Variability in storm climate along the Gulf of Cadiz: the role of large scale atmospheric forcing and implications to coastal hazards.

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Abstract

In the context of increased coastal hazards due to variability in storminess patterns, the danger of coastal damages and/or morphological changes is related to the sum of sea level conditions, storm surge, maximum wave height and run up values. In order to better understand the physical processes that cause the variability of the above parameters a 44 years reanalysis record (HIPOCAS) was used. The HIPOCAS time-series was validated with real wave and sea-level data using linear and vector correlation methods. In the present work changes in the magnitude, duration, frequency and approach direction of the Atlantic storms over the Spanish Gulf of Cadiz (SW Iberian Peninsula) were identified by computing various storm characteristics such as maximum wave height, total energy per storm wave direction and storm duration. The obtained time-series were compared with large-scale atmospheric indices such as the North Atlantic Oscillation (NAO) and the East Atlantic pattern (EA). The results show a good correlation between negative NAO values and increased storminess over the entire Gulf of Cadiz. Furthermore, negative NAO values were correlated with high residual sea level values. Finally, a joint probability analysis of storm and sea level analysis resulted in increased probabilities of the two events happening at the same time indicating higher vulnerability of the coast and increased coastal risks. The above results were compared with coastal inundation events that took place over the last winter seasons in the province of Cadiz.

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1. Introduction

Storm events are considered the main cause for shoreline change in many areas worldwide (Fenster et al., 2001). Depending on the magnitude of the event and the morphological characteristics of the coastline the change can be transient or persistent (Anderson et al., 2010). Hence, storms are attributed an essential role in coastal long-term evolution (decadal and centennial) despite the usually short time scale of their action (Morton et al., 1995). Nowadays there is a growing socioeconomic need towards innovative coastal management and evaluation of the risks associated with the development in coastal plains (Van Dongeren et al., 2014). The combined effect of storm activity and surge level changes can pose a risk on the coastal environment by eroding the upper beach and inundating any low lying backshore area. In coastlines exposed to long fetches, such as the Spanish Atlantic coast, large scale atmospheric phenomena are the main source of storminess variability. Hence, the seasonal and inter-annual variability or long term trends of the above processes can affect the risk distribution over a particular stretch of coastline.

Specific trends in wave heights have long been observed in the North Atlantic (Bacon and Carter, 1991; Carter and Draper, 1988; Dupuis et al., 2006; Feng et al., 2014b) and in the Northeast Pacific (Allan and Komar, 2000). Furthermore, Bacon and Carter (1991) observed a correlation between North Atlantic meridional atmospheric pressure gradient and wave height. Woolf et al. (2002) established a relationship between wave height anomalies and large scale atmospheric pressure patterns over the Northeast Atlantic on the basis of satellite altimetry. Finally, Dodet et al. (2010) presented a larger influence of the NAO over the south area of Europe using a 60 year long wave model forced by the 6h wind field from National Centres of Environmental Prediction (NCEP) reanalysis project (Kalnay et al., 1996).

Traditionally the NAO and other climatic indices have been mainly linked with temperature, precipitation and large scale circulation patterns. Recent studies have also focused on analysing the response of sea level variability to NAO, but always at a broad spatial scale (Efthymiadis et al., 2002; Tsimplis and Shaw, 2008; Tsimplis et al., 2006; Woolf et al., 2003). Main findings are that the NAO influence on sea level is dominant in winter and represents one of the causes of the high inter-annual variability of sea level during this season. Moreover, Woolf et al. (2003) suggest that NAO effects are probably similar in the open ocean and along coastlines in large geographical areas, although they might sometimes be masked by local phenomena.

Regional correlations between mean monthly wave height and NAO values have been performed by various researchers (Bertin et al., 2013; Dodet et al., 2010; Feng et al., 2014a; Rangel-Buitrago and Anfuso, 2013; Woolf et al., 2003). The impact of storms on shoreline variability has been widely demonstrated (e.g. (Cooper et al., 2004; List et al., 2006; Morris et al., 2001). However, a direct relationship between inter-annual wave variability and coastal response has not been linked to NAO because of the general lack of long-term and detailed coastal topographic data and the influence of other local and regional aspects on coastal changes, such as geological framework (Jackson et al., 2005). Recently O’Connor et al. (2011) demonstrated a tentative link between coastline topography and NAO-modulated external forcing, by focusing in small and well constrained tidal inlets of northwest of Ireland. Thomas et al. (2011) derived a negative correlation between beach rotation and volume with the NAO for the southwest Welsh coast. In the same way, using the SOI climatic index and the data-set of Narrabeen Beach in Australia a relation between beach rotation was linked to the variability of wave characteristics (Harley et al., 2009; Ranasinghe et al., 2004; Short et al., 2000).

Although it is well known that NAO affects the latitudinal aspects of storm tracks variability over the Atlantic (Keim et al., 2004; Rogers, 1997) an in-depth investigation of the joint influence on the wave storminess and sea surface height variability and the associated coastal hazards in the Gulf of Cadiz has not yet been undertaken. The present work focuses in the coastal area of the northern part of the Gulf of Cadiz and goes beyond the separate analysis of wave and residual sea level as it investigates the combined occurrence of the phenomena and their contribution to the increasing severity of coastal hazards. Furthermore, other aspects of wave storminess are examined such as the storm significant wave height, the storm direction and the total amount of storm hours in a month. A description of the study area, wave and sea level data sets is presented, followed by a section focused on the validation exercise and correction fittings that were employed in order to improve the results of the hindcast models for storm conditions. Then the inter-annual variability of the record is presented and correlations with large scale climatic indices are undertaken. Finally, discussion and conclusions of the results obtained and comparison with other work are detailed.

2. Study area

The Gulf of Cadiz is the sub-basin that connects the Atlantic Ocean with the Mediterranean Sea through the Strait of Gibraltar. Its northern and southern boundaries are, respectively, the southwest coast of the Iberian Peninsula and the Atlantic coast of Morocco (Figure 1). As an Atlantic exposed coast it is influenced by large scale oceanic weather systems that cross the North Atlantic following an eastward path, that determine the patterns of precipitation, wind and long-fetch waves. The storms generated by these systems are the principal cause of transient erosion in the area (Del Rio, 2007). On a local scale, the orientation of the coastline and the local physiographic characteristics result in sheltering effects to the north component winds and funnelling effects to south and east component winds due to the complex orography of the Strait of Gibraltar (Dorman et al., 1995). The prevailing wind and wave fields are from WSW directions (Figure 1), with a yearly average significant wave height of 1 m comprised of both sea and swell, generating a predominant longshore current towards the E and SE (Benavente et al., 2000). Waves with north component are not frequent within the Gulf of Cadiz due to the sheltering effect of the Cape St. Vincent (Figure 1) where significant diffraction and attenuation takes place. In general the northern Gulf of Cadiz is characterized by a lower energy wave climate than the western coastlines of the Iberian Peninsula (Loureiro et al., 2013). Storm events are generally frequent over the autumn and winter months with significant wave height values reaching up to 7m. The less frequent wave storm events in the Gulf of Cadiz have easterly directions (wave rose in Figure 1) but they do not produce high waves to the north-eastern part of the Gulf due to their limited fetch. Due to the coastline orientation, these limited fetch storm waves affect only the north-western part of the Gulf of Cadiz.

