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Proxy records of Holocene storm events in coastal barrier systems: Storm-wave induced markers

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ABSTRACT

Extreme storm events in the coastal zone are one of the main forcing agents of short-term coastal system behavior. As such, storms represent a major threat to human activities concentrated along the coasts worldwide. In order to better understand the frequency of extreme events like storms, climate science must rely on longer-time records than the century-scale records of instrumental weather data. Proxy records of storm-wave or storm-wind induced activity in coastal barrier systems deposits have been widely used worldwide in recent years to document past storm events during the last millennia.

This review provides a detailed state-of-the-art compilation of the proxies available from coastal barrier systems to reconstruct Holocene storm chronologies (paleotempestology). The present paper aims (I) to describe the erosional and depositional processes caused by storm-wave action in barrier and back-barrier systems (i.e. beach ridges, storm scarps and washover deposits), (ii) to understand how storm records can be extracted from barrier and back-barrier sedimentary bodies using stratigraphical, sedimentological, micro-paleontological and geochemical proxies and (iii) to show how to obtain chronological control on past storm events recorded in the sedimentary successions. The challenges that paleotempestology studies still face in the reconstruction of representative and reliable storm-chronologies using these various proxies are discussed, and future research prospects are outlined.

1. Introduction

Established along the depositional shorelines of wave-dominated coasts, beaches and barriers are in a foremost position to undergo the effects of storms (e.g. Masselink and van Heteren, 2014). Within the worldwide context of the recent and ongoing rising sea level (Church and White, 2011; Hay et al., 2015; Kopp et al., 2016), inundation levels are likely to increase in most coastal areas, even with an unchanged storminess regime (e.g. FitzGerald et al., 2008; Williams, 2013; Horton et al., 2014; Clemmensen et al., 2016). As a consequence, recurrence intervals of storm-induced inundation levels will likely be shortened (Masselink and van Heteren, 2014), thus threatening many coastal regions worldwide along which populations densities keep on increasing (FitzGerald et al., 2008). Storminess patterns in coastal regions are also likely to be strongly influenced by global warming. Many climate-modelling studies have been carried out (e.g. Oouchi et al., 2006; Vecchi and Soden, 2007; Gastineau and Soden, 2009; Fant et al., 2016) and showed that, as warming progresses, subtropical zones of dry-air subsidence may expand, most probably resulting in (i) a poleward shift of the mid-latitude storm tracks and (ii) a deepening of the convective tropical storms (Oouchi et al., 2006; Vecchi and Soden, 2007; Stephens, 2011). In Europe, for instance, extreme wind speeds are expected to increase between 45°N and 55°N (except over and south of the Alps, Beniston et al., 2007; Gastineau and Soden, 2009) and to become more northwesterly-orientated than nowadays. Some lowering of the mean sea-level pressure is also likely to occur, leading to more North Sea storms and a consequent increase in storm surges along coastal regions of Holland, Germany and Denmark, in particular (Beniston et al., 2007).

1.1. "Paleotempestology": previous achievements and present challenges

To be adequately addressed, the future changes in wind regimes (including in particular the frequency and magnitude of storm events) have to be placed in the context of long-time records of past storminess. Indeed, in most regions, storm magnitude and

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frequency are precisely known from instrumental and historical data for the last 50–200 years (e.g. Dawson et al., 1997; Clemmensen et al., 2014a). Some historical data contain additional information on storm events up to 1000 years B.P. (e.g. Lamb and Frydendahl, 2005; Clarke and Rendell, 2009). However, high-resolution data on storminess in coastal regions are scarce over longer (millennia) time spans. Data are often lacking to allow the estimation of the recurrence intervals of the most extreme events or even to document storminess variation on climate (decadal to century) time scale. Longer-term data on storm histories are therefore needed worldwide to improve the evaluation of the recurrence intervals of extreme storms and to subsequently more reliably assess future storm surge risk. Holocene coastal sedimentary sequences can enable this. As mobile barrier and back-barrier systems were emplaced worldwide during the Holocene, they were impacted by storm waves and surges, some of which left characteristic depositional or erosional evidences in the sedimentary records. Due to inland aeolian transport of sand particles during periods of active storminess, storms frequently also left traces within more landward terrestrial domains (e.g. De Jong et al., 2006; Orme et al., 2016; Nielsen et al., 2016).

Storm-wave induced markers represent highly-valuable geo-archives of past storminess and can form the basis for extended storminess chronologies in many coastal areas if they can correctly be identified, unambiguously related to storm events and dated with sufficient precision. This field of study, termed "paleotempestology" (Hippensteel, 2010), began to develop during the 1990's through studies on the sand barriers of the eastern coast of the USA and aimed to reconstruct past-hurricane events (e.g. Liu and Fearn, 1993; Hippensteel and Martin, 1999). Since then, most of the works conducted in this field have been centered on reconstructing past-hurricane and tropical cyclone activity (notably because these latter bear the most extreme impacts on very densely populated coastal areas, particularly along the eastern coasts of the USA). But interest in paleotempestology has also spread worldwide and additional proxies have been developed to reconstruct past storminess from a wider-array of sedimentary contexts and regions, including those concerned by extra-tropical storms. Most recently there has been an increase in studies of European coastal systems, and several robust Holocene storm chronologies have been extracted from barrier and back-barrier deposits including some from Atlantic (e.g. Billeaud et al., 2009; Cunningham et al., 2011; Van Vliet Lanoë et al., 2014), Mediterranean (e.g. Sabatier et al., 2010a, Sabatier et al., 2012; Degeai et al., 2015; Raji et al., 2015; Dezileau et al., 2016) and North Sea coasts (e.g. Fruergaard and Andersen, 2013). Promising attempts to reconstruct Holocene storminess chronologies have also been derived from the traces left by storm-induced aeolian processes within marginal coastal and terrestrial domains (e.g. Björck and Clemmensen, 2004; De Jong et al., 2006; Clemmensen et al., 2009; Parris et al., 2010; Costas et al., 2012; Forman, 2015; Nielsen et al., 2016; Orme et al., 2016).

Notwithstanding these recent efforts, high-resolution Holocene storm chronologies are still relatively few, notably with regard to extra-tropical storms. Patterns of past storminess are therefore still imperfectly understood and extrapolations towards reliable scenarios of future storminess patterns remain uncertain. A significant challenge paleotempestology faces is the question of the representativeness of local storms records relative to the regional and global processes causing storminess activity. As already noted by May et al. (2013), storm-reconstructions derived from local sedimentary archives essentially carry local significance and, similarly to what has been evidenced for relative sea-level reconstructions based on sedimentary evidences, should be extrapolated only with great care to provide regional to extra-regional significance (Otvos, 2011). Indeed, each site is characterized by specific sensitivity thresholds (defined by several parameters such as the local morphological configuration of the coast, sediment variability and dynamics) which determine the way it is impacted by a given storm and is chance to record and archive this event as a sedimentary feature. A common feature of past-storminess studies is that the recurrence periods of reconstructed events are often observed to be longer than can be expected from modern and historical records (Hippensteel, 2010). This discrepancy raises doubts on whether (i) every single storm has been recorded by the studied sites and on whether (ii) storm testimonies have progressively disappeared from the sedimentary records. Following an actualistic approach, most reconstructions of past storminess use observations on the impacts left by recent storms, to build modern analogs of storm indicators and try to understand the spatial and temporal patterns of the impacts lefts by storms of known track and intensity. Such an understanding of the spatial distribution of the impacts left by modern storms of known tracks and intensities is crucial in trying to figure out where the most complete past storm records can be located. This is all the more important as paleo-tempestology seeks to move towards higher temporal (ideally centennial to decadal) and spatial (regional to local) resolutions, in order to adequately model and forecast the parameters driving storminess.

1.2. Review rationale and outline

In recent years, the impacts of storms on coastal sedimentary systems and past-storminess reconstructions have received a considerable attention and several synthesis papers addressing various aspect of this emerging field of research have been published. For example, Masselink and van Heteren (2014) reviewed the response of wave and mixed-energy barriers to storms in a wide geographical framework, primarily looking at modern to subrecent coastal changes. Chaumillon et al. (2017) documented the use of historical archives, sedimentary proxies and numerical hydrological models outputs in an overview of storm-induced marine floods within the framework of future coastal management and adaptation. Other review papers have focused on specific sedimentary systems and/or on specific proxies for past-storminess reconstruction. Scheffers et al. (2012) and Tamura (2012) specifically focused on beach-ridge systems and their use to extract knowledge on pre-historic catastrophic events including storms. Otvos (2011) relied on several studies of the Holocene hurricane activity conducted along the coasts of the Gulf of Mexico, to present some of the proxies that can be used for storminess reconstruction, with a large space being devoted to discuss the limitations and pitfalls of each proxy. A useful review of the panel of available proxies was made by May et al. (2013) with a focus on tropical cyclones along the world's coastlines. Their work presents the use of washover sediments, beach ridges and high-resolution coral and speleothems proxy records to reconstruct past storminess, and draws up a valuable state-of-the art picture of the open questions and challenges faced by paleotempestology. However, to date, the characteristics and related dating methodologies of the full assemblage of proxies that can be used to document Holocene storm events from coastal barrier and back-barrier sedimentary sequences have not been reviewed in detail.

This review builds upon the works referenced above and aims (i) to summarize storm-induced wave impacts on coastal barrier systems and describe how these events are recorded in the sediments and (ii) to present the methodologies that can be used to extract proxy-data from the sedimentary records. Focusing on sandy and mixed sandy and gravelly barrier systems, we present the three main elements which can form during storm events; namely beach ridges, storm scarps, and sedimentary features related to washover events (Fig. 1). This review aims to describe the

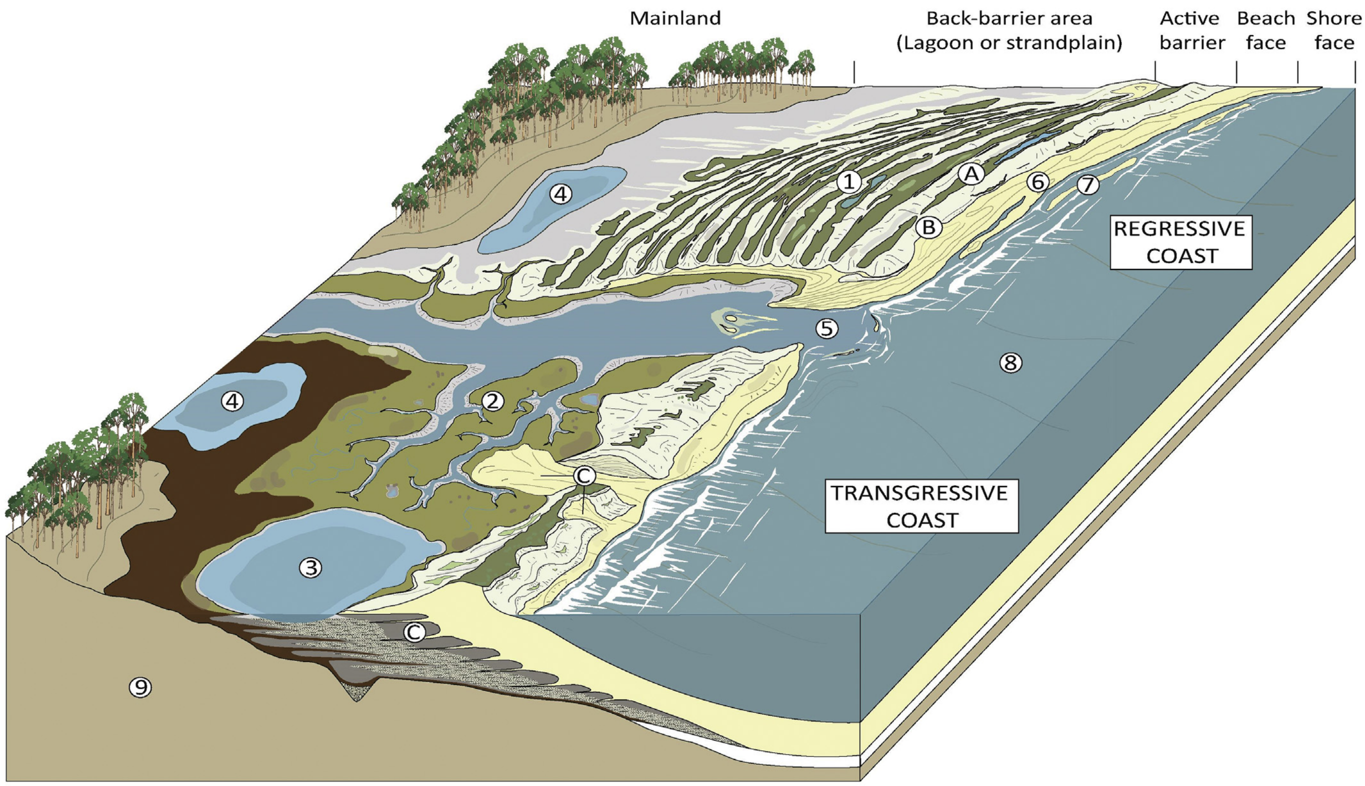


Fig. 1. Schematic illustration of transgressive and regressive costal barrier systems. Storm-wave induced markers are: A-Beach ridges, B-Storm scarps, C-Washover features. Coastal barrier sub-environments are: 1-Beach-ridge succession (strandplain); 2-Back-barrier lagoonal saltmarsh; 3-Brackish coastal mire; 4-Freshwater coastal lake, 5-Tidal inlet & flood/Ebb delta sedimentary features, 6-Beach berms, 7-Swash bars, 8-Shoreface, 9-Mainland/bedrock.

emplacement processes of each of these features, the principles guiding their identification within Holocene sedimentary archives and the methods used to date them. Finally, the limitations and challenges pertaining to the use of each proxy in interpreting past storm events are discussed.

2. The "Storm impact regimes" and influencing factors

During onshore storms, beach, barrier and back-barrier areas are influenced by the action of enhanced hydrodynamic processes, as they deviate from their fair-weather regimes (Masselink and van Heteren, 2014). Extensive description of the parameters controlling the impacts of storm waves and surges to beaches and barriers is outside the scope of this review and readers are referred to Donnelly et al. (2004, 2006), Anthony (2009) and Masselink and van Heteren (2014) for complete and extensive syntheses on the physical processes induced by storm waves. For a given event, and all other parameters being equal, the type of morphological impacts undergone by a coastal barrier will depend on the heights of the dune-foot and dune-crest relative to the maximum elevation reached by the storm water-level (Sallenger, 2000; Priestas and Fagherazzi, 2010; Masselink and van Heteren, 2014). Storm water-level is the sum of (i) the astronomical tide level, (ii) the height of the storm surge and (iii) the wave setup and runup (Fig. 2; Stockdon et al., 2006). It has been proposed that the types of impacts caused by storm waves on a barrier may fall into four main regimes (Sallenger, 2000) (Fig. 3), any of which being able to promote either erosional or depositional features (Morton and Sallenger, 2003). These four regimes have recently been re-described in detail by Masselink and van Heteren (2014) and are only very briefly described here:

(i) The "Swash regime" (Fig. 3a) is characterized by an elevation of the swash runup (or runup) remaining below the elevation of the dune foot. The "Swash regime" implies the storm-induced processes to be confined to the beach and foreshore domain. It typically causes erosion of the beach-profile but can also promote the accretion of beach berms and their migration towards the upper beachface.
(ii) The "Collision regime" (Fig. 3b) occurs when runup reaches the dune foot. A typical consequence of the collision regime is that wave and swash actions cut into the barrier, causing an erosional scarp often termed "storm scarp".
(iii) The "Overwash regime" (Fig. 3c) takes place when a beach-berm or, if present, a dune is overtopped by the wave runup. Energy dissipation is then mostly achieved through the overwash bore running over the barrier, flattening the barrier crest profile and depositing offshore originating material over the back-barrier zone. Ultimately, particularly intense or repeated confined overwash events can promote the opening of an inlet between the ocean and the back-barrier lagoon/coastal lake.
(iv) The "Inundation regime" (Fig. 3d) describes the situation when the storm level is so high that water completely submerges the barrier. Widespread inundation of the barrier and of the back barrier lowlands can then occur, possibly depositing extensive amount of sedimentary material over widespread barrier and back-barrier areas. The "Inundation regime" is also responsible for bringing massive volumes of salt-water into the brackish to freshwater marshes or lagoons, which can induce ephemeral ecological modifications.

It should be noted than the range of beach-barrier reactions to

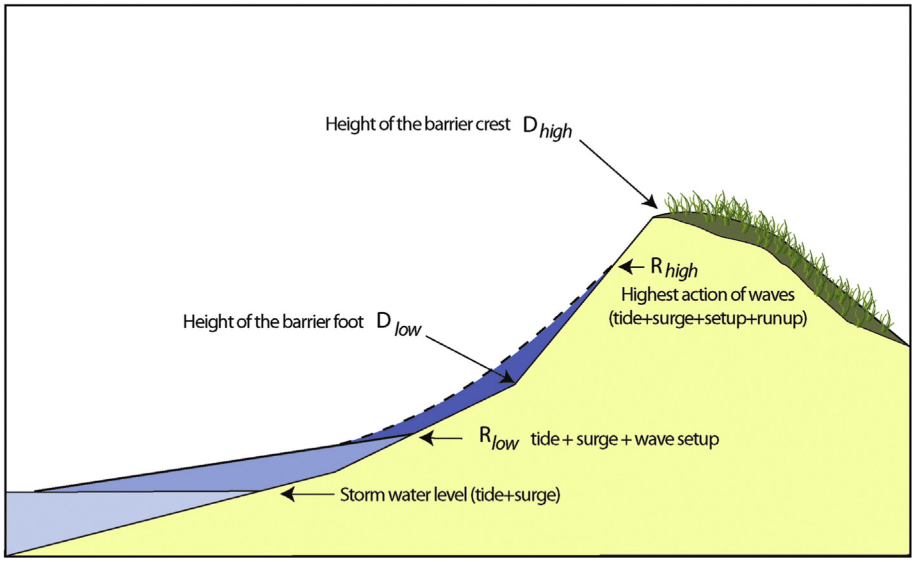


Fig. 2. Definition sketch of the parameters used to define the Storm Impact Scale of Sallenger (2000). Modified from Stockdon et al. (2006) after Sallenger (2000) and Masselink and van Heteren (2014). The dashed line represent the maximum swash excursion (runup <η>) from the wave setup level (solid line).

storm-wave action is much more complex than can be summarized by the four regimes above. As described in detail by Morton (2002) and Masselink and van Heteren (2014), the way and the magnitude a beach-barrier system is impacted by a storm (and thus the likelihood of preservation of this precise event) are highly dependent on the interplays between numerous forcing factors and are thus highly variable in time and space. The first group of factors characterizes the storm itself (wind direction(s), sustained wind speeds, maximum gust speed, orientation of the storm path relative to the coastline, speed of the storm displacement, atmospheric pressure variation and timing of the storm with regard to the tidal cycle). The second group of factors pertain to the location and configuration of the site (e.g. location of the site relative to the storm track, fetch distance, orientation with respect to the storm winds and waves) and to its morpho-sedimentary characteristics (topography of the beach and of the offshore shelf/platform, tidal regime, sediment availability, sediment texture, vegetation cover).

3. Beach-ridge systems

Storm waves can cause both erosional and depositional features in the beach and frontal dune domains. In the following section, we shall focus on the uses that have been made of prograding coastal barrier stratigraphies for reconstructing high-energy wave events by considering beach-ridges morphologies.

3.1. Definition, formation processes and characteristic features

Beach-ridge systems have now been studied for almost a century (e.g., Johnson, 1919, 1965; King, 1959; Psuty, 1967; Hine, 1979; Møller, 1995; Komar, 1998; Isla and Bujalesky, 2000; Otvos, 2000; Neal et al., 2003; Clemmensen et al., 2012) and have received quite a broad range of definitions (as listed by Otvos, 2000). We will retain that beach-ridge systems are progradational sedimentary systems which are composed of a succession of shore-parallel ridges (the beach ridges) of more or less pronounced topography typically intercalated with elongated relatively low-lying (lower than the surrounding ridge crests) and rather flat areas termed swales (e.g. Otvos, 2000; Clemmensen et al., 2012; Bendixen et al., 2013, Fig. 1). Most beach-ridge systems are either sandy (siliciclastic material) or gravel-rich morphological features. Systems composed of shell material or coral fragments or rubbles do also exist but are less frequently reported (e.g. Nott and Hayne, 2001; Nott, 2003, 2011; Spiske and Halley, 2014). Otvos (2000) and, later, Hesp et al. (2005) insisted on the necessity to distinguish between wave-built ridges and back-shore ridges influenced by aeolian processes; the "eolian beach ridges" of Otvos (2000). The present article follows the terminology of Otvos (2000) and Hesp et al. (2005) and uses the term "beach ridges" to refer to swash-aligned depositional features, formed above the mean high water spring tide level as the result of wave action and separated by swales. In agreement with Otvos (2000) and Bendixen et al. (2013) the term beach ridge here only encapsulates relict features separated from the shoreline by progradation; while active structures are termed berm ridges.

The mechanisms of beach-ridge formation have been considerably debated (Otvos, 2000; Hesp et al., 2005; Anthony, 2009; Tamura, 2012). The variety of processes reported may result from the fact that the formation of such ridges obey to an assemblage of several interdependent parameters (Scheffers et al., 2012) which can both differ from one system to another (Nott et al., 2009) and in between the different sets of ridges in one system (Anthony, 2009; Nott et al., 2015) as well as be variable in time. Two main modes of beach-ridges formation have been reported (Carter, 1986; Anthony, 2009). A first mode involves the progressive building and landward migration of beach berms under intermediate to dissipative conditions, following the landward migration of coalescent nearshore bars formed by constructive waves (Masselink et al., 2006). The bars eventually become welded to the beach (generally to the lower beachface) as incipient berms. Berms then grow both vertically and horizontally during successive high-water levels both under the action of repeated up-rush sediment transport in the surf zone or as a result of non-erosive overwash events depositing material on top and landward of the berms (Masselink et al., 2006; Weir et al., 2006; Bendixen et al., 2013) until they reach a "mature berm" or "berm ridge" structure (Fig. 6 of Otvos, 2000 and Fig. 4 of the present article). The development only stops when, because of coastline progradation, the berm reaches a position located too far from the shoreline to be reached by storm waves. This results in the berm being abandoned and turned into a beach ridge, while new

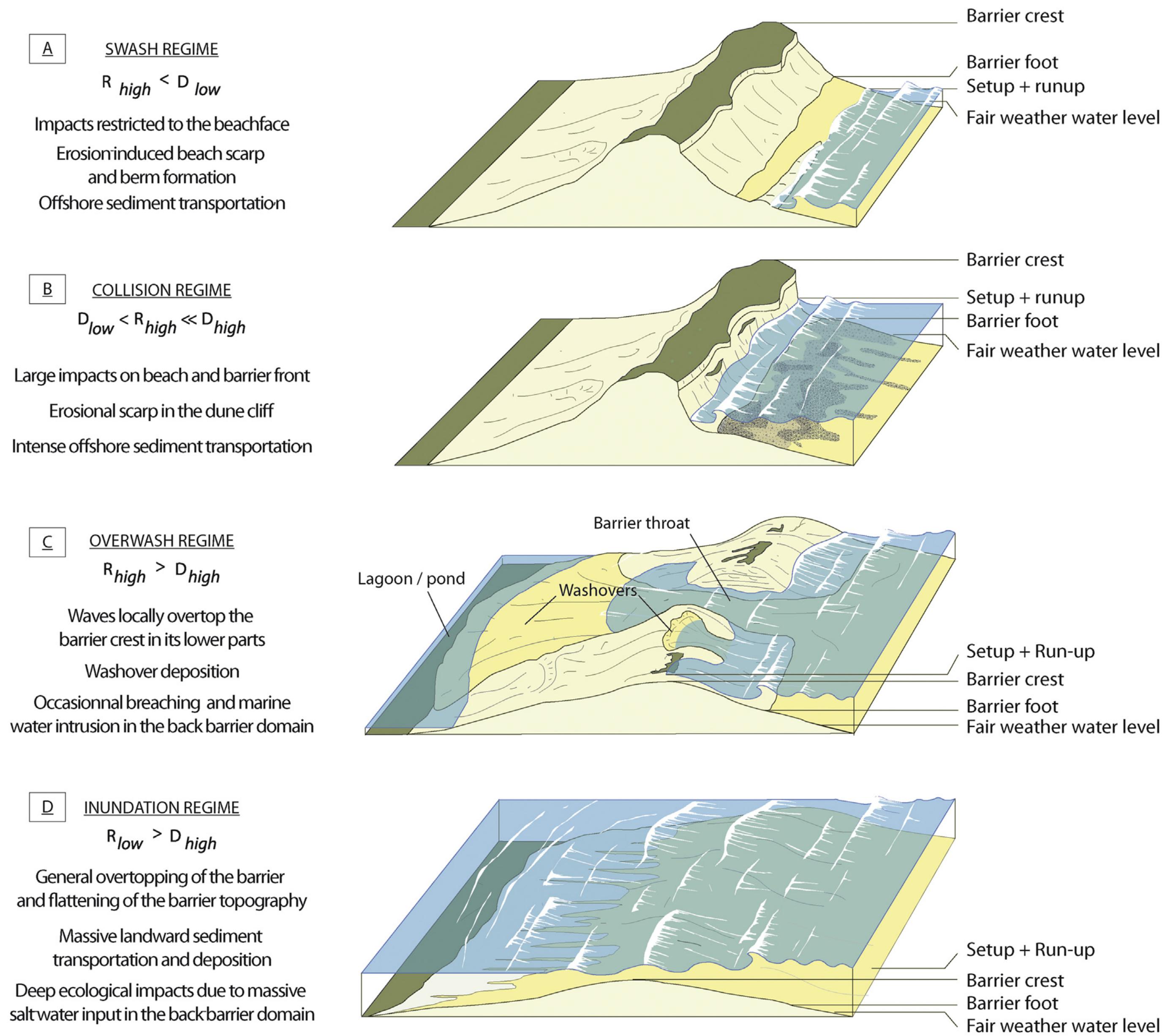


Fig. 3. Schematic illustrations of barrier systems under the four storminess regimes of Sallenger (2000). The topographical and hydrological conditions needed to enter these regimes along with the main impacts implied by each of them are summarized. RLow, RHigh, DLow, DHigh refer to the parameters defined in Fig. 2.

berms are being built seaward. A second mode generally considered to promote beach-ridge formation and progradation involves the progressive build-up and aggradation of a berm situated landward of the active beach by accumulation of material eroded from the beachface and deposited landwards without the need of bars to build-up and migrate on the beach face in the first place. This situation is normally encountered in more reflective situations where nearshore bars are generally absent, and is promoted either by grain-by-grain particle transport (Anthony, 2009) or more intensively by storm-wave action (Nott et al., 2013; Clemmensen et al., 2014b, 2016). It is important to consider that any of these modes can have relayed each other both in time and space due to complex feedback dynamics between beach morphology, sediment availability and wave energy, and this either between the different sets of beach ridges comprised in one single system, or even within one single set.

