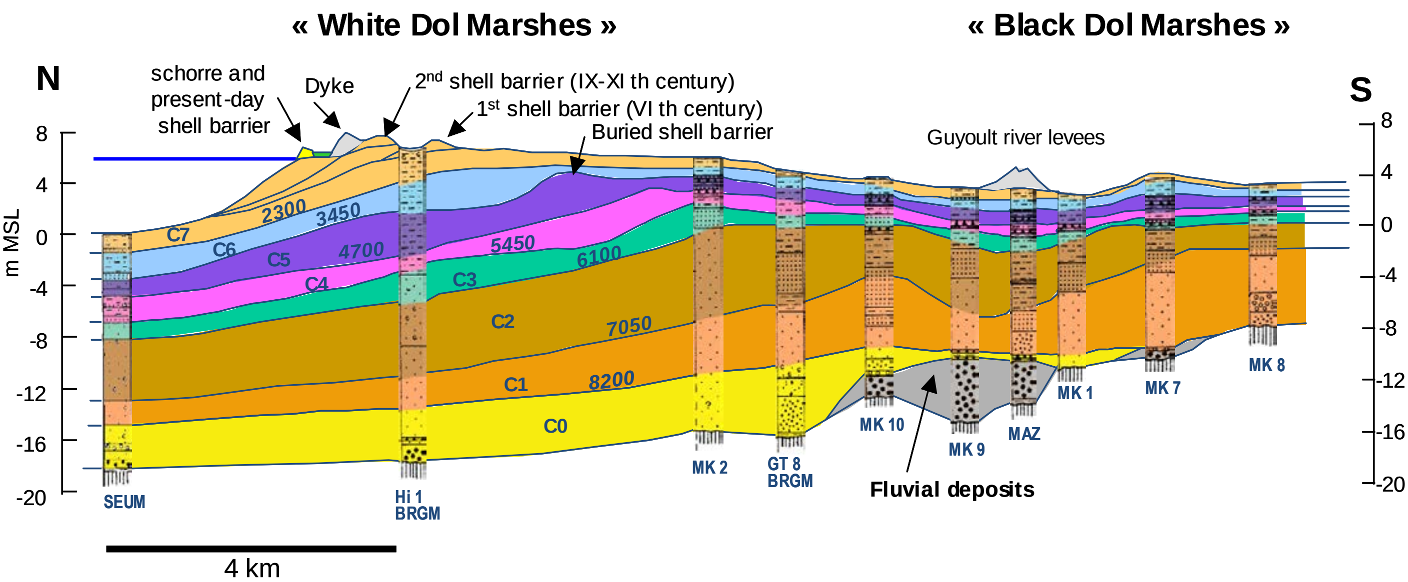


Figure 5. Schematic cross-section of the Holocene coastal wedge preserved below the Dol marshes (After L'homer et al., 2002). This reconstruction mainly based on drillholes into the Dol Marshes shows the passage from a vertical aggradation to a progradation at around 6500 y. BP, i.e. when rise in sea-level slowed down.

Stop 2 – The Pointe du Grouin

Western rocky coastline, tidal current pattern

With clear weather, “La Pointe du Grouin” provides a panoramic view on the whole MSMB, especially on the western rocky coastline and the western embayment.

The Northwestern extremity of the MSMB is characterised by very powerful currents during the ebb “flushing” due to the effect of narrowing between La Pointe du Grouin and l’Ile des Landes (Fig. 6). Tidal current velocity exceeds 1m/s during mean tidal range (*Fig. 1-2-2, p. 33*).

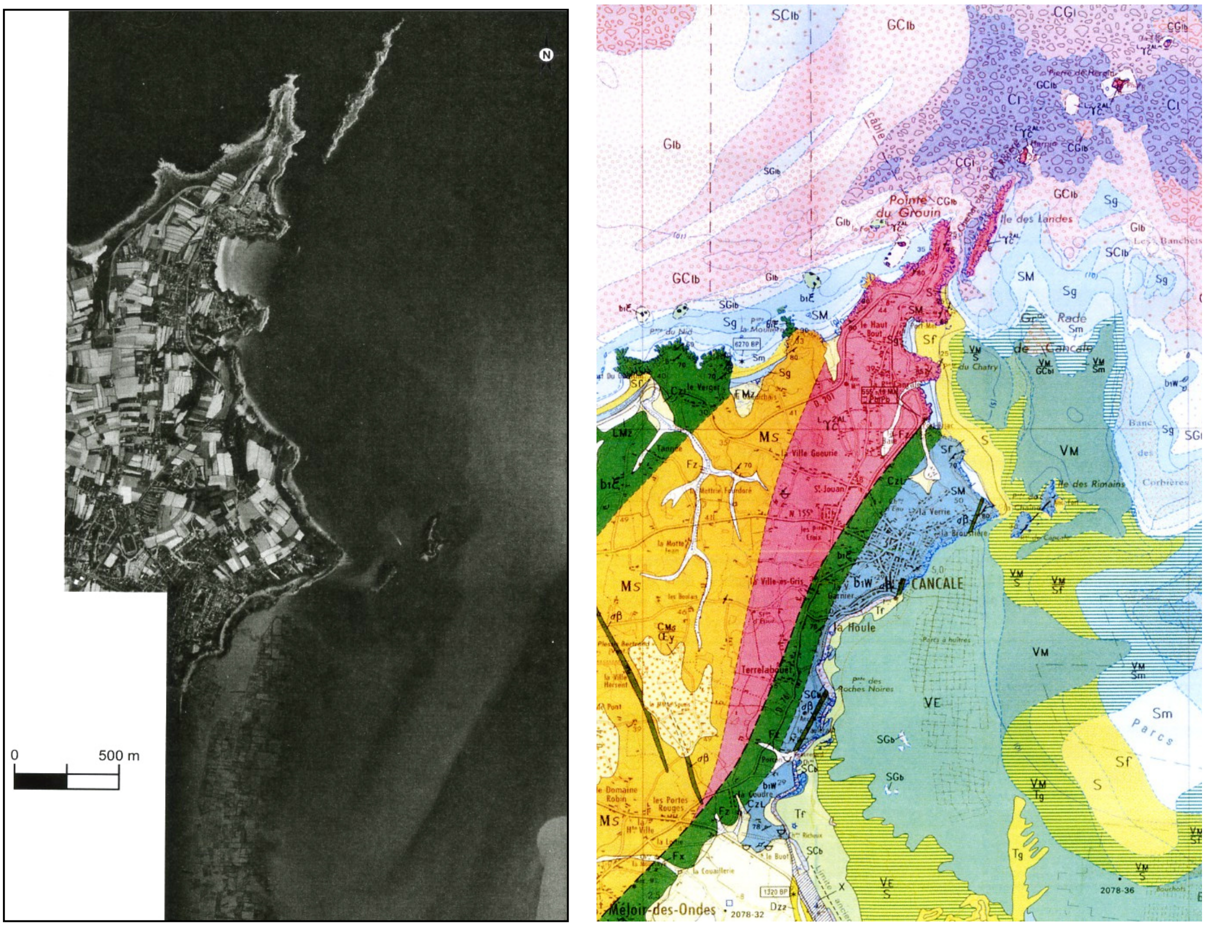


Figure 6. Aerial photograph (IGN, 1982) and geological map (1/50 000, BRGM) of the "Pointe du Grouin" area. The impact of the hydrodynamic regime on sediment distribution in this area is materialized by the sharp contact and rapid variation between the thin sheet of reworked relic deposits in the North (the western offshore entrance of the Bay of Mt-St-Michel) and the thick muddy and sandy deposits in the South (Bay of Cancale).

The western part of the MSMB corresponds to a large embayment stretching from Cancale to Cherrueix. Moderate tidal currents, of giratory type, occur in the Cancale bay which is moreover protected from dominant NW to W swells by « La Pointe du Grouin ».
The Cancale bay is characterized by the presence of extensive oyster beds (Fig. 7) settled on the lower and middle mud flat since the beginning of the XXth century *(Text § 3 .1, p. 99; Fig. 3.1.2. p. 102).*

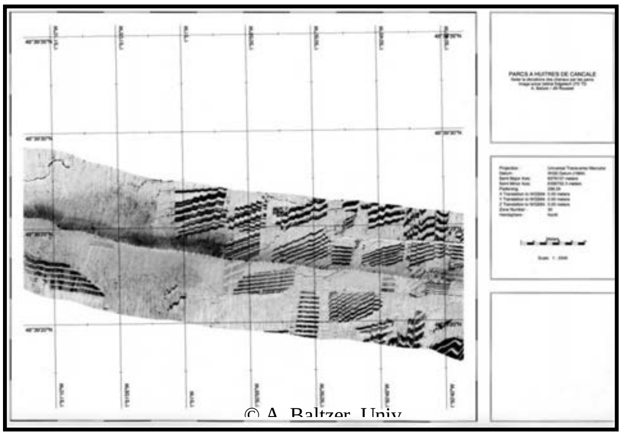


Figure 7. Aerial photograph (by C. Bonnot-Courtois) and side scan sonar image (by A. Baltzer, Univ. Caen/ CNRS) of the Oyster beds in the Bay of Cancale (western MSMB).

From Château Richeux westward, the coastline is outlined by the Dyke of Brittany, spreading over almost 20 km, and settled on a natural coastal barrier sheltering the Dol marshes from marine submersions. In this southern part of the MSMB, the upper tidal flat is marked by numerous shelly ridges concentrated in four sectors, respectively from west to east (Fig. 2 and Fig. 8): Saint-Benoît des Ondes, Vildé-Hirel (Stop 3), Cherrueix and Chapelle Sainte-Anne (cf. day 2, Stop 5).

The intermediate and lower parts of the tidal flat are occupied by two types of human installations: traditional fishing grounds and mussel farms arranged in regular lines parallel to the coastline (cf. Fig. 8) *(Text § 3 .1, p. 99; Fig. 3.1.4 & 3.1.5, p. 103).*

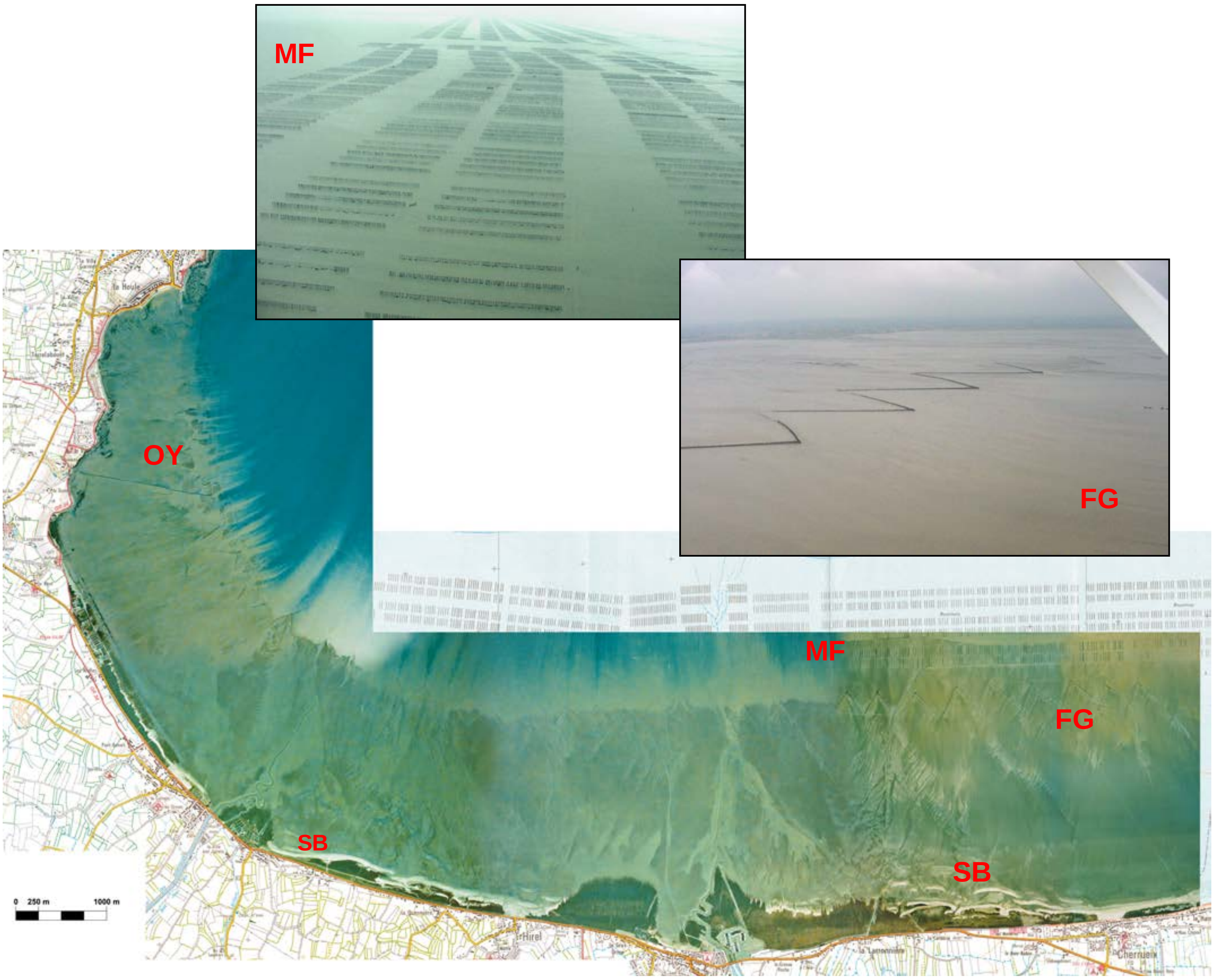


Figure 8. The Western embayment: composite aerial photograph (IGN, 1996) superimposed on the 1/25 000 IGN topographic map (A. Dréau). The photograph, taken at mid-ebb, displays particularly well some of the main features that characterize the western embayment: The oyster beds of the Bay of Cancale (**OY**), part of the mussel farms (**MF**), the old fishing grounds (**FG**), and the upper intertidal shell banks (**SB**) at Hirel – Vildé (Stop 3) and Cherrueix (Aerial oblique photos C. Bonnot-Courtois).

Stop 3 – Hirel-Vildé area

Shelly banks and mixed flat

As mentioned previously, the upper tidal flat of the southern embayment, is outlined by numerous shelly ridges that form and migrate progressively onshore under the action of swell and swash action. They are concentrated along four sectors from west to east: Saint-Benoît des Ondes, Vildé-Hirel, Cherrueix and Chapelle Sainte-Anne (*Text § 2.1.1., p.45-56*).
Stop 3 is dedicated to the Hirel-Vildé sector illustrated by the aerial photographs below.

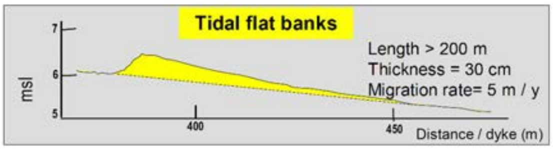
Aerial photos of the shell banks in the Hirel-Vildé sector

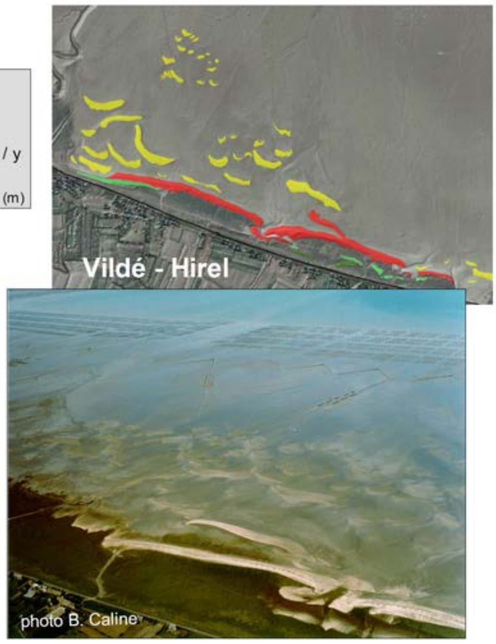
Two main aspects dealing with the shell banks are discussed during Stop 3:

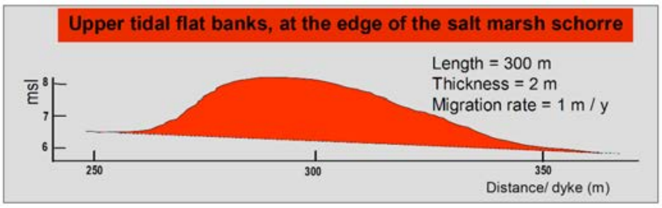
1) The general morphological characteristics and dynamics of migration of the banks based on recent lidar data and the works by Bonnot-Courtois et al., 2004 ; Bonnot-Courtois, 2012 (Plate 1).
2) The hydrosedimentary processes of construction, internal architecture and factors of evolution of the banks based on the PhD works of Pierre Weill (2011, Univ. Caen), Weill et al., 2010a, 2010b, in press (Plate 2).

