revue en ligne : http://geomorphologie.revues.org/

Géomorphologie : Relief, Processus, Environnement

vol. 23, n° 1 (2017), p. 37-50
Mise en ligne le 13 mars 2017 sur Revues.org
URL : http://geomorphologie.revues.org/11638
DOI : 10.4000/geomorphologie.11.638

http://edytem.univ-savoie.fr/gfg/

Development of an incipient foredune field along a prograding macrotidal shoreline, northern France

Formation d'un champ de dunes embryonnaires le long d'une côte macrotidale en accrétion, nord de la France

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ABSTRACT

In this paper, we examine the development of an extensive hummocky incipient foredune field of approximately 100,000 m² that developed along a short stretch of coastline on the macrotidal coast of northern France. Analysis of historical maps and documents, of a series of aerial photographs as well as recent LiDAR and topographic data allowed us to depict the formation and evolution of this incipient foredune field. Our results show that this coastline stretch is prograding seaward since at least the 16th century, this shoreline advance being related to the onshore welding of subtidal sand banks. During the early 20th century, isolated sand islets, surrounded by the sea at high tide, formed at the landward edge of a wide convex-shape sand flat. They progressively merged to the main shoreline during the mid-20th century, inducing a shoreline advance of up to 300 m between 1938 and 2012. Once merged to the main shoreline in the 1960s, a field of incipient foredunes, facing the dominant winds began to develop, inducing a shoreline progradation at a rate of 5 m.y^{-1} that resulted in a seaward widening of the hummocky dunefield. Although the field of incipient foredunes continued to expand since its initiation, the morphology of individual mounds, colonized by a sparse vegetation cover, tended to remain remarkably stable through time. In this area of high sediment supply, the formation of a hummocky dunefield is probably due to the fact that the incipient foredune do not have enough time to coalesce and to merge into a continuous foredune ridge as they rapidly become disconnected from their main sediment source.

Keywords: costal dunes, incipient foredunes, shoreline change, northern France.

RÉSUMÉ

Dans cet article, nous examinons la formation d'un vaste champ de dunes embryonnaires d'une superficie d'environ 100 000 m² qui s'est développé sur une section limitée du littoral macrotidal du nord de la France. L'utilisation de cartes et de documents historiques, de photographies aériennes ainsi que de récentes données LiDAR et topographiques a permis d'analyser la formation et l'évolution de ce champ de dunes embryonnaires. Nos résultats montrent que ce secteur côtier progresse vers la mer depuis au moins le XVIe siècle, cette avancée du littoral étant liée à l'accolement de bancs sableux sub-tidaux. Au début du XXe siècle, des îlots sableux, entourés par la mer à marée haute, se sont formés en haut de plage, au niveau d'un large replat sableux de forme convexe. Ils se sont progressivement soudés au rivage au milieu du XXe siècle, provoquant une avancée de la ligne de rivage de 300 m entre 1938 et 2012. Un champ de dunes embryonnaires, faisant face aux vents dominants, a commencé à se développer dans les années 60, entraînant une progradation du rivage à un rythme de 5 m/an ce qui s'est soldé par un élargissement vers le large de ce champ de dunes chaotiques. Bien que le champ de dunes embryonnaires ait continué à se développer depuis sa formation, la morphologie des monticules individuels, colonisés par une couverture végétale clairsemée, est restée remarquablement stable. Dans ce secteur caractérisé par d'abondants apports sableux, la formation d'un champ de dunes chaotiques est probablement due au fait que les dunes embryonnaires n'ont pas eu assez de temps pour coalescer et former une dune bordière continue du fait qu'elles aient été rapidement privées d'apports sédimentaires.

***Mots-clés :** dunes côtières, dunes embryonnaires, évolution de la ligne de rivage, nord de la France.*

INFORMATION SUR L'ARTICLE
Reçu le 01 décembre 2016.
Reçu sous sa forme révisée le 10 février 2017.
Définitivement accepté le 14 février 2017.

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1. Introduction

Most sandy beaches are bordered landward by some form of coastal dune built up by the onshore transport of beach sand by wind (Martinez and Psuty, 2004; Davidson-Arnott, 2010; Hesp and Walker, 2013). Depending on climatic, biological, geomorphological and/or oceanographic conditions such as dominant wind velocity and direction, precipitation, vegetation cover, sediment supply, grain-size of beach sediment, beach morphology and tidal range, dunes of various size and morphology are found worldwide (Carter, 1988; Hesp, 2011; Houser and Ellis, 2013; Ruz and Hesp, 2014). However, the most common coastal dunes that develop immediately landward of the active beach are foredunes, which are sometimes considered as the only distinctive coastal dunes (Bauer and Sherman, 1999). They usually form shore-parallel, convex, symmetrical to asymmetrical dunes developed in the upper-beach zone (Hesp, 2012). The morphology of foredunes is varied, but they can be classified into three main types: incipient foredunes, established foredunes, and foredune plains (Short and Hesp, 1982; Hesp, 2002).

Incipient foredunes, also called embryo dunes (Davidson-Arnott, 2010), are low dunes forming by aeolian sand deposition within pioneer plant communities on the backshore of beaches (Hesp, 2002, 2012). Their formation is due to an increase in surface roughness due to the presence of some elements at the surface of the backshore responsible for a reduction of wind flow velocities, resulting in sediment deposition. Vegetation is the most common roughness element that contributes to the formation of incipient foredunes. In their early stage of development, incipient foredunes may form mounds and nebkhas, often associated with shadow dunes (Hesp, 1981). They usually develop on the backshore on swash-aligned drift material or on storm debris lines, through alongshore growth of pioneer plant seedlings and/or by rhizome growth, forming a quasi-continuous line of low (50-100 cm high) vegetated mounds (Hesp, 1989).

Incipient foredunes are sometimes ephemeral features that can be partly eroded or completely destructed by waves during storms. However, they may also survive and grow to become an established foredune if they form sufficiently high above the high spring tide limit (Hesp, 1999, 2002). The morphology and size of foredunes reflect the short to long term surfzone-beach-dune processes operating on any particular beach (Hesp, 1988), and the largest foredunes are usually found on dissipative beaches associated with high wave-driven sediment supply and where wide and low gradient foreshores favour maximum aeolian fetch and high aeolian sand transport (Short and Hesp, 1982; Davidson-Arnott, 2010; Hesp, 2012). Because macrotidal sandy beaches are characterized by a wide exposed foreshore at low tide, they are commonly thought to represent optimal conditions for coastal dune development (King, 1972; Carter, 1988). There are nevertheless significant differences in aeolian dune development along low-gradient macrotidal beaches (Anthony et al., 2006, 2009) that can be fringed by massive coastal foredunes, but also by much smaller aeolian landforms (Battiau-Queney et al., 2001).

In this paper we examine the development of an extensive hummocky incipient foredune field that developed along a short stretch of coastline bordering a prograding macrotidal foreshore on the coast of northern France (Héquette and Aernouts, 2010; Anthony, 2013). The analysis of the evolution of this incipient foredune field at decadal and annual time scales using vertical aerial photographs, airborne LiDAR topographic data and field measurements allowed us to depict the formation and morphological evolution of these aeolian landforms and to get some insights into the mechanisms responsible for continuous incipient foredune development and shoreline progradation. This study was complemented by the analysis of historic nearshore and shoreline changes using marine charts and historical documents. This information enabled us to identify the conditions that lead to the development of an incipient foredune field instead of a continuous, shore-parallel, established foredune ridge that more commonly occurs at the landward edge of a beach.

2. Study area

The French North Sea coastline is a 55 km long almost continuous sand beach barrier facing the eastern English Channel and the Southern Bight of the North Sea (fig. 1). The coast mainly consists of wide, gently sloping, barred sandy beaches (Anthony et al., 2005; Reichmüth and Anthony, 2007) backed by coastal dunes that generally do not exceed 15 m high (Battiau-Queney et al., 2001; Ruz et al., 2005). The dunes border the French coastal plain which extends 10 to 20 km landward. The coastal plain is a low-lying reclaimed area (mean elevation of 2 m above mean sea level) and coastal dunes and dikes are the only protection against marine flooding.

The study area consists of a 350 to 500 m wide beach at low tide, east of Calais (fig. 1), characterized by a series of intertidal bars and an upper beach bounded landward by coastal dunes. The coastal dunes fringe a salt marsh that has been largely modified by man, with the excavation of numerous hunting ponds (fig. 1). Eastward, the intertidal zone forms a protruding and extensive sandflat. To the west, the upper beach consists of a 250 to 400 m wide and 1,500 m long sand platform extending to the port of Calais. The foreshore and upper beach platform exhibit large variability in surface sediment size, depending on morphology, elevation and exposure to waves. The intertidal bars and troughs are composed of well-sorted fine to medium sand (mean grain size: 0.2 to 0.33 mm), while the

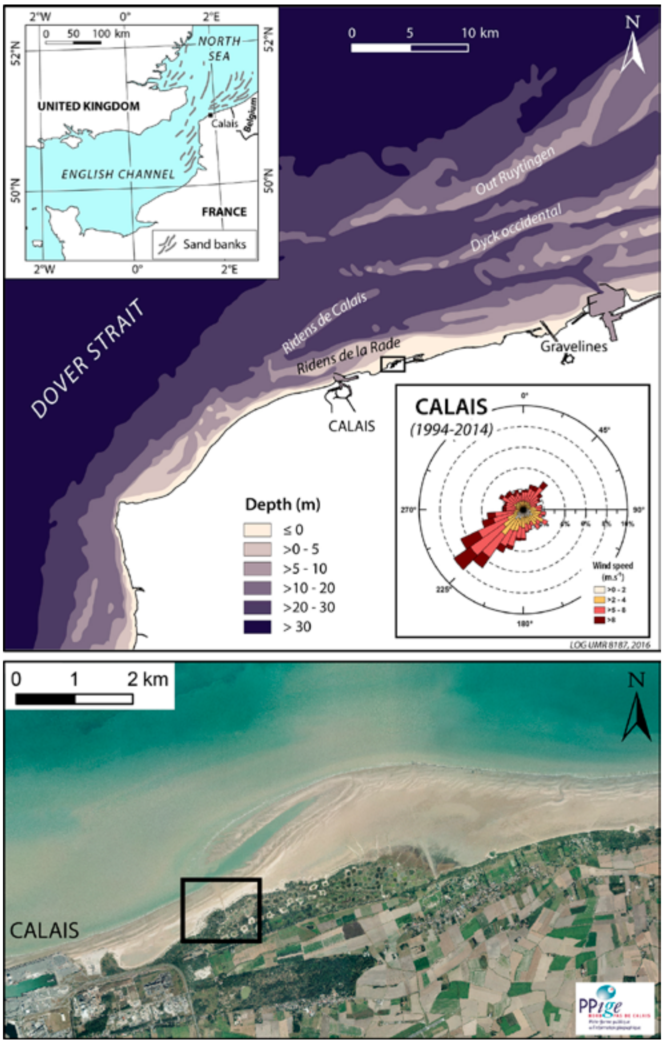

Fig. 1 – Location map of the study area.

