Coastline sand waves on a low-energy beach at “El Puntal” spit, Spain

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Abstract

Coastline sand waves have been observed at “El Puntal” spit, located on the north coast of Spain. The spit has been monitored by an Argus video system since 2003 and the formation and destruction of sand waves has been observed. Coastline data from the video images are analyzed by means of principal components analysis, obtaining a mean sand wave length of 125–150 m and a maximum amplitude of ≈ 15 m. It is also observed that sand waves reach their maximum amplitude at about 15 days. No propagation of these sand waves is noticed during the approximately two-month-long events analyzed. Sand wave formation and evolution are examined in relation with the prevailing local wave conditions during that period. Incident waves at the west end of the spit approach from the east–northeast, with a very high angle with respect to the shoreline. Field observations suggest that sand waves may result from an instability in alongshore sediment transport caused by moderate-energy waves with a high-angle incidence.

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1. Introduction

Shoreline sand waves are periodic features in the shoreline planform, with spatial and temporal scales of 10−1–102 km and 10−1–102 yr, that may propagate along the coast. These features have also been defined as longshore wave-like movements of the shoreline, measured in a horizontal plane (Verhagen, 1989). Sand waves have been observed on many coasts around the world although systematic field studies are relatively scarce. Sand waves along the Danish coast with lengths of 0.2–2 km and amplitudes of 60–80 m propagating in the same direction as the net longshore sand transport were described by Bruun (1954). Thevenot and Kraus (1995) identified eleven longshore sand waves along Southampton Beach in New York, with an average length of 0.75 km, an amplitude of 40 m, an average migration speed of 1.09 km yr−1 over the winter and a yearly average of 0.35 km yr−1. The generation of those features was attributed to the periodic opening of a small inlet and the subsequent welding of its ebb shoal to the beach. More recently, Davidson-Arnott and Van Heyningen (2003) have described longshore sand waves that occur at the downdrift end of a 40 km long spit in Lake Erie, Canada ranging in length from 350 m to more than 1500 m and showing an alongshore migration at rates of 100–300 m yr−1. They conclude that sand waves result from onshore migration and welding of inner nearshore bars, and that their development and growth is promoted by refraction of highly oblique waves. The Dutch coast is a quite well documented case. According to Ruessink and Jeuken (2002), the wavelength is about 3.5–10 km and the amplitudes are small, ranging from 20 to 50 m. The celerity ranges between 0 and 0.2 km yr−1. The results of the analysis by Verhagen (1989) are qualitatively in agreement with Ruessink and Jeuken (2002) but some quantitative differences arise. Larger sand wave amplitudes
ranging from 30 to 500 m are reported and the wavelengths are larger too, ranging from 2.5 up to 22 km. The specific case of the sand waves propagating northeasterly along the isle of Vlieland in the Wadden Sea was analyzed by Bakker (1968) who reported a migration speed of 0.17–0.33 km yr\(^{-1}\). Similarly, Guillet et al. (1999) investigated the specific case of the Holland coast. By analyzing the Jarkus data set during the period 1964–1992 they found southward propagating sand waves. Their wavelength, \(\lambda \approx 2–3\) km, is however in the lower limit of the range observed by Russink and Jeuken (2002) and the celerity is \(V \approx 0.15–0.2\) km/yr.

At smaller length and time scales (10\(^{-1}\) km, 10\(^{-1}\) yr) shoreline sand waves are often observed at El Puntal spit located on the northern coast of Spain (Fig. 1). Due to the reduced temporal and spatial scales of these morphological features, this location provides an ideal laboratory where direct measurements of the development of sand waves, by means of a video monitoring system, have been possible for the first time. A three-year long time series of video images contains several complete sand wave events including their formation and their destruction. The first aim of this work is to describe these small-scale coastline undulations or sand waves observed at El Puntal and to analyze their growth, evolution and eventual decay.

