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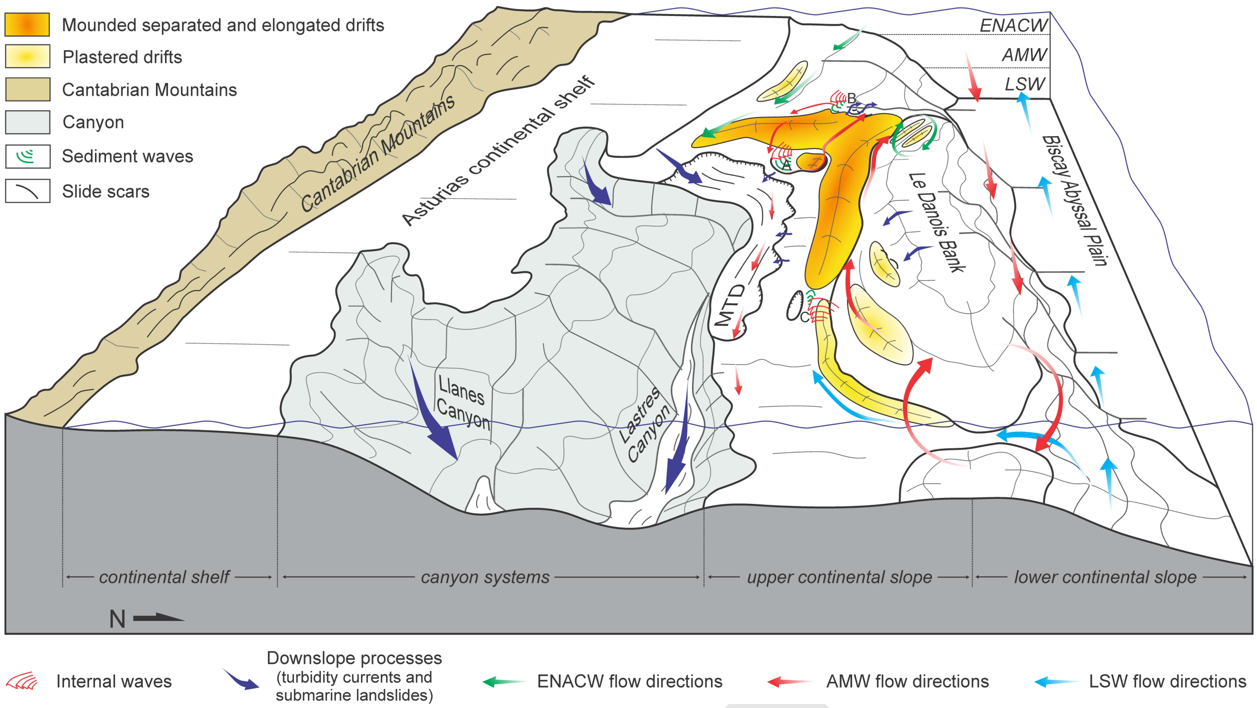

Mounded separated and elongated drifts
Plastered drifts
Cantabrian Mountains
Canyon
Sediment waves
Slide scars
ENACW
AMW
LSW
Cantabrian Mountains
Asturias continental shelf
Biscay Abyssal Plain
Le Danois Bank
Llanes Canyon
Lastres Canyon
MTD
A
B
C
continental shelf
canyon systems
upper continental slope
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N
Internal waves
Downslope processes (turbidity currents and submarine landslides)
ENACW flow directions
AMW flow directions
LSW flow directions

Morphological features and associated bottom-current dynamics in the Le Danois Bank region (southern Bay of Biscay, NE Atlantic): a model in a topographically constrained small basin

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Abstract

The present-day morphology of the Le Danois Bank region has been investigated based on bathymetric and high to ultra-high resolution seismic reflection data. The involved bottom-current processes are associated with the Eastern North Atlantic Central Water, the Atlantic Mediterranean Water and the Labrador Sea Water. Sediments originating from various canyon systems along the Cantabrian Margin and the Asturias continental shelf are transported by downslope and alongslope processes towards the Le Danois intraslope basin. The background flow velocities of bottom currents are all below the threshold (8-10 cm/s) of generating plastered and mounded geometries of contourite drifts. However, bottom currents are locally accelerated (up to 25 cm/s) due to the presence of the Le Danois Bank and the Vizco High, creating a furrow and three moats and generating six plastered drifts, three elongated mounded and separated drifts at different depth intervals. The extension and distribution of the drifts are controlled by slope morphology and/or bottom current velocities. Unlike contourite drifts along other continental slopes, a single contourite drift (the Gijón Drift) with a lateral variation in drift geometry and internal structure indicates the interaction of bottom currents with different flow dynamics. Additionally, scouring of active bottom currents and rapid sedimentation rate of contourite drifts may be at the origin of slope instability events. Besides contourite drifts, internal waves may have induced the formation of sediment waves. In the Le Danois intraslope basin, multiple sedimentary processes work together and shape the present-day seafloor. Bottom currents are focused due to deflection on complex topographical obstacles within a relatively small basin setting, and create a wide variety of sedimentary features, including contourite drifts. The resulting sedimentary features thus have more frequent lateral variations, a feature typical for topographically constrained small basins.

Keywords: bottom currents; contourites; southern Bay of Biscay; small basin.

1 Introduction

Bottom currents are currents flowing near the bottom of the ocean (Faugères and Stow, 1993) that can be classified based on their driving forces. Examples are thermohaline circulation, wind-driven currents, geostrophic currents, etc. (Rebesco et al., 2008). Bottom currents generated by thermohaline circulation are described as semi-permanent alongslope currents (Rebesco et al., 2014). Alongslope bottom currents are capable of suspending, transporting and/or controlling the deposition of sediments at the seafloor (Stow et al., 2002). Sediments deposited or significantly affected by bottom currents are known as contourites, and the sedimentary bodies they create are called contourite drifts (Faugères and Stow, 1993; Faugères et al., 1999; Stow et al., 2002). Bottom currents are also capable of creating erosional/non-depositional features such as contourite channels and moats (Rebesco et al., 2014; Hernández-Molina et al., 2016). The lateral and temporal variations of these contourite deposits (drifts) associated with a variety of erosional features lead to the definition of the Contourite Depositional System (CDS) (Hernández-Molina et al., 2008).

During the last two decades, numerous studies have focused their attention on large margin-scale contourite deposits, such as the well-studied Cádiz CDS along the SW Iberian continental margin (Llave et al., 2001, 2006; Stow et al., 2013), the Lofoten and Vesterålen Drifts along the NW European Atlantic margin (Laberg et al., 2005; Stoker et al., 2005) and the giant CDS along the Argentine continental margin (Preu et al., 2012, 2013). Bottom-current processes associated with these large-scale drift systems are being widely investigated (Maldonado et al., 2005; Hernández-Molina et al., 2010). However, small-scale bottom-currents in smaller basins and around topographical obstacles could strongly interact with seafloor morphology and oceanographic processes (e.g. internal waves/tides, eddies, Chen et al., 2019). The related processes received far less attention and their mechanisms are not yet fully understood (Turnewitsch et al., 2004; Rebesco et al., 2014; Van Rooij et al., 2016; Zhang et al., 2016).