The processes that govern the formation of swales remain poorly understood. Most probable hypothesis is that swales may result from the isolation of the beachface induced by the welding of a berm. This was confirmed by the observation in many systems of seaward dipping beachface deposits underlying the swales, indicating that the part of the backshore which turned into a swale originally was part of a low-relief beachface (e.g. Nielsen and Clemmensen, 2009). Swales may evolve during the inundation of the backshore during periods of elevated water levels (induced either by water overtopping the berm ridges or an elevated water table or both), with their flat shapes in sandy environments being subsequently shaped by aeolian deflation of the area (Clemmensen et al., 2012). Following coastal progradation, swales eventually become covered with water and turn into coastal mires, thus favouring peat formation (Clemmensen et al., 2001, 2012).

A few authors have reported beach-ridge formation to be related

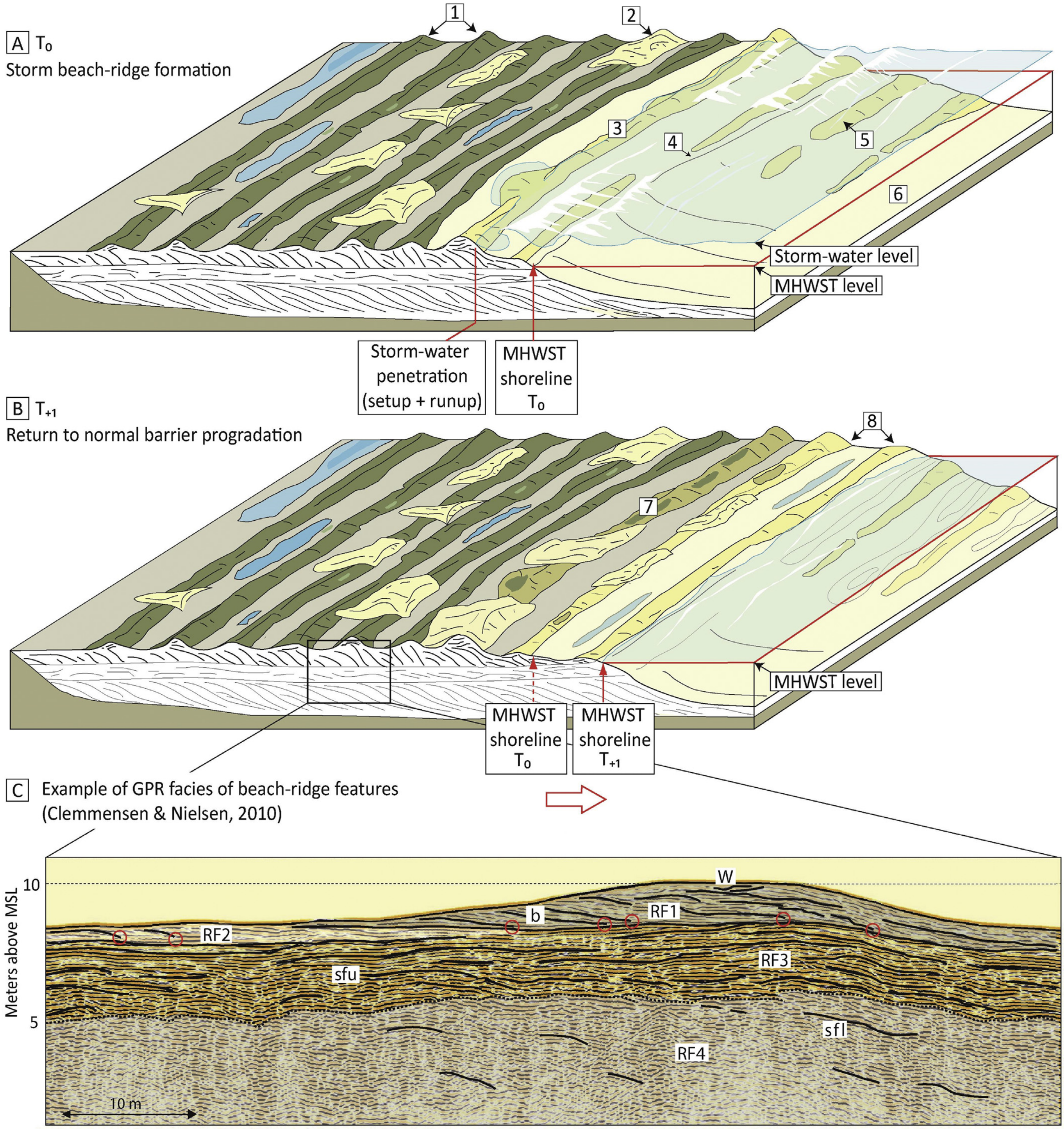


Fig. 4. Schematic illustration of the development of an idealized beach-ridge system and of the stratigraphical/sedimentological features usable for past-storminess reconstructions. **(A)** Building of a storm ridge under elevated water level and high wave energy involving or not the progressive migration of shoreface bars and **(B)** Return to the state of normal fair-weather barrier progradation. MHWST = Mean High Water Spring Tide. 1-Stabilized and vegetalized beach ridges, 2-Strandplain dunes, 3-Storm built ridge, 4-Upper-beach berm, 5-Migrating shoreface swashbars and berms, 6-Lower shoreface, 7-Recently fixed beach ridge. The red arrow shows coastal progradation. **(C)** GPR profile obtained across the beach-ridges succession on the southeastern part of the Island of Anholt (Northern Denmark) by Clemmensen and Nielsen (2010), showing the main GPR facies of a beach-ridge succession. Vertical exaggeration is 1:2. The following features are indicated: b = beachface deposits; w = washover deposits; sfu = upper shoreface deposits; sfl = lower shoreface deposits. Red circles show "downlap points". See Clemmensen and Nielsen (2010) for details on the description and interpretation of the radar facies RF1, RF2, RF3, RF4. Figure (C) reproduced with the permission of Elsevier. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

to fair-weather processes (e.g. Tanner, 1995; Goy et al., 2003), but most of the literature agreed on that elevated water levels produced by high-energy waves during onshore storms (even distant) were the most common mechanisms responsible for berm initiation and building (e.g., Johnson, 1919; Hine, 1979; Sandweiss et al., 1998; Komar, 1998; Isla and Bujalesky, 2000; Neal et al., 2002; Hesp et al., 2005; Weir et al., 2006; Nielsen and Clemmensen, 2009; Nott et al., 2009, 2013; Clemmensen et al., 2012; Nott et al., 2013; Bendixen et al., 2013). Accepting this as a starting point, a major matter of debate concerns whether storm-driven beach ridges are the result of single or multiple storm events. Such question is central since it will bear considerable differences when calculating recurrence intervals. Unique storm events could have been responsible for the emplacement of single shingle and shelly beach ridges beach ridges (Nott, 2003, 2011) but it appears that most sandy to gravelly beach-ridges are probably formed over time as the result of several storm events (e.g. Nott et al., 2009; Bendixen et al., 2013). In their study of the Feddet Spit system (south-eastern Denmark), Bendixen et al. (2013) reported that beach-ridges formation was observed to occur during only 20% of all storm events with a return period of once building event every three years in average. Considering that the ridges forming the Feddet system are separated by 180 years in average, Bendixen et al. (2013) concluded that each beach ridge could only be the result of several storm events. The authors suggested that berm heights would only probably reflect the height of the highest (or latest) inundation level over the period during which the considered ridge formed, as exemplified by the observation of the impacts the 1872 Baltic storm flood left on the site. Nott et al. (2013) documented accreting sandy beach ridges units in northeast Queensland (Australia). Thanks to eyewitness accounts and historical reports, each unit in the beach ridge could be related to a specific tropical cyclone and beach-ridge build-up was thus shown to be the result of a number of storm events.

3.2. Identification of beach ridges within the sedimentary archives and use as proxies of past-storminess

3.2.1. Principle of identification

Beach ridges have been shown to constitute valuable archives of coastal evolution and their morphology, and internal structures contain information on both past relative sea levels (e.g. Nielsen and Clemmensen, 2009; Clemmensen et al., 2012; Hede et al., 2013, 2015; Sander et al., 2016) and past storminess activity (e.g. Nott et al., 2009; Nott, 2011; Bendixen et al., 2013; Clemmensen et al., 2016). For such information to be retrieved, a precise mapping (and dating) of the beach-ridge successions is needed. The identification and mapping of beach ridges along shorelines that have long been prograding is generally relatively easy, thanks to the alternating ridge-swale morphologies. In most cases, however, aeolian sand covers prevent direct observations and mapping of the wave-build part of the ridges, so that non-destructive imagery methods and drillings are required (e.g. Clemmensen et al., 2014b). The advent of ground-penetrating radar (GPR) mapping (Fitzgerald et al., 1992; Neal et al., 2002, 2003; see Neal, 2004 for review) has made it possible to investigate in detail the internal structures of barrier and beach-ridge systems (e.g. Clemmensen et al., 1996, 2001; Jol et al., 1996; Meyers et al., 1996; Clemmensen et al., 2001; Bristow and Pucillo, 2006; Nielsen and Clemmensen, 2009; Hede et al., 2013, 2015). The aforementioned studies revealed that the internal structures of beach ridges are typically dominated by low-angle seaward dipping beachface deposits, sometimes (but less frequently) associated with landward dipping strata of overwash origin (Fig. 4, Nielsen and Clemmensen, 2009; Clemmensen and Nielsen, 2010; Clemmensen et al., 2014a,b). The transition between the beach-face and the shoreface domains was not only marked by a change in the sedimentary facies, but also revealed by a clear change in the dip of the strata from relatively steeply dipping beach-face strata to less steeply dipping upper shoreface strata defining a “downlap” point (Nielsen and Clemmensen, 2009; Hede et al., 2013; see Fig. 4). The precise mapping of the heights of this downlap level within the internal bodies of the beach-ridge systems permitted by the GPR was demonstrated to be able to produce Relative Sea-Level (RSL) data (Nielsen and Clemmensen, 2009; Clemmensen et al., 2012; Hede et al., 2013, 2015; Sander et al., 2016). As will be discussed below, since beach ridges are considered to be formed and shaped by storm waves, the size of these morphological features was also considered likely to provide information on past storminess intensities.

3.2.2. Usefulness for extracting information about paleo-storms intensities

Studies of the formation of modern beach ridges during storm conditions have shown that, in many cases, a direct relationship could be observed between the elevation of the ridges and storm wave runup (Bendixen et al., 2013; Masselink and van Heteren, 2014), thus suggesting that the elevation of both sandy and gravelly fossil beach ridges could be used to quantify the inundation level of past storm events. Bendixen et al. (2013) for instance showed that sandy berms formed under elevated water level and high-energy onshore wave conditions along a micro-tidal shoreline in southeastern Denmark, and demonstrated that the height of mature active berms was directly related to the height reached by storm wave induced runup. Clemmensen et al. (2016) reported that the elevation of a gravelly storm berm that developed during a major storm in 2013 along a micro-tidal shore in NW Zealand (Denmark) was the sum of the still water level in the adjacent sea (the Belt Sea) during the storm and of the wave runup at the shore. If the original morphology of a gravel-rich structure is being preserved during coastal progradation, then its elevation above past (contemporaneous) sea level can be considered a relatively precise measure of the inundation level during an onshore storm (Clemmensen et al., 2016). In meso- and macro-tidal environments, the amplitude of the tidal range and the timing of the high-tide level vs. the storm-surge occurrence will strongly control the height of the storm-surge level (Nott et al., 2009; Nott, 2011) and, thus, the influence storm waves will have on the berm topography. With the aim to relate beach-ridge topography to the magnitude of cyclones, Nott et al. (2009) investigated a succession of 29 shore-parallel sandy beach ridges located at Cowley Beach in north-east Queensland (North-eastern coast of Australia, Fig. 5). The crests of the ridges were observed to lie between 3.5 and 5.5 m (mean 5 m) above Australian Height Datum (0 m AHD corresponding to the mid-spring tide level). Storm surge and wave modelling results suggested that a cyclone of minimum category-4 strength would be required to build a ridge that formed at +5 m AHD, while a very intense storm could not be responsible for ridge formation at +4 m AHD (Fig. 5). Thus, the results of Nott et al. (2009) provided evidence of tropical cyclones of category 4–5 occurring at a centennial recurrence period during the last millennia, while analysis of historical records suggest that such cyclones were occurring at a millennial scale frequency. Nott et al. (2013) suggested that the lowermost units of the composite beach ridges they studied, whose crest are situated only a few meters above mean sea level, could have possibly deposited over a wide range of inundation levels (ranging from cyclonic extreme inundation levels to non-cyclonic ones for example driven by strong trade winds and/or high water spring tides). On the contrary, increasingly higher inundations and higher magnitude storms have been undoubtedly needed to deposit the uppermost units in the ridges. As such, Nott et al. (2013) concluded that, in any case, the elevations of beach-ridges crests only provide minimum estimates of storm-water levels and that the upper-units forming a beach-ridges were best to be used for this latter purpose.

However, if the elevations of beach ridges are to be used as a proxy of past storm intensities then any long-term trend that may

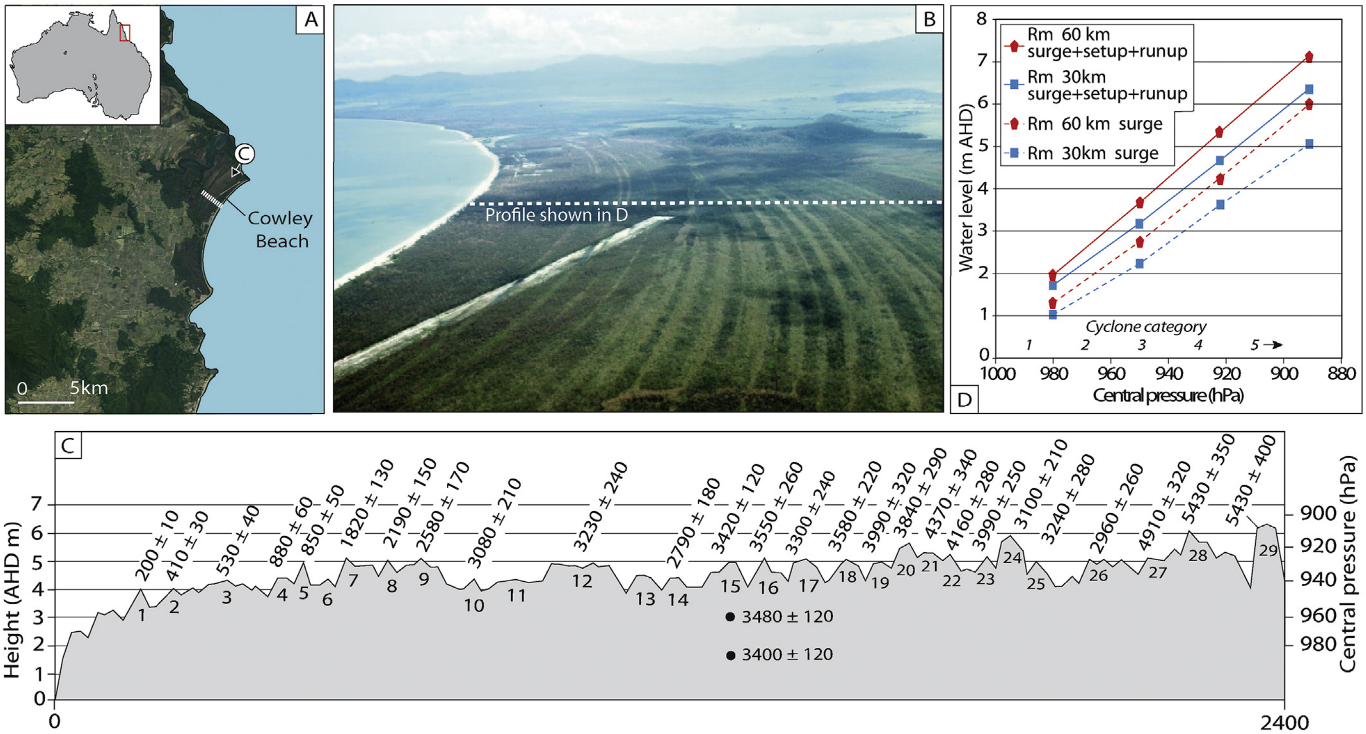


Fig. 5. **(A)** Location of the beach-ridge succession studied by Nott et al.(2009) at Cowley Beach (Queensland, north-eastern Australia). **(B)** Oblique aerial photograph of outer barrier (Holocene) beach ridge plain. Darker colours represent ridge crests (picture courtesy D. Hopley). View is to south. The dashed line depicts the location of the transect shown in C). **(C)** Topographic profile of the beach-ridge plain at Cowley Beach. Numbers correspond to beach-ridge individuals. Mean OSL ages are noted above each ridge. **(D)** Model outputs of storm water level (surge alone and surge + wave set up and wave run up) obtained at Cowley beach for hurricane ofdifferent pressure and intensity (Rm stands for radius of maximum winds). All figures modified from Nott et al. (2009) and reproduced with the permission of Elsevier.

• *Each beach ridge is here considered to have formed following several hurricane event. The number of beach ridges is thus interpreted to reflect the minimum number of hurricane events having made landfall in this region during the past ca. 6000 years. The height of each beach-ridge is proposed to reflect the water-level at the coast caused by the most intense hurricane and is related to cyclone intensity (Saffir-Simpson scale) through modelling. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)*

have modified the relationship between beach-ridge topography and water level should be understood and detrended before any analysis of the magnitude of past storm events can be done. As an example for the purpose of this review, we re-analyzed the topography of the raised beach-ridge system described by Clemmensen and Nielsen (2010) and Clemmensen et al. (2012) on the island of Anholt (Denmark, Fig. 6). This suite of progradational gravel-rich beach ridges formed during the last 7000 years and an overall fall in relative sea level fall, at an inferred rated of one beach-ridge emplaced every 15 years (Fig. 6). Using an average of the RSL data retrieved from this beach-ridge system by Clemmensen et al. (2012) on the basis of downlap points (Fig. 6), we detrended the height of the beach ridges from the RSL tendencies by subtracting the RSL curve from the beach-ridge topography. The beach-ridge topography corrected from the changes in RSL reveals three groups of beach ridges each characterized by very different morphologies and elevations (Fig. 6). In a first hypothesis, these three sets of beach ridges can be interpreted to reflect three periods of differing past-storminess climate between 6500 and 4300 yrs ago (unit 3), 4300-3500 yrs ago (unit 4) and after ca. 3500 yrs ago (unit 5). The beach ridges formed between 4000 and 3500 yrs ago particularly stand out suggesting that this period was characterized by storms of increased intensity (Fig. 6C).

3.3. Dating beach ridges

The determination of the age of individual ridges in a system can rely either on the setting up of an age model for evolution of the beach-ridge system (e.g. Hede et al., 2015) or by undertaking dating each beach ridge individually (Nott et al., 2009; Clemmensen and Nielsen, 2010; Nott, 2011). The dating of fossil beach ridges has been conducted by applying Optically Stimulated Luminescence (OSL) methods to the sand forming the ridges (e.g. Nielsen et al., 2016; Nott et al., 2009; Clemmensen and Murray, 2010; Reimann et al., 2011; Clemmensen et al., 2012, 2014b; Tamura, 2012; Hede et al., 2015; Rémillard et al., 2015; Clemmensen et al., 2016). Dates were also obtained by radiocarbon (^{14}C) dating of the organic material from soils and/or peat preserved in the swale deposits (e.g. Clemmensen et al., 2001) or of the shell material embedded within the beach ridges (e.g. Nott, 2011). However, several uncertainties limit the use of radiocarbon techniques to date beach-ridges successions. Especially, caution must be taken when dating organic-rich swale material, because these latter may have accumulated an unknown amount of time after the ridges formed, thus potentially leading to an under-estimation of the ridge ages. Conversely, Nott (2011) reported that dating shells embedded in the ridges may lead to and over-estimation of ridge age. Indeed, shells may reside for long periods within the near-shore domain or be eroded for older deposits before being incorporated in the ridge examined. It is thus considered desirable to conduct both radiocarbon dating and OSL dating on the same beach-ridge succession to obtain more accurate age determinations (e.g. Clemmensen and Murray, 2010).

Samples used for dating should preferably be taken in the topmost part of the wave-formed beach ridge, either by coring into the ridge body from the crest or by sampling from trenches excavated into the crest of the ridge (e.g. Nott et al., 2009; Rémillard

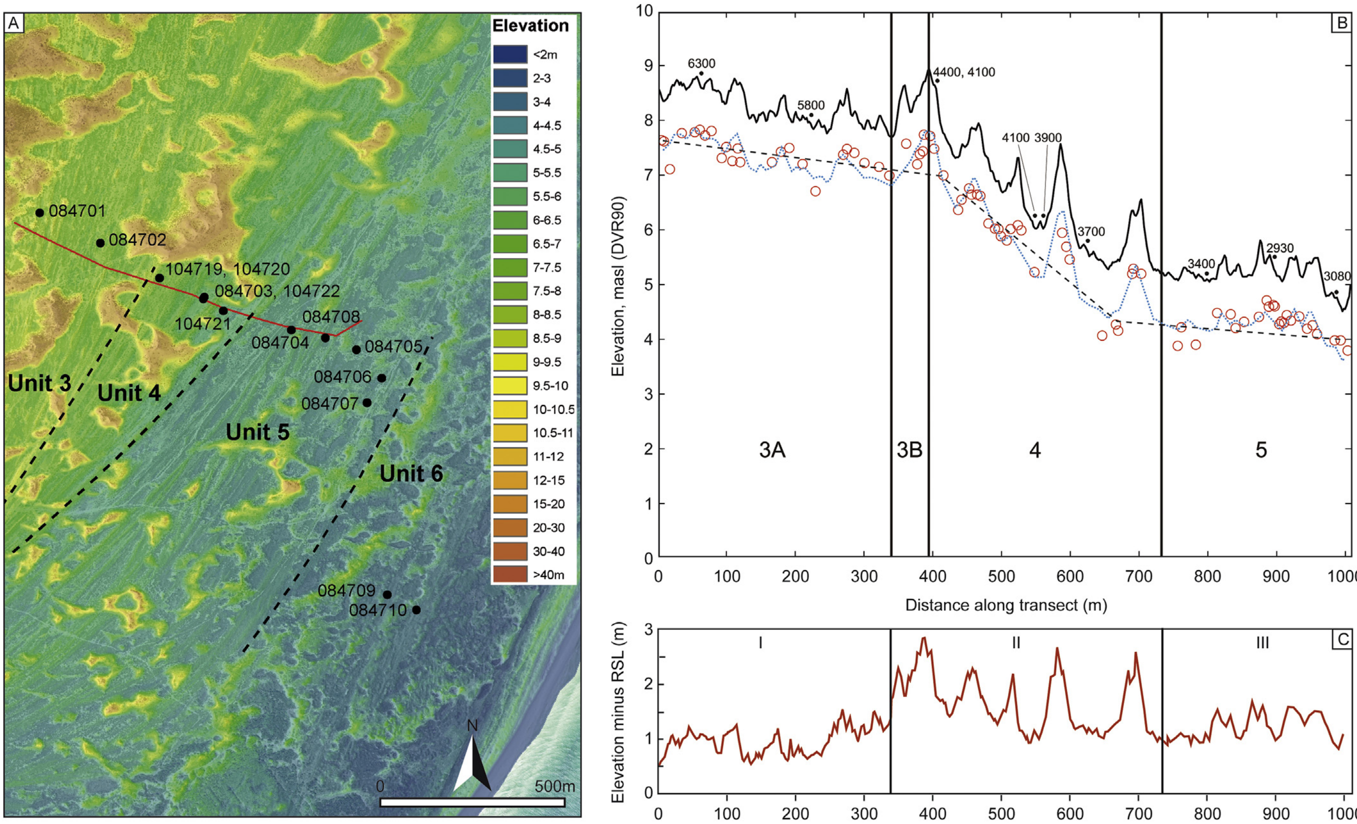


Fig. 6. **(A)** Topographic map (airborne laser scanning) of the southeastern part of Anholt (central Denmark) showing a beach-ridge succession divided into four morphological units (Unit, 3, 4, 5, and 6). OSL sample sites and GPR line (red line) are indicated. **(B)** Topographic profile extracted from the DEM model across the beach ridge plain (Unit 3 A, 3B, 4 and youngest part of unit 5). Red circles show the down-laps points observed in GPR data and used by Clemmensen et al. (2012) to derive RSL data. Dashed blue line correspond to the elevation points of swales displaced 0.9 m downward in order to reconstruct a curve of relative sea level change with time (see Clemmensen et al., 2012 for details). Dashed black line is the simplified RSL curve we used in the present article to detrend beach-ridges elevations from RSL changes. Age control (years ago) is given by OSL dates (small black dots). **(C)** Elevation profile of the beach-ridges detrended from local RSL changes (produced for the present article). Figures (A) and (B) modified from Clemmensen et al. (2012) and reproduced with the permission of Elsevier.

•Each beach ridge is here interpreted to have formed during several episodes of elevated storm water level. The three successions of beach ridges in Anholt (I, II, and III) are thought to reflect changes in dominating wind and wave conditions through time. The high (detrended) elevations of the beach-ridges crests for period II suggest a period of particularly intense storminess activity between ca. 5000 and ca. 3600 years B.P. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

et al., 2015). Yet, when this is not possible due to the presence of a too thick aeolian sand cover topping the ridge, samples from the aeolian sand can be used, assuming that the capping aeolian cover formed contemporaneously with the beach ridge (or with an insignificant time lag after ridge formation) (Nielsen et al., 2016; Reimann et al., 2011). (Clemmensen et al., 2009). Uncertainties in most OSL ages used for dating beach ridge and associated aeolian sediments lie between 6 and 8% and the uncertainties are dominated by experimental uncertainties in the measurement of dose and dose rates (Clemmensen et al., 2009; Hede et al., 2015). Dose rates are calculated using measured radionuclide concentration and in this process the mean water content of the sample during the burial period must be estimated and the mean burial depth of the sample must be known (see Clemmensen et al., 2009 for details). Most of the OSL datings obtained in the works referenced above are based on the use of quartz sand grains, but most recently feldspar grains have increasingly been used to provide supplementary age control (Madsen et al., 2011). Gravelly systems with little sandy matrix may be hard to date by traditional OSL methods, and much work has therefore been directed towards the development of a method that could date the age of the large clasts in these beach ridges (e.g. Sohbati et al., 2012).