In terms of tides the area can be described as semidiurnal with mean tidal range of 2.20 m decreasing towards the Strait of Gibraltar. Changes in shoreline orientation along the Gulf coast greatly influence the approach angle of waves, which diminishes progressively towards the Southeast, generating less significant littoral currents close to the Strait of Gibraltar and weaker longshore drift (Medina, 1991). The surface circulation over the continental shelf is mainly wind-driven but it is also affected by local forcing mechanisms, such as the Guadalquivir River discharge, and is subject to seasonal and inter-annual variations deeply related to the seasonal variability of the open sea circulation (Criado-Aldeanueva et al., 2009). The latter is greatly affected by the large-scale atmospheric patterns over the Atlantic Ocean, roughly represented by the NAO index (Criado-Aldeanueva et al., 2009).

3. Methodology

3.1 Data description

In the present work the nearshore nodes of a hindcast dataset (HIPOCAS data hereinafter), located over the north coast of the Gulf of Cadiz were used. The dataset consists of a 44-year reanalysis of meteorological, wave variables and sea level spanning between January 1958 and December 2001 (Sebastiao et al., 2008). For the wave simulation over the Gulf of Cadiz a grid of 5’ was linked to a larger WAM model (WAMDI-Group, 1988) of the Atlantic Ocean through a series of nested models (Gomez Lahoz and Carretero Albiach, 2005). The wave model was initially forced by the NCEP reanalysis wind fields. The ocean circulation and sea level variations were simulated with the HAMSOM model over a grid of 10’x 15’ taking into account wind and pressure forcing. The hydrodynamic model was forced with boundary conditions from the REMO atmospheric model (Sebastiao et al., 2008) that was in turn forced with NCEP data. The sea-level data represent the atmospherically-induced contribution and the associated storm surges and do not have a tidal signal, or the contribution of the steric effect on sea-level (Sebastiao et al., 2008). The output time-step was 3 hours for all the parameters. From the above data set five stations were selected that cover the entire north coast of the Gulf of Cadiz (Figure 1). From west to east the selected stations are: Faro, that represents the most exposed part of the Gulf of Cadiz with a narrow shelf and a steep continental slope; Huelva and Seville, located further to the east, at the widest part of the continental shelf and partially sheltered from the west and north component winds by the Cape St. Vincent; Cadiz station is located further to the southeast where the shelf width starts to reduce and the coastline is more exposed to the Atlantic storms; finally the Zahara station is largely dominated by the Gibraltar Strait conditions and is characterized by the absence of continental shelf and reduced tidal range. In terms of wave data the coastal buoy of Cadiz managed by Spanish Port Authorities and the buoy of Faro managed by the Portuguese Hydrographical Office (Figure 1) were selected for validation purposes since they provided directional wave measurement overlapping with the model data.

3.2 HIPOCAS data validation

An extensive validation exercise was undertaken by Mendez et al. (2006) between the HIPOCAS wave data and wave buoy data collected around the Spanish coasts. However, in order to optimize the results in the Gulf of Cadiz, a new correction was applied in the present study that consisted in: (i) fitting the model wave height to observations focusing mainly in the case of storm conditions; (ii) evaluating differences between model and measured wave directions using a vector correlation approach (Kundu, 1976). The wave height validation was evaluated by calculating the bias and the Brier Skill Score (BSS). The latter parameter relates the variance of the difference between data and model with the variance of the data. BSS=1 means perfect skill, BSS=0 means no skill (Roelvink et al., 2009). The wave direction validation was evaluated based on the Kundu coefficient and mean angle rotation (θ). Both wind waves and swell were analysed together since they coexist during storm events and no spectral information was available.

In order to obtain a statistically independent wave height data set of storm conditions a peak over threshold analysis (POT) was used for the period with simultaneous model and observation data (Kamphuis, 2000). The above analysis allowed producing a correction based on the peak values of each storm and was then applied to the entire data set. The threshold value for the POT analysis was set as 1.5m wave height (a threshold value proposed by the local authorities for civil protection), and a storm independence criteria (time between two consecutive independent storms) was calculated based on the integral time scales of the autocorrelation function (Emery and Thomson, 2001). Higher storm threshold values proposed for the area of Cadiz (Almeida et al., 2012; Del Rio et al., 2012; Ribera et al., 2011) were also used but the fitting coefficients obtained were not significantly different.

The corrected values of significant wave height (Hsc) improved the time series extracted from the reanalysis data for storms with Hs higher than 3m, where the HIPOCAS data had significantly and systematically overestimated the wave height approximately by 30%. A single model was constructed from both data series with a correlation coefficient of r=0.75 and applied to all stations. Similar correlation coefficients were also obtained for only the Faro data by Almeida et al. (2011). Separate analysis of Cadiz and Faro buoys resulted in correlations that had statistically no significant differences. Typical results are shown in Figure 2 for March 1995 for both stations, where it can be observed that the corrected HIPOCAS data show better agreement during storm conditions with the buoy data. For the data set relative bias values of 0.02 instead of 0.29 for the uncorrected data and BSS of 0.60 instead of 0.43 were obtained for the buoy of Faro. For the case of Cadiz buoy similar values were calculated with relative bias decreasing from -0.35 to -0.02 and BSS increasing from 0.41 to 0.66.

The agreement between the original (uncorrected) model data and buoy data, in both Faro and Cadiz wave buoys, was tested for wave height and direction simultaneously using the Kundu (1976) vector correlation approach. This correlation method produces a coefficient between the directional wave heights of the two time series and the main angle (θ) through which the first series would have to be rotated anticlockwise to match the direction of the second series. Although overlapping of directional wave data between the buoy of Cadiz and the HIPOCAS data only exists over part of 2001, the period is long enough to cover both calm and stormy conditions. For the case of Faro the overlapping period was much longer (1997-2001). The correction coefficients calculated for the two time series were not statistically different; hence a common correction equation was derived for both sites and then applied to all the wave data.

