Marine Geology 340 (2013) 49–56

Contents lists available at SciVerse ScienceDirect

Marine Geology

journal homepage: www.elsevier.com/locate/margeo

Morphological evolution of Cap Lopez Canyon (Gabon): Illustration of lateral migration processes of a submarine canyon

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ARTICLE INFO

Article history:
Received 3 September 2012
Received in revised form 3 April 2013
Accepted 15 April 2013
Available online 2 May 2013

Communicated by D.J.W. Piper

Keywords:
Gabon
Cap Lopez Canyon
Canyon head
Slope failures
Lateral migration
Meander development

ABSTRACT

Comparison of bathymetric data on short time scales (1959–2008) is used to provide new insights into the modern sedimentary dynamics of the Cap Lopez Canyon (Gabon, West Africa). The canyon head evolution is characterized by a north–eastward lateral migration of ~180 m between 1959 and 2008. The evolution of the coastline position over the same period highlights the strong relationship between the probable increase of the longshore transport, related to the smoothing of the coast and the morphological evolution of the canyon head. Lateral migration of the thalweg from the inner bend to the outer bend is estimated between 77 and 190 m (or between 1.6 and 3.8 m/year), leading to an increase of the sinuosity (from 1–1.5 in 1959 to 1.3–3.6 in 2008). The migration of the thalweg is associated with strong erosion on the outer bends (up to 60 m) and sediment deposition on the inner bends (up to 25 m). Both the absence of overbank deposits and the developing point-bar morphology may reflect that equilibrium flows are the most frequent flows encountered in the Cap Lopez Canyon. Although erosion of the outer bends is the result of slope failures and steady erosional processes, our study suggests that erosion related to the transit of sediment gravity flows would be the predominant process.

1. Introduction

Coastal submarine canyons incising a continental shelf are of particular importance to the continent-ocean sediment budget because they facilitate the transfer of sediment from terrestrial and coastal sources to the deep sea. In places where the longshore drift is captured by canyons, a large volume of sediments can be evacuated (Shepard and Dill, 1966; Burke, 1972; Paull et al., 2005; Covault et al., 2007; Smith et al., 2007). Sand, transported by wave-induced currents in the nearshore zone, is temporarily trapped in the head of the canyon, until it is transported suddenly down canyon (Mastbergen and Van Den Berg, 2003). In these types of environment, processes such as slope failures, channel erosion and deposition are significant.

Such sedimentary processes that shape submarine canyons may occur on very short time scales, as suggested by recent works on Monterey Canyon. Monitoring activities revealed the occurrence of numerous subannual transport events (Paull et al., 2003; Xu et al., 2004). Bathymetric comparison between 2002 and 2005 illustrated several significant morphological evolutions: active movement of crescent-shaped bedforms (Smith et al., 2005, 2007; Paull et al., 2010), oscillation of canyon-head rim position (Smith et al., 2007), high erosion on outer bends (Smith et al., 2007) or headward erosion of gullies (Smith et al., 2005).

The main objective of this study is to carry on those previous investigations and to bring complementary results about the sedimentary dynamics of submarine canyons on a short-time scale. Based on annual to multi-decadal bathymetric comparison of Cap Lopez Canyon (Gabon, West Africa), our work illustrates different examples of morphological evolution and evaluates the frequency of involved sedimentary processes. This study focuses more particularly on the interactions between the canyon head and the coastline and on the evolution of the thalweg position through time. Finally, we discuss the possible triggering mechanisms and the magnitude of flow events in the canyon.

2. Study area

The onshore part of the Gabon continental margin is drained by the Ogooué River which constitutes the third largest African fresh-water source into the Gulf of Guinea (Mahé et al., 1990). The river flows west–north–westward and discharges south of Port-Gentil, forming a large delta on both sides of Mandji Island (Fig. 1A). The Gabonese coast is essentially exposed throughout the year to southwesterly swells (Bourgoin et al., 1963; Actimar, 2004). The waves strike obliquely the coast and induce a northward longshore transport, first estimated between 300,000 and 400,000 m^3/year (Bourgoin et al., 1963, Fig. 1B). The

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http://dx.doi.org/10.1016/j.margeo.2013.04.014

