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The late Holocene coastal dunefield at Vejers, Denmark: characteristics, sand budget and depositional dynamics

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Abstract

The coastal dunefield at Vejers (west coast of Jutland), which is now stabilized by vegetation, covers an area of approximately 120 km^2 and occurs in a very high-energy wind regime. Along the coastline a narrow belt of 5–15 m high dune ridges occur. The dune ridges are gradually replaced inland by 10–20 m high parabolic or irregular dune forms. In the central part of the dunefield a large aeolian sand plain (ca. 50 km^2) occurs at the windward side of a large and up to 20 m high parabolic dune.

The 3D structure of the dunefield deposits have been studied by geomorphological analysis, sedimentological facies analysis of borings, trenches and natural exposures, and most importantly by georadar mapping. The aeolian deposits which overlie a middle to late Holocene barrier spit depositional system can be divided into a Lower unit that drapes the underlying barrier system topography, and an Upper unit that includes the present dunes. The boundary between the two aeolian units is a well-developed Phragmites peat. The Lower aeolian unit is composed of two aeolian subunits separated by an organic-rich horizon. Also the Upper unit is composite and composed of a basal aeolian sand cover and overlying dune or sand plain deposits. The dune deposits locally are composed of up to four depositional subunits separated by immature soils. The base of the Lower aeolian unit formed around 300 A.D., whereas the Phragmites peat at the base of the Upper aeolian unit has been dated to ca. 1000 A.D. From historical sources we know that the present dunefield primarily formed between 1550 and 1850 A.D.

The sand content in the Upper unit is estimated to 550×10^6 m^3, which yields sand transport rates between 25 m^3 (m width)$^{-1}$ yr^{-1} (accumulation in 1000 years), and 83 m^3 (m width)$^{-1}$ yr^{-1} (accumulation in 300 years). The sand in the dunefield originated from beach deposits. A large but pulsating supply of sand was supplied to the beaches by southwards running coastal currents.

The composition of the Vejers dunefield deposits indicate that periods of dunefield growth alternated with periods of dunefield stabilization during the last ca. 1700 years. The final and most important phase of dunefield growth took place during 'the Little Ice Age'. This period was characterized by an overall cold and stormy climate and a relative low sea level, and resulted in an increased availability of sand in the shorezone and a high influx of sand into the dunefield. Short periods of decreased storminess are recorded by the immature soils. Dunefield growth prior to 1000 A.D. was also linked to cold and stormy climatic intervals.

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1. Introduction

Coastal dunes are significant morphological elements of the west coast of Jutland, Denmark (Fig. 1). Coastal dunes form a number of more or less connected dunefields, the largest of which occur at Vejers, in Thy and on the Skagen Odde (Fig. 1). Various aspects of the late Holocene Vejers dunefield have been described since the area was first mapped by Jessen in 1925, but no modern account of the sedimentary characteristics and genesis of the dunefield exists. We, therefore, initiated a large-scale mapping program of the aeolian deposits in the dunefield in 1987 primarily with georadar analyses. Additional methods included sedimentological studies of natural profiles or trenches and geomorphological analyses.

The main purpose of the project is to map the large-scale 3D structure (sedimentary architecture) of the aeolian deposits and to develop an aeolian event stratigraphy for the late Holocene deposits. The second purpose is to calculate the sand budget of the dunefield and to discuss dunefield dynamics in relation to characteristics of the wind, sediment supply, sea level and human influence. The Vejers dunefield formed on a prograding shoreline supplied with large quantities of sediment from longshore drift (Nielsen et al., 1995). The dynamics of the dunefield are compared to the dynamics of other late Holocene coastal dunefields in northwest Europe.

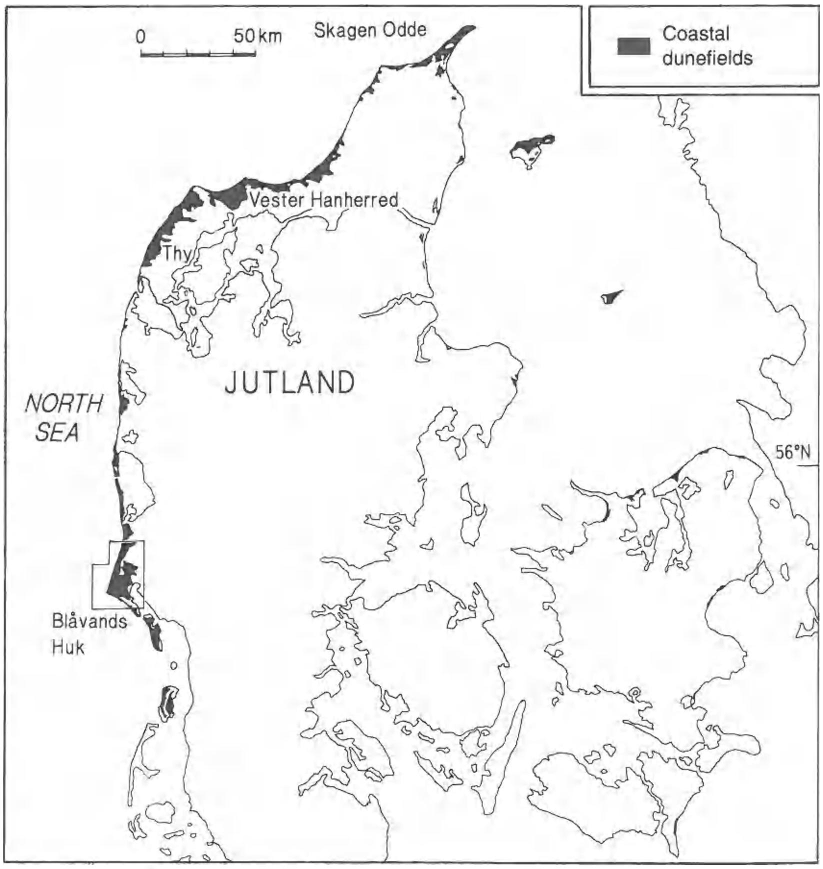

Fig. 1. Map showing the distribution of coastal dunes in Denmark. Map based on Kuhlman, 1968. The largest dunefields occur on the west coast of Jutland. The study area occurs just north of Blåvands Huk (see Fig. 2).