Small-scale contourite drifts have been documented at different continental margins (Tournadour et al., 2015; Juan et al., 2016). They are reported around the mud volcanoes of the Moroccan Atlantic margin (Vandorpe et al., 2016) and the Gulf of Cádiz (Palomino et al., 2016), around the Pontevedra obstacle in the Galicia Bank region (Ercilla et al., 2011; Zhang et al., 2016), associated with cold-water coral mounds in the Porcupine Seabight, SW of Ireland (Van Rooij et al., 2007), around a seamount in the South China

Sea (Chen, et al., 2014) and associated with the Le Danois Bank along the northern Iberian continental margin (Van Rooij et al., 2010). Whereas the steep northern flank of the Le Danois Bank is dominated by gravitational processes, the intraslope basin hosts small-scale contourite deposits. The related drift systems extend from the Le Danois Bank (in the north) to the Asturias continental upper slope (in the south) and are dominating the geomorphology of the present intraslope basin. Based on some sparsely distributed seismic profiles, these drift systems seem to be highly affected by the interaction between the intermediate water mass and the associated morphologies of the intraslope basin (Ercilla et al., 2008; Iglesias, 2009; Van Rooij et al., 2010). Types and dimensions of depositional, erosional and mixed features show remarkable variations: elongated mounded and separated drifts (the Le Danois and the Gijón Drifts), plastered drifts, moats (the Le Danois and the Gijón Moats) and slide scars have all been identified (Iglesias, 2009; Van Rooij et al., 2010). However, the dynamic interaction between the present-day bottom-current circulation and the detailed characteristics, as well as the spatial variability of the drift systems are still not fully understood.

Due to the presence of a topographic obstacle (e.g. a coral bank, slide scar or fault), bottom currents will enhance locally and have the capability of creating flow filaments, in turn creating their own erosional and depositional features (Stow et al., 2002, 2008; Rebesco et al., 2008, 2014). The Le Danois Bank region is ideal for investigating the relationship between various topographic obstacles, bottom current dynamics, and different contourite features. Mediterranean Outflow Water (MOW) is considered to be the most important water mass (Ercilla et al., 2008, 2011; Van Rooij et al., 2010), referring to the pure thermohaline outflow in the Gulf of Cádiz (Hernández-Molina et al., 2014). This water mass will now be referred to as the Atlantic Mediterranean Water (AMW), due to its different physical properties and dynamics (Rogerson et al., 2012; Flecker et al., 2015). The location of the Le Danois Bank region is thus key for better understanding the local interaction of the AMW with the continental slope sedimentary processes.

This study aims to provide a comprehensive description of the distribution and current morphology of various depositional and erosional features of the Le Danois Bank region. By combining previous observations with additional ones derived from more recent seismic datasets and improved insights regarding the present-day oceanographic setting, this study aims to improve the understanding of the recent interaction between topographic obstacles and bottom currents. As such, this study introduces a unique model related to the responsible bottom-current dynamics in a topographically constrained small basin.

2 Regional setting

2.1 Geology and morphology

The Le Danois Bank, an E-W narrow topographic high, is located at the Cantabrian continental slope, in the southern Bay of Biscay (Fig. 1). The northern flank (mean slope values between 18° and 20°, reaching 34° locally) of the Le Danois Bank steeply drops into the Biscay abyssal plain, whereas the southern flank (slope gradient varying from 0.8° to 15°) is connected to the upper continental slope (Figs. 2, 3). The presence of the Le Danois Bank creates an intraslope basin (about 65 km long and 15-25 km wide) between the bank and the Asturias continental shelf (Figs. 2, 3). This basin is bounded to the west by the Vizco High and to the southeast by the Lastres Canyon (Fig. 1b). The basin becomes shallower from the centre (1070 m) towards the bank (855 m) and the continental shelf (443 m).

The Cantabrian margin was deformed by compression during the Early Paleocene to the Eocene when the northward displacement of the African plate led to the Iberian-European collision, resulting in the Cantabrian-Pyrenean chain along the northern border of the Iberian plate (Boillot et al., 1979; Alvarez-Marrón et al., 1996). In the Early Oligocene, the subduction of the Iberian plate beneath the European plate, as well as shortening of the northern Iberia plate, resulted in regional compression in the Bay of Biscay (Fernández-Lozano et al., 2011). The subsequent mantle exhumation and crustal thinning beneath the Cantabrian continental margin induced the formation of the Le Danois Bank and the intraslope basin (Vissers and Meijer, 2012; Cadenas and Fernandez-Viejo, 2016). At the end of the Oligocene, the collision of the Iberian and European plates halted and there is no evidence of large-scale tectonic activity in the Le Danois Bank region from the Miocene onwards (Muñoz, 2002; Vergés et al., 2002; Roca et al., 2011; Tugend et al., 2014; Cadenas and Fernandez-Viejo, 2016).

The contourite deposits (drifts) are observed in the Pliocene unit (Van Rooij et al., 2010). Confined contourite drifts have been identified in the Upper Pliocene unit and the elongated mounded and separated drifts are generated in the Middle Pleistocene unit (Van Rooij et al., 2010). The sediment supply consists of fine-grained material eroded from the Cantabrian Mountain range (Gaudin et al., 2006; Ercilla et al., 2008). Gómez-Ballesteros et al. (2014) suggested that the sediments are transported into the ocean through the Narcea and Nalón rivers (Fig. 1a) and move along the Cantabrian continental slope eastwards by means of alongslope currents.

2.2 Oceanography

At present, three water masses are found below the surface mixed layer in the Le Danois Bank region: (1) The Eastern North Atlantic Central Water (ENACW) between 350 and 600 m water depth (McCartney and Mauritzen, 2001), (2) the Atlantic Mediterranean Water (AMW) between 750 and 1550 m water depth (Iorga and Lozier, 1999a, b), and (3) the Labrador Sea Water (LSW) between 1750 and 2000 m water depth (van Aken, 2000a). Between these water masses, two interfaces are located respectively at 600 to 750 m and 1550 to 1750 m water depth (van Aken, 2000a, b) (Fig. 4).

The ENACW originates at the northeast of the Azores (Pollard et al., 1996). Subtropical origin ENACW enters the Bay of Biscay as an eastward slope current along the northern Iberian continental slope (Pingree, 1993; Pollard et al., 1996; McCartney and Mauritzen, 2001). Its subtropical branch constitutes an eastward slope current along the northern Iberian continental slope and flows polewards (Pingree and Le Cann, 1990; van Aken, 2000b). The ENACW contains relatively small variations in temperature and salinity in the Bay of Biscay, ranging respectively between 11.8° to 12.2 °C and 35.53 to 35.58 (Haynes and Barton, 1990; Ríos et al., 1992; Pingree, 1993). The mean velocity is around 1 cm/s in the southern Bay of Biscay (Frouin et al., 1990; Lavín et al., 2006). In the Le Danois Bank region, the ENACW has a higher velocity (10-30 cm/s) and flows eastwards along the Asturias continental upper slope (Lavín et al., 2006; González-Pola et al., 2012). Along the rim of the eastern summit of the Le Danois Bank, an anticyclonic circulation with a velocity of about 15 cm/s has been observed at the ENACW depth range (González-Pola et al., 2012) (Fig. 1b). This circulation cell is most likely induced by a steady background flow impinging on the topographical barrier or interaction between periodically enhanced currents and the Le Danois Bank (González-Pola et al., 2012). Its core is located at 400 m water depth with a minimum salinity of 35.5 (Fig. 4).