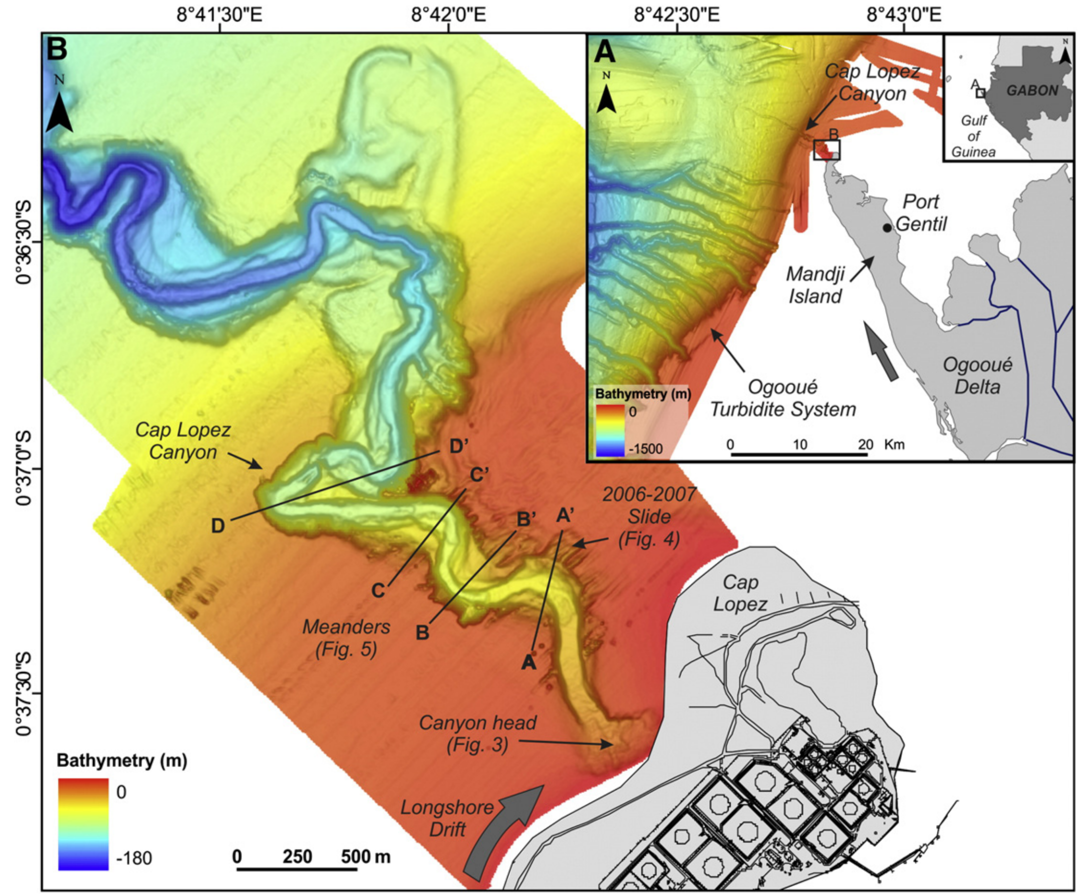

Fig. 1. Location of the study area. A) General bathymetric map showing the Ogooué turbidite system, Mandji Island and the Cap Lopez Canyon. Inset indicates the location of the Mandji Island on the Gabonese margin. B) Detailed bathymetric map of the upper part of the Cap Lopez Canyon obtained from multibeam echo-sounding. A–A', B–B', C–C' and D–D' are the locations of the bathymetric sections through the canyon meanders (Fig. 4). Grey arrows indicate the direction of longshore drift in the main figure and inset.

longshore transport is responsible for the formation of the Mandji Island, a sandy spit of 50 km long and 1–7 km wide located at the northern end of the Ogooué Delta (Fig. 1A & 1B). Sediment samples collected along the west coast of the Mandji Island are composed of well sorted fine to medium sand. Particles <63 μm are rare (<3%). Fine sedimentation is more present in the bay of Port Gentil (up to 20% of silty clay sediment; Artelia, 2012a).

Based on directional wave statistics and numerical simulations, recent work on the study area provided more detailed information about longshore transport (Artelia, 2012b). First, offshore wave data were obtained from the WANE2 wave database between 1999 and 2006 at the geographic point no. 25918, located 90 km south–westward from Cap Lopez (900 m water depth, Fig. 2). A second dataset, made up of in-situ measurements near Cap Lopez between 10 and 45 m water depth, was used to complete wave observations on the coast. This dataset covers short period (up to 4 months) between 2008 and 2010. Statistical analysis of offshore wave conditions shows a predominant direction between N190° and N200° (85%, Fig. 2). 70% of wave conditions are characterized by a peak period between 9 and 13 s and 17% show a peak period greater than 14 s. More than 90% of significant wave height ranges between 0.75 m and 2 m. 100-Year extreme sea waves simulated on a study area located ~20 km south of Cap Lopez at 12 m water depth show a significant wave height and a peak period around respectively 3.8 m and 15.8 s (Actimar, 2004). A wave refraction model of the dominant swells was calculated by wave propagation on the coast (Artelia, 2012b). Different results were highlighted: 1) swells tend to rotate around the Cap Lopez and especially as the peak period increases, 2) because of bathymetric effects, swells are gradually attenuated northward, 3) the Cap Lopez Canyon favours the local strengthening of swells south of the canyon and their attenuation in and just north of the canyon. Dominant wave statistics and simplified wave rays from wave refraction model are illustrated in Fig. 2.

Using the wave propagation model, the longshore sediment transport along the Mandji Island was calculated with the CERC formula (Artelia, 2012c, Fig. 2). South of Cap Lopez, longshore sediment transport is estimated between 670,000 and 1,120,000 m^3/year (Fig. 2). Refraction effect on the swells then increases northward, leading to a lower longshore transport compared to the south (mean 630,000–675,000 m^3/year). In the vicinity of the canyon head, transport efficiency becomes considerably weak (60,000–100,000 m^3/year), indicating that the canyon head capture a major part of the longshore transport. Finally, north of the canyon head, the change of orientation of the coast conduces to a local intensification of the longshore drift (90,000–150,000 m^3/year). Longshore transport estimates along the coast are synthetized in Fig. 2.

The strong sediment accumulation at the extremity of the Mandji Island is thought to have led to the formation of the Cap Lopez Canyon by retrogressive erosion (Le Fournier, 1972). The deep incision of the continental shelf by the canyon head (Fig. 1) combined with longshore drift favours the capture of sandy sediment by the canyon (Reyre, 1984).