2. Geological setting

2.1. Study area

The coastal dunefield at Vejers is situated at the southern end of the west coast of Jutland and faces the high-energy North Sea (Figs. 1 and 2). The shoreline is wave-dominated and waves have a significant height of approximately 1.6 m (cf. Bartholdy and Pejrup, 1994). The tidal range is between 1 and 1.5 m and storm surges typically reach 2–2.5 m above mean sea level. The strong, southwards directed littoral drift amounts to $0.5–1 \times 10^6$ m^3 yr^{-1} (Bartholdy and Pejrup, 1994). The beach flanking the dunefield is ca. 150 m wide and is composed primarily of fine-grained sand with scattered shells. A well-developed foreshore with ridge and runnel topography and a smooth and gradually sloping backshore is present that is strongly influenced by aeolian processes.

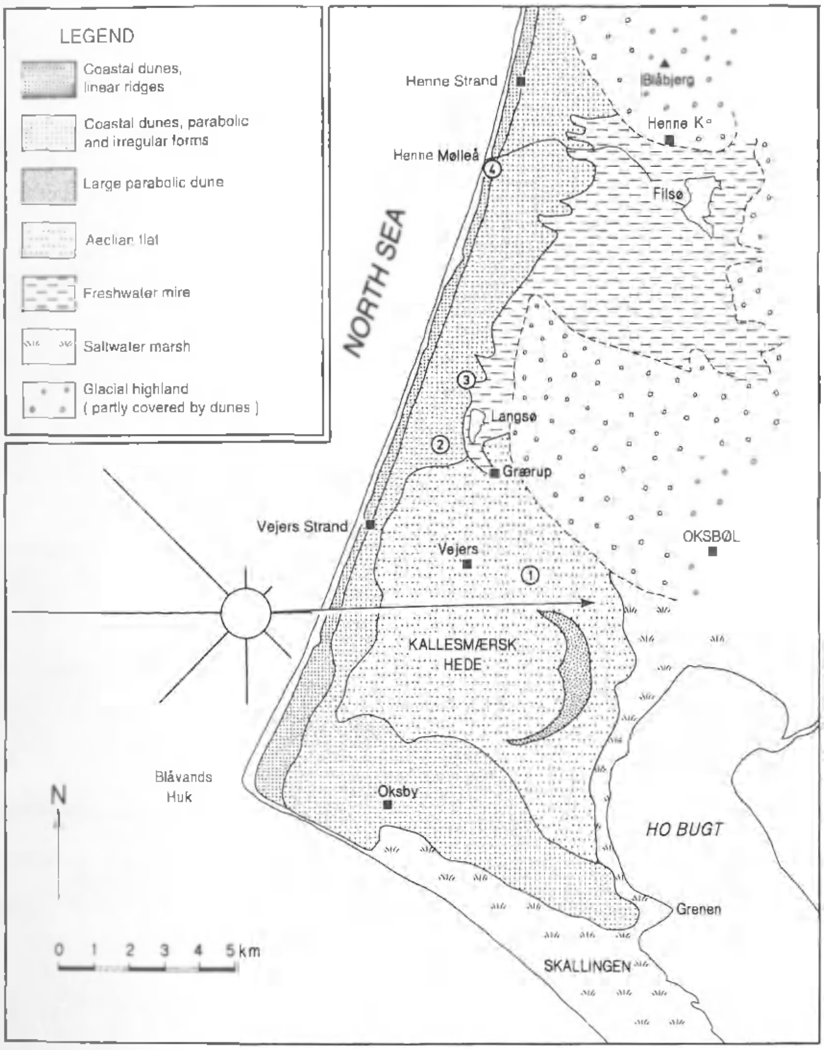

Fig. 2. Generalized geomorphological map of the modern Vejers dunefield. The dunefield is stabilized by vegetation. Numbers refer to studied exposures: *1* = Kallesmærsk Hede; *2* = Grærup Strand, gravel pit; *3* = Børsmose, gravel pit; *4* = Henne Mølleå. The sand rose at Blåvands Huk was calculated using the method described by Fryberger (1979). RDP is 55 m^3 (m width)$^{-1}$ yr^{-1} directed east.

The dunefield at Vejers, an area of ca. 120 km^2, is demarcated by the present shoreline to the west, by a line running from Blåvands Huk to Grenen to the south, by the Ho Bugt followed northwards by glacial hills at Oksbøl and the lake system Langsø and Filsø to the east, and finally by glacial hills at Blåbjerg to the northeast (Fig. 2). Aeolian dunes also cover parts of the glacial highlands at both Oksbøl and Blåbjerg, but these aeolian deposits are not included in the Vejers dunefield, which is defined here to cover only the low-lying Holocene barrier spit deposits.

The coastal dunefield at Vejers occurs in a very high-energy wind environment. Wind data compiled