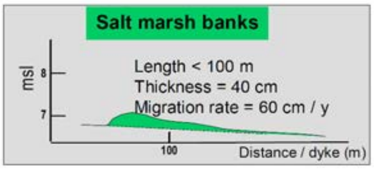
Plate 1 – Shell banks
Morphological characteristic and dynamics of migration

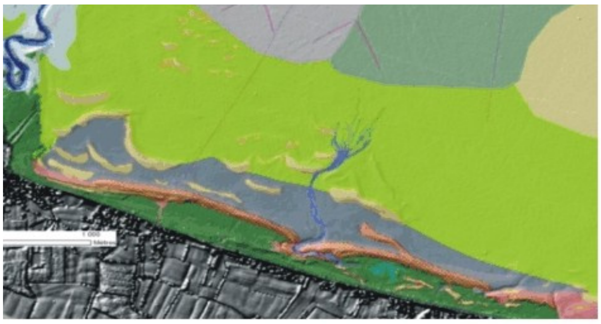
Three morphological types of shell banks can be distinguished: tidal flat banks, upper tidal flat banks and salt marsh banks (Bonnot-Courtois et al., 2004). Lidar data acquired in 2002 allowed defining accurately the altitude of the banks and by this way determining their dynamics with respect to tidal submersion (Lidar, 2002, © Fondation TOTAL et Ifremer, in Bonnot et al., 2007b, Bonnot-Courtois, 2012).



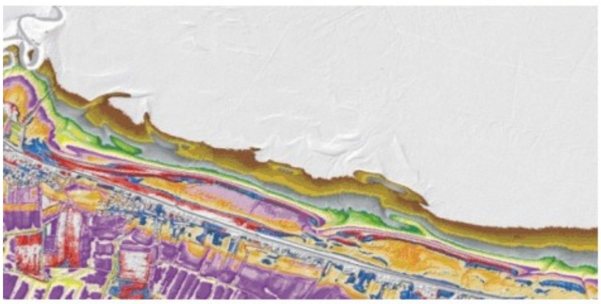




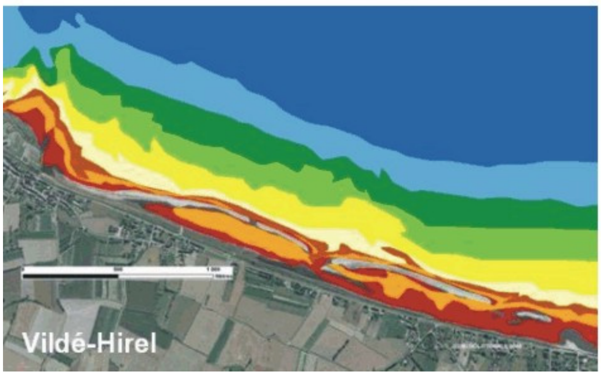




Morphosedymentary map of the upper tidal flat and banks in the Vildé-Hirel area (sedimento-logical data combined with lidar data.



MSL	Tidal range	% >/y
< 4 m		
4 - 4,5 m	60 (10,75 m CM)	65 %
4,5 - 5 m	70	50 %
5 - 5,5 m	75	42 %
5,5 - 6 m	80	36 %
6 - 6,5 m	90	20 %
6,5 - 7 m	100	8 %
7 - 7,5 m	110	3 %
7,5 - 8 m	115 (14,25m CM)	1 %
> 8 m		



Mean high tide levels
(neap to spring tides)

Coeff. 30
Coeff. 40
Coeff. 50
Coeff. 60
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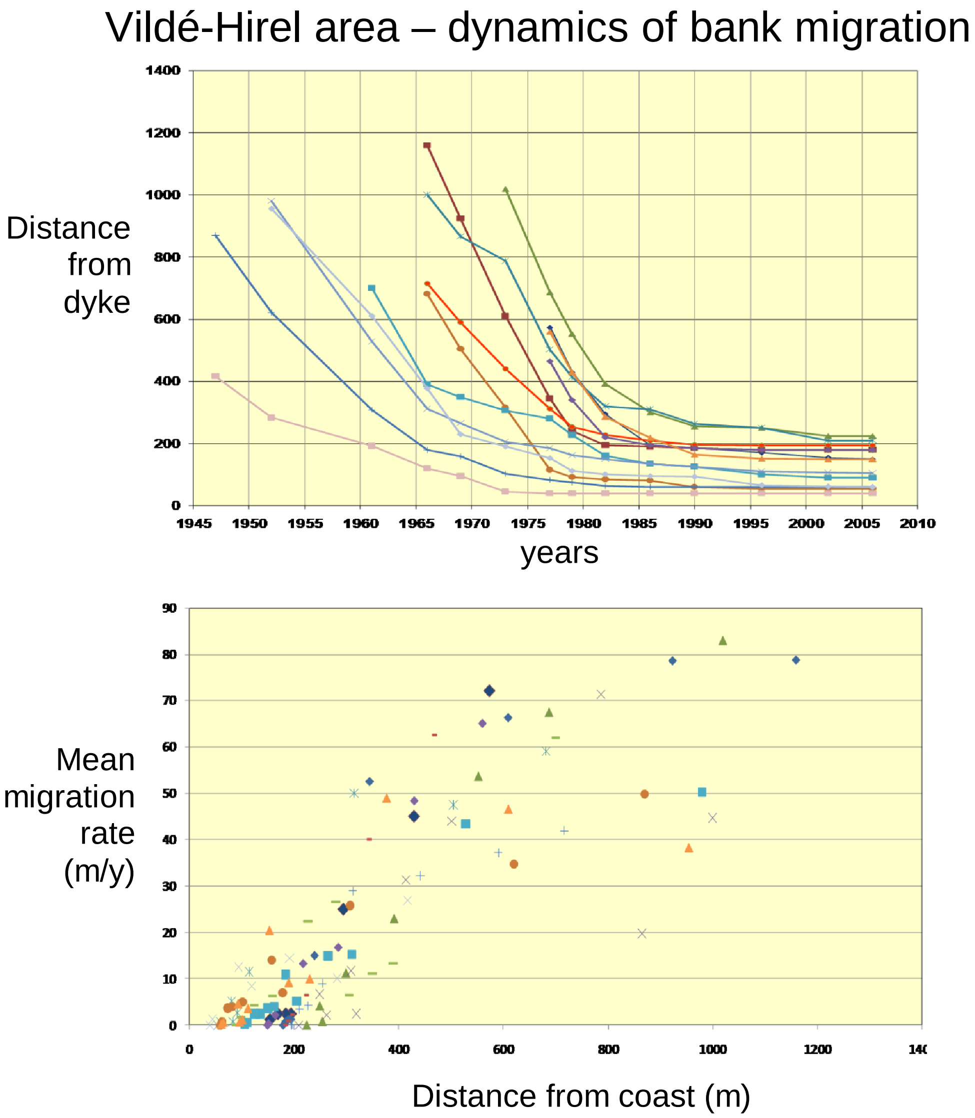


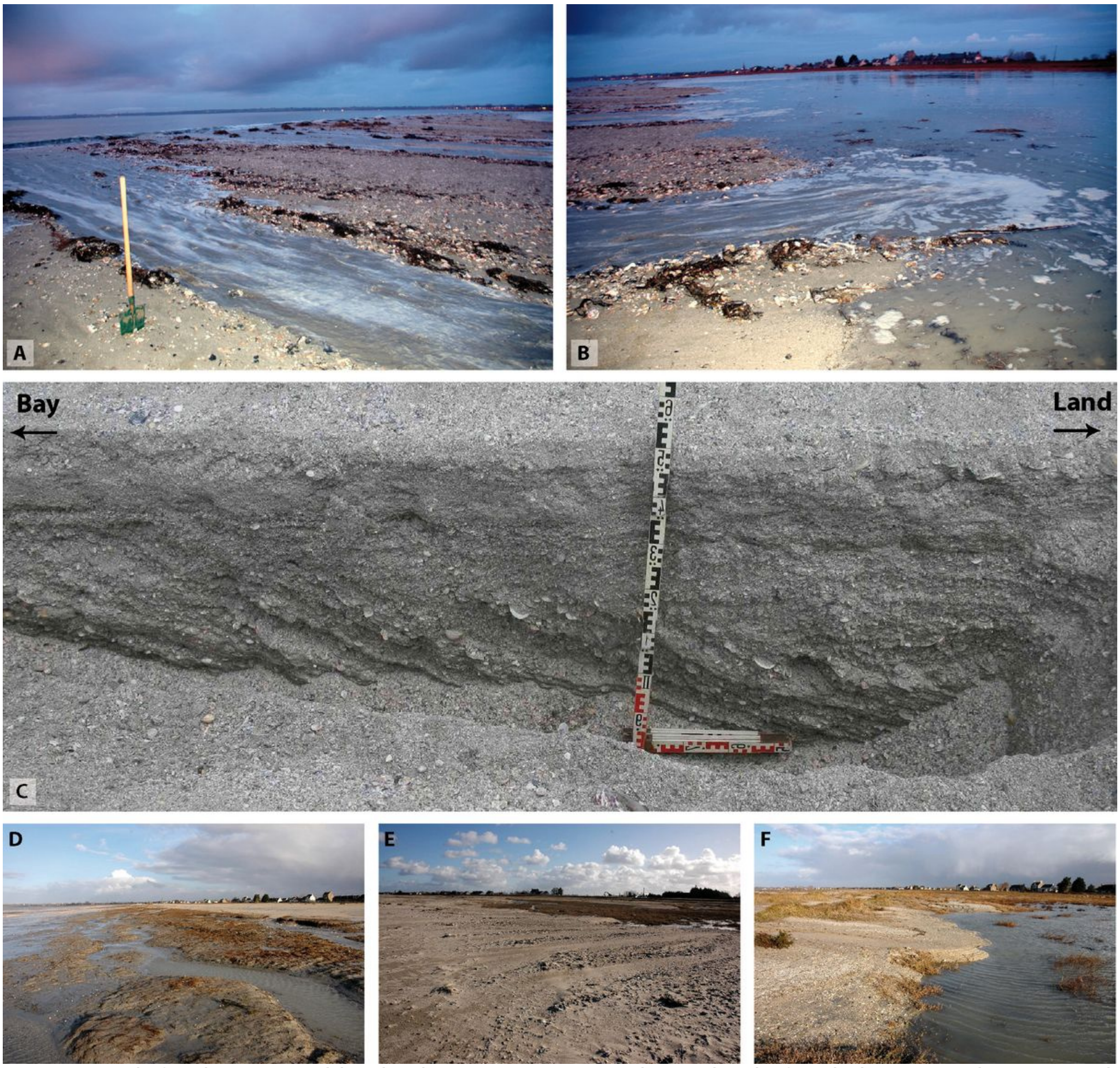
The comparison of successive aerial photographs since the 1940 allows specifying the dynamic of migration along the flat of the different types of banks. In the Hirel-Videl area, it appears that (:

- The dynamic of migration is regular and similar for tidal flat banks having the same origin on the mid flat;
- Migration rate slowed down around 1980. From that moment, banks are almost stabilized;
- The rate of migration of the tidal flat banks decreases from 50-80 m/y on the mid-flat to less than 10 m/y on the upper flat. When the banks reach the uppermost flat, their thickness increases and their migration rate slows down to only a few m/y. The oldest banks, anchored into the salt marshes, are almost totally stabilized;
- This dynamic of migration is partly controlled by the annual rate of submersion of the bank by tides, since sedimentary processes for bank migration are active only during high tides.

Plate 2 – Shell banks
Hydrosedimentary processes, Internal geometry, factors of evolution
Vildé – Hirel area

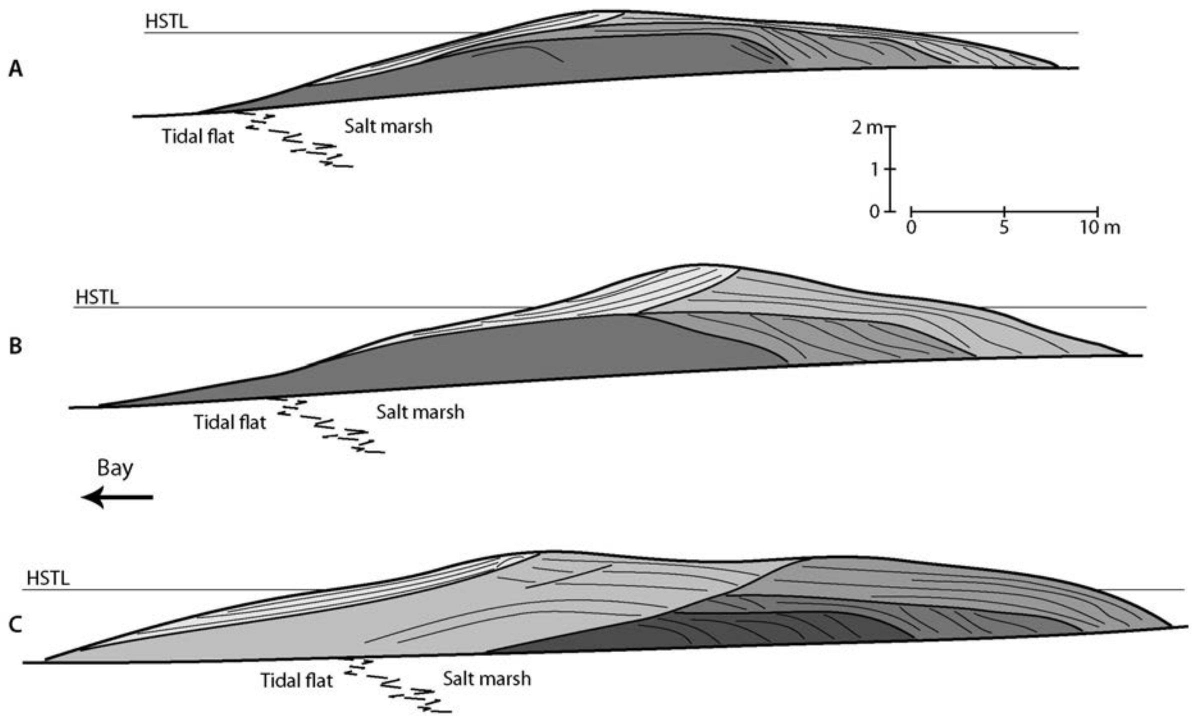
Extensive field work and flume experiments have been conducted on the upper tidal flat banks, also defined as beach ridges, to understand the hydrodynamic and sedimentary processes involved in their evolution and their internal architecture. The study included ground-penetrating radar (GPR) survey, trenching, CT-Scan, porosity and permeability measurements on cores, and wave flume modelling.

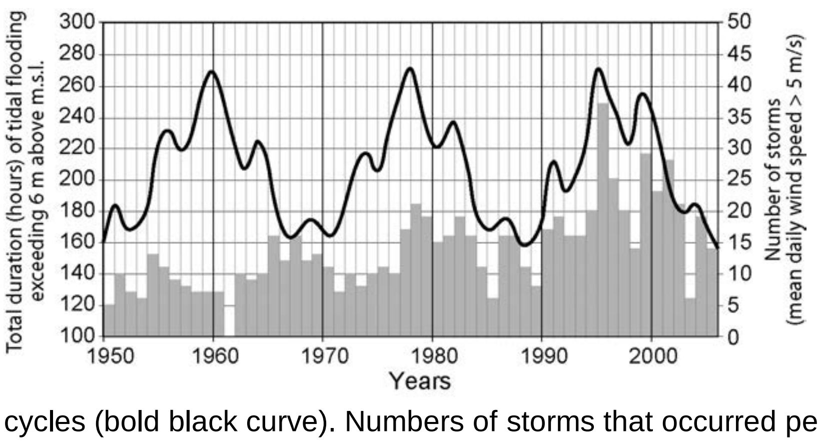
The main processes that results in the beach ridges landward migration and construction are i) wave breaking and swash currents acting on the foreshore, and ii) overwash events that flow on the ridge backslope and pour in the flooded back-barrier.



A- Overwash flowing along ridge backslope. B- Overwash pouring in flooded salt marsh. C- Trench along a washover lobe showing high-angle landward-inclined foreset strata (sub-aqueous sedimenation) overlay by low-angle landward-inclined washover sheets (sub-aerial sedimentation). D- Eroded salt marsh deposit at the toe of beach ridge. E- Washover run-off channels on the ridge backslope. F- Washover lobes covering the salt marsh.