The inset on the aerial photograph shows the location of the investigated coastal dunes. Wind rose diagram based on three-hourly mean wind speed and direction from the Météo-France station in Calais.

Fig. 1 – Localisation de la zone d'étude.

Le rectangle sur la photographie aérienne montre la localisation des dunes étudiées. La rose des vents est basée sur les données tri-horaires des vitesses moyennes et directions du vent enregistrées à la station de Météo-France de Calais.

upper beach platform mainly consists of fine sand (mean grain size: 0.23 mm). The coastal dunes in the backshore are composed of fine to medium sand (mean grain size: 0.17 to 0.31 mm), similar in size to the sand found in the intertidal bars (Anthony et al., 2006).

Offshore, tidal sand banks are widespread across the nearshore zone and the inner shelf where they form extensive linear sand bodies sub-parallel to the shoreline (Augris et al., 1990; Beck et al., 1991). Some of these shallow sand banks are actively migrating such as the Ridens de la Rade, a 13 km long and 1.4 km wide bank near Calais (fig. 1), that migrated alongshore and onshore during the 20th century and became eventually attached to the shore (Augris and Clabaut, 2001). Comparison of bathymetric charts revealed that the volume of this shore-attached sand bank increased by about 100×10^6 m^3 in the course of the 20th century (Héquette and Aernouts, 2010), forming an extensive sub-tidal sand source for the development of the intertidal sandflat and coastal dunes in the backshore (Anthony, 2013; Anthony et al., 2006), therefore favouring shoreline progradation.

The coast is dominantly exposed to shore-parallel winds from the southwest, followed by winds from north to northeast directed obliquely onshore (fig. 1). Winds are usually moderate, with more than 45% of winds with a mean velocity of less than 5 m.s^{-1}; strongest winds ($\geq$ 16 m.s^{-1}) occur less than 1% of the time. Associated with these winds, the offshore wave regime is dominated by waves from southwest to west, originating from the English Channel, followed by waves from the northeast to north, generated in the North Sea (Anthony, 2013). The coast is exposed to short-fetch, relatively low-energy waves punctuated by storm activity. Most waves have a significant height of less than 1 m and periods ranging from 4 to 8 s, but wave height may episodically exceed 5 m with periods of 9 to 10 s during major storms (http://candhis.cetmef.developpement-durable.gouv.fr/campagne/). Wave heights are much lower at the coast, however, due to significant refraction and shoaling over the offshore sand banks, resulting in modal inshore wave heights less than 0.5 m high (Sedrati and Anthony, 2007; Héquette et al., 2009).

The coast is also characterized by a large tidal range. The tidal regime in the region is semi-diurnal and macrotidal, the tidal range increasing from the North Sea to the English Channel, with a range of 6.5 m for spring tides at Calais. Due to the large tidal amplitude, tidal currents are strong along the northern coast of France, reaching maximum near-surface speeds of 1.5 m.s^{-1} during flood tide and 1.35 m.s^{-1} during ebb in the narrow interbank channels (Augris et al., 1990). Tidal currents are alternating in the coastal zone, flowing almost parallel to the coastline, flood currents being oriented towards the east-northeast and ebb currents towards the west–southwest. Because of the flood-dominated asymmetry of the tidal currents towards the northeast and of the dominant waves from the southwest, net sediment transport in the coastal zone is directed to the east-northeast (Héquette et al., 2008).

3. Methodology

Historical evolution of the coastal zone was analysed using historical maps, marine charts and from scientific papers published during the 19th and early 20th century in the journal "*Annales de la Société Géologique du Nord*", available online at the French National Library (http://gallica.bnf.fr/). Theses of geography published during the early 20th century (Blanchard, 1906; Briquet, 1930) provided highly valuable information about past shoreline geomorphology and evolution. We also used an Ordonnance Survey map (called "*carte d'Etat-Major*" in France) of 1835 at a scale of 1:40 000, georeferenced by the French National Institute of Geographic and Forestry Information (IGN). Detailed cadastral plans of 1820-1826 were also consulted on line (http://www.archivespasdecalais.fr/Archives-en-ligne/Plans-cadastraux).

Shoreline change since the 16th century was assessed based on the location of the low tide limit identified on historical maps and on the position of the French hydrographic datum levelled in 1902 on a marine chart published in 1922 and on 1977 hydrographic field sheets of the French Hydrographic Survey (*Service Hydrographique et Océanographique de la Marine* - SHOM). Based on previous investigations using the same data set, the error margin of the 1977 bathymetry was estimated at ± 0.6 m, which includes the error range of the instrument (echosounder) as well as errors due to inaccuracy in ship positioning and to tidal correction (Héquette and Aernouts, 2010). The error range in 1902 was estimated at ± 1.0 m, due to less precise depth sounding technique (line sounding) and less accurate ship positioning. Because the French hydrographic datum corresponds to an elevation slightly below the lowest astronomical tide level, its position can be considered as an approximate indicator of the lower tide limit on the hydrographic charts.

The analysis of the evolution of coastal dunes was based on the comparison of vertical aerial photographs orthorectified by IGN (years 1963, 1983, 2000, 2005, 2009, 2012 and 2015) and georeferenced and rectified aerial photographs using ArcGis software (years 1938, 1949, 1957, 1961, 1972, 1977, 1986, 1989, 1993, 1997), which were used for carrying out detailed geomorphological evolution of coastal dunes from stereo-pairs of aerial photographs. The error margin of the orthorectified aerial photograph is ± 2 m.

High-resolution topographic data of the beach and coastal dunes were obtained from airborne LiDAR surveys carried out in 2011 (03/21/2011) and 2014 (01/18/2014). The data were obtained using a Leica ALS60 LiDAR system that acquired topographic data with a density of 1.2 to 1.4 points per m². The LiDAR operating system was coupled with real time kinematic DGPS and inertial motion unit, ensuring a planimetric position accuracy lower than 0.5 m. The planimetric position accuracy of the data points during these two surveys ranged from ± 0.10 to 0.17 m with a vertical accuracy < ± 0.10 m as verified by several ground control points using a very high resolution Differential Global Positioning System (DGPS; see below). These vertical error ranges can easily increases to ± 0.25 m or more in areas covered by dense vegetation however (Saye et al., 2005). The LiDAR topographic data were filtered to remove vegetation, buildings and other objects. Filtered data were then used to create contour Digital Terrain Models (DTM) using Golden Software Surfer™ and calculate volume changes between each LiDAR survey. The DTMs were obtained by linear interpolation using a Delaunay triangulation resulting in a grid with a 1 m resolution, a grid cell resolution of 1 m² appearing to provide reliable representation of topography and accurate volumetric measurements in coastal dunes using LiDAR data (Woolard and Colby, 2002).

The airborne LiDAR surveys were complemented by ground topographic surveys using a differential GPS (Leica TPS Syst1200) with vertical and horizontal accuracy of ± 2.5 cm and ± 1.5 cm respectively. Cross-shore profiles were surveyed across the coastal dunes and the foreshore down to the low tide level of the day on 09/28/2011 and 04/16/2014. All the topographic measurements, including the LiDAR topographic data, were referenced to the French vertical datum (IGN 69).

4. Results

4.1 Historical shoreline changes

Between Calais and Gravelines (fig. 1), a series of low sandy ridges are found inland, parallel to the present day shoreline (fig. 2). These successive ridges, some capped by low dunes, now located 0.5 to 2 km inland have been identified as former shorelines developed since at least the 12th century (Houthuys et al., 1993; Sommé, 1975), testifying of a continuous shoreline progradation over centuries. Dubois (1926) and Briquet (1930) analysed historical data and depicted a detailed description of the evolution of this coastal area. Based on historical records (old topographic and hydrographic maps), it seems that this area experienced shoreline progradation since at least the Middle Ages.

According to Briquet (1930), shoreline progradation occurred since at least the 14th century (Briquet, 1930). To the East, between Calais and Waldam, a former shoreline (fig. 2A), is presumably the shoreline of the 15th or 16th century (Briquet, 1930). Then, as the shoreline prograded seaward, dikes were erected in order to gain new farm lands. According to Dumas-Vence (1869), the Robelin dike (now named the Royale dike) was constructed in 1630, and the Taaf dike in 1773 (fig. 2A). This dike appeared as the shoreline limit (limit with the sea) on a detailed cadastral plan of 1820-1826. Then, seaward of this dike, coastal dunes developed, isolating a saltmarsh. A line of new dunes, detached from the shore is shown by Briquet on his map of the shoreline at that time (ca. 1920-1925) (fig. 2A). Shoreline progradation was not uniform along this coastal sector. A comparison of the shoreline on the early 19th century topographic map and the present day shoreline (fig. 2B) shows that the coastline advanced seaward immediately east of Calais harbor, resulting in a protruding shoreline, while eastward, the coastline is fairly stable since the 19th century. This coastline evolution is associated with the migration of subtidal sand banks that welded to the shore.

During the 14th and 15th century, immediately east of Calais, a sand bank, called "*Het Nieuland*" or "*banc Braseux*" (fig. 3A), parallel to the shoreline, was emerged at low tide and was separated from the shoreline by a channel (Dumas-Vence, 1869). According to Dumas-Vence (1869), the channel was progressively infilled, presumably in response to the construction of a jetty to the west by the English in 1444. During the 16th century, the jetty was extended and the channel almost disappeared, due to continuing sand infilling, leading to the merging of this sand bank to the shore (Briquet, 1930), which resulted in the formation of a convex-shaped sand platform protruding seaward (fig. 3A). On its seaward limit, a screw-pile lighthouse, the Walde lighthouse, called "*feu de Waldam*" by Briquet (1930), was constructed in 1856. The foreshore, at that time was 1,500 m wide at low tide and the upper beach was extremely flat (Briquet, 1930). During the 19th century, another sand bank (Ridens de la Rade) has developed close to the shore seaward of Calais and

