

Holocene sea level and climate interactions on wet dune slack evolution in SW Portugal: A model for future scenarios?

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Abstract

We examine the Holocene environmental changes in a wet dune slack of the Portuguese coast, Poço do Barbarroxa de Baixo. Lithology, organic matter, biological proxies and high-resolution chronology provide estimations of sediment accumulation rates and changes in environmental conditions in relation to sea-level change and climate variability during the Holocene. Results show that the wet dune slack was formed 7.5 cal. ka BP, contemporaneous with the last stages of the rapid sea-level rise. This depositional environment formed under frequent freshwater flooding and water ponding that allowed the development and post-mortem accumulation of abundant plant remains. The wetland evolved into mostly palustrine conditions over the next 2000 years, until a phase of stabilization in relative sea-level rise, when sedimentation rates slowed down to 0.04 mm yr⁻¹, between 5.3 and 2.5 cal. ka BP. Later, about 0.8 cal. ka BP, high-energy events, likely due to enhanced storminess and more frequent onshore winds, caused the collapse of the foredune above the wetlands' seaward margin. The delicate balance between hydrology (controlled by sea-level rise and climate change), sediment supply and storminess modulates the habitat's resilience and ecological stability. This underpins the relevance of integrating past records in coastal wet dune slacks management in a scenario of constant adaptation processes.

Keywords

diatoms, geochemistry, Holocene, peat, sedimentology, wet dune slacks

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Introduction

Wet dune slacks are 'low-lying area[s] within the coastal dune systems where the water table is near or above the surface but subject to high seasonal fluctuation' (Grootjans et al., 1998: 96). Wet dune slacks ('humid slack' in the EU Habitats Directive terminology; European Commission, 2013) are a unique type of wetland ecosystem, highly rated in conservation for the presence of many rare plant species and fauna (Beaumont et al., 2014; Grootjans et al., 2004). Climate change poses serious threats to humid slack ecosystems (Jackson and Cooper, 2011), through impacts on dune water tables, predominantly due to changing patterns of rainfall (Davy et al., 2006). These habitats are particularly sensitive to variations in groundwater regimes, which play a key role in controlling processes within the dune slack (Berendse et al., 1998). Humid slacks are also particularly susceptible to the effects of changes in sea level (SL) and temperature and also to anthropogenic impacts. Increasingly, their hydrology is likely to be threatened by SL rise, especially through sea water intrusion. Nutrient enrichment brings major consequences for community development (Davy et al., 2006). Likewise, water abstraction, changes in land use, atmospheric nitrogen deposition and other forms of pollution have severe effects on the condition of these systems (Stuyfzand, 1993).

Humid slacks form an integral part of the extensive dune systems of the Atlantic and some Mediterranean biogeographical

regions (Houston, 2008). Nearly 28% of the total surface of this habitat in the Mediterranean is included in Natura 2000 and up to 14% in Portugal (Houston, 2008). However, most research on wet dune slacks has predominantly centred on northwest European coasts (e.g. Grootjans et al., 2004; Van Dijk and Grootjans, 1993). Comparatively, little work has been undertaken along Mediterranean coasts with focus on biological and environmental processes (Dimopoulos et al., 2006; Romo et al., 2016; Serrano et al., 2006).

The southwestern Iberian coast, at the transition between Atlantic and Mediterranean influences, has long been recognized as an area vulnerable to the impacts of climate change. This coast

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has been predicted to be severely impacted by future SL rise (Ferreira et al., 2008; Schmidt et al., 2014) and is likely to experience warmer and drier conditions in the future (Ramos et al., 2011). Humid slacks and wetlands in this area support a distinct and highly diverse plant community which constitute an important natural value for the preservation of listed plants (Beja et al., 2005). In addition, these small ecosystems are vital to promote natural re-vegetation of dunes via primary succession and provide important ecological services for both the overall landscape processes and endangered species. A better understanding of the impacts of these drivers in the past may help in the analysis of the current situation and also facilitate prediction of how global change will influence future conditions. Stratigraphic studies provide an objective mean to reconstruct past dynamics with respect to climate and SL change impacts on coastal regions (Bicket et al., 2009; Dickinson and Mark, 1994; Fontana, 2005; Freund et al., 2004), and humid slacks represent one of the few available sinks for sediment and organic matter (OM) in coastal areas. However, wet slacks have not been often used for palaeoenvironmental reconstructions along the SW Iberian coast (Ferreira, 2006; Freitas et al., 2007). The most important works dealing with the long-term evolution of humid slacks in Portugal do not consider the parameters and mechanisms involved in their formation from a sedimentological perspective (Mateus, 1992; Queiroz, 1999). In fact, research of wet slack systems incorporating data about Holocene stratigraphies and sedimentological models are uncommon (Freitas et al., 2007). All this generates uncertainties in the reconstructed palaeoenvironmental trends because of the lack of information regarding their sedimentological evolution.

Our main objective is to examine the Holocene environmental changes in a humid slack of the Portuguese coast (Poço do Barbaroxa de Baixo, BB) using a multi-proxy approach. This information is interpreted in terms of climatic variability and SL change, and their impact on environmental conditions. Using the proxy records in conjunction with the established Holocene climate and SL records we seek: (1) to propose a Holocene evolutionary pattern of a humid slack system in SW Portugal in response to climate and SL change and (2) to discern the most important drivers affecting its long-term evolution and their implications within the context of future SLR and climate change projections.

Study site

The study area is located in the southern part of the Tróia-Sines bay (SW Portuguese coast), about 80 km south of Lisbon (Figure 1). Barbaroxa de Baixo (38° 04' 45.00" N, 8° 48' 33.39" W) is a small (0.027 km²) and shallow (<1 m) wetland occupying the trough defined between N-S trending dune ridges of the Old Dune Complex (ODC) (Freitas et al., 2007). This humid slack is part of the Natural Reserve of *Lagoas de Santo André e da Sancha*, included in the list of Portuguese Wetlands of International Importance (the 'Ramsar List').

Field surveys conducted during 2012–2013 show that BB is a freshwater to oligohaline (0.2–3.8‰) and slightly acidic to slightly alkaline (pH 5.9–7.6) environment with medium to high mineralization (0.06 meq L⁻¹). Dissolved oxygen of 3–7 mg L⁻¹ and Eh indicate prevailing reducing conditions (–110 to +150 mV). The slack is usually flooded but the water level varies seasonally (~75 cm). During the rainy season, the water level rises (never exceeding 1 m depth) and drops during the dry season, causing occasional drying. It is fed by both the regional and local groundwater flow systems and surface water runoff. Additional inputs of marine water can be derived from overwash during severe storms. The hydrological regime of the humid slack results from a balance between precipitation, evapotranspiration and groundwater dynamics. The maintenance of the water table permanently

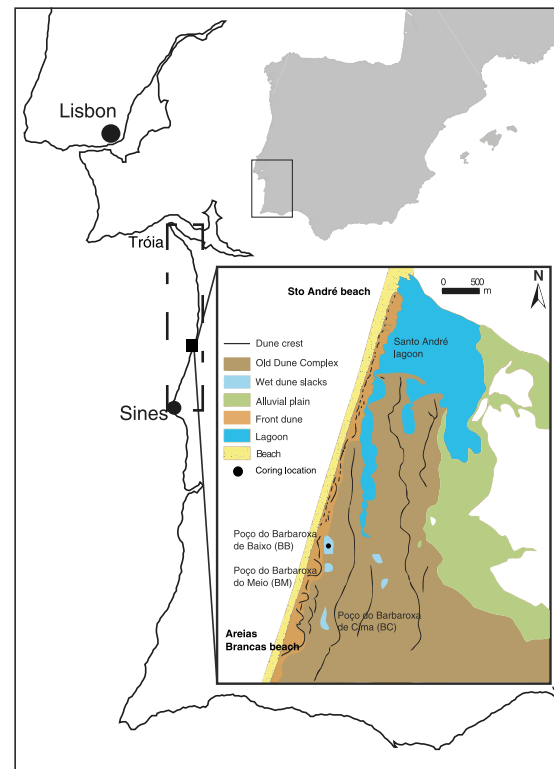


Figure 1. Geographical setting and location map of the Poços do Barbaroxa.

near or above the topographic surface provides conditions for growth of marsh plants (such as *Phragmites australis* and *Cladium mariscus*) along the wetlands' margin and basin, producing the accumulation of OM and development of peat.

Methods

Sampling, sedimentology, bulk chemistry, stable isotopes and dating

Two cores (BxBx3, 336 cm and BxBx5, 330 cm) located very close (4–6 m) to each other were taken in April 2010 using a Livingstone corer. The cores were opened in the laboratory, photographed and macroscopically described for texture and composition. Samples were taken every 2 cm and freeze-dried. In core BxBx5 subsamples were selected every 5–10 cm for bulk chemistry, stable isotopes and textural analysis. Whenever quartz sand was macroscopically detected, textural analysis was performed: size fractions above (coarse) and below (fine) 63 µm were separated by wet sieving using a 4 φ sieve and coarse fraction was classified according to mean diameter (Wentworth, 1922) and sorting (Friedman, 1962).

OM content was analysed following standard procedures for loss-on-ignition (Loh et al., 2008). Total organic carbon (TOC), nitrogen (TN), sulphur (TS) and stable isotopes ($\delta^{13}\text{C}$ and $\delta^{15}\text{N}$) were processed at the Servizos de Apoio á Investigación of the Universidade da Coruña. Subsamples were homogenized and weighed into tin capsules. Capsulated samples were analysed by Dumas combustion using a Thermo Flash EA 1112 coupled online to a Thermo Delta V Plus isotope ratio mass spectrometer. All carbon and nitrogen isotope ratios are expressed in conventional per mille (‰) notation (Hayes, 2004).

Accelerator mass spectrometer (AMS) ^{14}C dates were obtained on plant macrofossils and particulate OM from BxBx3 and BxBx5 cores. Macrofossils were extracted by wet sieving in distilled water, cleaned and sorted under a dissecting microscope, and identified when possible. A total of 18 AMS ^{14}C determinations

Table 1. AMS ^{14}C age data for Poço de Barbaroxa de Baixo, SW Portugal.