3. Data acquisition and analysis

Annual oceanographic surveys on the Cap Lopez Canyon were performed by the IOTA SURVEY Company between 2004 and 2008. Bathymetric data were collected with Odom Echotrac DF3200 MKII and 320M Knudsen single beams. Acquisition of bathymetric data for year 1959 and 1982 was performed by the French Navy and the

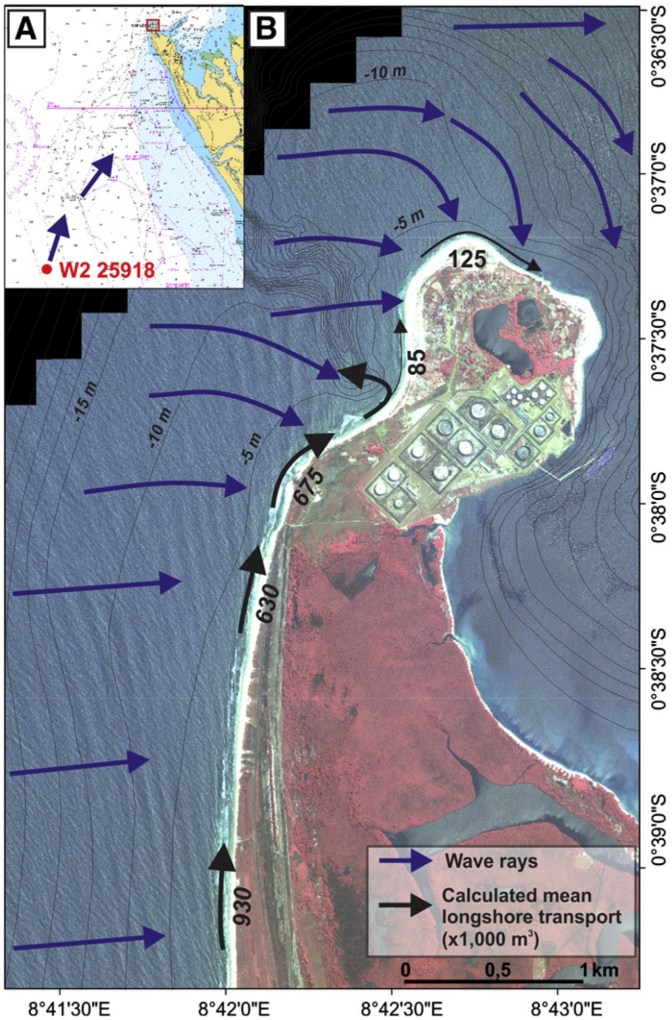

Fig. 2. Simplified wave refraction model and coastal longshore transport along the Cap Lopez spit (Artelia, 2012b). Wave rays are calculated based on long period southwesterly swells (D = 200°, T = 16 s). Basemap of Cap Lopez spit is derived from Ikonos satellite image (2001).

Comex Company, respectively. Line spacing of the survey for 2008, 1982 and 1959 datasets is around 15, 30 and 90 m, respectively.

To evaluate the morphological evolutions of the canyon, bathymetric comparisons were performed on the 1959–2008 single-beam datasets. The estimated errors associated with bathymetric comparisons are calculated upon the methodology established by Biscara et al. (2012). The mean estimated errors obtained for the bathymetric comparisons range between 0.8 and 1.2 m. A final digital elevation model (DEM) is generated for each survey, using the IDW interpolation method in ArcGIS software. The best results are obtained with a 10 m cell size and a 100 m fixed search radius for the 2004–2008 datasets and with a 20 m cell size with a 100 m fixed search radius for 1959 and 1982 dataset, reflecting the low bathymetric density soundings. For the sake of coherency, bathymetric comparisons are therefore performed at 10 m resolution between 2004 and 2008 and at 20 m resolution for older bathymetric data. Bathymetric evolution is calculated by subtracting similar DEMs extent (~2 km^2) to determine the depth difference for each grid cell. The depth difference in each cell is then multiplied by the cell surface and integrated to estimate the volume differences in specific regions. The impact of the interpolation error on the estimated volume change is calculated by multiplying the surface of the study area by the interpolation error of the compared DEM. Analysis of the thalweg migration through time is performed between 1959 and 2008 on several transects located on the meanders (Fig. 1). The line of the steepest slope of the thalweg is reconstructed for each year using the 0.5 m isobaths and its position is noted on the bathymetric sections (Fig. 4).

The volume of sediments deposited in the canyon by the longshore drift depends essentially on the proximity and the orientation of the shoreline, so we tried to link the morphological evolution of the canyon head along with the coastline. Coastlines originated from georeferenced satellite images and from topographic profiles were provided by the company Total and used to monitor the coastline position between 1959 and 2007. Although the coastlines result from different sources and resolutions, they provide useful information about the coastal dynamics.