3.4. Limitations in the use of beach ridges as indicators of past storminess

Several conditions should be met before reliable records of past storminess and inundation levels can be built using sandy or mixed sand-gravel beach ridges. First, one must ensure that the morphology of the storm ridge is well-preserved, as noted by Nott et al. (2009) and Nott (2011). This will rarely be the case for sandy ridges in coastal climates as only moderate winds are likely to modify considerably the original morphology. Bendixen et al. (2013) mentioned that the morphology of sandy ridges are likely to be modified after emplacement by aeolian action, either resulting in an erosive truncation of the ridge crest or in an accumulation topping this latter. Such modifications could have occurred even after the sandy berms have turned into beach ridges. Therefore, caution much be exercised when the height of sandy beach ridges is used as a source of information on the magnitude of past-storm events. Clemmensen et al. (2016) elaborated on these results by suggesting that, due to their higher preservation potential, gravelly beach ridges were more likely to provide adequate proxy records of past storminess. Gravel ridges are indeed more likely to survive aeolian modification than sandy structures, although ridges composed of mixed sand and gravel can also be deflated

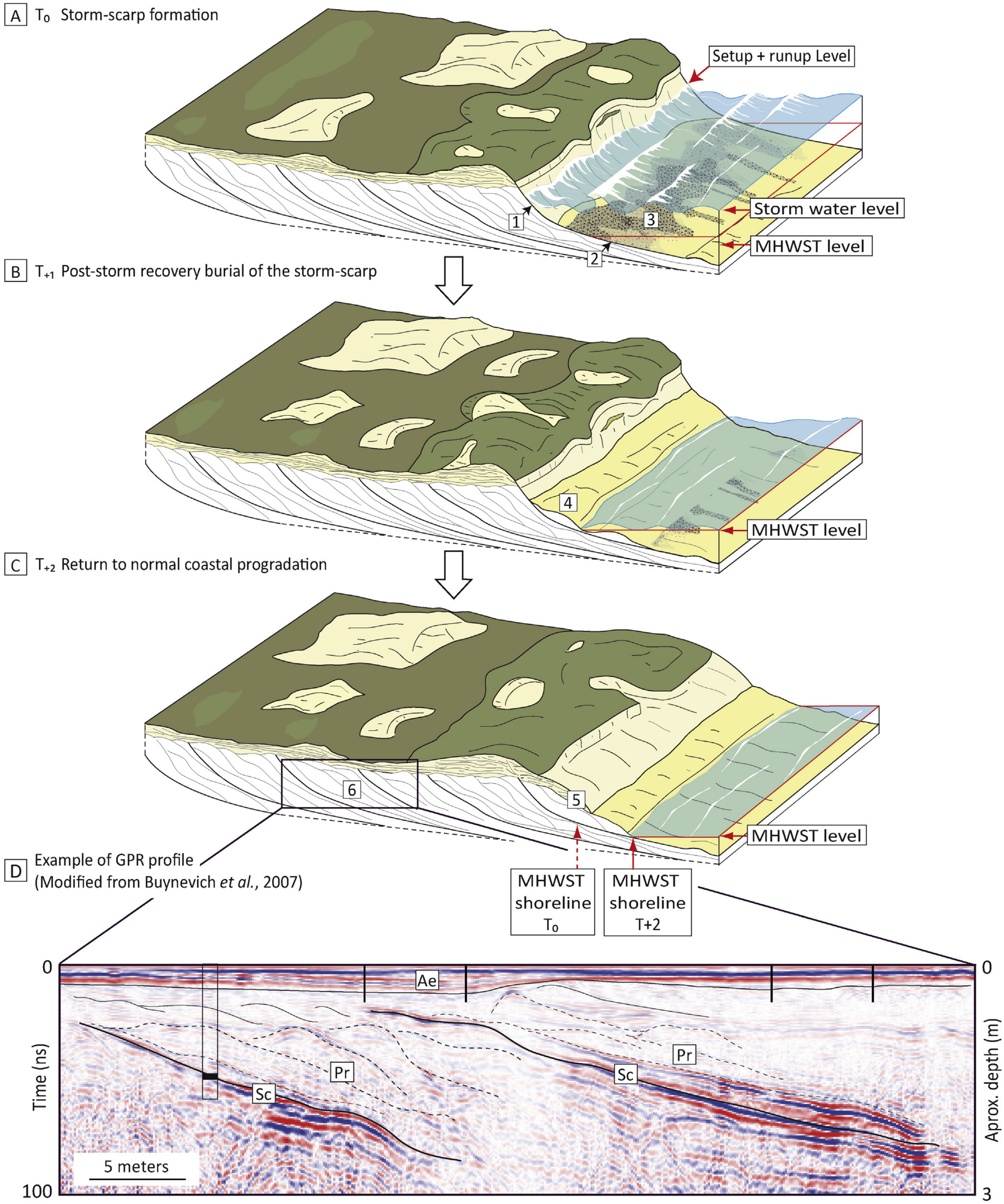
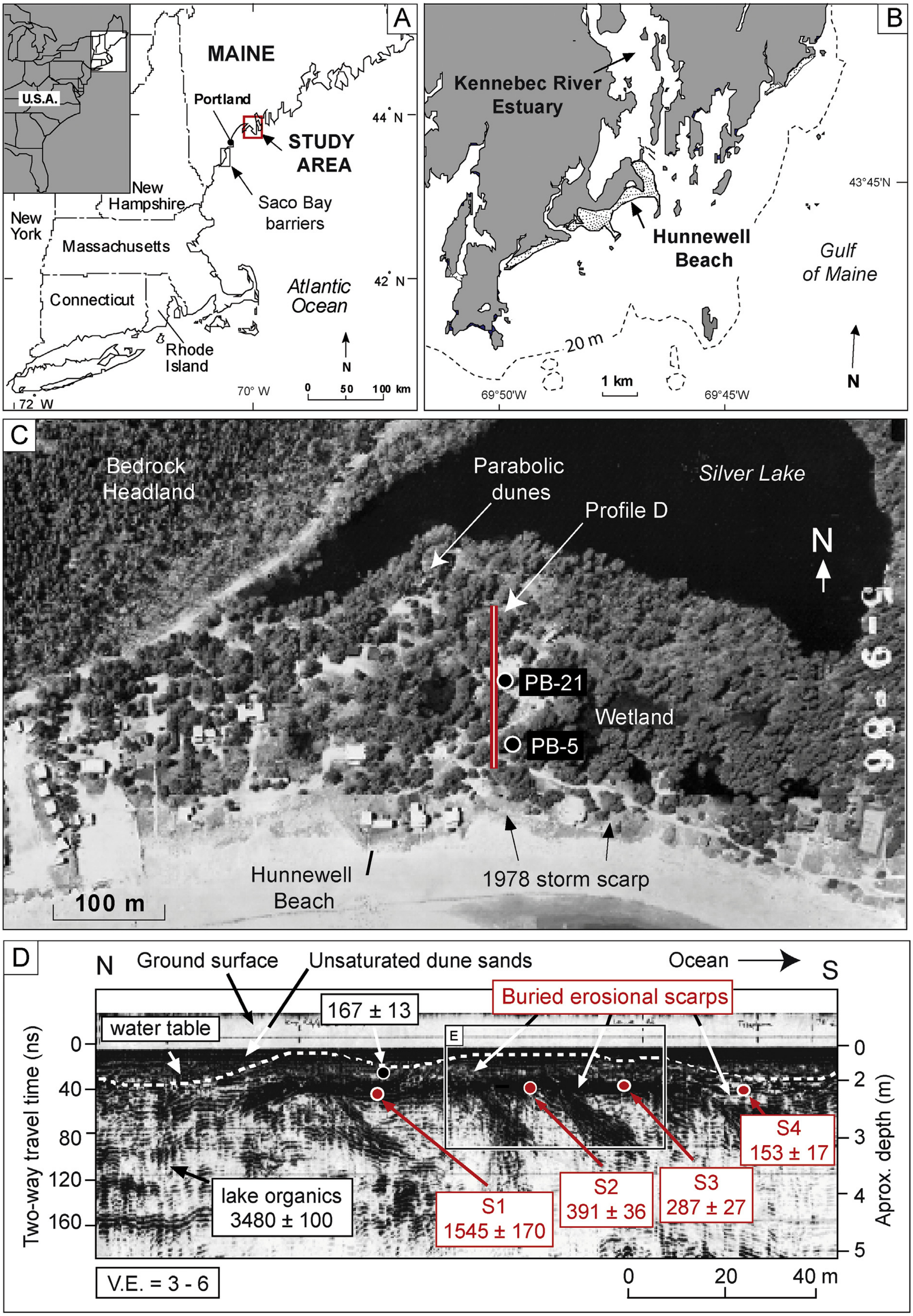


Fig. 7. Schematic illustration of the development of an idealized storm-scarps succession and of the sedimentary features usable for past-storminess reconstructions, along with pictures of the real expression of each state. **(A)** Cutting of the storm scarp (1) and storm beach profile (2) under storm-elevated water level conditions and collision regime. Deposition of Heavy Mineral lags and/or coarse-grained layers (3) on the beachface. **(B)** Return to fair-weather conditions and building of a post-storm recovery infilling (4). **(C)** Stabilization of the post-storm recovery infilling (5) and burial under continued system progradation. (6) Succession of storm scarps buried within the prograding barrier body. MHWST = Mean High Water Spring Tide. **(D)** GPR profile obtained across the Hunnewell barrier (Maine, USA) by Buynevich et al. (2007), exemplifying the main GPR facies of a storm-scarps succession. Storm-scarps are noted “Sc”. “Pr” mark landward-dipping, asymmetric reflections interpreted as accretionary nearshore sandbar units. “Ae” stand for aeolian sand deposits. Figure (D) modified from Buynevich et al. (2007) and reproduced with the permission of the Geological Society of America.


A
U.S.A.
MAINE
Portland
STUDY AREA
44° N
Saco Bay barriers
New Hampshire
New York
Massachusetts
Connecticut
Rhode Island
Atlantic Ocean
42° N
72° W
70° W
0 50 100 km
N
B
Kennebec River Estuary
Hunnewell Beach
Gulf of Maine
43°45'N
20 m
1 km
69°50'W
69°45'W
N
C
Bedrock Headland
Parabolic dunes
Silver Lake
Profile D
N
PB-21
Wetland
PB-5
Hunnewell Beach
1978 storm scarp
100 m
D
N
Ground surface
Unsaturated dune sands
Ocean
S
Buried erosional scarps
water table
167 ± 13
E
Two-way travel time (ns)
Aprox. depth (m)
lake organics
3480 ± 100
S1
1545 ± 170
S2
391 ± 36
S3
287 ± 27
S4
153 ± 17
V.E. = 3 - 6
0 20 40 m

(Clemmensen and Nielsen, 2010). Thus, gravel-rich beach-ridge systems with negligible aeolian sand cover and formed in a system with a high progradation rate should preferably be used in analyses of past storminess. Secondly, it should be determined if a given ridge can be considered as emplaced by a single storm event or if it records several storm events. In the latter case, the ridge elevation could either characterize the highest storm event (Forsyth et al., 2010; Nott et al., 2013) or the last storm event (Bendixen et al., 2013). The more rapidly the coastal system is prograding, the more plausible it is that the final and highest storm events coincide.

Finally, the best possible chronology of ridge formation must be established throughout the examined beach-ridge system if precise information about past storminess is to be derived. Apart from the precision of the dating technique in itself, only a precise dating can reveal the presence of hiatuses or gaps within the chronology of formation of the beach-ridge plains, which, if undetected, may induce considerable biases in the calculations of storm recurrence intervals (Nott and Forsyth, 2012; Nott et al., 2015). The behavior of beach-ridge sets towards post-depositional erosive events can be variable, ranging from a mere scarping of the most seaward beach ridge and a flattening of the beachface to massive removals of several rows of beach ridges at one event (Nott et al., 2015 and references therein). Nott et al. (2015) investigated the evolution of the Cungulla beach-ridge plain (Bowling Green Bay, North-Eastern Australia) and showed that gaps in the storm history could be either the result of (i) erosive processes responsible for the removal of some of the ridges or of (ii) periods when the building of beach ridges really stopped, likely due to a more quiescent period in storm activity. Erosive gaps would be signaled by angular unconformities in the beach-ridge series as well as by changes in the beach-ridge morphologies and orientation. True periods of less intense storm activity would, conversely have induced a slowing down in the development of beach ridges, associated with no evident sign of ridge truncation nor modification of the ridges morphologies. The slowing-down in the building of beach ridges was also suggested by Nott et al. (2015) to have been most likely fostered by a decrease in sediment supply to the coast due to less frequent river-floods during periods of reduced storminess.

In short, long-term records of Holocene storm events can be preserved in prograded beach-ridge systems. Best and most reliable records are given by rapidly prograding gravel-rich systems and dating of these is best carried out using OSL dating methods or combined OSL/radiocarbon dating. The data obtained can provide useful estimates of the number of storm events and hence, in the most ideal cases, of their recurrence intervals. However, in view of the case studies discussed above, it should be kept in mind that the individual beach ridges rarely will be linked to single storm events but more likely be the record of several storms. Furthermore, potential gaps in beach-ridges formation and/or coastal progradation should be accounted for as they may strongly influence the representativeness of the record. Complete and extensive beach-ridge systems may therefore contain evidence of systematic changes in storminess history on climate scale (30 years or longer) rather than on a weather or event scale.

4. The storm scarps

In the following section, we will focus on the uses that can be made of erosive events recorded in prograding coastal barrier stratigraphies for reconstructing high-energy wave events by considering storm-scarps morphologies.

4.1. Definition, formation processes and characteristic features

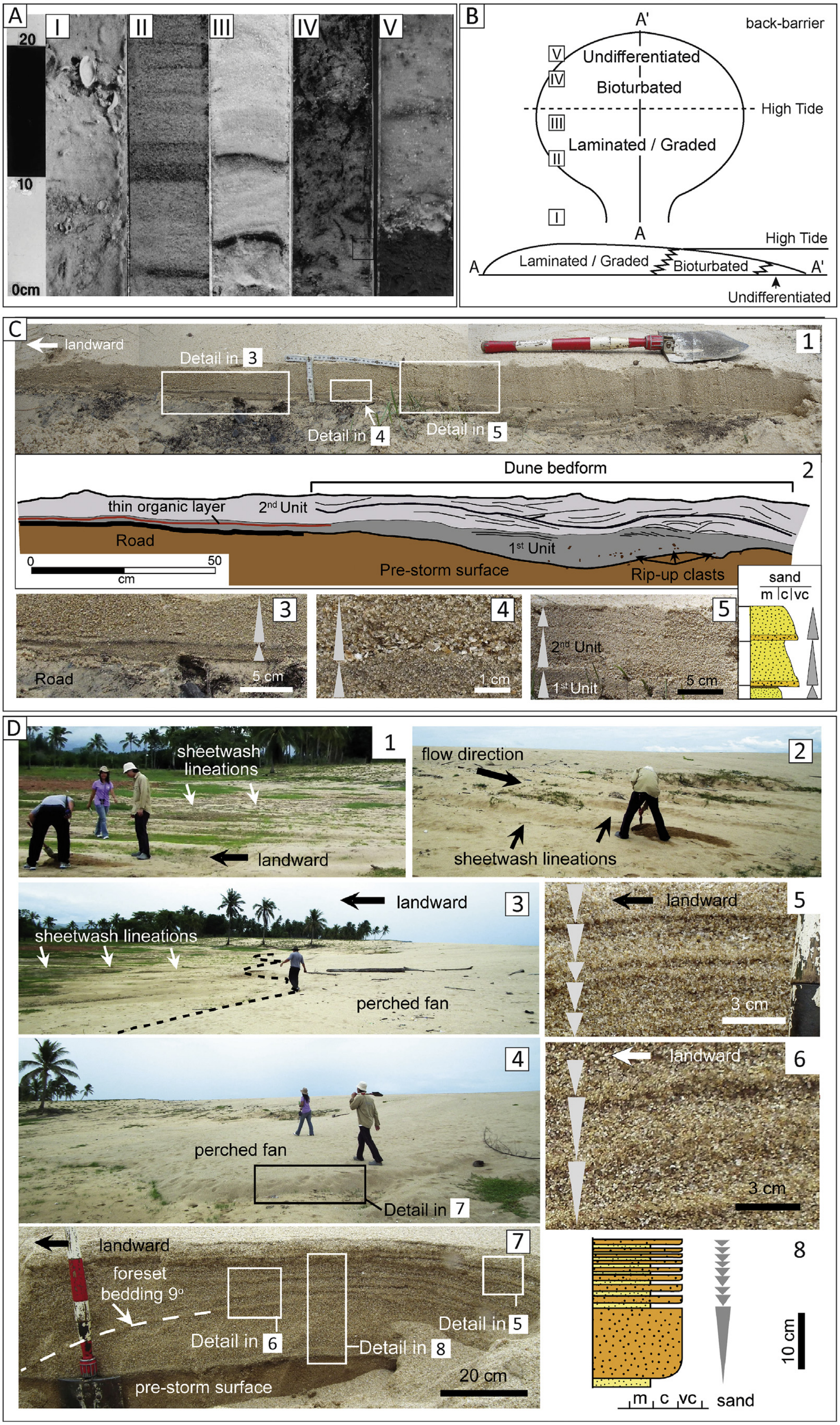
Storm scarps have quite early been recognized as storm-induced features (e.g. Bruun, 1954; Edelman, 1972). They consist of erosion features in beach and barrier deposits characterized by (i) notches excavated in the barrier front by breaking waves and/or runup swash action (the "dune scarps") and most often further developed by mass gravity slumping and by (ii) planar to concaved-up seaward dipping unconformities (Morton, 2002; Morton and Sallenger, 2003; Wang et al., 2006; Suanez et al., 2012) (Fig. 7). These latter unconformities are caused by undertow currents driven by plunging breakers, which may be strengthened by the reflective conditions induced by the escarpment of the upper-shoreface (e.g. Héquette, 2001; Castelle et al., 2007, 2015; Suanez et al., 2012). Storm scarps are generally accompanied by heavy-mineral concentrations (HMCs) topping the erosion surface on the foreshore (Smith and Jackson, 1990; Dougherty et al., 2004; Wang et al., 2006; Dougherty, 2014) and/or by coarse-grained material blankets covering the beach- and shore-faces (e.g. Orford and Carter, 1985; Dougherty et al., 2004; Clemmensen et al., 2016). HMCs layers have been considered to be deposited during the waning stages of storm events, when wave transport energy diminishes and fosters a selective density sorting of the material (Komar, 1998, Meyers et al., 1996; Moore et al., 2004; Wang et al., 2006). Deposition of HMCs have also been reported during fair-weather conditions and later exposed by more energetic events, as a result of the non-magnetic grains being preferably eroded from the upper-swash zone than the heavy minerals (Gallaway et al., 2012). Following the storm, a "post-storm recovery" stage is often observed causing re-deposition of sand on the upper beach area both by constructive wave action bringing back the sediments which were deposited on the foreshore during the storm and by aeolian deposition (Wang et al., 2006). This returning of material to the upper beachface then eventually forms the basis for sand accumulations in "incipient foredunes", growing at the dune foot (e.g. Wang et al., 2006; Suanez et al., 2012), and promoting the preservation of the storm profile (cut-and-fill features) as an archive of past storminess (Figs. 7 and 8).

4.2. Identification of storm scarps within the sedimentary archives and their use as proxies of past storminess

Storm scarps have been shown to be capable of providing information on past storminess frequency and strength (Smith and Jackson, 1990) and, as such, have been successfully used during the last decade for the recognition of Holocene extratropical storms and hurricanes. The difficulty of observing and mapping the geometry of scarp features in barrier sedimentary sequences has long hindered their use for paleo-environmental purposes (Meyers et al., 1996). The advent of GPR made the recognition of these features much easier. Several studies highlighted the potential of GPR to accurately trace storm scarps within the barrier sediments (Bristow et al., 2000; Lindhorst et al., 2008; Angulo et al., 2009; Buynevich et al., 2011; Barboza et al., 2013; Gontz et al., 2014) but

Fig. 8. **(A)** & **(B)** Maps showing the location of the Hunnewell barrier (Maine, USA) studied by Buynevich et al.(2004, 2007). **(C)** Location of the GPR transect and coring sites. **(D)** Interpreted GPR profile showing the set of storm scarps (S1 to S4). OSL ages obtained on the post-storm recovery sands are indicated, as well as a radiocarbon age obtained on freshwater peat underlying barrier sands. Ages in years cal. B.P. All Figures modified from Buynevich et al. (2007) and reproduced with the permission of the Geological Society of America.

• *Each storm scarp is interpreted to be the result of intense erosion of the dune barrier by elevated water during an extreme storm event (collision regime). The chronology of the scarps reveals storms separated by ca. 100 years during the last 500 years and a 1000 year gap in between 500 B.P. and 1500 B.P. This gap can reflect either a more quiescent period in storm activity, or erosive retreat of the coastline, that led to the disappearance of the storm scarps which may have formed within the 1500 to 500 B.P. period.*



A
I
II
III
IV
V
20
10
0cm
B
A'
back-barrier
Undifferentiated
Bioturbated
High Tide
Laminated / Graded
A
High Tide
A
Laminated / Graded
Bioturbated
A'
Undifferentiated
C
landward
Detail in 3
Detail in 4
Detail in 5
1
Dune bedform
2
thin organic layer
2nd Unit
Road
1st Unit
0
50
cm
Pre-storm surface
Rip-up clasts
sand
m c vc
3
Road
5 cm
4
1 cm
5
2nd Unit
1st Unit
5 cm
D
sheetwash lineations
1
landward
flow direction
2
sheetwash lineations
landward
3
sheetwash lineations
perched fan
landward
5
3 cm
landward
6
3 cm
4
perched fan
Detail in 7
landward
7
foreset bedding 9°
Detail in 6
Detail in 8
Detail in 5
pre-storm surface
20 cm
8
10 cm
m c vc
sand

did not specifically contemplate the use of these features to build records of past storminess. Systematic studies of storm scarps with this objective in mind were initially conducted on the coastal barriers of the eastern coasts of the United States (Buynevich et al., 2004, 2007; Dougherty et al., 2004; Dougherty, 2014) and are now being carried out elsewhere.

The recognitions and mapping of storm scarps using GPR traditionally lean on the fact that the post-storm heavy mineral layers frequently covering the beachface have high magnetic susceptibility. The HMCs density lags show intense reflectors which intensively contrast with the signal of surrounding sands (Moore et al., 2004; Van Dam et al., 2002, 2013). As a result, storm scarps can become relatively easy to map (Fig. 7). Contrasting GPR reflectors at the storm-induced surface can also be the result of some variation in the grain sizes (caused e.g. by post-storm coarse-sand and pebbles covering the beach-face, e.g. Dougherty et al., 2004, Moore et al., 2004) or of porosity changes resulting from the compaction of the beachface by storm breakers (Moore et al., 2004). Sediments associated with storm scarps are characterized by (Figs. 7 and 8) (i) steep seaward-dipping strata/reflectors marking the dune scarp and (ii) sweeping seaward-dipping strata/reflectors truncating horizontal to seaward low-dipping reflectors representing the upper- and lower-beach sediments, respectively (Bristow et al., 2000; Neal et al., 2002; Bristow and Pucillo, 2006; Gontz et al., 2014). Landward-dipping sediments can sometimes be observed in between these two latter sediment types; interpreted as aeolian “foredune” deposits capping the upper beach at the time of beach recovery (Bakker et al., 2012; Choi et al., 2014; Gontz et al., 2014). In ideal situations, modern or historical scarps can be used as modern analogs. Buynevich et al. (2004, 2007) for example found the four well-individualized couples of scarps and seaward steeply-dipping unconformities covered with HMCs lags they identified in the internal topography of the Hunnewell Barrier (Gulf of Maine, north-eastern USA) to show very close similarities with a scarp left on the site in 1978 by an easterly storm (Fig. 8).

4.3. Dating storm scarps

Precise chronologies of storm scarps formation have developed following the spread of the OSL method. Dating of the events responsible for storm scarp formation can be undertaken by dating the bottom surface of the post-storm recovery sand infill, following the assumption that this recovery normally occurs quite fast after the storm and thus approximate quite precisely the age of the event(s) responsible for the cutting. This approach was used for the first time by Buynevich et al. (2007) for dating four paleo storm scarps preserved within the stratigraphy of the Hunnewell barrier (Maine, USA). Considering the effect heavy minerals may have on the determination of the dose rates, the authors instead used the quartzose sand directly overlying the heavy-mineral lags. Dating the deposits directly underlying the erosive surface may also be used to provide an adequate maximum chronological control for the event responsible for the scarp formation. Nonetheless, using this latter approach may greatly overestimate the age of the storms if, for example, storm waves would have provoked a massive shoreline retreat and cut into much older sediments (Buynevich et al., 2007). Caution should be exercised when using coarse quartz samples for OSL dating. For example, sand originating from glacial inherited features can have poor luminescence characteristics (e.g. Fuchs et al., 2012) or incomplete bleaching of the sediments prior to deposition can occur (notably due to the erosion and rapid re-deposition of older deposits from the barrier or foreshore domains). These limitations potentially complicate the obtaining of accurate OSL dates on the post-storm sedimentary infills. Finally, organic or shell material present within the post-storm sedimentary infills may allow radiocarbon dating. However, there is a reasonable probability that this material could have been reworked from older deposits.

