Pockmarks on the South Aquitaine Margin continental slope: the 2 seabed expression of past fluid circulation and bottom currents

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Abstract

Inactive pockmarks of 100-200 m in diameter recently discovered on the South Aquitaine Margin continental slope are present between 350 m and 1150 m water depth in both interfluve and sediment wave areas. Water column and seafloor backscatter and sub-bottom profiler data do not exhibit present-day or past gas evidence, e.g. massive and continuous gas releases at the seabed and fossil methane-derived authigenic carbonates. To discriminate hypothetical pockmark populations, e.g. due to the nature of the involved fluids (gas, water), a detailed study of pockmark morphological characteristics based on a Geographical Information System was performed. Pockmarks are divided in two groups from sub-circular
to elongated shapes. It is proposed that pockmark elongation may mainly result from the dynamics of near bottom currents, inducing upwelling, having thus reshaped the initially-formed subcircular pockmarks.

**Keywords**

Pockmark, fluid, seabed morphology, Aquitaine slope, GIS, currents

**1. Introduction**

Pockmarks were first described by King and MacLean (1970) as seafloor morphological depressions, formed by fluid escapes. Pockmarks are commonly encountered and are worldwide related to fluid migrating upward (Hovland and Judd, 1988) and triggering-sediment resuspension during leakage and sediment collapse. These depressions are observed from shallow environments (Garcia-Gil, 2003; Josenhans et al., 1978; Rise et al., 2015) to deep bathyal environments (Gay et al., 2006; Pilcher and Argent, 2007). Pockmark morphologies can be associated with several processes and various types of fluids, e.g. small scale pockmarks can be related to a unique local gas source (Gay et al., 2007; Hovland and Judd, 1988), to dewatering of the sediments upon compaction (Harrington, 1985) and to freshwater seeps (Whiticar, 2002) while pluri-kilometre-scale pockmarks may indicate hydrate dissolution (Sultan et al., 2010). Pockmarks may occur as clusters (Hovland et al., 2010) or as strings of pockmarks (Andrews et al., 2010; Pilcher and Argent, 2007). Strings of pockmarks are commonly related to geological features focusing fluid flows, e.g. fractures and faults (Dimitrov and Woodside, 2003; Gay et al., 2007) and buried valleys (Baltzer et al., 2014; Gay et al., 2003; Gillet et al., 2008).

The modification of original pockmark morphologies will depend on internal factors such as successive fluid expulsion events (Hovland and Judd, 1988), the presence of methane-
derived authigenic carbonates (Andresen et al., 2008; Gay et al., 2006) and external factors such as bottom currents (Dandapath et al., 2010; Gafeira et al., 2012; Josenhans et al., 1978; Schattner et al., 2016); slumping and sedimentary destabilization along the slope direction (Brothers et al., 2014; Dandapath et al., 2010), presence of benthic fauna and debris accumulation (Webb et al., 2009), e.g. coarser sediments (Pau and Hammer, 2013). Bottom currents may contribute to elongate pockmarks along the direction of the currents by eroding sediments and preventing sedimentation over the pockmarks (Andresen et al., 2008; Dandapath et al., 2010). Bottom currents may induce upwelling within the pockmarks that would limit the sedimentation of fine-grained sediments, therefore maintaining the pockmark morphology (Brothers et al., 2011; Pau et al., 2014). Moreover, coalescent pockmarks (merging depressions) (Çiçı et al., 2003; Dupré et al., 2010; Gay et al., 2006) may be a result of successive fluid escapes or external processes as cited above and eventually form elongated ones. Pockmark morphological characteristics, accessible through their acoustic signature, may be used to determine potential activity (Dupré et al., 2010; Hovland et al., 2010), the nature of fluids involved (Gay et al., 2006; Hovland and Judd, 1988) and to address the relative timing of pockmark formation with regards to surrounding sedimentation (Bayon et al., 2009).

The present study mainly focuses on the geophysical characterization of a wide pockmark field discovered on the continental slope of the Aquitaine Margin (offshore France) in 2013. Semi-automated methods to map pockmarks, never applied in slope and heterogeneous seafloor morphology areas (Lecours et al., 2016), have been tested. In this work, the pockmark activity and the nature of the fluids involved in the pockmark formation are discussed. Particular attention is paid to the pockmark reshaping related to external factors such as bottom currents.
2. The setting

Related to the opening of the North Atlantic Ocean, the Bay of Biscay initially corresponded to a V-shaped rift, initiated during Late Jurassic and aborted in mid-Upper Cretaceous (Biteau et al., 2006; Roca et al., 2011). Its extensional phase was stopped during the Santonian age by the opening of the South Atlantic Ocean. The subsequent northward drift of the Iberian plate (Roca et al., 2011) and the strongly related compression phase led to Pyrenean orogeny (Ferrer et al., 2008; Roca et al., 2011). The Bay of Biscay is surrounded by different shelves, the large Armorican Shelf, the Aquitaine Shelf, the Basque Shelf and the Iberian Shelf (Fig. 1) with two main morphological highs, the Landes Plateau and Le Danois High. The hydrocarbon Parentis Basin, created during the Pyrenean Orogeny (Biteau et al., 2006), extends from the onshore to the offshore domain, in the south part of the Aquitaine Shelf (Fig. 1).