Various authors have recently explained the presence of large-scale shoreline features as a result of an instability induced by high-angle waves (waves approaching at an angle greater to 43 with respect to shoreline orientation). Some examples of these features are the sand waves on the Dutch coastline (Ashton et al. 2003; Falqués, 2006), the capes of the North Carolina coast of North America, and the spit-shaped shoreline features at the north shore of the Sea of Azov, Ukraine (Ashton et al. 2001; Ashton and Murray, 2006a,b). Ashton and Murray (2006b) have also stated that shoreline instability might play an important role in the behaviour of the Long Point sand waves observed by Stewart and Davidson-Arnott (1988) and Davidson-Arnott and Van Heyningen (2003). The instability mechanism is based on the coupling between littoral drift and coastal morphology. However, no direct observation and measurement of the development of the shoreline features driven by the instability have been made before, mainly because the length and time scales at which the instability is expected to develop on open ocean beaches are typically very large \(\approx 1–10\) km and 1–5 yr (Falqués and Calvete, 2005). It is worth noting however that according to these authors, the scale of such instability might be reduced on low-energy beaches and by a quite steep bathymetry, such as along sandy spits protected by headlands. These conditions along with a persistent high wave obliquity prevail on el Puntal spit so that the scales of the developing instability could be similar in this case to the scales of the observed coastline sand waves. Therefore, the second aim is a characterization and analysis of nearshore conditions and, in particular, wave conditions as a first step towards elucidating whether those sand waves could be related to that shoreline instability.

This paper is organized as follows. In Section 2, a description of the study site and the field data collected is presented, which includes coastline and deep water wave data. Then, the methodology applied for the coastline and wave data analysis is explained in Section 3, followed by an analysis of the correlation between coastline sand waves and water waves (Section 4). Finally, the discussion and conclusions are presented in Section 5.

### 2. Study area and field data

The beach of interest is located on the west end of a 2.5 km-long sand spit, called “El Puntal”, with a W–E orientation and located north of Santander Bay, on the Cantabrian coast of Spain, Gulf of Biscay (Fig. 1). The Cantabrian Coast, in northern Spain, is divided into a series of pocket beaches and small inlets isolated between rocky headlands. Waves on the Gulf of Biscay approach mostly from the northwest with a mean significant wave height, \(H_s\), of 1 m and a typical winter storm significant wave height of \(H_s \approx 5\) m. The tides are semi diurnal with a mean tidal range of 3 m and a spring tidal range of 5 m. Mean grain size along the spit is 0.3 mm.

The middle and eastern parts of the spit are fully exposed to the NW Cantabrian swell waves, and their beach morphodynamic state is, according to the Wright and Short...
Fig. 2. Wave propagation associated with typical storm wave conditions \((H_s=5\text{ m}, T_p=16\text{ s})\) from the NW during high tide, performed by Oluca-SP (GIOC, 2001).

Fig. 3. (a) Bathymetry of the study area and typical bathymetric profiles of (b) the western part of the spit (navigation channel area) and (c) the eastern part of the spit (open beach area). Depth is referenced to the lowest astronomical tide (2.5 m below MSL).

Fig. 4. Study area showing camera location (left), and video camera setup on top of a building 90 m above MSL (right).
Short (1984) classification, usually dissipative. On the contrary, the west end of the spit is characterized by a reflective beach in accordance with a reduced wave height due to its location in the sheltered area behind the Magdalena Peninsula. This section of the spit shows a propagation coefficient \(K_p = H_s / H_s^0\) of around 0.2. In addition, waves approach with high obliquity from the east-northeast sector, due to diffraction and refraction effects (Fig. 2).

The beach profile of the most western 500 m-section of the spit, characterized by a very steep slope, reaches the 15 m isobath within the first 100 m, due to the presence of an ebb tidal channel (Fig. 3b). This bathymetric profile shows an intertidal section that approximates to a Dean-type equilibrium beach profile (Dean, 1991) with a mean beach slope (shoreface slope) of around 0.04, and a channel section with a very steep slope of about 0.15. The beach on the rest of the spit (east side) shows a mild slope \(\approx 0.013\) bathymetric profile which represents a dissipative beach with a wide surf zone (Figs. 3c and 5).