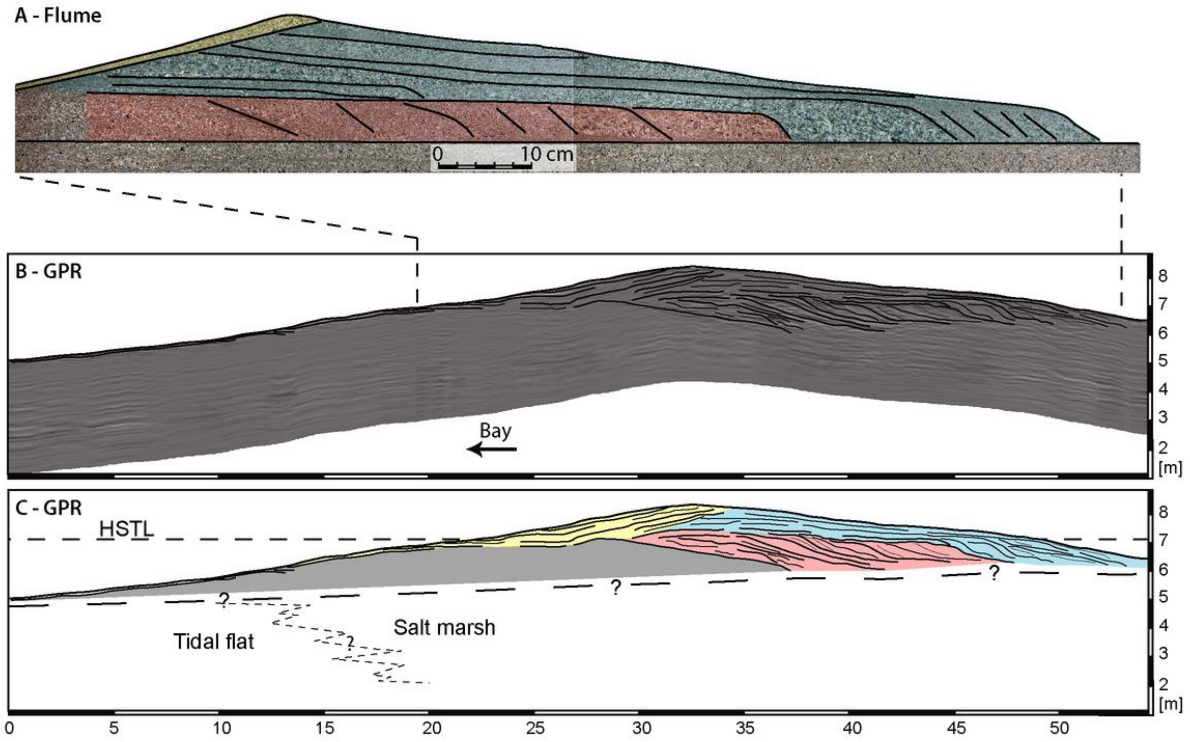
From GPR profiles, 3 stages of beach ridge evolution have been identidied (Weill et al. 2010b) : A) Early transgressive, B) late transgressive, and C) progradationnal stages. The evolution and internal architecture of the beach ridges are closely related to the level of tidal flooding during high spring tides (HST). Washover foresets are deposited as long as overwash occurs and pours in the flooded salt marsh. If ridge backslope is over-extended, overwash infiltrates before it reaches the salt marsh. Aggradation occurs on the backslope in the form of washover sheets. When vertical accretion prevent overwash to occur, beach ridges extend bayward by ridge ammalgamation in a progradational pattern.



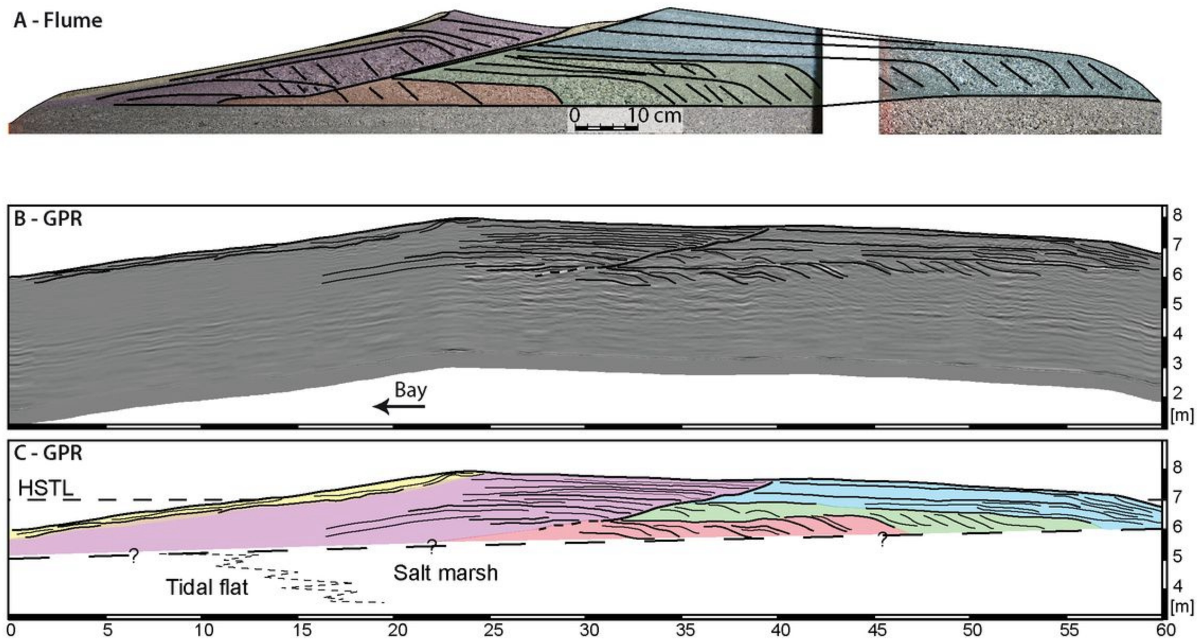


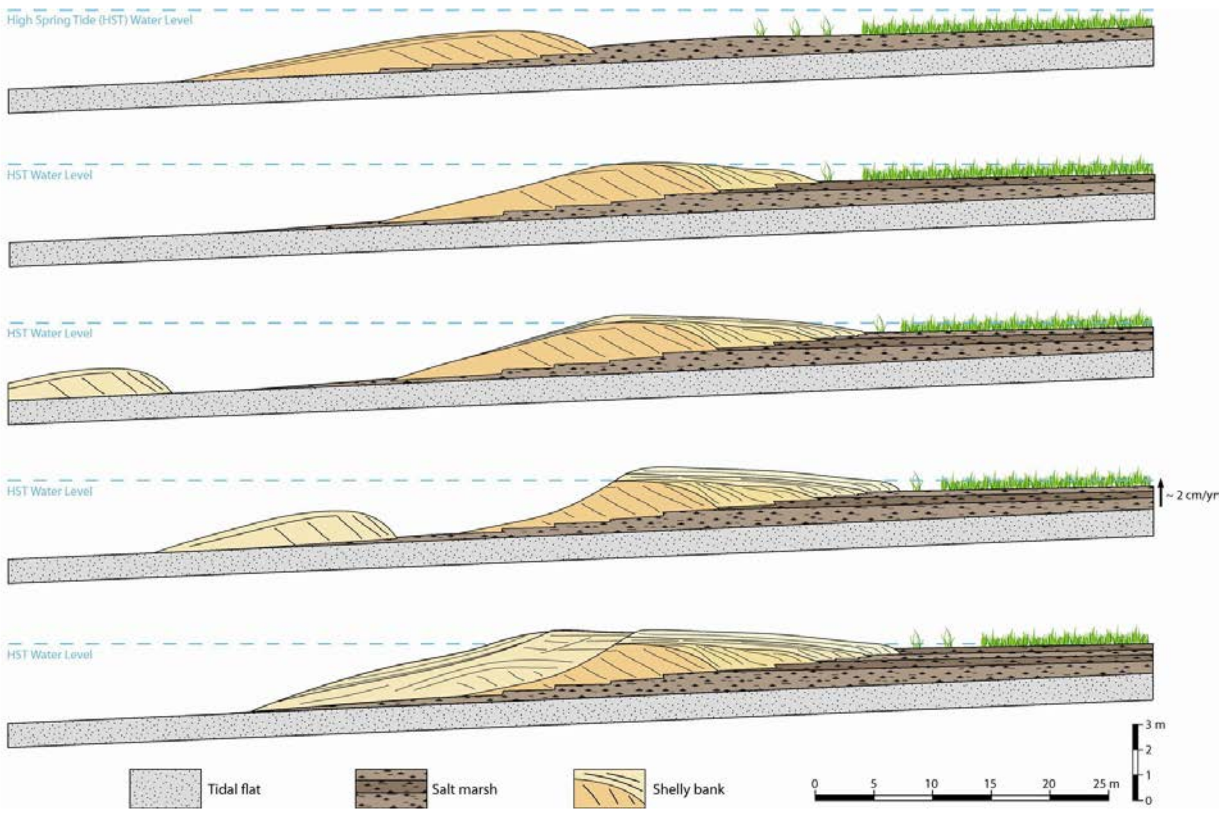
Beach ridges are activated only a few hours at high spring tides. Annual cumulated time of flooding exceeding 6 m above m.s.l., which corresponds to the level of beach ridge activation, is calculated. It oscillates between 160 and 270 h of flooding per year, following 4.4 and 18.6-yr tidal cycles (bold black curve). Numbers of storms that occurred per year during HST flooding are represented in light grey bars. It displays some variability and a global increase on the 1950-2000 period. Among other factors influencing beach ridge development (wave activity, sediment supply…), spring tide level is the parameter that shows the more variability on a multi-decadal time scale. It is thus supposed to have a major influence on the overall system dynamics.

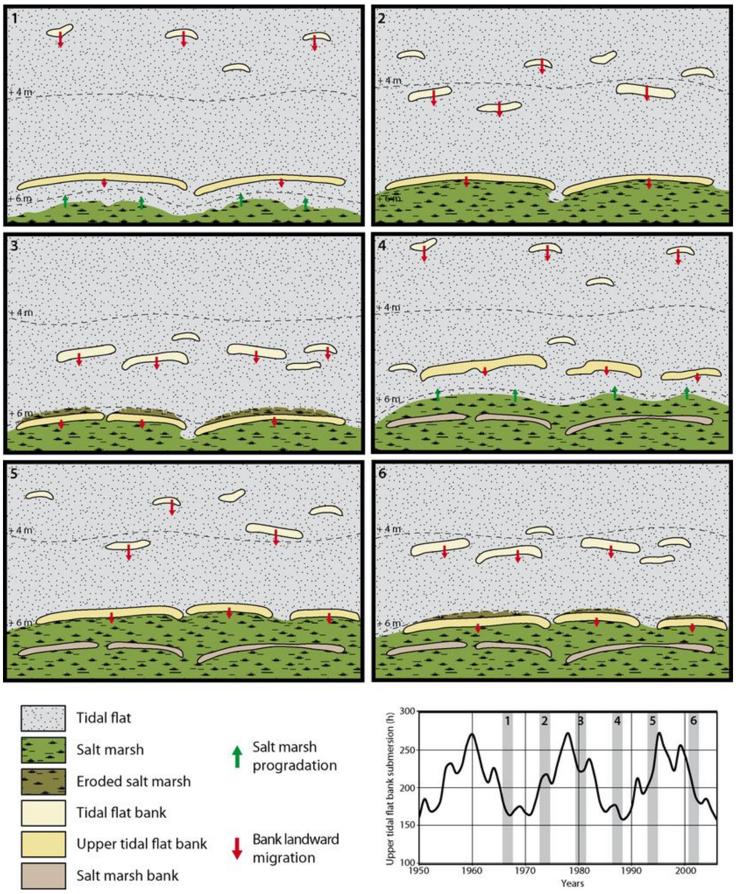
Flume experiments were conducted with natural sediment sampled in the field (Weill et al., in press). Constant wave parameters were used, and low-frequency water level fluctuations were generated to mimics the variations in the frequency and intensity of high spring tide flooding. Most of the landform morphologies and internal structure observed in GPR profiles have been reproduced in the flume. As low-frequency water level fluctuation was the only variable parameter in the experiment, it suggests that it is the main forcing parameter that controls beach ridges construction and evolution.



Comparison between flume experiment and GPR data for the late transgressive (top) and progradational (bottom) beach ridge stage. Red: Washover foresets. Green: Vertically stacked washover foresets. Blue: Washover sheets. Yellow: Beach deposits. Purple: Amalgamated sand ridge.







Based on field data and flume modeling results, a depositional model of the beach ridges influenced by low-frequency tidal level fluctuation is proposed. Periods of low frequency of HST flooding (troughs of 18.6-yr cycles) allow the stabilization of shell banks lower on the tidal flat. It creates favorable conditions for salt marsh progradation. Periods of high frequency of HST flooding (peaks of 18.6-yr cycles) trigger major reworking and landward migration of beach ridges by overwash events. 4.4-yr tidal cycles may be responsible for the formation of individual washover units. Of course, other parameters are involved in the process, such as storm wave activity, sediment supply, biological productivity, and human infrastructures (dykes).

Day 2

The embayment/estuary transition and tide-dominated inner estuary

Objectives

Focus on: - the litho-bioclastic ridges of the upper sandflats at the transition between the embayment and the estuary; - the sedimentary facies and sequences typical of the inner estuary (tidal rhythmites, tidal channel infill successions); - the operation of restoration of the maritime character of the Mont-Saint-Michel

Successive stops

Stop 4	08h00 – 10h30	The Mont-Saint-Michel (Tourism, Environmental management operation)
Stop 5	11h00 – 14h00	Chapelle Ste Anne: sandflat and litho-bioclastic banks
Stop 6	14h30 – 17h30	Gué de l'Epine – Pontaubault: Inner estuary

Stop 4 – The Mont-Saint-Michel

Tourism and restoration of the maritime character

The Mont-Saint-Michel, with around 3 millions visitors per year, is one of the major French touristic sites. In addition to the elegant gothic abbey settled on top of the granitic mount, the tides that surround regularly the site constitute the main attraction. However, due to the natural infilling of the estuary, dramatically enhanced by land reclamation since the XIXth century, rapid sediment and salt marsh accretion occur around the Mont-Saint-Michel (Fig.9, 10), so that the attractive spectacle of the incoming tidal bore is becoming very rare.

In order to restore the maritime character of the Mont-Saint-Michel, a research project was initiated in 1995, including hydrodynamical and sedimentological studies, numerical and physical models (Fig. 11). The project advancement, as well as the arising restoring operations that began in 2005 (Fig 12), are described in detail on the following web site: http://www.projetmontsaintmichel.fr (official web site of the "syndicat mixte Baie du Mont Saint Michel" for the restoration of the maritime character of the Mont-Saint-Michel).

Two aspects of the Mont-Saint-Michel:
a very busy touristic site (top) and a progressively "continentalized" environment (right).
Photos in http://www.projetmontsaintmichel.fr

The Mont-Saint-Michel and nearby Couesnon river and salt marshes during high spring tide. Because of sediment accumulation around the Mont, this spectacle is increasingly rare.

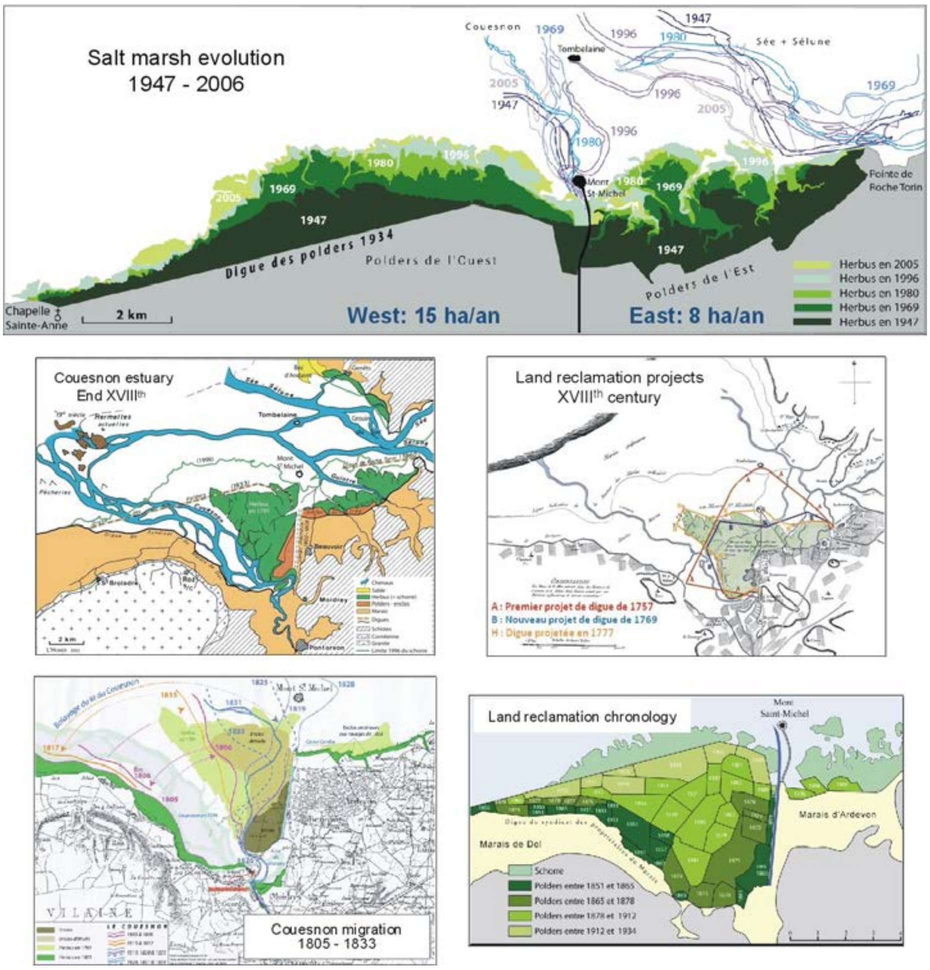


Figure 9. Evolution around the Mont-Saint-Michel since the XVIII[th] century. Before land reclamation operations, the Couesnon estuary was wider, with active migrating tidal channels. After unsuccessful attempts of land reclamation during the XVIII[th] century, the dyking of the Couesnon River was finally done to prevent migration and polder destruction. Land reclamation was achieved in the middle of the XX[th] century. Since that moment, salt marshes dramatically extended west and east of the Mont-Saint-Michel, leading to the present project of restoration of the maritime character (after Bonnot-Courtois, 2012).