At a depth of about 1000 m, a core of eastward-flowing AMW is observed hugging the northern Iberian continental slope with a mean velocity of 1-5 cm/s (Iorga and Lozier, 1999a). In the Le Danois Bank region, its maximum salinity reaches 35.8 while the temperature remains about 10°C (Fig. 4b). Most of the AMW flow penetrates along the northern flank of the Le Danois Bank and there is no dominant flow within the intraslope basin at present day (González-Pola, data not shown). The intraslope basin has the capability to develop its own recirculation pattern, due to the topographic constraints of the Le Danois Bank and the continental shelf (González-Pola et al., 2012). Anticyclonic recirculation during upwelling conditions and cyclonic recirculation during downwelling conditions have been observed at the AMW level between the bank and the continental shelf (González-Pola, personal communication) (Fig. 1b). At the eastern edge of the Le Danois Bank, a pronounced anticyclonic circulation, potentially caused by eddy shedding from the shelf, is also recorded at the AMW level by a single observation (González-Pola et al., 2012) (Fig. 1b). It symmetrically distributes over the bank and has a mean velocity of 10 cm/s (González-Pola et al., 2012).

Below the AMW, the LSW penetrates along the northern Iberian continental margin from east to west (Lazier, 1973; Gascard and Clarke, 1983; van Aken, 2000a). In the study area, the core of the LSW is recognized at 1800 m water depth with a minimum salinity of 35.05 (Figs. 4b, f). Due to the depth of the intraslope basin (max. 1720 m), the LSW is hindered and fails to penetrate the basin towards the west. Intense diapycnal mixing, resulting from the highly energetic internal tidal waves, has been observed between the LSW and the AMW over the Cantabrian continental slope (Pingree, 1993; van Aken, 2000a, b). This mixing action makes the Le Danois Bank region a focal point for the occurrence of the high-energetic turbulent events at the interface between the AMW and the LSW (van Aken, 2000b; Lavín et al., 2006).

3 Material and methods

Three sets of reflection seismic data have been used for this study (Fig. 1b). These datasets embrace different vertical resolutions, ranging from high to ultra-high. These different datasets contain nearly overlapping NNE-SSW to N-S orientated profiles, enabling a more detailed seismic interpretation of the morphological features. The ultra-high resolution TOPAS (topographic parametric sonar) PS18 profiles were obtained during the BIO Hespérides campaign MARCONI II in 2003 (Ercilla et al., 2008; Iglesias, 2009) and penetrates down to 200 ms two-way travel time (TWT) at full oceanic depth. The primary frequency was 15 kHz (secondary 0.5-6 kHz), while the maximum vertical resolution was about 0.2 m. Four S-N trending and three E-W trending TOPAS lines were acquired in the study area with a mean spacing of 25 km. The high-resolution single-channel sparker seismic profiles were obtained during the R/V Belgica cruise ST1118a in June 2011. A 500 J energy SIG sparker (800 Hz frequency) has been used, with a shot interval of 2.5 s. The penetration depth of the acoustic signal varies around 500 ms TWT and the vertical resolution varies around 1.5 m. Twelve NNE-SSW and fifteen W-E to WNW-ESE orientated sparker seismic lines with a spacing of 3-5 km have been acquired. The three-channel airgun seismic records have been acquired during the MARCONI II campaign. The seismic source, a 5-meter-long airgun array of 6 Sercel G-GUN II (140 bars, 80 Hz frequency), was located at 2.5 m depth with a shot interval of 6 s. The receiver system was a 150 m SIG streamer with 3 sections of 40 hydrophones each. The penetration of the acoustic signal is about 1.5 s TWT. The standard vertical resolution was about 4.5 m. Seventeen NNE-SSW and nine WNW-ESE orientated airgun seismic lines with a spacing of 10-15 km were surveyed.

The TOPAS and sparker seismic data have been processed using the DECO Geophysical RadexPro Software. They were corrected for spherical divergence, amplitude loss and burst noise. The sparker seismic data was further processed using a swell filter and an Ormsby bandpass filter (160 to 250 Hz and 1400 to 1500 Hz). The airgun seismic data was processed on board using the Delph Seismic Plus Software and included a bandpass filter (2.5 kHz high pass and 80 kHz low pass). Afterwards, the data were processed by applying a spherical divergence correction, an interactive velocity analysis, and a burst noise removal.

The multibeam data had been acquired with the RV Vizconde de Eza in 2003 within the framework of the ECOMARG project (http://www.ecomarg.com/). The positioning data was provided by Seapath 204, while the SIMRAD EM 300 provided the bathymetric data. This system operated at a water depth ranging between 200 and 3000 m and had a vertical error of less than 1 %. The sampling frequency was set at 30 kHz. The footprint is 15 x 15 m at 500-600 m and 25 x 25 m at about 1000 m. The data covered an area of 5530 km^2. It encompasses the entire study area including the Le Danois Bank, the intraslope basin, the Vizco High, the Lastres Canyon and the adjacent continental shelf (Figs. 2, 3). Vertical CTD profiles were extracted from the World Ocean Database (2013) (https://www.nodc.noaa.gov/OC5/WOD13/). The salinity and temperature cross sections (Figs. 4, 9) were made using Ocean Data View (ODV) software.

4 Results

In the Le Danois Bank region, the most pronounced physiographic features were already reported by Ercilla et al. (2008) and Van Rooij et al. (2010). Based on these previous results, the combination of three currently available seismic datasets, the multibeam data and the updated classification of contourite drifts presented in Rebesco et al. (2014), new insights have been provided on morphosedimentary characteristics of alongslope, downslope and mixed features and their distribution.

4.1 Alongslope features

4.1.1 Elongated mounded and separated drifts

The SW-NE oriented Le Danois Drift is located between 790 and 1080 m water depth (Figs. 5, 6a, b, c). The drift encompasses an area of 242 km^2 and is about 42 km long and 4-11 km wide and 50 m high. The main element characterizing this drift is the high mounded geometry (slope gradient of 0.5° to 2.8°) (Fig. 3). Internally, it comprises continuous parallel-stratified reflections, onlapping upslope with a sigmoidal-oblique progradational pattern towards the Le Danois Bank (Fig. 6c). Towards the south, the mean thickness laterally decreases from 280 to 130 ms TWT. At the western limit of the Le Danois Drift, the thickness reaches its minimum (10-20 ms TWT).

The NW-SE oriented Gijón Drift is located at the southernmost part of the intraslope basin between 320 m and 1060 m water depth (Figs. 5, 6a, b, d). The drift is 34 km long and 2 to 13 km wide. It displays a broad mounded geometry (slope gradient of 1° to 3°) and a basal unconformity with onlap terminations (Fig. 6d). The mounded part rises 80 m above the surrounding seafloor. The seismic facies of the Gijón Drift is characterized by stratified layers of continuous sigmoidal reflections interbedded with discontinuous low-amplitude chaotic reflections (Figs. 6a, d). Towards the north, the sedimentary bodies of the Gijón and the Le Danois Drifts overlapped, where the seismic reflections are continuous parallel-even configurations (Fig. 6a). The thickness of the Gijón Drift gradually decreases from 320 to 30 ms TWT towards the southeast. The southeast extension of the Gijón Drift is limited by the presence of the Lastres Canyon (Fig. 5). Towards the northwest, the geometry of the Gijón Drift evolves from elongated and mounded (Figs. 6a, d) to confined and mounded (Fig. 6b). The change of the geometry is associated with a reduction in thickness (from 220 to 110 ms TWT). The internal structure of this part of the Gijón Drift is characterized by convex-upward seismic patterns with high- to medium-amplitude sigmoidal reflections thinning towards the edges (Fig. 6b).