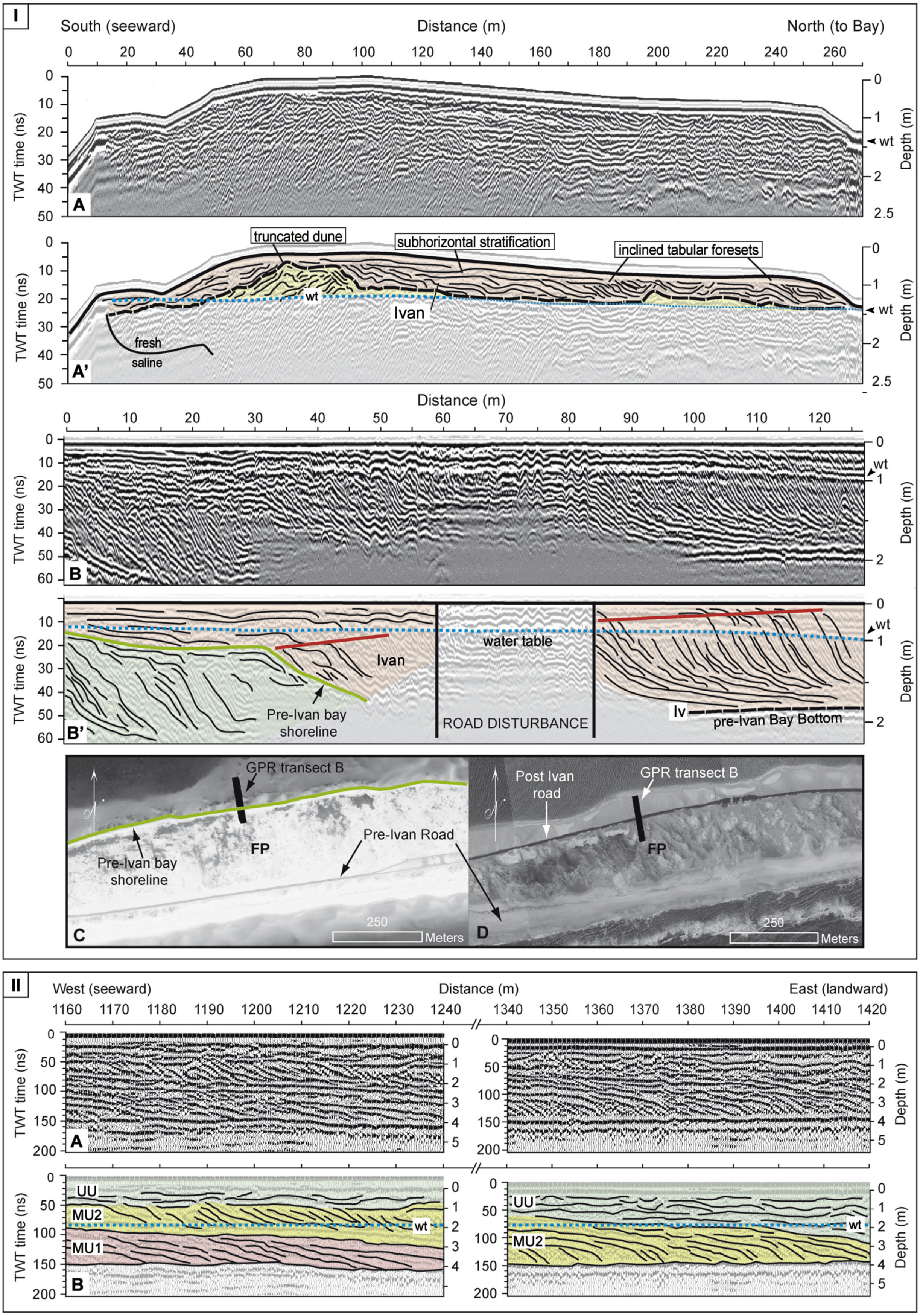
4.4. Limitations in the use of storm scarps as indicators of past storminess

Dune scarps and associated storm-shaped beachface markers are produced by several complex and interdependent processes relating to various beach and shoreface morphodynamics that are still uncompletely understood (De Alegria-Arzaburu et al., 2013). Therefore, building reliable and representative past storminess records on the sole basis of storm scarps still remains challenging.

4.4.1. The importance of local morpho-dynamic factors in the occurrence of storm scarps

The formation of storm scarps is closely related to (i) the dune and beach characteristics (height, curvature, slope) and (ii) the storm wave climate (waves period and height). Alongshore variations of these latter induce variable probabilities that storm scarps are produced and subsequently preserved. Weymer et al. (2015) for instance reported highly varying post-storm beach and dune morphologies in relation to alongshore variations in the height of the dunes, resulting in more or less sensitivity to a presumed spatially-constant storm water level. Storm scarps were reported absent from the lower part of the systems, where dunes were instead overtopped by waves. By monitoring strom-related intertidal profile changes on a beach of the Netherlands, Aagaard et al. (2005) also documented alongshore spatial variability, characterized by short-scale variations between reflective and dissipative sectors. Large differences in the morphological responses were observed between the embayments, showing intensive erosion of the beachface and offshore sediment exportation at high tide, and the salients where storm impacts were much reduced. These different behaviours are (at least partly) conditioned by pre-storm beach and shoreface morphologies and may be further fostered during the storm through morpho-dynamic feedbacks (e.g. Regnauld and Louboutin, 2002). A negative storm feedback would make a beach-dune system more susceptible to undergo storm impacts as the result of the effects previous events had on the morphology. A steepened upper-beach profile can for instance facilitate erosion by (i) increasing the water-depth within the surf-zone, thus allowing an onshore shift of the wave plunge-line and (ii) by strengthening backwash currents, thus making erosion worse (Aagaard et al., 2005; Erikson et al., 2007; Suanez et al., 2012;

Fig. 9. **(A)** Pictures of cores through washover deposits showing their sedimentary characteristics (Sedgwick and Davis, 2003): (I) stratified sand with high macrofauna shell contents, (II) discrete layers (dark-colored) rich in dark-heavy minerals, (III) washover sand separated by dark-colored algal mats (barrier and back-barrier small ponds), (IV) bioturbated washover sand (back-barrier marsh facies), (V) washover sand unconformibly overlying back-barrier marsh deposits. **(B)** Schematic model showing the idealized repartition of facies across a wash-over deposit (Sedgwick and Davis, 2003). The locationof the facies depicted on (A) is indicated.**(C)** 1: Cross-section through a washover deposit (Phantuwongraj et al., 2013) showing the internal structure of a washover deposit. (C) 2: Line-drawing interpretation of (1). 3, 4, and 5: Close-up pictures showing two stacked normally-graded washover deposits. Termination of the first unit is marked by a heavy mineral layer, while a shell lag forms the base of the second layer (fining upward). **(D)** 1 and 2: Pictures of washover sheetwash lineations deposited under an inundation regime (Phantuwongraj et al., 2013). 3 and 4: Pictures of perched washover fans deposited under an overwash regime (Phantuwongraj et al., 2013). 5 and 6: close up of reversely-graded stacked washover layers within a perched fan (stratigraphic position shown on D-7). 7: internal structure of the landward termination of a perched washover fan. Note the landward dipping foreset bedding. 8: line drawing illustrating the reverse-graded nature of the stacked washover deposits shown in 5. All figures reproduced with the permission of Elsevier.


I
South (seaward)
Distance (m)
North (to Bay)
TWT time (ns)
Depth (m)
wt
A
truncated dune
subhorizontal stratification
inclined tabular foresets
Ivan
fresh
saline
A'
B
water table
Pre-Ivan bay shoreline
ROAD DISTURBANCE
Iv
pre-Ivan Bay Bottom
B'
GPR transect B
Pre-Ivan bay shoreline
FP
Pre-Ivan Road
250
Meters
C
Post Ivan road
D
II
West (seaward)
East (landward)
UU
MU2
MU1

Blaise et al., 2015). Nonetheless, as suggested by Hesp (2002) and shown by e.g. Suanez et al. (2012), post-storm incipient foredunes are likely to develop at the dune foot, following particularly large erosive events. Post-storm recovery beach profiles often exhibit dissipative states, which have been shown as more favorable to aeolian transport and to foster an elevation of the dune toe, thus possibly reducing the impacts of subsequent storms.

4.4.2. Presence of heavy minerals

As stated above, recognizing scarps within Holocene coastal successions by GPR investigations requires the presence of HMCs lags and/or of layers of coarse material covering the beach-face after the occurrence of energetic events. As for heavy-minerals, their amount supplied to the system can be highly variable through time. Dougherty et al. (2004) for instance investigated the sedimentary sequence of the Castle Neck barrier (Massachusetts, USA) and found a well-preserved set of uniformly spaced piled-up storm scarps clearly popping in the GPR profiles thanks to the high reflectivity of the HM blankets. The frequency of the scarps was observed to diminish seawards, thus *a priori* hinting for a decrease in storm activity towards younger ages. Yet, more than a true decrease in storminess, this was seen by the authors to be more probably a consequence of the interplays of several changes in the barrier dynamics and a progressive change in the sediment sources: as the barrier prograded seaward, the main heavy-minerals suppliers (here the inherited glacial drumlins) were progressively shielded from coastal erosion, until they became completely disconnected from the shoreline. The sediment source then became dominated by river sediments poor in heavy minerals, making scarps more difficult, not to say impossible, to detect and individualize.

4.4.3. Post-storm recovery

To be of some use for paleo-environmental purposes, storm scarps must be preserved within the barrier deposits. On short time scales, the preservation of storm scarps is thus dependent on the resilience of the system *i.e.* its ability (i) to return to its pre-storm state after each high-energy event (Masselink and van Heteren, 2014) and (ii) to be concerned by large enough post-storm sedimentary infills to prevent a subsequent erosion of the scarps by next storm events. The timing of post-storm recovery processes is highly variable. Normally quite fast, it can, in some cases, take several years (Morton et al., 2007; Suanez et al., 2012) or even up to a decade (Houser et al., 2015), leaving the barrier vulnerable to several storm events before the recovery is completed. Timing and volume of the recovery mostly hinge on the sediment supply to the upper-beach, which is mainly nourished by material originating from a landward migration of the nearshore bars (Maspataud et al., 2009; Houser et al., 2015) and aeolian redistribution of this material (Aagaard et al., 2004, 2007). Yet, storms of exceptional energy convey sediments down to depths below the mean storm wave base (depth of closure). These sediments are non-available for post-storm recovery for a considerable amount of time, making subsequent storms more likely to attack the system. Finally, Weymer et al. (2015) and Houser et al. (2015) recently showed that post-storm recovery rates and volumes kept pace with the height and volume of the dune within a single system. These latter authors proved that the regions where dunes were the largest showed the least and the longest post-storm recovery capability and suggested that storm scarps would be potentially less completely preserved within these systems.

4.4.4. Long-term preservation potential

On longer time scales, the quality of the recording and preservation of storm scarps within a barrier heavily depends on the erosion/progradation ratio of the considered barrier, and as such on the sediment budget. In other words, for storm scarps to be well-preserved within a barrier it is critical that the rates of progradation have kept pace with the landward erosion rates caused by extreme events. The results obtained by Buynevich et al. (2004) when investigating the Hunnewell barrier (Gulf of Maine, north-eastern USA) illustrate this limitation. This study revealed the presence of four well-defined couples of scarps and seaward steeply-dipping unconformities covered with HMCs lags. OSL dating of the scarps showed that the easternmost (seaward) features could be linked to storm events respectively dated to ca. 160, 290 and 390 yrs B.P., while the westernmost (landward) scarp was dated to 1500 yrs B.P. (Fig. 8; Buynevich et al., 2007). The ca. 1000 yr gap was suggested to either reflect a drop in storm activity during the Medieval Warm Period or, more probably to be the result of subsequent erosion. Therefore the record was considered to only represent a minimum estimation of the real number of events (Buynevich et al., 2007) with the youngest 390-yrs B.P. erosive event being probably responsible for the removal of almost 1000 years of the late Holocene storm history recorded at this site (Tamura, 2012). At Flinders Beach (Australia), changes in the ebb-tidal dynamics of the near delta were suggested to have caused modification of the sediment supply, thus influencing the preservation of storm scarps (Gontz et al., 2014). Bristow and Pucillo (2006) exemplified another major aspect of coastal progradation susceptible to play an important role in the ability of storm scarps to keep pace with the storm history. These authors identified six generations of beach ridges at Guichen Bay (south-east Australia). Each set of beach ridges were shown to contain equivalent volumes of sediment and, as such, were considered to have formed under constant sediment supply. Yet, chronological control demonstrated of dramatic decrease of the net shoreline progradation rates with time, due to the barrier building over increasing depths of the sea floor and i.e. increasing accommodation space. Considering this, the barrier must have undoubtedly been more and more exposed and impacted by erosive events which are likely to have partly erased some testimonies of barrier progradation. Such dynamic

Fig. 10. **I)** GPR profiles of Ivan hurricane deposits on Santa Rosa Island (Florida, USA) by Wang and Horwitz (2007). **(A)** GPR profile 1 of Ivan washover blanket at Beasley Park study site (see Wang and Horwitz, 2007 for detail on the location of the study sites and transects) and **(A′)** its line-drawing interpretation showing: (1) the water table (wt; dashed blue line); (2) the base of the Ivan washover deposit (heavy black dashed line); (3) a truncated dune (yellow colored area); (4) washover sediments deposited over the barrier interior platform (light red colored area); and (5) tabular foreset bedding showing the progradation of washover deposits over a low-lying subaqueous area (e.g. marsh, pond). **(B)** GPR profile 2 of Ivan washover depositsat Fort Pickens study site and **(B′)** its line-drawing interpretation showing: (1) the water table (wt; dashed blue line); (2) the base of the Ivan washover deposit ("Iv", heavy black dashed line) and (3) Ivan washover deposits (light red colored area) characterized by sigmoidal, tangential and downlapping reflections. Similar reflective patterns at the bottom of the profile (green colored area) are seen to represent an earlier washover fan. The downwarping of the "Iv" horizon at 30 m correlates with the position of the pre-Ivan back-bay shoreline (green line) shown on (C) (Wang and Horwitz, 2007). The red line stands for the "Topset-foreset Break in Slope" (see text for explanations). **(C) & (D)** Aerial photographs of the study site taken pre-and post-Ivan hurricane, respectively. **(II) (A)** GPR profiles from the Holmsland barrier (western Denmark) by Møller and Anthony (2003) and **(B)** their line drawing interpretations. The seaward section of the transect shows two sets of prominent, steeply dipping reflections interpreted as two stacked sets of washover deposits (MU1, MU2) covered with low-dipping recent washover deposits (UU), while the landward portion of the transect show prominent steeply dipping reflectors, interpreted as delta foresets deposited in the distal part of the washoverfan MU2, and covered with recent washover deposits (UU). Stippled blue line indicates the water table. Figures from **(I)**and **(II)** modified from Wang and Horwitz (2007) and Møller and Anthony (2003), respectively, and reproduced with the permission of John Wiley & sons and the Geological Society of London. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

competes with the capability of a barrier to keep a constant record of cut-and-fill features as it progrades, and thus may greatly bias the building of storm-frequency reconstructions on such proxies.

In short, the use of storm scarps as proxies for paleotempestology can be considered to be at an incipient stage. Storm scarps can be regarded as promising and complementary storm archives. Indeed, they enable *a priori* storm events to be recorded in barrier deposits, where and when barrier systems are too high and/or too wide to have undergone overwashing and, thus, to have recorded past-storm events in the form of washover deposits (see following dedicated section 5 "*Washover deposits and features*"). Studies of storm scarps thus potentially contribute to the representativeness of the storm record, for example by incorporating storm events of relatively moderate strength (which did not lead to barrier overwash) in the storm archive. Yet, it should be kept in mind that the preservation potential of storm scarps can vary through time for a number of reasons, and storm recurrence intervals derived from these features are probably longer than the "true" recurrence intervals.

5. Washover deposits and features

Storms may leave different types of imprints on the barrier and back-barrier domains that can be used to document past storminess. This section describes the characteristics of the diverse storm-related features that can be found in these domains and how these latter can be used to reconstruct past storminess. The "back-barrier" area is defined as the domain located between the most seaward and active part of the barrier complex (beach and frontal dunes) and the mainland (Fig. 1). The back-barrier area is mostly characterized by low-energy conditions and associated sedimentary environments, such as lagoonal sandflats and salt-marshes as well as coastal lakes or mires. In progradational systems the strandplain can be considered to be part of the back-barrier system (Fig. 1). Storm imprints on the barrier and back-barrier domains can include stratigraphic markers, either resulting from the landward deposition of allochtonous sediments originating from offshore areas or from erosive events (caused by runup surges or inlet opening) and modifications of the ecology of the back-barrier domain, related to ephemeral or longer-lasting saltwater intrusions.

5.1. Definition, formation processes and characteristic features

The sediments transported by the overwashing fluxes are deposited landward in the barrier and back-barrier areas, following the decrease of the flow velocity with distance from the coastline induced by the friction against the bottom and vegetation as well as the percolation through the barrier sediments. Exhaustive reviews of the hydraulic processes involved in overwashing can be found in Donnelly et al. (2006), Carruthers et al. (2013) and Masselink and van Heteren (2014). The consequences of overwash events mainly depend (i) on the height of the wave runup relative to the height of the barrier crest, (ii) on the magnitude of the surge (in terms of duration and energy) and (iii) on the spatial scale over which it extends (Morton and Sallenger, 2003; Phantuwongraj et al., 2013). Overwash processes are traditionally reported to fall into three main categories which can be defined as follow:

- Runup overwashing refers to the state when the storm water level (setup + runup) only slightly and locally exceeds the height of the overall barrier crest. It preferably occurs where the beach topography promotes confined hydraulic flows and where the profile of the frontal barrier is uneven and locally exhibits lows or gaps (Donnelly et al., 2006; Phantuwongraj et al., 2013). Runup washovers deposits normally consists of small, lobate and/or elongated splays of limited extent (both laterally and across the barrier) that are sometimes referred to as perched fans (Morton, 2002). Locally, closely situated perched fans can nonetheless merge into larger features coined as washover sheets (Otvos and Carter, 2008) or washovers terraces (Morton, 2002).
- Inundation overwashing takes place when the storm water level is much higher than the barrier crest. It causes the barrier to be submerged, either rather locally, taking advantage of some topographic lows, or completely (Donnelly et al., 2006). This later scenario takes place if the alongshore topography of the barrier crest is constantly low and/or during persistent or exceptionally high-energy storm events. Inundation promotes unconfined flows which generally lead to an extensive deposition of sedimentary blankets across the barrier and back-barrier areas. Termed sheetwashes, these deposits can be either confined to the vicinity of barrier throats or cover extensive areas of a barrier (Donnelly et al., 2006).
- Breaching can occur as an ultimate state of barrier inundation or following channel incision through the barrier due to the intensive funneling of the overwash flows into specific throats (Morton and Sallenger, 2003). It fosters the most massive and widespread impacts to the barrier and back-barrier areas, consisting of (i) large morphological changes, (ii) massive inputs of allochtonous sediments into the back-barrier areas, (iii) changes of the hydrodynamic regime of the back-barrier domain, resulting in potential reorganizations of the marshes and channels and, finally (iv) potential disruptions of the ecological state of the back-barrier, as a consequence of more or less temporary changes in the salinity regime (e.g. Sabatier et al., 2008).

Runup washover perched fans, terraces and inundation sheetwashes share common attributes. All form under high-energy water flow conditions (Sedgwick and Davis, 2003) and are characterized by material originating from offshore, shoreface, beach and ridge areas accumulating over the pre-storm barrier and back-barrier deposits. A wide range of sedimentary structures and types of stratigraphical architectures have been observed to result from overwash deposition (Leatherman and Williams, 1977; Sedgwick and Davis, 2003), this diversity being promoted by the complex and multiple parameters which influence the overwash flow. This wide array of sedimentary features, together with the similarities washover deposits share with other features (such as tidal deltas and tsunamis blankets) sometimes makes overwash deposits problematic to identify within sedimentary sequences. Several attempts were nonetheless made to sub-specify washover facies by their stratigraphical and sedimentary characteristics (e.g. Leatherman and Williams, 1977; Schwartz, 1982; Sedgwick and Davis, 2003; Wang and Horwitz, 2007; Phantuwongraj et al., 2013).

Washover deposits normally consist of one or multiple layers composed of foreshore and beachface material, often accompanied by marine shells (Sedgwick and Davis, 2003; Cunningham et al., 2011), heavy-minerals (Sedgwick and Davis, 2003; Switzer and Jones, 2008; Phantuwongraj et al., 2013) and macro-vertebrates (Morton, 2002) (Fig. 9). As the material constituting overwash deposits may be the product of the scouring and erosion of the proximal parts of the barrier by the high-energy water flux (including self-cannibalism of the proximal parts of the washover itself), the washover deposits may also contain both dune and terrigenous shell debris. In the case of extensive washovers, lateral changes in the sedimentary characteristics of the deposits are commonly observed between the proximal and the distal parts of the deposition lobes. The proximal parts of washover fans are often

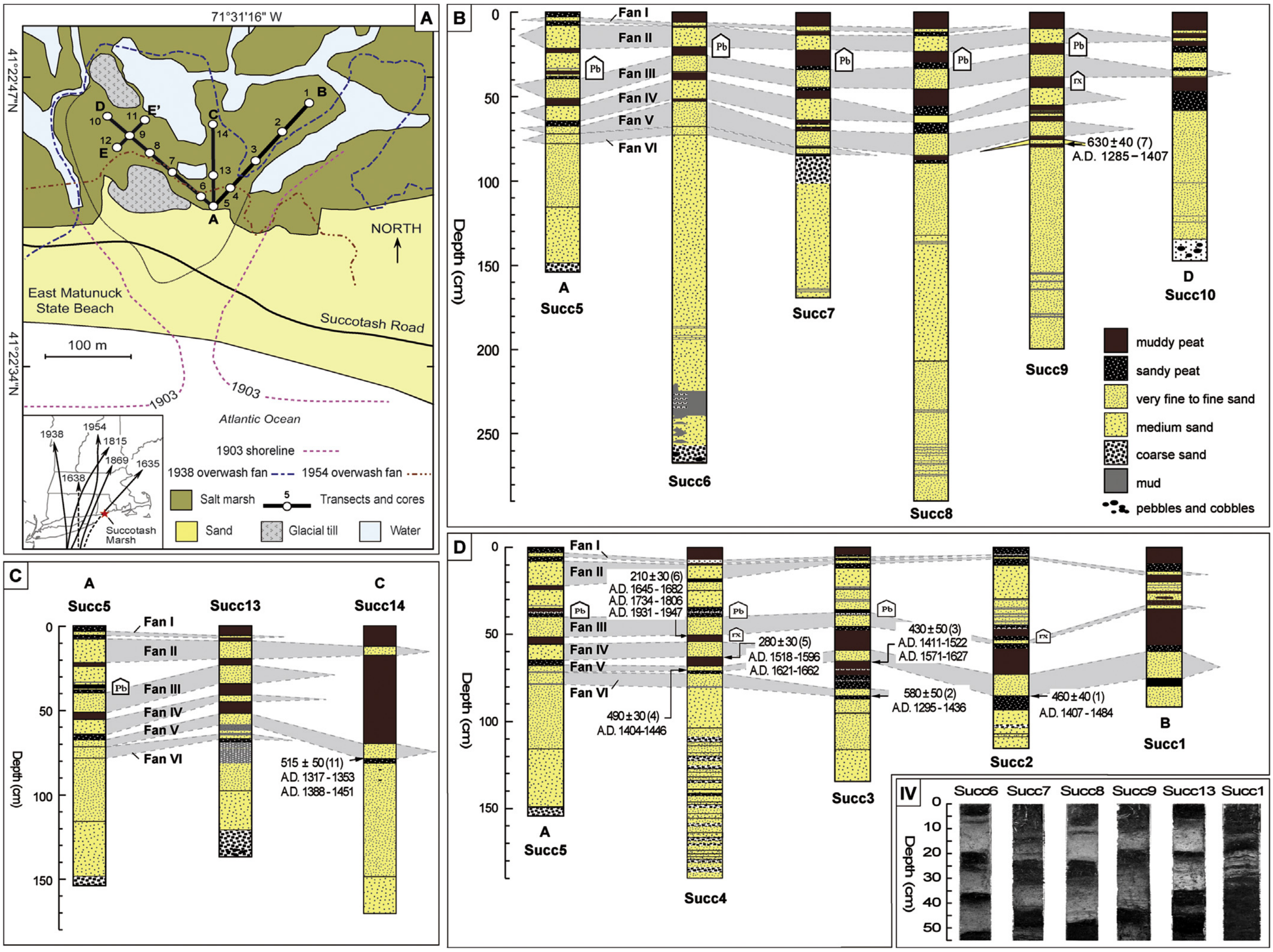
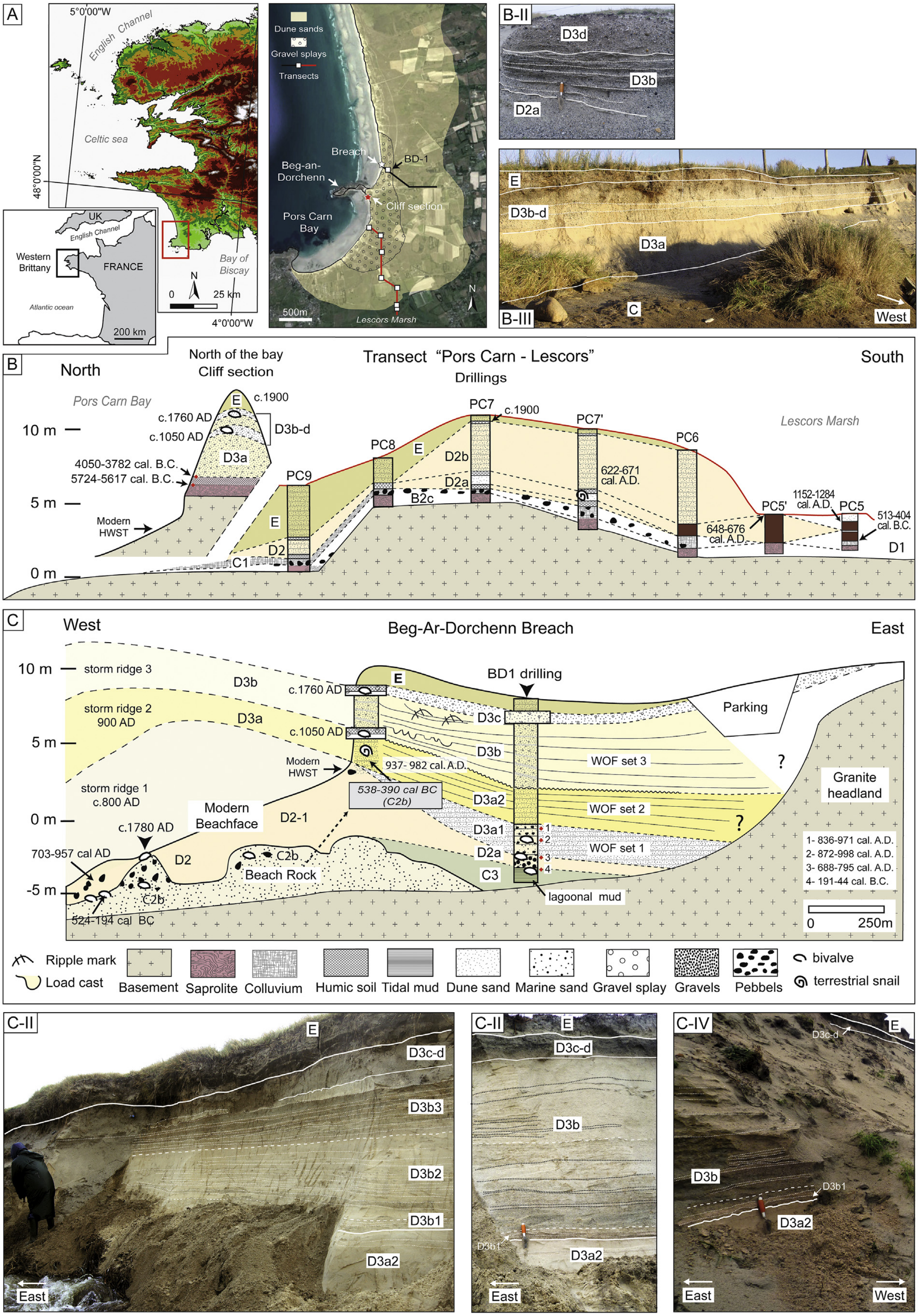


Fig. 11. **(A)** Location map of the Succotash barrier system (southern New England, USA) studied by Donnelly et al. (2001a). Open circles and numbers indicate core sites in the marsh area. **(B), (C) and (D)**: Stratigraphic logs of the cores taken in the Succotash marsh. Note the alternations between peat layers (brown to black) and washover sand layers (yellow). Overwash fans correlated between cores are shown in grey and labeled fans Ito VI. Radiocarbon ages are shown in cal. B.P. ages. (IV) Photographs of the upper 55 cm of selected cores. Dark units are salt-marsh peat. Light units are very fine-to medium-grained sand. All figures modified from Donnelly et al. (2001a) and reproduced with the permission of the Geological Society of America.