FIG 1

The study area is located in the French EEZ (Exclusive Economic Zone) on the continental slope of the Aquitaine Shelf, from 200 m to 1600 m water depths, 60-80 km westward of the coastline, between the Cap Ferret Canyon (44°40’ N) and the Capbreton Canyon (43°30’ N) (Fig. 1). Although the whole continental slope of the Bay of Biscay is characterised by a large canyon density (Bourillet et al., 2006), the canyons are only restricted to the half northern part of the study area (Fig. 2). There, the inter-canyon areas are kilometre wide in N-S (Fig. 2a) and are affected by slope instabilities within a context of silt dominated sedimentation (Schmidt et al., 2014). In contrast to this northern part, the southern part is characterised by a wide sediment wave field (Fig. 2), with a surficial sandy silt sedimentation, extending from the shelf break to the foot slope (Faugères et al., 2002; Gonthier et al., 2006). Sediment wave morphologies, with wave lengths between 800 m and 1600 m and heights from 20 m to 70 m show crests slightly oriented at an oblique angle of the main slope,
between 010°N and 035°N (Faugères et al., 2002). The influence of bottom currents in the formation processes of sedimentary waves (U3 unit older than 120-150 ky) along the Aquitaine slope has been indicated (Faugères et al., 2002; Gonthier et al., 2006). The sedimentary waves are covered by a thin homogenous layer corresponding to U4 unit, 12-15 metre thick (Gonthier et al., 2006). The surficial sedimentary cover of the Aquitaine Shelf is mainly composed of sand and silty sand (Castaing and Weber, 1982; Cirac et al., 2000).

**FIG 2**

The hydrology regime of the study area appears to be very complex due to the semi-enclosed morphology of the Bay of Biscay and the interaction between different currents of different time scales, meso-tidal currents (Batifoulier et al., 2012; Charria et al., 2013; Le Boyer et al., 2013; Pingree et al., 1986), contour currents (Van Aken, 2000) and some temporary currents related to wind-forced events (Kersalé et al., 2016).

Inactive pockforms and pockmarks have been described on the Landes Plateau (Baudon et al., 2013; Iglesias et al., 2010) and on the Basque Shelf (Gillet et al., 2008), respectively. Recently, Dupré et al. (2014) described an active cold seep system at the edge of the Aquitaine shelf without any pockmarks.

### 3. Data and methods

#### 3.1. Geophysical data acquisition and processing

High-resolution marine geophysical data were acquired during BOBGEO2 cruise in 2010 and more significantly during the GAZCOGNE1 survey in 2013 covering 3200 km² of the seafloor at water depths ranging from 130 m to 1600 m (Fig. 2). During GAZCOGNE1 cruise, multibeam bathymetry, water column and seafloor backscatter and seismic reflection (sub-bottom profiler) data were acquired simultaneously along 2868 km of survey profiles, mostly oriented north-south (Fig. 2) and spaced from 380 m (at 220 m water depth) to 2800 m.
(at 1200 m water depth). Bathymetry, water column and seafloor backscatter were collected onboard the R/V Le Suroît at a speed of ~8 knots with a Kongsberg EM302 ship-borne multibeam echosounder operated at a frequency of 30 kHz with celerity profile calibrated with ©Sippican shots. Raw bathymetric data were processed through Caraïbes software (©IFREMER) with application of bathymetric filters and correction of position, pitch, roll and tide effects. Seafloor backscatter data were processed in Caraïbes software with the generation of a compensation curve to harmonize values along the survey lines. Both bathymetry and seafloor backscatter processed data were mainly exported to mosaic grids of 15x15 (with some backscatter maps at 10x10 m cells). Water column backscatter data were only recorded during the GAZCOGNE1 marine expedition (2013). They were processed in SonarScope software (©IFREMER) and then interpreted in GLOBE/3DViewer (©IFREMER) (Dupré et al., 2015).

The sub-bottom profiles were recorded with the ship-borne sub-bottom profiler ECHOES 3500 ©T7iXblue emitting a linear frequency modulated signal, ranging from 1.8 to 5.3 kHz (central frequency of 3.5 kHz), with a vertical resolution of twenty centimetres and a maximum vertical penetration of 100 m. A 2D sub-bottom profiler insonified a surface at the seafloor defined by Fresnel equation and may record lateral reflexions from close-by 3D features, as well as artefacts. These artefacts may be displayed as diffraction hyperbola (Dupré et al., 2014b) and triplication points so-called "bow ties" (Moss et al., 2012) . Raw data were processed with QC-SUBOP software (©IFREMER) before being exported in SEG-Y and then interpreted in ©Kingdom software (Fig. 3).