At the centre of the intraslope basin, a small elongated mounded and separated drift is identified (Fig. 6c). Based on the proximal physiographic feature, the Asturias continental shelf, this drift is referred to as the Asturias Drift (Fig. 5). The Asturias Drift covers an area of 21 km^2 at water depths between 900 and 1080 m. It is characterized by an unconformity surface with onlap and downlap terminations and mounded geometry (slope gradient of about 1°) (Figs. 3, 6c). The drift displays a sigmoidal-oblique seismic stacking pattern (Fig. 6c). The maximum thickness is 90 ms TWT, decreasing towards the southwest.

4.1.2 Plastered drifts

Along the upper continental slope, WNW-ESE trending plastered drift 1 is identified between 200 and 500 m. It encompasses an area of about 48 km^2 and has a length of about 7.5 km and a width of 1.5-2.5 km (Figs. 3, 5, 6d). The drifts are characterized by sigmoidal reflections draped over the upper continental slope and is about 10 to 90 ms TWT thick. At the southern flank of the Le Danois Bank, four E-W trending narrow plastered drifts are recognized at different water depths. Plastered drifts 2 (9 km long and 2.4 km wide, from 528 to 642 m) and 3 (9.3 km long and 0.8 km wide, from 715 to 782 m) are located at the western edge of the Le Danois Bank and are separated by a ridge (Figs. 5, 7a). The ridge (10 km long, 600 m wide) rises 20 m above the surrounding seafloor with a WNW-ESE orientation. Plastered drift 4 (from 870 to 1100 m) is positioned at the upper southeast flank of the bank and has a length of 10 km and a width of 0.8 to 1.1 km (Figs. 3, 7b, c). Plastered drift 5 (from 1350 to 1780 m), located at the lower southeast flank of the bank, is 7 km long and 1.8-3.9 km wide (Figs. 3, 5, 7d). Along the southeast foot of the Le Danois Bank, the SW-NE oriented plastered drift 6 is located at water depths between 1760 and 1920 m (Figs. 3, 5, 7e). This plastered drift covers an area of 56 km^2 and is about 18 km long and 1.8 km wide. The thickness ranges from 20 to 60 ms TWT. The internal structure is characterized by sigmoid reflection configurations and an unconformity surface with onlap and downlap terminations.

4.1.3 Moats and furrows

The Le Danois Moat parallels the southern foot of the Le Danois Bank between 866 and 1530 m water depth (Figs. 6a, b, c). It separates the Le Danois Drift from the bank along a WNW-ESE linear trend (Fig. 5). The Le Danois Moat is about 42 km long. It has a U-shaped profile with widths ranging from 1.9 to 2.2 km and depths from 20 to 80 m (Figs. 6a, c). Towards the west, the moat narrows to a minimum width of 400 m.

The NW-SE oriented Gijón Moat separates the Gijón Drift from the upper continental slope (Fig. 5). This moat (from 320 to 1090 m) starts at the southern foot of the Vizco High and extends towards the Lastres Canyon. It deflects to the southeast and loses expression to the east. The northwest part of the moat shows an asymmetric U-shape profile (Figs. 6a, b, d) and has a depth ranging from 25 to 110 m, a width of about 2.3 km. Towards the southeast, the moat narrows to a minimum width of 800 m and the depth reduces to 10 to 20 m.

The Asturias Moat is associated with the Asturias Drift (Fig. 5). It is oriented in a W-E direction in the western part and becomes NNW-SSE towards its NE limit. The moat is about 6 km long, 400 m wide, 80 m deep and displays a V-shape profile (Fig. 6c).

A NW-SE oriented furrow occurs at the southwest edge of the Gijón Drift (Fig. 5). It can be identified on the multibeam bathymetry (Figs. 2, 4a) and seismic reflection profiles (Fig. 6b) by its vertical incision, ranging from 10 to 20 m. The length is 4.1 km and the width is 0.8 km. It gradually loses its expression towards the southeast, as well as towards the Vizco High.

4.2 Downslope features

Four isolated small slide scars are identified along the southern flank of the Le Danois Bank (Fig. 5). They possess relatively steep slopes (values between 6° and 18°) compared to the surrounding seafloor (values between 1° and 4°) (Fig. 3). They are characterized by crescent shapes, lengths varying between 1.3 and 3.2 km, depths ranging between 20 and 80 m and south-dipping orientations. All of the slide scars are present between 544 and 1020 m water depth (Fig. 3). The slide scars located at plastered drift 4 (from 840 to 877 m, slope gradient of 12°) and central part of the southern flank of the bank (from 803 to 854 m, slope gradient of 15°) are namely documented by Figures 6c and 7a.

Along the southern rim of the top of the Le Danois Bank, one large isolated slide scar with a steep slope gradient of 15° to 20° has been identified (Figs. 3, 5, 7b, 8a). The headwall, between 565 and 681 m, is characterized by a linear geometry (Fig. 5). This W-E orientated slide scar is about 33 km long and has a mean depth of 90 m.

Along the southern flank of the Le Danois Moat, between 990 and 1120 m water depth, two slide scars are identified based on their arcuate morphology (Figs. 3, 5, 7b). Compared to the entire southern flank (mean slope gradient of 2.5°) of the Le Danois Moat, both slide scars are located on the steepest slopes (7° to 9.5° slope). The eastern scar (2.2 km long and 60 m deep) has a north-dipping trend and an average slope gradient of 8.2°. The western one (3.4 km long and 30 m deep) displays a lower slope gradient of 7.8° with a northeast-dipping trend. Their associated slide deposits are characterized by onlap and downlap terminations overlying an unconformity surface (Fig. 7b).

At the centre of the intraslope basin, a 2.7 km^2 isolated ellipse-shaped slide scar is shown on the bathymetric and seismic data (Figs. 3, 5, 8b). This well-developed seafloor scarp shows a higher slope gradient (3.5°) compared to the surrounding slopes (1.5°). It has a northeast-dipping trend and gradually loses its expression from southwest to northeast (Figs. 3, 5). The headwall, located between 750 to 777 m water depth, is about 2.5 km long and displays truncated seismic reflection configurations in the seismic records. At the bottom of this slide scar, chaotic-transparent seismic reflections are present (Fig. 8b).

A series of isolated slide scars have been identified southeast of the Vizco High between 870 and 1280 m water depth (Figs. 5, 8c). They are positioned on an erosive unconformity surface and display a step-like pattern with at least 5 levels (at 1070, 1010, 960, 940 and 920 m) (Fig. 8c). These slide scars have an east-dipping trend and gradually lose their expressions from the west to the east. The seismic facies of these features are truncated reflections at their headwalls and chaotic-transparent reflections at the bottom. They have small variations in shape (arcuate), size (2.8 to 3.2 km in diameter, 20 to 60 m deep) and orientation (N-S) (Fig. 5). The slope gradient of the headwalls ranges from 7° to 13° (Fig. 3).

A large area (233 km^2) of mass-transport deposit is located between the Le Danois Drift, the Gijón Drift and the Lastres Canyon (Fig. 5). The identification is based on a distinctive seafloor irregularity and discontinuous low-amplitude chaotic and transparent reflections in the seismic records (Fig. 8d). The boundary of these mass-transport deposits has a relatively sharp slope gradient (about 4.5°) compared to the surrounding area (about 0.8°) (Fig. 3).