4. Results and discussion

4.1. Coastline position

The analysis of the coastline position between 1959 and 2007 indicates a significant westward rotation of the coast, resulting in erosion of ~200 m in the vicinity of the lighthouse while the extremity of the island straightened northward (accretion up to 270 m, Fig. 3D). However, comparison of the coastline position at shorter time scale shows that this evolution results from variable accretion and erosion rates. The erosion of the coast around the lighthouse has been relatively progressive (~5 m/year between 1959 and 1982 and ~3.5 m/year between 1982 and 2007, Fig. 3B & C), although a short cycle of enhanced erosion was reported between 1978 and 1984 (Latteux, 2007). On the contrary, the coastline position immediately north of the lighthouse has been characterized by significant accretion between 1959 and 1982 (>10 m/year) and a period of relative stability between 1982 and 2007 (Fig. 3D). However, the apparent stable position of the coastline during this period reflects erosion and accretion cycles on short-time scales (Latteux, 2007).

The modification of the coastline position may have a strong impact on the propagation of waves on the coast and consequently on the longshore transport. In 1959, the rectilinear morphology of the coast south of the lighthouse favoured the sedimentation in the continuity of the coast, forming a submarine sandbar located west of the canyon head (Fig. 3A). North of the lighthouse, the canyon morphology and the SW–NE orientation of the coast tend to modify the direction and the intensity of propagating swells and probably contribute to a weaker longshore transport. Between 1959 and 1982, erosion of the lighthouse headland modified the sedimentary dynamics, and sediments which previously accumulated to form a sandbar were finally transported alongshore, favouring the accretion of the coast north of the canyon (Fig. 3C). With the smoothing of the coast, the longshore drift north of the lighthouse is expected to become more and more intense.

4.2. Canyon head

The bathymetric comparison between 1959 and 2008 highlights the north–eastward shift of the canyon head without retrogressive erosion (Fig. 3G). The SW flank of the canyon head as defined by the 10 m isobath moved by 180 m (or by 3.7 m/year, Fig. 3D). The lateral migration of the canyon head is characterized by deposition on the SW flank and erosion along the NE flank (+6 m and −4 m in average, respectively Fig. 3G). However, on a shorter time scale, bathymetric data show that the lateral migration of the canyon head happened in two steps. First, significant deposition in the canyon head (600×10^3 m^3 ± 200×10^3 m^3) occurred between 1959 and 1982, leading to a displacement offshore of the 10 m isobath (30–40 m, Fig. 3B & E). Then, between 1982 and 2008, the canyon head experienced deposition on the SW flank (~400×10^3 m^3 ± 100,000 m^3) while erosion occurred on the NE flank (350×10^3 m^3 ± 100,000 m^3 Fig. 3F). This morphological evolution led to a 140 m north–eastward shift of the 10 m isobath (Fig. 3D). On a shorter time scale (2004–2008),