The water current data were acquired during the ASPEX2010A mooring survey (Kersalé et al., 2016; Le Boyer et al., 2013) with an Acoustic Doppler Current Profiler (ADCP) operated at a frequency of 75 kHz and recording every 2 minutes. The data discussed in this paper come from mooring 10 located at 44°00.069' N - 02°08.644' W at 450 m water
depth in the sediment wave field (Figs. 2a and c). Water current data were recorded over more than 6 months (18th July 2009 - 30th January 2010). Current velocities were integrated between 17 m and 33 m above the seafloor and averaged every 20 minutes. Classical harmonic tide analyses were conducted on ASPEX current data to extract tide-related signals from the raw signal (Lazure et al., 2009).

3.2. Morphometric methods

3.2.1. Morphology detection

Three methods to map pockmark have been tested, two semi-automated and one manual to check the validity of the two previous. The Fill method (Gafeira et al., 2012) involves pockmark extraction based on a succession of Geographical Information System (GIS) operations focused on the numerical filling of depression and then subtraction of filled bathymetry. The second method called the BPI (Bathymetric Position Index, Wright et al., 2012) is based on the calculation of differential bathymetry cells side by side and seafloor roughness analysis. Both semi-automated methods map a large number of depressions which are not pockmarks. Around 500 times more features than manually mapped pockmarks have been detected with the Fill method and 300 times more with the BPI method. For both semi-automated methods, the detected features have been filtered with correction based on the pockmark surface and the surface/perimeter ratio (Gafeira et al., 2012). The number of remaining features is 10 and 20 times higher than the number of manually mapped pockmarks, with Fill and BPI methods respectively. Therefore, in order to minimize the biases observed with semi-automatic methods, all pockmarks have been manually delimited, identified by their rim on the slope grid processed at 15 m (calculated with Slope function in Spatial Analyst toolbox from Arcmap 10.2, ©ESRI). It is worth noting that below the bathymetry resolution (15 m), detection cannot be performed effectively. In other words, small pockmarks of diameter <30 m, if present, could not have been mapped.
3.2.2. Calculation and extraction of morphological attributes

Morphometrics of pockmarks is, in most studies, based on their diameter, internal depth, surface, slope angle (Andrews et al., 2010; Hovland and Judd, 1988; Moss et al., 2012) and elongation (Andresen et al., 2008; Gafeira et al., 2012; Schattner et al., 2016).

Eleven morphological attributes were extracted from GIS for each pockmark: its surface, perimeter, surface/perimeter ratio, internal depth, minor and major axis lengths, major axis direction, elongation (major/minor axis length ratio), bathymetry, slope and morphological domain. Pockmark internal depth has been calculated in two ways, using the Fill method developed in Gafeira et al. (2012) and by calculating the difference between the maximum and minimum bathymetric values over the delimited pockmark surface. These attributes are available online as a SEANOE public database with information on pockmark location and sedimentary facies (Michel et al., 2017).

4. Results

4.1. Pockmark spatial distribution

The studied portion of the Aquitaine continental slope is characterised by a mean smooth slope of ~3°, deepening westward from 220 m down to 1200 m water depth. Deeper, the foot slope is characterised by a gentle slope of 1.5° down to 1600 m water depth. The study area can be divided into two main domains showing different seafloor morphological features. The northern part, from 44°35'50''N to 44°11'44''N latitude, is deeply incised by E-W oriented canyons with heads rooted at the shelf break edge. The southern part, from 44°11'44''N to 43°52'37''N latitude, does not show any canyons, only some landslide scarps located at 230 m water depth and a wide sediment wave field (Fig. 2) located between 250 and 1000 m water depth. Two main sedimentary facies characterised the Aquitaine slope, the sandy silt facies and the silt dominated facies (Faugères et al., 2002; Gonthier et al., 2006;
Michel et al., 2017; Schmidt et al., 2014). The inter-canyon area is dominated in the northern part by the silt facies, and the southern part by sandy silt. The sediment wave field is dominated in the upper slope (between 250 and 500 m water depth) by silt facies and deeper than 500 m by sandy silt facies.

606 pockmarks have been discovered and manually mapped. They are exclusively located on the continental slope, between 350 m and 1150 m water depth, covering 800 km² (Fig. 2). The oceanward extension of the pockmarks is limited by the survey acquisition (Fig. 2). The mapped pockmarks are relatively large, with regards to known pockmarks (Hovland and Judd, 1988; Pilcher and Argent, 2007), with a rough diameter from 52 to 330 m and an internal depth up to 42 m for the largest pockmarks (Fig. 4a). The majority of the pockmarks (80%) have a rough diameter comprised between 100 and 200 m for an averaged internal depth of 15 m (Michel et al., 2017) (Fig. 5).