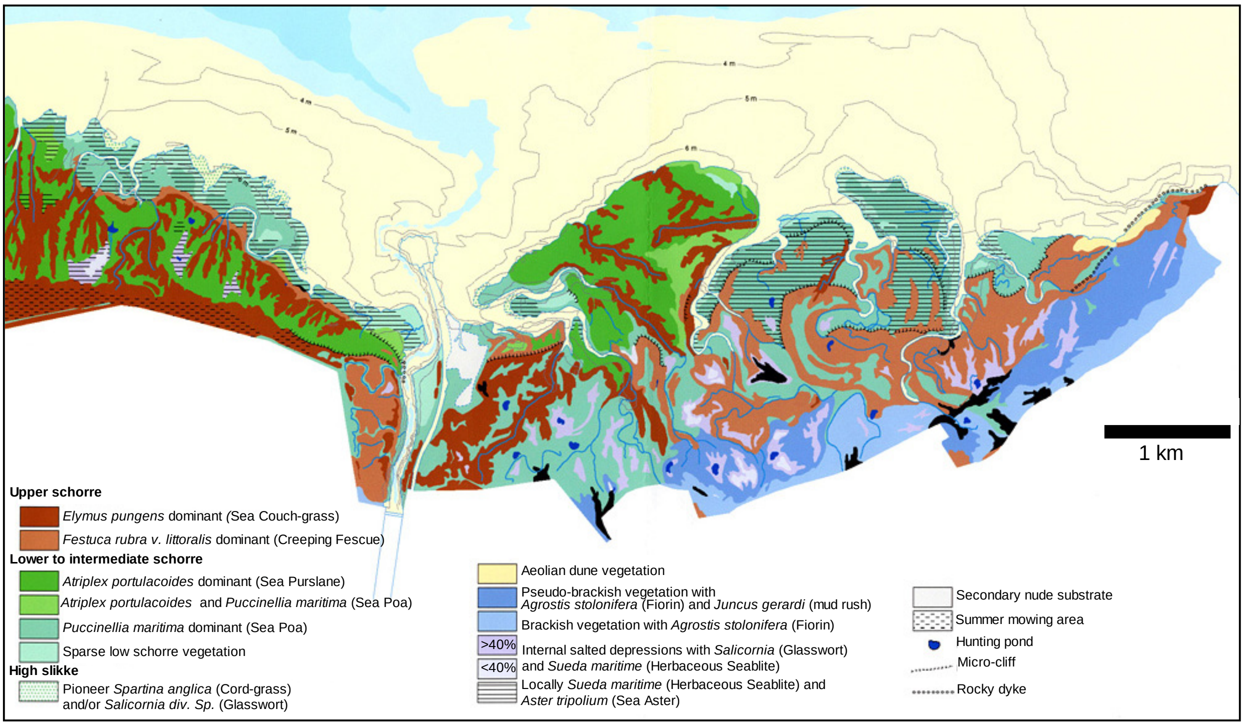


Figure 10. Salt marsh vegetation nearby the Mont Saint-Michel in 1999
(from Bonnot-Courtois and Levasseur, 2000)

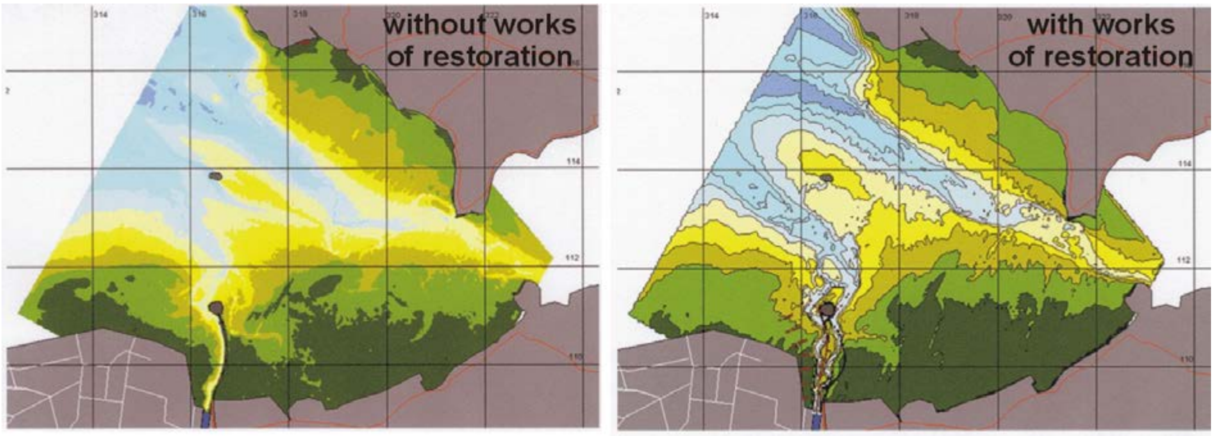


Figure 11. Some results of the numerical model developed in the framework of the project of restoration of the Mont Saint Michel's maritime character. The use of the model has been critical for defining the works to be done around the Mont.

Achievements at the project's halfway point (in http://www.projetmontsaintmichel.fr) (cf. Figure 12)

• **2009 -** The construction of the dam over the Couesnon, officially launched by the French Prime Minister in June 2006, is complete. This work, which is the cornerstone of the project's hydraulic aspect, began to remove sand from around the rock in May 2009. The public service delegation for visitor parking and transport was also awarded at the start of autumn 2009.

•**2010-2011 -** These years saw the start of the reception work (landscaped car park, reception and service buildings) and access work for the Mont (pedestrian footbridge and causeway from 2011) enabling a completely new approach to the rock. These years also marked the start of the hydraulic developments upstream and downstream of the dam (2011-2015) which will restore the Couesnon's hydraulic capacity to move sediment away from the rock.

•**2012 -** The new car park on the continent and the public transport shuttles are commissioned to bring visitors to the Mont.

•**2014 -** The pedestrian footbridge is open to visitors, pedestrians and shuttles, but also logistics (outside busy periods) and the Mont's permanent security services.

•**2015 -** More symbolically still, the operation is completed with the destruction of the causeway, which is over 100 years old (1879), which brings visitors from the continent to the Mont. The works to restore the Mont-Saint-Michel's maritime character are then complete. It will then take a few years for a wide strand to form around the rock and for the Mont to regain its full maritime landscape for many years to come.

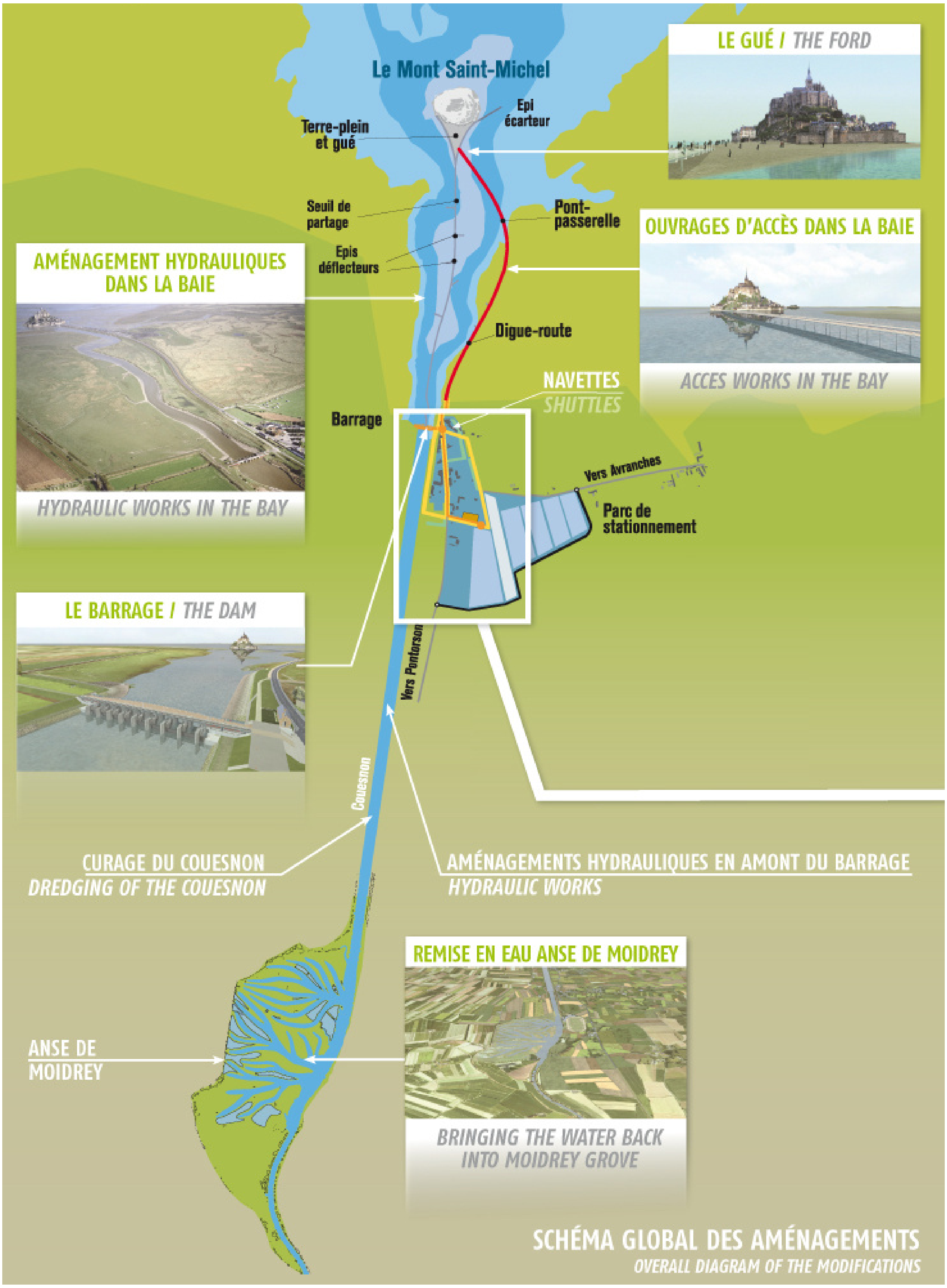


Figure 12. Diagram of the main operations planned to achieve the restoration of the maritime character of the Mont-Saint-Michel (http://www.projetmontsaintmichel.fr)

Visit of the monastery of the Mont-Saint-Michel

Some History

Welcome to the abbey of Mont-Saint-Michel, a monument of legend...

Rising up from an indefinable point in the sand and waves, Mont-Saint-Michel appears like a human challenge to the elements and time. A lost rock in a landscape otherwise made featureless by the power of the wind. Perched on its summit, the abbey is an invitation to come and discover its builders' wild inspiration - one which has driven so many to make this isolated place a universal meeting point for 708 years.

From up there, everything comes as a surprise: the amazing feats of medieval architecture, the power of nature, the light... everything. The abbey of Mont-Saint-Michel is listed by **UNESCO** as part of the World Heritage.

Saint Aubert's dream

How the cult of St. Michel came into being is told in a Xth century manuscript, La Revelatio ecclesiae sancti Michaelis. According to this original document, one night in 708, Aubert, Bishop of Avranches had a dream in which he saw the Archangel Michael who commanded him to build a church on the rock. And when the bishop dared to express doubts, the Archangel thrust a finger through his skull! Aubert therefore had a chapel erected on the site. In the troubled and uncertain times of the Merovingian era, the faithful were quick to embrace the cult of the Archangel and the rock became a place of pilgrimage, soon to be known as Mont-Saint-Michel.

The Wonder of the Western World

The Mount became part of Normandy in 933, and the number of pilgrims steadily grew from the XIth century on, as the number of miracles recorded there increased. Donations and legacies enabled the monks to build a Roman abbey with a large church on the very summit of the rock. By the end of the XIIth century, the number of Benedictine monks living there had risen to 60.

In 1204, the Capetian King Philippe Auguste took the Duchy of Normandy and the Mont Saint Michel by force, which caused some damage to the edifice. To make amends, the King ordered restoration work to be carried out and had new buildings erected, so that the work on Mont Saint Michel seemed never-ending. During the XVth century, during the Hundred Years' War, the English sieged and assaulted the island many times. But to no avail. The abbey's defences were reinforced and the Mont Saint Michel remained invulnerable.

The dark days of the Mont Saint Michel

In the XVIth century, the Mont began to fall into disrepair due to a lack of maintenance work and the decline of the monastic life. The situation was made worse by Protestant attacks during the Religious Wars, so that by the end of the XVIIIth century less than a dozen monks remained.

In 1793, the revolutionaries turned the abbey into a prison, and it was not until 1863 that an imperial decree put an end to this sacrilege.

Renewal

In 1872, the restoration of the Mont was entrusted to an architect working for the Monuments Historiques, Edouard Corroyer. For more than a century, a series of architects and wardens worked at fitting out and renovating the interior and exterior of the abbey. Since 1983, the upkeep and enhancement of the monument have been the responsibility of Pierre-André Lablaude.

Text in: **http://www.monum.fr/m_stmichel/indexa.dml?lang=en**

Stop 5 – The Chapelle Sainte-Anne

Litho-bioclastic banks and sandflat at the embayment / estuary transition

In the center of the MSMB, off "la Chapelle Ste-Anne" locality, at the transition between the embayment and the estuary, a spectacular bio-construction made by annelids (*Sabellaria alveolata* or "Hermelles") is settled on the lower tidal flat (Fig. 13). The reef system constitutes an area of important sediment storage, available for bank construction. The dynamic of shoreward migration of these banks is fairly similar to that of the banks of Vildé-Hirel and Cherrueix areas. Migration rate can reach locally 100m/y on the mid flat (Fig. 13). The banks that finally reach the upper tidal flat are composed of bioclasts (such as in the banks located west, at Vildé-Hirel and Cherrueix localities), but are characterized also by a high content in lithoclasts. This is the result of the selection by the annelids of bioclastic particles for construction of the reef, so that the sand that is not trapped into the reef, is enriched in lithoclastic elements.

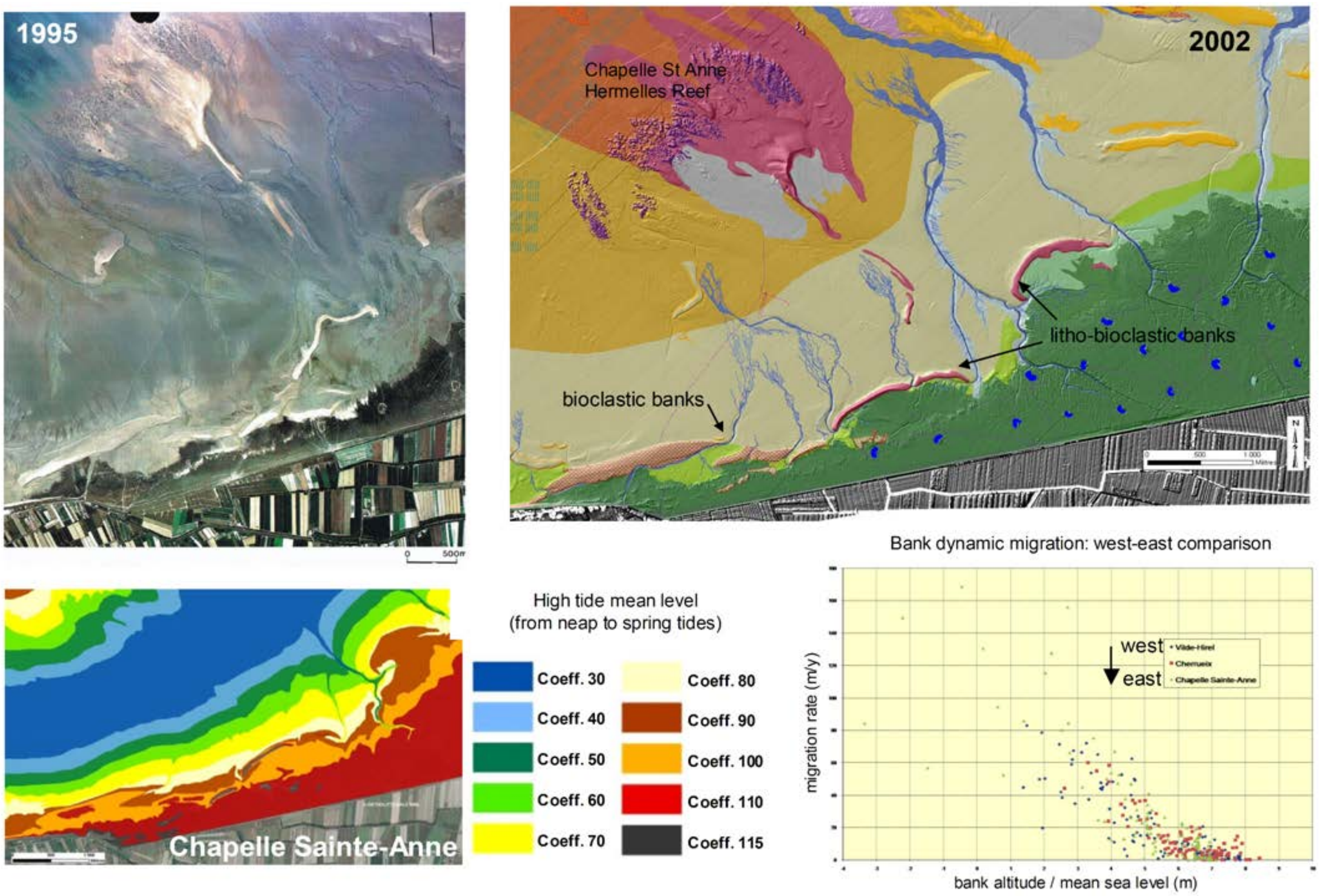


Figure 13. The litho-bioclastic banks of « La Chapelle Sainte-Anne ». These banks are supplied by sediments that are reworked from the Annelid reef area, located on the lower sand flat. The relative lithoclast richness into the banks is due to the impoverishment in bioclasts, these latter being used for the reef construction (after Bonnot-Courtois et al., 2007a; Bonnot-Courtois, 2012).