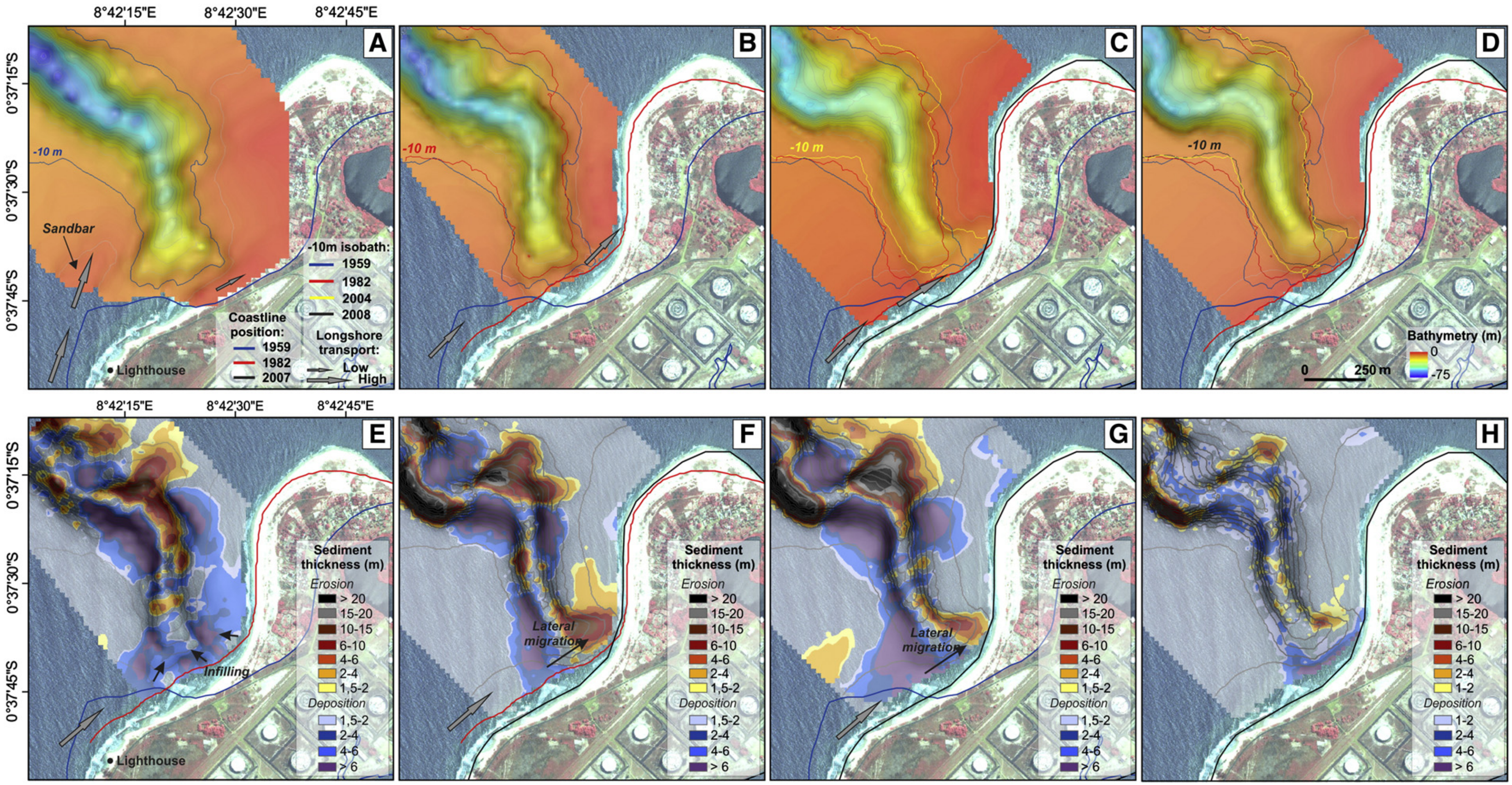

Fig. 3. Coastline and bathymetric evolutions on the area of the canyon head between 1959 and 2008. 1959, 1982 and 2007 coastlines are obtained from georeferenced satellite images and topographic profiles. A: 1959 bathymetric map; B: 1982 bathymetric map; C: 2004 bathymetric map; D: 2008 bathymetric map; E: difference map 1959–1982; F: difference map 1982–2008; G: difference map 1959–2008; H: difference map 2004–2008.

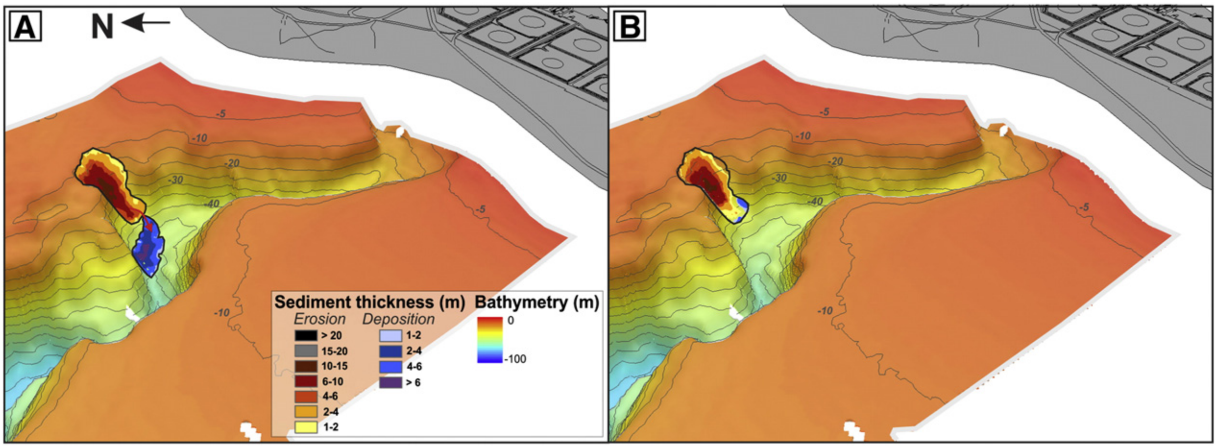

Fig. 4. Slide initiation and evolution on the outer bend of the Cap Lopez Canyon. A: 2006–2007; B: 2006–2008. Vertical exaggeration: ×3.

significant morphological changes of the head are also visible (Fig. 3H), suggesting that head migration through time is essentially progressive.

Because of the strong relationships between littoral inputs and feeding of coastal submarine canyons (e.g. Chamberlain, 1964; Smith et al., 2007), morphological evolution of the canyon head must be related to beach geomorphic change. From the comparison of coastline position between 1957 and 2007, it is proposed that the longshore transport increased progressively in relation with the erosion of the lighthouse headland. Sediments which initially accumulated to form a sandbar were therefore transported alongshore leading to the partial infill of the canyon head and to the accretion of the coastline north of the canyon. In response to the increasing longshore transport, the canyon head migrated downdrift and limited the westward accretion of the coast. The imbalance between strong sediment supply (up to 80% of the longshore transport, Artelia, 2012c) and erosion processes may explain why the SW flank experienced infilling and is not subject to recurrent erosion. However, as the canyon head becomes narrower, the opposite flank is eroded. This observation suggests that down-canyon flows are sufficiently recurrent to maintain the morphology.