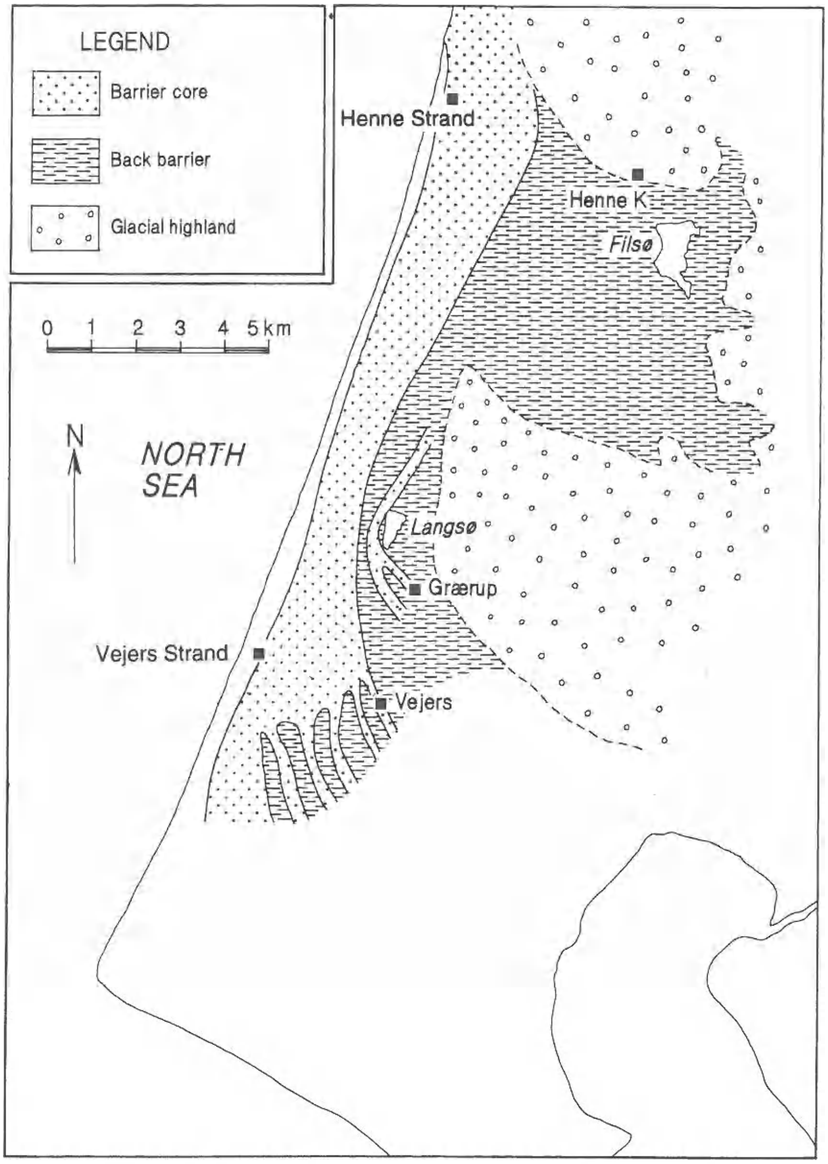

Fig. 3. Reconstruction of the middle–late Holocene barrier spit system that underlies the Vejers dunefield. Based on georadar mapping.

by Frydendahl (1971) for the period 1931–1960 indicate that winds from southwestern, western and northwestern directions dominate throughout the year and the resultant drift direction (RDD) has been calculated to be towards 88° (Fig. 2).

Only small variations in the wind regime occur throughout the year. Mean wind velocity varies between 4.9 m/s and 5.8 m/s, being strongest in the winter, and weakest in the summer. Winds of wind force 9 on the Beaufort scale (19.3 m s^{-1}) and more, however, are rare in the summer, but relatively frequent in the autumn and winter with a maximum of 0.6 percentages of all wind events in October and January. The resultant wind direction is the most variable factor, being southwesterly in winter and autumn and westerly in spring and summer. Storm winds (wind force 9 and more), however, are most common from western or northwestern directions throughout the year. Data from Kristensen and Frydendahl (1991) indicate that significant short-term variations occur in storminess.

The Vejers dunefield receives a moderate amount of precipitation (650 mm a year) with most rain falling in late summer and autumn. Evaporation commonly exceeds precipitation in late spring and early summer, whereas the opposite is the case in the rest of the year. Mean monthly temperatures are between 0.5°C in February and 16.5°C in July.

2.2. Depositional system

The barrier spit deposits. Mapping of the study area by Jessen (1925) revealed the existence of a number of north–south trending gravel-rich beach ridges underlying the aeolian dunefield. These beach ridges occur from the glacial highland at Blåbjerg and southwards to Vejers. Our georadar analyses confirmed the existence of a well-developed system of gravel-rich beach ridge deposits that form the barrier core part of the barrier spit system (Fig. 3; Nielsen et al., 1995).

The barrier core deposits, which are now covered by aeolian dunes, are separated from the mainland to the east by relatively flat-lying terrains with the present lakes Filsø and Langsø. This area constitutes the back barrier area (Fig. 3).

The evolution of the barrier spit system has been much debated (e.g. Jessen, 1925; Jonassen, 1957; Krog, 1979). According to the most recent view (Nielsen et al., 1995) the Filsø basin was a marine bay during the initial stage of the post-glacial transgression and became a lake between late Atlantic and late Subboreal time because of southwards progradation of the gravel-rich barrier spit. In the final stage of barrier spit evolution (1200 A.D. to present) the coarse-grained sediment source was cut off, but much sand was still added to the system. In this period the barrier spit system was covered by large dunes.

The dunefield deposits. The aeolian deposits seem to have formed in the final stage of barrier spit evolution, and datings of the initial aeolian layer in the area indicate that the dunefield is very young (after 300 A.D.). The bulk of the aeolian deposits, however, formed after 1200 A.D.

From historical sources we know that important periods of sand drift and coastal dune formation occurred in Denmark between ca. 1550 A.D. and ca. 1850 A.D. (cf. Pontoppidan, 1769; Brüel, 1918; Hansen, 1957), and in the studied area it appears that sand drift was most severe in the seventeenth century (Brüel, 1918). The historical sources do not mention any episodes of significant sand drift prior to ca. 1550 A.D. It is likely, however, that dunes had already formed in the coastal area, but that the dunes were largely stabilized by vegetation until 1550 A.D. The sand drift in Denmark between 1550 and 1850 A.D. is commonly explained as the result of overgrazing in the coastal area and forest cutting (e.g. Brüel, 1918). Since ca. 1750 A.D., the planting of Ammophila and later forests have gradually damped the sand drift, and the dunefield at Vejers is now stabilized. The partly active dunefield was mapped in 1804 by Videnskabernes Selskab.

In a recent paper Christiansen et al. (1990) describe three periods of dune formation in Denmark (after 400 B.C., after 400 A.D. and between 1450 and 1750 A.D.) and explain coastal dune building in Denmark as a result of low sea level and increased availability of sand in the shore zone. The present study supports these ideas and further suggests that the final period of dune formation in the Vejers area was episodic and that dunefield dynamics primarily were related to fluctuations in the characteristics of the wind.

3. Methods

Information on the large-scale sedimentary accumulation and stratigraphy of the Vejers dunefield is based on several types of investigation. The morphology of the stabilized dune landscape was studied in the field and on topographic maps. The sedimentary characteristics of the deposits were studied in natural profiles, in trenches and in samples from borings. The internal stratigraphy of the aeolian and underlying deposits was examined by georadar mapping (Fig. 4, Nielsen et al., 1995).

Georadar mapping of the coastal deposits was carried out in two periods. During the first period in 1987 and 1989, most of the area was covered by georadar lines (Fig. 4). Several of these lines as well