Of the 606 pockmarks manually mapped, 72% (434 units) occur in the inter-canyon areas (574 km²) and 25% (153 units) in the sediment wave field (374 km²) (Figs. 2 and 5). Pockmark density in the inter-canyon domain is twice as high as in the sediment wave field. The 3% (19 units) remaining pockmarks are located at the foot slope deeper than 1200 m water depth (Fig. 2). In the inter-canyon domain, pockmarks are randomly dispersed on inter-canyons which are oriented east-west (Figs. 2a,b). There is no preferential distribution with regards to the summits of the antiforms and to the borders of the canyons. A few numbers of pockmarks clusters are observed (Fig. 2) with densities up to 12 pockmarks per km². Pockmarks in the sediment wave field are located both on the wave crests (36%) and between the crests (48%) as noticed by Baudon et al. (2013) for similar pockmarks located on the upper slope of the Aquitaine slope south of the studied area. Locally, a few pockmark strings (maximum 8 depressions along 2 km), only concerning less than 13% of the 153 pockmarks mapped in the sediment wave field, are observed related to sediment wave orientation (Figs
The 16% remaining pockmarks are located on flat areas without any spatial alignments. On the whole, less than 10% of the pockmarks are locally aligned (considering at least 4 pockmarks).

4.2. Pockmark characterization

4.2.1. Acoustic signature of water column and surficial sediments

The EM302 water column backscatter data from the GAZCOGNE1 marine expedition do not exhibit any amplitude anomaly in the water column related to gas escaping bubbles, and this throughout the pockmark field and over the 6 days of the acoustic survey (28th of July to 2nd of August 2013).

The averaged seafloor backscatter amplitude within the pockmarks ranges from -34.5 to -21.8 dB in the inter-canyons with a mean value of -27 dB (Fig. 4b). The seafloor backscatter amplitude values vary from -31.6 to -21.9 dB with a mean value of -26.9 dB in the sediment wave field (Fig. 4b). The seafloor backscatter of surrounding sediment, calculated within a 100 m buffer around the pockmark rim, vary from -34 dB to -22 dB with a mean value of -26.9 dB. The EM302 seafloor backscatter values in the majority of the pockmarks are similar to the ones of surficial sediments around wherever pockmarks are located in inter-canyon or sediment wave field domains. Only 15% of the pockmarks exhibit within part of the depression either high or low seafloor backscatter amplitudes, in the northern and southern study area, respectively.

4.2.2. Seismic investigation at the seabed and inside the sediment pile

The acquired sub-bottom profiler lines only cross 38 pockmarks, i.e. 6.3% of all pockmarks. They display the same seismic signature for each pockmark (Fig. 3). The profiles do not exhibit any high seafloor amplitude anomalies, e.g. enhanced reflectors, or high amplitude anomalies within the uppermost 100 m of sediment. Only triplication points due to geometry artefacts below pockmarks are observed. The sedimentary records below and
besides the pockmarks are not disturbed. Moreover, no distinct draped sediment layers are observed within the depressions in regards of the tens of centimetres resolution from the sub-bottom profiler.

FIG 4

4.2.3. Pockmark morphometry

Pockmark surface. The pockmark surface ranges from $0.29 \times 10^4$ m$^2$ to $7.49 \times 10^4$ m$^2$ at the inter-canyon area and from $0.25 \times 10^4$ m$^2$ to $6 \times 10^4$ m$^2$ in the sediment wave field area. The mean value of pockmark surface is $1.8 \times 10^4$ m$^2$ in the inter-canyon and $1.7 \times 10^4$ m$^2$ in the sediment wave field (Fig. 4a). The variations in pockmark size are similar in both morphological domains. A general increase in pockmark surface is observed at shallower water depths but no linear trend is observed (regression line, $R^2 = 0.1259$ in the inter-canyon area and $R^2 = 0.1895$ in the sediment wave field) (Fig. 5). The largest pockmarks ($>3.10^4$ m$^2$ with a rough diameter $>200$ m) occur preferentially in the sandy silt facies with 55 occurrence while only 5 pockmarks occur in the silt facies (Michel et al., 2017).

FIG 5

Pockmark internal depth. The calculation of pockmark internal depth based on the method by Gafeira et al., (2012) leads to strongly minimize the internal depth of the studied pockmarks with results showing that most of the pockmarks (82%) have an internal depth $<1$ m and 8% have an internal depth of 0 m ! Instead, the method based on the difference between maximum and minimum bathymetry provides realistic values. Thus, studied pockmarks have an internal depth ranging from 4 to 42 m with a mean value of 15 m (Fig. 5). The deepest internal depth values correspond to the largest pockmarks ($>200$ m in diameter) with a mean value of 22 m.
**Pockmark elongation.** The pockmark elongation defined as the major/minor axis length ratio values range from 1 to 5.7 with a mean value of 1.4 on the inter-canyon area and from 1 to 2.7 with a mean value of 1.4 in the sediment wave field (Fig. 4c). Most of the pockmarks (88%) are elongated with an elongation superior to 1.1 while only 12% are sub-circular (elongation between 1 and 1.1) (Fig. 2). Elongation values <1.1 are considered as sub-circular shapes in order to take into account potential mapping biases and calculation approximation. Among the elongated pockmarks a majority has an elongation between 1.1 and 1.5 (66%) while 19% have an elongation between 1.5 and 2.2. The most elongated pockmarks with an elongation >2.2 are less common (3%) and mainly correspond to coalescent pockmarks (Fig. 2b, most south-eastern pockmarks). In both morphological areas, elongated pockmarks and sub-circular pockmarks are randomly located (Fig. 2). The sub-circular pockmarks are observed within the sandy silt facies (62%) and the silt facies (30%) whereas the most elongated pockmarks (elongation >2.2) are clearly and only associated with the sandy silt facies (Michel et al., 2017).