Differences in wave propagation direction between the measured and modelled data are presented against the significant wave height in Figure 3. There is a large scatter in directions for wave heights lower than 1m in both sites. However, as it can be seen from the point's density, large differences are only present over few events, while the majority of the data show small deviations. Such deviations are reduced to a variance of 45 deg for waves between 1-1.5m, also with larger densities concentrated in small differences. Data from Faro present slightly larger deviations for the lower wave heights. This is probably related to the diffraction processes that waves undergo at the Cape St. Vincent, which are probably not fully resolved by the model resolution. For the higher wave heights the directional scatter is minimized and the HIPOCAS data present a good agreement with the observed data.

Using the vector correlation approach, the correlation coefficient obtained for the Cadiz buoy was 0.71 with an average angle difference of 4.7 deg. For Faro the correlation obtained was 0.76 with an average angle difference of -5.2 deg. In Figure 4 the E-W (zonal) and N-S (meridional) components of the waves are plotted for both sites. A good agreement is observed between both components, particularly for large storm events of southwest direction, which is the main oceanic storm approach direction. Similar results were also observed by Ribera et al. (2011) for the same data set. The locally generated small storms with predominant southeast directions are not well represented by the model probably due to spatial constraints of the atmospheric forcing that cannot resolve the local east wind acceleration over the Strait of Gibraltar (Figure 4, around 11/08/2001) and the short fetch of those waves. These locally generated storms are not affecting the eastern coastline of the Gulf of Cadiz, because of the short fetch and its general orientation. However, the above events can generate coastal erosion events further west, over the coast of southern Portugal (Garcia et al., 2005).

No validation to the sea level data was applied here because the calibrated time series (Sebastiao et al., 2008) are the only available in the area for the reanalysis period. The sea level reanalysis carried out for the HIPOCAS data along the coasts of the Iberian Peninsula presented good results (Sebastiao et al., 2008). For example, in the area of the Gulf of Cadiz (tidal station Seville) the results were under-predicting the observations with a RMSE of 12.61only for the extreme peak of the storms. This discrepancy could be due to the tidal gauge location (close to estuary mouth) where the measured water levels are locally affected by the river discharge; however, this do not influence the general surge level on the continental shelf (Laiz et al., 2013).

3.3 Data analysis

The corrected time series were used to calculate the monthly, seasonal and annual values of the wave heights and directions and sea level in order to identify the wave climatology in the area. Furthermore, storminess indices were calculated on a monthly basis in order to be compared with climatic indices. The wave storminess indices obtained were: the number of individual storms per month (storm number, SN); the storm significant wave height H_{st} which corresponds to the monthly average ($\overline{H_s}$) of wave heights (H_s) for the values above the threshold (H_{th})

$$H_{st} = \overline{H_s}\,|H_s > H_{th}$$

Finally, the number of hours when the storm threshold was exceeded divided by the total number of hours per month (Storm Duration Ratio, SDR) was also calculated

$$SDR = \frac{\sum_{i=1}^{N} t_i | H_i > H_{th}}{\sum_{i=1}^{N} t_i}$$

where N is the number of model output per month and H_{th} is the storm threshold. For both indices the wave height threshold and storm independence criteria remained the same as in the POT analysis. SDR is a value similar to the percentage of run length introduced by Feng et al. (2014a). Additionally the total storm energy per month was calculated as the sum of wave energy above the storm threshold.

The computed indices were correlated with the climatic indices that influence weather patterns over Europe, namely the North Atlantic Oscillation (NAO), the East Atlantic Pattern (EA), the Scandinavia Pattern (SCAN) and the Polar/ Eurasia Pattern (POL). The monthly data corresponding to the HIPOCAS dataset were obtained from NOAA climate centre and were calculated from the rotated EOF of the 500 hPa geopotential height. The SCA and POL did not present any significant correlation; hence the results are not presented. Significant differences between two correlation coefficients were tested using the Fisher r-to-z transformation (Fisher, 1970). This method converts first each correlation coefficient into a *z*-score. Then, making use of the sample size employed to obtain each coefficient; these *z*-scores are compared.

Furthermore, a detailed study of surge levels given a specific wave height ($f_{\mathrm{S|Hs}}$) was studied at a storm event timescale. For the joint probability estimation of the residual sea level (SLres) and Hsc were jointly used in POT analysis. For each storm event identified by the POT analysis the peak Hsc and peak SLres were selected in order to construct a contingency table. Tidal variations were not taken into account since in the Iberian Peninsula tidal–surge energy transfer is low (Ratsimandresey et al., 2008). Combined wave and SLres probability function ($f_{\mathrm{SLres,Hs}}$) events was undertaken parting from the assumption that the surge elevation probability function (f_{S}) is not independent of the significant wave height probability function (f_{Hs}) for waves above the threshold.

$$f_{Hs,RSL}(Hs,RSL) = f_{RSL|Hs}(RSL|Hs).f_{Hs}(Hs)$$

For the calculation of join probability the initial POT analysis results were used in order to construct a contingency table with storm events and the associated SLres.

4. Results and Discussion

4.1 Wave and Residual Sea Level Climate

The mean annual cycle for the corrected significant wave height (Hsc) and the associated wave directions for the coastal area of the northern Gulf of Cadiz are presented in Figure 5a and 5b. The average wave heights over the area are higher during the winter months and part of the autumn. From the comparison of corrected significant wave heights (Hsc) and their associated directions it can be seen that there is a high energy period starting in November and extending up to March, when the mean wave heights are higher (Hsc ≈ 1m) and from more southerly directions. The same pattern is present for all stations with some differences in Huelva and Seville where the mean significant wave height is lower and mean direction of propagation has a stronger southern component, due to the shelter effect of the Cape St Vincent. These values classify the coastline as a mixed energy, wave-dominated coast according to (Davis and Hayes, 1984). Over the rest of the year mean wave height is significantly lower with more westerly directions. This annual variability represents the typical wave climatology of the region. Based on the above results the wave climate in the Gulf of Cadiz can be separated in a storm (November - March) and a calm (April - October) season. These results are in agreement with previous wave climate studies (Dodet et al., 2010; Lozano et al., 2004) on the area and with the seasonal morphological behaviour of the beach (Benavente et al., 2002). In terms of peak period (T_p, Figure 5c) the seasonal pattern is repeated with higher values of T_p during the storm season and lower ones during the calm season. All stations present the same mean monthly values with the only difference of Huelva (the most protected station) where T_p values are smaller.

On the other hand the monthly residual sea level (SLres) variation (Figure 5d) presents an inverse image, with higher SLres during the calm period and lower values during the storm period. These results are in agreement with previous analysis of SLres records in the area that show a seasonal cycle with a minimum in February and a maximum in October, consistent with atmospheric pressure forcing in this region (Laiz et al., 2013). The SLres annual signal ranges between about 4–6/5–6 cm from tide gauges/altimeter measurements, respectively, the latter always showing slightly larger amplitudes (Gomez-Enri et al., 2012; Laiz et al., 2013; Marcos et al., 2011; Marcos and Tsimplis, 2007).