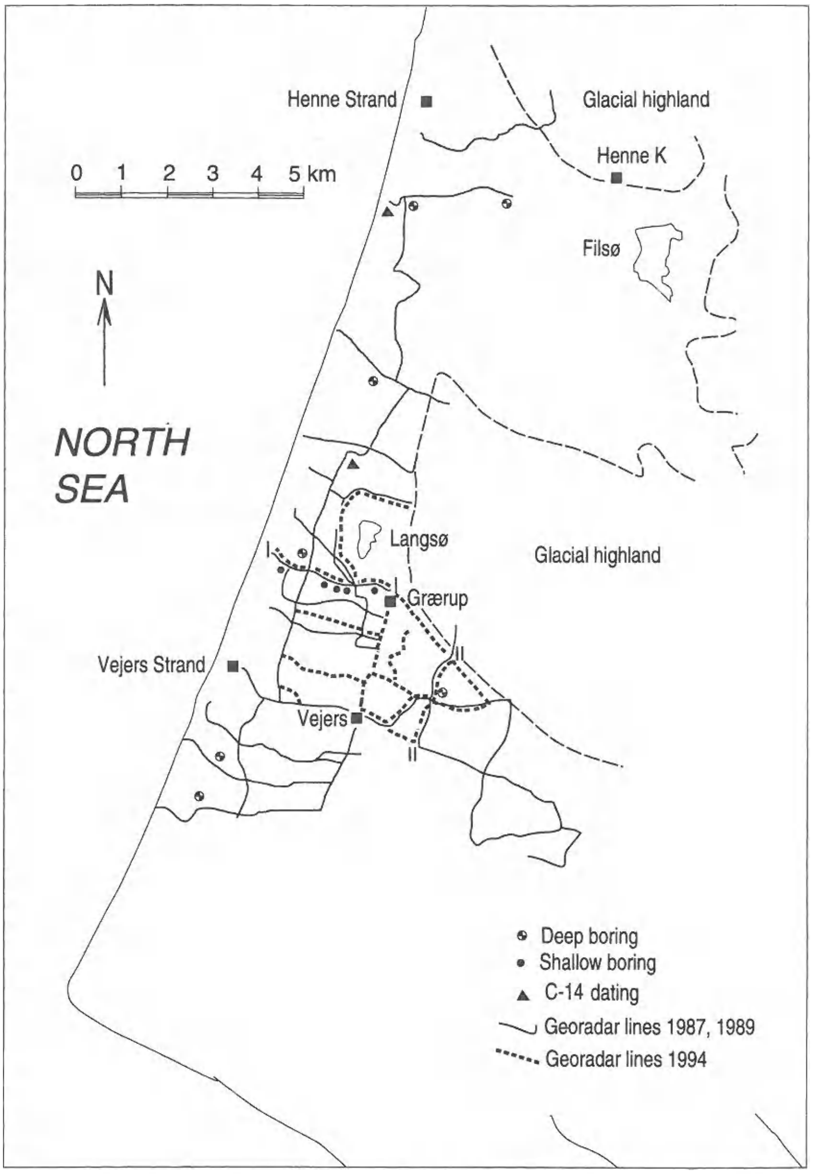

Fig. 4. Georadar lines in the Vejers dunefield. The lines cover the northern and part of the central area in the dunefield. The locations of borings are also shown.

as new lines were studied in 1991 and in 1994 with improved equipment (Fig. 4). The new lines were digitally recorded and collected with 40 MHz and 100 MHz antennae.

4. Dunefield characterization

4.1. Geomorphology

The Vejers dunefield can be divided into northern, central and southern parts (Fig. 2). The northern part (i.e. north of Grærup) is relatively narrow and has an average width of 2 km (Fig. 5). It is composed of an outer belt of more or less continuous coast-parallel dune ridges (width of belt is 200–300 m). Inland, a gradual transition to a belt of parabolic and irregular dunes (width of belt is 1–3 km) occurs. All dunes except those in the outermost dune ridge are completely covered by vegetation. The dunes rest on a slightly undulating aeolian sand cover and the basis of the dunes lies ca. 6 m above sea level. The coast-parallel dune ridges have typical heights between 5 and 10 m, whereas the parabolic and associated irregular dunes have heights between 10 and 20 m. A maximum dune height of 25 m occurs at the transition zone between the two dune belts ca. 4 km south of Hennestrand. The northern part of the dunefield is sharply demarcated by the flat-lying plain of the Filsø basin to the east, and the more southern Langsø basin. The water level in the Filsø basin has been artificially lowered since 1848 (Jessen, 1925).

The central part of the dunefield has a width up to 9 km and is also composed of a relatively narrow

Fig. 5. The dunefield in the northern part of the area, south of Henne Mølleå. View is towards the west (the North Sea). In the foreground part of an elongate, compound parabolic dune occurs (dune height ca. 12 m above sand plain). In the far distance the outermost coastal dunes are visible.

outer belt of transverse dune ridges (dune belt mostly 200–300 m wide, which gradually widens southwards (Fig. 2). Inland a belt of parabolic and irregular dunes (width of belt 0.5–1 km) occurs. Dunes in the outer belt have typical heights between 5 and 10 m (Fig. 6), whereas dunes in the inner belt are typically between 5 and 15 m high. The eastern part of the dunefield is here composed of a large aeolian sand plain with few and scattered small dunes (the Kallesmærsk Hede, Fig. 7), and near the eastern margin of this aeolian sand plain is a very large parabolic dune.

The southern part of the dunefield is composed of two well-defined coast-parallel dune ridges separated by a relative wide and densely vegetated low-lying corridor. Inland a 10.5 km long (in the windward direction) and 1.5–3.5 km wide area of irregular and parabolic dunes has formed. A maximum dune height of 18 m is seen at two parabolic dunes east of Oksby. The central and southern part of the Vejers dunefield is bounded by glacial hills towards the northeast (actually dunes have transgressed part of this landscape), and by saltmarsh areas towards the east along the margin of the Ho Bugt.

4.2. Stratigraphy

Knowledge of the stratigraphy of the barrier spit system is primarily based on georadar mapping. The georadar data indicate that the barrier spit system is composed of lowermost shoal sand deposits overlain by beach and shoreface gravel (to the west) and lagoonal sediments (to the east), and topped by aeolian sediments (cf. Nielsen et al., 1995). The

Fig. 6. Coast-parallel dune ridges in the outer coastal dunes belt north of Vejers Strand (central part of area). Gravel-rich sediments from the underlying barrier core occur at the floor of the 'interdune' corridor. View is towards the northwest. Dune forms reach a height of 5–10 m above the 'interdune' corridor.

aeolian sediments can be divided into a Lower and Upper unit separated by a well-developed peat or organic-rich horizon. The Upper unit can be subdivided into a basal aeolian sand cover, aeolian sand plain deposits, parabolic dune deposits, and coastal dune deposits.

The stratigraphy of the area is also known in some detail from a number of borings (e.g. Jessen, 1925; Geologisk basisdatakort. 1113 IV Varde, D.G.U., 1983). In 1987 seven relatively deep borings were carried out (Ribe Amtskommune et al., 1989). Five of the borings were placed in the barrier core area. The sequence of units was here: at the base 3–6 m of marine sand, then 10–12 m of gravel-rich shoreface and beach deposits and at the top 2–4 m of well-sorted aeolian sand. Two borings were placed in the back-barrier area. The sedimentary succession in the southern boring is interpreted to consist of lagoonal sediments (min. 5 m) overlain by aeolian sediments (8.5 m). The aeolian sediment contains four organic-rich horizons and it is suggested that only the uppermost part belongs to the Upper aeolian unit. Finally, five shallow borings were made along line I in 1994 in order to examine the thickness of the aeolian deposits and the lithology of internal reflectors within the aeolian deposits identified on georadar profiles.