Lab number	Core	Depth (cm)	Material dated	Age (^{14}C yr BP \pm 1 σ)	Calibrated age (cal. yr BP \pm 2 σ)
Beta-393517	BxBx5	74–75	Organic sediment	2650 \pm 30	2765–2730
Beta-393518	BxBx5	100–101	Organic sediment	2140 \pm 30	2300–2255 2160–2040 2015–2010
Beta-393519	BxBx5	199–200	Organic sediment	5340 \pm 30	6265–6250 6210–6000
Beta-393520	BxBx5	269–270	Organic sediment	6070 \pm 30	7000–6855 6810–6805
Beta-393521	BxBx5	299–300	Organic sediment	6250 \pm 30	7255–7160
Poz-51897	BxBx3	21–23	Seeds	775 \pm 25	673–729
Poz-51898	BxBx3	82–84	Cyperaceae seeds	1950 \pm 30	1825–1950 1960–1970 1980–1983
Beta-294416	BxBx3	92–94	Seeds	1530 \pm 30	1352–1424 1426–1445 1453–1522
Beta-309757	BxBx3	102–104	Seeds	2500 \pm 30	2487–2733
Poz-51893	BxBx3	108–110	<i>Cladium</i> seeds	1970 \pm 30	1868–1992
Poz-51894	BxBx3	115–116	<i>Cladium</i> rhizome	4690 \pm 35	5319–5427 5431–5478 5538–5577
Poz-52107	BxBx3	118–120	<i>Cladium</i> rhizome	4525 \pm 35	5049–5194 5212–5308
Poz-51896	BxBx3	128–130	Degr. seeds, leaves	2020 \pm 30	1894–2051
Beta-326248	BxBx3	142–144	Wood	4880 \pm 30	5587–5654
Beta-294417	BxBx3	162–164	Plant material	5040 \pm 40	5664–5690 5709–5901
Beta-294418	BxBx3	212–214	Plant material	5630 \pm 40	6315–6486
Beta-294419	BxBx3	282–284	Pine needles	6180 \pm 50	6949–7179 7197–7240

AMS: accelerator mass spectrometer.

(13 from BxBx3 and 5 from BxBx5) were obtained from two different laboratories (Table 1): Beta Analytic Inc. (USA) and Poznan Radiocarbon Laboratory (Poland). A depth-age model was built using BxBx3 data by fitting the age–depth data using linear interpolation between neighbouring levels using the *Clam* code version 2.1 routine (Blaauw, 2010) developed for the R software (R Core Team, 2010). This code calibrates radiocarbon ages using the standard IntCal09 calibration curve (Reimer et al., 2009; Stuiver et al., 1998) and generates Monte Carlo age–depth fits through the calibrated age probability distributions. It then calculates the ‘best-fit’ age–depth curve as the weighted average of 1000 iterations. The confidence intervals are calculated at the 2 σ range centred at the mean value of the iterations. The age–depth model adopted in this study represents the best compromise between: maximizing goodness of fit to the data set as indicated by *Clam*’s best-fit parameter; avoiding changing accumulation rates where the studied proxies indicate continuity of environmental conditions; avoiding samples representing incorporation of younger carbon where bioturbation is not visible. Precise correlation between BxBx3 and BxBx5 was carried out with the help of the AMS radiocarbon dating and using distinct stratigraphic markers easily recognized in both cores.

Diatoms and other biological proxies

Diatom remains were used for facies characterization and palaeo-ecological reconstruction because of their relevance in coastal wetlands (Trobajo and Sullivan, 2010). Early studies illustrated the abundance of diatoms in humid slacks and their usefulness in monitoring trends in habitat quality (Round, 1958) and as palaeoenvironmental proxies (Denys, 2003). Samples (0.1 g dry weight)

were processed after Renberg (1990) and mounted onto microscope slides with Naphrax (RI = 1.74).

Taxonomic identification and quantification of diatoms was carried out in 96 samples (every 5 cm, approximately, between 330 and 100 cm depth, and every 2 cm until top of the BxBx5 core). Identifications were performed at a magnification of 1000 \times under a Nikon Eclipse 600 microscope with Nomarski differential interference contrast. Whenever possible, at least 300–500 valves were counted in each sample. When diatom content was lower than 300, counting continued until the frequency of the taxa in the sample stabilized in relation to the overall sample size (Battarbee, 2000; Battarbee et al., 2001). In any case, at least 100 valves were considered an adequate number to make samples statistically valid (100 ± 0.93 ; $\alpha = 0.05$) (Fatela and Taborda, 2002). Taxonomic identifications were based on Hofmann et al. (2011), Krammer (2000, 2002, 2003), Lange-Bertalot and Levkov (2009), Lange-Bertalot et al. (2011) and Witkowski et al. (2000). Diatom valves (DVs) showed a high degree of dissolution and/or fragmentation, and low abundance in some of the samples. Nevertheless, the most representative taxa could be identified reliably, while for most of the other taxa, easily identifiable fragments remained relatively well preserved, and hence preservation did not interfere with counting accuracy. A DV preservation index was calculated as defined by Ryves et al. (2001). This index varies between 0 (when all valves present dissolution signals or are broken) and 1 (when all valves are pristine). In addition, the presence and abundance of chrysophyte cysts, sponge remains and non-pollen palynomorphs were also recorded.

Statistically constrained cluster analysis was performed using CONISS (Grimm, 1987) with Bray-Curtis distance on square-root transformed abundance data to define intervals containing

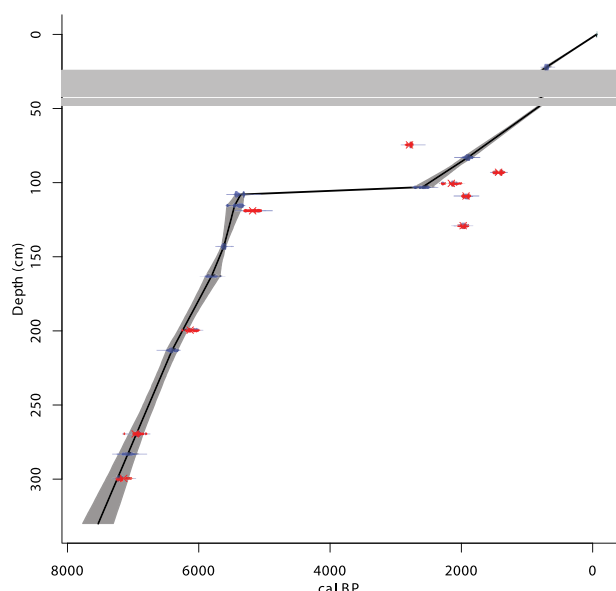


Figure 2. Chronological model of the studied sequence based on AMS ^{14}C dates (grey frame indicates a rapid sedimentary event).

similar species assemblages and to identify zonation in the taxonomic profile. Species with a relative abundance less than 5% were not included in the analysis. Prior to analysis, species relative abundance data were square-root transformed. The effect of the transformation is to maximize the signal-to-noise ratio in the data by upweighting subdominant taxa and is considered most appropriate for species abundance data (Overpeck et al., 1985). The number of statistically significant zones was determined by the broken-stick model (Bennett, 1996). The remaining assemblage was graphed using the software package C2 version 1.4 beta (Juggins, 2004).

Palaeoenvironmental interpretation was based on the diatom species' environmental preferences (salt, brackish or freshwater), habitat and lifeforms (benthic, tychoplanktonic or planktonic) following Denys (1992 [1991]), Vos and de Wolf (1993a), Van Dam et al. (1994) and Witkowski et al. (2000). A salinity index was constructed specifically using autoecological information calculated by summarizing the relative frequencies of occurrence in all salinity classes defined by Hustedt (1953, 1957). The salinity index was calculated as the ratio between the halophobous/oligohalobian and mesohalobian/polyhalobian diatom forms present.

Results

Dating, sedimentology and bulk chemistry

Radiocarbon dating results are given in Table 1 and Figure 2. Some age inversions were found.

According to the age-depth model, the base of the Holocene succession was dated to 7550 cal. yr BP. A two-stage sedimentation history can be considered, separated by a period characterized by a pronounced drop in the sedimentation rate (Figure 2). Sedimentation rates are higher and quite constant ($\sim 1 \text{ mm yr}^{-1}$) prior to 5230 cal. yr BP. They drop considerably to 0.04 mm yr^{-1} in the next 2730 years, until 2500 cal. yr BP. From that time onwards, sedimentation increases to 0.29 mm yr^{-1} . The only exceptions correspond to two abrupt sedimentary events (interpreted from sedimentological data), at 24–42 and 43–48 cm, tentatively dated to c. 750 and 800 cal. yr BP, respectively.

During the low sedimentation phase (between 5230 and 2500 cal. yr BP), the sequence of ^{14}C dates includes several reversals and it is difficult to adopt a robust age model between 92 and 130

cm depth below surface. The age difference between the large pieces of *Cladium* rhizome and smaller plant remains immediately below and above them suggests that these robust macrofossils became embedded within younger material after that an erosional event had removed their original matrix. This can provide incorrect age data to the dated horizon incorporating these plant remains (Verschuren, 2001). Thus, we interpret this particular section of the record as resulting from the combination of two components. First, an older and essentially non-depositional component and time lapse, which corresponds to the no or negligible accretion under stable SL conditions. And, second, an erosive component generated by reduced water level in the slack, once the wetland sedimentation was catching up with the preceding rise in SL and grew vertically to exceed the water table and became drier. The major drop in TOC values between 108 and 100 cm (Figure 3) may help to constrain the range of this interruption/slowdown in deposition to between 5230 and 2500 cal. yr BP.

The sediments are hyperacid to sub-acid (Freitas et al., 2007) and carbonate free (Ferreira et al., 2006). Sedimentological and geochemical results (Figure 3) allow us to distinguish seven units in the infill of BB wetland:

Unit I (330–310 cm; 7550–7350 cal. yr BP) – this unit consists of moderately sorted medium quartz sand. OM is vestigial in the base and increases in the upper part, reaching almost 10%. $\delta^{15}\text{N}$ and TN are below detection level and the C/N ratio varies 10-fold, ranging between the lowest value and one of the highest in the whole core (Figure 4).

Unit II (310–300 cm; 7350–7250 cal. yr BP) – this unit is essentially formed by fine peaty deposits with occasional wood fragments (30–70% OM). A decrease in OM to c. 9–15% at the top of the unit is due to the presence of sand. $\delta^{15}\text{N}$ is high and both C/N ratio and $\delta^{13}\text{C}$ values are within the range found in Unit I.