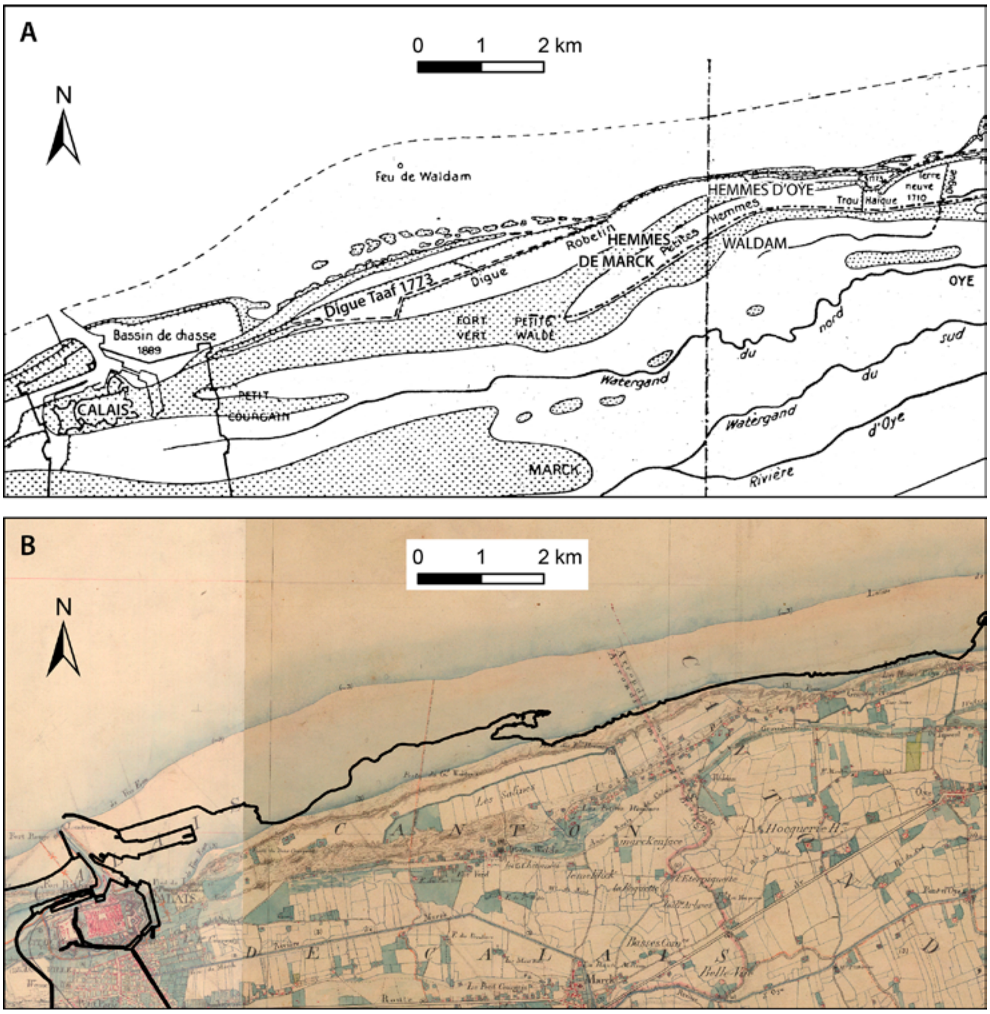

Fig. 2 – Historic maps of the coastal area east of Calais.

A. Schematic map of the distribution of coastal dunes (dotted grey areas) in the early 20th century (from Briquet, 1930); the approximate position of the low tide limit (dashed line) is also shown. B. Ordonnance Survey map of 1835 showing the configuration of the coastal zone in the early 19th century overlain by the 2005 shoreline extracted from aerial photographs (solid black line).

Fig. 2 – Cartes historiques de la zone côtière à l'est de Calais.

A. Carte schématique de la répartition des dunes côtières (surfaces en pointillés gris) au début du XXe siècle (Briquet, 1930) ; la position approximative de la limite des basses mers (ligne en tireté) est également indiquée. B. Carte d'Etat-Major de 1835 montrant la configuration de la zone côtière au début du XIXe siècle sur laquelle a été reporté le trait de côte extrait des photographies aériennes de 2005 (ligne noire continue).

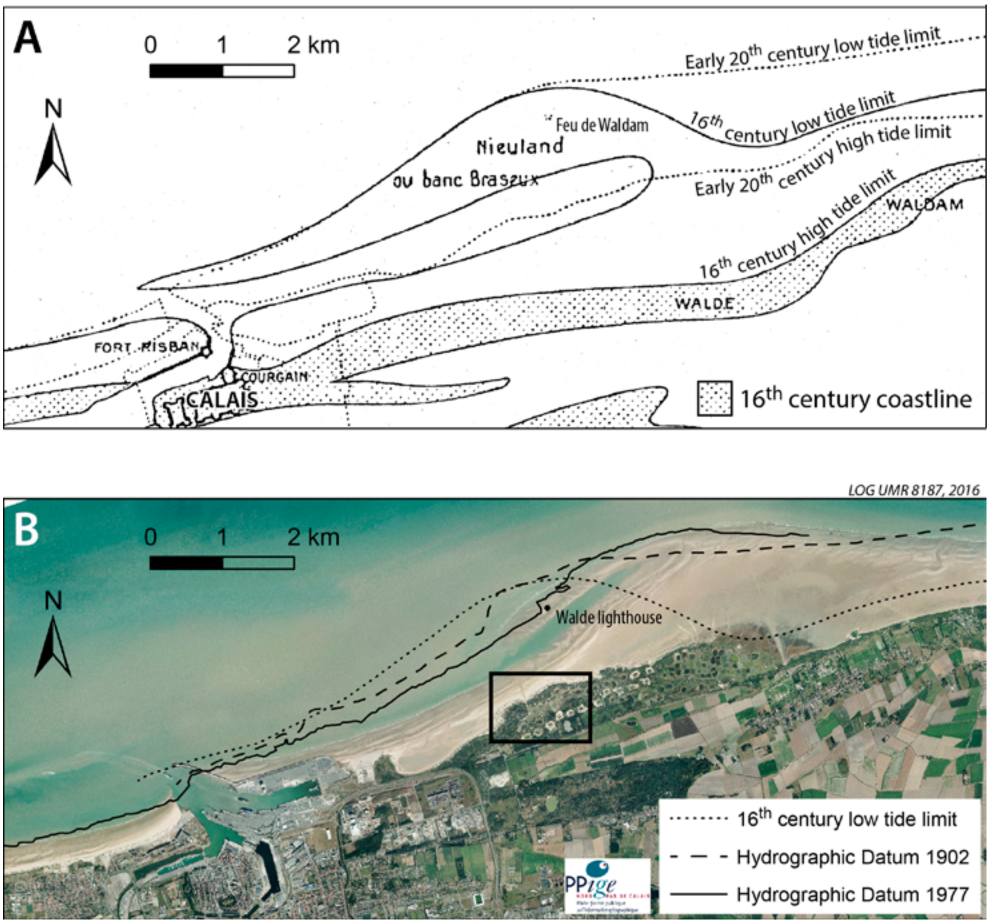

Fig. 3 – Changes in the position of tide levels since the 16th century.

A. Evolution of low and high tide limits east of Calais since the 16th century based on the map of Briquet (1930). B. Position of 1902 and 1977 hydrographic datum extracted from SHOM hydrographic field sheets and 16th century low tide limit east of Calais overlain onto 2012 aerial photograph. The inset on the aerial photograph shows the location of Figure 4.

Fig. 3 – Variations de la position des niveaux de marée depuis le XVIe siècle.

A. Variations de la position des limites de basse mer et de haute mer à l'est de Calais d'après la carte de Briquet (1930). B. Positions du zéro hydrographique 1902 et 1977 extrait des minutes bathymétriques du SHOM et de la limite de basse mer au XVIe siècle à l'est de Calais reportées sur la photographie aérienne de 2012. Le rectangle sur la photographie aérienne montre la localisation de la Figure 4.

migrated onshore. The merging of these subtidal banks to the shore resulted in the formation of a prominent sand flat where active aeolian sand transport favored rapid coastal dune formation and shoreline advance (Anthony, 2013).

Comparison of the location of the hydrographic datum on a marine chart dating back to 1902 and on 1977 hydrographic field sheets with the position of the 16th century low tide limit depicted by Briquet (1930), showed that this convex sand platform migrated eastward since its formation during the 16th century (fig. 3B). This eastward migration resulted in reduction in beach width during the 20th century on the western side of the sand platform while the beach considerably widened to the east. To the west, the decrease in beach width was further enhanced by the migration of a channel separating the foreshore and the remaining part of the banc Braseux. Because the aerial photograph shown on Figure 3B was not taken at low tide, the sand/water interface, which almost matches the 1977 hydrographic datum at the seaward edge of the sandflat, cannot be directly compared with the (near) low tide limits inferred from hydrographic charts. The aerial photograph was in fact taken near mid tide on 12 August 2012 when the water level was 4.42 m above spring low tide level. The low tide level was therefore located seaward of the instantaneous water level visible on the aerial photograph, which is consistent with a continuing seaward development of the sand platform between 1977 and 2012.

4.2 Recent coastal dune evolution: Formation of a field of incipient foredunes

On the 1938 aerial photograph, three small sandy islands, or islets, 190 m to 360 m long and 150 m to 210 m wide, detached from the main shore, are clearly visible on the upper beach platform (fig. 4). These sandy islands with chaotic aeolian dune morphology and sparse vegetation cover were separated from each other by channels and from the main shoreline by a wet upper beach up to 110 m wide, indicating tidal incursion around and past them at high tide. The main shoreline consisted of low vegetated coastal dunes, with a hummocky topography, and to landwards, reclaimed salt marshes backed against the Taaf dike.

According to Briquet (1930), coastal dunes in this area were not continuous, but formed an archipelago of small sandy islands separated by narrow channels and spaced one hundred metres or more apart (fig. 2A). In his description of the shoreline in ca. 1920-1925, Briquet (1930) called these coastal features, developed on the upper beach, "*insular dunes*" as they formed isolated mounds, detached from the main shoreline and surrounded by the sea at high tides. According to Briquet (1930), these types of coastal dunes were observed all along the coast, between Calais and Gravelines (fig. 1). Some of them separated salt marshes, like between Calais and "*Hemmes de Marck*" (fig. 2A) and some developed on the seaward face of dikes and rapidly merged into a continuous foredune, like close to "*Hemmes d'Oye*" where coastal dunes are backed by the Robelin dike (fig. 2A). It is likely that the three sandy islets visible on the 1938 aerial photographs (fig. 4) correspond to these "*insular dunes*" described by Briquet (1930).

Between 1938 and 1963, these islets continued to expand and progressively coalesced with the main shore (fig. 5). The development and coalescence of these small islands was responsible for a seaward shoreline movement, the advance of the shoreline having been the greatest on the west side of the westernmost island where the shoreline in 2012 was up to 300 m beyond its 1938 position (fig. 5). At the northern edge of the islands, the shoreline advance was more limited, ranging from approximately 20 to 90 m between 1938 and 2012. Once the sandy islands were connected to the main shore, two different types of foredunes developed. To the north, an established foredune ridge began to form and very large isolated nebkhas, up to 4 m high and 20 m wide, formed seaward (Ruz et al., 2017), while westward, a hummocky dunefield developed.

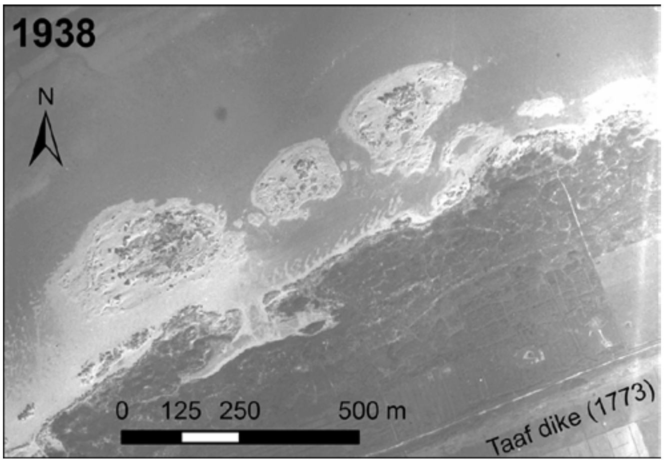

Fig. 4 – 1938 aerial photograph showing sandy islets on the upper beach east of Calais.