4.3. Age

The age of the aeolian deposits was determined by several methods. The most direct is dating of peat layers at the boundary between the two aeolian units. One such layer from the central part of the dunefield at the abandoned Børsmose gravel pit (Fig. 2) has been dated by the Carbon-14 Dating Laboratory in

Fig. 7. The aeolian sand plain (Kallesmærsk Hede) and distant parabolic dune in Bordrup Plantage. View is towards the east. Width of sand plain seen on photo is ca. 4 km.

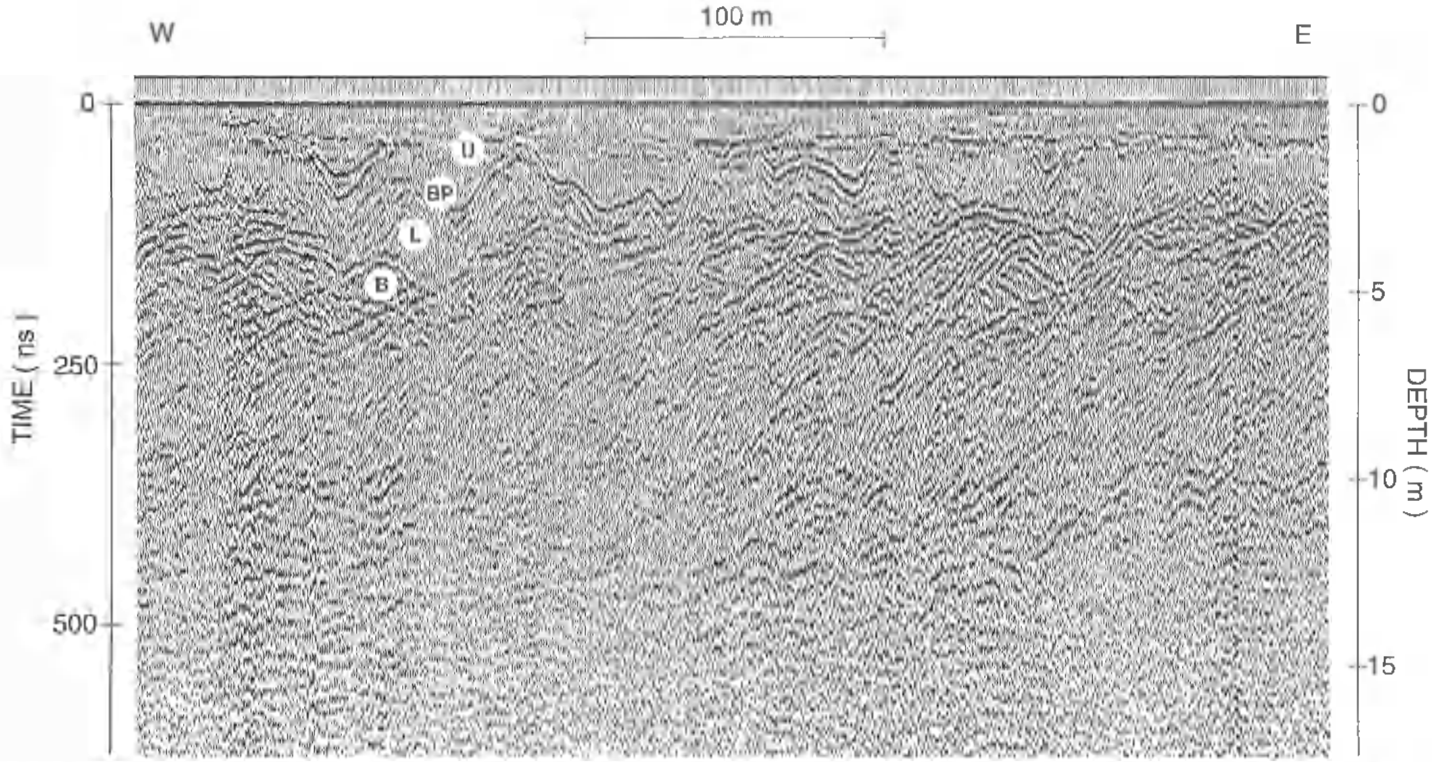

Fig. 8. Georadar profile from barrier core area (line I, data from 1994). Area is presently covered by coastal dunes (not portrayed on georadar profile). Note the well-developed ridge and swale topography in the beach gravels (*B*) and the occurrence of several small dune forms in the Lower aeolian unit (*L*) covered by peat (*BP*). Also the lowermost part (basal sand cover) of the Upper aeolian unit (*U*) is portrayed. Groundwater table occurs ca. 1 m below the surface. Road level in +6–7 m.