4.3 Topographic irregularities and morphological depressions

At the northeast edge of the Gijón Moat, an ellipse-shaped morphological depression (about 20 m deep) is identified between 680 and 770 m water depth (Fig. 8e). It is 5.5 km in diameter and 30 to 80 m deep with a NW-SE orientation. The scarp is characterized by an erosive unconformity surface and truncated reflections. Undulating wavy-oblique reflections make up the succession between the seafloor and the unconformity surface in the seismic records. The wave-crest features are about 1-2 km long, 200 m wide and rise 10 m above the surrounding areas with a trend of upslope migration. The wave ridges have a NE-SW orientation.

At the centre of the intraslope basin, a remarkable asymmetric ear-shaped depression is identified between 990 and 1080 m water depth (Figs. 5, 8f). It is 6 km in diameter and has a northwest-dipping trend. The upslope facing flank is 20 to 90 m deep with a slope gradient of about 18° (Fig. 3). It is characterized by truncation of parallel-stratified reflections (Fig. 8f). This depression delimits an oblique terrace (2.3° slope) with wavy morphology along the downslope facing flank. The related terrace displays high-amplitude chaotic reflections overlain by continuous wavy-stratified reflections (Fig. 8f). These wave-like features are about 2.5-3.5 km long, 400 m wide and rise 30 m above the surrounding seafloor. The wave ridges have a N-S orientation.

Several topographic irregularities and morphological depressions (from 1540 to 2100 m) occur between the Le Danois Drift, plastered drift 6 and the Lastres Canyon. The identification is based on their irregular shapes (circular, crescent and ellipse-shaped), sizes (0.8 to 3.7 km in diameter, 40 to 80 m deep), slope gradients (values between 15° and 22°) and orientations (WSW-ENE to NW-SW trending) (Figs. 3, 5). Only a few seismic profiles document these depressions. They are characterized by truncated, chaotic, semi-transparent or wavy-stratified reflections (Fig. 7e).

5 Discussion

5.1 Present-day bottom-current implications on contourite features

The present-day bottom-current circulation within the Le Danois intraslope basin is dominated by three water masses: the ENACW, the AMW and the LSW. Due to their different physical characteristics and circulation patterns, they may play a significant role in shaping the present-day depositional and erosional contourite features. The combination of the CTD data (World Ocean Database, 2013) and the interpreted seismic profiles (Figs. 6, 7, 8) allow gaining more insights regarding the interaction between each water mass and its impact on the local seabed (contouritic) processes.

5.1.1 ENACW related processes

The ENACW (flowing between 200 and 570 meter water depth) mainly interacts with the upper continental slope and the upper southern flank of the Le Danois Bank, where plastered drifts 1 and 2, 3 are respectively located (Figs. 9a, b, c). Present-day bottom currents along the southern flank of the bank in that depth interval are approximately 15 cm/s (González-Pola et al., 2012). Along the Asturias continental slope, the ENACW generally has a mean velocity of 10-30 cm/s (González-Pola et al., 2012). All of these values meet the conditions (Stow et al., 2002, 10-30 cm/s) for generating plastered geometry of contourite drifts 1, 2, 3, and related bottom currents are most likely resulted from the ENACW circulation documented by González-Pola et al. (2012) (Fig. 10). Additionally, plastered drifts 2 and 3 are distributed along the northern and southern foot of a ridge (Fig. 7a) with similar orientations, shapes and lengths. These features illustrate similar bottom-current processes resulting from interactions between small flow filaments and the adjacent topographic obstacle. The related bottom-current dynamics could be compared with the present-day anticyclonic circulation cell along the western rim of the Le Danois Bank (González-Pola et al., 2012) (Fig. 10).

Gentle slope morphology is one of the main elements responsible for the formation of plastered drifts as well (Laberg and Camerlenghi, 2008). At the eastern edge of plastered drifts 2 and 3, the slope gradient abruptly increases to 8.5°. This steep slope is maintained to the southeast flank of the bank (Fig. 3). Higher slope gradients could accelerate bottom currents, in turn shifting sedimentary processes from deposition to non-deposition/erosion (Rebesco et al., 2014). These higher slope angles could inhibit the generation of plastered drifts along this part of the Le Danois Bank. Towards the eastern boundary of plastered drift 1, the slope gradient maintains but the space between the lower continental slope and the Asturias continental shelf is widened (Fig. 3). The wide morphology will decelerate bottom currents (Faugères and Stow, 2008). Slower flows could limit the lateral extension of plastered drift 1 towards the east.

5.1.2 AMW related processes

The AMW mainly interacts with the southern flank of the Le Danois Bank and the intraslope basin between 750 and 1500 m water depth (Fig. 9). Plastered drift 4 (from 870 to 1100 m) is positioned along the upper southeast flank of the Le Danois Bank (Figs. 7b, 9d), where bottom currents are estimated at 10-

15 cm/s (González-Pola et al., 2012). This drift covers a gentle slope (mean slope gradient of about 3.6°) (Fig. 3) and is shaped by the AMW (Fig. 10). Along the entire boundary of plastered drift 4, the slope gradients abruptly increase to 8-15° (Fig. 3), delineating its extent. Considering all, slope morphology is the main controlling factor for the spatial distribution of plastered drift 4.

The Le Danois Drift (from 790 to 1080 m) is suggested to be generated by the AMW as well. The associated Le Danois moat indicates a focused flow pathway of bottom currents all along the southern foot of the Le Danois Bank. This feature matches with the present-day oceanographic data documented by González-Pola et al., (2012). A 12-month long mooring record (at 44°02.33', N 4°49.33' W) indicates a persistent westward near-bottom flow along the southern foot of the bank at the AMW level (González-Pola et al., 2012). Direct current measurements at this level, consisting of few snapshots made by landers, fit with the velocity in the range of 10-25 cm/s (González-Pola et al., 2012). This velocity is sufficient to generate associated contourite drifts (Stow et al., 2002, 2009). Towards its western limit, the Le Danois Drift extends to an open area between the lower continental slope and the intraslope basin (Fig. 5). In this area, the Le Danois bank is absent and bottom currents most likely will drop to the reported background values, which are less than 5 cm/s (Iorga and Lozier, 1999a). Towards the eastern limit of this drift, the presence of a group of morphological depressions limits the distribution of the Le Danois Drift (Fig. 5). Consequently, slower bottom currents and seabed morphology limit the distribution of the Le Danois Drift. The Le Danois Drift and plastered drift 4 have similar orientations and are located at the same water depths (Figs. 5, 10). The distribution suggests that the associated bottom-current processes of both drifts are related to the same current (Fig. 10). The presence of two types of contourite drifts along the AMW pathway can be explained by changes in current velocities (Faugères and Stow, 2008). This implies accelerated and then decelerated processes from the southeast towards the southwest flank of the Le Danois Bank.

The Gijón Drift and the Moat extend from 320 to 1090 m water depth. A CDS is usually associated with one water mass (Rebesco et al., 2014). However, at the present day, the Gijón Drift is situated within the boundaries of two different water masses, being the ENACW and the AMW (Figs. 9a, b, c). Along the Cantabrian continental upper slope, the ENACW does not reach to 1000 m for prolonged periods of time (McCartney and Mauritzen, 2001), while the AMW could only reach up to 400 m during interglacial climate cycles (Zhang et al., 2016; Kaboth et al., 2016, 2017). Additionally, shape and morphology of the

Gijón Drift display different features in different depth intervals. Within the ENACW, the Gijón Drift has elongated and mounded geometry (Figs. 9b, c), whereas the part within the AMW is confined and mounded (Fig. 9a). Different shapes of contourite drifts are related to different bottom-current conditions (Faugères and Stow, 2008). As such, it is possible that the ENACW and the AMW interacted with the upper continental slope during different climate intervals and both are responsible for the Gijón Drift.