4.2 Correlations with climate indices

The inter-annual variability of the storm significant wave height and storminess indices did not show significant correlation with climatic indices. These results have been reported before for the Faro station and NAO (Almeida et al., 2011) and seem to extend over the entire Gulf of Cadiz. Other studies have concluded that the inter-annual variability of Hs is partially controlled by NAO (Dodet et al., 2010; Woolf et al., 2002) but Bertin et al. (2013) show that this relationship has a large spatial variability that results in non-significant correlations over the Gulf of Cadiz. Also the same studies concluded that no linear trend is present for Hs over the study area. This paper is focusing on the monthly variability both for the storm and calm period as well as their climatological anomalies.

Both the storm mean monthly values (mean value of significant wave height that exceeds the storm threshold, 1.5 m) and their anomalies for the whole year and for the storm seasons were tested against all indices that are linked with climate variability over Europe, in order to investigate possible correlations with the wave parameters and SLres in the Gulf of Cadiz. Significant correlations were obtained for the NAO and EA pattern. Data showed correlation between the annual Hsc and NAO for all sites with values between -0.27 for Faro and -0.34 for Seville, increasing for the storm season months to values of around -0.58 (Table 1, Figure 6). However, the correlations between the mean monthly anomalies and NAO show significant increase ($p<0.01$) for the full year where the correlation coefficient almost doubles, suggesting that NAO has an effect on the wave variability that prevails during calm season as well. This variability could be due to waves formed by local wind or to swell field being determined by the NAO over the North Atlantic. This can have implications on the medium term shoreline evolutions since the wave period characteristics during the calm season are responsible for the beach accretion and recovery after the storms. For the storm season the anomaly correlations are also higher but differences are small and not statistically significant. In terms of spatial distributions there are no significant differences observed for the various locations between NAO and the annual Hs for all cases. The physical explanation of the above correlations could be attributed to the higher abundance of low pressure systems and the increase in wind speed linked with the NAO and attributed to the southern shift of the Atlantic storm tracks (Dodet et al., 2010; Keim et al., 2004).

Significant correlations but with low correlation coefficient values were observed between NAO and monthly wave directions for the full year; the best correlation was observed in Huelva but the coefficient did not exceed -0.20. Over the storm season correlations did not improve but the significance levels decreased due to the lower number of observations (220 instead of 528) that influences the degrees of freedom of the correlations. The spatial distribution again did not show any significant variation.

The residual water levels also presented strong correlations with NAO especially during the storm season months, showing a high correlation value between mean storm residual water levels and NAO (-0.66 on average, $p<0.01$). These results are in agreement with previous studies in the area using the same reanalysis data for southern Europe (Marcos et al., 2009). The main mechanism that drives the residual sea level response to the NAO is both hydrostatic and non-hydrostatic (Woolf et al., 2003). Considering this, the correlation between mean storm residual water level and NAO is possibly related to the fact that the HIPOCAS dataset includes both processes, as it was modelled using a barotropic version of the HAMSOM model (Ratsimandresy et al., 2008). All correlations have a negative sign because wave height and residual sea level increase with negative NAO values (Table 1).

For the case of EA the above parameters explain a small but statistically significant part of the variability (Table 2) among which the most pronounced is that of the significant wave height during the storm months. In all cases the correlation coefficients were smaller than with NAO, as expected, since the EA is the second prominent mode of low-frequency variability over the North Atlantic (Barnston and Livezey, 1987). For this index the correlations are positive for wave height and direction but negative for SLres. Positive EA values are responsible for zonally extended storm tracks that affect the southern coasts of Europe (Wettstein and Wallace, 2010). Significant differences between stations were only present for wave height correlations, where the lowest value obtained (Huelva) was significantly different from the rest of the stations.

Correlations between wave directions during storms were stronger in the case of EA than with the NAO. Correlations were positive with mean values around 0.20 to 0.3 and were constant both for mean values and anomalies (Table 2). Apart from these correlations, EA index presented negative correlations (between -0.36 and -0.46) with the wave direction standard deviation during the storm season (Figure 7). These correlations suggest a focusing of the storm around west direction during the positive EA and a greater dispersion during negative EA. This dispersion is represented in the Gulf of Cadiz with increased southern direction because of the negative skewness inherited to the data by the coastline orientation.

Because the NAO and the EA represent different modes of the atmospheric variability i.e. the percentage of variability expressed by the two indices is uncorrelated, the above results suggest that storminess during negative NAO and positive EA phases can be further increased. However, on a seasonal scale the indices can be correlated; hence, part of the explained variance can be common (Martinez-Asensio et al., 2014). The physical mechanism during NAO-negative and EA-positive phases is that the orientation of the boundary between the positive and negative pressure anomalies crosses the North Atlantic from northwest (60N, 60W) to southeast (45N, 10W), which is likely to influence the meridional circulation intensity (Nesterov, 2009) and direct the storm tracks towards south Europe and into the Gulf of Cadiz. Furthermore, average wind during winter NAO-negative and EA-positive phases reveals patterns of westerly wind (Martinez-Asensio et al., 2014) that can induce a net mass flux in the Gulf of Cadiz and at the same time generate increased Hs (Fukumori et al., 2007). Similar average wind patterns are also generated during positive SCAN phases but with a more pronounced northern component; however, any generated waves by this wind pattern are not affecting the northern part of the Gulf of Cadiz due to the sheltering effect of the Cape St. Vincent.

Apart from the mean monthly values and anomalies, specific storm indices calculated above were also correlated with the climatic indices. Similar correlations (-0.52 and -0.43, $p<0.01$) were obtained between the SDR, which is a measure of the total time of the Atlantic oriented storm activity per month, and the NAO for the storm season (Figure 8). Higher values are observed for the more exposed stations (Faro, Cadiz and Zahara). Special cases for the Atlantic storms were selected by gradually restricting the direction of the incoming storm to pure westerly directions (data not shown). For these cases the correlation coefficients remain practically unchanged in all stations except Seville and Huelva, where the number of events drops dramatically due to the sheltering effect. The Storm number also presented similar patterns to the SDR both in terms of NAO correlations and spatial variability (data not shown).

The monthly storm wave height (H_{st}) obtained from individual storms produced a weaker correlation (-0.41, $p<0.01$) with NAO over the storm season. Similar results were obtained for the total energy of the storm waves for each month where the correlation coefficient with NAO was -0.45 ($p<0.01$). However the correlation between SDR and the mean monthly Hs was of the order of -0.90 for the study area, with no significant variations between the stations. The above results suggest that although negative NAO values increase the storminess over the study area, they do not control the magnitude of the wave height which is probably affected by synoptic atmospheric patterns. On the other hand there is a correlation with the number and total storm duration arriving to the Gulf of Cadiz. Similar results were presented for the Norwegian Sea where no statistical correlation was obtained between NAO and waves with low probability of exceedance (largest waves) (Feng et al., 2014b); however it has to be noted that NAO correlations with the largest 1% of Hs can reach $r=0.83$ over the Northwest of Scotland (Wang and Swail, 2001; 2002) but these correlations present large spatial variation over the North Atlantic.