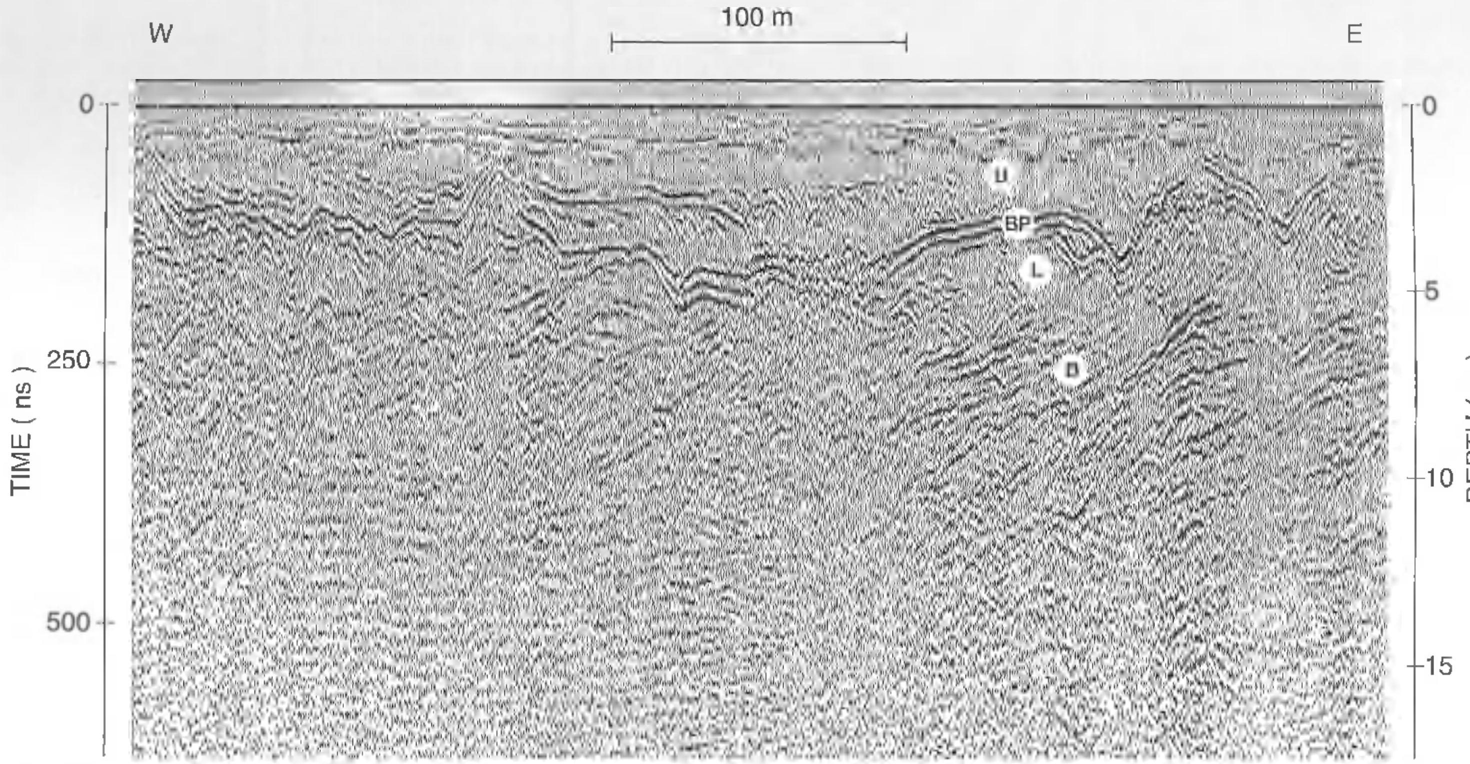

Fig. 9. Georadar profile from barrier core area (continuation of Fig. 8, line 1, data from 1994). A large low lying area in the barrier core topography is filled in with aeolian sediments. A well-developed peat layer (*BP*) marks the boundary between the Lower (*L*) and Upper aeolian (*U*) units. Note the occurrence of inclined reflectors in the Upper unit suggesting eastwards migrating dune forms. Groundwater table occurs ca. 1 m below the surface. Road level in +6–7 m.

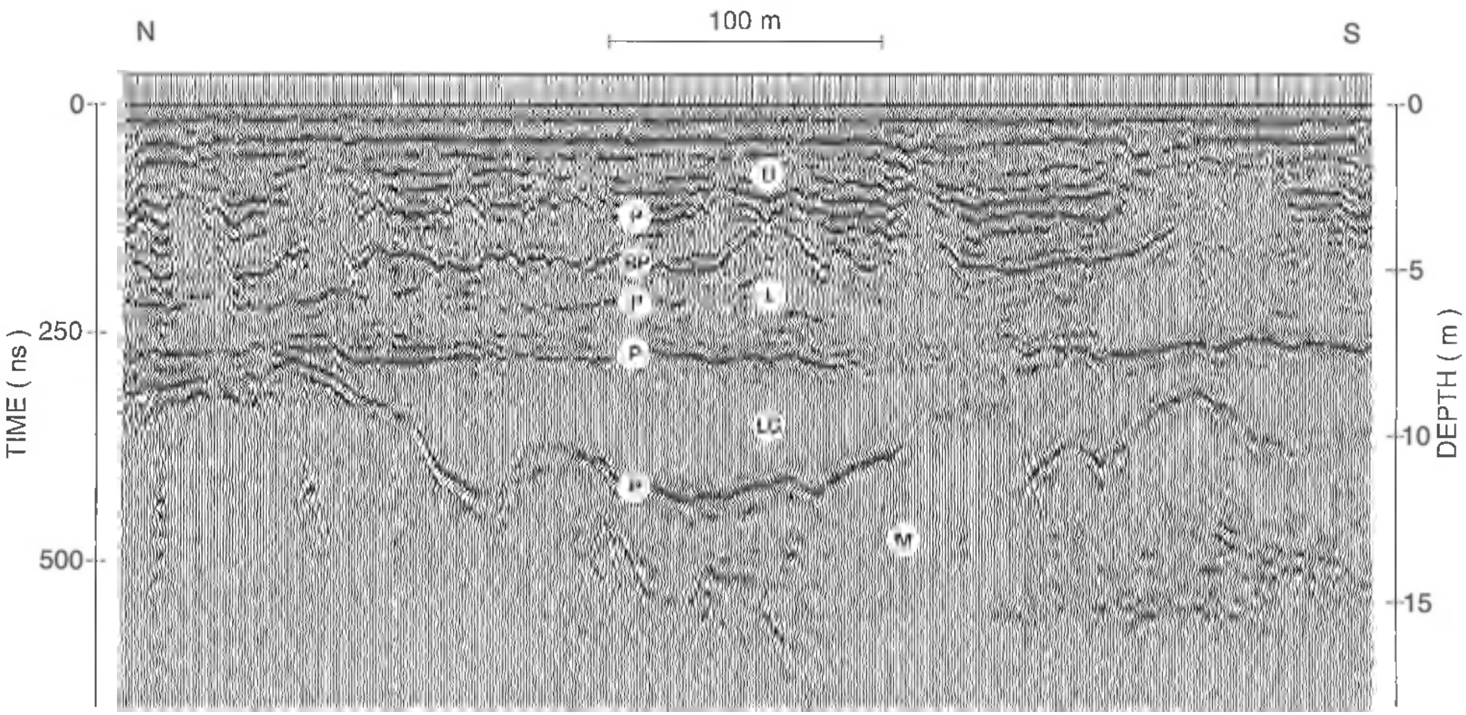

Fig. 10. Georadar profile from back-barrier area (line II, data from 1994). The aeolian succession is underlain by presumed lagoonal deposits (*LG*) and is composed of a Lower unit (*L*) and an Upper unit (*U*) separated by an organic-rich horizon (*BP*) in +4.5–5 m. Additional organic-rich layers are marked (*P*). At the base of the profile marine (*M*) deposits occur. Groundwater table occurs ca. 1 m below the surface. Road level in + ca. 10 m.