See Figure 3 for location.

Fig. 4 – Photographie aérienne de 1938 montrant la présence d'îlots sableux sur le haut de plage à l'est de Calais.

Voir la Figure 3 pour la localisation.

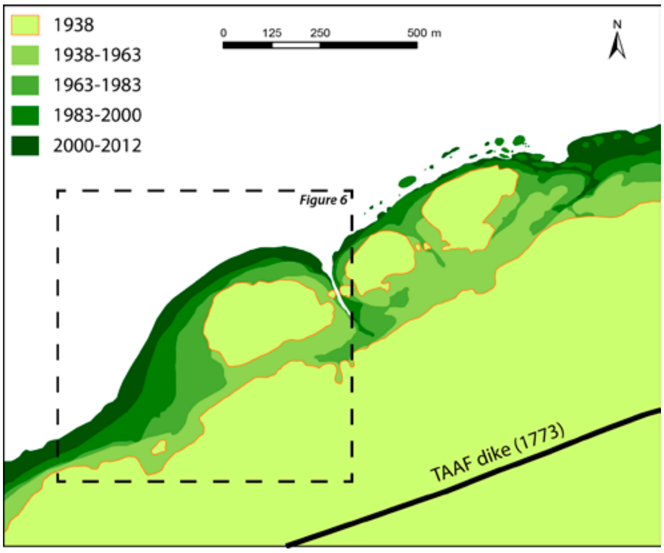

Fig. 5 – Representation of shoreline change east of Calais between 1938 and 2012 based on aerial photographs interpretation. The dashed line square indicates the position of Figure 6.

See Figure 1 for location of Figure 5.

Fig. 5 – Représentation des variations de la position du trait de côte à l'est de Calais entre 1938 et 2012 établies à partir de photographies aériennes. Le rectangle en tireté indique la localisation de la Figure 6.

Voir la Figure 1 pour la localisation de la Figure 5.

A close look at the westernmost one reveals that the merging of the islet with the main shoreline landward, took place sometime between 1949 and 1963 (fig. 6). On the 1949 air photograph, the islet was still clearly detached from the main shore, being separated from the vegetated coastal dunes by an area of bare sand with some small, partly vegetated incipient foredunes. In 1961, the eastern part of the islet was connected to the main shoreline, but a channel remained westward. By 1963, incipient foredunes had increased in size and new ones had appeared across that area, along the main shoreline as well as around the islet, which resulted in the progressive coalescence of the islet with the main shore as shown by the continuous extent of dry sand surrounding the former islet (fig. 6). The coalescence of the islet resulted in the formation of a small embayment facing southwest. After 1963, the density of vegetation cover, likely marram grass and shrubs, increased significantly, and by 1972, almost all the surface of the former island and adjoining bare sand area was covered by vegetation. From that time on, a considerable development of incipient foredunes was observed, resulting in the formation of an arcuate field of incipient foredunes, consisting of roughly circular and partly stabilized nebkhas, separated by deflation corridors and deflation hollows (fig. 6). In 1983, this field of incipient foredunes was about 600 m long and 40 m to 100 m wide. This dunefield continued to develop during the following decades, reaching a maximum width of more than 150 m in 1993 and nearly 200 m in 2012 (fig. 6). The development of this dunefield resulted in a shoreline advance at a rate of up to 5.25 $m.y^{-1}$ between 1963 and 2012.

Although the field of incipient foredunes continued to expand since its initiation during the 1960s, the morphology of individual nebkhas, colonized by a sparse vegetation cover, tended to remain remarkably stable through time (fig. 7). At the seaward edge of the dunefield, however, the progressive coalescence of the seawardmost incipient foredunes resulted in the initiation of a discontinuous foredune ridge visible on the 2005 aerial photograph (fig. 8). From 2005 to 2015, continuing sediment accumulation resulted in the progressive development of the foredune ridge at the seaward limit of the hummocky dune field and in the formation of a new line of incipient foredunes (fig. 8). This resulted in a seaward movement of the shoreline ranging from approximately 25 to 30 m to the north and of more than 50 m to the west where the dunefield is the wider. By 2015, the dunefield reached a total area of about 100 000 m^2.

Two DGPS topographic profiles surveyed on September 2011 and April 2014 provide an example of the recent evolution of the incipient foredune field and the foreshore (fig. 9). The profiles have different orientations with respect to the general orientation of the shoreline in order to assess the influence of the exposure of the coast to winds on the evolution of the incipient foredunes. Profile 1 is oriented perpendicular to the shoreline, and profile 2 is oriented east-west, obliquely to the beach and shoreline, parallel to the upper beach platform extending westward. Between 2011 and 2014, no major morphological changes are discernible inland in the hummocky dune field, although slight accumulation occurred. Close to the upper beach, however, accumulation prevailed on the newly developed incipient foredunes (fig. 9). On profile 1, a maximum vertical accretion of about 1.5 m was measured on the developing foredune ridge, but erosion was also observed in the inter-dune swale between the foredune ridge and the new line of

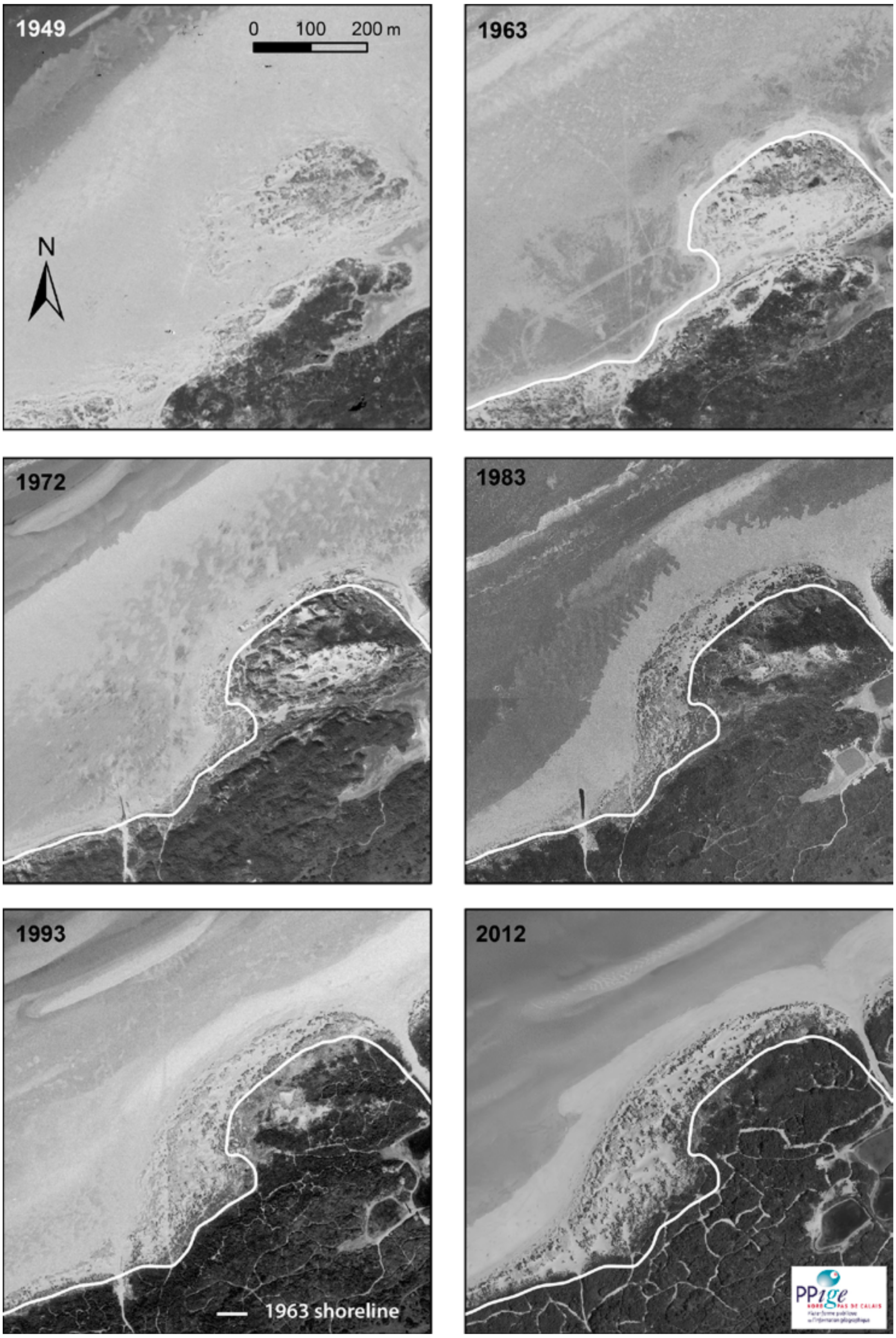

Fig. 6 – Series of aerial photographs from 1949 to 2012 showing the progressive merging of the sandy islets and the development of the incipient foredune field.
See Figure 5 for location.

Fig. 6 – Série de photographies aériennes de 1949 à 2012 montrant la coalescence des îlots sableux et le développement du champ de dunes embryonnaires.
Voir la Figure 5 pour la localisation des photographies aériennes.

incipient foredunes (fig. 9A). While the backshore was characterized by a mean sand accumulation of +0.24 m, the foreshore underwent erosion (mean erosion: -0.18 m) (fig. 9A), possibly due to the series of severe storms that caused coastal erosion in the region during winter 2013-2014 (Crapoulet et al., 2017). More extensive sediment deposition occurred along profile 2, not only in the developing incipient foredunes where a mean accumulation of approximately +0.47 m was measured, but also over the wide upper beach platform where sediment deposition was about +0.13 m on average (fig. 9B). Because the altitude of the upper beach platform is close to the mean spring tide level (MHWS), it is only episodically submerged during high amplitude tides or during high water level events associated with storm surges. This low frequency of submergence favors aeolian sediment transport over this wide sand surface that constitutes an extensive source of dry sand. As a result, nebkha and shadow dunes are actively forming on the upper beach in that area

Fig. 7 – Photographs of the incipient foredune field in (A) 2006 and (B) 2014.

Note the maintenance of the morphology of the incipient foredunes and progressive colonization by vegetation (marram grass). Photographs: M.H. Ruz.

Fig. 7 – Photographies du champ de dunes embryonnaires en (A) 2006 et (B) 2014.

Noter la permanence de la morphologie des dunes embryonnaires et la colonisation progressive par la végétation (oyat). Clichés : M.H. Ruz.