Unit III (300–108 cm; 7250–5230 cal. yr BP) – this unit is essentially formed by peaty deposits (OM mainly above 80%), the organic component being made of fermented plant debris (fragments of stems, leaves and roots) of variable dimension. Rhizomes of *Phragmites* were occasionally found. The presence of sand induces a decrease in OM to values of 60–70% at 202–204, 184–186, 179–181 and 162–166 cm and to lower values (23–36%) at 200–202 cm. OM is absent in a millimetric-thick whitish muddy lamina at 290 cm. Values of TOC and TN are substantially higher than in underlying units. In broad terms, their vertical profiles mirror the OM variations, though TN is somewhat higher at the middle section of the unit. The base (below 240–250 cm; Unit IIIA) and top (above 150 cm; Unit IIIC) of this unit yielded lower $\delta^{13}\text{C}$ values (means of -28.87% and -27.82% , respectively), whereas in the middle section (Unit IIIB) this parameter is higher (mean of -27.52%). Similarly, the C/N ratio is lower in the middle section (<36) than at the base and top. $\delta^{15}\text{N}$ varies between -1.9% and $+0.7\%$. TS peaks at 180–181 and 113–114 cm, with values of 8.4% and 7.6%, respectively. TOC/TS varies between 4 and 32.

Unit IV (108–100 cm; 5230–2500 cal. yr BP) – this unit essentially consists of coarse peaty materials with sand dispersed along the unit (OM 20–60%) and concentrated in some levels (105–108 and 100–102 cm; OM $<10\%$). Values of TOC, TN and TS are lower than in the underlying unit and TOC/TS ratio varies between 7 and 17. C/N ratio and $\delta^{13}\text{C}$ values are compatible with C3 terrestrial plants at the base of the unit and with freshwater DOC at the top (Figure 4). $\delta^{15}\text{N}$ values are higher than in the previous unit, between $+0.8\%$ and $+1.73\%$.

Unit V (100–48 cm; 2500–800 cal. yr BP) is essentially formed by fine peaty deposits at the base (100–90 cm) and coarser texture peaty material up unit. Sand occurs concentrated in some levels (e.g. 77 cm) and scattered between 70 and 65 cm. OM varies accordingly, decreasing upwards (essentially from 74% to 50%) and decreasing to 12% where sand is present. TOC, TN and

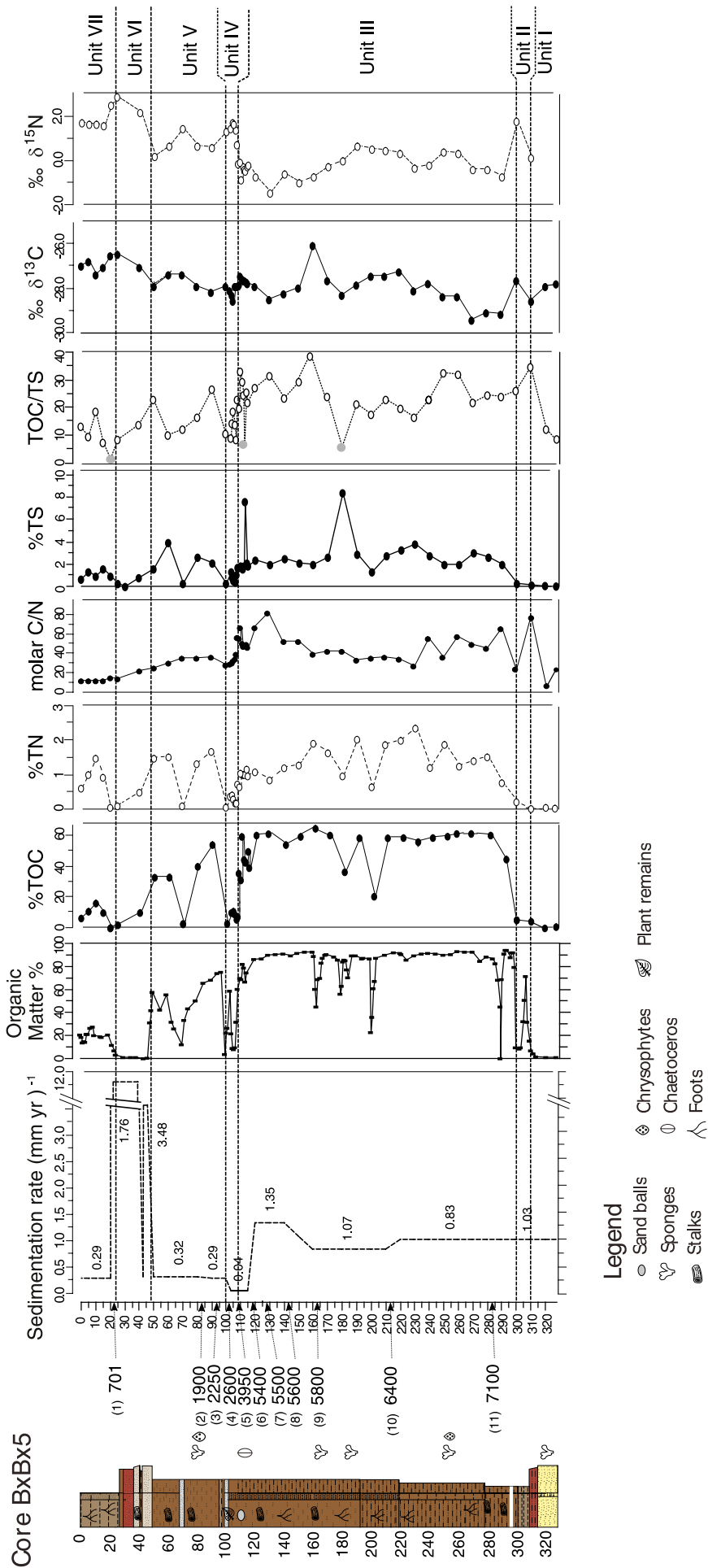


Figure 3. Sedimentary log showing sedimentation rates (mm⁻¹ yr), total organic carbon (TOC) and total nitrogen (TN), molar TOC/TN ratio, total sulphur (TS) and TOC/TS ratio, stable isotopes data (δ¹³C and δ¹⁵N) and main lithological units from core BxBx5. The available dates are indicated. Vertical scale is in cm.

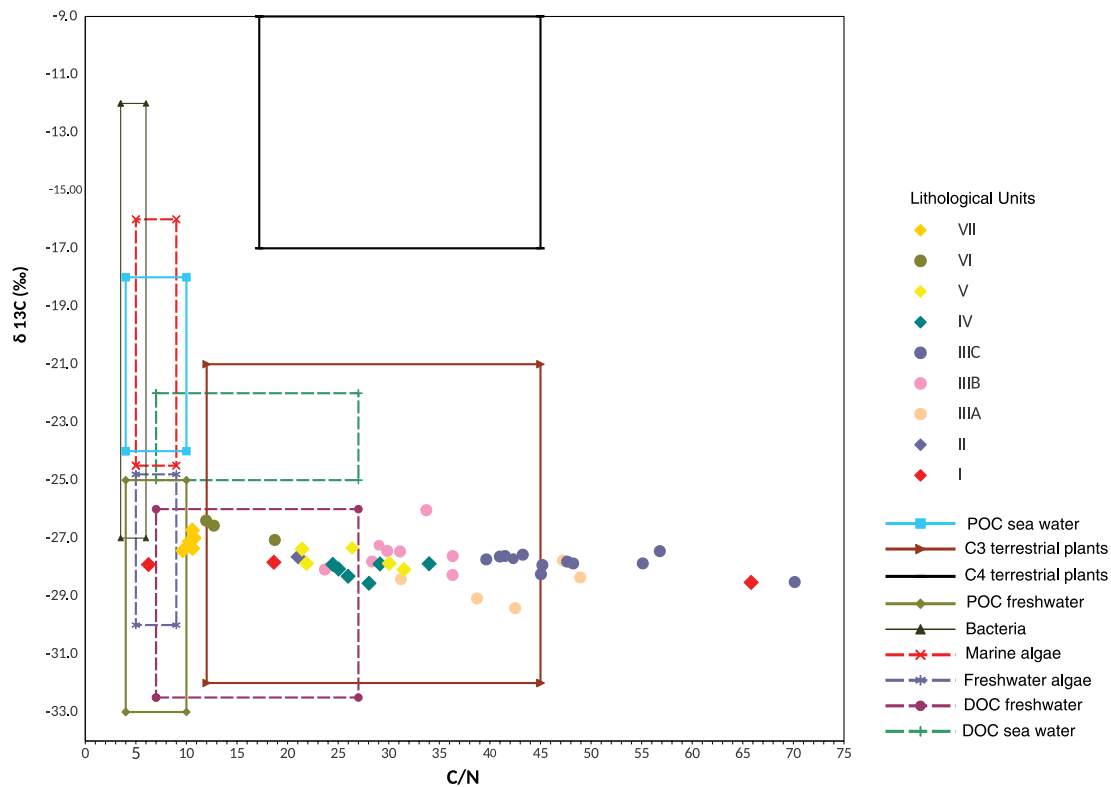


Figure 4. Biplot showing the distribution of C/N values versus stable isotope ($\delta^{13}\text{C}$) signatures of organic matter from the BB sediment core (Bx5). Lithological units are shown. The data areas are based on data presented in Lamb et al. (2006).

TS patterns mimic OM variations. Values of C/N decrease up the unit and high values of $\delta^{15}\text{N}$ are similar to those found in the underlying unit. C/N ratio and $\delta^{13}\text{C}$ values (-28.13‰ to -27.34‰) are compatible with C3 terrestrial plants at the base of the unit and with freshwater DOC at the top (Figure 4).

Unit VI (48–24 cm; 800–750 cal. yr BP) – at 48 cm there is a sharp shift to sand-dominated sediments, interrupted by a thin sandy organic peaty layer that spans the interval from 42 to 43 cm. Overlying the peat layer there is a second sandy unit up to 24 cm. Sandy sediments are OM depleted, varying from zero, increasing upwards, reaching 3% at the top. Close examination of both sand laminae shows the graded nature of the contact between the upper sand and the overlying organic mud, indicating progressive return to more organic-rich sedimentation. In contrast, the lower sand shows neat lower and upper contacts. Sand fines upwards, with mean diameter varying between 1.25 and 1.97. These sand laminae are interpreted as representing abrupt sand transfer episodes landward. TN, TOC and C/N are low and decrease up the unit, whereas isotopic values ($\delta^{15}\text{N}$ and $\delta^{13}\text{C}$) are higher and increase higher up. C/N ratio and $\delta^{13}\text{C}$ values correspond to C3 terrestrial plants and freshwater DOC.