The morphology of El Puntal spit has been substantially modified in the past two centuries, mostly due to a decrease in tidal prism caused by a land reclamation project in Santander Bay in 1875. As a result, the spit end has turned towards the bay and the mean shore normal in that section is now oriented towards the north. A detailed morphological evolution of ‘El Puntal’ spit is presented in Losada et al. (1991).

Coastline undulations are usually observed on the west end of the spit during winter, for periods that last approximately two months. These coastline features appear in the most western 300 m-long section, in which the bathymetric profile is a channel-profile (Fig. 3b). The length and time scales of these undulations are of the order of 100 m and weeks. This allows for the first time to measure and analyze complete shoreline sand wave events since their formation until their complete destruction.

2.1. Coastline data

The coastline evolution data of El Puntal spit was obtained from video images collected by an Argus video monitoring system (Holman et al., 1993) installed in front of El Puntal since 2003. The Argus video system consists of four video cameras pointing to El Puntal, which take images every 30 min. The study area is captured by cameras c1 and c2 (Figs. 4 and 5), which are located \(\approx 800\) m from the shoreline, with a pixel resolution of 0.5 m in the cross-shore direction and 1.5 m in the alongshore direction (Medina et al., 2007). Three types of images are available on the Argus image archive (www.thecoastviewproject.org): snapshot (instantaneous oblique camera shots), time exposure images (average of the intensities seen by the cameras over a time period of 10 min with a sampling rate of 1 s), and variance images (variability in intensity seen in the image over a 10-minute period).

Two sand wave events of approximately two-month-long durations were observed on this set of video images (Fig. 6),

![Fig. 5. Timex video image of El Puntal showing the area captured by each camera.](image-url)

![Fig. 6. Video images of the study area during the (a) first and (b) second event, (1) at the beginning (absence of sand waves), (2) at the middle (presence of sand waves), and (3) at the end (absence of sand waves), all at mean sea level (MSL).](image-url)
the first one from November 6th 2004 to January 4th 2005, and the second one from December 7th 2005 to February 19th 2006. Two sand waves are clearly identifiable on the coastline (Fig. 6 a and b, panel (2)). Time exposure images (timex) corresponding to mean sea level during both events were selected, with the aim of comparing coastline changes at the same tide level.

The coastline is detected from the oblique timex images, captured at mean sea level, by means of the Physical and Statistical Detection Model (PSDM). The PSDM (Osorio, 2006) combines six different edge detection algorithms, including the SDM (Shoreline Detection Model), developed by Aarninkhof et al. (2000, 2003), which identifies colour differences between wet and dry beach, based on the PIC (Pixel Intensity Clustering) technique. The PSDM was validated with field measurements of intertidal bathymetry at El Puntal, obtaining a mean vertical error of \( \mu_v = 0.36 \) m with \( \sigma_v = 0.10 \) m (Osorio, 2006). Once the coastline is detected from the oblique image, it must be rectified into \((x, y)\) coordinates (Fig. 7) for further analysis.
2.2. Wave data

Deep water wave data \((H_s, T_p, \theta)_\infty\) were obtained from the oceanographic database of the Spanish Port Authority (www.puertos.es), specifically from the WANA data set, which provides time series of wave and wind parameters obtained by numerical modelling. Wave fields are generated by means of the Wave Model (WAM), which is driven by wind fields. The WAM-model (WAMDI Group, 1988) is a third-generation wave model which solves the energy balance equation explicitly for the two-dimensional surface wave spectrum, without making any presumptions on the shape of the wave spectrum. The WAM cycle 4 (Günther et al., 1991), was introduced operationally at the European Centre for Medium-Range Weather Forecasts (ECMWF) (Günther et al., 1992) in November 1991. Later on, Gómez and Carretero (1997) modified the nesting procedure of the model for its application to the Spanish coast.

The WAM grid has a resolution of 0.25° (30 km) in the Atlantic, and of 0.125° (15 km) in the Mediterranean. The grid point WANA-1066075 (Fig. 8), located at coordinates 43.75° N 3.5°W (approximately 35 km offshore of the study area), was selected for the present work.