4.3 Joint Probability

Despite the opposite seasonal pattern followed by the Hsc and the SLres presented in Figure 4, sea level values during storms were found to deviate substantially from the average seasonal cycle especially during the storms. The joint probability between these parameters is presented in Figure 9, with similar patterns observed at all stations. A large proportion of the events identified by the POT (50%) correspond to relatively low energy events (<2.5 m Hsc), for which the SLres showed a large spread that is mainly concentrated in positive values between 0 and 15cm. For the rest of the events a clear trend is obtained where higher wave heights are observed together with positive SLres, with values up to 35 cm for the extreme wave height events of 4-6.5 m that have a return period in the area of Cadiz between 3 and 4 years respectively (Del Rio et al., 2012).

The stations of Seville and Huelva that are situated at the shadow zone of the Cape St. Vincent receive less storm activity both in terms of number and magnitude but follow the same probability patterns as the rest of the area. Besides, this zone shows the highest values of storm surge probably due to the larger width of the continental shelf. In contrast the areas of Zahara and Faro that are characterised by a relatively narrower continental shelf present much lower surges for the similar or higher wave heights. These results show dependence between the SLres and the peak storm Hsc and have wide implications on the coastal hazards and the associated risk of coastal erosion and inundations of the coastal plain (Del Rio et al., 2012).

In accordance with the correlation presented above between the storm variables and NAO, the joint probability analysis was undertaken separately for positive and negative NAO events. In general the ratio between storm events occurred during a positive NAO and events occurred during a negative NAO phase is close to 1 for all sites (Table 3). For NAO phases with an index higher/lower than +/-1 and +/-1.5 it can be seen that the negative NAO phases present almost twice the events than the positive ones for the central part of the Gulf (Seville and Huelva). On the contrary the two sites located at the extremities of the Gulf of Cadiz (Faro and Zahara) do not show this pattern. This is partly due to the strong easterly winds that can also create short-fetch storms for these areas, such events are more frequent during positive NAO (Dorman et al., 1995). These events are present in the wave record of Zahara due to the proximity to the Strait of Gibraltar and in Faro due to the orientation of the coastline and the considerable easterly fetch.

Differences are not present in extreme NAO phases (-2 > NAO > +2) probably due to the small number of events (Table 3).

The joint probability analysis for positive and negative NAO events with index higher/lower than ±1.5 is presented in Figure 10, where it can be seen that during strong positive and negative phases of NAO the joint probability of the wave–surge follows a different pattern. Positive NAO events (Figure 10 left panels) are concentrated in weak storm events (Hsc<2.5m) with mainly small SLres. On the other hand, for the negative NAO events (Figure 10 right panels) the same pattern that was observed in the full data analysis is repeated with a positive trend between storm wave heights and SLres. The above results corroborate with the NAO correlations of Table 1; where during storm season Hsc and SLres show significant correlations with NAO.

The joint probability results emphasize the importance of the NAO on coastal hazards. The Hs and storm surge height drive the morphological evolution and coastal hazard estimation in the Gulf of Cadiz shores according to Del Rio et al. (2012). This way, severe coastal erosion and flooding events have been recorded in the area during negative NAO phases, with a great socioeconomic impact. This impact is related to both direct damage to coastal infrastructure and undesirable morphological changes in the coastal area, such as long-term reduction in beach width or damage to dune ridges (Del Rio et al. 2012). One of the most significant periods in this respect occurred in the 2009-2010 winter season, when a peak in negative NAO index over the last 190 years was recorded (Osborn, 2011). In that period a number of energetic storm events caused widespread beach and dune recession and coastal flooding along the Gulf of Cadiz (Benavente et al., 2013; Del Rio et al., 2010; Rangel-Buitrago and Anfuso, 2013; Vousdoukas et al., 2012). Maximum wave heights of up to 8.4m were observed and SLres up to 0.50m were recorded at the tidal station of Cadiz. In terms of Hs the storm had a return period of 20years; however, taking into account the prolonged duration of the storm (more than 20 days of storm conditions) the total event as a group of storms should have a much greater return period according to Ferreira (2006). The associated damage on coastal assets generated important economic losses, which for instance in Cadiz city beaches were around 157,000 € (Lopez-Doriga et al., 2010) For this reason, the fact that Hs and storm surge height have a high joint probability especially during the negative NAO phases is of great importance and can be very useful at the design stage of coastal protection systems and civil protection plans.

5. Conclusions

Reanalysis of wave and sea level data for a period of 44 years (HIPOCAS data) were used to investigate the connection between large scale atmospheric circulation patterns (NAO, EA) and the wave climate and sea level in the Gulf of Cadiz. Significant improvement of the storm wave record was obtained after applying correction functions derived from the coastal wave buoys in the area of Cadiz and Faro. In general, the HIPOCAS data correctly represented the directional storm climate in the Gulf of Cadiz and mainly the one coming from the northwestern Atlantic. The locally accelerated easterly winds in the area of the Strait of Gibraltar are not well represented due to the relative large scale of the re-analysis data. However, these events are not affecting storm-related hazards along the coastline of the study area except for the zone of Faro, as they are mostly related to high atmospheric pressure situations; furthermore, shoreline orientation and short fetch determine a negligible impact of easterly waves along the northeastern coast of the Gulf of Cadiz.

In terms of wave activity two seasons can be distinguished: the storm and the calm season. The former extends from November to March and shows higher mean monthly significant wave height and distinguishable wave period and direction than the calm season. Based on these results further analysis was undertaken following the above seasonal pattern and not the atmospheric season convention. NAO presented negative correlations with the monthly parameters of the storm season. When the mean wave climatology was subtracted from the data this correlation was extended to the entire year (anomalies) suggesting influence of the NAO to the calm wave conditions . Positive correlations were obtained with the EA pattern that probably represents the zonal extension of the storm tracks over the study area during positive EA phases. Better correlations were identified for the total storm hours (Storm Index) and the residual mean sea level but not with the maximum wave height. The above results suggest that although negative NAO values increase the storminess over the study area they do not control the magnitude of the wave height, which is probably affected by mesoscale atmospheric patterns. The combined NAO and EA patterns explain a large part of the mean wave variability, also positive EA patterns are correlated with more westerly directions of the storm waves.