The spatial variation of the Gijón Drift could be controlled by the slope morphology as well. On the multibeam bathymetry, the Gijón Moat (NE-SW trending) does not fit the alongslope distribution compared to the Asturias continental slope (WNW-ESE trending) (Figs. 4a, 5). The obliquity may be caused by the presence of the Gijón Canyon (NE-SW trending) (Figs. 5, 9f). Interactions between canyon channels and bottom currents are possible to provoke streamline distortions and accelerate current flows (Holland, 1972). The related distortion and acceleration enable bottom currents flowing upwards along canyon channels (Jackson et al., 2006). When the AMW encounters the Gijón Canyon, bottom currents could follow the canyon morphology (Allen and De Madron, 2009; Muench et al., 2009) and arise towards the shallower continental slope (320 m water depth). The variation of the Gijón Drift could also link with the interaction between the Gijón Canyon and the AMW.

The Asturias Drift lies around a buried structural high at the centre of the intraslope basin (Figs. 5, 6c, 9c). The presence of a topographical obstacle can create faster currents, in turn generating contourite depositional and erosional features (Ercilla et al., 2016; García et al., 2016). The associated Asturias Moat indicates the pathway of bottom currents. The W-E to NNW-SSE deflection of the moat may result from the morphological control of the associated depression (Fig. 10). As such, the Asturias Drift and Moat result from interactions of the AMW with a buried structural high and a morphological depression. Related bottom-current processes could be linked with the present-day AMW circulation cell (González-Pola et al., 2012) in the intraslope basin (Fig. 10).

At the southeast edge of a group of slide scars, a furrow is located in the Gijón Drift at water depths of 844 to 870 m (Figs. 5, 6b). Present-day bottom currents associated with the AMW in the intraslope basin (from 10 to 20 cm/s, González-Pola et al., 2012) do not have sufficient energy to create erosive furrows, which require at least 30 cm/s (Stow et al., 2009). However, it is well documented that the presence of topographic obstacles or irregularities can cause locally intensified bottom currents (Rebesco et al., 2008, 2014). Examples include the Galicia Bank region where (numerically modelled) intensifications from 7 cm/s up to 35 cm/s are recorded (Zhang et al., 2016), the Porcupine Abyssal Plain where accelerations reach 15 cm/s (Turnewitsch et al., 2004), and the Xisha Trough where (also numerically modelled) enhancements from 15 cm/s up to 30 cm/s are documented (Chen et al., 2016). Similarly, bottom currents can recirculate around the Vizco High, possibly speeding up from 10-20 cm/s to over 30 cm/s in this region. The presence of the furrow indicates faster bottom currents resulted from strong interactions with the Vizco High.

5.1.3 LSW related processes

The LSW is only present between the Lastres Canyon and the lower southern flank of the Le Danois Bank in the study area (below 1750 m water depth) (Figs. 4f, 9). Plastered drifts 5 is located along the lower southern flank (2.4° slope) while plastered drift 6 lies along the southeastern foot (1.2° slope) of the bank (Figs. 3, 5, 9e). Both lay within the depth windows of the LSW. Along the Cantabrian continental slope, the bottom current velocities (2-6 cm/s; Speer et al., 1999; Friocourt et al., 2007) are below the threshold for depositing plastered drifts (Stow et al., 2002, 2009). Consequently, the presence of these two plastered drifts indicates local intensification. Since there are no known measurements of current velocities at this depth interval in the study area, tentative bottom-current velocities generating these plastered drifts have to be estimated in order to better understand the acceleration of bottom currents in a topographically constrained basin. Since the average threshold for deposition of plastered drifts has been indicated at 10 cm/s (Stow et al., 2009), we thus estimate the local current velocities must exceed this value. Along the entire boundary of plastered drift 5, slope gradients increase to 6°-10° (Figs. 3, 5). As such, steeper slope morphologies control and limit the distribution of plastered drift 5. The extension of plastered drift 6 is limited by seafloor depressions positioned between the bank and the Lastres Canyon (Fig. 5). Towards the northeast boundary of plastered drift 6, the presence of a topographic high locally creates slopes from 1° to 11° (Fig. 3). These slopes limit the distribution of plastered drift 6.

5.2 Interaction with slope instability processes

In the study area, plenty of slide scars occur at the surfaces of contourite drifts and along the southern flank of the Le Danois Bank (Figs. 2, 3, 5). The main causal factors for submarine slides include high slope angles, seismic activity, volcanic activity, gas charging and rapid sediment accumulation (Locat and Lee, 2002; Sultan et al., 2004; Verdicchio and Trincardi, 2008; Miramontes et al., 2016; Rashid et al., 2017). Since no seismic or volcanic activity and gassy sediments have been observed in the study area, the triggering mechanisms for these slide scars are either rapid sediment accumulation or oversteeping.

5.2.1 Increased sediment accumulation

An isolated slide scar is positioned at the northeast part (3° slope) of the Gijón Drift (Figs. 2, 3, 8b). The sedimentation rate of the Gijón Drift is unknown. However, mounded drifts are known to have relatively high sedimentation rates (5-60 cm/ka) compared to pelagic (<2 cm/ka) and hemipelagic (5-15 cm/ka) sediments (Stow et al., 2008). During the build-up of the Gijón Drift, high sedimentation rates are capable of decreasing shear strength of sediments, favouring the formation of mass-wasting events (Baeten et al., 2013). Since the slide scar lies between the Asturias Drift and the Gijón Drift (Fig. 5), two drift systems can deliver sediments to this location and drastically increase the sedimentation rate. As such, the occurrence of a slope instability event at this location is more likely, compared to other areas of the Gijón Drift.

Opposed to the Gijón Drift, no slope instability processes occur at the surface of the Le Danois Drift (Figs. 2, 3, 5). The difference between both drifts may be related to their internal depositional structures. Interbedded continuous sigmoidal reflections and discontinuous low-amplitude chaotic reflections within the Gijón Drift are interpreted as interbedded mass-transport and contourites deposits (Figs. 6a, d), whereas only contourites are present within the Le Danois Drift (Figs. 6a, c). These interbedded features are mainly located at the Gijón Moat, indicating multiple shifts between alongslope and downslope processes over the Gijón Drift. Different degree of sedimentary sorting, resulting from distinct sedimentary processes, could lead to the deposition of different sedimentation layers (Wilson et al., 2004). These sedimentation layers generally have higher sensitivity, due to their distinct physical properties (Laberg and Camerlenghi, 2008). Higher sensitive layers can effectively reduce shear resistance strength of sedimentary bodies and thus a dynamic slide process can initiate (Kvalstad et al., 2005). Similar examples have been documented in the vicinity of the Storegga Slide region (glacigenic/contouritic deposits), offshore Norway (Bryn et al., 2005) and in the Afen Slide area (mud/sand contouritic deposits), offshore UK (Wilson et al., 2004). Within the Gijón Drift, multiple shifts between alongslope and downslope processes create high sensitive layers and induce mass movements. Conversely, the shear strength of the Le Danois Drift is not low enough to trigger slope instability events.