Stop 6 – "Gué de l'Epine – Pontaubault" area

Tidal facies and sequences in the inner estuarine environment

The Stop 6 "Gué de l'Épine- Pontaubault" area (*Text § 2.2.2, p.83*) belongs to the inner estuary (Fig. 14A). It is located on the right bank of the Sélune river where the estuarine system is made up of a single channel, bordered by the slikke and schorre domains which extend over a width of about 1km (this part of the estuary corresponds to the straight-to-meandering fluvio-tidal transitory domain of the morphosedimentary model for tide-dominated estuary by Dalrymple *et al.* (1990), Fig. 14B). Some 1-2 km downstream, the estuary consists in a vast area of sandy-silty slikke covered by a dense networks of tidal channels bordered by megaripples (the braided tidal system according Dalrymple's model, Fig. 14B).

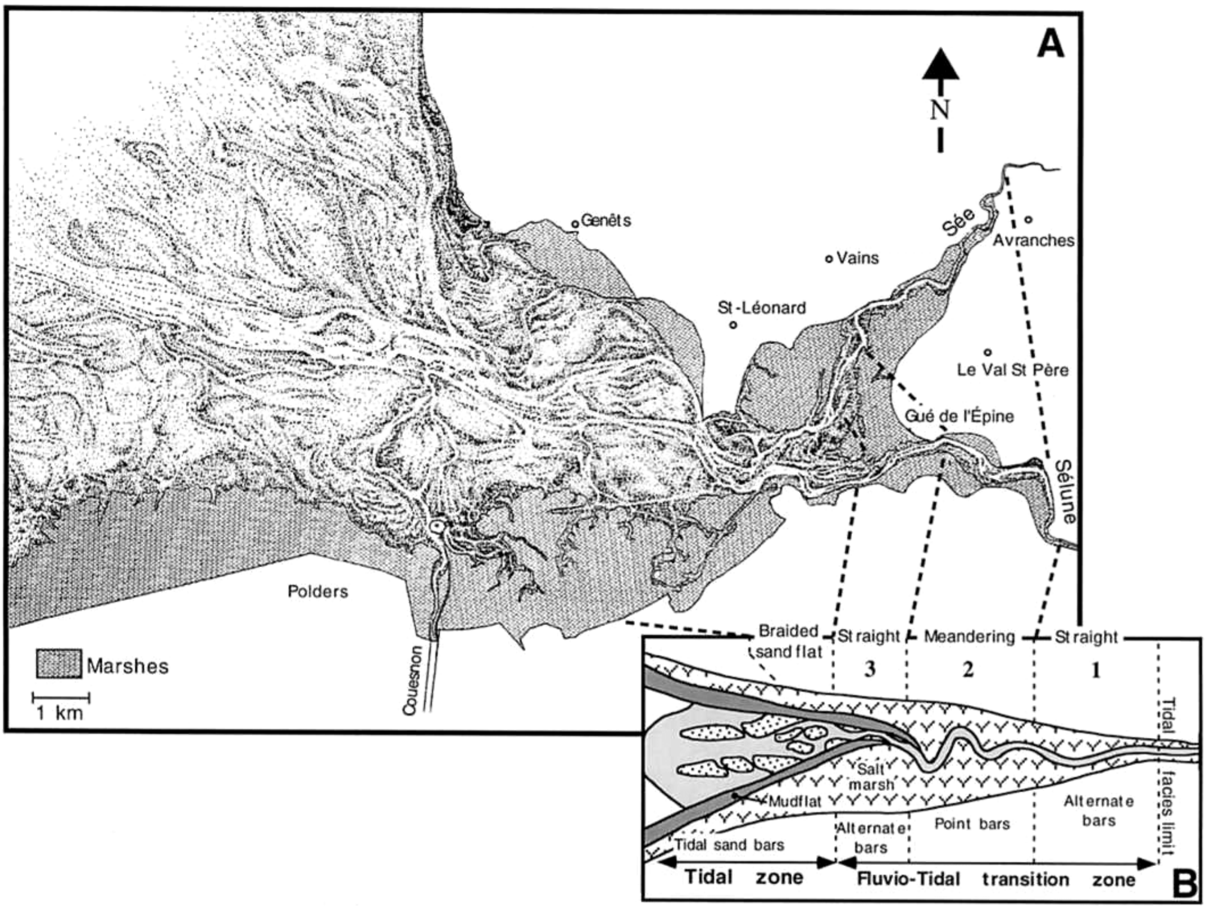
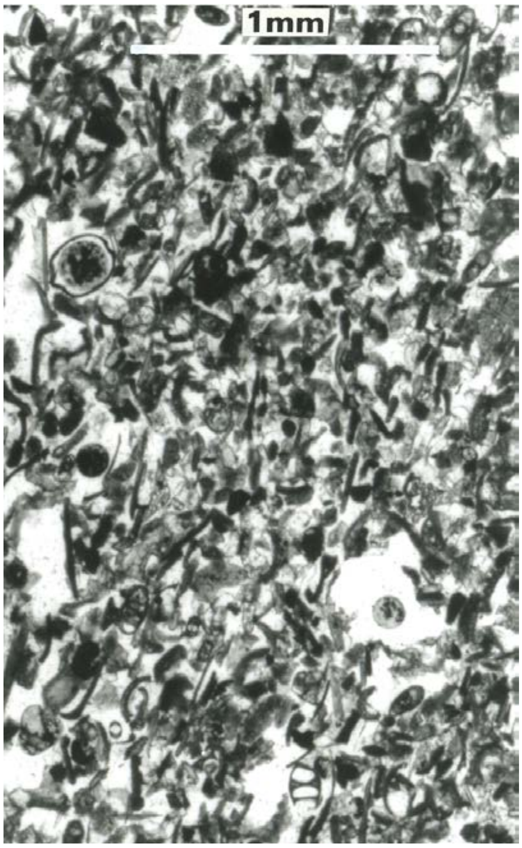


Figure 14. Morphosedimentary map of the estuarine system (A)
with indication of the concordances with Dalrymple's model (B)

The hydrodynamic of the site is controlled by tidal currents and in particular by the flood, faster than the ebb. The flood arrival is accompanied by a tidal bore (reaching about 80 cm during spring tides) which spreads into the channel with a velocity in the order of 3 m/s. The bore reworks a high quantity of sediment as it passes, contributing in increasing considerably the turbidity of the penetrating water in the estuary. Once the channel is filled, the flood overflows and sweeps progressively through the slikke, then the upper slikke if the tidal range is high enough. The velocity of the flood thus decreases very quickly and is only 0.5 m/s on average. The ebb is, as a rule, the subordinate current. Nevertheless, it is locally dominant assuming specific trajectories.

The migration of the channel in time and space constitutes the main morphosedimentary process of Gué de l'Épine – Pontaubault area (cf. Fig. 16 for example of cores collected in the meandering zone). This migration controls the dynamic of the point bars and the development of the banks associated to the channel, determining the distribution of the areas of sedimentation and erosion.

The inner estuary is characterized by a **specific sediment** called "**tangue"** (regional name), grey in colour, and generally described as a sandy to silty mud. Its mean grain size ranges from 0.03 to 0.09 mm. The tangue is mixed silico-bioclastic sediment, containing about 50 % of biogenic carbonate represented by a fine-grained mixture of mollusca, foraminifera, ostracods, coccoliths and bryozoair fragments. The mineral fraction consists mainly of quartz, mica and heavy minerals. Due to its physico-chemical properties, linked to its composition, grain-size and texture, the tangue is a sediment that can be drained easily and compacts quickly. It is also thixotropic, making its reworking by tidal currents easy. It is, therefore a mobile sediment, favourable to the formation and preservation of numerous figures and depositional structures. In cross section, the tangue displays a bedded appearance, made up of alternating layers, few millimetres to few centimetres thick, of sandy silt or muddy silt. Flaser-, wavy- and lenticular-bedding, as well as planar bedding of low energy, are the most represented beddings (Plate 3, photos 1, 2, 3). Climbing ripple bedding can also be frequent in short cut or levee position. Due to the tidal bore passage, freshly deposited tangue succession along tidal channel, can be deeply convoluted (Tessier & Terwindt, 1994) (Plate 3, photos 4, 5, 6).



As wave activity is almost negligible, and thanks to the tangue occurrence and properties, the inner estuarine domain is the most favourable area in the MSMB to observe and analyse **tidal facies** and more specifically, **tidal rhythmites**.

In the lower intertidal area (lower slikke), in cross section, facies are quite homogenous represented by fine grained sand with occasional ripple cross bedding and upper flow planar bedding. Generally no rhythmicity can be observed at these low topographic levels on the edge of the active channel (Fig. 15, and Fig. 16 for example of cores). The energy is too high, inducing intense erosion processes, which prevent the record of the cyclic character of tidal dynamics.
Towards the upper intertidal (high slikke) and supratidal areas, the tangue facies become finer and the preservation of mud drapes increases. Flaser-, wavy - and lenticular beddings appear progressively. A high variety of beddings is well-preserved and tidal rhythmite facies are observable.

In cross-section, from upper intertidal to supratidal facies, **two main type of tidal rhythmites (TR)** are distinguished (Plate 4 and 5):

- Semi-lunar TR that record the neap-spring-neap tidal cycle of 14 days (Tessier, 1993). They are the best-developed TR and are essentially preserved in the upper intertidal deposits. Different types of beddings form semi-lunar TR (Plate 4). The most frequent develop in planar bedding made up of sand- or mud-dominated, millimetric to centimetric doublets. In ripple bedding, neap/spring cycle is characterised by a variation in the thickness of the successive doublets, but also by an evolution in the type of bedding materialising the energy evolution during the semi-lunar cycle. Finally, semi-lunar TR are expressed quite frequently in climbing ripple bedding (Lanier & Tessier, 1998) in the ebb-dominated short-cut channels above point bars, or in flood levee facies (Plate 5, photos 1, 2, 3, 4).

- Annual TR that record the highest equinoctial tides of the year. They are preserved exclusively in supratidal (salt marsh) facies and consist in a few cm thick sequences made up of very thin (mm) tidal couplets (Plate 5, photos 5, 6, 7). Annual sequences are formed by a succession of undisturbed sand-dominated couplets and mud-dominated couplets heavily disturbed by roots. The later materialize the very low energy summer sedimentation of the marshes when grass develops, the sand dominated episode being related to the higher energy winter/spring sedimentation.

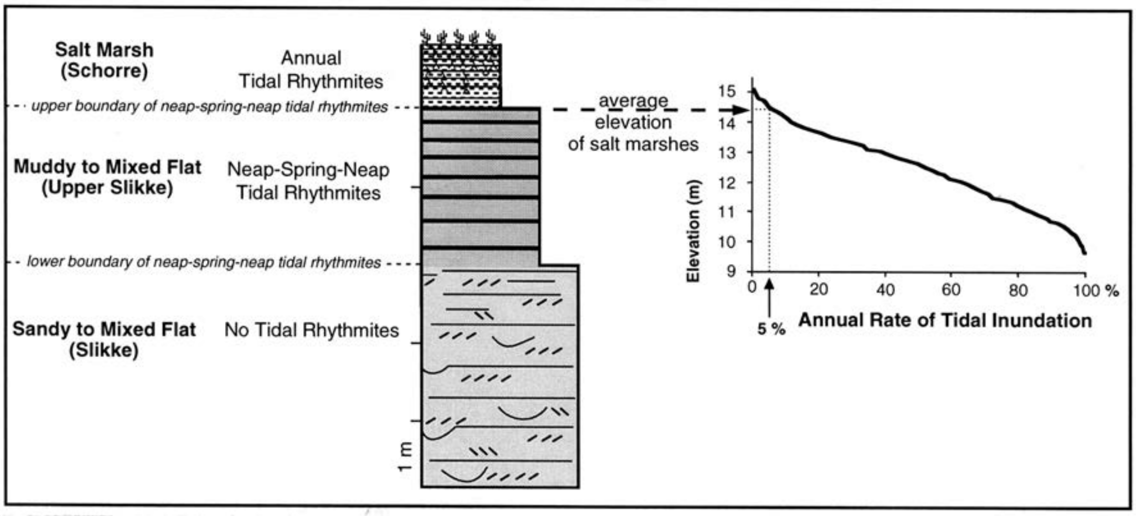


Figure 15. Ideal facies succession in the internal estuary based on tidal rhythmite (TR) occurrence and evolution from lower intertidal (no TR), to upper tintertidal (semi-lunar TR) and supratidal (annual TR) deposits (after Tessier, 1998)

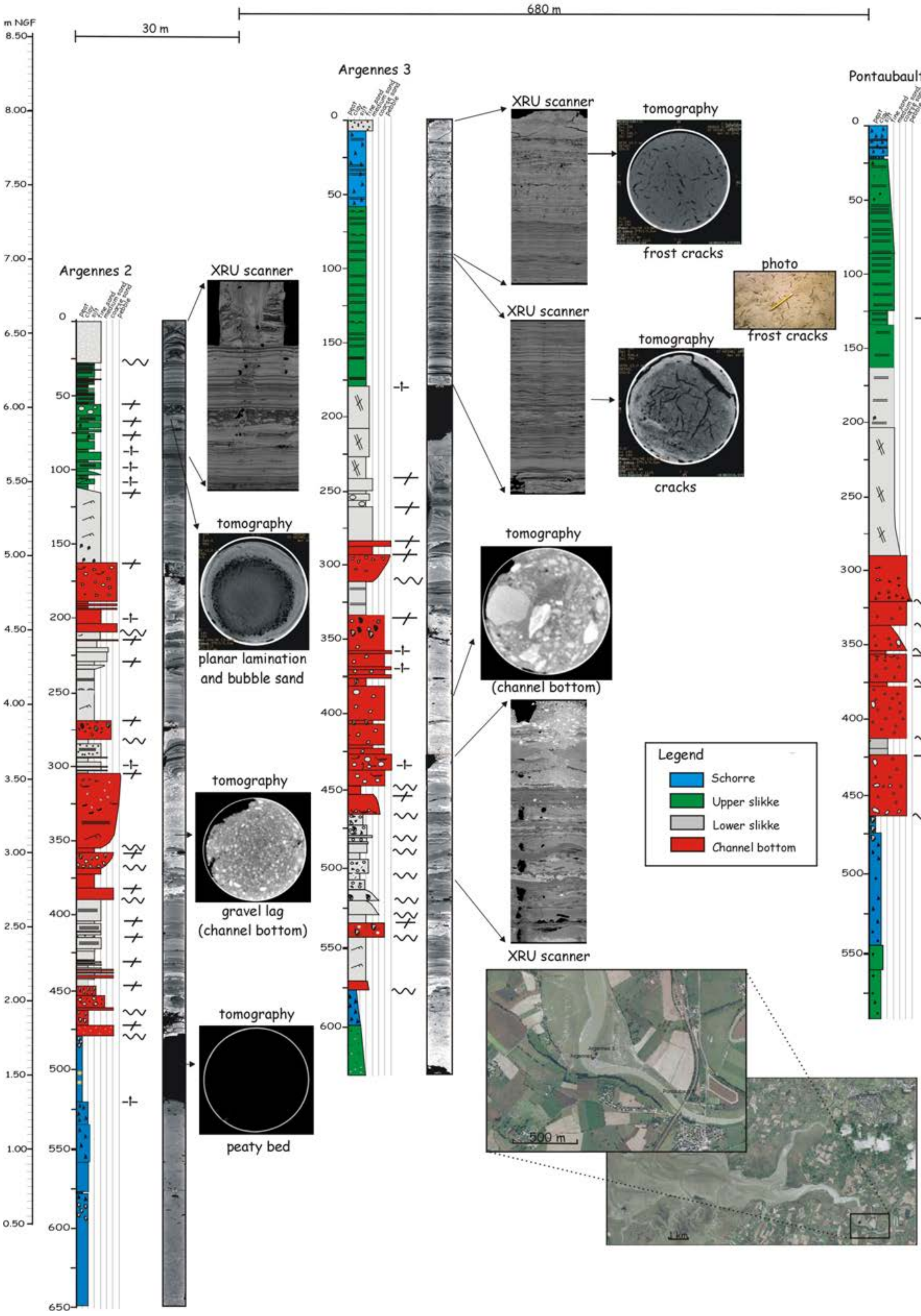
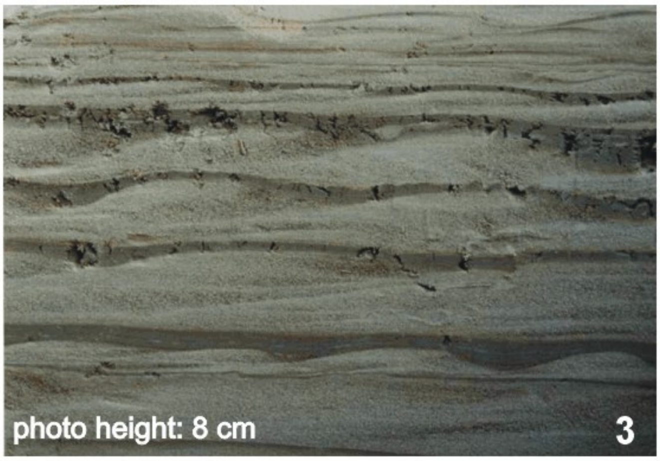
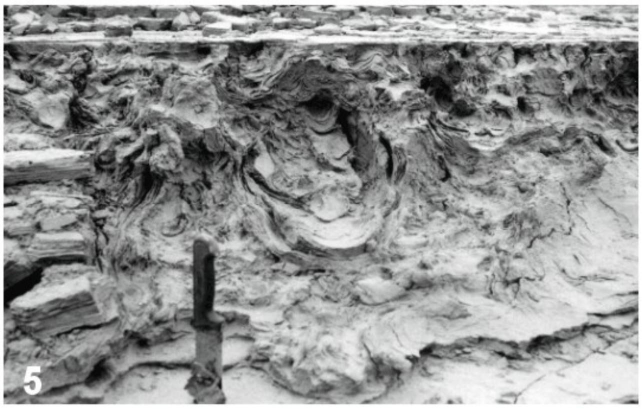