**Pockmark direction.** The major axis direction of the pockmarks with elongation values >1.5 (134 units) has been compared to the surrounding slope value (Fig. 6). These pockmarks correspond to 92 depressions in the inter-canyon domain and 42 in the sediment wave field (Fig. 6). In the inter-canyon domain (Fig. 6a), local slope orientation around the pockmarks is mostly E-W while the pockmark major axis is mostly NW-SE, with 40% of them oriented N150-330 and 35% others oriented N120-300. In the sediment wave field (Fig. 6b), the local slope around the pockmarks is oriented around N300 and the pockmark major axes are mostly oriented WNW-ESE, 40% of them oriented N100-280 and 22% others oriented N120-300.
4.3. **Bottom currents in the sediment wave field**

Current direction and amplitude distributions are displayed in current roses (Fig. 6) with E-W and N-S current components (Fig. 7). Current velocities derived from the raw signal are mostly smaller than 10 cm/s (90% of the records for the E-W component and 81% for the N-S component) (Fig. 6c) with the maximum amplitude reaching 31 cm/s (Fig 7). Currents vary on different time scales (Fig. 7), associated with different forcing factors. High-frequency large amplitude tidal signals coexist with weaker and longer period signals. The tidal signal is mostly oriented E-W, and exhibits a significant cross-slope component (Fig. 7a). The longer-period component (red curves) is oriented along-slope (Fig. 7b) due to the geostrophic constraint, as evidenced by the red dots in Fig. 6. Its cross-slope component is always smaller than 5 cm/s (Fig. 7a). The along-slope component (Fig. 7b) is almost always weaker than the tidal current (for 81% of the records), but can reach high instantaneous values during specific events (higher than 15 cm/s, 6% of occurrence).

**FIG 7**

5. **Discussion**

5.1. **Pockmark detection and characterisation**

*Pockmark mapping.* Both semi-automatic methods and manual picking show advantages and drawbacks. Semi-automatic methods are based on a succession of quick numerical calculations, but most of these latter have to be manually checked to limit the number of artefacts. 5433 features were detected as depressions with the “Fill” method (Gafeira et al., 2012) and 10437 with the BPI method (Wright et al., 2012) whereas the manual picking only gives 606 pockmarks. The elimination of a large amount of artefacts is time-consuming, hence defeating one of the main advantages of semi-automatic methods. Although manual picking is considered as time-consuming method, it is much more appropriate in the case of
complex seafloor morphologies due to human capability to focus on features of interest.

Indeed, along the Aquitaine slope, there is the superimposition of different scale morphologies such as slope, canyons and sediment waves that prevent the semi-automated detection process from being accurate. Thus, semi-automatic methods should be used in relatively flat bathymetry areas to obtain successful results, e.g. at continental the shelf (Gafeira et al., 2012), bays (Andrews et al., 2010) and in basins (Geldof et al., 2014). For large extents and huge densities but of similar features, the automatic methods are clearly efficient (Andrews et al., 2010; Gafeira et al., 2012; Geldof et al., 2014).

Pockmark internal depth. It is clear from our results that the Fill method is not able to calculate the effective infilling of the studied pockmarks, most likely because of their irregular morphology (e.g. collapsed flank) and regional slope of 3°. Thus, this method suits uniform areas with well-shaped pockmarks (Gafeira et al., 2012; Geldof et al., 2014) but does not fit with complex morphologies with slopes. In the latter case, it is more appropriate to calculate the internal depth by subtracting the maximum bathymetry over the entire pockmark from the minimum one.