3. Field data analysis

With the aim of investigating the coastline evolution in relation with the prevailing local wave conditions, coastline and wave data must be first treated separately. Coastline data is analyzed by means of a principal component analysis in order to examine their variability modes, and wave data is propagated to the shore with the purpose of obtaining the wave conditions close to the study area. Possible correlations between coastline evolution and wave conditions are further described in Section 4.

3.1. PCA method

The principal component analysis (PCA) method was performed with the aim of separating spatial and temporal
variability of coastline data. This method, also known as the empirical orthogonal function (EOF) method has been widely applied in coastal morphology, mostly focused on cross-shore variability (Winant et al., 1975; Dick and Dalrymple, 1984; Medina et al., 1994), and more recently on longshore variability (Wijnberg and Terwindt, 1995; Muñoz-Pérez et al., 2001; Miller and Dean, 2007). The purpose of the method is to describe coastline changes by using the least number of functions, called eigenfunctions. The first eigenfunction represents the dominant mode which accounts for the greatest variance. Successive eigenfunctions represent the greatest amount of the remaining variance. Considering \( x \) the cross-shore axis and \( y \) the along-shore axis, \( xyt \) is explained by the summation of eigenfunctions multiplied by coefficients, 

\[
x(y, t) = \sum_{n=1}^{N} C_n(t)e_n(y)
\]

where \( e_n(y) \) represents the spatial eigenfunction \( n \) evaluated in the alongshore position \( y \). The constant \( C_n(t) \) represents the coefficient for eigenfunction \( n \) at time \( t \). Original coastline data can be reconstructed by adding all eigenfunctions multiplied by their coefficient. In the present work, two sets of coastlines were analyzed, each of them corresponding to a different time period (November 2004 to January 2005 and December 2005 to February 2006). A total of 22 coastlines were analyzed for the first event (Fig. 9a), and 25 for the second (Fig. 9b). The number of coastlines depends on the number of images captured during mean sea level and from which the coastline could be detected (captured when not foggy, during daylight, etc.). Coastline changes are referenced to the coastline at time \( t_1 \) characterized by the absence of sand waves.

The coastline data, and consequently the spatial eigenfunctions and their temporal evolution, are very similar for both events (Fig. 9). The first eigenfunction (solid line), which accounts for the greatest variance (>70%) represents two waves of different amplitude, the westward being almost two times greater, with a temporal evolution that grows from zero to approximately the same value in both cases, and goes back to zero at the end. The second eigenfunction (dashed line), with \( \approx 10\% \) of the variance, shows a spatially increasing trend, with a temporal evolution that could be represented by a sinusoid-like curve, more clearly in the first event than in the second. And finally, the third spatial eigenfunction (dotted line) explains less than 8% of the variance.

The first spatial eigenfunction (Fig. 9a and b, middle panels, solid line) represents the mean shape of the coastline sand wave. Therefore, by performing this analysis, a mean wavelength of the sand wave together with the time evolution of its amplitude (Fig. 9a and b, lower panels, solid line) can be obtained. A more detailed description is presented later in Section 4.

### 3.2 Wave propagation

The wave time series, corresponding to the same period during which video images are available, was propagated from the offshore to the coast by means of the Oluca-SP model (GIOC, 2001) which solves the parabolic approximation of the mild-slope equation for spectral waves. Then, from the propagation grid, a point close to the study area \( p \) (Fig. 3) was selected to represent the wave conditions in the sheltered area (Fig. 10). Point \( p \) is located at a depth of \( \approx 15 \) m from MSL and at a distance of \( \approx 100 \) m from the shore. Seasonal variation of significant wave height, \( H_s \), and peak period, \( T_p \), is evident.

![Fig. 10. Significant wave height, \( H_s \), peak period, \( T_p \), and wave angle, \( \theta \), time series at a location offshore of ‘El Puntal’ spit and at point \( p \), from January 2002 to March 2006.](image-url)
both at the offshore and at point \( p \), being smaller in summer and higher during winter.