• *Each sand layer represents a washover fan that was deposited upon the back-barrier marsh under hurricane-induced elevated water level and massive inundation of the back-barrier area, thus suggesting six hurricanes to have made landfall in this region during the last 600 years. Washover fans are observed to disappear landwards, showing the landwards loss of transport velocity. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)*

characterized by clear laminations of sand, shell lags and HM concentrations induced by the rhythmic pattern of short-period surges (and were thus coined as "stratified sands" by Sedgwick and Davis, 2003, Fig. 9). Progressing landward, washover deposits can be characterized by either positive or negative gradients in grain-sizes. The causes for a preferred deposition in either one of the other sorting mode remain poorly understood (Sedgwick and Davis, 2003; Wang and Horwitz, 2007; Phantuwongraj et al., 2013). If it fines upward, the washover is termed "normal-graded" and a shell or gravel lag then often forms the lower boundary of the deposit (Sedgwick and Davis, 2003; Phantuwongraj et al., 2013; Fruergaard et al., 2015). Washover deposits coarsening upward are referred to as either "Reverse-graded" (Sedgwick and Davis, 2003) and normally show a concentration of fine sediments at their base, often composed of heavy-mineral rich material. In any case, algal mats or vegetation may develop on top of the washover deposit, favoured by the presence of ephemeral ponds (Wang and Horwitz, 2007). If deposited upon back-barrier areas, the distal margins of washovers are most often characterized by mixed sand and mud layers, unconformably overlying the low-energy, organic-rich back-barrier sediments. Intense bioturbation is commonly observed and can annihilate the grading within the unit, forming the bioturbated muddy sand unit, Sedgwick and Davis, 2003, (Fig. 9).

These changes in the sedimentary characteristics of washovers between their proximal and distal parts are generally accompanied by a landward sorting of the material. The mean grain-size of the siliciclastic washover material is normally seen to progressively diminish landward in response to the decrease in the inundation flow velocity (Woodruff et al., 2008; Brandon et al., 2014). The deposition of particles traveling landward into a lagoon in an overwash flow has been described by advective-settling models of sediment transport (Moore et al., 2007; Woodruff et al., 2008). The distance to which the particles are transported landward behind the barrier and the distribution of the particle sizes along a barrier-to-lagoon transect were shown to depend (i) on the height of the overwash surge and (ii) on the average flow velocity. However, an overall landward thinning of washover deposits is not


A
5°0'00"W
English Channel
Celtic sea
48°0'00"N
UK
English Channel
Western Brittany
FRANCE
Atlantic ocean
200 km
Bay of Biscay
N
0 25 km
4°0'00"W
Dune sands
Gravel splays
Transects
Breach
BD-1
Beg-an-Dorchenn
Cliff section
Pors Carn Bay
500m
N
Lescors Marsh
B-II
D3d
D3b
D2a
B-III
E
D3b-d
D3a
C
West
B
North
North of the bay Cliff section
Transect "Pors Carn - Lescors" Drillings
South
Pors Carn Bay
Lescors Marsh
c.1900
c.1760 AD
c.1050 AD
D3b-d
D3a
4050-3782 cal. B.C.
5724-5617 cal. B.C.
Modern HWST
PC9
PC8
PC7
PC7'
PC6
PC5'
PC5
D2b
D2a
B2c
622-671 cal. A.D.
648-676 cal. A.D.
1152-1284 cal. A.D.
513-404 cal. B.C.
D1
D2
C1
10 m
5 m
0 m
C
West
Beg-Ar-Dorchenn Breach
East
BD1 drilling
storm ridge 3
storm ridge 2 900 AD
storm ridge 1 c.800 AD
D3b
D3a
c.1760 AD
c.1050 AD
D3c
Parking
WOF set 3
WOF set 2
WOF set 1
937- 982 cal. A.D.
538-390 cal BC (C2b)
Modern HWST
Modern Beachface
D2-1
D3a2
D3a1
D2a
C3
Granite headland
c.1780 AD
703-957 cal AD
D2
C2b
Beach Rock
524-194 cal BC
lagoonal mud
1- 836-971 cal. A.D.
2- 872-998 cal. A.D.
3- 688-795 cal. A.D.
4- 191-44 cal. B.C.
0 250m
-5 m
Ripple mark
Load cast
Basement
Saprolite
Colluvium
Humic soil
Tidal mud
Dune sand
Marine sand
Gravel splay
Gravels
Pebbels
bivalve
terrestrial snail
C-II
E
D3c-d
D3b3
D3b2
D3b1
D3a2
East
C-II
D3c-d
D3b
D3b1
D3a2
East
C-IV
E
D3c-d
D3b
D3b1
D3a2
East
West

systematically observed. The monitoring of washover deposits left by Hurricane Irene along the north-eastern coasts of the USA for instance showed a great variability in textural sorting patterns among the washover layers (Williams, 2015). These authors suggest that lateral changes in the textural/sedimentary characteristics of washover deposits may also be linked to changes in the sediment sources during the course of the overwash event.

The internal architecture of a washover layer depends on (i) whether the overwash reaches the back-barrier flat or remains confined to the barrier domain and (ii) in the case it extended to the back-barrier area, if it prograded over a subaerial flat or into subaqueous environment (a pond, a lake or a lagoon that was flooded at the time of deposition). Generally, subaerial overwashing leads to the deposition of horizontal to slightly landward-dipping sedimentary layers (Schwartz, 1975; Sedgwick and Davis, 2003; Wang and Horwitz, 2007; Weill et al., 2012). If the washover progrades into a subaqueous environment that is sufficiently deep to provoke a sudden deceleration of the overwash flow, then the horizontal stratification abruptly transforms into a more steeply dipping stratification (tabular delta foresets) generally reaching slopes near or at the angle-of-repose, i.e. of approximately 30° (Fig. 10; Schwartz, 1975; Sedgwick and Davis, 2003; Wang and Horwitz, 2007; Weill et al., 2012; Phantuwongraj et al., 2013; Shaw et al., 2015; Clemmensen et al., 2016). This transition between horizontal bedding and tabular delta foresets has been has been termed Topset-Foreset Break in slope (noted TFB hereafter; Shaw et al., 2015, Fig. 10). These downlapping tangential contacts can be used to differentiate between several washover deposits stacked one upon each other (Møller and Anthony, 2003; Wang and Horwitz, 2007, Fig. 10).

5.2. Identification of overwash events within sedimentary archives and use as proxies of past-storminess

5.2.1. Introduction

The identification and mapping of washover deposits within barrier and back-barrier sequences rely on corings, trenching (or occasionally natural exposures) and non-destructive imaging methods such as GPR (e.g. Møller and Anthony, 2003; Wang and Horwitz, 2007; Weill et al., 2012; Shan et al., 2016). GPR allows the investigation of large-scale features, and is therefore useful in imaging washover units over large areas and in depicting large-scale stratigraphical patterns (Møller and Anthony, 2003; Switzer et al., 2006; Wang and Horwitz, 2007). The efficiency of GPR in imaging sandy washover deposits is generally fostered by the large reflectivity contrasts between the pre-storm/post-storm deposits and the washover layers, induced by the differences in organic content, grain-size and presence of macro-debris such as shells and heavy-mineral concentrations. This proves true either where washovers are separated by clear intercalated organic layers (such as soil or organic marsh horizons) or where piled-up washover units are only separated by coarse sand or shell layers (e.g. Cunningham et al., 2011). Erosive surfaces, such as often characterizing the dune ridge and of the proximal parts of a washover also normally show up as well-defined GPR reflectors (Switzer and Jones, 2008). An additional proof-checking using invasive techniques such as trenching and coring is often necessary, not only to obtain material for analyses and dating, but also because GPR is only moderately appropriate for the observation of sequences that comprise thick peat/organic-rich layers such as in back-barrier areas. Furthermore, the finest structures of washover deposits may be overlooked by the GPR which normally only allow strata thicker than approximately 10–15 cm to be distinguished along the profiles (cf. Fig. 10; Møller and Anthony, 2003; Wang and Horwitz, 2007; Nielsen and Clemmensen, 2009; Hede et al., 2015).

Once the Holocene barrier and back-barrier sequences are outlined, the identification and characterization of washover deposits can rely on several indicators. The following sections will review the stratigraphical, sedimentological, mineralogical, macro- and micro-paleontological and geochemical indicators that can be used to extract past storminess records from washover deposits along sandy and mixed sand-gravel barriers and back-barrier domains.

5.2.2. Stratigraphical and sedimentological markers of overwash events

5.2.2.1. Principles of use. The most widespread use of washovers for paleotempestology has consisted in seeking alternations between fine-grained back-barrier sediments and washover deposits within Holocene back-barrier sequences. During the last decades, intensive research efforts were conducted on past overwash processes and their impacts on coastal systems, particularly in relation to the study of hurricane cyclogenesis, notably along the eastern coasts of the United-States (Liu and Fearn, 1993, 2000; Donnelly et al., 2001a, 2001b, 2004). Pioneer studies relied on the visual identification of sand layers embedded within fine-grained and organic lagoonal or coastal lake sequences. Liu and Fearn (1993) for instance identified several sand layers within the lake gyttja and lagoonal clay deposits preserved in Lake Shelby (Western Gulf of Mexico, USA). The deposition of these layers, by means of "tidal overwash" and by waves entering the lagoon at the time it was still open to direct marine influences, was considered by Liu and Fearn to be a direct consequence of hurricanes of categories 4 and 5 (Saffir-Simpson scale). Washovers were dated back to ca. 3600-3200, 2600, 2200, 1400 and 800 yrs cal. B.P., thus implying a recurrence period of ca. 600 yrs for hurricanes of these magnitudes. The same approach was subsequently applied to coastal lakes of Florida (Liu and Fearn, 2000), New Jersey (Donnelly et al., 2001a, 2004) and New England (Donnelly et al., 2001b). In this last study, fourteen cores were retrieved from the Succotash back-barrier marsh (Rhode Island, New England, USA). Up to six evenly spaced conspicuous sand layers could be observed within the 700 yrs long sequence, all abruptly truncating autogenic *Spartina* sp. peat marsh deposits, thus revealing sudden onsets of higher-energy events in the marshes (Fig. 11). All sand layers showed an upward gradual increase in organic content, suggesting a gradual return to marsh conditions and peat deposition after washover formation (Fig. 11). More recently, a high-resolution study of the sedimentary wedge preserved in the Mont-Saint-Michel Bay (Brittany, France) allowed

Fig. 12. **(A)** Location map of La Torche site (western Brittany, France) with stratigraphic transects studied by Van Vliet Lanoë et al. (2014). Red and black lines on the aerial photograph show the location of the stratigraphic profiles shown in (B) and (C), respectively. The white squares show the location of the drillings. **(B)** Stratigraphic profile obtained from drilling and exposures in the southern part of the area (Pors-Carn site) showing successive washover layers either composed of alternation of stratified fine-and coarse-grained sand (picture B-III) or of stacked coarse sand to gravel deposits (picture B-II). **(C)** Stratigraphic profile showing a succession of three generations of stacked washover deposits, materialized by well stratified successions of sand and gravelly sand. Ages are reported in years cal. B.P. All figures modified from Van Vliet et al. (2014) and reproduced with the permission of Sage publications.

• *Each washover body is interpreted as the result of high water levels induced by extreme storm events (and/or synchronous high tide and stormy conditions). Storm waves are suggested to have overtopped the barrier crest and to have deposited offshore originating sand and gravelly sand over the back-barrier flanks and lagoon/coastal mires. Presence of stratified dune sands in between the washover layers suggests the onset of dune formation during post-storm recover phases. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)*

the reconstruction of extratropical past storminess frequency (Billeaud et al., 2009). The sedimentary sequences comprise four sedimentary cycles of back-barrier mudflat sedimentation, interrupted by fine-grained sand deposits interpreted as washovers layers. These washover events (or periods of increased washover activity) were dated back to 5500–5800, 4000–4500, 3000, and 1000–1200 cal Yrs B.P. These dates interestingly suggested that periods with enhanced storminess and massive coastal disruption returned at a period of 1000–2000 years during the Holocene, in accordance with the cold events evidenced in ice-rafted debris by Bond et al. (1997). Such information is of particular interest, as it was obtained from a sedimentary record which was deposited under a macro-tidal regime (up to 15 m tidal range).

Although back-barrier investigations of washover deposits dominate the literature devoted to Holocene storminess studies, some research efforts have also been conducted to reconstruct Holocene storminess from the coastal sand-ridges themselves. The coasts of Europe are flanked by several coastal ridge systems either (i) mainly composed of dune-sand accumulations interrupted by storm-surges deposits (e.g. Jelgersma et al., 1995) or (ii) composed of piled-up washover deposits with local intercalations of aeolian deposits and peaty/soils layers (e.g. Van Vliet Lanoë et al., 2014). In both cases, the ridges have formed simultaneously to a decelerating RSL rise around the mid-Holocene period and are generally topped by Late Holocene aeolian sand dunes (with the last and most impressive dating back to the Little Ice Age; Lindström, 1979; Cunningham et al., 2011; Van Vliet Lanoë et al., 2014, 2016). Most of these ridges are transgressive features, and washover features can often be observed in bluff-exposures, providing excellent data for the reconstruction of Holocene storm activity. Cunningham et al. (2011) described shell beds embedded in dune sands at the Heemskerk coast (west Netherlands). These shell lags were seen by the authors to have been emplaced as perched fans by storm-surges invading the low-lying areas of the dunes and provided strong reflectors easily readable on the GPR profiles. Dense OSL dating made it possible to define very precisely the chronology of a storm surge that occurred around 1760–1785 A.D. in accordance with a massive storm event that hit the region in 1775 A.D., as reported by historical sources. Cunningham et al. (2011) concluded that the dating of storm-surge deposits by OSL in dune bluffs exposures could yield a temporal resolution of storm-events higher than the one obtained from back-barrier successions of organic and sand deposits. Van Vliet Lanoë et al. (2014) established a chronology of Holocene storm-events from the dating of perched washover terraces and sheet-washes preserved in coastal ridge deposits in Audierne Bay (Western Brittany, France, Fig. 12). Stacked sets of stratified coarse-grained sand and gravel layers and shell lags were observed to truncate peat deposits. These latter were considered as representing pre-storm surfaces and dated as such, while sediments topping the washover units where seen as representing the upper-age boundaries of the storm events (Fig. 12). The compound stratigraphy of the ridges made it possible to identify and date 21 storm events that struck the coast since ca. 7000 yrs B.P., a fair correlation being observed between the stronger events and the colder periods of the Holocene. Nonetheless, the washover units contain mixed material of beach and barrier domains, suggesting that the ridges have been continuously re-trimmed and reworked by Holocene storms, within a globally transgressive context, thus only resulting in a discreet preservation of distal washover deposits. The progressive disappearance of the older washover units, due to the landward roll-over of the ridges, was proposed as a crucial factor explaining why the storm frequencies reconstructed for the mid-Holocene period (8000–3000 yrs B.P.) seem lower than the frequencies reconstructed for the more recent times (Van Vliet Lanoë et al., 2014).

5.2.2.2. Usefulness for extracting information about paleo-storms intensities. Some attempts were made to try and relate sedimentological and stratigraphical features of the washover sequences to parameters of the storm event responsible for their deposition. Shaw et al. (2015) for instance studied the internal stratigraphy of a washover deposits formed by Hurricane Ike (2008) along the Matagorda peninsula (South-eastern Texas, USA) and measured the changes in elevation of the Topset-foreset Breaks in slope (TFBs) within the washover along a seaward to landward transect. It showed that the trajectories of TFBs elevation were usable features to track the elevation of the storm surge water-level over the back-barrier area and were related to the timing of the event. The authors suggested this approach could be applied to fossil washover stratigraphies (see an example of the tracking of the TBD on Fig. 10I–B').

5.2.3. High-resolution grain-size markers of overwash events

5.2.3.1. Principles of use. Visual inspection of successions with alternating washover and back-barrier deposits is often sufficient to identify sand deposits formed by major overwash events. Yet, visual inspection alone can overlook thin sand layers that may have been deposited by storms of less magnitude, at sites which were not located on the track of maximum intensity of the storms or from core taken in more distal location with regards to the former barrier position. Overlooking these thinner layers potentially induces biases in the calculations of storm frequencies. Also, in lagoonal configurations, background sedimentation may already include a significant amount of sand, so that the discrimination of individual storm layers can prove to be more difficult (Ercolani et al., 2015). To get round these limitations, high-resolution grain-size analyses have been be used to identify and characterize overwash events (Donnelly and Woodruff, 2007; Sabatier et al., 2008; Woodruff et al., 2008; Boldt et al., 2010; Wallace and Anderson, 2010; Lane et al., 2011; Sabatier et al., 2012; Brandon et al., 2013, 2014; Dezileau et al., 2016).

The identification of overwash events on the basis of grain-size analyses relies on the detection of peaks in the mean grain-size data. Grain-size measurements can be made (i) on the bulk sediment (Donnelly and Woodruff, 2007; Boldt et al., 2010; Naquin et al., 2014), (ii) on a specific grain-size fraction (whose choice will depend on the background grain-size characteristics of the sediment of each study site, e.g. < 1 mm, Sabatier et al., 2008), or (iii) on sediments sieved to get rid of the noise induced in the grain-size signal by macro-detritus such as shells (e.g. Sabatier et al., 2012) or treated to remove the organic fraction (e.g. Brandon et al., 2013). Since the average grain size decreases towards the distal margins of the washovers, grain-size peaks should be considered with respect to the location of the coring site relative to the estimated former position of the barrier at the time of the washover deposition (Donnelly and Woodruff, 2007). Donnelly and Woodruff (2007) established a chronology of hurricane landfall in Puerto Rico by analyzing peaks in the bulk coarse grain-size data obtained from cores retrieved from a back-barrier lagoon (Fig. 13). Such a high-resolution record allowed the authors to identify control-links between the hurricane activity in the Western Atlantic during the last 6000 years and the El-Ninõ/Western African monsoon patterns. Boldt et al. (2010) examined the occurrence of washover layers in cores retrieved from a back-barrier salt-marsh in New England (USA). High-resolution grain-size analyses conducted on the inorganic fraction of the sediments showed unambiguous peaks in the D_{90} grain-size that correlated well with the sand layers observed on digital X-ray logs of the cores. These analyses allowed the identification of 30 high-energy events having occurred during the last 2000 yrs, with an average occurrence frequency of ca. 1.5 events per century. Interestingly, this high-

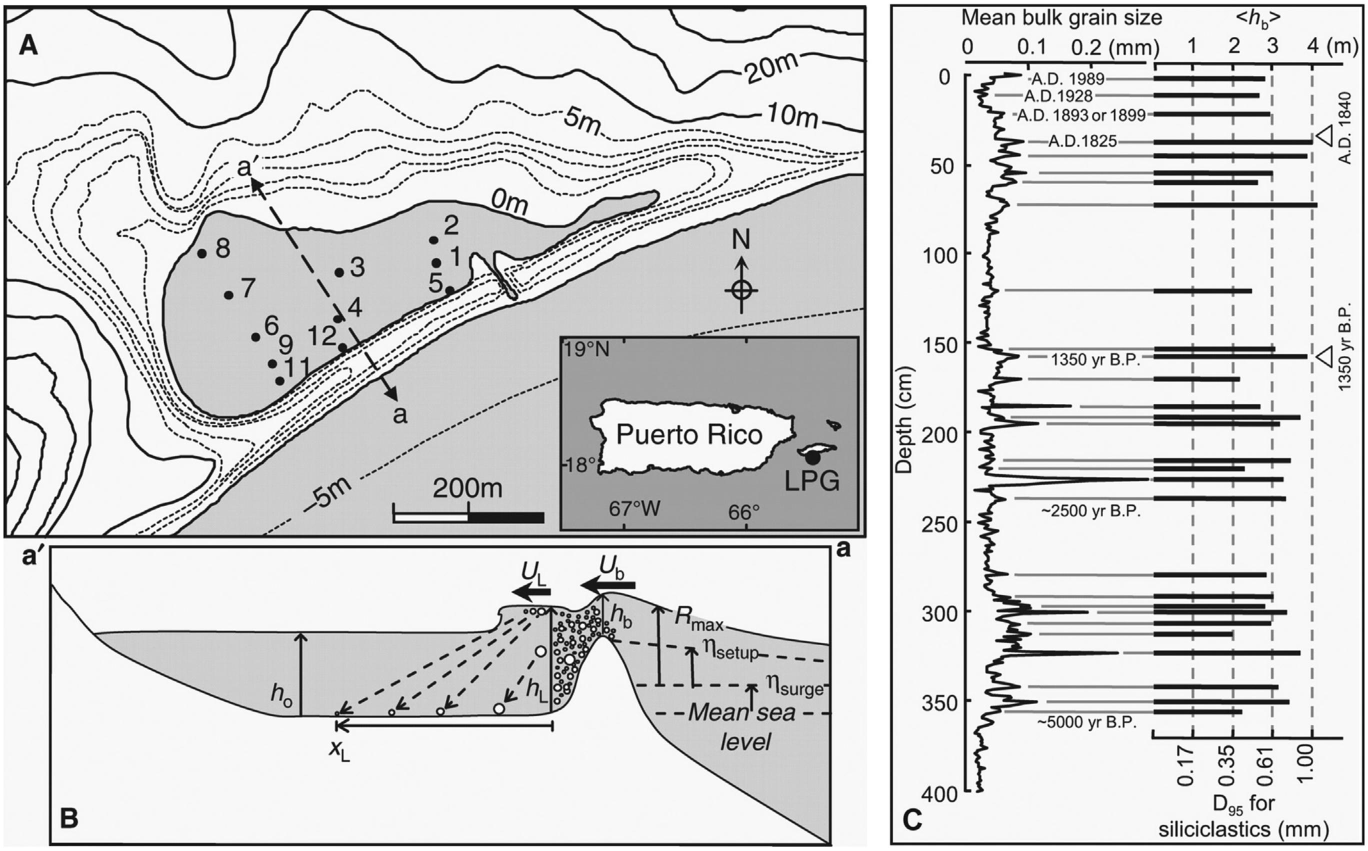


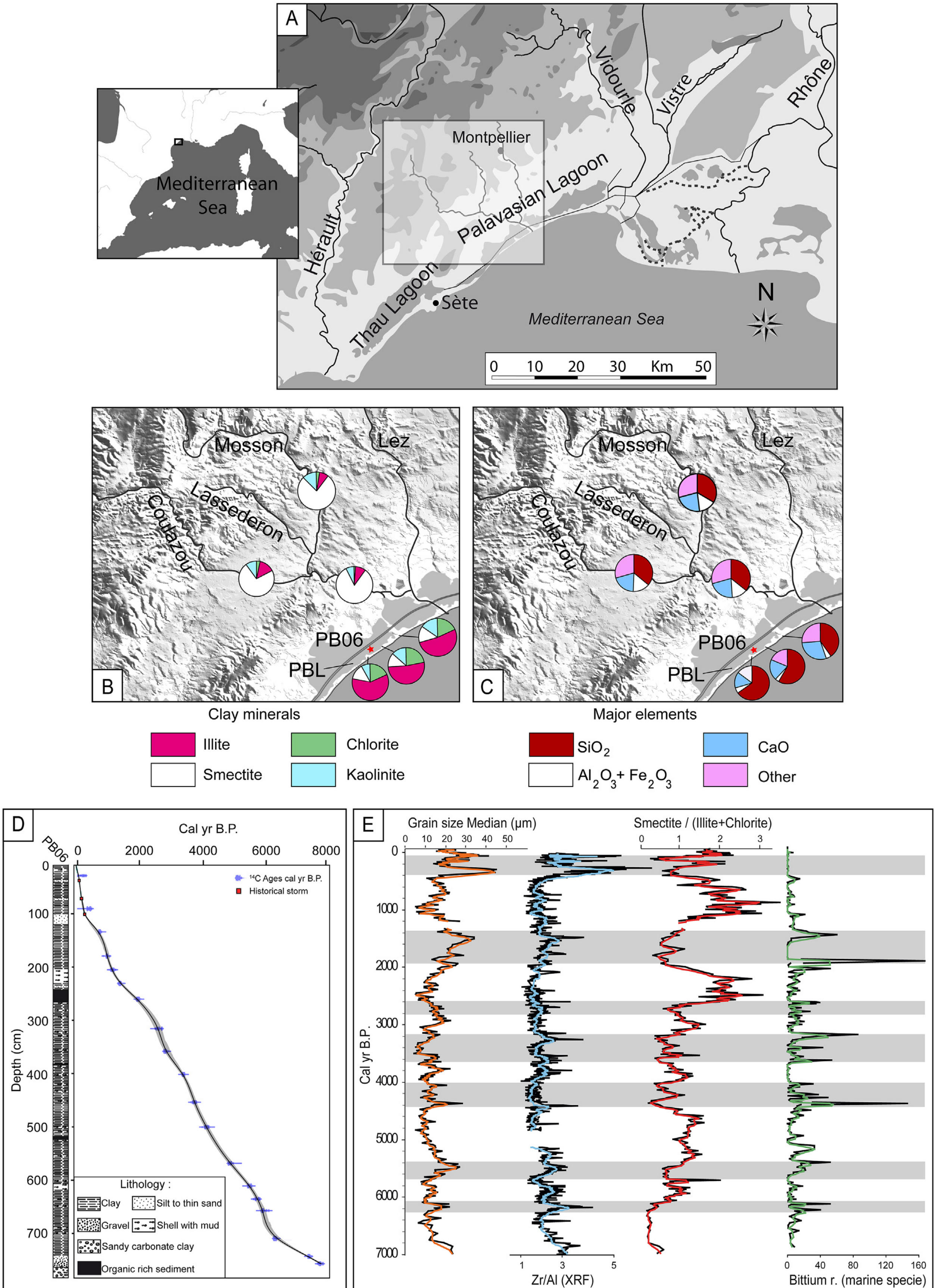
Fig. 13. **(A)** Location map of Laguna Playa Grande (LPG) (Vieques, Puerto Rico) studied by Woodruff et al. (2008). Dots indicate the location of coring sites and thedashed line (a-a') shows the location of cross section shown in (B). **(B)** Shore-normal cross section illustrating the overwash process described by the advective-settling model. <ηsetup> and <*ho*> represent time-averaged wave setup and lagoon water depth, respectively, while <hb>stands for the flow depth over the barrier. See Woodruff et al. (2008) for details and definitions of the others variables. **(C)** Down-core mean bulk grain size data for core 3 along with estimates of corresponding <*hb*> for isolated event layers. Equivalent D95 siliciclastic grain sizes used to calculate incremental values of <*hb*> are noted. All figures modified from Woodruff et al. (2008) and reproduced with the permission of the Geological Society of America.