Figure 16. Vibrocores collected at Argennes-Pontaubault locality, in the lower to mid intertidal area of the meandering zone of the fluvio-tidal transitional channel that extends from Gué de l'Epine (downstream) to Pontaubault (upstream). Facies successions made of gravels and coarse sands and then sandy to silty tidal bedding are interpreted as the result of the channel migration and infilling.
In Argennes 1 core, this succession lays abruptly on a well developed peat layer
(I. Billeaud, 2007, PhD, Univ. Caen; Billeaud et al., 2007)

Plate 3 – Tidal facies in the inner estuary
Tidal beddings and tidal bore–induced deformation

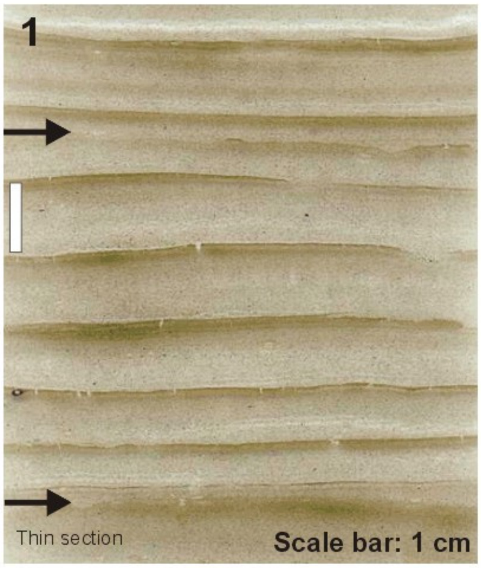


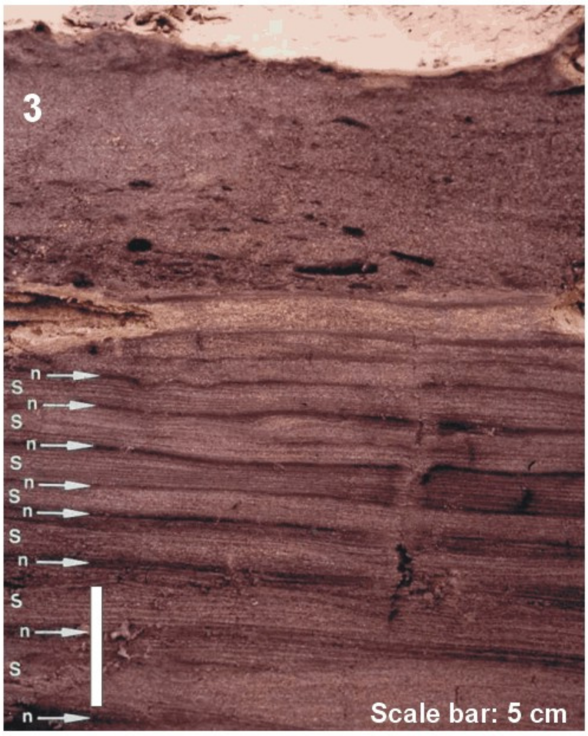
The typical laminated aspect of the tangue in cross-sections. The tangue is constituted by the superposition of sand/mud couplets of varying thickness and shape.

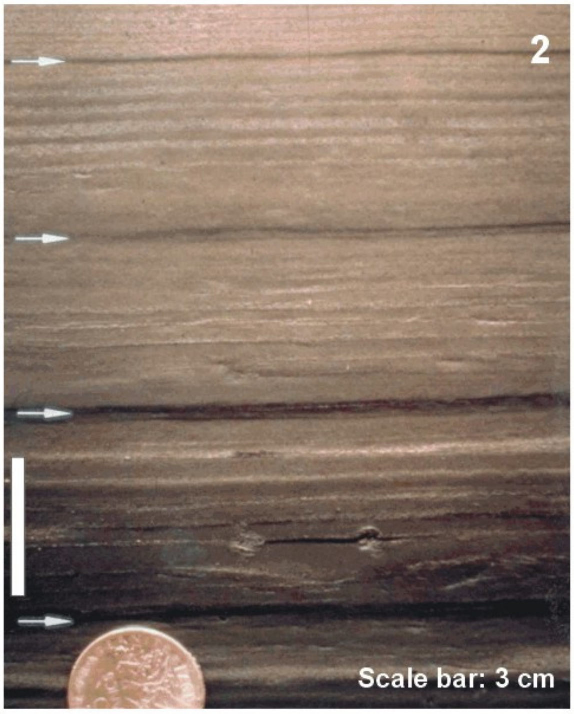


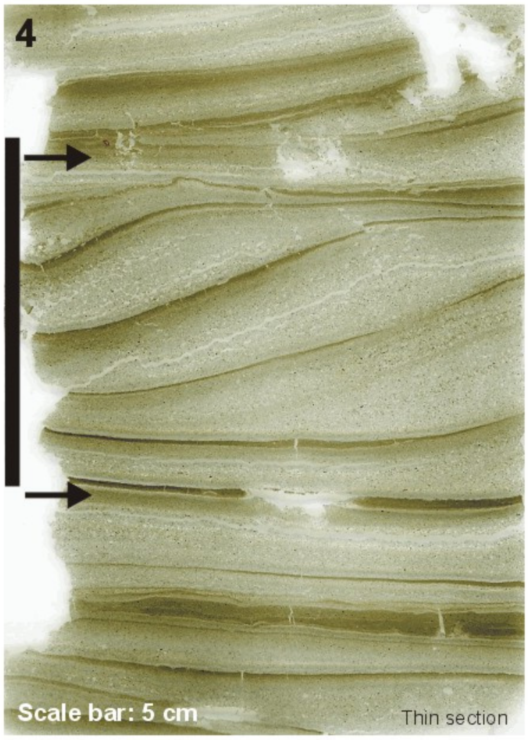
Syn-sedimentary deformations in tangue deposits due to the tidal bore passage over the flat along the main estuarine channel (Grouin du Sud). (Tessier & Terwindt, 1994).

Plate 4 – Tidal facies in the inner estuary spring – neap Tidal rhythmites



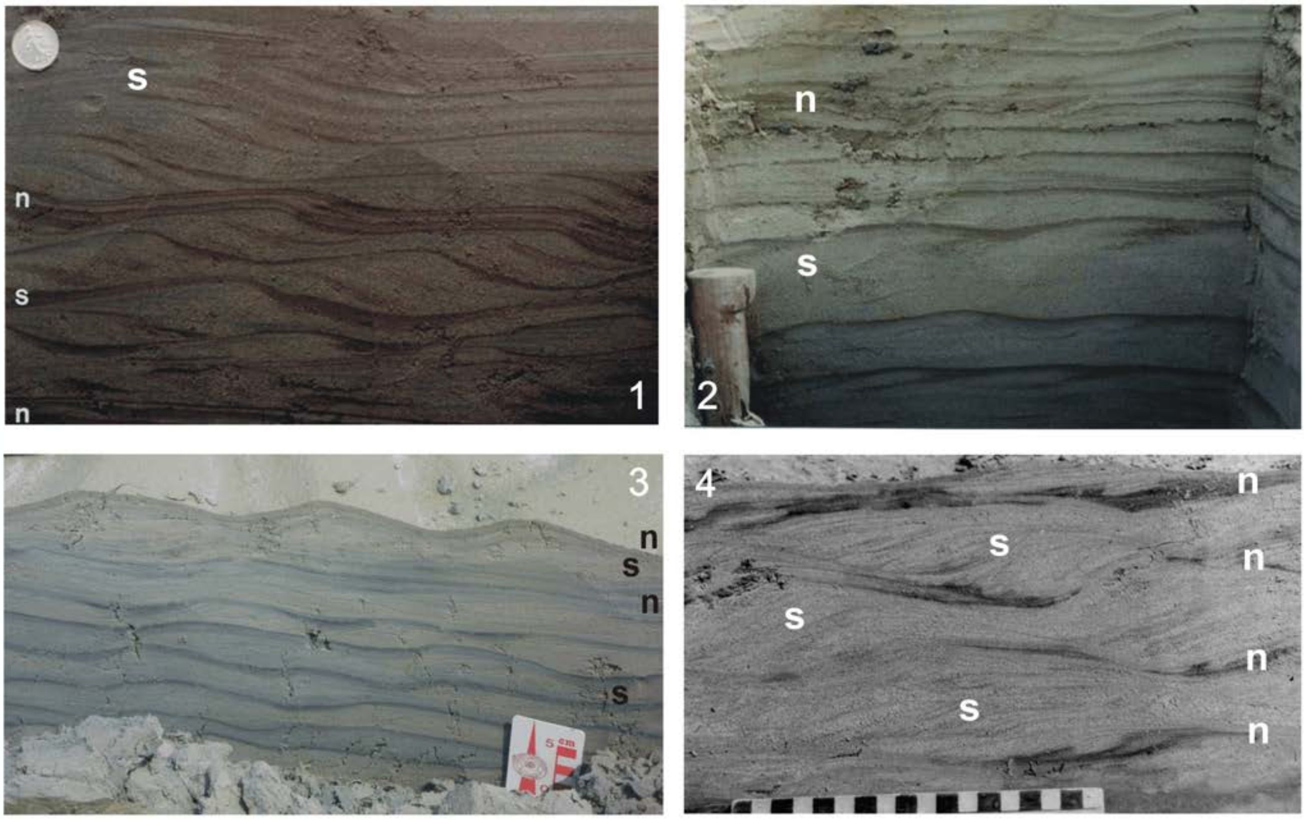




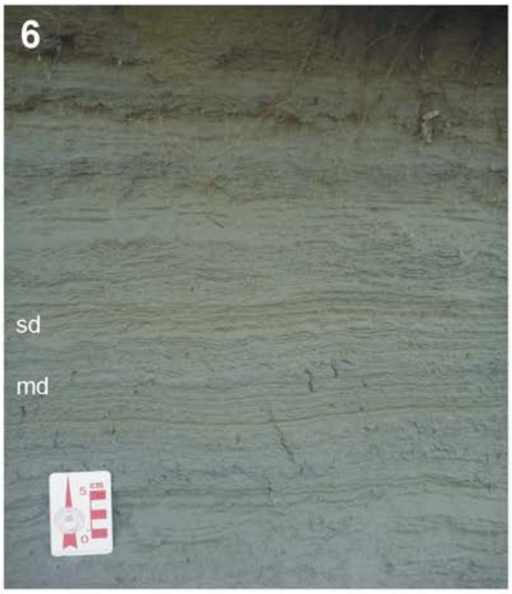


Examples of semi-lunar tidal rhythmites recorded in upper intertidal deposits of the inner estuary (arrows indicate neap tide stages (n) that correspond to a 5-6 days period of emersion. Spring tides (s) only are recorded). These examples illustrate TR recorded in planar-, wavy- and flaser-beddings (P, WB, FB) (in Tessier, 1993).

Plate 5 – Tidal facies in the inner estuary
spring – neap and annual Tidal rhythmites



Examples of semi-lunar tidal rhythmites (TR) recorded in upper intertidal deposits of the inner estuary
These examples illustrate TR recorded in Climbing-ripple beddings (n: neap tide stages; s: spring tides).
(1, 2: Flood-dominated facies of flood levees; 3, 4: ebb-dominated facies in short cut channels)
(After Lanier & Tessier, 1998)



Annual cycles recorded in salt marsh deposits of the inner estuary. Sand-dominated couplet succession (sd, light color - winter equinoctial tides) alternate with highly rooted mud-dominated succession (md, dark color - low energy summer deposition) (in Tessier, 1998).

Day 3

The northeastern wave-dominated coastline

Objectives

In addition to take advantage of spring tides to observe the tidal bore at the entrance of the inner estuary, and to examine at low tide bio-construction due to marine annelid worms, the aim is to focus on: - The wave-dominated NE coastline dynamics and recent evolution (sandspits and back barrier facies, beach accretion / retrogradation); - The holocene evolution and infill (seismic and core data, stratigraphic reconstruction); - The impact of Holocene climate changes on the infill dynamics

Successive stops

Stop 7	08h30 – 09h30	Grouin du Sud: Tidal bore passage
Stop 8	10h00 – 12h00	Dragey: wave-dominated sandy barrier and back barrier system.
Stop 9	12h30 – 14h30	Saint-Jean-Le-Thomas: Holocene evolution
Stop 10	15h00 – 17h00	The Champeaux Hermelle reef (polychaete bioherm)

Stop 7 – The Grouin du Sud

Tidal bore and panoramic view on the transition between the outer and inner estuary

The « Grouin du Sud » offers one of the most beautiful panoramic views on the MSMB, especially onto the estuarine domain. The narrowing of the estuary from this point leads to an acceleration of tidal current velocities into the channels of the Sée and Sélune rivers. During spring tides nice tidal bores commonly develop into these channels. Hence, the site is as well an appropriate place to observe this spectacular tidal process. A research study on the tidal bores in the MSMB is in course since January 2011 (Plate 6).

Photo in: http://www.projetmontsaintmichel.fr

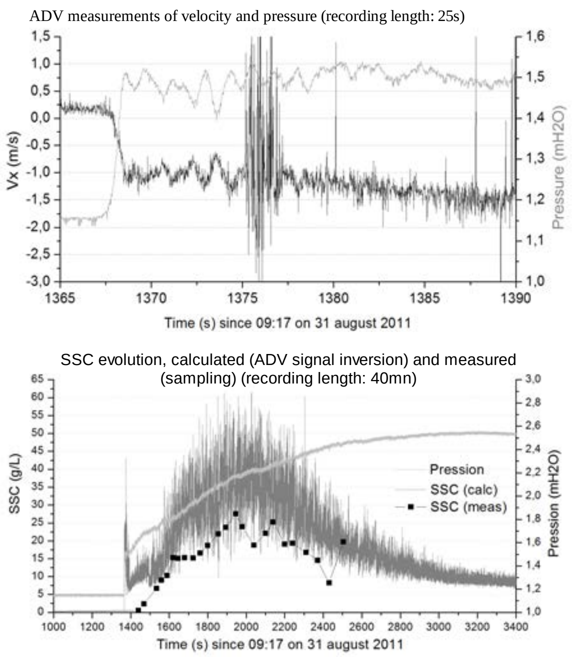
The tidal bore at the Grouin du Sud (October 2010, photo B. Tessier)

Plate 6
Study of tidal bore dynamics in the MSMB estuary

The tidal bore is a common phenomenon observed in about 80 estuaries around the world. All are hypersynchronous estuaries with low slope and tidal range exceeding 6 m. In spite of this relative commonness, the tidal bore phenomenon is still poorly studied and understood. However, since a few years, several projects aiming in studying the hydrosedimentary processes induced by tidal bores emerge (i.e. Fan et al., 2012 – Qiantang River; Bonneton et al., 2011 – Garonne River; Chanson et al., 2011 – Garonne River; Mouazé et al., 2010 – Sélune River).