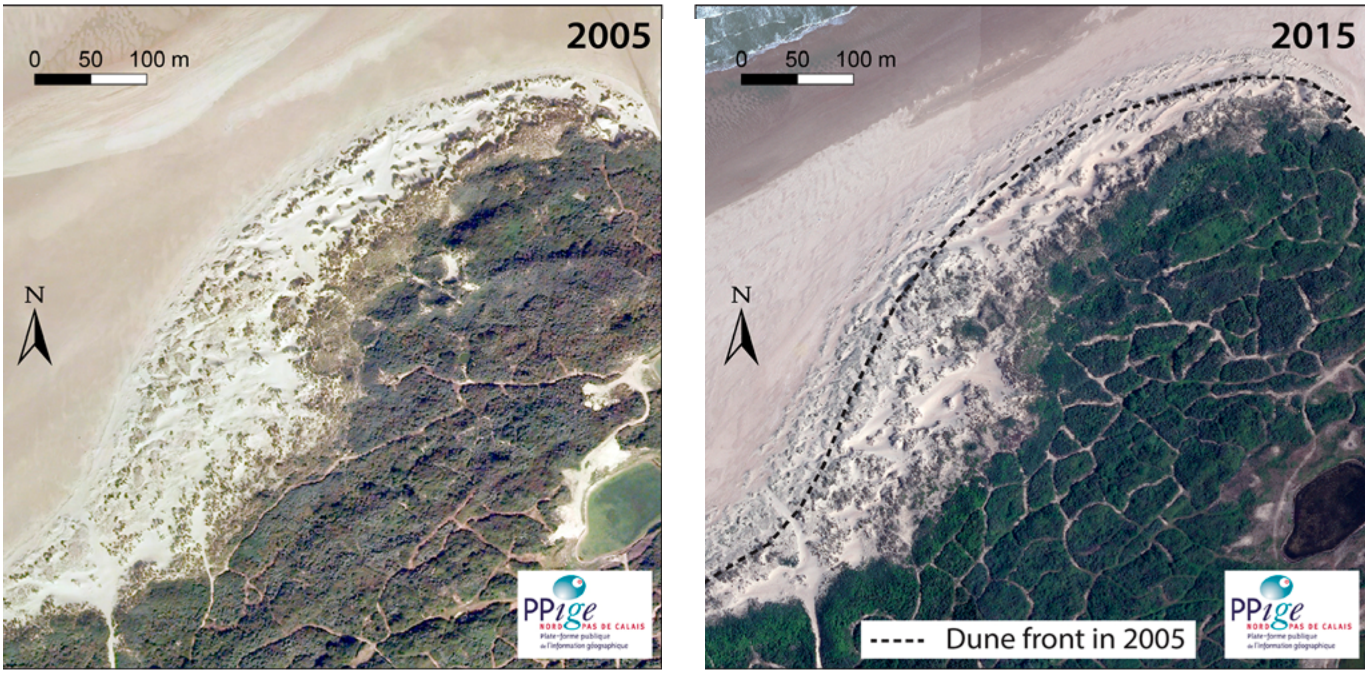

Fig. 8 – Aerial photographs showing the seaward development of the incipient foredune field from 2005 to 2015.

Fig. 8 – Photographies aériennes montrant la progression vers la mer du champ de dunes embryonnaires de 2005 à 2015.

(fig. 10A) facing the prevailing southwesterly winds (fig. 1). Due to their location behind the high tide limit and because wave energy is strongly dissipated over the wide and low gradient upper beach sand platform (slope < 0.5%), it appears that these incipient foredunes were not eroded during the winter storms of 2013-2014.

Airborne LiDAR topographic data collected in March 2011 and in January 2014 were used for analysing the morphological change and evaluate sediment volume variations throughout the study area between the two surveys. Detailed DTMs enabled us to discriminate the spatial extent of the hummocky dune field (#3) and of a zone of developing foredunes (#2) (fig.10C). The comparison of the DTMs show that between 2011 and 2014, the studied area lost sediment, with a net volume change of -40,014 m^3 corresponding to a mean volume change of about -0.08 $m^3.m^{-2}$ when averaged over the complete study area (fig. 10B), which is slightly less than the error margin of ± 0.1 $m^3.m^{-2}$. Erosion, which was possibly due to the impacts of the 2013-2014 winter storms, essentially took place over the foreshore that experienced a sediment loss of approximately -77,330 m^3 (0.21 $m^3.m^{-2}$) (fig. 10B). Except for the upper beach platform that remained fairly stable, the coastal dunes mostly underwent sediment deposition although some erosion locally occurred in blowouts and deflation hollows (fig. 10C). In the hummocky dunefield (#3) (fig. 10C), a mean accumulation of 0.26 $m^3.m^{-2}$ was recorded, and the maximum sedimentation took place in the zone of developing foredunes (#2) (fig. 10C), with a mean volume change of about 0.62 $m^3.m^{-2}$ (fig. 10B) revealing that most of the wind-blown sand transported from the beach is deposited in this zone of growing and rapidly forming shadow dunes and nebkha mounds (fig. 10D-E).

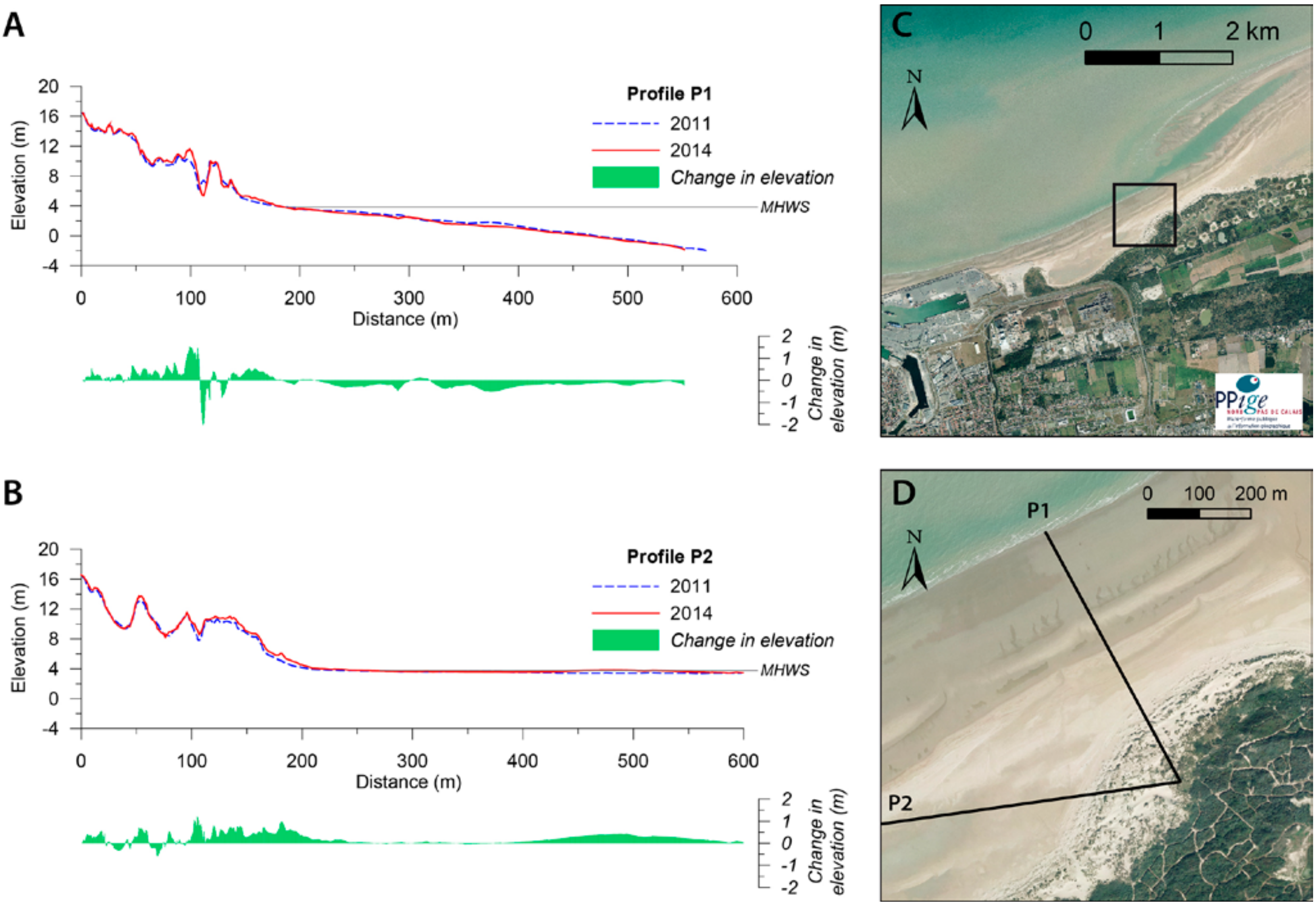

Fig. 9 – Cross-shore DGPS topographic profiles of the beach and coastal dunes east of Calais.

A and B. Changes in surface elevation between September 2011 and April 2014 (MHWS corresponds to the mean high water spring tide level). C. Location of the study area (the black square indicates the location of the aerial photograph shown in D). D. Location of the topographic profiles.

Fig. 9 – Profils topographiques DGPS de la plage et des dunes côtières à l'est de Calais.

A et B. Variations de la topographie entre septembre 2011 et avril 2014 (MHWS correspond au niveau moyen des pleines mers de vive eau). C. Localisation du site d'étude (le carré noir indique la localisation de la photographie aérienne apparaissant en D). D. Localisation des profils topographiques.

5. Discussion

Incipient foredunes are usually considered as the initial dunes that develop above the high tide limit. In their early stage of formation, they form nebkha and shadow dunes, and then they usually adopt one of three forms: ramps, terraces, and ridges (Hesp, 2002). Because they are potentially exposed to wave action at high tide, incipient foredunes can be eroded, but they may also survive and eventually evolve as an established foredune (Hesp, 1999, 2002). Once formed, with continuous sediment supply, foredunes may gradually or rapidly become isolated from accretion and erosion processes by the seaward development of a new incipient foredune, which itself may evolve into an established foredune. The original foredune then becomes a relict foredune as it is largely or totally removed from a foremost beach position. Systematic beach progradation over time may lead to the development of wide foredune plains (Sanderson et al., 1998; Hesp, 2012). Here, since the merging of sandy islets to the main shoreline, a field of incipient foredunes rapidly developed, forming a low-lying platform consisting of hummocks or nebkhas. Such fields of incipient foredune or nebkha can be found in other coastal environments (Hesp and Walker, 2013), especially in arid or semi-arid regions (Zahran, 1993; Cabrera and Alonso, 2010) although they can also be observed in cold-climate regions (Mountney and Russel, 2006; Ruz and Hesp, 2014). They usually consist of incipient foredunes that form inland of the beach as a result of landward-directed aeolian sediment transport from the beach. Conversely to what we observed near Calais, where the formation of a field of incipient foredunes resulted in seaward shoreline progradation, the incipient foredunes found in other coastal settings generally develop at a distance inland from the beach where wind-blown sands are deposited.