Unit VII (24–0 cm; 750 cal. yr BP to present) – this unit consists essentially of organic muds (OM 7–27%). Elemental values (TS, TN, TOC and C/N) are higher than in sandy materials but comparatively lower than in peat. C/N ratio and $\delta^{13}\text{C}$ values vary within the values described for freshwater DOC and POC (Lamb et al., 2006).

Diatoms

The integrity of the sedimentary diatom record of BB is interrupted to some degree at certain levels (grey frames in Figure 5). From bottom of the core to 200 cm depth diatoms were well preserved. From 200 to 70 cm some of the samples contained too few diatom fragments or/and valves for quantitative analyses. Diatoms were well preserved from 70 cm to the surface. A total of

119 diatom taxa was identified, but only 31 were common ($>5\%$ in at least one sample). Four distinct breaks were identified at 265, 95, 75 and 43 cm based on the diatom assemblage changes, producing five diatom zones (DAZ; Figure 5).

DAZ 1 (330–265 cm); interpolated age c. 7500–6900 cal. yr BP. This zone is dominated by brackish/freshwater diatoms. *Pseudostaurosira* species (~60%) and *Pseudostaurosira trainorii* (~10%) are the dominant taxa. These species form colonies which are either attached to the substratum at one end of the frustule by a mucilage pad or planktonic. Brackish epipsammic diatoms are also common. Among the latter *Halamphora borealis*, a benthic brackish water species widely distributed in Europe, is the main species. The characteristics of this periphytic diatom assemblage suggest a shallow wetland environment with slightly brackish, stagnant water during the initial flooding. The dominance of diatoms from the *Fragilaria* complex (made up of *Fragilaria* sensu stricto, *Pseudostaurosira*, *Pseudostaurosira*, *Staurosira* and *Staurosirella*) points to environmental instability (Stabell, 1985). At the top of this zone, some levels show decreased preservation of diatom frustules (Figure 5) while others are barren (285 cm – grey frame, Figure 5). Megascleres of the freshwater sponge *Racekiela ryderii* (Potts) constitute a minor component of the assemblages, being the earliest dated record in Portugal.

DAZ 2 (265–95 cm); interpolated age c. 6900–2300 cal. yr BP. This zone is still characterized by the dominance of brackish/freshwater tychoplanktonic diatoms. However, some variations were found indicating differences between the bottom and the top of the zone. The lower section includes samples between 260 and 160 cm (c. 6900–5775 cal. yr BP) and is still representative of the baseline wetland conditions. The peak of the subdominant epipellic *Mastogloia smithii* probably reflects higher sediment availability. The relatively low values of the epiphytic (~15%) component suggest moderate macrophytic development. Between

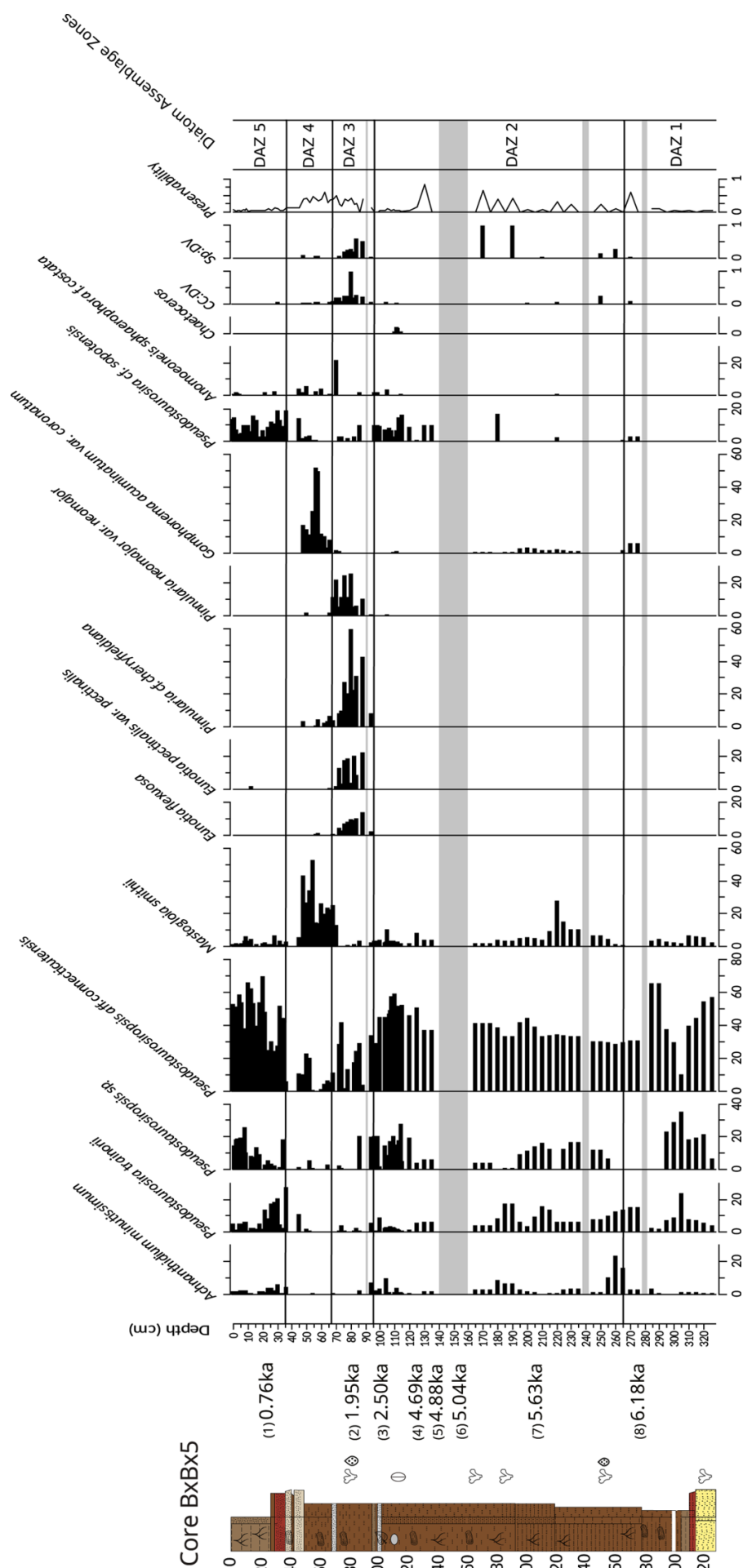


Figure 5. Diatom percentage diagram for selected taxa ($\geq 5\%$ abundance in at least one sample) from core BxBx5. Shaded areas indicate the parts of the record that lacked sufficient diatoms for quantification.

120 and 95 cm (c. 5450–2300 cal. yr BP), the brackish *Pseudostaurastira sopotensis* contributes to the co-dominant group of the whole assemblage. This diatom can form small chain-like colonies attached to sand grains (Ribeiro, 2010). Within this zone there are two gaps with no diatoms, between 235 and 245 cm and between 140 and 160 cm. Spicules of *R. ryderii* are more abundant and less fragmented.

DAZ 3 (95–75 cm); interpolated age c. 2300–1650 cal. yr BP. Freshwater benthic diatoms, mainly epipellic, are the main component of this zone, ranging in abundance from 50% to 75%. The dominant species is *Pinnularia* cf. *cherryfieldiana* (~70%) along with *Pinnularia neomajor* and several *Eunotia* species. This zone is also characterized by the abundance of megascleres and gemmoscleres of the freshwater sponge *R. ryderii* particularly between 85 and 75 cm. This sponge species prefers slightly acidic environments (Økland and Økland, 1996) and is most abundant when *Eunotia* species are also numerous. The part of the sequence at c. 2150 cal. yr BP shows high richness of *Pseudoschizaea* paly-nomorphs. The high abundance of macrofossils from terrestrial plants, as well as the increase in the acidophilous diatoms *Eunotia* sp. and *Pinnularia* sp., suggests a very shallow environment. The halophobous nature of these diatoms and the increase in chrysophyte cysts, compared with the dominance of oligohalobous diatoms in the underlying zone, also indicate water freshening at this site. Sediment between 92 and 90 cm is devoid of diatoms.

DAZ 4 (75–43 cm); interpolated age c. 1650 to ~800 cal. yr BP. The most significant feature in this zone is the reduction in freshwater epipsammic and epipellic diatoms in favour of both brackish water epipelon and fresh/brackish episammon (mainly represented by *M. smithii* and *Anomoeoneis sphaerophora* f. *costata*, respectively) and fresh to brackish epiphytic diatoms (*Gomphonema acuminatum* var. *coronatum*). *M. smithii* is the dominant diatom in periphyton mats of alkaline, freshwater to brackish wetlands, preferring fairly mineralized waters (β -mesohalobe forms). *Gomphonema acuminatum* var. *coronatum* is another cosmopolitan species found in usually slightly acidic to weakly alkaline waters, characteristic of low-electrolyte, oligotrophic environments. The dominance of epiphytic and epipellic assemblages, as well as the elevated abundance of macrofossils from terrestrial plants, suggests an encroachment of wetland and terrestrial plant communities in the BB pond. A special category of the epipellic diatoms in this zone consists of aerophilous diatoms, which live on muddy soils and are adapted to irregular flooding. The overall character of diatom assemblages indicates stronger sedimentation conditions and at least seasonal desiccation events, reflecting the progressive terrestrialization of the water-body margins.

DAZ 5 (43 cm to surface); interpolated age c. 800 cal. yr BP to present. The uppermost sediment is characterized by the importance of the brackish tychoplanktonic component, which could reflect higher salinities than in the previous zone. *Pseudostaurastira* sp., *P. sopotensis* and *P. trainorii* replace the strictly periphytic diatoms of the former zone. The tychoplanktonic character of these diatoms suggests the development of a shallow environment with open waters. Also present and common in the recent sediments is *Halampyrea borealis*. This zone shows a strong resemblance to the base of the core. As in DAZ 1, the diatom assemblage suggests the development of wetland conditions subject to permeating sea water, sea spray or occasional episodes of extreme high water.