5.2. Pockmark inactivity and nature of the fluids involved

Free gas leakage produces clear water column backscatter anomalies commonly used to attest seepage activity (Dupré et al., 2015, 2014a; Klaucke et al., 2006). During the GAZCOGNE1 survey, no water column acoustic anomalies corresponding to gas bubbles were detected in the whole slope area including the 606 pockmarks. Although the temporal variability of seepage activity may be invoked, the 6 days of the acoustic survey are sufficient to cover the time window for the tidal cycle which could be a possible triggering mechanism (Baltzer et al., 2014). Thus, pockmarks along the Aquitaine slope are interpreted as currently inactive in terms of free gas seepage.
Considering the sediment cover, methane-derived authigenic carbonates are considered as confident indicators of long-term gas circulation (Bayon et al., 2013, 2009). Outcrops and sub-outcrops of carbonate structures are easily detected on seafloor backscatter data as occurrence of high amplitude anomaly patches (Dupré et al., 2010; Klaucke et al., 2006). The lack of high seafloor backscatter values within the pockmarks and the similarity of seafloor acoustic signature between the pockmarks and the surrounding sediments clearly provide evidence for the absence of methane-derived authigenic carbonates along the Aquitaine slope.

Within the uppermost 100 m of the sediment, sub-bottom profiles across pockmarks do not exhibit any enhanced reflectors and diffracting points at the seabed pile that carbonates would seismically produce if present (Dupré et al. 2010). No acoustic blanking, blank chimneys and any other seismic evidence of gas accumulations (Brothers et al., 2011; Løseth et al., 2009) within the vertical resolution limit of twenty centimetres are observed. In other words, at the present day the absence of acoustic anomalies within sedimentary records excludes the occurrence of 1) layers charged with free gas, 2) buried pockmarks and 3) carbonates underlying or disconnected from the present-day seafloor pockmarks.

Based on these observations and interpretations, the pockmarks along the Aquitaine slope may have been formed by dewatering (Harrington, 1985), fresh water expulsion (Whiticar, 2002) or short-duration gas escapes, associated with a relatively shallow source level (the pockmarks being rooted a few metres to maximum a few tens of metres below the seafloor) (Judd and Hovland, 1992). Indeed, gas releases over a long period of time (order of ky years) would have led to authigenic carbonate precipitation (Andresen et al., 2008; Bayon et al., 2009). Although the pockmarks along the Aquitaine slope are located away from the hydrate stability zone, it is unlikely with regards to the absence of fluid evidence that they have been formed by gas hydrate dissociation as suspected along the U.S. Atlantic continental
Moreover, the morphology and acoustic signature of the studied pockmarks do not fit those of hydrate-bearing pockmarks (Davy et al., 2010; Riboulot et al., 2016; Sultan et al., 2010). The latter are generally kilometre large depressions with internal filling of disturbed sediments caused by hydrate destabilization. A few smaller pockmarks may be associated with these mega structures but exhibit disturbed sediments underneath (Davy et al., 2010).

Based on sub-bottom profiler data displayed in Gonthier et al. (2006) and in accordance with the seismic signature of pockmarks from our dataset, we suspected the occurrence of pockmarks within the recent sedimentary cover, which corresponds in the sediment wave field mainly to the so-called U4 unit (Faugères et al., 2002). The formation of the pockmarks appears therefore to postdate the sediment wave formation (U3 unit). Based on the age of the base of the 12-15 m thick U4 unit, which depends on the sediment rates, 10 cm/ky (Winnock 2013) or 100 cm/ky (Schmidt et al., 2014, 2009), the pockmarks along the Aquitaine slope may have been initiated 120-150 ky BP or more recently 12-15 ky BP, respectively. Within this context, sea level falls may have triggered fluid escapes and initiation of pockmarks in the Aquitaine Basin as evidenced e.g. in the Gulf of Lions (Riboulot et al., 2014) and offshore West Africa (Andresen and Huuse, 2011; Plaza-Faverola et al., 2011; Riboulot et al., 2013). But without any detailed seismic data and dating of long cores through the Aquitaine slope, it is impossible to conclude.

With regards to the available data, the non-conclusive nature of the fluids and because of their morphology and repartition, it is rather difficult to establish any link between the pockmarks along the Aquitaine slope and other known, but not much documented, fluid systems of the Bay of Biscay: 1) the Capbreton Canyon area where size-differentiated pockmarks are related to different migration pathways (Baudon et al., 2013; Gillet et al.,
2008), 2) deeper offshore mega-pockforms on the Landes Plateau (Baudon et al., 2013; Iglesias et al., 2010) and 3) gas emissions at the Aquitaine Shelf (Dupré et al., 2014a).

5.3. **Origin of pockmark elongation: slope, coalescence, currents?**

Elongation is the only attribute able to differentiate pockmark groups along the Aquitaine slope. As it is assumed that pockmarks initially have a sub-circular shape (Hovland and Judd, 1988), why the majority of the pockmarks (88%) located along the Aquitaine slope (deeper than 350 m water depth) are elongated?

With regards to the inactivity and the absence of present and past fluid evidence, it is unlikely that successive fluid releases have occurred, and even less that this has been able to reshape the pockmarks. The slope along which pockmarks may become elongated and open downslope (Brothers et al., 2014) may be another explanation for pockmark elongation. This may be possibly for some pockmarks in the sediment wave field area but is unlikely in the inter-canyon domain. Coalescence of several pockmarks may in places explain some of the observed elongated pockmarks along the Aquitaine slope, especially in the northern part.