In the MSMB a study is performed since January 2011 in the framework of the ANR project “Mascaret” (http://mascaret.enscbp.fr/) involving the M2C Lab (CNRS/university of Caen). Part of the study is devoted to in situ measurements* in order to better understand the processes involved in the formation of a tidal bore, and its impacts on sediment transport (Lucille Furgerot’s PhD work, M2C / Univ. Caen). The main measurement site, “Le Bateau”, is located in the inner estuary, along the Sée River, in a linear portion of the channel where tidal bores develop during each high spring tide period. Depending on the channel cross-section and river water level, tidal bores from undular to breaking type (photos) can develop in this segment of the River.

Ten field campaigns have been carried out during high spring tides since January 2011. Various measurements into about 35 tidal bores were performed using ADV and ADCP (current velocities), CTD (temperature, salinity), OBS and direct sampling (suspended sediment concentration - SSC).



Measurements (Velocity and SSC)
during the passage of an ondular tidal bore
(Furgerot et al., 2012a)

Some preliminary conclusions can be drawn (Furgerot et al., 2012a, 2012b, 2012c):
- Tidal bores that propagate in the inner estuary are made of successive fronts: 1) a tidal wave front (the bore itself), 2) a turbidity front, arriving about ten seconds after the bore, and 3) a marine front recorded 3 to 16 mn after the bore.
- OBS measurements and direct samplings, as well as ADV signal inversion calculation show that SSC can reach up to 40 g/l in the middle of the water column.

These first results already demonstrate the major role of tidal bores on estuarine sediment transport. In complement with field measurements, experimental studies in flume will be performed at M2C Lab in order to better constrain the tidal bore fluid dynamics.

The M2C team wishes to warmly thank Jean-Yves Cocaign, Director of the Ecomusée of the MSMB, for his welcome at the Museum during field surveys, and for his very useful advices for selecting the most suitable site of measurements.

Stops 8 and 9 – Dragey and Saint Jean-Le-Thomas beaches

Wave-dominated coastline and Holocene infill reconstruction

The northeastern littoral of the MSMB is represented by a sandy coastline that experiences some spectacular and contrasted processes of erosion (coastline retreat) and sedimentation (coastline progradation).

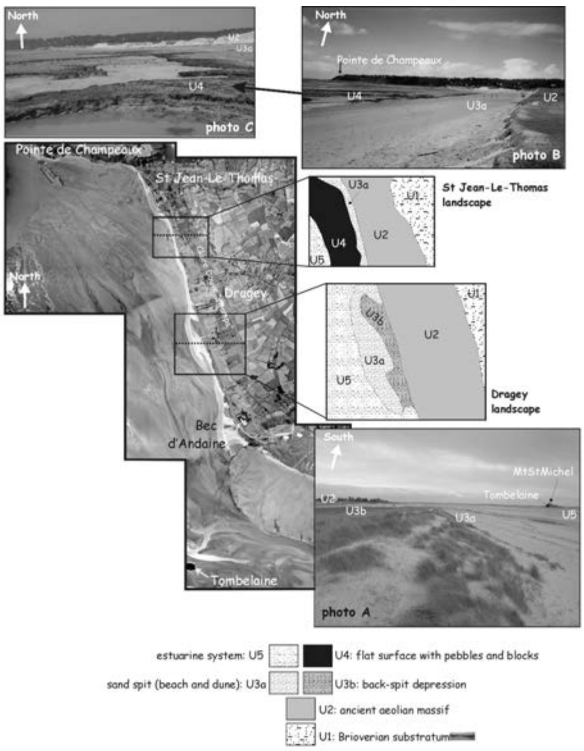
Studies performed by comparing aerial photos since 1947 have demonstrated that the beach at Saint Jean-le-Thomas locality has retreated of about 200 m while, southward, at the Bec d'Andaine or Dragey sites, beaches advanced of about the same value (*Text § 2.2.3, p. 89 and Fig. 2.2.3.3, p. 93*) (Compain et al., 1988).

These evolution and coastline organization are mainly related to wave action and sediment supply distribution. The NE entrance of the MSMB is exposed to prevailing W to NW swells that induce severe erosion at St Jean-le-Thomas. The eroded sand material is transported southward by the littoral drift until Dragey and Bec d'Andaine localities where it is fixed as sand spits.

The sand spits isolate from the open sea energy, sheltered depressions. These back-barrier systems are invaded only by spring tides, and are filled by fine-grained tidal facies evolving progressively to salt marsh deposits.

Northward, at Saint Jean-le-Thomas, the high energy swell action induces a severe beach and dune (barrier) retreat. The process produces a wave ravinement surface eroding older back-barrier sediment successions that deposited when the barrier was located more offshore than the present-day one. This barrier began to construct some 6500 y. ago, when the rise in sea level slowed down significantly. Since that period, sea level rose slowly and the barrier has retreated until its present-day position.

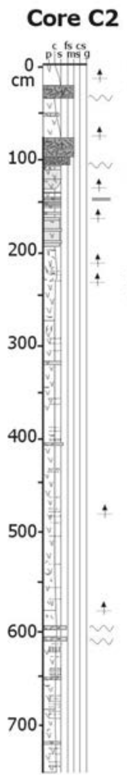
The NE littoral of the MSMB can be considered as a wave-dominated environment. However, its evolution is significantly influenced by the tidal dynamics of the adjacent estuary since the main estuarine channel borders the coastline. When the channel migrates northward and tends to approach the shore, the Bec d'Andaine and Dragey sand spits experience erosion due to tidal ravinement processes.

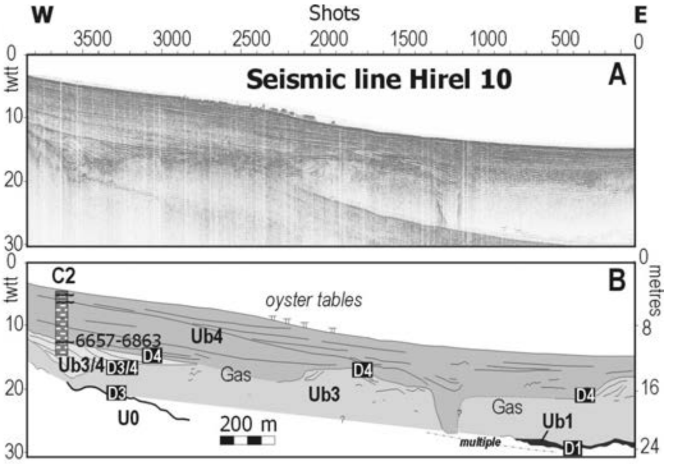


These different aspects of the NE littoral evolution and functioning are summarized on Fig. 17. Stops 8 and 9 offer also the opportunity to discuss about the Holocene sedimentary infill of the MSMB. In the framework of Isabelle Billeaud's PhD work (I. BIlleaud, 2007, Univ. Caen), some 50 vibrocores and 600 km of VHR seismic profiles were acquired, that allowed reconstructing the Holocene infill architecture (Billeaud et al., 2007; Tessier et al., 2010) (Plate 7), as well as demonstrating that Holocene rapid climate changes have significantly impacted the dynamic of infill of this megatidal environment (Billeaud et al., 2009) (Plate 8).

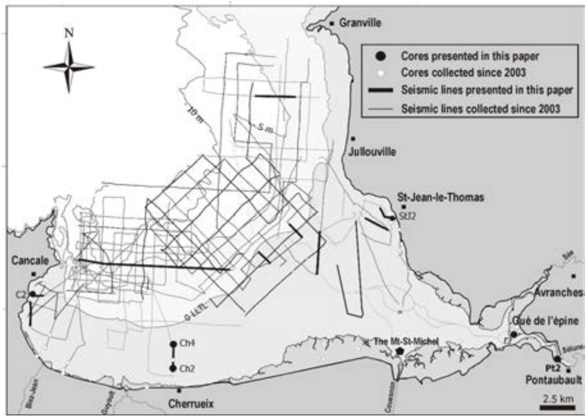
Figure 17. The different landscapes and morphosedimentary units of the NE shoreline of the Bay of Mont-Saint Michel (Aerial Photo © IGN 1999) (in Tessier et al., 2006).

Plate 7
Reconstruction of the Holocene sedimentary infill
(all figures in Tessier et al., 2010 and after Billeaud, 2007)

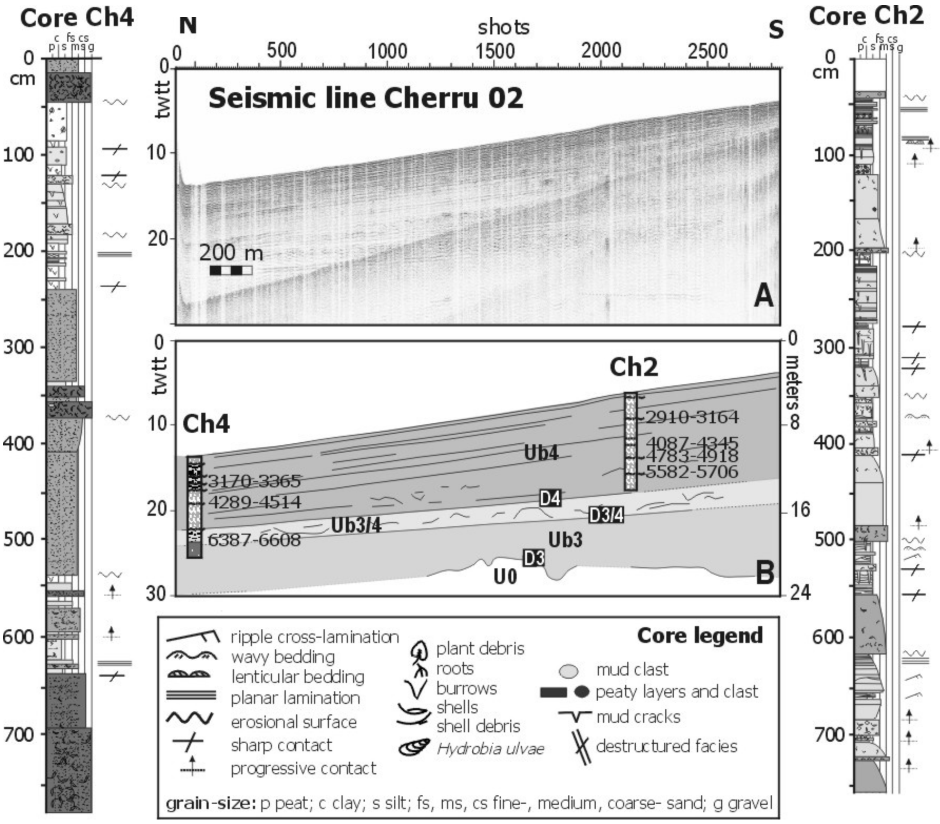




Examples of seismic lines and cores acquired in the embayment, on the mudflat of Cancale (above – Hirel 10) and on the sandflat of Cherrueix (below – Cherru 2). Seismic data in the illustrate that the HST (Ub4) is an aggradational depositional unit, with locally internal erosional surfaces (Hirel 10). In both sectors, the TST (Ub3) is represented by bioclastic sands and locally migrating bedforms such as dunes or banks. Cores and 14C dating demonstrate that the passage from the TST to the HST occurred around 6500 y. BP.



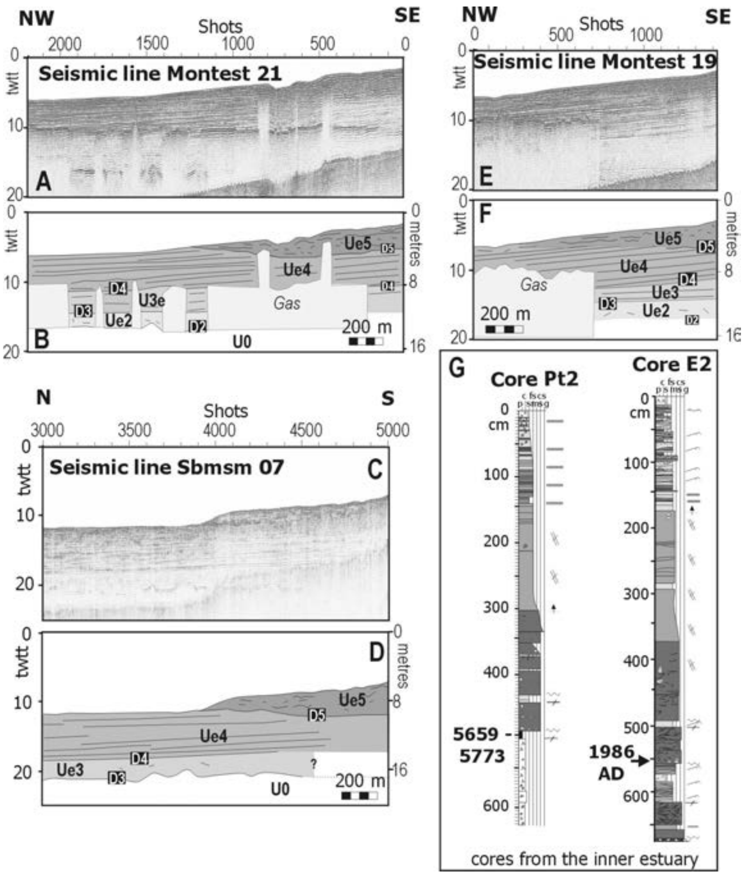
Location of seismic lines (~ 600 km) and vibrocores (~50, 7 m long) collected in the MSMB



Above the bedrock, two depositional units are identified:

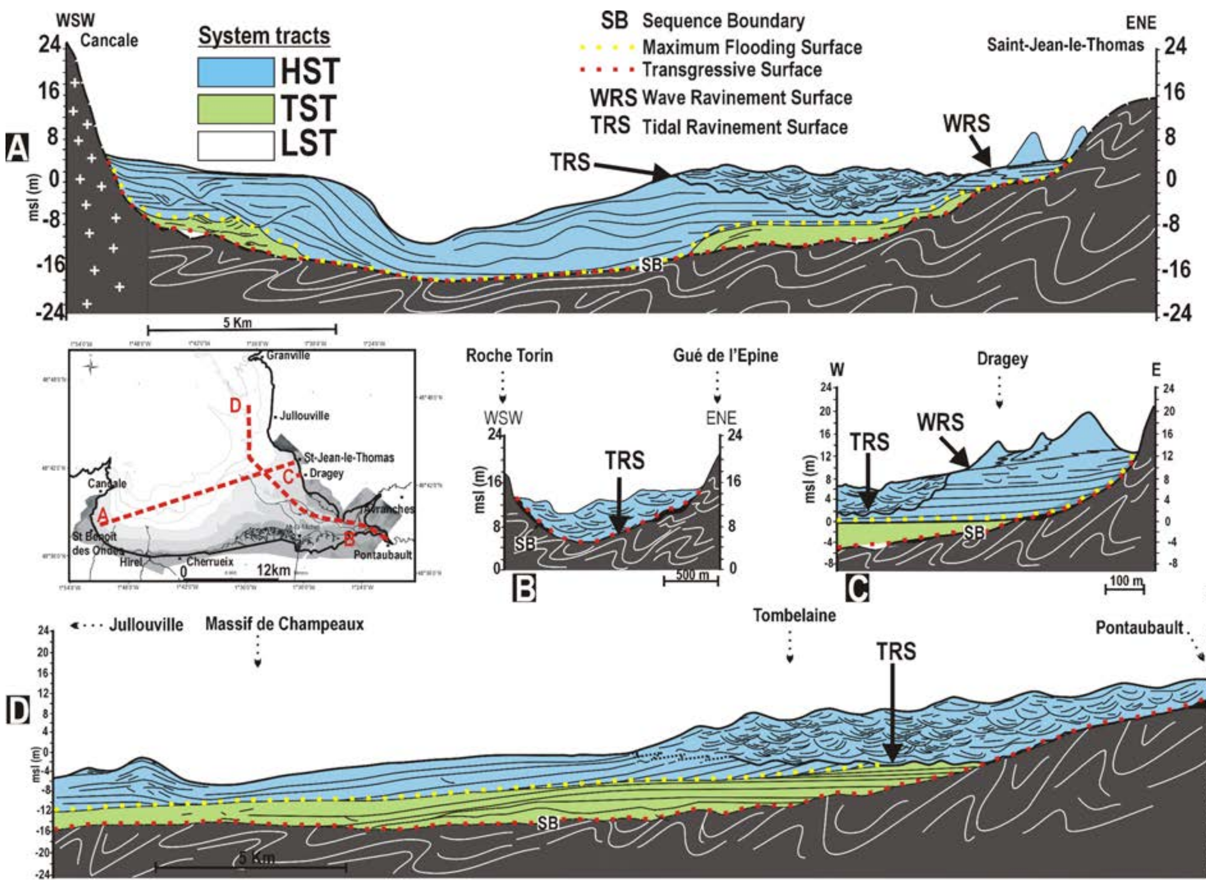
The TST is made of aggradational fine-grained facies in the estuarine domain, while it is composed of high energy coarse-grained deposits into the embayment.