Discussion

The palaeoenvironmental reconstruction is based on microfossil analyses and diatom zonation, taking into consideration changes in the sedimentary facies, the OM content and the physical and

geochemical proxies. Sediment supply clearly varied during the accumulation of sediment of Units II to VII, as evidenced by the distinct contributions of organic material and sand. Peat/organic mud deposits are the product of autochthonous wetland accumulation from plant growth added by local in-washing of fine-grained particles, whereas the sand beds or laminae either wind-blown or transported sediments from the adjacent ODC and present-day foredune. The increased sand content in the upper part of the sediment record is common to all three Barbaroxa ponds (Freitas et al., 2007).

Palaeoenvironment reconstruction

The combination of several proxies allows recognition of six main phases in the evolution of the BB sediment record (Figures 6 and 7).

Phase 1: c. 7550 to c. 7250 cal. yr BP. The onset of wetland conditions occurred in this time window and was synchronous with Holocene postglacial SL rise. SL must have been sufficiently high by c. 7550 cal. yr BP to have induced a change in the local base-level allowing for water ponding in this interdune trough (whose base was at c. 2.3 m below present-day mean SL, mSL) and for OM accumulation. Contemporaneously, a beach-dune barrier system developed that persisted until present. The earliest infill of the humid slack is essentially peaty (Unit II) and rests upon a basal unit (Unit I) of slightly organic medium sand (Unit Ib in Freitas et al., 2007), similar in texture to the modern surface sediment of the western adjacent ridge (Ferreira et al., 2006). The change in facies between the lower sand and upper organic materials evidence of sediment impeded drainage. Frequent flooding and ponding inhibited post-depositional OM oxidation and allowed the development and accumulation of abundant plant remains and debris over the surface of the depression.

OM content is initially low ($\text{TOC} \leq 1\%$) and increases up unit. In this case, TOC/TN total values are unreliable indicators of OM sources (Meyers, 2003). The $\delta^{13}\text{C}$ content shows little variation, ranging between -27.65% and -28.53% . These values are consistent with material derived from C3 plants (-22% to -30%) (Maslin and Swann, 2006). *Phragmites australis* (C3), a common and abundant plant in Atlantic Iberian coastal wetlands, yielded $\delta^{13}\text{C}$ values of -30.6% and -31.3% (Bao et al., unpublished data). The range of $\delta^{15}\text{N}$ values at the base of the core (0.16% – 1.83%) is characteristic of groundwater and terrestrial N inputs (McClelland et al., 1997; Middelburg and Nieuwenhuize, 1998). It is plausible that the sandy sediments of the basal unit allowed for terrestrial and groundwater infiltration and percolation, which contributed N to BB.

The diatom assemblages consist predominantly of small-sized fragilarioid taxa. These typical pioneering diatom communities colonize newly formed lacustrine habitats (Reed et al., 1999) and have an advantage in rapidly changing environments (Yu et al., 2004). Thus, their abundance can be regarded as the marker of the onset of wetland conditions. The predominance of *Pseudostaurastira* aff. *connecticutensis* suggests the existence of a shallow wetland environment. Autoecological information on *P. connecticutensis* is rather limited. Although originally found in low abundances in lotic freshwater systems in North America (Morales, 2001), its presence can be considered as indicative of a predominantly freshwater environment with a narrow salinity variability.

Phase 2: c. 7250 to 5500 cal. yr BP. During this phase, BB developed mostly palustrine conditions, with changes in OM origins and preservation, as well as in water depth and salinity. There are several gaps in the diatom record during this interval. The absence of diatoms or poor preservation has been attributed to different environmental conditions in the sediment of coastal environments

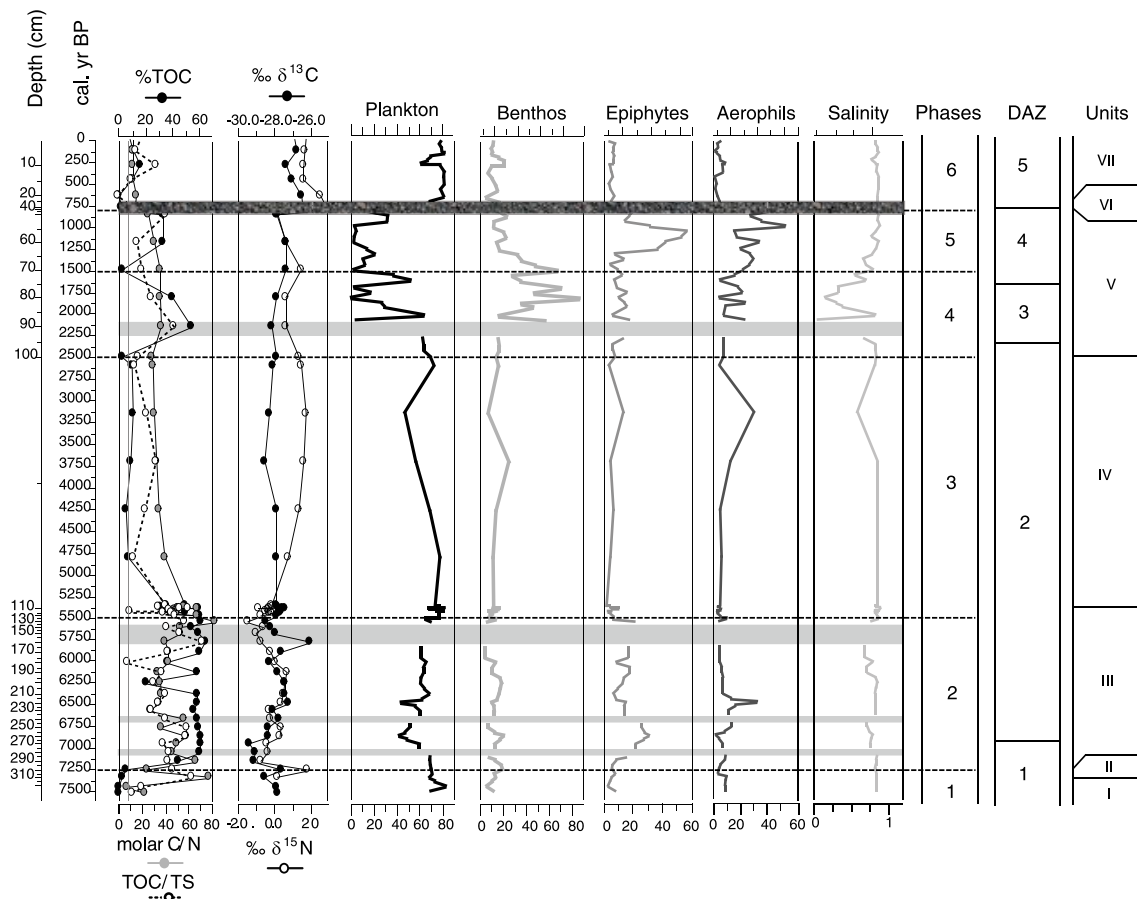


Figure 6. Summary diagram for selected data from BxB5 sequence plotted against age. Diatom taxa were grouped according to their life form and salinity affinities (see text for details). Salinity index was calculated as diatom assemblage index. The index is based on percent abundances of diatom species that were divided into five indicator categories: (1) halophobous; (2) oligohalobian – indifferent; (3) oligohalobian – halophilous; (4) mesohalobian; and (5) polyhalobian. Shaded areas indicate the parts of the record that lacked sufficient diatoms for quantification. The grey dotted area marks the abrupt sand deposition event c. 750–800 cal. yr BP.

(Taffs et al., 2008). Desiccation, leaching, salinity and temperature all have an effect on the biogenic silica cycle (Natori et al., 2006; Roubex et al., 2008).

The shift to higher $\delta^{13}\text{C}$ values, associated with relatively low contribution of aquatic vegetation to the sediment carbon inputs (Aichner et al., 2010), is interpreted as a higher OM input from aquatic macrophytes in a land-plant derived OM-dominated environment. Nevertheless, it is noteworthy that sediments during this period show the lowest values of $\delta^{15}\text{N}$, close to atmospheric N_2 values ($\delta^{15}\text{N} = 0\text{‰}$). These values can be explained by the cyanobacterial fixing activity with relatively little isotope fractionation to produce OM with $\delta^{15}\text{N}$ typically in the range -3‰ to $+1\text{‰}$ (Fogel and Cifuentes, 1993). This possibility is supported by high C/N ratios, as a result of depressed OM degradation and microbial nitrogen fixation in anoxic conditions (Meyers and Teranes, 2002). Changes in TS can be used to provide a signature of palaeosalinity or OM oxidation. Very high TS concentrations at 180 cm point to a brackish/marine depositional environment (Casagrande et al., 1977) due to the higher availability of sulphate in sea water compared to freshwater (e.g. Chagué-Goff, 2010; Chagué-Goff et al., 2000, 2002). Oxidation and cycling could also account for much of the TS excess in highly organic sediments. During summer, water table levels in these shallow water bodies decline significantly, exposing previously anoxic peat to air and allowing oxidation of reduced S compounds (Eimers et al., 2007). When normal hydrologic conditions resume, newly produced S is reincorporated into surface waters increasing S values in sediment deposited afterwards. Although it seems difficult to reach values as high as those found in our record.

The co-occurrence of tychoplanktonic and brackish/freshwater benthic diatoms points to environmental instability (Stabell, 1985). The combination of diatom assemblages of different optima such as the halophobous and the oligohalobous halophilous taxa, probably resulted from periods of freshwater inputs, allowing for temporary development of freshwater flora, followed by subsequent periods of evaporative concentration and higher salinities. The ratio of cysts to diatoms displays some peaks during this phase. Chrysophyte cysts are rare or absent in saline waters (Cumming et al., 1993). Thus, in coastal shallow temperate systems as BB, maximum values of this ratio are consistent with transition to freshwater conditions.

Phase 3; c. 5500 to 2500 cal. yr BP. This period encompasses a distinct change in both the origin of OM and accumulation rate. Accumulation rates are extremely low, influenced not only by SL stabilization but also by drier conditions and lack of accommodation space. This phase represents a period of shallowness.

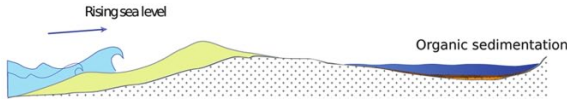
The $\delta^{13}\text{C}$ values are characteristic of both DOC and C3 land plants. TOC/TN values confirm contributions of terrestrial organic material (Emerson and Hedges, 1988; Meyers, 1994; Meyers and Lallier-Vergès, 1999) and the presence of macrophyte material (Meyers, 2003). $\delta^{15}\text{N}$ is also compatible with higher influence of terrestrial inputs, although similar values may result from marine (Maksymowska et al., 2000) and atmospheric inputs (Valiela et al., 2000).