Joint probability analyses showed dependence between storm conditions and positive residual mean sea level on the basis of 367 events. This dependence is more pronounced over storm events with large wave heights. Study of storm events over distinct NAO index values showed a dominance of storm events during negative NAO phases. At extreme negative NAO phases the coexistence of large SLres and large storm events are present. This is not the case in positive NAO phases, where small storm events are present with disperse SLres response. In terms of coastal hazards and risk the coexistence of storm events and high SLres can potentially increase the vulnerability of the coastal areas to erosion and/or flooding episodes. The fact that these two parameters have a high joint probability especially during the negative NAO phases is of great importance and can be very useful at the design stage of coastal protection systems and civil protection plans. Furthermore, such result provides valuable information for understanding and reconstructing the long-term coastline evolution of the Gulf of Cadiz due to the long record of NAO index.

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6. References

Allan, J.C. and Komar, P.D., 2000. Are ocean wave heights increasing in the eastern North Pacific? Eos, Transactions, American Geophysical Union, 47: 561-567.

Almeida, L.P., Ferreira, O., Vousdoukas, M.I. and Dodet, G., 2011. Historical variation and trends in storminess along the Portuguese South Coast. Nat. Hazards Earth Syst. Sci., 11(9): 2407-2417.

Almeida, L.P., Vousdoukas, M.V., Ferreira, O., Rodrigues, B.A. and Matias, A., 2012. Thresholds for storm impacts on an exposed sandy coastal area in southern Portugal. Geomorphology, 143-144(0): 3-12.

Anderson, T.R., Frazer, L.N. and Fletcher, C.H., 2010. Transient and persistent shoreline change from a storm. Geophys. Res. Lett., 37(8): L08401.

Bacon, S. and Carter, D.J.T., 1991. Wave climate changes in the North-Atlantic and North-Sea International Journal of Climatology, 11(5): 545-558.

Barnston, A.G. and Livezey, R.E., 1987. Classification, Seasonality and Persistence of Low-Frequency Atmospheric Circulation Patterns. Monthly Weather Review, 115(6): 1083-1126.

Benavente, J., Del Río, L., Anfuso, G., Gracia, F.J. and Reyes, J.L., 2002. Utility of morphodynamic characterization in the prediction of beach damage by storms. Journal of Coastal Research, SI36,: 56-64

Benavente, J., del Rio, L., Plomaritis, T.A. and Menapace, W., 2013. Impact of coastal storms in a sandy barrier (Sancti Petri, Spain). In: Conley, DC, Masselink, G., Russell, PE and O'Hare, TJ (eds.), Proceedings 12th International Coastal Symposium, Journal of Coastal Research, Special Issue No. 65: pp. 666-671, ISSN 0749-0208.

Benavente, J., Gracia, F.J. and Lopez-Aguayo, F., 2000. Empirical model of morphodynamic beachface behaviour for low-energy mesotidal environments. Marine Geology, 167(3-4): 375-390.

Bertin, X., Prouteau, E. and Letetrel, C., 2013. A significant increase in wave height in the North Atlantic Ocean over the 20th century. Global and Planetary Change, 106(0): 77-83.

Carter, D.J.T. and Draper, L., 1988. Has the north-east Atlantic become rougher? Nature, 332(6164): 494-494.

Cooper, J.A.G., Jackson, D.W.T., Navas, F., McKenna, J. and Malvarez, G., 2004. Identifying storm impacts on an embayed, high-energy coastline: examples from western Ireland. Marine Geology, 210(1-4): 261-280.

Criado-Aldeanueva, F., Garcia-Lafuente, J., Navarro, G. and Ruiz, J., 2009. Seasonal and interannual variability of the surface circulation in the eastern Gulf of Cadiz (SW Iberia). Journal of Geophysical Research-Oceans, 114.

Davis, R.A. and Hayes, M.O., 1984. What is a wave-dominated coast? Marine Geology, 60(1-4): 313-329.

Del Rio, L., 2007. Riesgos de erosión costera en el litoral atlántico gaditano. Ph.D. Thesis, University of Cadiz, 485 p.

Del Rio, L., Plomaritis, T.A., Benavente, J., Valladares, M. and Ribera, P., 2012. Establishing storm thresholds for the Spanish Gulf of Cadiz coast. Geomorphology, 143-144(0): 13-23.

Del Rio, L. et al., 2010. The impact of two different storm seasons on a natural beach of the Gulf of Cádiz (Spain): high versus low energy events, Geophysical Research Abstracts 12, pp. 15118. EGU2010.

Dodet, G., Bertin, X. and Taborda, R., 2010. Wave climate variability in the North-East Atlantic Ocean over the last six decades. Ocean Modelling, 31(3-4): 120-131.

Dorman, C.E., Beardsley, R.C. and Limeburner, R., 1995. Winds in the Strait of Gibraltar. Quarterly Journal of the Royal Meteorological Society, 121: 1903-1921.

Dupuis, H., Michel, D. and Sottolichio, A., 2006. Wave climate evolution in the Bay of Biscay over two decades. Journal of Marine Systems, 63(3-4): 105-114.

Efthymiadis, D., Hernandez, F. and Le Traon, P.Y., 2002. Large-scale sea-level variations and associated atmospheric forcing in the subtropical north-east Atlantic Ocean. Deep-Sea Research Part Ii-Topical Studies in Oceanography, 49(19): 3957-3981.

Emery, W.J. and Thomson, R.E., 2001. Data Analysis Methods in Physical Oceanography (second edition revised). Elsevier Science, Amsterdam.

Feng, X., Tsimplis, M.N., Quartly, G.D. and Yelland, M.J., 2014a. Wave height analysis from 10 years of observations in the Norwegian Sea. Continental Shelf Research, 72(0): 47-56.

Feng, X., Tsimplis, M.N., Yelland, M.J. and Quartly, G.D., 2014b. Changes in significant and maximum wave heights in the Norwegian Sea. Global and Planetary Change, 113(0): 68-76.

Fenster, M.S., Dolan, R. and Morton, R.A., 2001. Coastal storms and shoreline change: Signal or noise? Journal of Coastal Research, 17(3): 714-720.

Ferreira, O., 2006. The role of storm groups in the erosion of sandy coasts. Earth Surface Processes and Landforms, 31(8): 1058-1060.

Fisher, R.A., 1970. Statistical Methods for Research Workers. Oliver & Boyd.

Garcia, T., Ferreira, O., Matias, A. and Dias, J.A., 2005. Coastal Hazards Representation for Praia do Barril (Algarve, Portugal). Journal of Coastal Research: 28-33.

Gomez-Enri, J., Aboitiz, A., Tejedor, B. and Villares, P., 2012. Seasonal and interannual variability in the Gulf of Cadiz: Validation of gridded altimeter products. Estuarine Coastal and Shelf Science, 96: 114-121.