• *Storms which brought coarser material into the lagoon are identified by grain-size anomalies (coarser sediments) in the cores. The distance to which offshore originating sand grains are transported by the flow inside the lagoon in interpreted to reflect the intensity of the flooding caused by storm-waves. Comparison with historical events of known intensity allows former hurricanes to be associated to a flooding-height, thus suggesting former hurricane intensities or distance to the hurricane track.*

resolution sedimentological record showed that no substantial gaps could be identified in the history of hurricane activity in the region, contrary to what was previously inferred from stratigraphical reconstructions (Liu and Fearn, 2000; Scileppi and Donnelly, 2007).

The robustness of using grain-size data as proxies of overwash events can be made stronger by using statistical analyses, in order to remove the mean background grain-size of the embedding sequence so that coarse-grained anomalies can stand out (Ercolani et al., 2015). Sabatier et al. (2008, 2012) for instance used the standard deviation of several grain-size classes to identify the one showing the larger variability through time relative to the mean grain-size of the succession. More sophisticated analyses were conducted by Lane et al. (2011) to reconstruct a high-resolution chronology of hurricane activity from sediments preserved in a Florida sinkhole. High-energy events were identified by running spectral analyses (Power spectra Thompson's Multi-Taper Method, “MTM”) on the bulk inorganic and on the >63 μm sand fractions sieved from the Loss-On-Ignition residues (see Thomson, 1982; Percival and Walden, 1993 for a detailed description of MTM spectral analysis methods). Coarse fraction time series were then high-pass filtered to isolate the storm-events from the variations in background sediments generated by longer time-scales (>30 years) environmental changes. This way, Lane et al. (2011) proposed an updated high-resolution chronology of hurricane history showing a recurrence interval of 3.9 events per century, much shorter than what Liu and Fearn (2000) previously inferred from the same site using stratigraphical evidences. Dezileau et al. (2016) studied the recurrence of extreme storms during the last 6.5 kyrs in the Western Mediterranean by analyzing a core taken from the Mar-Menor Lagoon (south-eastern Spain). Storms were identified on the basis of a combined use of sedimentological, mineralogical and macrofaunal proxies (the results of the two latter proxies will be described below in sections 5.2.4 and 5.2.5). Peaks in coarse-grained material were used as an unambiguous sign of storm-induced marine water intrusions into the lagoon, otherwise characterized by of quiescent silty and clayey sedimentation. Statistical analyses were carried out on the time series of sand percentage along the sequence using both “MTM” and Maximum Entropy (MEM) spectral analysis methods. They evidenced recurrence periodicities of significantly storminess activity of 1228 ± 327, 732 ± 80, 562 ± 58, and 319 ± 16 yrs.

5.2.3.2. Usefulness for extracting information about paleo-storms intensities. Spatial variations within the grain-size characteristics of washover deposits have been used to evaluate the magnitude of past storm events. Building on the study of Donnelly and Woodruff (2007) from Puerto Rico lagoon, Woodruff et al. (2008) established


A
Mediterranean Sea
Montpellier
Vidourle
Vistre
Rhône
Palavasian Lagoon
Hérault
Thau Lagoon
Sète
Mediterranean Sea
N
0 10 20 30 Km 50
B
Mosson
Lez
Lassederon
Coulazou
PB06
PBL
C
Clay minerals
Illite
Smectite
Chlorite
Kaolinite
Major elements
SiO_2
Al_2O_3+ Fe_2O_3
CaO
Other
D
Cal yr B.P.
PB06
Depth (cm)
^{14}C Ages cal yr B.P.
Historical storm
Lithology :
Clay
Silt to thin sand
Gravel
Shell with mud
Sandy carbonate clay
Organic rich sediment
E
Grain size Median (µm)
Smectite / (Illite+Chlorite)
Cal yr B.P.
Zr/Al (XRF)
Bittium r. (marine specie)

a relationship between the D_{95} grain-size (*i.e.* 95% of the grain-size distribution of the examined sample is finer than the considered grain size) and the height of the flow-depth over the barrier (Fig. 13B). Using the deposition from the last historical hurricane of 1928 as a modern analogue, Woodruff and colleagues built an advective-settling model describing the spatial sorting of overwash deposits with regards to the storm water level and the water depth within the flooded lagoon. This model was then inversely applied to calculate past inundation heights of 29 events identified in the lagoon core and dating back to 5000 yr B.P (Fig. 13C). However, the authors noticed that such results did not account for local changes in the barrier morphology along the last millennia as well as RSL rise whose changes surely impacted inundation levels and thereby the sensitivity of the site towards hurricane impacts. As stated by Brandon et al. (2013), an increase in grain-sizes towards the younger ages of a lagoonal succession may be seen as an increase in the magnitude of the storms while it may only reveal that coring location is getting closer to the coastline. To account for this, Brandon et al. (2013) developed an upgraded model in which changes in the grain-size of the siliclastic material advected to the back-barrier are corrected from regional RSL changes. In this latter study, nested models were used to establish quantitative relationships between the grain-sizes of the material carried within Spring Creek Pond (Florida, USA) during hurricanes, transport competences and, as a result, wind speeds and hurricanes intensities. An exponential relationship was observed to link wind speed and surges capable of transporting grains of size >63 μm. Results of the modelling experiments were validated against the sedimentary testimonies left by historical hurricanes of known magnitudes and surge heights. The inverse model was then subsequently applied to a Holocene core and used to reconstruct the magnitude of hurricanes dating back to ca. 2000 yrs B.P.

5.2.4. Mineralogical evidences of overwash events

5.2.4.1. Principles of use. Recently, clay mineralogy has been explored as an alternative way to trace storm events from back-barrier sedimentary successions. Needing less energy for transport, clay minerals would allow either (i) to trace back events either of lesser magnitude that did not transport sand into the back-barrier domain or (ii) to extract storm signal from locations that would have been too far from the coastline to receive sand inputs (Sabatier et al., 2010a, 2012). An additional benefit of the use of clay mineralogy is that clay material may be composed of several minerals, whose relative abundances can be used as markers of different sediment sources. Sabatier et al. (2010a) for example found clear differences between two sediment sources in the Pierre Blanche lagoon (Palavasian lagoonal complex, western Mediterranean Sea, France). The clay comprised in the sediments sampled in the watershed basin of the river that flows into the lagoon shows a large domination of smectite (73–81%), brought by the erosion of smectite-rich Cenozoic conglomerates in the upper part of the drainage basin (Fig. 14). On the other hand, the sandy barrier fronting the lagoon yields material showing high illite (45–59%) and chlorite (17–26%) concentrations within the clay fraction, nourished by detrital material from the Gulf of Lions platform that mostly originates from the Alps through the Rhône river. As such, the ratio smectite/(illite + chlorite) could thus be used as a proxy of past storminess, with minima showing changes in the sedimentation sources from the drainage basin to the marine environment and thus standing for periods of increased offshore material transport into the lagoon (Fig. 14). This approach, applied to a 7000 yrs-long sedimentary sequence retrieved from the Pierre Blanche lagoon, allowed Sabatier et al. (2012) to identify conspicuous past-storm events in good accordance with those evidenced by other independent sedimentological, macrofaunal and geochemical proxies (Fig. 14E).

5.2.4.2. Usefulness for extracting information about paleo-storms intensities. As it comes to evaluating storm intensities, Sabatier et al. (2010a) concluded that the smectite/(illite + chlorite) ratio was apparently not be an adequate indicator as the ratio did not keep pace with the strength of historic events of known intensities. An explanation could be that, as a ratio, it is not only sensible to the amount of barrier-originating material entering the lagoon (and thus to the intensity of the flooding) but also to the amount of material released by the rivers which is likely to have been heavily influenced by anthropogenic activities in the watershed on millennial to historical scales.

5.2.5. Macro- and micro-fauna fossils as proxies of overwash events

5.2.5.1. Principles of use. Redeposited macro- and micro-fossils have received a sustained attention during the last decades as indicators of high-energy marine events (from tsunamis to tropical and extra-tropical storms). By mobilizing sediments originating from the shoreface (either from fine-grained shoreface sediment or from coral reefs in tropical regions, e.g. Pilarczyk and Reinhardt, 2012) and barrier areas, storm overwash dynamics also displace macro- and micro-fauna assemblages indigenous to these domains and deposit them in the back-barrier areas as part of the washover layers. The presence of marine macro- and micro-fauna of offshore origin within barrier, lagoonal or coastal lake sequences is thus a powerful indicator of marine incursions into these environments, possibly caused by past hurricane or storm events.

A typical feature of washovers is the deposition of shells and shell debris mixed with the coarse-grained material or concentrated in well-defined blankets on top of the fine-grained deposits backing either sandy barriers or shelly ridges (e.g. Ruiz et al., 2007; Cunningham et al., 2011; Van Vliet Lanoë et al., 2014). Taxonomic analyses of these macro-faunal assemblages show that they constitute either of (i) offshore to shoreface bivalves and mollusks species, indicating a marine origin of the washover material and/or the reworking of former washover deposits (Sedgwick and Davis, 2003; Sabatier et al., 2008; Billeaud et al., 2009; Cunningham et al., 2011; Dezileau et al., 2016), (ii) terrestrial and freshwater/brackish snails, characteristic of dune and back-barrier environments indicating some reworking of the barrier and back barrier sediments (Jelgersma et al., 1995; Sedgwick and Davis, 2003; Van Vliet Lanoë et al., 2014), or (iii) fossils of both origins. Ruiz et al. (2007) and Engel et al. (2015) showed that a prominent characteristic of storm beds was a major increase in the abundance and diversity of mollusc species compared to the surrounding fair-weather deposits, revealing the reworking by storm waves of

Fig. 14. **(A)** Location map of the Pierre Blanche lagoon (southern France) studied by Sabatier et al. (2010a, 2012). **(B)** and **(C)** respectively show the concentrations of clay minerals and selected major elements in suspended sediment collected in the Mosson River watershed during a flood event and in sediment from the sandy barrier. The small red star indicates the location of the PB06 core retrieved fromthe lagoon. **(D)** Stratigraphic log and age/depth model of core PB06. **(E)** Down-core data from core PB-06, showing (from left to right) variations in grain size, XRF ratio Zr/Al, clay minerals content and marine indicative macrofossils individuals. Shaded areas show the areas identified as high storm activity periods. All figures reproduced from Sabatier et al. (2010a,b, 2012) with the permission of Elsevier.

• *Well-contrasting signatures in clay minerals and major elements are observed between the sediments originating from the coastal barrier and from the watershed of the river supplying the lagoon. Such indicators are used along with macrofauna (marine shells) as source tracers for sediments and allow to identify offshore originating material that was deposited into in the lagoon during storm-induced inundations. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)*

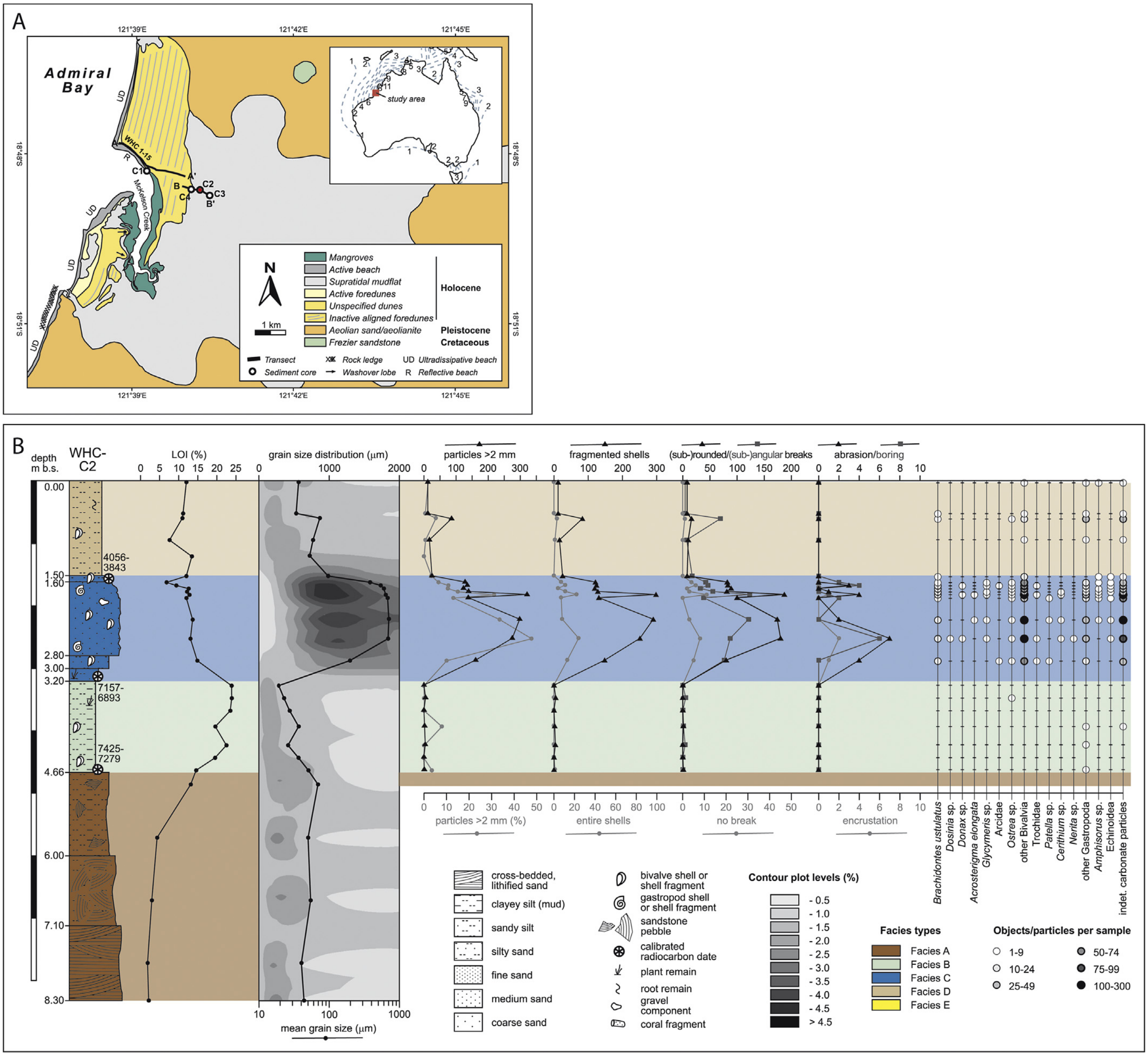


Fig. 15. **(A)** Map the McKelson Creek estuary area (Western Australia) studied by Engel et al. (2015), showing the location of the studied cliff exposure and of the drilling transects. The location of core-C2 shown in (B) is indicated by the red dot. **(B)** Synopsis of sediment core WHC-C2 with sedimentological results [loss-on-ignition (LOI), grain size distribution, shell taphonomy of carbonate macro-remains, state of preservation]. Radiocarbon dates are indicated on the stratigraphical column in calibrated B.P. ages. All figures reproduced from Engel et al. (2015) with the permission of John Wiley & Sons.
• *The massive presence shell macro-remains of mixed species and various taphonomic states can be used to characterize storm-flood episodes which would have accumulated shell-rich beds in the back-barrier area barriers and inside the swales during episodic breaching events. The presence of high percentages of complete shells with no breakage is proposed to mark particularly intense storm events whose waves would have activated deep coastal source areas where shells are not normally exposed to mechanical stress from wave action. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)*

sediment originating from several depositional environments and their homogenization within the storm layers. Sabatier et al. (2008) used mollusc assemblages to identify past storm events within Holocene successions in the Pierre Blanche lagoon (Western Mediterranean sea, France). Periods of high-energy conditions and marine water incursions were identified by pulses in the presence of marine malacofauna (*Bittium reticulatum* and *Rissoa ventricosa*) and deposition of fine sand, as opposed to normal conditions for which autochthonous lagoon species (*Hydrobia acuta, Cerastoderma glaucum and Abra ovata*) dominated the assemblages in silty sediments. While the occurrences of marine species in the successions appeared quite sudden, revealing the quasi-instantaneous character of the marine inundations, the return towards normal lagoonal conditions shown to be more gradual, thus showing the progressive closure of the storm-induced inlets (Sabatier et al., 2008). In their investigation of the washover features preserved in the coastal barriers of western Netherlands (see section 5.2.2 for further details), Jelgersma et al. (1995) and later Cunningham et al.

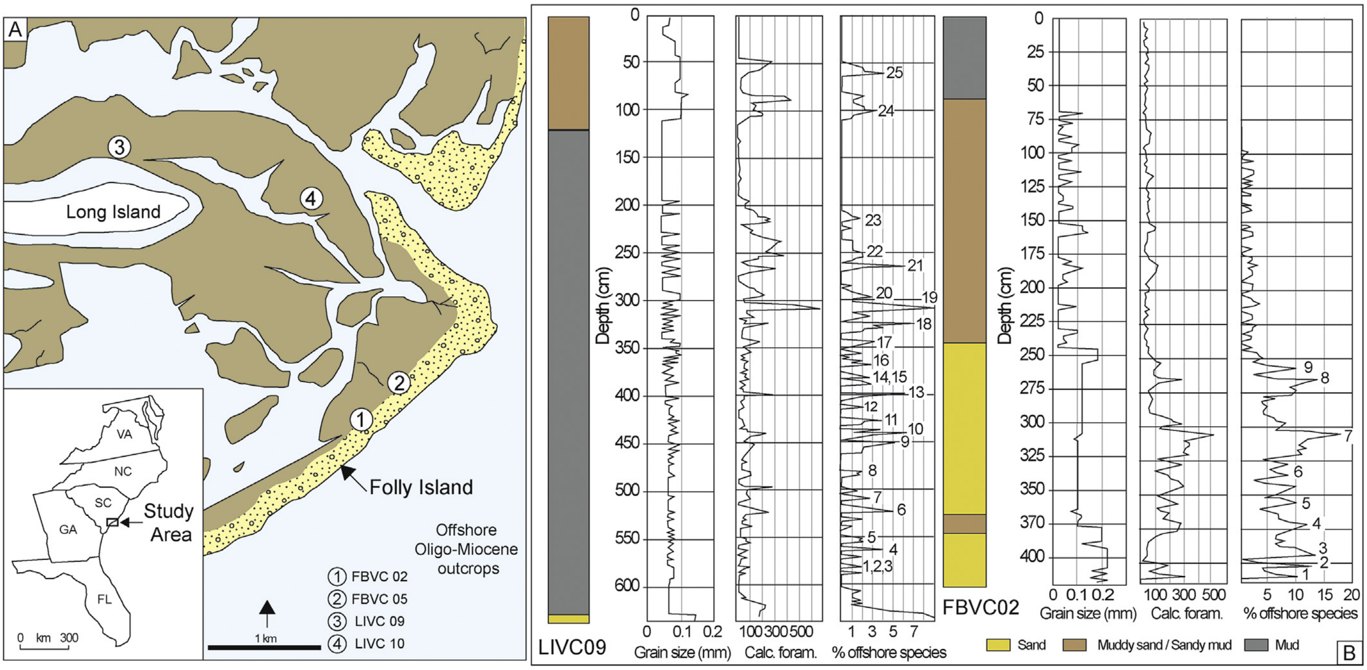


Fig. 16. **(A)** Location map of Folly Island (southCarolina, USA) studied by Hippensteel and Martin (1999). Numbers show the location of the vibracores retrieved from the back-barrier marsh area. **(B)** Stratigraphic logs of cores LIVC09 and FBVC02, with indications of mean grain-size, number of calcareous foraminifers and percentage of offshore foraminifera species. All figures modified from Hippensteel and Martin (1999) and reproduced with the permission of Elsevier.
• *Peaks in percentages of offshore foraminifera species within back-barrier sedimentary sequences are interpreted as signs of storm-wave inundation. In this study, the presence of foraminifera species originating from offshore Oligo-Miocene outcrops in the back barrier sediments allow to attest for a marine origin of the material and suggest deep wave action only permitted by extreme storm events. Differences in the numbers of peaks in between the two cores suggest complex preservations issues and possible redistributions of the foraminifera individuals upon the marsh due to tidal channel action.*

(2011) studied the taxonomy of the shells embedded within washover lags exposed in frontal dune deposits. The shell assemblages mostly consisted of mollusc species characteristic of shoreface (within a water depth of 15 m) to tidal-channel domains, with some occurrences of offshore species typical of deeper open sea origin and occasional terrestrial gastropod shells (Jelgersma et al., 1995). Such shell taxa indicate offshore origin of the material and aeolian transport can be ruled out due to the high number of the deposited shells. Engel et al. (2015) studied the shell populations preserved within the stratigraphy of relict Holocene foredunes in McKelson Creek (Western Australia) with some focus made on describing their taphonomic characteristics and defining taphofacies. High-energy backshore and washover deposits were shown to carry comparable macro-faunal assemblages. Nonetheless, washover layers were characterized by a higher percentage of complete shells, compared to the beach and tidal-channel deposits (Fig. 15). This was interpreted by Engel et al. (2015) to indicate that the shells preserved in the washover lags originated from deeper marine areas, where they had not been regularly exposed to strong mechanical stresses as opposed to the high-energy water motion of wave and current action characteristic of the beach and tidal channel domains.

Foraminifera and diatoms are the most commonly used microfossils for characterizing washover deposits preserved within the Holocene back-barrier successions (e.g. Parsons, 1998; Collins et al., 1999; Zong and Tooley, 1999; Hippensteel and Martin, 1999, 2000; Dawson et al., 2004; Andrade et al., 2004; Ruiz et al., 2007; Leorri et al., 2010). Other microfossils such as ostracods (Ruiz et al., 2007) are less commonly used. Pilarczyk et al. (2014) give a detailed review of the use of microfossils to document past extreme events, including paleostorms. With regards to foraminifera, washovers deposits are typically characterized by an enrichment in allochtonous offshore-specific species and mixed calcareous benthic species that contrasts with the marsh assemblages dominated by agglutinated species (e.g. Culver et al., 1996; Collins et al., 1999; Hippensteel and Martin, 1999, 2000; Andrade et al., 2004; Hawkes and Horton, 2012; Pilarczyk and Reinhardt, 2012; Hippensteel and Garcia, 2014). Hippensteel and Martin (1999) for instance investigated the Holocene back-barrier stratigraphy preserved at Folly Island (South Carolina, USA). Peaks in calcareous foraminifera and in the percentage of offshore species, termed “impulse” layers by the authors (*i.e.* revealing a short-lived event) were interpreted to indicate overwash events. As such, these allowed the identification of 8–9 overwash events within a ca. 1.5 m thick sand layer (Fig. 16). For some reasons most probably related to the acidity of local waters and marsh environments dissolving calcareous test of the offshore species, lack of nutrients availability or ecological parameters, sandy washover deposits can also be totally devoid of foraminifera and, this way, may also stand out from the microfossil-rich marsh assemblages (e.g. Horton et al., 2009; Williams, 2015). Using diatoms, marine water overwash events are identified both by the presence of peaks in Polyhalobous (planktonic and benthic) and lows in Halophobous diatom species (Zong and Tooley, 1999; Andrade et al., 2004). Zong and Tooley (1999) published an extensive study based on diatoms found in the Holocene succession of the Roudsea Marsh (Morecambe Bay, northwest England). After the site was shielded from direct marine influence due to barrier closing, the diatoms population became dominated by halophobous species reflecting the dominating freshwater context. Nine flooding events were identified within this freshwater succession between ca. 6000 and 5700 yrs B.P., characterized by clear peaks in the presence of marine planktonic or tychoplanktonic *Polyhalobous* and *Oligohalobous* diatoms species (dominated by *Cymatosira belgica*, *Paralia sulcata* and *Podosira* *stelliger*) synchronous with minima in autochthonous halophobe specimens.