The HST into the estuary is represented by a sand-dominated tidal channel-and-shoal belt, whereas into the embayment, it is an aggradational unit, consistent with the present-day tidal flat landscape.



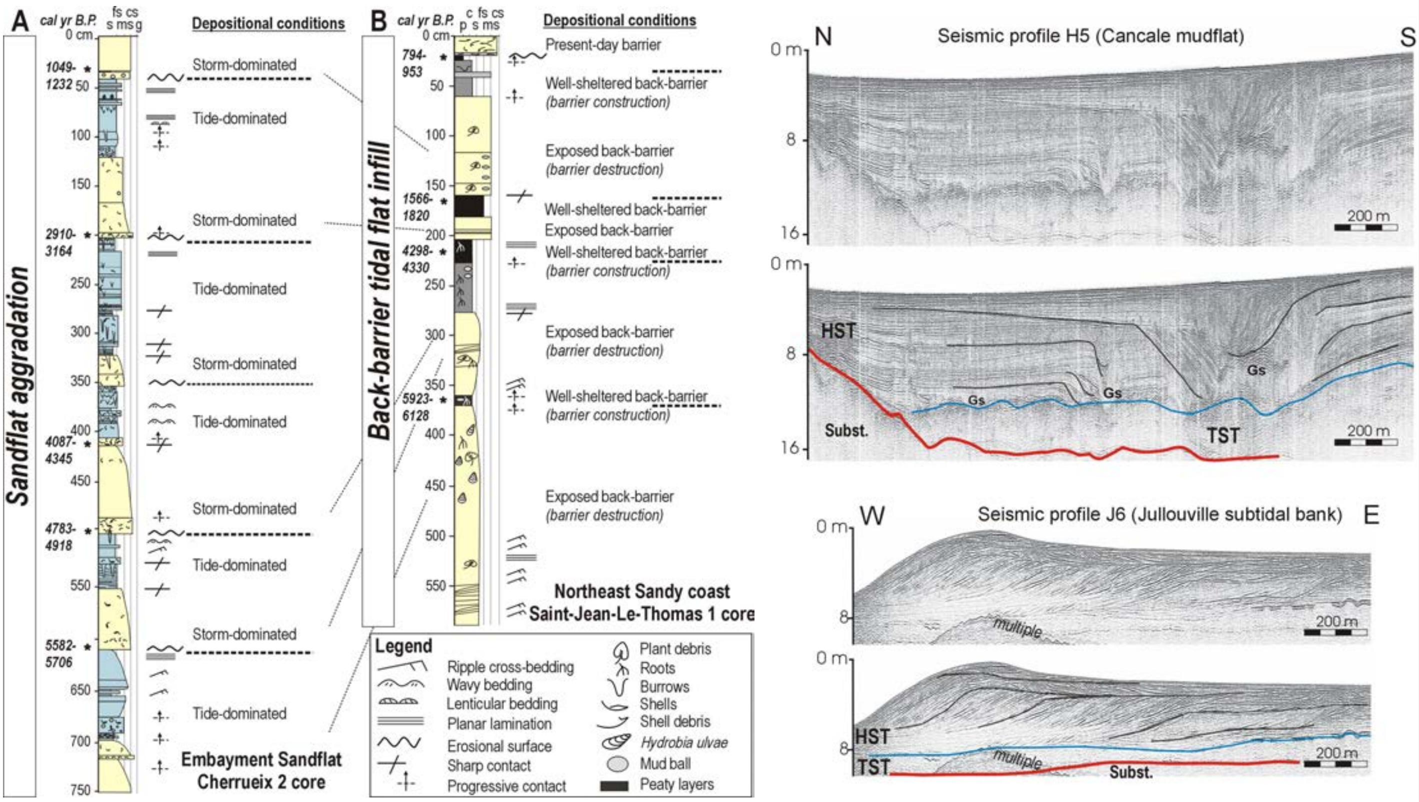
Examples of seismic lines and cores collected into the estuary. The HST is composed of 2 units: an upper unit (Ue5) containing multiple erosional surfaces (estuarine channels), and that progressively pinches up seaward. It rests through an erosional surface (tidal ravinement) on a seaward progradational unit (Ue4) interpreted as the bottomset termination of the estuarine body. The TST is made of an aggradational unit composed of fine-grained deposits that produce biogenic gas.

Synthetic cross-sections illustrating the main geometrical patterns of the MSMB sedimentary infilling.

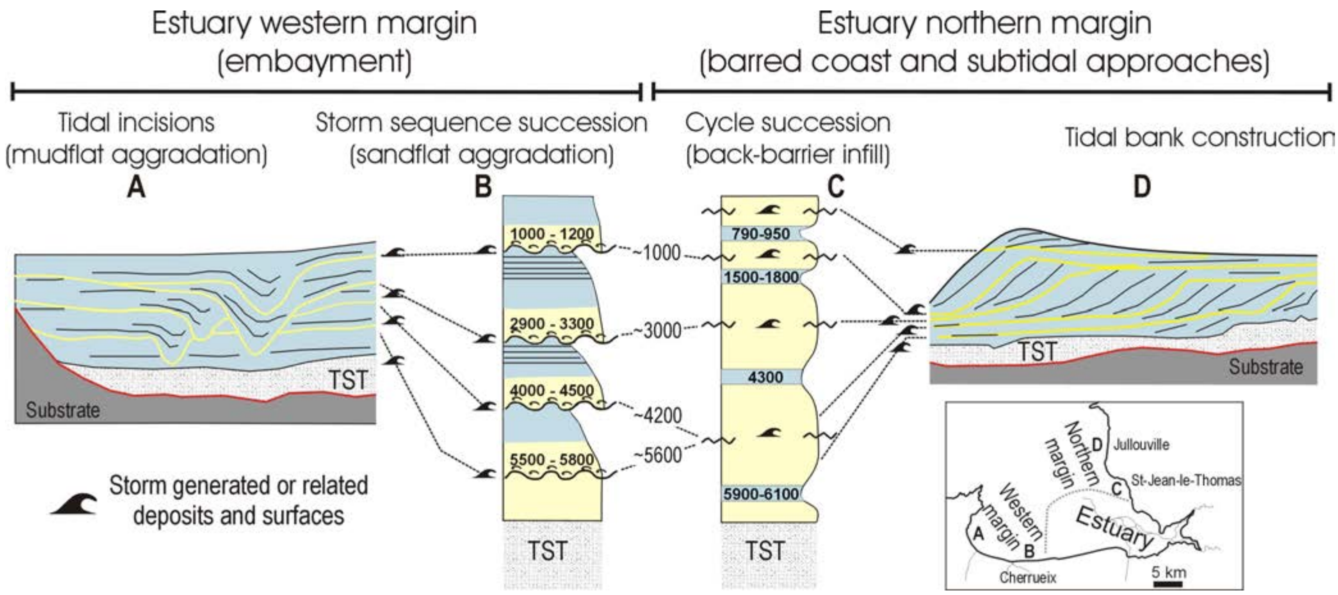


The HST represents the main unit of the infilling of the MSMB. It is composed of aggrading tidal flats in the embayment, and of aggrading / wandering channel-and-shoal body in the estuary. This body, fed exclusively by marine source, translates progressively seaward, as the estuary fills. Because of the shallowness of the bedrock compared to the deepness of the tidal ravinement, the HST occupies most of the initial accommodation into the estuary.

Plate 8
Holocene rapid climate change impacts
(all figures after Billeaud et al., 2009)



Cores and seismic lines acquired in the MSMB, illustrating the different types of records of Holocene climate changes into the HST: storm-dominated facies in the sandflat successions into the embayment; barrier destruction phases along the NE shoreline; tidal incisions into the mudflat of the embayment induced by barrier destruction; storm-generated erosional flat surfaces into subtidal banks.



The records of periodic environmental changes are preserved into the HST deposits of the MSMB sedimentary infill. 14C dating demonstrates these changes can be correlated at the scale of the bay. They are attributed, in all cases, to storm impacts, which occur with a millennial time-scale periodicity. Four to five significant changes occurred. The time of these climatic crises was ~ 5500-5800, 4000-4500, ~ 3000, 1000-1200 yr B.P. and matches the time of the Bond's events.

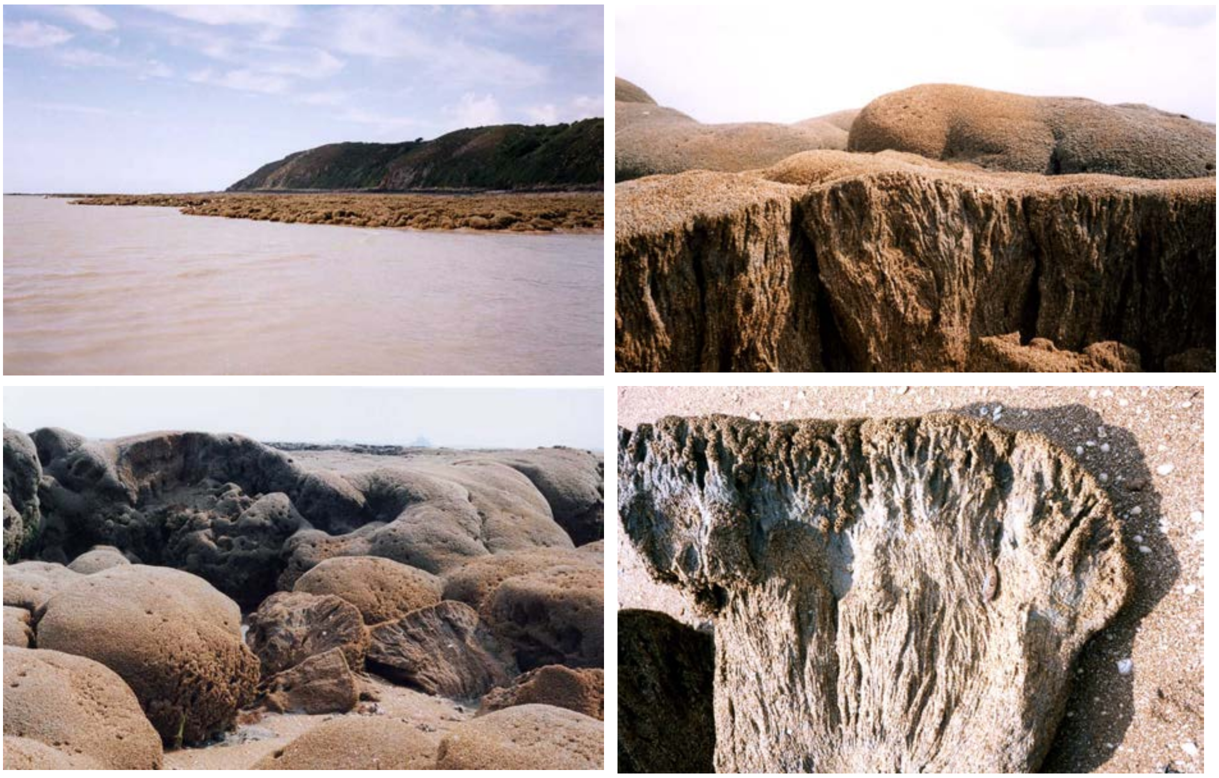
Stop 10 – Champeaux rocky flat

The “Hermelles” (polychaete worms) reef

At two places in the Bay of Mont-Saint-Michel, bioconstructions made by polychaete worms develop on the margins of the estuarine system in the lower intertidal area. The best developed and known, called the “Banc des Hermelles” is located about 4 km off the Chapelle Sainte Anne at the transition between the estuarine domain and the embayment (*Text 2.1.3, p.61*). The second reef is implanted on the rocky flat at the foot of **Champeaux cliff** (cf. photos below and Fig. 18), in the north of the Bay, at the junction between the estuarine entrance and the open marine domain. This reef is easily accessible at low tide and, although of less extension than the “Banc des Hermelles”, displays very spectacular shapes and development.

The “Hermelles”, or ***Sabellaria alveolata***, are gregarious and sedentary **annelid polychaete** that construct massive reefs made of contiguous arenaceous tubes 5 to 10 mm in diameter. The reefs constitute arborescent structures reaching up to 1.2 m above the floor. Densities are very high (15,000 to 60,000 individuals/m^2).

The settlement of the juvenile polychaete requires a hard or sufficiently stable substrate. In the case of Champeaux, the bioherm naturally settles on the outcropping rocky substrate as well as on the walls of ancient permanent fishing grounds (Fig. 18). The nature of the substrate supporting the so-called “Banc des Hermelles” is not clearly defined. It is probably constituted by natural oyster beds.



The “Hermelles” reef of Champeaux

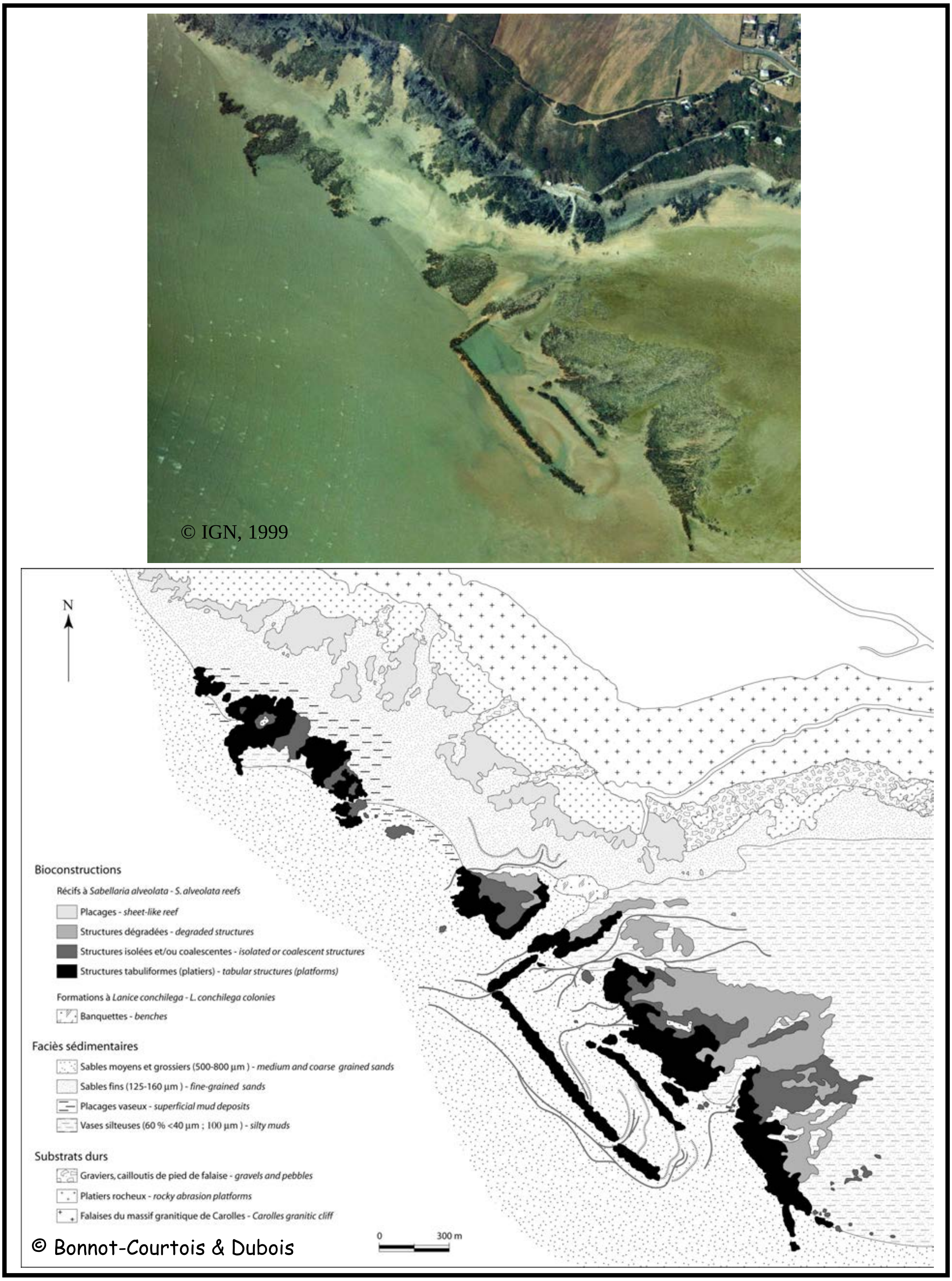


Figure 18. Distribution of the "Hermelles" Reef nature
and of associated sedimentary facies at Champeaux locality
(C. Bonnot-Courtois & S. Dubois)

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Ce document a été imprimé par
L’Université Lille 1 – Service de Reprographie
59 655 – Villeneuve d’Ascq Cedex

ISSN 0990-3925
ISBN 2-907205-71-4

Dépôt légal : 3ème trimestre 2012

8th International Conference on Tidal Environments

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