Gomez Lahoz, M. and Carretero Albiach, J.C., 2005. Wave forecasting at the Spanish coasts. Journal of Atmospheric & Ocean Science, 10(4): 389-405.

Harley, M.D., Turner, I.L., Short, A.D. and Ranasinghe, R., 2009. Interannual variability and controls of the Sydney wave climate. International Journal of Climatology, 30(9): 1322-1335.

Jackson, D.W.T., Cooper, J.A.G. and del Rio, L., 2005. Geological control of beach morphodynamic state. Marine Geology, 216(4): 297-314.

Kalnay, E. et al., 1996. The NCEP/NCAR 40-year reanalysis project. Bulletin of the American Meteorological Society, 77(3): 437-471.

Kamphuis, J.W., 2000. Introduction to Coastal Engineering and Management. Advanced Series on Ocean Engineering, Vol. 16. World Scientific Publishing, Singapore, 437 pp.

Keim, B.D., Muller, R.A. and Stone, G.W., 2004. Spatial and temporal variability of coastal storms in the North Atlantic Basin. Marine Geology, 210(1-4): 7-15.

Kundu, P.K., 1976. Ekman Veering Observed near the Ocean Bottom. Journal of Physical Oceanography, 6(2): 238-242.

Laiz, I., Gomez-Enri, J., Tejedor, B., Aboitiz, A. and Villares, P., 2013. Seasonal sea level variations in the gulf of Cadiz continental shelf from in-situ measurements and satellite altimetry. Continental Shelf Research, 53: 77-88.

List, J.H., Farris, A.S. and Sullivan, C., 2006. Reversing storm hotspots on sandy beaches: Spatial and temporal characteristics. Marine Geology, 226(3-4): 261-279.

Lopez-Doriga, U., Benavente, J. and Plomaritis, T.A., 2010. Natural recovery processes in an urban beach, La Victoria (Cádiz, SW Spain). In Hmimsa Y. et al. (eds.): Book of abstracts - 1er Colloque International Littoraux Méditerranéens: états passés, actuels et futurs.

Loureiro, C., Ferreira, O. and Cooper, J.A.G., 2013. Applicability of parametric beach morphodynamic state classification on embayed beaches. Marine Geology, 346: 153-164.

Lozano, I., Devoy, R.J.N., May, W. and Andersen, U., 2004. Storminess and vulnerability along the Atlantic coastlines of Europe: analysis of storm records and of a greenhouse gases induced climate scenario. Marine Geology, 210(1-4): 205-225.

Marcos, M., Jordá, G., Gomis, D. and Pérez, B., 2011. Changes in storm surges in southern Europe from a regional model under climate change scenarios. Global and Planetary Change, In Press, Corrected Proof.

Marcos, M. and Tsimplis, M.N., 2007. Variations of the seasonal sea level cycle in southern Europe. Journal of Geophysical Research-Oceans, 112.

Martinez-Asensio, A. et al., 2014. Impact of the atmospheric climate modes on Mediterranean sea level variability. Global and Planetary Change, 118(0): 1-15.

Medina, J.M., 1991. La flecha de El Rompido en la dinámica litoral de la costa onubense. Ingeniería Civil, 80, : 105-110.

Mendez, F.J., Menendez, M., Luceno, A. and Losada, I.J., 2006. Estimation of the long-term variability of extreme significant wave height using a time-dependent Peak Over Threshold (POT) model. Journal of Geophysical Research-Oceans, 111(C7).

Morris, B.D., Davidson, M.A. and Huntley, D.A., 2001. Measurements of the response of a coastal inlet using video monitoring techniques. Marine Geology, 175(1-4): 251-272.
Morton, R.A., Gibeaut, J.C. and Paine, J.G., 1995. Mesoscale transfer of sand during and after storms - Implications for prediction of shoreline movement. Marine Geology, 126(1-4): 161-179.
Nesterov, E.S., 2009. East Atlantic oscillation of the atmospheric circulation. Russian Meteorology and Hydrology, 34(12): 794-800.
O'Connor, M.C., Cooper, J.A.G. and Jackson, D.W.T., 2011. Decadal Behavior of Tidal Inlet-Associated Beach Systems, Northwest Ireland, in Relation to Climate Forcing. Journal of Sedimentary Research, 81(1): 38-51.
Osborn, T., 2011. Winter 2009/2010 temperatures and a record-breaking North Atlantic Oscillation index. Weather, 66(1): 19-21.
Ranasinghe, R., McLoughlin, R., Short, A. and Symonds, G., 2004. The Southern Oscillation Index, wave climate, and beach rotation. Marine Geology, 204(3β€“4): 273-287.
Rangel-Buitrago, N. and Anfuso, G., 2013. Winter wave climate, storms and regional cycles: the SW Spanish Atlantic coast. International Journal of Climatology, 33(9): 2142-2156.
Ratsimandresy, A.W., Sotillo, M.G., Carretero Albiach, J.C., Alvarez Fanjul, E. and Hajji, H., 2008. A 44-year high-resolution ocean and atmospheric hindcast for the Mediterranean Basin developed within the HIPOCAS Project. Coastal Engineering, 55(11): 827-842.
Ribera, P. et al., 2011. Reconstruction of Atlantic historical winter coastal storms in the Spanish coasts of the Gulf of Cadiz, 1929-2005. Nat. Hazards Earth Syst. Sci., 11(6): 1715-1722.
Roelvink, D. et al., 2009. Modelling storm impacts on beaches, dunes and barrier islands. Coastal Engineering, 56(11-12): 1133-1152.
Rogers, J.C., 1997. North Atlantic Storm Track Variability and Its Association to the North Atlantic Oscillation and Climate Variability of Northern Europe. Journal of Climate, 10(7): 1635-1647.
Sebastiao, P., Guedes Soares, C. and Alvarez, E., 2008. 44 years hindcast of sea level in the Atlantic Coast of Europe. Coastal Engineering, 55(11): 848.
Short, A.D., Tremblanis, A.C. and Turner, I.L., 2000. Beach oscillation, rotation and the Southern Oscillation, Narrabeen Beach, Australia, 27th International Conference on Coastal Engineering. ASCE, Sydney, pp. 2439-2452.
Thomas, T., Phillips, M.R., Williams, A.T. and Jenkins, R.E., 2011. Short-term beach rotation, wave climate and the North Atlantic Oscillation (NAO). Progress in Physical Geography, 35(3): 333-352.
Tsimplis, M.N. and Shaw, A.G.P., 2008. The forcing of mean sea level variability around Europe. Global and Planetary Change, 63(2-3): 196-202.
Tsimplis, M.N., Shaw, A.G.P., Flather, R.A. and Woolf, D.K., 2006. The influence of the North Atlantic Oscillation on the sea-level around the northern European coasts reconsidered: the thermosteric effects. Philosophical Transactions of

the Royal Society a-Mathematical Physical and Engineering Sciences, 364(1841): 845-856.