Copenhagen to 1080 ± 70 ^{14}C years B.P. (K-5876). This date was calibrated to 890–1015 A.D. (Stuiver and Pearson, 1986). A peat layer from the same stratigraphic level at Henne Mølleå (Fig. 2) has been dated by the AMS ^{14}C-dating facility in Århus to 950 ± 90 ^{14}C years B.P. (AAR-1404). This date was calibrated to 1010–1210 A.D. Finds of a cup and planks from a ship in the gravel-rich barrier core deposits at Vejers (cup dated to 1100–1200 A.D.) also suggest that the formation of the uppermost aeolian dune unit is a very young geological event. Two immature soils in the dune deposits at Henne Mølleå have also been dated by the AMS ^{14}C-dating facility in Århus. These organic-rich horizons were dated to 1680–1955 A.D. and 1685–1955 A.D. indicating that at least the upper part of these dune deposits formed after ca. 1680 A.D. As already stated, historical sources report periods of severe sand drift and dune formation between 1550 and 1850 A.D. The Lower aeolian unit is more difficult to date. Based on pollen analysis, Jonassen (1957) dates the earliest aeolian deposits in the central part of the back-barrier area to approximately 300 A.D. Carbon-14 dating of the basal peat layer and overlying organic-rich layer are, however, needed in order to better document the early phases of sand drift and dune formation in the area.

4.4. Lower aeolian unit

This aeolian unit comprises a 1–5 m thick sheet of fine- to medium-grained sand that drapes the underlying barrier core and back-barrier morphology (Figs. 8–10). The lowermost boundary is defined by a change from marine or lagoonal deposits to aeolian sand. This boundary is most easily detected on georadar profiles in the barrier core area where gravel-rich beach ridge and upper shoreface deposits are sharply overlain by the aeolian deposits. The boundary is undulating and follows the ridge and swale topography of the beach ridge system. In the back-barrier area the lower boundary of the aeolian unit is placed at the top of a well-defined peat layer or organic-rich horizon formed on top of massive fine- and medium-grained silty sand of lagoonal origin. When the basal peat layer is missing or less well developed, the boundary is difficult to place.

The upper boundary of the Lower aeolian unit is an undulating peat layer of ca. +4.5 to +5 m. This peat layer occurs both in the barrier core and back-barrier area. The peat layer has been observed in natural exposures at Henne Mølleå, in two abandoned quarries, in two shallow borings along line I and in a deep boring at line II. The peat layer occurs in swales between relatively small dunes (1–3 m high and 20–40 m wide). The dunes are well portrayed in the georadar profiles as small symmetrical to asymmetrical hummocks with internal reflectors that probably represent dune stratification (Figs. 8, 9, 10). An internal organic-rich horizon is commonly seen on the georadar profiles from the back-barrier area as a semi-continuous reflector (Fig. 10). In the barrier core area similar reflectors are absent or very poorly defined.

The aeolian sand cover on top of the barrier core gravel has a thickness between ca. 1.0 m and ca. 3.0 m. The aeolian sand filling the back-barrier area is significantly thicker with thicknesses between ca. 2.5 m and ca. 5.0 m. The mapped area of the Lower aeolian unit comprises ca. 50 km^2, but the aeolian unit probably continues into the Kallesmærsk Hede and most likely occupies a total area of ca. 120 km^2. This yields a total amount of 250×10^6 m^3 sand in the Lower aeolian unit.

This unit constitutes the initial aeolian deposit on top of the barrier spit system. The aeolian deposits in the barrier core area define a number of low and commonly regularly spaced dune ridges. These low dunes formed intermittently at the shore during a period of overall seaward (westwards) progradation of the shoreline. The small size of the dunes suggests a relatively low influx of sand. In the back-barrier area the aeolian deposits form two subunits separated by an organic-rich horizon. The internal structures of these subunits suggest that the aeolian sediments primarily formed low dunes. The back-barrier area, which originally formed part of a lagoon, dried up probably in connection with a minor sea-level fall and was filled in with aeolian sediments in two episodes. The trapping of this relatively thick succession of aeolian deposits in the back-barrier area resembles the situation described from the arid-climate Guerre Negro barrier island in Mexico (Fryberger et al., 1990). The Lower aeolian unit seems to have formed episodically between ca. 300 A.D. and 1000 A.D.

4.5. Upper aeolian unit

Basal aeolian sand cover. This genetic unit overlies the low dune forms in the Lower aeolian unit sharply (Figs. 8, 9). At many places, and especially in depressions between the underlying dunes, a well-developed Phragmites peat occurs at the boundary. The reflectors on the georadar profiles suggest that the aeolian sand cover primarily is flat-bedded. At some places, and particularly in larger depressions, inclined reflectors suggest the occurrence of eastwards migrating dunes. The upper boundary of this subunit is for practical reasons defined as the road level in +6 to +10 m. This level is almost identical to the level that defines the dune slacks and sand plains of the youngest aeolian deposits.

After the formation of the Lower aeolian unit a phase of dune stabilization and peat formation occurred in low-lying areas around 1000 A.D. The boundary peat lies in ca. +4.5 to +5 m and records the formation of extensive Phragmites swamps probably in relation to a sea level highstand. This eustatic sea level highstand has been documented by Shennan (1992). According to his sea level curve, the sea level around 1000 A.D. was slightly higher than the present sea level.

The subsequent formation of aeolian sand sheets and low dune forms is probably related to the following sea level fall, which culminated around 1500 A.D. (cf. Tanner, 1993). Such a sea level fall would cause an increase in sediment availability in the shorezone, which could trigger an episode of aeolian sand drift (cf. Christensen et al., 1990). Historical sources do not mention any periods of sand drift prior to 1550 A.D., but it has been suggested by Hansen (1957) that aeolian sand drift along the west coast started already around 1200 A.D.

Aeolian sand plain. This genetic unit occurs on top of the basal aeolian sand cover. The aeolian sand plain (sand sheet) (Kallesmærsk Hede and adjoining areas) is a flat aeolian terrain with small and scattered dunes and only local appearance of larger dunes. The sand plain occupies a large portion of the central and eastern part of the Vejers dunefield. The plain lies in ca. 6 to 10 m above sea level and gradually rises towards the east. The sand plain can be divided into a southern part (Kallesmærsk Hede), which is very flat and lies on the windward side of the large parabolic dune in Bordrup Plantage, and a northern part, which has a more undulating topography and is less clearly associated with a large parabolic dune. Small lakes and mires are locally developed.

The aeolian sand plain covers an area of ca. 50 km^2. Its genesis is intimately linked with that of the associated large parabolic dune(s). This is particularly well-demonstrated in the southern part of the plain (Kallesmærsk Hede) where the large parabolic dune in Bordrup Plantage forms the eastward margin of the gradually rising sand plain. The sand plain and associated sediments appear to form the trailing edge deposits of the eastward migrating large parabolic dune(s). The sand plain primarily functioned as an area of bypass, but some sediment was trapped because of vegetation or moisture content of the substrate, and the topographic level of the sand plain surface seems to be controlled by the groundwater level.