The monitoring of modern storms has highlighted that when a storm surge does not possess sufficient energy to bring extensive amounts of sand material to the back-barrier, it can nevertheless transport lighter material, such as microfossil tests. Collins et al. (1999) for instance studied the traces lefts by Hurricane Hugo (1989) in a pond of South-Carolina (USA) and showed that, in some areas of the pond, the overwash induced by this event could only be identified by layers enriched in offshore foraminifera. Microfossils can also be instrumental in identifying washover sediments that deposited upon lagoonal environments whose coarse-grained background sedimentation makes washover deposits more difficult to recognize on the basis of sedimentological markers alone (Pilarczyk et al., 2011). Another strength of microfossil proxies is that they can also provide insights on sediment sources, that may be used for cross-checking information given by stratigraphical and sedimentary proxies and help to alleviate doubts left about the overwash origin of some sand layers. Within lagoons, the dominating presence of offshore-originating foraminifera species in a suspected storm sand deposit can for instance allow to discriminate between washover and tidal creek deposits, the latter being characterized by species specific of lagoonal flats and tidal creeks (Hippensteel and Martin, 1999). Nonetheless, caution must be exercised as the transport of foraminifera tests as suspended matter is highly variable and strongly depends on the size and specific density of each specimen. Some foraminifera, notably those characterized by thick tests have densities very similar to that of quartz grains (Horton et al., 2009). Also diatoms were found useful to demonstrate the reworking of dune barrier sediments by storm surges and their incorporation within washover lags, thus providing an additional proof of the offshore origin of sandy material preserved in back-barrier successions. On the contrary, diatoms can also show that some sand material may not have been transported by overwash processes. Dawson et al. (2004) for instance showed the dominating presence of brackish and freshwater diatom species characteristic of the modern beach and dune areas in an extensive sand layer of some Holocene back-barrier successions of the Outer Hebrides islands (NW Scotland). This led the authors to propose that the sand layer was rather wind-blown by storm winds than deposited by marine inundation. Recently, some developments have been made in using foraminfera taphofacies analyses to provide additional information about the hydrodynamic regimes of the overwash events. Pilarczyk and Reinhardt (2012) described the taxonomic and taphonomic characteristics of reef-autochthonous *Homotrema rubrum* (Lamarck) foraminifera found in the modern sub-environments of hypersaline lagoons of Anegada Island (British Virgin Islands, Caribbean) and compared them to observations made on a washover layer that was deposited during the 17th-18th century. Very similarly to the modern beach, reef-flat, and storm wrack deposits, the washover layer showed the presence of “well-preserved” and “exceptionally preserved” (unaltered) *Homotrema* tests, thus unambiguously showing an offshore overwash origin of the deposit.

5.2.5.2. Usefulness for extracting information about paleo-storms intensities. As most of the macro- and micro-fauna species have well-defined habitats determined by ecological parameters (water-depth, hydraulic agitation, exondation and inundation frequency …), the observation of particular assemblages of species of shells or microfossils in the washover layers can yield supplementary information on storm magnitude e.g. by measuring the depth of wave action (Hawkes and Horton, 2012; Engel et al., 2015). Also, the co-existence, within a same sequence, of washover deposits either containing foraminifera or devoid of them has been proposed as potentially providing insights on both the sediment sources and the wave-regime of a particular storm. Indeed, as stated by Williams (2015), sediments originating from the sub-to inter-tidal foraminifera-rich domains would be more likely to provide fossil specimens to the washover than material originating from the supra-tidal beaches and dunes, which are environments naturally devoid of foraminifera.

5.2.6. Elemental analyses and organic geochemistry as proxies of overwash events

5.2.6.1. Principles of use. Recent advances in paleotempestology research have been achieved through the use of elemental analyses (major and trace elements) and organic geochemistry as new proxies of past storminess in coastal back-barrier environments. These new proxies represent valuable supplements to aforementioned stratigraphical, sedimentological, mineralogical or paleontological proxies of past storminess, as they bring additional information regarding the sediment sources.

The use of elemental analyses in the context of paleotempestology studies lean on the fact that, depending on where they originate from, sediments can be characterized by very different concentrations in major and trace elements. Therefore, elemental analyses can reveal information about the origin of the material preserved in Holocene sedimentary successions and bring some enlightenment on the mechanism by which they were emplaced. Differences in the sediment composition result, for example, from lithological differences, from the presence/absence and importance of the biogenic fraction and from the presence/absence and proportion of heavy minerals (e.g. Sabatier et al., 2012; Degeai et al., 2015; Dezileau et al., 2016). The contrasting geochemical characters of terrestrial detrital and of offshore-originating sediments can thus be used to document the origin of the sediment inputs within the lagoons and coastal lakes (and thus allow to discriminate, for example, between periods of marine inundations and/or periods of heavy rainfalls triggering inputs of terrestrial material). By allowing an evaluation of the importance of the terrestrial sedimentary inputs to the systems, elemental analyses can help to lift potential doubts about the origin of some of the sand layers, when this cannot be obtained on the sole basis of stratigraphical or sedimentological indicators. Results of elemental analyses are generally reported as the ratio of two elements, the denominator being relevant to a conservative element (aluminum is often used for its near-constant concentration in rocks, Sabatier et al., 2012) or presented as plots of the concentration of one or several elements which are characteristic of a particular sedimentary environment. Strontium (Sr) has often been used as an indicator of overwash dynamics because marine-sourced material is generally enriched in this element, due to the presence of biogenic calcium carbonate originating from shells and shell fragments, coral or algal material (Woodruff et al., 2009; Degeai et al., 2015; Raji et al., 2015). As each site is characterized by specific sediment sources, the geochemical “signature” of local sediments must be locally calibrated by the study of modern sediments in order to determine which relevant major and trace elements are pertinent to be used within the paleo-environmental reconstructions (Fig. 17, Degeai et al., 2015). In their study of lakes Namakoiike and Kaiike (southwestern tip of Japan), Woodruff et al. (2009) reported Sr concentrations up to 4 times larger in the sand of the barrier fringing the lakes than in sediment sampled along the watershed and within the tributaries feeding the lakes. Simultaneous peaks in coarse-grained material and Sr concentrations within the Holocene sequences of both lakes were related to periods of active marine inundation, most likely induced by enhanced typhoon activity around 4800-4200, 3600-3100, 2800-2500 and 1000 yrs B.P. Close correlations were observed between these periods and periods of

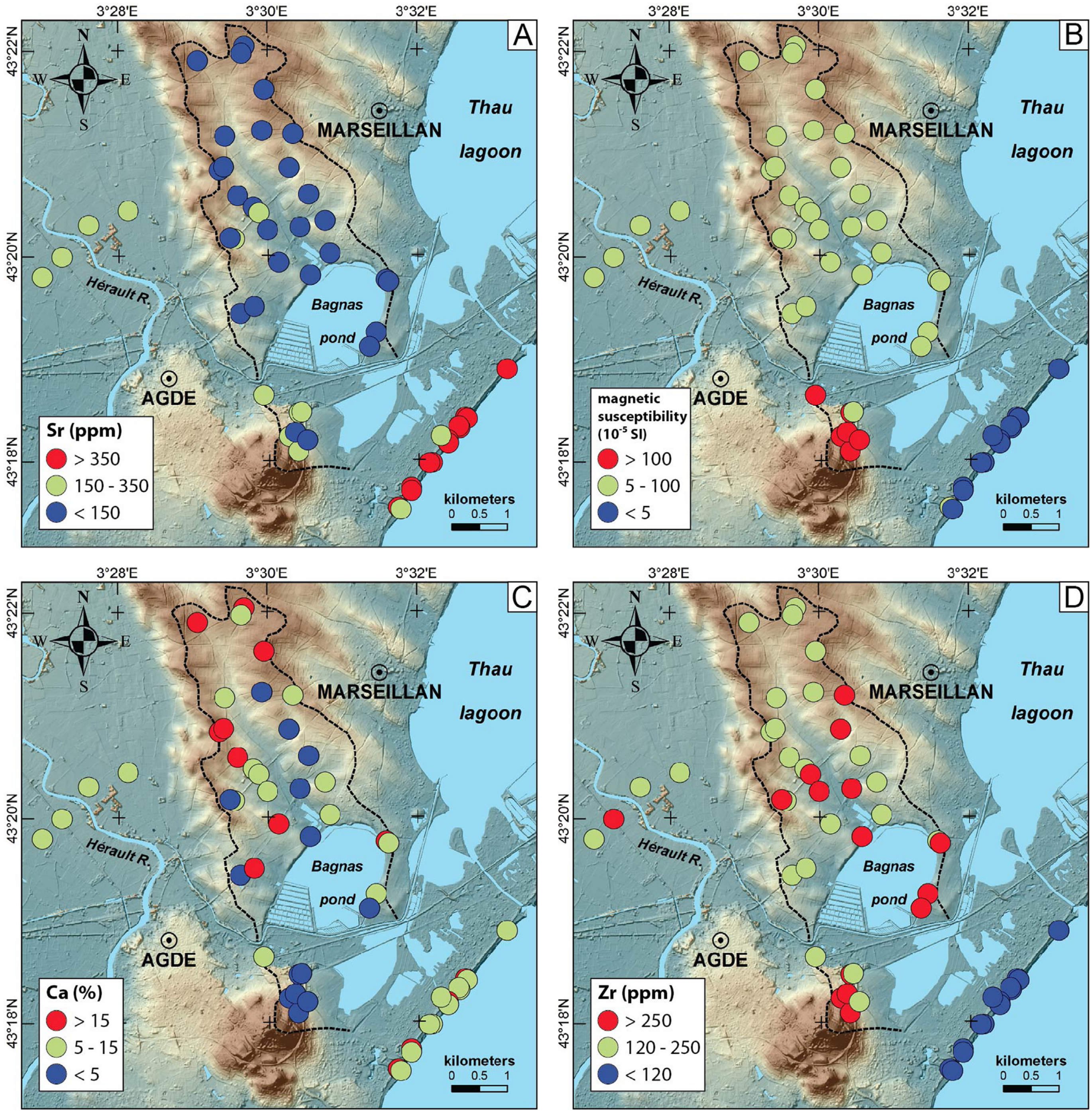


Fig. 17. Maps of **(A)** strontium (Sr) contents, **(B)** magnetic susceptibility values, **(C)** calcium content (Ca) and **(D)** zirconium (Zr) content obtained on samples taken from the watershed, pond and coastal barrier areas around the Bagnas pond (southern France) by Degeai et al. (2015). Figure reproduced from Degeai et al. (2015) with the permission of Elsevier.

• *Clear differences are observed between the Sr, Zr and magnetic susceptibility values given by the Bagnas watershed/Hérault river floodplain sediments (mainly inherited from Pleistocene alluvial terraces) and the ones obtained on the sediments sampled on the Holocene barrier. Such spatial discrimination promotes the use of these indicators as source tracers of the sediments in the back-barrier lagoon. It allows for the discrimination between terrigenous detrital material brought in by river floods and offshore originating material deposited under storm flooding regimes.*

moderate to strong El-Niño events, thus advocating for a storm origin over a tsunami origin of the identified events. Sabatier et al. (2010a,b; 2012) analyzed the elemental composition of modern sediments sampled within the main sedimentary suppliers to the Pierre Blanche Lagoon (western Mediterranean Sea, France). Analyses were made on barrier sands and on suspended sediment collected during a flood event in the Mosson River and other tributaries of the lagoon system. Highly contrasting signatures were obtained for these two sediment sources (Fig. 14): material from the watershed was shown to be dominated by Al_2O_3, Fe_2O_3 while barrier sediments expressed high concentrations in SiO_2, Na_2O, CaO, Sr and Zr characteristic of the Holocene marine sediment of the Gulf of Lions platform (Sabatier et al., 2010a, 2012). Peaks in the Si/Al and Zr/Al ratios along a sedimentary core retrieved from the lagoon showed close correspondence with peaks in the presence of the marine shell species (*Bittium reticulatum*, cf. section 5.2.5 of this paper). These provided supplementary evidence of the opening of major breaches in the barrier suspected to be related to seven storm events during the past 6300 yrs B.P. (Fig. 14).

Organic Geochemistry Proxies (OGPs, as coined in Lambert et al., 2008) refer to carbon and nitrogen stable isotope ratios ($\delta^{13}C$ and $\delta^{15}N$, respectively), Total Organic Carbon and Total Nitrogen contents (TOC and TN, respectively) as well as to the Carbon to Nitrogen (C/N) atomic weight ratio. During the last decades, OGPs have been fruitfully used for paleo-environmental reconstructions (see Lamb et al., 2006 for a review) and provided valuable information about the organic matter sources in the coastal zone. These methods have also made it possible to decipher between several depositional environments and has for instance been used for the reconstruction of Holocene RSL changes from salt-marsh environments (e.g. Wilson et al., 2005a, 2005b; Kemp et al., 2010, 2012; Engelhart et al., 2013; Khan et al., 2015; Goslin et al., 2015, 2017). A few studies have also highlighted the potentialities of OGPs as proxy indicators of past flooding events. Particularly, these indicators have been suggested to possibly yield valuable information about past storm events even when coarse-grained storm deposits were absent or indistinguishable in the sediment record (Lambert et al., 2008; Das et al., 2013). The organic geochemical signature of a sedimentary environment is determined by the type and origin of the Organic Matter (OM) entering and depositing into this environment (or more generally by the relative amount of the different OM sources). OM can originate from several sources, either autochthonous (such as vascular plants growing in situ on the sediment surface), allochtonous sources (e.g. brought by rivers or tide, such as particulate or dissolved organic carbon) or a mix of both. Any of these sources are characterized by specific signatures in $\delta^{13}C$, TOC, TN and C/N ratios. The largest differences occur between the marine- and terrestrial-originating OM, the former being remarkably more enriched in $\delta^{13}C$ and usually showing lower C/N ratios than that characterizing the OM originating from terrestrial vascular plants. Based on this, the rationale for using OGPs as proxies of storm occurrence within back-barrier sedimentary sequences is dual:

(i) As marine water inundates the back-barrier domain, it brings material of marine origin whose geochemical signature is remarkably different from the ones provided by the brackish to freshwater material that are normally deposited within the lagoon and back-barrier ponds (marine-originating material is typically characterized by higher $\delta^{13}C$ and $\delta^{15}N$ values, Lambert et al., 2008; Das et al., 2013).

(ii) By introducing massive amounts of salt water in what is normally a freshwater environment, storms can provoke short-lived dramatic changes in the ecological state of back-barrier coastal lakes, which in turn induce excursions in the OGP values. In their study of the response of the freshwater coastal Lake Shelby (north-eastern Gulf of Mexico) to storm events, Lambert et al. (2008) described that during intense marine inundation events, lakes may enter a “flooded” state. Seawater being characterized by high concentrations of nutrients (in the form of nitrate and Dissolved Inorganic Carbon), massive marine inundation may foster a temporary and rapid eutrophication of the lake leading to algal blooms (Lambert et al., 2008). Such a shift in the ecological regime of the lake would provoke intense excursions on the OGPs values, marked by clear increases in $\delta^{13}C$ and $\delta^{15}N$ values and drops in TOC, TN and C/N ratio values. Lambert et al. (2003, 2008) proposed that high-resolution storm-related signals could be recorded over an entire back-barrier basin by the use of OGPs proxies. Indeed, using this approach, Lambert et al. (2008) found evidence of 11 storm events that influenced Lake Shelby during the past 682 years, leading to estimate a mean recurrence rate of ca. 62 yrs, ten times shorter than the storm recurrence interval of ca. 600 yrs previously obtained by Liu and Fearn (1993) on the same site using stratigraphical evidences. Das et al. (2013) applied the methodology of Lambert et al. (2008) to two cores taken in back-barrier lakes of the Gulf of Mexico (Fig. 18, Western and Eastern lakes, Florida, USA). These authors identified several layers showing positive shifts in $\delta^{13}C$ and $\delta^{15}N$, accompanied with small decreases in the C/N ratios, again interpreted as the consequences of storm events having turned the lakes into a marine flooded state. Up to 22 and 35 of these layers were observed along two cores taken in the Eastern Lake, which would have been respectively deposited during the last 2400 years in one of the cores and during the last 2900 years in the second one (Fig. 18). In the Western Lake, up to 45 storm layers were identified using OGPs along a sedimentary sequence extending back to ca. 3900 yrs B.P. These results allowed the authors to calculate comparable storm recurrence intervals of around 85 years for the two lakes for the last 2900 and 3900 yrs, respectively, much shorter than the recurrence interval previously deduced from the same sites by using stratigraphical proxies (Liu and Fearn, 2000).

5.3. *Dating washover deposits*

Obtaining Holocene chronologies of the deposition of washover material has mainly relied on the use of radiocarbon-based (^{14}C) dating. Some chronologies have been established on the basis of ^{14}C dating shells found within the washover sand layers (Williams, 2013) but this approach requires to ensure that shells have not undergone extensive reworking prior to their deposition by the storm surge. This pre-requisite is rather tricky to perform, as it is likely that older shells may be released during the erosion of the barrier sediments and incorporated in the washover deposits. Small shells are more fragile and, if they are unbroken, can be of some help in achieving such a check and should be preferably chosen for dating (Williams, 2013). Most of the time, organic datable material is lacking within the sandy washover deposits and chronology of deposition is determined by dating the bounding back-barrier organic deposits. Some studies have dated the topmost pre-storm deposits (Scileppi and Donnelly, 2007), others the lowest part of post-storm organic deposits overlying the washover sediments (e.g. Kiage et al., 2011; Williams, 2013), while some works relied on both pre-storm and post-storm sediments to bracket the age of washover deposition (e.g. Donnelly et al., 2001a; Dawson et al., 2004). Dating the top of the underlying sediment layer (pre-storm surface) can give quite a good approximation of the age of washover deposition if this latter has deposited above an active depositional surface (e.g. a back-barrier marsh) and if the surface has not been eroded during the storm event. Dating the overlying marsh deposit may lead to some imprecision due to the time needed for organic sediment to reform again over the washover layer and thus gives a maximum estimates of the age of the washover (Donnelly et al., 2001a). Some studies have nonetheless documented sharp contacts between the washover layers and overlying marsh deposits, suggesting that fine grained sedimentation may return quite fast after the storm event (Kiage et al., 2011).

A basic assumption of radiocarbon dating is that dated sampled have incorporated carbon in equilibrium with its contemporaneous

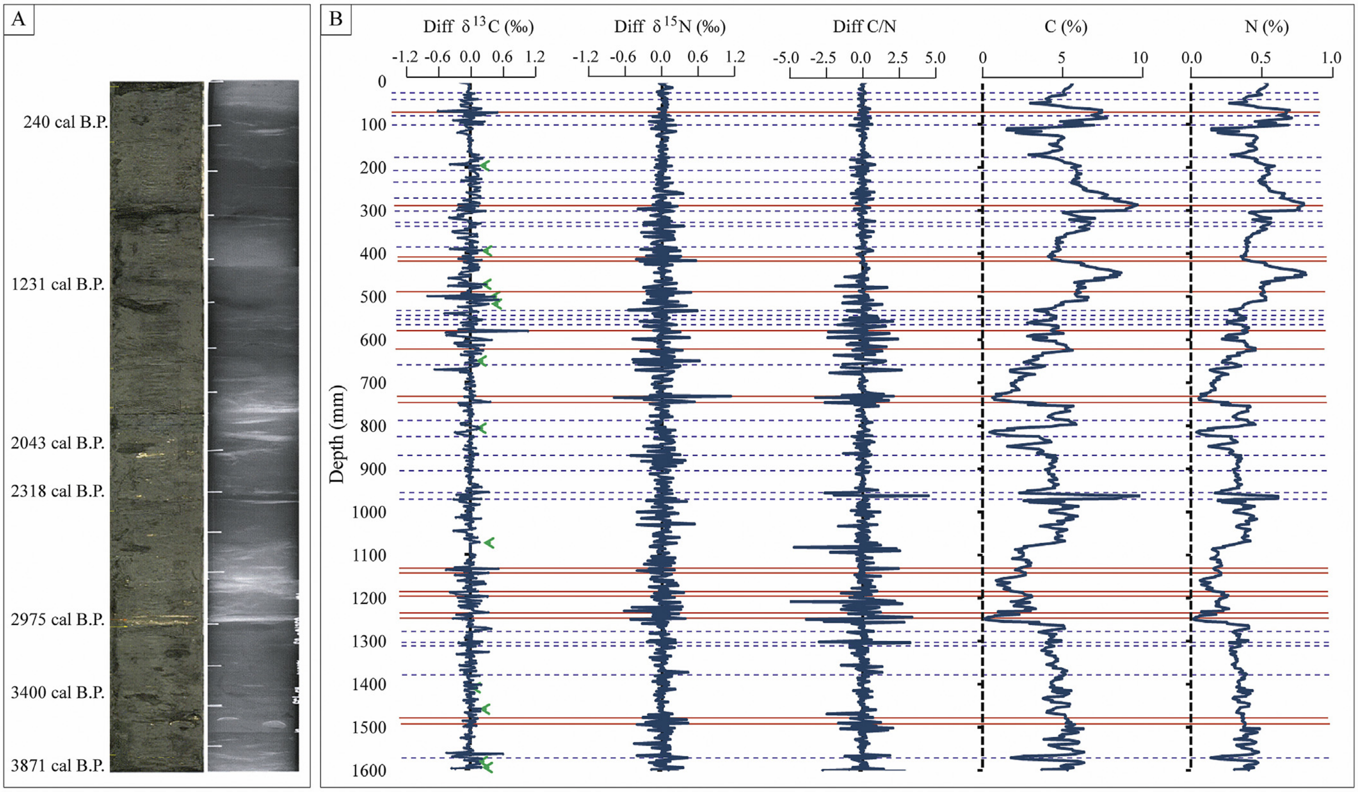


Fig. 18. **(A)** Digital image and X-ray image of a core taken by Das et al. (2013) in Western Lake (Florida, USA). Ages are reported in cal. years B.P. **(B)** Down-core detrended values of $\delta^{13}C$, δ^{15} N, C/N ratio, along with original C (%) and N (%) values. Note that $\delta^{13}C$, δ^{15} N, C/N ratio were detrended to remove any long-term trend by subtracting the next value from the current value. Figure reproduced from Das et al. (2013) with the permission of Elsevier.
• *Concurrent positive shifts in $\delta^{13}C$ and $\delta^{15}N$ and negative shifts in C/N ratio are interpreted to indicate flooding of the back-barrier lake area by storm event. These result either from (i) massive inputs of marine originating (large overwash events) material whose organic geochemical signature is characterized by lighter $\delta^{13}C$and $\delta^{15}N$ values and smaller C/N ratios than the terrestrial originating material into the lagoon or from (ii) smaller inputs of seawater (discrete overwash events) suggested to provoke eutrophication spikes in the lake due to input of fresh marine nutrients enriched in ^{13}C and ^{15}N.*

atmosphere. When dating marine-originating material or sediments from enclosed systems such as lakes and even lagoons by radiocarbon dating methods, anomalously old ages can arise from the reservoir effect, as well as from the incorporation of autochthonous old material in the sediments (Björck et al., 1998; Björck and Wohlfarth, 2001; Sabatier et al., 2010b). The general reservoir effect (also called "soft-water" reservoir effect, Björck and Wohlfarth, 2001) is linked to the fact that marine of freshwater material (either bivalves or aquatic plants) will acquire carbon directly from the water who carry lower ^{14}C levels than the atmosphere. The "hardwater" reservoir effect is due to the incorporation of old ^{14}C free inorganic carbon to the environment, thus "contaminating" the age of the water pool. Such old carbon can for instance be brought in by the erosion of old carbonate rocks from the watershed, by inputs of glacial material and/or melt glacial water as well as by groundwater seepage. Both soft- and hardwater reservoir effects lead to anomalously old ages if aquatic plants or shells which incorporate CO_2 from the water are used for dating. The reservoir effect has long been evaluated and taken in account for marine environments (e.g. Stuiver et al., 1986) but has frequently been quite overlooked within the context of lagoonal and lacustrine environments. Sabatier et al. (2010b) evaluated the reservoir effect of the Pierre Blanche lagoon (France) by comparing the ages obtained on two recent mollusk shells using AMS ^{14}C dating to chronologies derived from $^{210}Pb/^{137}Cs$ measurements and to the chronological controls provided by historical storm events (Sabatier et al., 2008). The results showed that (i) that the reservoir effect obtained from the lagoon was very large (943 ± 25 years), (ii) that is was higher by up to 600 years than the average regional marine reservoir effect and (iii) that it increased towards the late Holocene as the lagoon got progressively isolated from the sea due to the growth of the barrier. Thus, a thorough assessment of the reservoir effect must be established if the storm chronology is to rely on an age model built on either ^{14}C ages of bulk sediments, aquatic plants or aquatic shells. The incorporation of old organic remains such as charcoals, drift wood or pollen grains released to the systems after the erosion of old formations is also to be taken in account as it is known to cause serious biases towards erroneously old ^{14}C ages (Björck and Wohlfarth, 2001). To get round both the reservoir effects and the potential contamination of the sediment by old organic material, dating of bulk sediments is best avoided. If present in sufficient quantity in the sediment and adequately preserved, selected macrofossils of identified terrestrial or subaerial salt-marsh plants (which take their CO_2 from the atmosphere) must always be preferred for dating (Björck and Wohlfarth, 2001).

Radiocarbon dating of washover deposition has often been cross-checked by using alternative age-controls: bio-stratigraphic markers such as pollens periods (Donnelly et al., 2001a,b; Scileppi and Donnelly, 2007); heavy-metals pollution horizons related to anthropogenic industrial activities such as lead (Donnelly et al., 2001a, 2001b), lead and copper (Scileppi and Donnelly, 2007), titanium (Woodruff et al., 2008), or mercury and zinc (e.g. Brandon et al., 2014); ^{137}Cs radionuclide activity (injected in the atmosphere following nuclear weapon tests and nuclear accidents, Donnelly et al., 2001a,b, 2006, Boldt et al., 2010). More recently, chronologies of washover deposition have been achieved, for the last centuries, by precise absolute chronologies of deposition obtained from ^{137}Cs and ^{210}Pb radionuclides decay (e.g. Sabatier et al., 2008; Lane et al., 2011; Ercolani et al., 2015). Alternative dating approaches have also recently been developed in relation with OSL-dating of sands (Madsen et al., 2009; Davids et al., 2010; Clemmensen et al., 2014b). Using OSL directly on sandy deposits offers major advantages over the other dating methods (Madsen et al., 2009): (i) it provides an absolute chronology for washover deposits, (ii) it does not suffer from the calibration problems which sometimes plague radiocarbon dating and (iii) it is applicable to a broad time scale from the very recent periods to several millennia. Erosion and reworking of older sediments from the barrier during the overwash can nonetheless be a concern when using OSL dating. Reworked sediments may not have been exposed to light long enough before deposition to ensure that prior existing luminescence signal has been reset (insufficient “bleaching”), thus leading to potentially overestimated washover ages. The sensitivity of OSL results to water-content at the time of deposition and during the burial history was also reported to be an issue by Madsen et al. (2009). To try to overcome these limitations, Davids et al. (2010) compared the results of the dating of washover sands in New England (USA) given by OSL on quartz grains, InfraRed Stimulated Luminescence (IRSL) from K-feldspar and subtraction method (details in Davids et al., 2010). Quartz OSL and K-feldspar IRSL based dates were coherent, K-feldspar IRSL giving the more precise ages, while the subtraction method produced the larger uncertainties.