This phase shows scarce diatoms, but the occasional presence of *Chaetoceros* sp. resting spores (110–113 cm depth) suggests the sea influence into the slack restricted to periods of enhanced

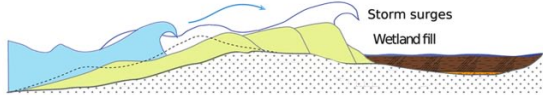
Phase 1: 7.5 - 7.3 cal ka BP, wetland onset



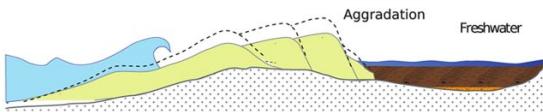
Phase 2: 7.3 - 5.5 cal ka BP, wetland development



Phase 3: 5.5 - 2.5 cal ka BP, stationary sea level



Phase 4: 2.5 - 1.5 cal ka BP, wetland reactivation



Phase 5: 1.5 - 0.8 cal ka BP, aggradation/progradation



Phase 6: 0.8 cal ka BP - present

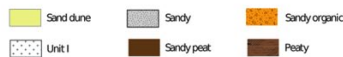
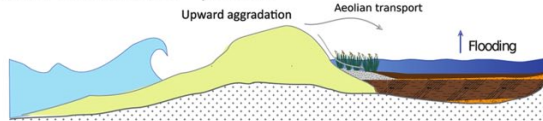


Figure 7. Schematic representation for Barbaroxa de Baixo system evolution since its origin.

storminess (Witak et al., 2011). *Chaetoceros* sp. resting spores, an allochthonous component that could, effectively, be transported by sea spray (Lee and Eggleston, 1989). However, the TS reaches values too high to be justified by just sea spray with TOC/TS values between 0.5 and 5 characteristic for marine sediments, while lake/freshwater sediments typically present values of 40–120 (Bernier and Raiswell, 1984). This suggests that the beach-foredune barrier was permeable enough to allow for the inflow of sea water during catastrophic storms at the time of the maximum reach of the Holocene transgression. Low sedimentation rates during this period have been recorded elsewhere in SW Portugal (Queiroz and Mateus, 2004; Schneider et al., 2010, 2016). Queiroz (1999) considered a sedimentation hiatus at the nearby dune trough of Barbaroxa de Cima (BC) commenced at sometime around 4700–4800 cal. yr BP. Also in the sedimentary record of BB a decrease in the sedimentation rate of similar magnitude is observed, although slightly earlier in time (5500 yr BP). In a shallow sedimentary environment such as BB, there are expected gaps in the depositional record due to the lack of sedimentation and they actually occurred in the sedimentary column treated here.

Phase 4: c. 2500 to 1500 cal. yr BP. This period is characterized by stabilization and aggradation/progradation of the bounding

sand dunes at the interface with the ocean and increased freshwater influence, marked by fine-grained sedimentation. The *Pseudostaurosira* dominated diatom assemblage was substituted by freshwater epismmon after 2300 cal. yr BP. Diatoms such as *Pinnularia* sp. and *Eunotia* sp. are typical of stagnant freshwaters with considerable macrophytic development (Vos and de Wolf, 1993b) and more acidic conditions (Hofmann et al., 2011; Krammer, 2000; Lange-Bertalot et al., 2011), congruent with paludification. This coincides with an increase in accretion rate. The paludification process requires a positive hydrological balance during the growing season. The increase in precipitation from middle to late Holocene created optimal conditions to re-initiate peat growth (starting at 2300 cal. yr BP) and promoted the development of shallow temporary pools. A wetter period between 2350 and 1670 cal. yr BP was registered in wet dune slacks located along the Portuguese SW coast (Cruces et al., 2010). The abundance of *Pseudoschizaea* palynomorphs from c. 2200 cal. yr BP points to a peaty environment under marshy conditions with fresh stagnant water (Luz et al., 2011; Medeanic et al., 2008). This is further supported by the occurrence of chrysophycean cysts and the high spicule and gemmosclere numbers (Harrison, 1974). Bao et al. (2007) document a prevalence of increased freshwater conditions with rare episodes of some degree of salinity for the same period in the Traba coastal lake (NW Spain). However, the presence of *Pseudostaurosira* sp. still indicates environmental instability.

Phase 5: c. 1500 to 800 cal. yr BP. This stage is characterized by abundant periphytic taxa, such as *M. smithii* and *Gomphonema acuminatum* var. *coronatum*, and small-sized fragilarioid species, such as *Pseudostaurosira* species. Their fresh brackish to brackish water affinity suggests slight brackish conditions with extensive macrophytic development, which may have been caused by enhanced storminess and/or a minor sea-level rise impulse. The concurrent decline in TOC/TN corroborates a shift from largely terrestrial to increasingly aquatic sources of OM (Meyers, 1994; Meyers and Lallier-Vergès, 1999; Meyers and Teranes, 2002). The stable isotope values are slightly higher from this moment onwards, increasing gradually to the top of the sequence. There are two possible causes for $\delta^{13}\text{C}$ increase over this period. The first is an increase in primary production by algae (Brenner et al., 1999; Cifuentes et al., 1988; Schelske and Hodeli, 1991), thus driving a decreasing fractionation of C fixed by algae. The second is an increase in inputs from emergent and floating leaved macrophytes, which are within the range of -23‰ and -30‰ (Smith and Epstein, 1971), very similar to that of terrestrial C3 plants. On the other hand, $\delta^{15}\text{N}$ increased slightly during this period, which may represent an influx of nutrients (Bratton et al., 2003; Savage et al., 2004).

Phase 6: c. 800 to present cal. yr BP. This phase represents a return to more inorganic sedimentation conditions, probably in response to episodes of increased reactivation of aeolian processes or displacement of the foredune. $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ are much the same as during the previous phase, pointing to similar sources of N and C. There is a sharp change in the diatom assemblages, now dominated by *Pseudostaurosira* species. The co-dominant species *Pseudostaurosira trainorii* is an opportunistic brackish/freshwater tycho planktonic diatom, common in small water bodies with neutral pH (Edlund et al., 2006; Morales, 2001) and abundant in freshwater coastal lakes and ponds (Ferreira, 2013). The increase in tycho planktonic diatoms and the compositional and geochemical data (lower values of TC, TOC/TN and $\delta^{13}\text{C}$) evidence shallow conditions with abundant littoral vegetation, not unlike those at present. The dominance of small diatoms of the *Fragilaria* (s.l.) group

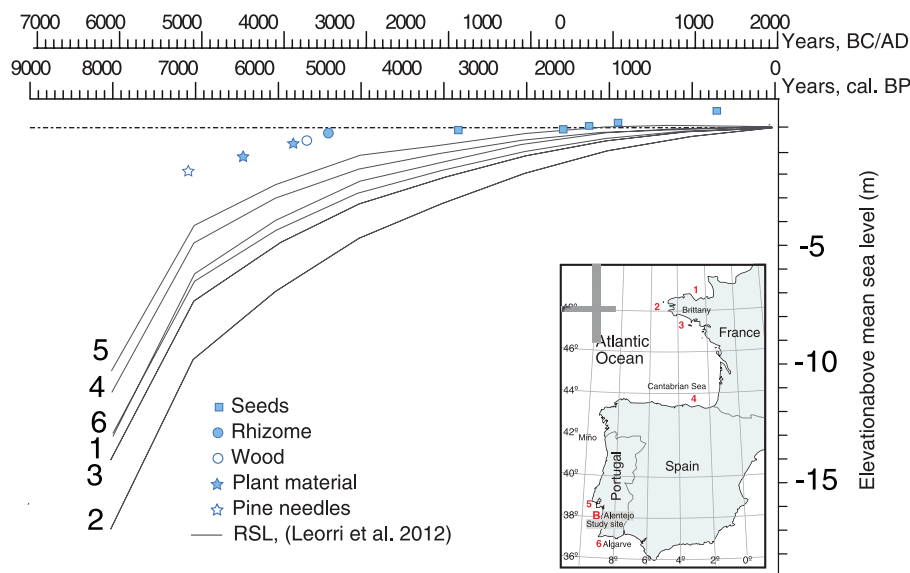


Figure 8. Sea-level change: 1, 2, 3, 4 and 5 grey lines correspond to relative sea-level curves along the Atlantic coast in SW Europe (Leorri et al., 2012). 1–5 positions are indicated in the map; letter B indicates the position of Barbaroxa wet slacks. Geometric symbols represent age data (AMS ^{14}C). All heights refer to the Alicante Mean Sea-Level – Spanish Datum System (AMSL).

includes the species *P. connecticutensis* and *P. trainorii*. This group of diatoms is typical of shallow but open waters and littoral zones, favoured by alkalinizing base cations released from the catchment (Axford et al., 2009; Reed et al., 1999). The dominance of these diatoms would indicate increased erosion in the catchment that enhanced runoff and cation input, probably related to foredune instability and vegetation clearance through a strong agro-pastoral pressure (Queiroz, 1999). The high abundance of *Achnanthes minutissima* might also be indicative of a higher accretion rate (Earle et al., 1986).

Environmental drivers in the evolution of BB

Several episodes occurred throughout the middle and late Holocene characterized by a reduced water table and/or lower organic content and/or low sedimentation rate in BB. These episodes occurred between 7000 and 6400 cal. yr BP, 6000 and 5500 cal. yr BP, 5250 and 2500, at around 2100 and at 1750 cal. yr BP and again between 1500 and 750 cal. yr BP. Some of them are partially coincident with absence or decreasing degree of diatom preservation in the sediment that may represent adverse environmental conditions. Climate and SL changes are crucial drivers of environmental changes. In this way, the Holocene history of BB can be tentatively analysed and interpreted in correlation with SL change (Figure 8) and climatic events (Figure 9).

Sea level. Changes in SL will influence the elevation of both the piezometric surface of the sandy coastal aquifer and water level in sandy troughs. The study of the sedimentary sequence allowed the identification of two phases characterized by pronounced variation in sedimentation rates in response to the slowing down of SL rise rate.