Between the northwest end of the Gijón Drift and the Vizco High, a series of slide scars overlay a steep slope (10°) (Figs. 2, 3, 8c). Thick contourite deposits of the Gijón Drift, as well as steep slope morphology could have a significant potential for producing mass movements.

5.2.2 Scouring of active bottom currents

Two slide scars occur at the southern flank of the Le Danois Moat (Figs. 2, 3, 5). Both slide scars are distributed along the steepest (9°) parts of the moat (Fig. 3). The focused flow within the Le Danois Moat could be held responsible for the formation of these slide scars. Similar to the Cape Basin, where a slide scar is induced by scouring of bottom currents (Weigelt and Uenzelmann-Neben, 2004), higher angles and active bottom currents can provide favourable conditions for slope instability processes in this part of the moat. The associated slide deposits display internal structures with onlap and downlap terminations and their geometry resemble patched drift features (Fig. 7b), hinting towards reworking of bottom currents. The observed features in the study area are similar to recent discoveries in the Guadalquivir Bank, where reworking of bottom currents reshaped the slide deposits (García et al., 2016). In conclusion, focused bottom currents within the Le Danois Moat locally scour the steeper flank, this may lead to undercutting of the slope, triggering mass movements.

5.2.3 Mixed processes

In the Le Danois Bank region, a large area of mass-transport deposits lies between the Lastres Canyon, the Gijón and the Le Danois Drifts (Fig. 5). The associated scarp is about 10 km away from the Lastres canyon channel (Figs. 2, 3). The seafloor morphology of the mass-transport deposits shows an upslope trending from the canyon wall towards the scarp. As such, turbidity currents associated with the Lastres canyon are not possible to penetrate (10 km) upslope undercutting the scarp, and are most likely not responsible for the formation of these mass-transport deposits. The slope gradient (9.5°) of this location is much higher than the surrounding seafloor (2.8°) (Fig. 3). High slope angles lead to downslope processes at this particular location. Towards the east, the orientation of this mass-transport deposits area gradually changes from downslope (N-S trending) to alongslope (W-E trending) directions (Fig. 5). This deflection indicates the interaction between mass movements and AMW currents (Fig. 10). Characteristics of these mass-transport deposits can be compared to those at the southeast Grand Banks (Rashid et al., 2017) and along the Mid-Norwegian Margin (Bryn et al., 2005), where the area of mass-transport deposits is elongated and enlarged and the related distribution and location are highly influenced by bottom currents. As such, a high slope gradient could be predisposing factors for the formation of these mass-transport deposits and bottom currents resulting from the AMW highly controls their distribution.

5.3 A genetic model for the sediment waves of the Le Danois Bank region

Within the intraslope basin, wavy features are identified in the seismic records (Figs. 7e, 8e, f). These features are interpreted as sediment wave fields (respectively A, B and C on Fig. 5) based on the traceable and continuous seismic reflections, as opposed to slope failure deposits with sharp and acoustically incoherent reflections (Gardner et al., 1999; Lee et al., 2002; Mosher and Thomson, 2002; Wynn and Stow, 2002). The formation of sediment waves has three possible causes, being turbidity currents, bottom currents or internal waves (Ercilla et al., 2002; Ribó et al., 2016). Turbidity-current related sediment waves generally occur on the back-slopes of channel levees and in turbidity-current channels (Wynn and Masson, 2008). Since the study area with sediment waves lacks the presence of channel-levee systems and turbidity-current processes, these sediment wave fields most likely result from bottom currents or internal waves.

Sediment wave field A is located within the AMW level and is induced by internal waves. Bottom currents can be ruled out as the moat in the vicinity of sediment wave field A suggests N-S oriented bottom currents (Fig. 10), which is parallel to the wave crests. This orientation does not fit the bottom-current induced sediment waves (oblique or perpendicular orientations of the wave crests; Masson et al., 2002; Wynn and Masson, 2008).

Sediment waves created by internal waves are documented in several regions (Pomar et al., 2012; Delivet et al., 2016; Ribó et al., 2016). When internal waves propagate down a sloping bottom, three possible reflection conditions exist based on the propagation angle of the internal wave (c) and the bottom slope angle (γ) (Cacchione et al., 2002). The relationship was established by Cacchione and Wunsch (2006) and can be written as:

$$c = arctan\left[\left(\frac{\sigma^2 - f^2}{N^2 - \sigma^2}\right)^{\frac{1}{2}}\right]$$

where σ is the internal wave frequency, f is the local Coriolis (inertial) frequency, and N is the Brunt–Väisälä (buoyancy) frequency. Different values of γ/c correspond to subcritical ($\gamma/c < 1$), critical ($\gamma/c \approx 1$) and supercritical ($\gamma/c > 1$) reflection conditions (Lamb, 2014). In the study area, high-frequency internal waves have been observed near the centre of the intraslope basin at the level of the AMW (González-Pola et al., 2012). Based on the local flow properties, González-Pola et al. (2012) proposed the following parameters: f as 1.013 x 10^{-4} /s, σ as 0.09 cph and N as 1.5 to 2.5 x 10^{-3} s^{-1}. Thus, internal wave reflection condition in sediment wave field A (1.6° slope) is subcritical. Subcritical conditions of internal waves are generally characterized by smaller wave heights and upslope migration (Lamb, 2014; Delivet et al., 2016; Ribó et al., 2016). And this is true for sediment wave field A, which migrates upslope and their wave heights is relatively small (10 m) compared to those in the Argentine Basin (30 m) (von Lom-Keil et al., 2002), the Gulf of Valencia (50 m) (Ribó et al., 2016) and the Bahama Outer Ridge (60 m) (Flood and Giosan, 2002).

Sediment wave fields B (Fig. 9f) and C (Fig. 9g) are respectively located at the ENACW/AMW and the AMW/LSW interfaces. No mooring measurements have been performed at these interfaces and the present-day oceanographic dynamics are unknown. However, the mean salinity values at the ENACW/AMW (35.6) and the AMW/LSW interfaces (35.2) (Fig. 9) indicate mixing processes as the ENACW has a mean salinity of 35.5, the AMW 35.8 and the LSW 35.1 (Fig. 4b). Turbulent mixing at the interface between water masses generally produces relatively high energetic currents associated with internal waves, which are capable of transporting and depositing sediments (Preu et al., 2013; Ercilla et al., 2016; Juan et al., 2016). Based on the local flow properties (González-Pola et al., 2012), internal wave reflection conditions in sediment wave fields B (1.6° slope) and C (1.7° slope) are subcritical. The relatively small wave heights (7 and 10 m for wave fields B and C) and the observed upslope migration both indicate a subcritical reflection parameter (Lamb, 2014). As such, sediment wave fields B and C are most likely related to internal waves as well. Both sediment wave fields are located within morphological depressions (slopes of 8°-10°) (Figs. 3, 5) and overlie irregular unconformity surfaces (Figs. 7d, 8c), which are the prerequisites for the formation of sediment waves (Aghsaee et al., 2010).