On the other hand, wave direction does not show a clear seasonal variation. Waves offshore approach mostly from the west–northwest sector and also from the north–northeast. There are also a few waves registered approaching from the south which are considered as calm periods at point \( p \). All waves arrive at point \( p \) from the east–northeast sector and significant wave height never reaches 2 m, due to the shelter provided by the Magdalena Peninsula. Peak periods are relatively large, ranging from around 5 to 20 s. This leads to large Iribarren numbers (\( \text{Ir} = \tan \beta / \sqrt{H_b/L_s} \)), also known as surf similarity parameter (Battjes, 1974), which means waves of low steepness on a steep beach, with a predominant collapsing or surging breaking type.

### 4. Relationship between coastline evolution and wave conditions

From the PCA analysis (Section 3.1) it is observed that coastline sand waves from the two events present similar features. The wavelength of the sand wave is of the same order of magnitude in both events, being smaller in the first event, with \( \approx 125 \) m, than in the second event (\( \approx 150 \) m). The sand wave amplitude shows a maximum of the same magnitude in both events (\( a = 15 \) m) which may imply that saturation of sand wave amplitude has been reached at that point. The time at which the sand wave reaches a maximum amplitude is around 15 days for both events. The western crest of the mean shape of the sand waves is approximately two times larger than the eastern crest for both events. No sand wave propagation is observed in any of the two events, the more evident wave crests are observed in the same alongshore position throughout the duration of the events. In the second event, a second sand wave mode (\( \lambda = 85 \) m) is present for a short time.

With the purpose of analyzing the wave conditions at the different stages of the coastline sand waves, the wave parameter time series at \( p \) corresponding to each event were divided into three different stages according to the temporal evolution of the first eigenfunction, \( C_1(t) \) (Fig. 11). The first stage is characterized by a continuous growth of the sand wave amplitude, while the last stage is characterized by a continuous decrease of the amplitude, ending with the total destruction of the sand wave. The maximum sand wave amplitude is reached (saturation) during or just in the limit with the intermediate stage. After reaching its maximum value, the amplitude oscillates for some time around a certain value before showing a sustained decrease (decay stage).

Wave conditions during each stage were plotted separately for each event in order to distinguish the most dominant values of wave parameters (Figs. 12 and 13). The wave angle is defined as the angle between the wave crests and the shore, assuming that the mean shore normal within the study area is oriented to the north. Fig. 12 shows the wave angle, \( \theta \), as a function of significant wave height, \( H_s \), and peak period, \( T_p \), while the most dominant wave angle with its corresponding wave height during each stage are plotted on Fig. 13. For the first event (Figs. 12a and 13a), wave conditions are ‘moderate’, with \( H_s \) always
smaller than 0.8 m, and a dominant $\theta$ around 50–55° during the growth stage. The saturation stage shows the smallest dominant wave heights ($H_s \approx 0.25$ m) with a greater range in wave angle, from 50 to 80° (Fig. 13a, 2)). More energetic and more oblique waves are present during the decay stage, with a dominant angle around 85° that corresponds to a $H_s$ around 1 m (Fig. 13a, 3)).

During the second event (Figs. 12b and 13b), wave characteristics corresponding to the growth and decay stage are similar to those of the first event, being ‘moderate’ at the growth stage, and showing the highest dominant wave angle during the decay stage, although wave height is smaller (Fig. 13b, 3)). However, in the intermediate stage the main difference is a storm event that occurred around January 1st, time at which the sand wave starts to show an unsteady behaviour, exhibiting an overall decreasing trend all the way to the end (Fig. 11). The wave angle range during that intermediate stage is similar to that of the first event; however, the wave height range is larger (Fig. 12b, 2)).

In addition, wave parameter time series, together with the alongshore, $F_s$ component of the wave energy flux ($F_s = EC_s \sin \theta$, where $E$ is the wave energy and $C_s$ the wave group velocity), were compared to the first derivative of $C_1(t)$ (sand wave growth velocity) through the correlation coefficient, $r_{xy}$, given by,

$$r_{xy} = \frac{\text{cov}(x, y)}{\sigma_x \sigma_y} = \frac{\sum_{i=1}^{N} (x_i - \bar{x})(y_i - \bar{y})}{\sigma_x \sigma_y},$$

where $\sigma_x$ and $\sigma_y$ are the standard deviations of the data sets being correlated ($x$ and $y$). The objective is to measure the linear association between shoreline changes and wave conditions. The correlation between two data sets is perfect when $r_{xy} = \pm 1$, varying inversely with one another if the coefficient is negative, and is absent when $r_{xy} = 0$.