5.4. Limitations in the use of overwash testimonies as indicators of past-storminess

5.4.1. The importance of local morphodynamic factors in overwash occurrence

The occurrence or non-occurrence of overwash events and the subsequent deposition of washover layers over the barrier and back-barrier areas are the result of an assemblage of highly-complex and interdependent processes. First, for a given storm, the probability that a site is concerned by overwash processes will be highly dependent on the distance separating the site and the location of the storm-surge maximum height. The site-specific storm-surge water level is one of the critical aspects of storm-reconstructions based on wave induced processes (Morton, 2002). Plant et al. (2010) and Hawkes and Horton (2012) for instance reported the very rapid spatial dampening of the height of the storm-surge and of the impacts caused by hurricane Ike within only tens of kilometers from the location where the eyewall of Ike made landfall. The spatial variability of storm impacts with regard to the distance to the storm-path can become problematic when dealing with paleo-storm reconstruction because the precise location of where the storms made landfall is unknown (Collins et al., 1999; Hippensteel et al., 2013). Deposition of washovers assumes that the storm-surge level was higher than the barrier crest. Along tidal coasts, for a given elevation of the barrier crest, the probability that the storm-surge is high enough to overtop the barrier will largely depend on the synchronism between the moment the storm hits the coasts and the time within the tidal cycle. Thus, as was previously noted, regions concerned with reduced tidal-ranges are the most likely to have recorded storm signals in the form of washover deposits. At a local scale, studies of modern washovers have demonstrated that the occurrence of washover deposits, their spatial distributions and the characteristics of the deposits (lateral and longitudinal extension, thickness and internal organization) on a given site are highly determined by the beach, barrier and back-barrier characteristics such as offshore bathymetry, pre-storm beach profile, pre-existing morphologies of the barrier and back-barrier domains and vegetation cover (as extensively described in Morton, 2002; Morton and Sallenger, 2003; Donnelly et al., 2006; Wang and Horwitz, 2007; Phantuwongraj et al., 2013; Williams, 2015). It is very unlikely that a given barrier would have kept a constant configuration throughout the Holocene. The worldwide reorganization of barriers following RSL changes, variations in sediment availability, sources and dynamics, have undoubtedly caused an evolution of the sensitivity of barrier systems to overwash processes, thus most probably inducing biases in the storm records reconstructed from such features (Buynevich and Fitzgerald, 2003; Sabatier et al., 2012).

5.4.2. Limitations of the use of stratigraphical and sedimentological markers for the identification of overwash events

A major question when using stratigraphical and sedimentological indicators to infer washover deposition concerns the ability to unambiguously relate these layers to overwash processes, as opposed to non-storm processes (Collins et al., 1999; Liu and Fearn, 1993; Hippensteel, 2008). This was especially pointed out as a potential problem in settings where the deposition of sand laminae upon the back-barrier marshes could be the result of processes different from storm-related overwash (Woodruff et al., 2008; Otvos, 2011) such as (i) important freshwater discharges to the system or terrestrial runoff during heavy rain episodes, (ii) incorporation and deposition of sandy material by high-tide and/or (iii) potential remobilization of intra-back barrier sand bodies and redistribution across the marshes (e.g. Otvos, 2000, 2011). The opening and functioning of tidal inlets during barrier histories and the temporary character of these (both in time and location) and the reworking of previously-emplaced sand bodies have especially been considered problematic (Otvos, 2000, 2011). Aeolian transport of sands from the dune fields and aeolian reworking and redistribution of washover sheets have been shown as important processes to be considered in back-barrier sedimentation. Indeed, aeolian deposition of sand across the back-barrier marshes may blur the time-signal of washover deposition as it has been observed several hundreds of meters landwards of the barrier over the back-barrier domain (Anderson and Walker, 2006; Rodriguez et al., 2013) or even further back (Björck and Clemmensen, 2004; De Jong et al., 2006) and, even if generally minor in volume, can have fairly constant deposition rates. For example, after Foxgrover (2009) demonstrated that the high-elevation of the regressive barrier located in the eastern part of Onslow Beach (North Carolina, U.S.A) precluded any washover deposition during the last 100 yrs, Rodriguez et al. (2013) proposed that the sand-layers at the back of this barrier most probably originated from aeolian deposition. Nonetheless, as observed by Rodriguez et al. (2013), stratigraphic (graded vs. non-graded), textural (presence/absence of heavy-minerals) and grain-size characteristics generally allow to differentiate between wave-deposited and wind-deposited sand layers. In regions potentially exposed to tsunamis, a challenging aspect of paleo-storminess reconstruction is to discriminate between the deposits left by storms and those brought by tsunamis (Phantuwongraj et al., 2013). Considerable research efforts focusing on tsunamis have been conducted during the recent years and led to important methodological developments on the detection and characterization of past-tsunami deposits in Holocene coastal sedimentary records (e.g. Goff et al., 2004; Kortekaas and Dawson, 2007; Morton et al., 2007; Switzer and Jones, 2008; Chagué-Goff et al., 2011; Fruergaard et al., 2015). The distinction between storm-generated and tsunami-generated coastal deposits remains debated and we invite the reader to refer to the above-cited papers and references therein for extensive details about the possible criteria that can be used to achieve such a distinction.

It should be kept in mind that transgressive Holocene coastal barriers underwent continuous erosion and roll-over so that the oldest washover stratigraphies are prone to have been erased. Also, considering that washover deposits thin landwards, coring locations may be located in the distal end of the oldest washovers lags (or even out of reach of them) so that no clear stratigraphical and sedimentological washover testimonies may be observable for the oldest storms (e.g. Sabatier et al., 2012; Dezileau et al., 2016). Finally, bioturbation can strongly degrade washover deposits. Bioturbation has been shown to be especially active within the back-barrier marshes upon which washovers sediments are deposited. The mid-to low-marsh intertidal to subtidal zones are mostly concerned by the action of infaunal organisms which may cause a random blend of material and erase depositional structures (Sedgwick and Davis, 2003) while the supratidal high-marsh domain is affected by the bioturbative effects of fiddler crabs, which have a more easily identifiable effect on the deposits. This facies-dependent bioturbation activity makes the high-marsh domain the most suitable area for the preservation of washover sand layers (Hippensteel and Martin, 1999; references therein; Sedgwick and Davis, 2003; Hippensteel, 2008; Leorri et al., 2009). Such a spatial partitioning of the bioturbation activity between the different marsh domains should be carefully considered when evaluating the representativeness and the reliability of a storm record build upon the identification and the dating of washover layers.

5.4.3. Limitations in the use of macro- and micro-fauna fossils markers for the identification of overwash events

One of the main difficulty in the use of macrofauna to extract information about washover deposition pertains to post-depositional taphonomic process, and particularly to corrosion and dissolution phenomena which lead to poor preservation of the shells (Jelgersma et al., 1995), as well as carbonate leaching (Andrade et al., 2004). The possibility that shells contained in previously-deposited sedimentary bodies could be reworked during a storm is also a major concern. Nonetheless, severe reworking of shells will most probably lead to their fragmentation, thus allowing their differentiation from living shells uprooted from their living environments and transported landward with the overwash flows.

Building past-storminess records from foraminifera assemblages raises two main concerns, as summarized by Hippensteel and Garcia (2014): (i) the original habitat of the foraminifera specimens found within the sediments and (ii) the quality of the post-depositional conservation of the assemblages. A main concern is how to deal with the presence/absence of specimens within the sediments and how to evaluate the representativeness and reliability of these assemblages. Hippensteel and Garcia (2014) retrieved cores from three back-barrier lagoonal marshes of Southeastern North Carolina (USA) and analyzed them in search of Holocene overwash events. A prominent variability in the washover events evidenced by foraminifera analyses was observed both regionally between the three marshes as well as locally in between the cores obtained within each of the marshes. Sixteen peaks in calcareous foraminifera were for example picked from one core retrieved in the central part of the bay (Alligator Bay marsh), while the adjacent cores only showed two to five of them. A lesser variability was conversely observed for the peaks in coarse-grained sediments, which more consistently evidence for overwash events throughout the different cores. Washover layers completely devoid of foraminifera have also been reported, such as within the sand layers which the clustered hurricanes Katrina and Rita deposited along the coast of Mississippi and Alabama (USA) in 2005 (Horton et al., 2009). The absence of microfossils within the washover layers may be the product of a combination of factors ranging from ecological factors influencing the microfossil population present when the storm hit (e.g. high environmental stress or poor reproductive conditions) to post-depositional bioturbation or taphonomic processes responsible for the partial blur or even complete disappearance of the microfossils after deposition. Offshore foraminifera, mainly characterized by calcareous tests, are sensitive to dissolution within the relatively acid marsh environments so that the fossil washover assemblage can be found partly or completely destroyed with time (Andrade et al., 2004; Horton et al., 2009 and references therein). Hopefully, storm layers often may contain enough organic-poor sediments and calcareous material, so that the dissolution remains minor and calcareous foraminifera are preserved (Scott et al., 2003). Post-depositional bioturbation can induce a mixing of the salt-marsh and washover assemblages, resulting in artifact subsidiary peaks in offshore foraminifera percentages unrelated to storm events and thus in an overestimation of the number of storm-events (Hippensteel and Martin, 1999; Hippensteel et al., 2005). As underlined by Leorri et al. (2009), bioturbation has received little attention with respect to the importance it bears for salt-marsh microfauna-based reconstructions. This phenomenon seems to have only a noticeable effect on foraminifera assemblages of the lower marsh domains (Hippensteel and Martin, 1999; references therein, Hippensteel et al., 2005; Leorri et al., 2009) and as such high-marsh sedimentary sequences may constitute better-preserved archives of washover deposition. Nonetheless, transgressive/regressive shifts in saltmarsh sedimentary environments during the Holocene can have induced, for a given location, some increase/decrease of bioturbation activity which could in turn misleadingly suggest respectively a decrease or an increase in washover deposition and, consequently, in storminess activity.

A further limitation of foraminifera proxies in paleo-storm records is linked to uncertainties in the sources of the foraminifera specimens. Back-barrier sediments may be concerned by inputs of offshore foraminifera unrelated to storm events or, at least, not directly related to overwash processes. The opening of tidal inlets can, for example, also cause offshore foraminifera to enter the back-barrier tidal creeks at flood-tides. Changes in the back-barrier internal morphodynamics, such as channel migration, elevated water levels and/or internal storm waves within large back-barriers lagoons can induce passive displacement of offshore specimens (Collins et al., 1999; Hippensteel et al., 2005; Schafer and Medioli, 2009). Wind can also transport offshore foraminifera into the back-barrier area. Such processes are commonly active along carbonate shorelines where extensive dune accumulations (carbonate aeolianites) typically develop (Frébourg et al., 2008). In such setting it may be difficult to differentiate between wind-deposited and storm-wave deposited strata only based on paleontological evidence.

5.4.4. Limitations in the use of elemental and organic geochemical markers for the identification of overwash events

Using major and trace elements as tracers of past storminess within back-barrier sedimentary sequences requires that some offshore sedimentary sediment has been brought to the back-barrier domain. As discussed by Sabatier et al. (2012), the grain-size of the material brought by waves in the system will determine, at least partly, the geochemical signature of the overwashes which can be identified. Indeed, changes in trace elements ratios related to inputs of offshore material result from the transport and deposition of silts and sands carrying quartz and heavy minerals. These inputs thus require waves of sufficient energy for such material to be transported to the back-barrier area. When obtaining data from XRF core scanner (Croudace et al., 2006), several problems may arise especially linked to down-core changes in the organic matter content, variations in the water content or grain-size variability (e.g. Tjallingii et al., 2007; Löwemark et al., 2011; Chawchai et al., 2015; and references therein). These problems are amplified in sedimentary sequences characterized by very variable lithologies such as those that are most of the time retrieved from coastal sedimentary systems. Caution must hence be adopted when working with elemental data obtained from XRF core scanner and procedures of normalization and standardization XRF-scanner data is to be performed before these latter can be trusted for any paleoenvironmental reconstruction (Löwemark et al., 2011).

When using OGPs as paleo-environmental indicators, a significant concern is the difficulty to estimate whether these indicators are correctly preserved within Holocene sediments. Indeed, the question is still open on whether the organic components that make up for $\delta^{13}C$, $\delta^{15}N$, TOC and TN undergo significant alteration both during their sinking down the water column and the early/late diagenesis within the sediment (Wilson et al., 2005a,b; Lamb et al., 2006; Goslin et al., 2017). Particularly, C/N ratios potentially undergo an artificial increase during diagenesis due to nitrogen being preferably and more rapidly lost than carbon following microbial degradation. The OM embedded in the sediment is thus likely to yield values of the C/N ratio higher than the "fresh" OM deposited at the surface (Wilson et al., 2005a,b; Goslin et al., 2017) and may lead to erroneous interpretations of the organic matter sources. Nonetheless, as underlined by Lehmann et al. (2002) and by Lambert et al. (2008), OM is significantly more resistant to alteration within anoxic conditions (such as those that can be found within back-barrier lakes) than in open-oxic conditions, and thus proposed that clear trends in OGPs variations could draw reliable tendencies in anoxic environments. Lambert et al. (2008) put forward several arguments in favor of the good state of conservation of OGPs within the sequences they studied. The most important arguments may be that (i) quasi-permanent anoxic conditions were observed at the lake bottom; (ii) no clear trend could be observed with increasing depth along the core, that would evidence an artificial increase of the C/N ratio and (iii) the shift in the depositional environment from a lagoon/estuary to a closed lake observed in the lithology was also clearly evident also in the OGPs.

6. Discussion and future research prospects

This review has presented the multidisciplinary pool of proxies that the paleo-tempestology/coastal science community has to extract past-storminess records from barrier and back-barrier sedimentary successions. Along with "traditional" stratigraphical, sedimentary or micro-faunal proxies, promising innovative proxies of storm occurrence such as the "organic geochemistry proxies" open new possibilities for extracting discrete storm records from barrier and back-barrier sedimentary systems. However, all proxies of storm occurrence provided by wave-induced erosional or depositional markers carry drawbacks and limitations (Table 1). Apart from the limits specific to each proxy, many of the limitations are the consequences of the inherent highly variable and dynamic context of the coastal barrier systems, which induces much uncertainty in how a given storm would impact the site and in which manner these impacts are adequately recorded and subsequently preserved. The study of past storminess would thus gain from being part of integrated paleo-environmental and paleo-geographic reconstructions carried out at a local to regional-scale and from working across disciplines. In this context and when hoping to derive recurrence intervals of past storm events, a second challenge the community faces is to obtain precise and reliable chronologies of deposition. The following sections discusses some research paths which appear to be promising towards (i) the obtaining of more reliable and better defined Holocene storm chronologies and (ii) their use in the understanding of regional to extra-regional atmospheric patterns.

6.1. Taking advantage of the pool of proxies

A single past-storminess proxy that behaves independently to local morphodynamics and with good preservation potential (what Hippensteel, 2010; coined as the "perfect paleo-storm proxy") has not yet been identified for barrier and back-barrier environments. However, making a use of several proxies from the large set of proxies we described across this review appears promising. Beach-ridges systems stand out as the proxy with the highest preservation potential when the studied system has long been prograding over the last millennia, preferably at high and constant rates. Also, beach ridges also offer the possibility of obtaining simultaneous data on both storminess and RSL changes, thus opening toward studies fully integrating these two agents. Yet, it is still unclear whether beach ridges primarily are built on an event-scale or are the results of several events. Also, site-scale sediment budgets play a major role in the building of beach ridges, and knowing how this drives progradation rates, and in turn may lead to the potential misidentification of storm quiescent periods is still incompletely understood. Storm scarps require very specific conditions to be adequately recorded and preserved in the barrier systems. Only systems characterized by fast and constant progradation rates are likely to deliver representative data about past storminess from storm scarp markers. Washover deposits clearly offer the advantage over the other proxies to be a widespread feature and to be identifiable by a large set of various indicators which offers the possibility of identifying paleo storm events from a wide array of site configurations. Nonetheless, a drawback of washover deposits is that their occurrence depends on a complex assemblage of parameters that have surely varied throughout the Holocene. Also, thin overwash deposits have low preservation potential and may be difficult to date. "Organic geochemistry proxies" have shown promising results for the identification of discreet overwash events, opening new paths for the identification of either smaller storm events or when the study site was in marginal position relative to the storm-track of maximum water level. As such, these latter deserve more widespread exploration.

A robust approach to increase the reliability of Holocene storm chronologies would be to use two independent sets of proxies eventually complementing each other. A first set of proxies would include those directly linked to storm-induced wave markers, as can be observed locally in each barrier system and extensively reviewed in the present article. A second set would include, at a regional scale, proxies unrelated, or not directly related, to the wave-related processes treated here. In this second set, proxies of storm-induced activity related to aeolian processes (e.g. Björck and Clemmensen, 2004; Clemmensen et al., 2009; Orme et al., 2016), would form an ideal supplementary data base for the reconstruction of past storminess, along with data obtained from proxies such as speleothems (e.g. Frappier et al., 2007; Nott et al., 2009), tree ring and lake-varve records (e.g Dean, 1997) providing additional paleoclimate constraints.

6.2. Evaluating the temporal and spatial significance of storm events in the context of complex local systems

In the previous sections, the complex suite of interdependent local parameters which control the changes a barrier system could undergo during and following a storm event was considered. References to various studies, most of which are recent, enabled an

Table 1
Summary of the characteristics of the proxies of past storm-wave activity reviewed in the manuscript along with key references.

Proxy	Strength	Limitations	Methods of age control	Key case studies & references
Regressive - progradational systems				
Beach-ridges (sand & mixed sand-gravel)	Direct relationship of the feature with water-level elevation	-Sensible to coastline progradation rates -Representativeness of beach-ridges: Single-storm or several-storms? -Post-depositional alteration of the beach-ridges (e.g. aeolian processes, subsequent erosion) -Contain long-term changes in water level (e.g. RSL) -Highly sensible to non-storm processes (e.g. sediment supply)	-Age models based on benchmark dating or dating of individual beach-ridges -OSL dating of material sampled from inside the ridge -Best using mixed OSL/^{14}C dating	Otvos (2000); Nott et al. (2009); Forsyth et al. (2010); Tamura (2012); Nott et al. (2015); Bendixen et al. (2013); Clemmensen et al. (2016)
Storm scarps	Easy to map features. Straightforward relationship with extreme wave events. May record smaller events	-Require precise water-level conditions - Require fast post-storm recovery infilling - Preservation potential very sensible to progradation rates of the system and erosion by subsequent events. - Recognition sensible to the presence of heavy minerals.	- OSL dating of bottom surface of post-storm infilling or surface directly underlying the scarp - 14C dating of shells embedded in the post-storm recovery infilling	Gontz et al. (2014); Dougherty et al. (2004) Dougherty (2014) Buynevich et al. (2007) Buynevich et al. (2004)
Transgressive - receding systems				
Markers of overwash events	Wide spread features. Unambiguous relationship with elevated water-levels	- Strongly dependent on water level vs local topography. - Overall high spatial variability	- Radiocarbon or radionuclide dating of pre-storm/post-storm organic back-barrier sediments - OSL dating of washover sand layers	Morton and Sallenger (2003); Phantuwongraj et al. (2013) Sedgwick and Davis (2003)
Stratigraphic evidences	Discreet markers of washover deposition	- May only represent largest events. - Prone to post-depositional reworking		Liu and Fearn (2000); Donnelly et al. (2001b); Cunningham et al. (2011); Van-Vliet Lanoë et al. (2014)
Grain-size anomalies	Recognition of smaller events and/or from landward located positions. Evaluation of the water-flow depth	- Prone to post-depositional reworking - Potential by massive aeolian sand inputs		Woodruff et al. (2008); Boldt et al. (2010); Brandon et al. (2014); Dezileau et al. (2016); Ercolani et al. (2015); Lane et al. (2011)
Mineralogical evidences	Recognition of smaller events and/or from landward located positions. Proof-checking of sediment sources	- Sediment sources must be characterized by well distinct mineralogies. - Potential blurring by massive aeolian sand inputs.		Sabatier et al. (2010a,b); Sabatier et al. (2012)
Macrofauna	Evaluation of storm wave conditions. Proof-checking of sediment sources	- Prone to post-depositional taphonomic alterations. - High spatial variability. - Potential post-depositional redistribution		Sabatier et al. (2008); Engel et al. (2015)
Microfauna	Evaluation of storm wave conditions. Proof-checking of sediment sources	- Prone to post-depositional taphonomic alterations. - High spatial variability. - Potential post-depositional redistribution		Pilarczyk et al. (2014); Hippensteel and Martin (1999); Zong and Tooley (1999); Pilarczyk and Reinhardt (2012)
Major and trace elements	Efficient tracers of sediment sources	- Best if barrier and lagoonal areas are concerned with inputs of well-distinct sediment sources		Woodruff et al. (2009); Sabatier et al. (2010a,b, 2012); Degeai et al. (2015); Dezileau et al. (2016)
Organic geochemistry	Allow high-resolution reconstructions. Do not require inputs of offshore material to the back-barrier area (if freshwater lake)	- Prone to post-depositional alteration of the isotopic and organic compounds		Lambert et al. (2008); Das et al. (2013)

insight into the different approaches which can be used to build long-term chronologies of past-storminess activity from observations gathered on a barrier and back-barrier system.

To the first order, the impacts of a storm event on a barrier system appear mostly determined by the relationships between the storm water level and the pre-storm topography of the barrier. When trying to build long-term chronologies of past-storminess activity, it should be kept in mind that all parameters driving both these variables may have varied considerably through time, thus resulting in equally considerable changes in the sensitivity of barrier systems to storms. Barrier systems are dynamic systems whose behavior adapts to autogenic and allogenic processes. As a consequence, obtaining storm chronologies records from the study of barriers and back-barriers systems is often made complex by over-printing phenomena inherent to the dynamic character of these particular sedimentary systems. Moreover, as can be appreciated by studying modern storms, storm trajectories undergo large temporal and spatial shifts that are still imperfectly understood. The storminess history deduced from a given site can therefore only be given a local significance, also underlined by most of the case studies considered here, and it is clear that event-scale past-storminess chronologies cannot be resolved without addressing the importance of local fingerprinting. To do so requires working across disciplines and integrating the full range of parameters that may have driven the local morphology of the sites from which the sedimentary records are retrieved (for example, a reliable regional RSL history and an understanding of the morpho-sedimentary dynamics of the sites and of how these evolved during the Holocene). However, observations made at a regional scale (i.e. by integrating several local-scale studies) are also needed to avoid possible artefacts caused by local processes. In this sense, paleotempestology faces the trade-off between local and regional study scales, a problem somewhat equivalent to problems encountered by relative sea-level reconstructions.

If the aim of building exhaustive and broad regional past-storm records is considered unrealistic, then aggregating data obtained from several sites picked at a local scale for the various morpho-sedimentary configurations they offer should be preferred to define a regional signal of past-storminess. Where and when possible, working at a local scale on multiples sites showing equivalent exposures towards storm impacts but characterized by differing morphologies may provide complementary records of past storminess from which variability linked to site-scale morpho-sedimentary parameters could potentially be identified and filtered. Resolving the trade-off between storminess and the forcing factors responsible for site-scale morphodynamics (sediment variability, relative sea level) is crucial to move forward the evaluation of the quality of the past-storm records reconstructed from barrier and back-barrier sedimentary archives. We suggest that storm records should preferably be sought in sites for which the best possible knowledge is available on the regional Holocene RSL history and the stratigraphic emplacement and evolution of the sedimentary wedge.

6.3. Building reliable chronologies

Increasing the accuracy of the dating of storm events is a keystone if recurrence intervals are to be derived. Developments in radiocarbon AMS dating mean less and less material is required, opening the possibility of dating of thinner sedimentary layers and thus of reducing uncertainties in the obtained ages. Nonetheless, the use of ^{14}C dating remains complicated by several potential biases such as the pollution by old or recent carbon, or reservoir effects in enclosed basins and material used for dating must be chosen very carefully in a rigorously known context. In recent years, OSL has developed considerably and now offers the possibility of dating sand layers with much reduced age uncertainty. Single-grain OSL dating opens the potential to date small quantities of sand material, but the origin of the grain must be ascertained to ensure that the sample has been exposed to light long enough to be "re-zeroed". Improving the reliability of storm chronologies may lie in the joint-use of ^{14}C and OSL dating techniques to balance the source of uncertainties inherent to each method, combined with multiple dates made on single layers and narrowing of the age uncertainties through the use of Bayesian statistical methods.

6.4. Working at event or climate-scale?

Building storm-chronologies and evaluating future extreme-events is also closely dependent on the temporal scales that are considered relevant. With the advent of more and more precise dating methods, the community has moved towards reconstructions of past storminess on an event (weather) scale, calculations of return periods and has tried to establish controlling links between large-scale forcing agents and past-storm events (e.g., Donnelly and Woodruff, 2007; Wallace and Anderson, 2010; Van Vliet Lanoë et al., 2014; Degeai et al., 2015). Such goals are appealing as they would help decision-makers building sustainable coastal management scenarios. Nevertheless, considering the serious limitations weighting on the many proxies that we synthetized in this paper, being able to attribute robust return periods to past storm-events and eventually forecast future storms of a certain size and strength still seems quite out of reach, due to the extreme complexity of the numerous factors interplaying at the event scale. We consider that, in many settings, it may be more realistic to progress with reconstructing climate-states of past storminess, *i.e.* by identifying wider periods of decreasing or increasing storm frequency in the records, detrended from the records available for one or several forcing agents (such as relative sea-level evolution).

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