In shallow water, canyon currents are related to surface waves, wind and high tides (e.g. Shepard et al., 1975; Inman et al., 1976). During storms, nearshore circulation cell consisting of the longshore and rip currents pulsate with the frequency of the group of high waves (in Inman et al., 1976). As the longshore current pulsate, strong rip currents flow seaward in the canyon and depose sand in the submarine canyon head and along its rim (Inman et al., 1976). Several works have reported resuspended sediment visible at the sea surface associated with rip currents (e.g. Reimnitz, 1971; Inman et al., 1976; Okey, 1997), suggesting large-scale mixing during the initiation of sediment transport. Strong current action, accompanied by larger volumes of sediment-laden water (Inman et al., 1976) or breaching of sand trapped in the canyon head (Mastbergen and Van den Berg, 2003) would then be conducive to the formation of turbidity currents ("flushing events"; Okey, 1997). Until the supply of sand is depleted, the turbidity current is sustained and its erosive capacity may increase (ignitive turbidity current; Parker, 1982; Fukushima et al., 1985). When no more sand is incorporated in the turbidity current, acceleration stops but the flow may persist for a long distance (Mastbergen and Van den Berg, 2003). These events are however not necessarily related to storm conditions: previous works have reported the influence of heavy runoff, high swells or tides in the formation of these flows (in Shepard et al., 1975). Occasional current measurements on the canyon floor at 45 m water depth showing down-canyon currents with velocity peak up to 0.7 m/s (Artelia, 2012a) may indicate similar processes, however further work is needed to confirm this hypothesis.

4.3. Slope failures

Visual comparison and subtraction of 2006–2007 bathymetric grids reveal a small submarine slide located on the northern flank of the canyon (Fig. 4A). The slide scar shows an amphitheatre-shape and extends over 175 m between 10 m and 40 m water depth. Its maximum width of 150 m decreases gradually as water depth increases. The headwalls of the scar are up to 15 m high and exhibit steep slopes (up to 12°). The bottom of the depression displays a U-shape, again with a steep slope (9°) and a relatively smooth sea-floor relief. The apparent slide volume (SVa) between 2006 and 2007, calculated using the methodology of Smith et al. (2007), is estimated to be 95,000 $m^3 \pm 15{,}000$ m^3 over an area of 0.02 km^2, corresponding to an average remobilized sediment thickness of 5 m (Fig. 4A). The comparison of bathymetric data shows the existence of a small depositional area in the canyon thalweg between 40 and 50 m water depth (35,000 $\pm$ 5000 m^3; Fig. 4A). Assuming that these deposits are related to the slide, only 40% of the slide volume is preserved at this location, suggesting that a part of the deposits was quickly remobilized.

To understand the triggering of this small slide, we analysed the oceanographic conditions measured offshore Cap Lopez (NOAA database: http://www.sat-ocean.com/squalls.html) during the studied period. Two large storms occurred between 2006 and 2007. Significant wave height (average of the upper third of the highest waves, Hs) reached nearly 3 m, corresponding to an estimated maximum wave height of about 6 m. Because the upper slide head is located in shallow water (10 m), the bathymetric increase related to wave height may have generated excess pore pressure in the seafloor and a decrease of the sediment strength leading to failure (Prior et al., 1989).

A similar comparison was performed between years 2006 and 2008 (Fig. 4B). The SVa between 2006 and 2008 (70,000 $m^3 \pm 15{,}000$ m^3) is reduced by around 25% and corresponds to an average remobilized sediment thickness of about 3.5 m. The major change when compared to years 2006–2007 is the presence of a depositional area that partially fills the lower part of the slide scar between 30 m and 40 m water depth (Fig. 4B). Assuming a constant sedimentation rate, the slide scar could therefore be completely filled in approximately 4 years. Even if the infill of the slide scar is not performed homogeneously due to overflow processes, these observations highlight that slope failures on the canyon flanks may quickly disappear from the morphology.

4.4. Migration of the thalweg

Throughout the 49 years between 1959 and 2008, significant meandering behaviour of the canyon thalweg has been observed (Fig. 5). Bathymetric comparison highlights the canyon morphology evolving toward an asymmetrical profile with steep banks on the outer bends and smoother inner bends (Fig. 5A). By comparing the bathymetric sections, the lateral migration of the thalweg from the inner bend toward the outer bend is estimated between 77 and 190 m (or between 1.6 and 3.8 m/year, Fig. 5A) and is associated with an increase of the sinuosity (from 1–1.5 in 1959 to 1.3–3.6 in 2008; Fig. 5A). The data also reveal the intensity of the sedimentary processes associated with the migration: erosion on the outer banks results in a retreat of several tens of metres of the canyon flanks (Fig. 5A). Thickness of eroded deposits on the outer bend can reach 60 m while deposition of up 25 m occurs on the inner bend. Despite the low density of bathymetric echo-sounding data collected in the 1959 survey, the depth of the thalweg seems relatively constant through time (Fig. 5A). Deepening of the canyon is generally low except for the last meander (10 m). The slope of the thalweg is constant (1.9°) between 1959 and 2008.

Similar morphological changes are visible at annual to multi-annual time scales (Fig. 5B). Comparison of annual bathymetric sections displays the general lateral migration of the thalweg from inner toward outer bends (2–14 m/year). These values are comparable to the average lateral-migration rates of river channels (0.7–9.4 m/year; Hickin and Nanson, 1984). The largest displacement are observed in meanders where the canyon is the widest, suggesting that the amplitude of displacement would be inversely related to the increasing confinement of the canyon. At this time scale, the lateral displacement leads to a slight increase of the sinuosity (up to + 0.07 between 2004 and 2008). Although the thalweg position moves naturally toward the outer bend, annual bathymetric data also illustrate displacement on the opposite direction (Fig. 5B). The displacement toward the inner bend varies from 9 to 36 m and is mainly observed between 2006 and 2007 on the two