5.4 Sediment sources

The sediments constituting the various drifts in the region are mainly transported by downslope processes to the continental slopes and caught by bottom currents from adjacent areas towards the Le Danois Bank region (Gómez-Ballesteros et al., 2014). One year long current meter data, obtained from the Avilés Canyon (west of the Le Danois area, Fig. 1a) indicates direct delivery of river-sourced (the Narcea and Nalon Rivers) material into the canyon and its adjacent continental slope (Rumín-Caparrós et al., 2013). Frequent severe storms and repeated cycles of semidiurnal tides regionally enhanced the bottom currents, assuring a permanent amount of sediment in suspension in the canyon region (Rumín-Caparrós et al., 2016). Additionally, the Narcea and Nalon Rivers are generated before the formation period (the Neogene, Gómez-Ballesteros et al., 2014) of the Avilés canyon (Fernández-Viejo et al., 2014). The suspended material is possible to be picked up by eastward moving water masses towards the Le Danois Bank since the intensification of the AMW (during the late Pliocene, Hernández-Molina et al., 2014) and the establishment of the Iberian Poleward Current (during the late Pleistocene, Mena et al., 2018). Consequently, the Narcea and Nalon Rivers could provide sediments for identified contourite deposits for prolonged periods of time (Gómez-Ballesteros et al., 2014; Rumín-Caparrós et al., 2016).

The Avilés Canyon is not the only sediment source for the contourite deposits. Current meter data reveal a long-term (12-month) persistent westward flow within the Le Danois Moat (González-Pola et al., 2012), which has an opposite direction compared to the major branch of the AMW (Fig. 10). Additionally, the LSW flows towards the intraslope basin in a westward direction as well (Fig. 10). These observations suggest an additional eastern sediment source should be present. The Torrelavega Canyon is located in the vicinity of the study area and fed by sediments from the Besaya River (Fig. 1a) (Caballero et al., 2014). Eddies extending from the surface down to 3500 m water depth have been observed above the Torrelavega canyon (Caballero et al., 2014), which remained stationary for a long time (locally up to 7 months) (Pingree and Le Cann, 1992; Caballero et al., 2014), creating the possibility for associated energetic current patterns to erode the seafloor and suspend sediments (Shanmugam, 2013). As such, suspended sediments can be transported towards the Le Danois Bank area by bottom currents associated with the AMW and the LSW.

The Gijón and the Lastres Canyons can be possible sediment sources as well. Few studies have documented these two canyon systems. Based on their locations, the Sella River is suggested to be the sediment supply for both canyons (Fig. 1a). Erosive features at the canyon walls have been displayed in the seismic records (Fig. 9f). Due to the consistent AMW flowing above the intraslope basin, eroded sediments can contribute to the sediment accumulation of the prevailing contourite drifts. Finally, mass-transport deposits are interbedded within the Gijón Drift (Figs. 6a, d) and indicate the interaction between bottom currents and gravitational processes along the upper continental slope. These processes may add sediments from the Asturias continental shelf to the contourite drifts.

6 Conclusion

The spatial variability of contourite drifts, slope morphology and small-scale bottom-current processes are all tied together in this topographically constrained small basin. Bottom currents associated with the ENACW, the AMW and the LSW are more focused and strongly intensified (estimated acceleration up to 25 cm/s) due to the morphological constraint of the Le Danois Bank and the Vizco High. Changes in velocities of bottom currents and slope gradients result in variations in types, shapes and spatial distributions of contourite drifts, especially along the current pathways of the AMW. Unlike typical elongated and mounded drifts, the Gijón Drift exhibits a lateral variation (evolving from elongated to confined geometry) at different depth intervals, which evidently indicate different bottom-current dynamics associated with one contourite drift, suggesting specific geometries and shapes of contourite drifts in topographically constrained small basins. Despite the influence of large topographic obstacles, small steep (slopes of 8°-10°) morphologic irregularities may interact with internal waves, generating small-scale (wave heights 7-10 m) upslope-migrated sediment waves in the intraslope basin. Enhanced bottom-current processes interplay with internal waves and slope instability events, and determine the complex morphology of the present-day seafloor. Compared with large contourite features resulted from deep water masses, small basin-scale contourite features of the Le Danois Bank region is an exquisite example of multiple processes interacting in a topographically constrained basin at an intermediate water depth. The spatial variability of the contourite features is more frequent than anticipated so far and can have far-reaching implications for comparable topographic and oceanographic settings all over the world.

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Figures

Fig. 1. Location of the study area: A) Position with respect to the North Iberian continental margin (Ercilla et al., 2008), contour lines every 500 m. Location of the rivers is based on Prego et al. (2008); B) Available geophysical datasets for this study. The present-day oceanographic circulation pattern is modified from González-Pola et al. (2012). The main morphological expressions are shown (contour lines every 100 m). ENACW = Eastern North Atlantic Central Water; AMW = Atlantic Mediterranean Water; LSW = Labrador Sea Water.

Fig. 2. Multibeam bathymetry map of the study area, with the location of displayed seismic profiles.

Fig. 3. Slope map (in degrees) of the study area. The identified contourite deposits (outlined in red) and slide scars are indicated.

Fig. 4. (a) 3D colour shaded relief multibeam high-resolution bathymetric map of the study area (ECOMARG project, IEO), with the location of oceanographic cross-sections A-A', B-B' and C-C (World Ocean Database, 2013). (b) Potential temperature versus salinity diagram from the water masses in the study area (World Ocean Database, 2013). CTD stations (red triangles) and their respective locations are indicated in cross-section profiles (c, d, e, f). Oceanographic cross-sections from the Le Danois Bank towards the Cantabrian continental margin. The water column colour ranges indicate salinity (c, e, f) and temperature (d).

Fig. 5. Morphosedimentary map of the Le Danois Bank region, based on the interpretation of multibeam bathymetry and seismic profiles. Numbers (1-6) and letters (A, B, C) respectively denote plastered drifts and sediment wave fields.

Fig. 6. Interpreted sparker seismic profiles showing the morphological features of elongated mounded and separated drifts and a plastered drift. Onlap and downlap terminations (red arrows) are indicated. The location the seismic lines is indicated on the multibeam bathymetry map (Fig. 2).

Fig. 7. Interpreted sparker (a, b), TOPAS (c, d) and airgun (e) seismic profiles showing the morphological features of plastered drifts and slide scars. Onlap and downlap terminations (red arrows) are indicated. The location of the seismic lines is indicated on the multibeam bathymetry map (Fig. 2).

Fig. 8. Interpreted sparker (a, b, d, e) and airgun (c, f) seismic profiles showing the morphological features of slide scars, mass-transport deposits and sediment waves. The location of the seismic lines is indicated on the multibeam bathymetry map (Fig. 2).

Fig. 9. Seismic and oceanographic profiles in the Le Danois Bank region. The depth intervals of the water masses and contourite features show the dynamic interaction between water masses and the present-day sedimentary regime. The locations of airgun seismic profiles are indicated in the bathymetric map. The junctions of Line a, b and f; and Line e and g (dotted black lines) are indicated in the seismic profiles. The CTD stations (red triangles) and their respective locations are indicated in the seismic and oceanographic profiles.

Fig. 10. Sketch of the recent sedimentary processes within the Le Danois Bank region. This sketch has been produced based on the morphosedimentary features defined in the morphosedimentary map (Fig. 5).

Highlights:

- Present-day morphology of the Le Danois Bank region is shaped by alongslope, downslope and mixed processes.
- Due to the morphological constraint of the Le Danois Bank and the Vizco High, bottom currents are more focused and strongly intensified (estimated acceleration up to 25 cm/s).
- Contourite drifts are suggested to have specific geometries and shapes in topographically constrained small basins.
- More frequent lateral variations are typical features for contourite drifts generating in topographically constrained small basins.