Similarly to others areas, the influence of the bottom currents on the pockmark morphology, namely their elongation, is questioned.

Current-induced processes, such as high density flow on the slope (Kuhnt et al., 2013) and internal tide impacting the seabed (Pingree et al., 1986), producing strong shear stress on the seafloor (Durrieu De Madron et al., 1999; Kuhnt et al., 2013), may influence seafloor morphology. However, the Aquitaine slope, except within the canyons, does not appear to be greatly affected by internal tide (Durrieu De Madron et al., 1999; Kuhnt et al., 2013).

Two main current regimes are evidenced along the Aquitaine slope, one driven by the tide and mostly oriented east-west and a second long-period current mostly oriented north-
south, both mainly showing velocity values lower than 10 cm/s. The elongated pockmarks along the Aquitaine slope are not aligned along the present-day bottom current direction. Some higher velocity values, reaching up to 31 cm/s in the along-slope S/N direction are observed similarly to other near-bottom current measurements across continental shelves (Schattner et al., 2016) and slopes (Tallobre et al., 2016). These events have been associated to westerly-wind pulses occurring along the Cantabrian Slope (Batifoulier et al., 2012). It is worth noting that the recorded current velocities at the mooring between 17 to 33 m above the seafloor provide an order of magnitude but are likely to be a strong overestimate of flow velocities near the seafloor. Along the Aquitaine slope, current velocities of 10 cm/s are sufficient to limit sedimentation for silt and mud (Migniot, 1977) therefore preventing pockmark filling. Moreover, these relatively low current velocities may induce upwelling within the depressions preventing fine sediments from being deposited (Brothers et al., 2011). This would not exclude the accumulation of coarser sediments within the pockmark as observed in several pockmarks located in an inter-canyon area. On the other hand, in order to remobilize consolidated silt and mud, velocities higher than 30 cm/s are necessary (Migniot, 1977). The sediment cover from silt to silty sand along the Aquitaine slope may be considered as less consolidated than pure silt or pure mud. But present-day tide velocity and N-S current velocity are not strong enough to suspend sediment along the Aquitaine slope. Moreover, elongated pockmarks occurring randomly amidst sub-circular ones are not coherent with the influence of a regional bottom current. Considering that pockmarks along the Aquitaine slope have been formed at the same time, the post-formation processes that have reshape and elongate the pockmarks along the WNW-ESE axis may be related to a former different current regime than the present-day one. Even if the current regime itself or more likely the impacts of bottom currents on the seafloor morphology did change in the Bay of Biscay since the initiation of the pockmarks, it is
443 proposed that upwelling induced by near-bottom currents within the pockmarks may
444 contribute to modify the depressions (Brothers et al., 2011; Pau et al., 2014). Upwelling
445 within the pockmark may contribute to elongate the pockmark and maintain its shape,
446 preventing sedimentation by winnowing out the fine grained sediments. Relatively weak near-
447 bottom currents (20 cm/s), as it is along the Aquitaine slope, appear sufficient to induce such
448 upwelling (Brothers et al., 2011). This upwelling does not exclude the occurrence locally of
449 stronger current events that could have also contribute to modify the seafloor morphology.

6. Conclusion

450 The geophysical survey conducted on the Aquitaine slope revealed numerous
451 pockmarks (606) over 800 km² occurring on canyon interfluves and in the sediment wave
452 field at water depths deeper than 350 m. These pockmarks are relatively large with the
453 majority of them having a rough diameter comprised between 100 and 200 m for an average
454 internal depth of 15 m. Multi-data analyses point to an absence of present-day free gas
455 releases and free gas accumulations within the sedimentary column and to an absence of hard
456 substrates that could account for methane-derived authigenic carbonates, usually associated
457 with long-term seepage. Therefore at present time, pockmarks are confidently interpreted as
458 inactive fluid-escaping structures. These pockmarks were most likely formed by past short-
459 duration fluid-release events associated with gas or water, and appear to be mainly comprised
460 within the upper most sedimentary pile (a few meters to a few tens of metres thick)

463 Semi-automatic methods to map pockmarks are not appropriate in the study area
464 because of the complex bathymetry inherited from several orders of morphologies, the slope
465 angle and the presence of for example canyons and sediment waves. Pockmark morphometry
466 was therefore based on manual mapping.
Pockmarks along the Aquitaine slope are randomly distributed with regards to the water depth and the slope and the surrounding morphology, namely the antiform structures in the inter-canyons areas and the crests and inter-crests in the sediment wave field, as observed elsewhere (Gafeira et al., 2012; Rise et al., 2015). Shelf-indenting submarine canyons are pockmark free zones as observed along other shelves (Brothers et al., 2014). Moreover there is no positive correlation between the dimension of the pockmark and the water depths (Gafeira et al., 2012; Schattner et al., 2016). Thus, the spatial distribution of pockmarks along the Aquitaine slope does not seem to be driven by seafloor morphology and structural feature, although a few pockmark strings are observed in the sediment wave field. As pockmark size is often influenced by the nature and thickness of sediments (Baltzer et al., 2014; King and MacLean, 1970; Rise et al., 2015), it appears that the sandy silt facies, in contrast to the silt facies, may favour the occurrence of only the largest pockmarks (with diameter >200 m) along the Aquitaine slope.