The Barbaroxas were far above SL near the start of the Holocene, when they stood ~27 m below present-day mSL. From c. 10,000 until c. 7000 cal. yr BP, SL rose rapidly up to ~5 m below mSL and experienced a first attenuation phase (Leorri et al., 2012). About this time, SL (and the correlated base-level) was already high enough to decrease drainage efficiency and cause stagnation of water in morphological lows such as BB. Sedimentary sequences from SW Iberia suggest a regional pattern of pronounced SL rise until 7000 cal. yr BP and transition from a transgressive to a SL highstand phase after that date until 5000

cal. yr BP (Fletcher et al., 2007; Lario et al., 2002). This explains the peat development observed in BB and elsewhere along the southwest and west Iberian façade after 7500 cal. yr BP (Figure 7) during the Holocene transgressive maximum (Freitas et al., 2007). Palaeotopography allowed for earlier peat accumulation in nearby Barbaroxa de Cima and Barbaroxa de Baixo (c. 7500 cal. yr BP), while this started much later than in Barbaroxa do Meio (c. 4100 cal. yr BP) and Lagoa de Sancha (c. 5400 cal. yr BP), because of their higher elevation (Freitas et al., 2007). At present, the surface of sediment carpeting the Barbaroxas stands 2 m above mSL and is fairly horizontal. Assuming similar hydrological relationship in the past (Freitas et al., 2007), by 7500 cal. yr BP, mSL should have been positioned at about 5 m and rose to almost its present position by 4000 cal. yr BP (Teixeira et al., 2005). These figures are in general agreement with the SL curves independently reconstructed for the area (Figure 8). Coastal wetlands maintain an elevation in equilibrium with SL by accumulating sediment (Redfield, 1965, 1972). As SL rises, their long-term survival mostly depends on the water table level, which maintains the flooding within a relatively narrow range of vertical fluctuations. If the rate of accretion is greater than the combined effects of SL rise and subsurface compaction, the surface will remain above water under aerobic conditions. On the other hand, when the accretion rate falls below the rate of SL rise, the dune slack becomes waterlogged, the vegetation will shift to species more tolerant of anoxic waterlogged soils and the dune slack eventually converts to a pond. While the accretion rate keeps pace with SL rise, it will be permanently flooded. The palynological study of the neighbouring pond of Barbaroxa de Cima (Mateus, 1992) indicates that flooding depended on freshwater. This is a reflection of the direct relationship between the mean water level in the dune slack, the piezometric surface of the free aquifer formed by the coastal sand, and SL.

By 6000 cal. yr BP, SL rise had slowed down initiating a second deceleration phase. It is precisely this decline which propitiates the development of coastal features (Bao et al., 1999; Días et al., 2000) and the expansion of salt marshes (Schneider et al., 2016). Dune aggradation began, as indicated by the establishment of detrital barriers that enclosed and confined previous estuaries (Albufeira and Melides) or open embayments (Santo André), defining lagoonal environments which evolved as such until now. This geomorphological threshold corresponds to the upper limit

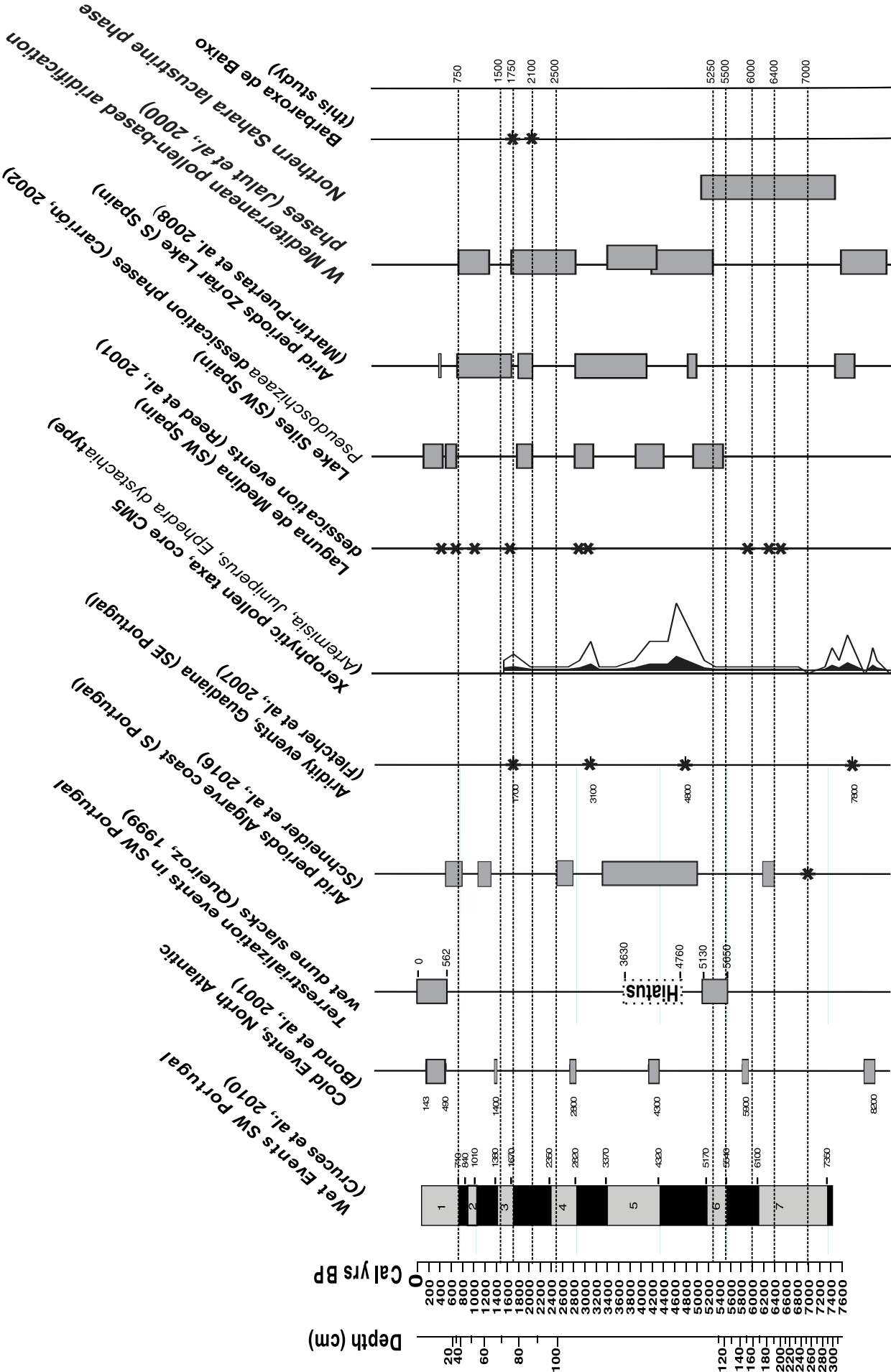


Figure 9. Selected sedimentary records from Southwestern Europe for the Holocene compared to the Bx5 core record.

of the time interval characterized by this mid-Holocene deceleration and stabilization of the SL rise (Freitas et al., 2002). After c. 5200 cal. yr BP, the progressive infilling of estuaries in SW Portugal is documented by palynological evidence indicating expansion of salt marshes. At around 5000 cal. yr BP, there is a major decrease in the sedimentation rate in BB by approximately two orders of magnitude (0.04 mm yr^{-1}). From this moment on, local factors begin to exert a greater influence.

This interruption or slow down in sediment deposition is ubiquitous along the Iberian south coast (Boski et al., 2008; Dabrio et al., 2000; Delgado et al., 2012; Lario et al., 2002; Mateus, 1992; Queiroz, 1999; Queiroz and Mateus, 2004; Schneider et al., 2010, 2016) and has been attributed to a regional change in climate (drier) around 4000–3000 yr BP (cf. Queiroz and Mateus, 2004). In fact, this episode occurred within the range of the aridity events defined for SW Europe (Carrión, 2002; Fletcher et al., 2007; Jalut et al., 2000; Martín-Puertas et al., 2008; Reed et al., 1999). However, the effect of SL rise on the water table is the most decisive factor determining succession in coastal wet dune slacks. This distinct drop in accumulation rate observed between 5230 and 2500 cal. yr BP is most probably a local effect related to the strong attenuation of the SL rise rate when wetland infilling was catching up with the preceding rise in SL. The wet slack should have experienced a relative decrease in the elevation of the depression in response to the stabilizing SL. As SL rise decelerated, productivity and sedimentation led to a shallowing trend and compromised the wetland's development under steady conditions of flooding. This resulted in a change of total wetland area and, consequently, total deposition rates, causing a lower materials input. Eventually, the water table may fall in relation to the accreting surface, so that the wet dune slack becomes shallower and eventually it may dry out.

Rise in SL and a reduced sediment budget must have decreased the robustness of the beach foredune barrier to wave-driven breaching dependent on the coastal sediment budget and wave regime (Figure 7). Under such conditions, significant storm surges may have provoked flooding of the low-lying wetland region, by overwash and inundation of wetland areas behind the dune barrier. This statement is supported by the exclusive presence of the diatom *Chaetoceros* and the low TOC/TS values (<5 , Berner and Raiswell, 1984) in this period (~5400 cal. yr BP). The TOC/TS ratio also experienced low values between 5230 and 2500 cal. years BP suggesting marine influence. Extreme wave events and facies of correlated sediments in the 4th millennium BP (~4000, ~3550 and ~3150 cal. yr BP) are coincident with this low sedimentation period at Barbarroxa de Baixo (Rodríguez-Ramírez et al., 2015).