Van Dongeren, A. et al., 2014. RISC-KIT: Resilience-Increasing Strategies for Coasts - toolkit. In: Green, A.N. and Cooper, J.A.G. (eds.), Proceedings 13th International Coastal Symposium (Durban, South Africa), Journal of Coastal Research, Special Issue No. 70: 366-371.

Vousdoukas, M.I., Almeida, L.P.M. and Ferreira, O., 2012. Beach erosion and recovery during consecutive storms at a steep-sloping, meso-tidal beach. Earth Surface Processes and Landforms, 37(6): 583-593.

WAMDI-Group, 1988. The WAM Model - A Third Generation Ocean Wave Prediction Model. Journal of Physical Oceanography, 18(12): 1775-1810.

Wang, X.L. and Swail, V.R., 2001. Changes of Extreme Wave Heights in Northern Hemisphere Oceans and Related Atmospheric Circulation Regimes. Journal of Climate, 14(10): 2204-2221.

Wang, X.L. and Swail, V.R., 2002. Trends of Atlantic Wave Extremes as Simulated in a 40-Yr Wave Hindcast Using Kinematically Reanalyzed Wind Fields. Journal of Climate, 15(9): 1020-1035.

Wettstein, J.J. and Wallace, J.M., 2010. Observed Patterns of Month-to-Month Storm-Track Variability and Their Relationship to the Background Flow*. Journal of the Atmospheric Sciences, 67(5): 1420-1437.

Woolf, D.K., Challenor, P.G. and Cotton, P.D., 2002. Variability and predictability of the North Atlantic wave climate. J. Geophys. Res., 107(C10): 3145.

Woolf, D.K., Shaw, A.G.P. and Tsimplis, M.N., 2003. The influence of the North Atlantic Oscillation on sea-level variability in the North Atlantic region The Global Atmosphere and Ocean System, 9(4): 145-167.

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Table 1: **Comparison of the correlation coefficients between mean monthly values and anomalies and NAO index for wave height (Hsc), wave direction (Dir) and residual sea level (SLres). Significance levels are <99%.**

		H_{annual}	H_{storm}	Dir_{annual}	Dir_{storm}	$SLres_{annual}$	$SLres_{storm}$
Cadiz	Mean Values	-0.32	-0.58	-0.11	-0.136*	-0.37	-0.67
	Anomalies	-0.43	-0.59	-0.12	-0.133*	-0.42	-0.70
Faro	Mean Values	-0.27	-0.54	-0.09*	-	-0.36	-0.66

	Anomalies	-0.42	-0.56	-0.11*	-	-0.42	-0.70
Zahara	Mean Values	-0.28	-0.54	-0.10	-0.17	-0.37	-0.67
	Anomalies	-0.42	-0.56	-0.12	-0.17	-0.42	-0.70
Seville	Mean Values	-0.34	-0.60	-0.06*	-	-0.38	-0.65
	Anomalies	-0.44	-0.61	-	-	-0.42	-0.70
Huelva	Mean Values	-0.31	-0.61	-0.16	-0.22	-0.38	-0.67
	Anomalies	-0.43	-0.62	-0.20	-0.23	-0.42	-0.70

* Significance level of <95%

Table 2: **Comparison of the correlation coefficients between mean monthly values and anomalies and EA for wave height, wave direction and residual sea level. Significance levels are <99%.**

		H_{annual}	H_{storm}	Dir_{annual}	Dir_{storm}	$SLres_{annual}$	$SLres_{storm}$
Cadiz	Mean Values	0.20	0.37	0.17	0.25	-0.16	-
	Anomalies	0.25	0.35	0.18	0.26	-0.16	-
Faro	Mean Values	0.20	0.35	0.19	0.35	-0.18	-0.14*
	Anomalies	0.27	0.36	0.25	0.36	-0.18	-
Zahara	Mean Values	0.19	0.34	0.13	0.21	-0.17	-
	Anomalies	0.24	0.36	0.15	0.21	-0.17	-
Seville	Mean Values	0.20	0.34	0.12	0.19	-0.16	-
	Anomalies	0.23	0.35	0.15	0.19	-0.15	-
Huelva	Mean Values	0.13	0.23	0.15	0.32	-0.17	-
	Anomalies	0.14	0.24	0.20	0.33	-0.17	-

* Significance level of <95%

Table 3: **Number of storm events identified for different positive and negative NAO thresholds.**

	0	0.5	1	1.5	2

Cadiz	NAO +	230	145	70	32	11
	NAO -	250	174	87	39	7
	Ratio (+/-)	0.92	0.83	0.80	0.82	1.57
	Total	**480**	**319**	**157**	**71**	**18**
Faro	NAO +	306	199	101	50	20
	NAO -	270	172	82	33	7
	Ratio (+/-)	1.13	1.16	1.23	1.51	2.8
	Total	**575**	**371**	**183**	**83**	**27**
Zahara	NAO +	278	178	81	41	15
	NAO -	279	181	89	38	8
	Ratio (+/-)	1.0	0.98	0.91	1.08	1.87
	Total	**557**	**359**	**170**	**79**	**23**
Seville	NAO +	125	71	35	16	8
	NAO -	191	131	86	38	7
	Ratio (+/-)	0.65	0.54	0.40	0.42	1.14
	Total	**316**	**202**	**121**	**54**	**15**
Huelva	NAO +	173	101	58	24	10
	NAO -	209	149	83	38	6
	Ratio (+/-)	0.83	0.68	0.70	0.63	1.5
	Total	**382**	**250**	**141**	**62**	**16**

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Fig. 1: **Bathymetric map of the study area showing the HIPOCAS data points and the location of the coastal buoys of Cadiz and Faro. Superimposed wave rose presents the annual wave height (m) at the central part of the Gulf of Cadiz.**

Fig. 2: **Comparison between modelled (HIPOCAS), measured (buoy), and corrected data for significant wave height (Hs) for a) Cadiz and b) Faro. Station locations in Figure 1**

Fig. 3: **Comparison of the difference between mean wave direction for the corrected (HIPOCAS) and measured (buoy) data for 2001 for the buoys of Cadiz (top panel) and Faro (bottom panel). Colour scale represents the data density.**

Fig. 4: **Comparison between modelled (HIPOCAS), measured (buoy), and corrected for the North-South and East-West components of significant wave height for Cadiz (a, b) and Faro (c, d).**

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Fig. 9: **Observed joint occurrence of storm wave heights and residual mean sea level. Colour scale indicates the probability of occurrence.**

Fig. 10: **Observed joint probability distribution of storm wave heights and residual mean sea level for: (a) storm events during NAO>+1.5 and (b) storm events during NAO<-1.5. Colour scale represents the probability of occurrence.**