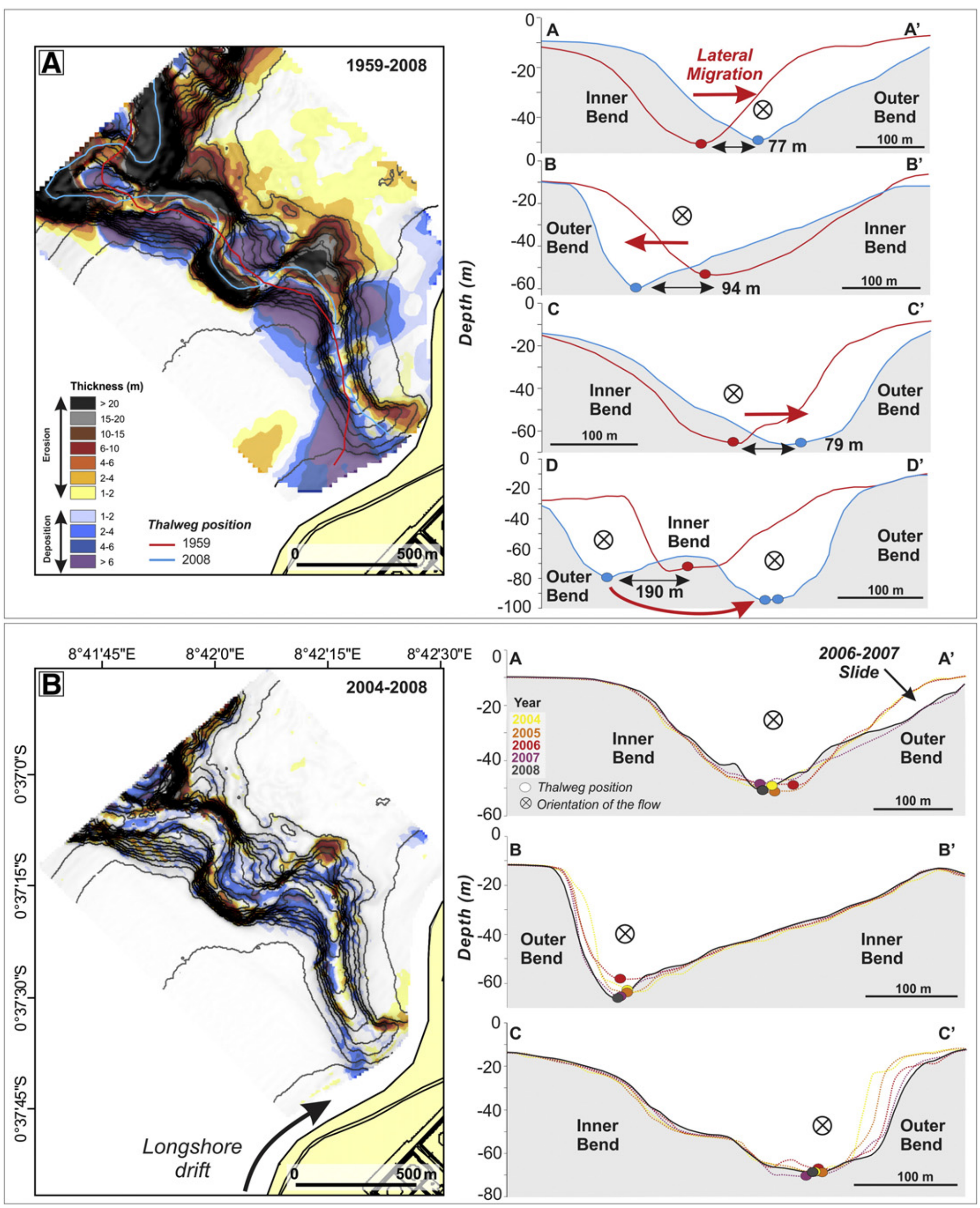

Fig. 5. Multi-scale evolution of the Cap Lopez Canyon and associated bathymetric profiles. A: Multi-decadal evolution (1959–2008); B: Multi-annual evolution (2004–2008). Bathymetric profiles are located in Fig. 1.

meanders where the canyon is the widest. This kind of morphological evolution suggests chute cut-off, a relatively common phenomenon in river systems and which has already been observed in the submarine environment (Lonsdale and Hollister, 1979). Part of the slide deposited on the canyon floor on the most coastal meander probably deflected the sediment gravity flows (Kneller and McCaffrey, 1999) and conduced to the shift of the thalweg toward the inner bend. On the most distal meander, particular high-energy events may explain the generation of the meander chute cut-off. Similarly to the longer time scale, erosion along the outer bend is more intense than deposition along the inner bend (Fig. 5B). Eroded thicknesses on the outer bend are very significant and can reach up to 36 m. Such erosional rates are similar to those observed in the Monterey Canyon (Smith et al., 2007). Erosion of the outer bends is the result of two different processes: (1) mass-wasting and (2) steady erosional processes (Fig. 5B). The relative uniformity of the slope and the absence of numerous recent slide scars on the outer bends suggest that erosion related to the transit of sediment gravity flows would be the predominant process (Fig. 5B). Depth variations of the thalweg are significant between two consecutive surveys (up to 6 m), but reduced on a 4-year scale (<3 m), suggesting that the apparent stability of the canyon floor results from a dynamic balance between erosion and deposition through time.