The formation of a field of incipient foredunes east of Calais occurred in a context of massive sediment supply from the shoreface through the onshore migration of an extensive sand bank that resulted in the development of a prominent sand flat during the last centuries (Briquet, 1930; Héquette and Aernouts, 2010). Once formed, this sandflat represented a source of sand for the formation of large sandy islets (Briquet, 1930). The coalescence of these aeolian landforms through time was responsible for a significant advance of the shoreline that resulted in the formation of a protruding headland in the study area (fig. 6). The formation of this sandy headland induced a major change in shoreline orientation, forming a small embayed shoreline facing the southwest which progressively became filled-up by aeolian sediment (fig. 6).

The presence of a 250 to 400 m wide and 1,500 m long upper beach platform (fig. 10A), over which aeolian deflation can be significant under the action of the dominant southwesterly winds, favors onshore aeolian sediment transport along this coastal stretch.

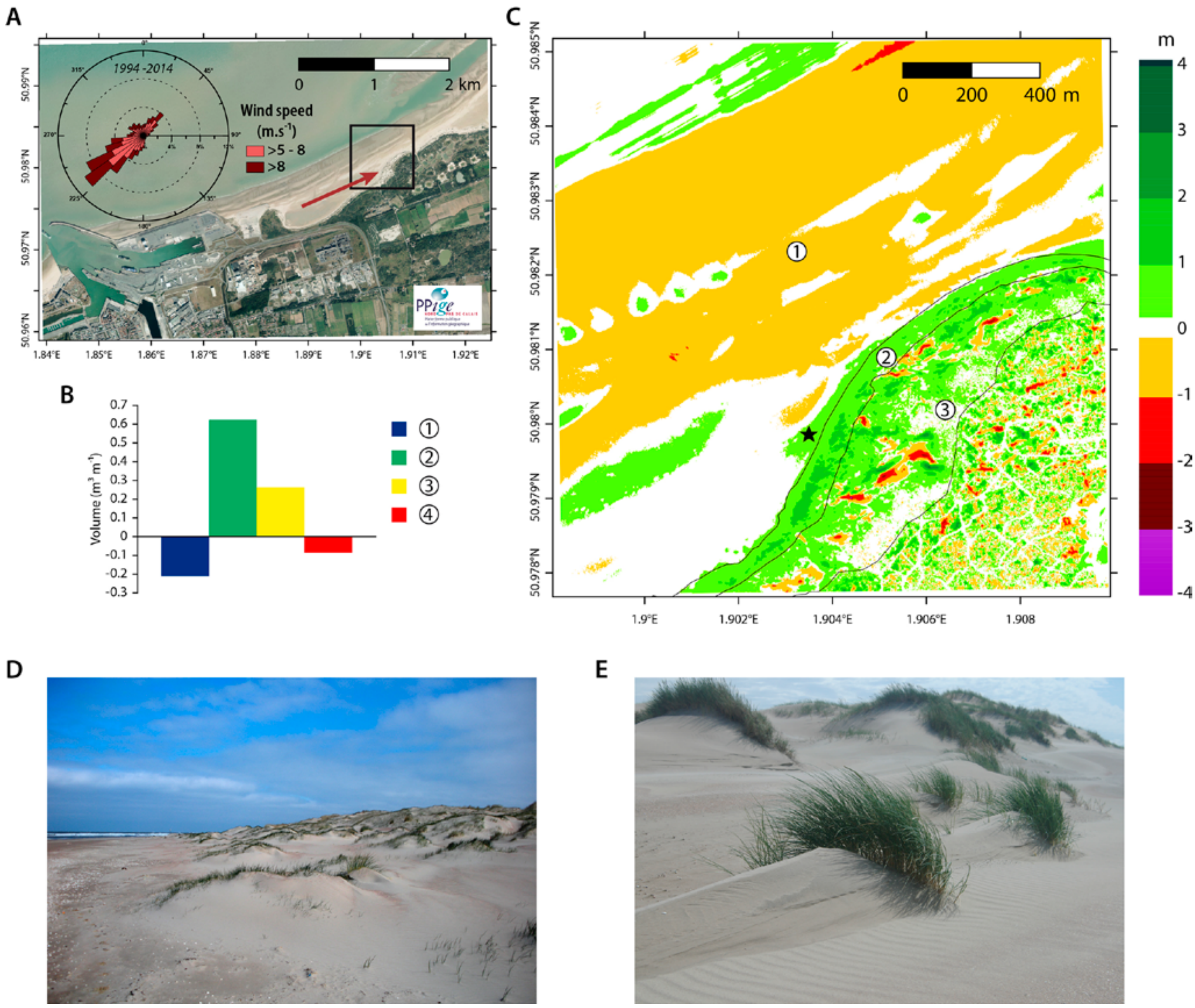

Fig. 10 – Topographic evolution and volume change in the study area between 2011 and 2014.

A. Aerial photograph of the study area showing the potential fetch length (arrow) over the sand platform and rose diagram of efficient winds for aeolian sand transport (5 $m.s^{-1}$ corresponds to the critical wind speed necessary for initiating sand transport. Winds with speed > 8 $m.s^{-1}$ represent the most efficient wind for aeolian transport according to *in situ* measurements; Anthony et al., 2007). B. Sediment volume variation per unit surface area between March 2011 and January 2014 in the different coastal sub-environments and over the complete area shown in C (1. Foreshore, 2. Zone of developing incipient foredunes, 3. Hummocky dunefield; 4. Complete study area covered by 1, 2 and 3). C. Net changes in surface elevation over the study area between March 2011 and January 2014 based on LiDAR topographic data (the star shows the location of the photographs D and E). D and E. Photographs of nebkha mounds and shadow dunes developing on the upper beach (photographs: D. M.H. Ruz, E. P.A. Hesp).

Fig. 10 – Variations de la topographie et des volumes sédimentaires dans le secteur d'étude entre 2011 et 2014.

*A. Photographie aérienne de la zone d'étude montrant la longueur potentielle du fetch éolien (flèche) sur la plateforme sableuse et rose des vents efficaces pour le transport de sables (la vitesse de 5 $m.s^{-1}$ correspond à la vitesse de vent minimale pour initier le transport de sables. Les vents ayant une vitesse > 8 $m.s^{-1}$ représentent les vents les plus efficaces pour le transport éolien d'après des mesures réalisées in situ ; Anthony et al., 2007). B. Variations du volume sédimentaire par unité de surface entre mars et 2011 et janvier 2014 dans les différents sous-environnements littoraux présentés en C (1. Plage intertidale, 2. Zone de formation de dunes embryonnaires, 3. Champ de dunes chaotiques, 4. Ensemble de la surface couverte par 1, 2 et 3). C. Variations de la topographie de l'ensemble de la zone d'étude mars et 2011 et janvier 2014 déterminées à partir de levés LiDAR topographiques (l'étoile montre la localisation des photographies D et E); D et E. Photographies de nebkhas et de dunes en queue de comète (*shadow dunes*) se développant sur le haut de plage (photographies :* D. M.H. Ruz, E. P.A. Hesp*).*

It is acknowledged that the length of the deflation surface is not the only factor that can limit or enhance aeolian sand transport, because the critical fetch distance above which aeolian sediment flux becomes saturated is commonly reached over distances of only a few tens of meters on natural beaches (Nordstrom and Jackson, 1992; Bauer et al., 2009). Nevertheless, this extensive sand platform represents a significant sand source that may supply substantial quantities of sand to the coastal dunes during dominant wind conditions due to infrequent submergence. Anthony et al. (2007) carried out aeolian sand transport measurements using a series of sand traps and anemometers on the upper foreshore and on the upper beach platform, close to the incipient foredunes described in the present paper. These aeolian sand transport measurements revealed that under dominant southwesterly winds, sand transport rate was on the order of 100 $kg.m^{-1}.h^{-1}$ under mean wind speed of 10-12 $m.s^{-1}$. Under direct onshore winds from north to northwest, with lower mean wind speed (8 $m.s^{-1}$) and higher beach surface moisture, sand transport rates were significantly lower (< 10 $kg.m^{-1}.h^{-1}$), which is explained by lower wind speed but also by a smaller dry deflation surface to the north responsible for shorter wind fetch.

The higher rates of dune progradation to the west and southwest compared to the seaward progression of the incipient foredunes to the north can certainly be attributed to higher sediment fluxes induced by the dominant southwesterly winds (fig. 10A), but also to a larger accommodation space to the west. The initial arcuate shoreline and wide upper beach platform to the west (fig. 6) provided extensive surfaces over which incipient foredunes could develop out of reach of waves.

Once formed, the shape of the incipient foredunes tended to remain fairly stable while new incipient dunes were forming seaward. This may be explained by the rapid development of new incipient foredunes that trapped onshore-transported aeolian sand at the expanse of the older incipient foredunes which gradually became isolated from their sand source. This evolution resulted in the formation of a hummocky dunefield rather than an established foredune. The fact that the incipient foredune field is exposed to dominant winds from two main directions (southwest and north-northeast), responsible for frequent dune reworking, probably also restrict the merging of the nebkhas in a single continuous foredune. Consequently, the individual aeolian mounds tend to grow up and maintain their shape (fig. 9), which is also favored by increasing vegetation cover (fig. 7).

Most studies on shoreline evolution in northern France have been focused on coastal erosion and shoreline retreat (Corbau et al., 1993; Vasseur and Héquette, 2000; Aernouts and Héquette, 2006; Chaverot et al., 2008; Ruz et al., 2009), but there are several examples of stable or even advancing shorelines along the coast of the region (Battiau-Queney et al., 2003; Chaaban et al., 2012; Crapoulet et al., 2015), although they are generally of limited extent. The rapid shoreline progradation associated with extensive incipient foredune development observed east of Calais took place at the only site along the Flemish coastal plain where coastal dunes are facing the dominant wind, which is blowing mainly alongshore everywhere else. Such rapid development of coastal dunes was also favored by a high sediment supply from the shoreface since at least the beginning of the 20th century (Héquette and Aernouts, 2010). The exact mechanisms and processes responsible for the transfer of sediment from the shoreface to the intertidal zone are not entirely understood yet (Sabatier et al., 2009), but they are likely related to the combination of onshore-directed wave-induced transport during storm events and alongshore tidal flows that resulted in a massive supply of sand to the beach (Anthony, 2013). It is noteworthy that shoreline advanced seaward at that site despite a sea level rise that was estimated at about 15-20 cm during the 20th century in the region (Haigh et al., 2009). The predictions for coastal change under scenarios of global sea-level rise usually depict increased coastal erosion, accelerated shoreline retreat and accompanying damages to coastal infrastructures (Nichols and Cazenave, 2010; Kron, 2013). As suggested by Psuty and Silveira (2014), however, much of the predictions of the impacts of sea-level rise tend to ignore the role of sediment budget in the maintenance of coastal morphology and the dynamics of sediment transfers in the beach-dune sand-sharing system. As stressed by these authors, sediment budget and accommodation space are key factors in the maintenance of coastal dunes under rising sea level. The example reported in the present paper shows that shoreline can significantly advance seaward during a period of rising sea level if there is sufficient sediment supply. However, it is not known if incipient foredunes will continue to develop and extend seaward with increasing rates of sea level rise in the future.