In order to calculate the correlation, and since the shoreline response is not instantaneous, a period of 1 day of wave parameters prior to each shoreline measurement was considered, and the mean of those daily sets of nearshore parameters was calculated. The resulting time series of the daily mean of each parameter (Fig. 14) was finally correlated with the derivative of $C_1(t)$.

Fig. 12. Wave angle, $\theta$, as a function of significant wave height, $H_s$, and peak period, $T_p$, during the three characteristic stages of each event: (1) growth, (2) saturation, and (3) decay. (a) November 2004 to January 2005, (b) December 2005 to February 2006).
As expected, the derivative of $C_1(t)$ (not shown) is positive when sand wave amplitude grows and negative when it decreases. Therefore, it shows the greatest positive values during the growth stage, then it oscillates around zero during the intermediate stage, and is negative during the decay stage. Energy flux is mostly in the alongshore direction ($F_x \gg F_y$) due to the high-wave obliquity, and its trend is similar to $H_s$, since it depends on $H_s^2$ (Fig. 14).

Lags from zero to 5 days were considered (Table 1), indicating that wave parameters time series preceded morphological changes. Most correlation coefficients obtained are negative, which means that overall, even though correlations are weak, parameters are inversely proportional to the derivative of $C_1(t)$. Strongest relationships for the first event occur for the wave angle $\theta$, significant wave height $H_s$, and consequently, for the alongshore component of the energy flux, $F_x$, which depends on $\theta$ and $H_s^2$, all within a lag from 0–2 days. These relationships imply that higher energy episodes led to the disappearance of the sand waves, and that not much energy is required for the growth of the sand waves. For the second event, the higher absolute values of the correlation coefficient are observed for the peak period ($T_p$, lag = 2) and wave angle ($\theta$, lag = 0). The peak period has an overall increasing trend during the second event and the derivative of $C_1(t)$ presents an overall decreasing trend, which means that higher periods are related to destruction and lower periods to sand wave formation (Fig. 14). Also, higher wave angles are related to the decay stage, and lower angles to formation and evolution, as occurs in the first event.

Time series of daily maximum and minimum values of wave parameters were also considered; however correlation did not improve significantly. Also, correlation was calculated for the wave steepness, $s = H/L$, and the wave energy flux in the cross-shore direction, $F_y = EC_g \cos \theta$, resulting in very low values as

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Fig. 13. Percent of occurrence of wave angle, $\theta$, and significant wave height, $H_s$, during the three characteristic stages of each event: (1) growth, (2) saturation, and (3) decay. ((a) November 2004 to January 2005, (b) December 2005 to February 2006).
well (not shown). In general, correlation coefficients are very low for all parameters and lags considered. Therefore, the growth, evolution, and decay of coastline sand waves cannot be explained by means of a direct linear relationship with the wave parameters.

Tidal currents are also an element to be considered, especially in inlet areas. Time series of tidal current velocity for the study area during both events were obtained by means of the long-wave propagation numerical model H2D (GIOC, 1990), calibrated with field data collected during a campaign that took place at El Puntal under the framework of the CoastView Project (www.thecoastviewproject.org). Tidal current velocity at point $p$, $|V_p|$, ranges from 0–0.5 ms$^{-1}$, and its alongshore component, $V_x$, is two orders of magnitude larger than the cross-shore component, $V_y$ (Fig. 15). The highest and lowest tidal current velocity episodes during each event, related to spring and neap tides respectively, seem to have no direct relation with the sand wave amplitude evolution ($C_1(t)$) or sand wave growth velocity ($\Delta C_1(t)/\Delta t$). Even so, the tidal current velocity time series, and its longshore, $V_x$, and cross-shore, $V_y$, components time series for each event were also correlated with the derivative of $C_1(t)$. As expected, all the correlation coefficients obtained are very low ($r<0.08$), implying that tidal currents have no direct linear relation with sand wave formation or destruction.