These comparisons at different time scales reveal that areas of deposition are restricted to the inner bends. Works of Kane et al. (2008) highlight that channel fill architecture relates directly to the degree of flow by-pass, in turn largely determined by the degree of confinement. Channel planform geometry may therefore reflect the most frequent flow magnitude to pass through it. Equilibrium flows (flow height <3.3× confinement depth) have the potential to bypass much greater volumes of sediment and deposit at the inner bend to form point bars. For disequilibrium flows (flow height >5× confinement depth), channel acts as an obstacle to flow rather than confining them, conducing potentially to enhanced overbank deposits. With an aspect ratio varying between 7.7 and 10.5, our study area is in the morphological range of modelled channels by Kane et al. (2008). The absence of overbank deposits and the developing point-bar morphology may therefore reflect that equilibrium flows are the most frequent flows in the Cap Lopez Canyon. Based on their results, the height of equilibrium flows would not exceed ~170 m. However, because the canyon is located in very shallow water (canyon flanks are around 15–20 m water depth), the flow height must be much shorter than expected.

5. Conclusions

The aim of this study was to provide new results about the modern sedimentary dynamics of a modern submarine canyon from a study of morphological changes of the Cap Lopez Canyon (Gabon, West Africa). Based on the comparison of bathymetric data and on the evolution of coastline position over a short-time scale (~50 years), the following results were illustrated:

1) Canyon head evolution is characterized by a north–eastward lateral migration between 1959 and 2008 (~180 m) which was performed in two steps. Because of the erosion of the lighthouse headland, longshore transport increased progressively northward, leading to the partial infill of the canyon head and to the accretion of the coastline north of the canyon (1959–1982). In response to the increasing longshore transport, the canyon head was then subject to a downdrift migration, limiting the westward accretion of the coast (1982–2008). The disequilibrium between strong sediment supply and erosion processes may explain why the SW flank experienced infilling and is not subject to recurrent erosion. It is proposed that recurrent flows which periodically move sediment from the canyon head to deeper water contribute to the progressive erosion of the NE flank. Small but frequent turbidity currents might be therefore the dominant sediment transport in the upper part of canyon. However, the triggering mechanisms producing these flushing events are still unclear.
2) By comparing the bathymetric data between 1959 and 2008, lateral migration of the thalweg from the inner bend toward the outer bend is estimated between 77 and 190 m (or between 1.6 and 3.8 m/year) and is associated with an increase of the sinuosity (from 1–1.5 in 1959 to 1.3–3.6 in 2008). The lateral migration of the thalweg is associated with strong erosion on the outer bends (up to 60 m) and sediment deposition on the inner bends (up to 25 m). Based on works of Kane et al. (2008), the absence of overbank deposits and the developing point-bar morphology may reflect that equilibrium flows are the most frequent flows encountered in the Cap Lopez Canyon.
3) Erosion of the outer bends is the result of two different processes: (1) mass-wasting and (2) steady erosional processes. The relative uniformity of the slope and the absence of numerous recent slide scars on the outer bends suggest that erosion related to the transit of sediment gravity flows would be the predominant process.
4) Although the thalweg position naturally moves from the inner bend toward the outer bend, bathymetric data reveal that fluctuations exist at least at an annual scale.
5) Comparison of a slide scar on the canyon flanks indicates a reduction by around 25% of the apparent slide volume between 2006 and 2008. Slope failures in such dynamic sedimentary environments may therefore quickly disappear from the morphology.

All these observations highlight that the Cap Lopez Canyon is a very active system which plays a role in the transfer of large volume of coastal sands to deep water. Active canyons are dependent on the nearby position of the shoreline for their sediment supply but also on surf-zone-related processes to initiate the seaward transport of the sediment. Due to this strong feedback, any modification of the coastline position, which controls the intensity of the longshore transport, will have deep consequences on the canyon head alimentation. On the other hand, sand transferred from the coastal area to the submarine canyon is definitively lost from the littoral system and must have an impact on the coast morphology. Due to human development and activities along the coast, it is a prerequisite to study the sedimentary cell formed by the interaction between the coastline and the canyon to assess the relevance of coastal geological hazards.

Even if morphological similarities to river channels are noticed, major differences remain. Differences in density contrasts of flows relative to ambient fluids, effects of centrifugal and Coriolis forces on flows, frequency, volume and duration of steady vs. catastrophic flows, modes of sediment transport, and effects of sea level changes appear to have caused the main differences in the internal architectures and lateral migration or aggradation of fluvial and deep-water channels (Kolla et al., 2007). Moreover, some sedimentation patterns are thought to be unique to subaqueous channels (e.g. Kane et al., 2008; Huang et al., 2012), suggesting significant differences between river and submarine flow dynamics. Although monitoring activities are particularly difficult in such type of environment, they remain until to present the only way to compare and to validate results from numerical models or flume experiments.

Acknowledgements

We are grateful to the SHOM, TOTAL and IOTA SURVEY for having made the bathymetric and topographic data available. We gratefully acknowledge H. Gillet, L. Troudet and R. Pennel for their critical and constructive comments and S. Laugier for English polishing. Finally, we benefited from constructive reviews by D. Smith, P. Hill and the Editor-in-Chief D.J. Piper. Laurie Biscara's PhD thesis is funded by a DGA (French Ministry of Defence)-CNRS doctoral fellowship. A part of this work was funded by the INSU (Institut National des Sciences de l'Univers) Program “Actions Marges”.

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