Pockmarks along the Aquitaine slope are divided in sub-circular (12%) and elongated (88%) pockmarks including some coalescent. The slope, as the primary controlling factor, constrains the elongation of part of the pockmarks. But for the other elongated pockmarks, present-day bottom-currents (velocity and direction) are not compatible with their morphology. It is proposed that upwelling induced by relatively weak near-bottom current along the Aquitaine slope contributes to the elongation of the pockmarks and the maintenance of their shape (Brothers et al., 2011; Pau et al., 2014).

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References


Dandapath, S., Chakraborty, B., Karisiddiaiah, S.M., Menezes, A., Ranade, G., Fernandes, W.,


Gonthier, E., Cirac, P., Faugères, J.C., Gaudin, M., Cremer, M., Bourillet, J.F., Faugeres, J.C.,


Figure captions

Fig. 1: Location map of the main morphological areas of the Bay of Biscay: Armorican Shelf, Aquitaine Shelf, Cantabrian Shelf and Landes Plateau. The study area (red rectangle) covers the western extension of the Parentis Basin (from Biteau et al., 2006) and the eastern Landes Plateau. Isobath lines are extracted from compiled bathymetry by Sibuet et al. (2004). The three main regional current regimes are displayed, tidal current (Batifoulier et al., 2012; Charria et al., 2013), contour current (Van Aken, 2000) and wind forced currents (Kersalé et al., 2016).

Fig. 2: a) Detailed shaded bathymetry map of the Aquitaine Margin with main seafloor morphologies: pockmarks, canyons and sediment waves. Data acquired during GAZCOGNE1 (2013) and BOBGE02 (2010) marine expeditions. Background bathymetry from EMODnet Bathymetry portal (http://www.emodnet-bathymetry.eu) compiled at 250 m grid size. Aspex current mooring 10 is located at a water depth of 450 m in the sediment wave field. Slope focus on b) elongated pockmarks in the northern study area and c) sub-circular pockmarks in the sediment wave field.

Fig. 3: Processed sub-bottom profiler line displayed in envelope in ©Kingdom Software. X axis corresponds to Shot Point (SP) and Y axis to depth in seconds in Two Way Time (TWT). The profile is displayed with a Vertical Exaggeration (VE) of 7, calculated with a seismic wave velocity of 1500 m/s, with indication of slope angle. This sub-bottom profiler line crosses two pockmarks (see location in Fig. 2a) without any fluid evidence and is exhibiting only triplication points so-called "bow tie" artefacts.

Fig. 4: Box plots of 3 attributes extracted from GIS (Michel et al., 2017) for both domains of inter-canyons and sediment waves: a) pockmark surface area with indication of diameter with regards to...
724 pockmark surface (a circular pockmark with a diameter of 200 m corresponds to a surface of
725 $3 \times 10^4 \text{m}^2$), b) pockmark seafloor backscatter amplitude from the 30 kHz EM302 multibeam data and
c) pockmark elongation (major/minor axis length ratio). Bathymetry, slope (15 m grid size) and
727 seafloor backscatter (10 m and 15 m grid size) were acquired during the GAZCOGNEI expedition (30
728 kHz EM302). Red curves stand for the contour of pockmarks. The legend of the box plots is displayed
729 in Fig. 4c, with representation of the minimum, maximum, first quartile (Q25), second quartile (Q50
730 or median), third quartile (Q75) of the series and serie outliers.

731
732 Fig. 5: Scatter plots of pockmark surface versus bathymetry with internal depth as point colour, for
733 both domains a) inter-canyon and b) sediment wave areas with regression lines and determination
734 coefficients ($R^2$).

735
736 Fig. 6: Rose diagrams of the pockmark major axis direction (green arrows) and local slope direction
737 around pockmarks (orange shape) for a) inter-canyon and b) sediment wave areas. Arrow and shape
738 lengths are proportional to the number of pockmarks involved. Recorded velocity and orientation of
739 currents from ASPEX mooring 10 are shown in black and red dots for raw and tide-filtered data,
740 respectively. c) Diagram of spatial distribution of current velocities and orientations of raw signals.

741
742 Fig. 7: Bottom current velocity, a) east-west ($U_E$) and b) north-south ($U_N$) components, recorded with
743 ASPEX mooring 10 (see location in Fig. 2). Current velocities are integrated between 17 m and 33 m
744 above the seafloor. Recorded velocity and orientation of currents are shown in blue and red curves for
745 raw and tide-filtered data, respectively.
a) Inter-canyon

b) Sediment wave