6. Conclusion

This study shows how a field of incipient foredunes of approximately 100,000 m^2 can develop in only a few decades. This work underlines the significant role of sand supply in the dynamics of the incipient foredunes that did not evolve in a continuous established foredune, but rather formed a hummocky dunefield composed of nebkha mounds. Such evolution is linked to the presence of an extensive upper beach sand platform that serves as a source of sediment for onshore-directed aeolian transport, providing sediment to the developing incipient foredunes which face the dominant southwesterly winds. Our observations show that once incipient foredunes formed, new ones rapidly developed seaward which trapped most of the aeolian sand transported from the beach. Continuous sediment supply induced rapid incipient foredune development that resulted in seaward shoreline displacement at a rate of about 5 m.y^{-1}. This evolution that leaded to the seaward widening of a hummocky dunefield rather than the formation of a foredune plain, which may be expected in the case of high aeolian sand supply, is probably due to the fact that the incipient foredune do not have enough time to coalesce and to merge into a continuous foredune ridge as they rapidly become disconnected from their main sediment source. The morphology of these nebkhas largely remains at this stage as they undergo partial reworking by southwest and northeast winds and become progressively stabilized by vegetation.

This example is certainly among the only ones occurring along the coast of the Flemish coastal plain which can probably be explained by the rare occurrence of such high sediment supply from the shoreface induced by the welding of a nearshore sand bank and by the fact that the coastline is locally facing the dominant longshore winds. Our results also highlight that coastal dunes can build up and develop seaward, even under conditions of rising sea level if sediment supply is sufficiently high and if there is enough accommodation space, which opens perspectives in the context of future sea level rise.

Acknowledgements

This study was partly funded by the French Centre National de la Recherche Scientifique (CNRS) and by the Ministère de l'Education Nationale et de l'Enseignement Supérieur through operating research funds to the Laboratoire d'Océanologie et de Géosciences. The LiDAR data were obtained through the interregional program CLAREC funded by the Regional Councils of Normandy, Picardie and Nord-Pas de Calais and by the French Centre National de la Recherche Scientifique (CNRS). F. Levoy and P. Bretel of the CLAREC program are thanked for the acquisition and processing of the raw LiDAR data. Vertical aerial photographs were obtained from the French National Institute of Geographic and Forestry Information (IGN). Water level and bathymetry data were obtained from the French Hydrographic Survey (SHOM - contract n°E70/2001). Wind data were obtained by courtesy of Météo-France though a research agreement with the Université du Littoral Côte d'Opale. The authors would like to thank F. Sabatier and an anonymous reviewer for their constructive comments and suggestions.

References

Aernouts D., Héquette A. (2006) – L'évolution du rivage et des petits fonds en baie de Wissant pendant le XXe siècle, Pas-de-Calais, France. Géomorphologie : relief, processus et environnement, 12 (1), 49-64.
DOI : 10.4000/geomorphologie.477

Anthony E.J. (2013) – Storms, shoreface morphodynamics, sand supply, and the accretion and erosion of coastal dune barriers in the southern North Sea. Geomorphology, 199, 8-21.
DOI : 10.1016/j.geomorph.2012.06.007

Anthony E.J., Levoy F., Montfort O., Degryse-Kulkarni C. (2005) – Patterns of short-term intertidal bar mobility on a ridge and runnel beach, Merlimont, Northern of France. Earth Surface Processes and Landforms, 30 (1), 81-93.
DOI : 10.1002/esp.1129

Anthony E.J., Vanhee S., Ruz M.H. (2006) – Short-term beach-dune sand budgets on the north sea coast of France: Sand supply from shoreface to dunes, and the role of wind and fetch. Geomorphology, 81 (3-4), 316-329.
DOI : 10.1016/j.geomorph.2006.04.022

Anthony E.J., Vanhée S., Ruz M.H. (2007) – Embryo dune development on a large actively accreting macrotidal beach : Calais, North Sea coast of France. Earth Surface Processes and Landforms, 32 (4), 631-636.
DOI : 10.1002/esp.1442

Anthony E.J., Ruz M.H., Vanhée S. (2009) – Aeolian sand transport over complex intertidal bar-trough beach topography. Geomorphology, 105 (1-2), 95-105.
DOI : 10.1016/j.geomorph.2007.12.013

Augris C., Clabaut P. (2001) – Cartographie géologique des fonds marins côtiers Exemples le long du littoral français. Editions Ifremer, Brest, 76 p.

Augris C., Clabaut P., Vicaire O. (1990) – Le domaine marin du Nord-Pas-de-Calais – Nature, morphologie et mobilité des fonds. Edition IFREMER – Région Nord-Pas-de-Calais, 93 p.

Battiau-Queney Y., Fauchois J., Deboudt P., Lanoy-Ratel P. (2001) – Beach-dune systems in a macrotidal environment along the northern French coast. Journal of Coastal Research, Special Issue 34, 580-592.

Battiau-Queney Y., Billet J.F., Chaverot S., Lanoy-Ratel P. (2003) – Recent shoreline mobility and geomorphologic evolution of macrotidal sandy beaches in the north of France. Marine Geology, 194 (1-2), 31-45.
DOI : 10.1016/s0025-3227(02)00697-7

Bauer B.O., Sherman D.J. (1999) – Coastal dune dynamics: problems and prospects. In: A.S. Goudie, I. Livingstone and S. Stokes (Eds.), Aeolian Environments, Sediments, and Landforms, Wiley, Chichester, 71-103.

Bauer B.O., Davidson-Arnott R.G.D., Sherman D.J., Nordstrom K.F., Ollerhead J., Jackson N.L. (2009) – Aeolian sediment transport conditions on a beach: Surface moisture, wind fetch, and mean transport. Geomorphology, 105 (1-2), 106-116.
DOI : 10.1016/j.geomorph.2008.02.016

Beck C., Clabaut P., Dewez S., Vicaire O., Chamley H., Augris C., Hoslin R., Caillot A. (1991) – Sand bodies and sand transport paths at the English Channel-North Sea border: morphology, dynamics and radioactive tracing. Oceanologica Acta, 11, 111-121.

Blanchard R. (1906) – La Flandre. Etude géographique de la Plaine Flamande en France, Belgique et Hollande. Thèse de doctorat, Université de Lille, 530 p.

Briquet A. (1930) – Le Littoral du Nord de la France et son évolution morphologique. Armand Colin, Paris, 439 p.

Cabrera L.L., Alonso I. (2010) – Correlation of aeolian sediment transport measured by sand traps and fluorescent tracers. Journal of Marine Systems, 80 (3-4), 235-242.
DOI : 10.1016/j.jmarsys.2009.10.012

Carter R.G.W. (1988) – Coastal Environments – An Introduction to the Physical, Ecological and Cultural Systems of Coastlines. Academic Press, London, 617 p.

Chaaban F., Darwishe H., Battiau-Queney Y., Louche B., Masson E., El Khattabi J, Carlier E. (2012) – Using ArcGIS® Modelbuilder and Aerial Photographs to Measure Coastline Retreat and Advance: North of France. Journal of Coastal Research, 28, 1567-1579.
DOI : 10.2112/jcoastres-d-11-00054.1

Chaverot S., Héquette A., Cohen O. (2008) – Changes in storminess and shoreline evolution along the northern coast of France during the second half of the 20th century. Zeitschrift für Geomorphologie, Suppl. 52 (3), 1-20.

Corbau C., Clabaut J.P., Tessier B., Chamley H. (1993) – Modifications morphosédimentaires historiques et récentes du domaine côtier dunkerquois (France). Comptes Rendus de l'Académie des Sciences, Paris, Série 11, 316, 1573-1580.

Crapoulet A., Héquette A., Levoy F., Bretel P. (2015) – Evaluation de l'évolution du trait de côte et du bilan sédimentaire littoral en baie de Wissant (nord de la France) par LiDAR aéroporté. Géomorphologie : relief, processus, environnement, 21 (4), 313-330.
DOI : 10.4000/geomorphologie.11146

Crapoulet A., Héquette A., Marin D., Levoy F., Bretel P. (2017, in press) – Variations in the response of the dune coast of Northern France to major storms as a function of available beach sediment volume. Earth Surface Processes and Landforms.
DOI : 10.1002/esp.4098

Davidson-Arnott R.G.D. (2010) – Introduction to Coastal Processes and Geomorphology. Cambridge University Press, Cambridge, 442 p.

Dubois, G. (1926) – Dunes et cordons littoraux dans l'agglomération de Calais. Annales de la Société Géologique du Nord, 51, 129-136.

Dumas-Vence M. (1869) – Notice sur les ports de la Manche et de la Mer du Nord, 1ère partie: De la frontière belge au Cap Gris-Nez. Imprimerie administrative de Paul Dupont, Paris, 213 p.

Haigh I., Nicholls R., Wells N. (2009) – Mean sea level trends around the English Channel over the 20th century and their wider context. Continental Shelf Research, 29 (17), 2083-2098.
DOI : 10.1016/j.csr.2009.07.013

Héquette A., Aernouts D. (2010) – The influence of nearshore sand bank dynamics on shoreline evolution in a macrotidal coastal environment, Calais, Northern France. Continental Shelf Research, 30 (12), 1349-1361.
DOI : 10.1016/j.csr.2010.04.017

Héquette A., Hemdane Y., Anthony E.J. (2008) – Sediment transport under wave and current combined flows on a tide-dominated shoreface, northern coast of France. Marine Geology, 249 (3-4), 226-242.
DOI : 10.1016/j.margeo.2007.12.003

Héquette A., Ruz M.H., Maspataud A., Sipka V. (2009) – Effects of nearshore sand bank and associated channel on beach hydrodynamics: implications for beach and shoreline evolution. Journal of Coastal Research, Special Issue 56, 59-63.

Hesp P.A. (1981) – The formation of shadow dunes. Journal of Sedimentary Petrology, 51, 101-112.
DOI : 10.1306/212f7c1b-2b24-11d7-8648000102c1865d

Hesp P.A. (1988) – Surfzone, beach, and foredune interactions on the Australian South East Coast. Journal of Coastal Research, Special Issue 3, 15-25.

Hesp P.A. (1989) – A review of biological and geomorphological processes involved in the initiation and development of incipient foredunes. In: C.H. Gimingham, W. Ritchie, B.B. Willetts and A.J. Willis (Eds.), Coastal Sand Dunes, Proceedings of the Royal Society of Edinburgh 96B. Royal Society, Edinburgh, 181-201.

Hesp P.A. (1999) – The beach backshore and beyond. In: A.D. Short (Ed.), Handbook of Beach and Shoreface Morphodynamics, John Wiley & Sons, Chichester, 145-170.