Furthermore, field observations of coastline sand waves during a complete tidal cycle show that during high and low tide, when tidal currents are null, sand waves are also identifiable, being more evident at high tide (Fig. 16). This suggests that tidal currents have no effect on sand wave formation.

5. Discussion and conclusions

Two sand wave events were identified from video images and further analyzed by means of a principal components analysis. Similarity between both events is remarkable, mainly with respect to the mean wavelength of the sand wave ($\lambda \approx 125–150$ m), the maximum sand wave amplitude ($a \approx 15$ m), the time

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</table>
at which maximum amplitude is reached ($\approx 15$ days), the western crest being approximately two times larger than the eastern crest, the duration of the event ($\approx 2$ months), and the fact that both events happened during winter. Also, the effect of particular wave characteristics on the coastline sand waves is very similar. Formation of coastline sand waves at El Puntal spit happens when the predominant waves are of ‘moderate’ (among high-angle waves) wave obliquity ($50^\circ \leq \theta_0 \leq 60^\circ$) and moderate wave height ($0.1 \leq H_s \leq 0.5$ m), and decay with very high obliquity ($75^\circ \leq \theta_0 \leq 85^\circ$) and higher waves ($0.5 \leq H_s \leq 1.5$ m). The differences in sand wave evolution between the two events are related to differences in wave conditions. The sand waves of the second event start decaying at an earlier time because of the storm event at the beginning of January. Also, the second mode ($\lambda \approx 85$ m) highlighted by the first spatial eigenfunction (not all the time) of the second event could be related to the abrupt change in wave conditions during that period.

The observed persistency of high wave obliquity ($\theta > 43^\circ$ all the time) in the study site, draws attention to the high-angle waves instability mechanism (Falguès and Calvete, 2005) to explain the origin of the coastline sand waves. High-angle wave instability occurs because of an instability in the sediment transport. When waves break at oblique angles, they produce an alongshore current that can transport sediment along a sandy beach. The magnitude of this sediment transport rate depends on the angle between wave fronts in deep water and the local coastline orientation (Komar, 1998). Sediment flux rises from zero at normal wave incidence ($\phi_0 = 0^\circ$) up to a maximum at $\phi_0 \approx 43^\circ$ (Ashton et al., 2001) and is null again for $\phi_0 = 90^\circ$.
(Fig. 17a). Even though longshore sediment transport tends to
smooth out coastline irregularities, an investigation by Murray et al. (2001) has shown that this is true only when waves approach
at a relative angle in deep water smaller than the one that maxi-
mizes the alongshore transport (Fig. 17b). Therefore, any pertur-
bation in the plan-view shoreline shape will grow if waves
approach at an angle greater than that which maximizes along-
shore transport (high-angle waves), as a result of a convergence of
sediment flux along the crest of the perturbation (Fig. 17c).

Both the scale of the features, much smaller in the case of El
Puntal, and the non-propagating condition of the coastline sand
waves at El Puntal are different from prior observations and
predictions carried out at other sites (Ashton et al., 2001, 2003;
Ashton and Murray, 2006b; Falqués and Calvete, 2005;
Falqué, 2006). However, because of the very peculiar local
wave climate at El Puntal, characterized by small-amplitude and
long-period waves (along with the quite steep underlying
bathymetry) the scales might be much shorter in this case. The
fact that these features might not be present during summer,
even though high-angle waves prevail all the time, could be due
to the smaller wave heights and periods during this season.
These waves might not be able to transport sufficient sediment
alongshore in order to cause the formation of sand waves.
Therefore, observations suggest that sand waves present at
El Puntal spit, and probably at other spits of the same
characteristics, might be mainly due to the instability in
sediment transport induced by high-angle moderate-energy
waves. Work is underway to implement existing models to test
the stability of the coastline at El Puntal.

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