Climatic induced changes. On the Portuguese south coast, the coastal dunes that started to accumulate during the mid Holocene as the rate of SL rise decelerated, experienced two more phases of accumulation after about 3000 yr BP (Moura et al., 2007). These more recent episodes of dune building prevented storm-associated overwashing and promoted new peat formation during a period of more humid conditions (Figure 7). Actually, higher sedimentation rates detected after 2500 cal BP broadly coincide in time with a wetter period in Europe between 3000 and 2000 cal. yr BP (Giraudi, 1989; Leira, 2005) and in the Mediterranean area (Harrison et al., 1999; Roca and Juliá, 1997; Sadori et al., 2004). A more humid period occurred during 2600–1600 cal. yr BP is identified by Martín-Puertas et al. (2008) in the Zoñar Lake. Increase in precipitation during the late Holocene has probably created optimal conditions to re-initiate peat growth in a slightly elevated, well-drained settings (e.g. Winkler et al., 2001) thereby increasing the dune slack hydroperiod in the absence of a rising SL. Increased organic sedimentation and well-preserved DVs during the period 2500–2250 cal. yr BP correlate with an increase

in wetland plant species recorded in some SW Portugal coastal wetlands (Schneider et al., 2016). More frequent floods were common in several Spanish rivers during this period (Macklin et al., 2006) and some lakes in northern Africa (Sidi Ali Lake; Lamb et al., 2006) experienced high levels. The humid period 2600–1600 cal. yr BP was interrupted by one arid interval between 2100 and 1900 cal. yr BP (Martín-Puertas et al., 2008). The sterile level at 2200–2150 cal. yr BP in BB (90–92 cm depth) correlates with the abrupt decrease in flooding episodes at about 2350–2000 cal. yr BP (Macklin et al., 2006) and sharp lake-level decrease in African lakes (Lamb et al., 2006). A second period between 2100 and 1650 cal. yr BP corresponds with an increase in freshwater proxies and renewed sedimentation. Deposition during this second stage corresponds with increased arboreal pollen in the Tablas de Daimiel National Park (Gil García et al., 2007) from 2100 to 1680 cal. yr BP, climate improvement in NW Spain (Desprat et al., 2003) and increased flooding episodes in Iberian rivers (Macklin et al., 2006).

From 1650 to 800 cal. yr BP, BB experienced a large hydrological change. This is evident by a higher sand input and the increase in brackish diatoms, suggesting a progressive overall decline in freshwater influence and/or increasing proximity of the coastline. This is also detected in the Holocene record of the lagoons of Melides and Santo André (Cabral et al., 2006; Freitas et al., 2003). This signal is associated with the appearance of sand grains dispersed in the organic peat, which may indicate reactivation of aeolian activity and shifting of the bounding westward dunes towards the slack. Mayewski et al. (2004) documented a relatively weak climate change event characterized by cool poles and dry tropics occurring at 1200–1000 cal. yr BP, which coincides with the ice rafted debris (IRD) event that took place at 1400 yr BP (Bond et al., 1992) (Figure 9).

Warmer conditions and widespread aridity occurred in southern Europe after the Roman period, during the mediaeval climate anomaly (MCA). Lower average annual precipitation during the MCA is recorded in Soreq Cave (Bar-Matthews et al., 1999) and lower lake levels are inferred in central Italy (Dragoni, 1998; Issar, 2003) and northern Africa (Lamb et al., 2006). In Northern Spain, evidence for lower lake levels and decrease in floods during the 9th–11th centuries AD was found in the Iberian range (La Cruz Lake, Juliá et al., 1998; Taravilla Lake, Valero-Garcés et al., 2008) and the Pre-Pyrenean range (Morellón et al., 2007; Riera et al., 2004).

The low organic carbon content, dominance of benthic diatoms and the highest values of *P. trainorii* point to a drop in productivity and probably greater environmental instability after 800 cal. yr BP. This period represents a large hydrological change at about the transition between two well-documented global scale climatic events, the Dark Ages Cold Event (Keigwin and Pickart, 1999) and the MCA (Bradley et al., 2003; Diaz et al., 2011). The sand unit deposited at about 800 cal. yr BP is probably the result of a rapid event. The sand was sourced in the limiting dunes and entrained and transported by aeolian processes, or emplaced by abrupt collapse of the dune over the seaward margin of the slack, induced by slope instability. This is in agreement with contemporaneous phases of high aeolian activity that have been documented in different regions along Western Europe such as England (Bateman and Godby, 2004; Wilson et al., 2001), France (Bertran et al., 2011; Clarke et al., 2002) and Portugal (Costas et al., 2012). Sand deposition on Barbarroxa pond system can act as disturbance agent. Response to these sand depositional events are indicated at disturbance horizons by high percentages of small benthic fragilaroid diatoms, which are opportunistic and pioneering taxa, colonizing newly formed lacustrine habitats (Reed et al., 1999). The continued presence of scattered sand within the organic sediment suggests that these events have been persistent until present. This and the very broad ecological tolerances of species within this

complex (Reed et al., 1999; Stabell, 1985) indicate that the recent evolution at BB is marked by environmental instability and the resilience of the system to these episodes.

Implications for future evolutionary scenarios

Future climate scenarios for the Mediterranean region predict an increase in average temperatures, greater over the winter months, a decrease in annual precipitation and higher frequency of storm events (IPCC, 2013). Although coastal wet dune slacks develop beyond the direct influence of sea water, changes in SL may alter erosion/accretion patterns and affect groundwater levels. Hesp (2002) proposed different physical responses of foredunes to SL rise. SL rise and increasing frequency of surges and storm events are expected to increase dune erosion in Western Europe.

Empirical models for Portugal project an SL increase between 0.5 and 1.50 m for the period 2000–2100, with a central value of 0.95 m (Antunes et al., 2013). This rise will be accompanied by an increase in the mean relative SL (RSL) rise rate on the SW Portuguese coast of 1.9 mm yr^{-1} for the period 1920–2000 and 3.6 mm yr^{-1} for the period 2000–2010 (Antunes, 2013). According to our data, the sedimentation rate over the last 600 yr in BB was estimated to be 0.29 mm yr^{-1} . The long-term evolution of BB is controlled by the interaction among absolute SL, land elevation and sedimentation rate towards a balance with the RSL. This balance is adjusted upwards by increased sedimentation rate and downwards by increase in RSL rise rate. The predicted RSL rise over the 21st century exceeds the recent sediment accumulation rate; thus, sedimentation rate may be overcome by the rise in groundwater level leading to more frequent inundation. However, sedimentation is also likely to increase in response to increased inundation.

Sea-level changes since 11,200 cal. yr BP have been reconstructed for SW Portugal using data retrieved from coastal systems. Data from the Sado estuary yielded an RSL rise of 1.7 mm yr^{-1} between c. 7200 and 2800 cal. yr BP (Psuty et al., 2000), while results from the Tagus document a rapid rise from 12,000 to 7000 cal. yr BP followed by a negligible rise since then (Vis et al., 2008). Data from the southern coast (Teixeira et al., 2005) also suggest an attenuation at c. 7000 cal. yr BP, when mSL stood c. 2.5 m below its present position, and a definitive stabilization at 5000 cal. yr BP. Costas et al. (2012) document a continuous but slow SL rise after 6500 cal. yr BP with an accumulated change in elevation of about 2 m over this time. Evidence presented in Cruces (2016) indicates that the Barbaroxa system was able to withstand RSL rise rates up to 1.2 mm yr^{-1} , suggesting that this system is able to maintain against changes in relative mSL as long as water table elevation surpasses sediment loading. For BB the lowest limiting rate of RSL rise can be estimated to be at least 0.3 mm yr^{-1} . In fact, Cruces et al. (2010) have estimated a rate of 0.1 mm yr^{-1} for the period 5000 to 3500 cal. yr BP, coincident with the sedimentation hiatus observed between 5230 and 2500 cal. yr BP in BB, and far lower than the current rate of RSL rise. Considering the magnitude of the forecasted SL rise, it is likely that it will affect the long-term evolution of the system, inducing rejuvenation of the wet slack, if shoreline retreat allows it. Based on the sedimentary record, there is apparently no upper limit for the rate of RSL at BB. However, it is reasonable to expect that higher rates of SL rise will translate in more rapid and higher elevation of water surface, facilitating the turnover in the biotic communities out of their range of tolerance.

Water tables may rise due to SL rise, but changes in net precipitation due to climate change may further affect the groundwater level. Effective rainfall is the dominant influence on dune water table fluctuations in the short term (Clarke and Sanitwong Na Ayutthaya, 2010; Jones et al., 2006). Forecasted future decreases in rainfall and altered seasonality will have an impact on lower

dune slack water tables. The associated drying out will result in a loss of many rare species and may cause release of stored soil carbon due to faster decomposition. Depending on local context, it will also bring associated changes in the volume and timing of freshwater discharge with the potential to alter salinity regimes. This effect can be taken further by the rise in SL (Saye and Pye, 2007) and cause an increase in salinity in coastal slacks such as BB, depending on the pace of the rise (Greaver and Sternberg, 2006). Combined with prediction of decrease precipitation (De Castro et al., 2005), it is expected that dune ecosystems accumulate ocean water in the absence of rain during dry periods, therefore elevating salinity levels.

Sandy shores will face a concomitant enhancement of the erosion in response to the predicted changes in wave climate (Carter, 1991) that may be most evident during storm surges (Wigley, 1999). Beach erosion is the dominant trend for all mainland Portuguese beach–dune systems, with a mean change rate of $-0.24 \pm 0.01 \text{ m yr}^{-1}$ over the last half century (Lira et al., 2016). The Tróia–Sines coastal segment shows the highest average progradation trend for all sediment cells in mainland Portugal ($+0.45 \pm 0.01 \text{ m yr}^{-1}$) although this result is strongly biased by the extreme progradation observed at its northernmost tip. This coastal stretch has been alternating low-magnitude erosion and progradation throughout the last half century with no clear long-term trend. So, it is likely that foredunes will move both upwards and landwards as SL rises. In the worst-case scenario, RSL rise may contribute to dune degradation and the increased possibility of extreme coastal flooding and overwashing during storms so that dune wetlands behind coastal dune ridges may be exposed to marine processes.

Here, we have used palaeoecological techniques to understand environmental change specifically in dune slacks. Primarily, Barbaroxa de Baixo development and evolution is related to coastal processes but the response to environmental change is expressed in the geomorphology, hydrological processes, hydrochemistry, sedimentary infill and nutrient content of the wetland. Overall, the results show the resilience and relative stability of this ecosystem to RSL rise and climate change on millennial timescales. During the mid and late Holocene, the environment of the Barbaroxas lowlands experienced periods of stability interspersed with episodes of rapid change, in which SL rise rates and changes in sediment supply are the most important processes. Perhaps more importantly, records such as the one studied here provide relevant information about how SL rise and climate change may affect these systems. Knowledge of their past and related evolution patterns and trends may help us to anticipate responses to future changes. Notwithstanding, the impacts of future changes in climate expected by the end of the 21st century will most probably match similar intensity to those observed in the past but much faster; therefore, the exact nature of the impacts remain still uncertain and further research is required in order to increase confidence in modelled forecasts, making them progressively more useful to support objective adaptation measures. The nearby deposits located at different topographic elevation and distance from the coastline offer a good opportunity to improve our understanding of these systems.

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