Hesp P.A. (2002) – Foredunes and blowouts: initiation, geomorphology and dynamics. Geomorphology, 48(1-3), 245-268. DOI : 10.1016/s0169-555x(02)00184-8

Hesp P.A. (2011) – Dune coasts. In: E. Wolanski and D.S. McLusky (Eds.), Treatise on Estuarine and Coastal Science, vol. 3, Academic Press, Waltham, MA, 193-221.

Hesp P.A. (2012) – Surfzone-beach-dune interactions. In: W.M. Kranenburg, E.M. Horstman and K.M. Wijnberg (Eds.), NCK-days 2012, University of Twente, Department of Water Engineering & Management, Enschede, The Netherlands, 35-40.

Hesp P.A., Walker I.J. (2013) – Aeolian environments: coastal dunes. In: J. Shroder, N. Lancaster, D.J. Sherman, A.C.W. Baas (Eds.), Treatise on Geomorphology, vol. 11, Aeolian Geomorphology, Academic Press, San Diego, 109-133.

Houser C., Ellis J. (2013) – Beach and Dune Interaction. In: J.F. Shroder (Ed.), Treatise on Geomorphology, vol. 10, Academic Press, San Diego, 267-288.

Houthuys R., De Moor G., Sommé J. (1993) – The shaping of the French-Belgian North Sea coast throughout recent geology and history. In: R. Hillen and H.J. Verhagen (Eds.), Coastlines of the Southern North Sea, American Society of Civil Engineers, New York, 27-41.

King C.A.M. (1972) – Beaches and coasts (2nd ed). Arnold, London, 570 p.

Kron W. (2013) – Coasts: the high-risk areas of the world. Natural Hazards, 66 (3), 1363-1382.
DOI : 10.1007/s11069-012-0215-4

Martinez M.L., Psuty N.P. (2004) – Coastal Dunes, Ecology and Conservation. Ecological Studies, Springer-Verlag, Berlin Heidelberg, 386 p.

Mountney N.P., Russell A.J. (2006) – Coastal aeolian dune development, Sólheimasandur, southern Iceland. Sedimentary Geology, 192 (3-4), 167-181.
DOI : 10.1016/j.sedgeo.2006.04.004

Nicholls R.J., Cazenave A. (2010) – Sea-level rise and its impact on coastal zones. Science, 328 (5985), 1517-1520.
DOI : 10.1126/science.1185782

Nordstrom K.F., Jackson N.L. (1992) – Effect of source width and tidal elevation changes on aeolian transport on an estuarine beach. Sedimentology, 39 (5), 769-778.
DOI : 10.1111/j.1365-3091.1992.tb02152.x

Psuty N.P., Silveira T.M. (2014) – Global climate change: an opportunity for coastal dunes? Journal of Coastal Conservation, 14 (2), 153-160.
DOI : 10.1007/s11852-010-0089-0

Reichmüth B., Anthony E.J. (2007) – Tidal influence on the intertidal bar morphology of two contrasting macrotidal beaches. Geomorphology, 90 (1-2), 101-114.
DOI : 10.1016/j.geomorph.2007.01.015

Ruz M.H., Hesp P.A. (2014) – Geomorphology of high-latitude coastal dunes: A review. In: I.P. Martini and H.R. Wanless (Eds). Sedimentary Coastal Zones from High to Low Latitudes: Similarities and Differences, Geological Society of London, Special Publication 388, 199-212.

Ruz M.H., Anthony E.J., Faucon L. (2005) – Coastal dune evolution on a shoreline subject to strong human pressure : the Dunkirk area, northern France. In: J.-L. Herrier, J. Mees, A. Salman, J. Seys, H. Nieuwenhuyse and I. Dobbelaere (Eds.). Proceedings 'Dunes & Estuaries 2005' – International Conference on Nature restoration Practices in European Coastal Habitats, Koksijde, Belgium, 19-23 September 2005, 441-449.

Ruz M.H., Héquette A., Maspataud A. (2009) – Identifying forcing conditions responsible for foredune erosion on the northern coast of France. Journal of Coastal Research, Special Issue 56, 356-360.

Ruz M.H., Héquette A., Marin D. (2017) – Development of large nebkhas along an accreting macrotidal coastline, Northern France. Aeolian Research, 24, 1-14.
DOI : 10.1016/j.aeolia.2016.11.002

Sabatier F. Anthony E.J., Héquette A., Suanez S., Musereau J., Ruz M.H. and Régnauld H. (2009) – Morphodynamics of beach/dune systems: examples from the coast of France. Géomorphologie: relief, processus, environnement, 15 (1), 3-22.
DOI : 10.4000/geomorphologie.7461

Sanderson P.G., Eliot I., Fuller M. (1998) – Historical development of a foredune plain at Desperate Bay, Western Australia. Journal of Coastal Research, 14, 1187-1201.

Saye S.E., Van Der Wal D., Pye K., Blott S.J. (2005) – Beach–dune morphological relationships and erosion/accretion: An investigation at five sites in England and Wales using LIDAR data. Geomorphology, 72 (1-4), 128-155.
DOI : 10.1016/j.geomorph.2005.05.007

Sedrati M., Anthony E.J. (2007) – Storm-generated morphological change and longshore sand transport in the intertidal zone of a multi-barred macrotidal beach. Marine Geology, 244 (1-4), 209-229.
DOI : 10.1016/j.margeo.2007.07.002

Short A.D., Hesp P.A. (1982) – Wave, beach and dune interactions in South Eastern Australia. Marine Geology, 48 (3-4), 259-284.
DOI : 10.1016/0025-3227(82)90100-1

Sommé J. (1975) – Les plaines du Nord de la France et leur bordure, étude géomorphologique. Thèse de doctorat, Université de Paris I, 801 p.

Vasseur B., Héquette A. (2000) – Storm surges and erosion of coastal dunes between 1957 and 1988 near Dunkerque (France), southwestern North Sea. In: K. Pye and J.R.L. Allen (Eds.): Coastal and Estuarine Environments, Geological Society, London, Special Publications 175, 99-107.

Woolard J.W., Colby J.D. (2002) – Spatial characterization, resolution, and volumetric change of coastal dunes using airborne LIDAR: Cape Hatteras, North Carolina. Geomorphology, 48 (1-3), 269-287.
DOI : 10.1016/s0169-555x(02)00185-x

Zahran M.A. (1993) – Dry coastal ecosystems of the Asian Red Sea coast. In: E. van der Maarel (Ed.), Dry coastal ecosystems: Africa, America, Asia and Oceania, Ecosystems of the World 2B, Elsevier, Amsterdam, 17-29.

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La plupart des plages sableuses sont bordées par des dunes côtières qui résultent d'un transport par le vent de sables de la plage vers le haut de plage. Bien que les dunes côtières puissent avoir des formes et des dimensions variables, tout dépendant d'une série de facteurs climatiques, biologiques, géomorphologiques et/ou océanographiques, les dunes les plus communes qui se développent immédiatement à l'arrière des plages sont les dunes bordières qui forment généralement des cordons dunaires plus ou moins symétriques, parallèles au rivage. La morphologie des dunes bordières est également variée, mais elles peuvent être classées en trois types principaux : les dunes embryonnaires, les cordons bordiers uniques et les dunes bordières successives (Hesp, 2002). Dans cet article, nous examinons la formation d'un vaste champ de dunes embryonnaires (≈ 100 000 m²) qui s'est développé dans un secteur limité du littoral du nord de la France caractérisé par une accrétion sableuse et une avancée du trait de côte. Des cartes anciennes et des documents historiques, des photographies aériennes ainsi que des données topographiques récentes acquises par GPS différentiel et par LiDAR aéroporté ont été utilisés pour analyser la formation et l'évolution de ce champ de dunes embryonnaires à des échelles de temps annuelle à pluri-décennale. Ces informations ont permis d'identifier les conditions qui ont conduit au développement d'un champ de dunes embryonnaires plutôt qu'un cordon bordier continu, parallèle au rivage, que l'on rencontre plus communément en bordure de haut de plage.

Le site d'étude, situé à l'est de Calais (fig. 1), correspond à une large plage macrotidale caractérisée par une série de barres sableuses intertidales et une vaste plateforme sableuse de haut de plage à la limite de laquelle des dunes côtières se sont formées. Au large, des bancs sableux de plusieurs kilomètres de longueur, subparallèles au rivage, parsèment les fonds (fig. 1). L'analyse de documents historiques et de cartes anciennes montre que dans ce secteur du littoral du nord de la France, le rivage progresse vers la mer depuis au moins le XVI^e^ siècle (fig. 2). Cette avancée du littoral serait liée à l'accolement de bancs sableux pré-littoraux ayant migré jusqu'à la côte pour former une vaste plateforme subtidale (fig. 3) ayant servi de source de sédiments pour l'estran et les dunes côtières en haut de plage, ce qui a favorisé la progradation du rivage.

L'analyse de photographies aériennes a montré qu'au début du XX^e^ siècle, des îlots sableux, de 190 m à 360 m de long et de 150 m à 210 m de large, se sont formés en haut de plage, au niveau d'un large replat sableux. Initialement entourés par la mer à marée haute, ils se sont progressivement soudés au rivage au milieu du XX^e^ siècle. À partir des années 60, un développement très important de monticules dunaires grossièrement circulaires et partiellement stabilisées par de la végétation s'est produit au contact d'un de ces îlots soudés au rivage, ce qui a conduit à la formation d'un champ de dunes embryonnaires de forme arquée (fig. 6) faisant face aux vents dominants (fig. 10A). En 1983, ce champ de dunes embryonnaires atteignait 600 m de longueur pour une largeur de de 40 à 100 m. Le champ de dune a continué à se développer pendant les décennies suivantes pour atteindre une largeur de près de 200 m en 2012, ce qui s'est traduit par une avancée du trait de côte à un rythme d'environ 5 m/an.

Des mesures topographiques obtenues sur toute la zone d'étude en mars 2011 et en janvier 2014 par LiDAR aéroporté ont révélé que l'estran avait connu une érosion de 0,21 m³/m² (fig. 10C) qui s'est probablement produite lors de la série de tempêtes qui a frappé l'ouest de l'Europe fin 2013, entrainant un recul du trait de côte sur plusieurs sites côtiers du nord de la France (Crapoulet et al., 2017). Ces levés LiDAR ont également montré que, contrairement à la plage, les dunes côtières sur ce site ont surtout connu de l'accumulation pendant cette même période, avec un taux moyen d'accrétion de 0,26 m³/m². L'accumulation maximum (0,62 m³/m²) a été mesurée dans la zone de développement de nouvelles dunes embryonnaires, ce qui indique que la plus grande partie du sable transporté de la plage vers les dunes par le vent est capté dans cette zone au détriment des monticules dunaires plus anciens situés à l'arrière. Dans ce secteur caractérisé par d'abondants apports sableux, la formation d'un champ de dunes chaotiques est probablement due au fait que les dunes embryonnaires n'ont pas assez de temps pour coalescer et former une dune bordière continue du fait qu'elles sont rapidement privées d'apports sédimentaires par de nouvelles dunes embryonnaires qui se forment successivement sur le haut de plage.