Large inland parabolic dune. The inland parabolic dune of Bordrup Plantage occurs 7.5 km from the present dune front at the shore and is of impressive size. The dune has a width of ca. 4 km and length also of ca. 4 km, and is, thereby, one of the largest parabolic dunes in the world (cf. Breed and Grow, 1979). The overall shape of the dune is lobate (cf. Pye, 1993). The axis of the dune runs west–east indicating dune migration towards the east. The nose area of the parabolic dune has heights between 20 m and 30 m (10 and 20 m above the sand plain), whereas the dune arms have heights up to 24 m (14 m above the deflation plain), the southern arm being higher and better developed than the northern arm. Both arms are of a complex morphology and typically secondary parabolic dunes occur. The nose area of the parabolic dune is of a complex morphology, and stoss side, crest and lee side are covered by 1–5 m high dunes and associated scour pits.

The dune is now completely stabilized by a forest (planted between 1853 and 1917) and no natural exposures exist, but the internal structures of the uppermost portion of the dune stoss side and crest have been examined by georadar mapping. These studies indicate that the uppermost 2–3 m of the dune is composed of a series of low- to medium-angle dipping (trough-formed) cross-beds with dip azimuths towards the east. These structures primarily reflect the eastward migration of the superimposed dunes and associated scour pits.

The great size of the Bordrup parabolic dune reflects dune formation in periods of abundant sand supply. Much sand probably originated from erosion of the underlying aeolian deposits, and erosion of this material was probably favoured by periods of reduced plant cover and/or increased storminess. Such events took place between 1550 and 1850 A.D. partly because of overgrazing in the area, and the dune gained its final form in this period of time. Assuming that formation of the parabolic dune was initiated at the coast shortly after 1000 A.D. and that the dune stopped migrating around 1850 A.D., the annual migration rate of the parabolic dune equals a minimum of 8.8 m/yr. If, however, the parabolic dune first was initiated at the shore around 1550 A.D., an annual migration rate of 25.0 m/yr results. These figures appear both very reasonable when compared to measured annual migration rates at the active parabolic dune Råbjerg Mile on the Skagen Spit. This latter dune has migrated ca. 10.0 m/yr since 1887 A.D. (Anthonsen et al., 1996).

Coastal dunes. The 5–20 m high stabilized or semi-stabilized coastal dunes constitute the youngest aeolian unit in the barrier core area. The outer belt is typically composed of two coast-parallel dune ridges, but local topography varies considerably. The dune ridges are separated by a 10–50 m wide 'interdune' corridor in which the gravel-rich sediments of the underlying barrier core deposits occasionally are exposed (Fig. 6). A gradual transition between the second dune ridge and the inner belt of coastal dunes is dominated by parabolic dunes, and commonly the second dune ridge contains incipient parabolic dunes. At many places the outermost coastal dune ridge is overprinted by recent sand shadow dunes which run inland parallel to the dominating storm winds, and can reach sizes up to ca. 3 m (cf. Clemmensen, 1986). At other places large blow-outs have formed in the vegetated dune ridges. Occasionally the internal structures of the dunes can be studied in the walls of the blow-outs. One example occurs just south of Henne Mølleå (Fig. 11). This exposure reveals lowermost gravel-rich beach ridge deposits (forming the base of the blow-out) overlain by ca. 10 m aeolian sand primarily representing the coastal dune unit. The aeolian sand of the coastal dune unit displays large-scale cross-bedding dipping towards the east and contains three internal immature soils. The basal soil (S1) appears horizontal, but the soils S2 and S3 dip towards the east and apparently drape the lee-side of previous dunes. The S2 and S3 soils are simple and well-defined in the dune foreset part,

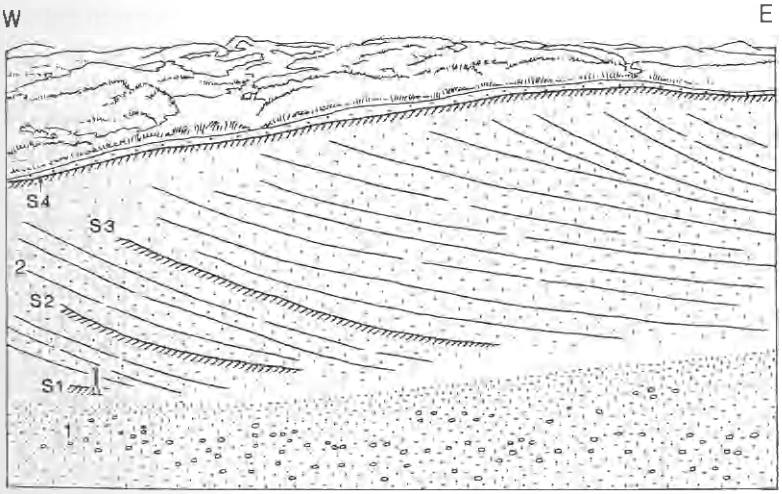

Fig. 11. Sedimentary build-up of large dune exposed in blow-out structure just south of Henne Mølleå (location 4 in Fig. 2). Gravel-rich barrier core deposits (*1*) occur at the base of the blow-out and are overlain by rather structureless aeolian sand and by a basal soil horizon (*S1*). The overlying cross-stratified dune sand (*2*) contains two internal soil horizons (termed *S2* and *S3*) and a very immature soil horizon (*S4*) at the top. The lowermost three (*S1, S2, S3*) soils are characterized by scattered, dark-coloured, branch and root fragments.

but they both split up into two or three sub-soils in the dune toeset part.

The inner belt of the coastal dunes consists of parabolic dunes and dunes of more irregular morphology as well as blow-outs. Just south of Henne Møllеå a relatively large elongate parabolic dune (length 2 km) occurs. The parabolic and related dunes are commonly associated with small sand plains. The parabolic dunes all indicate sand transport towards the east-southeast (90–115°). The internal build-up of these dunes was studied in two abandoned gravel pits in the central part of the dunefield. The gravel deposits lie just below groundwater table, but the overlying aeolian deposits are still partly exposed in some of the quarry walls. The aeolian deposits in both quarries are initiated by a 10–20 cm thick peat layer, which forms the boundary to the underlying Lower aeolian unit. On top of the peat layer follow several metres of aeolian dune sand with some cross-bedding and three immature soils. All soils and in particular the upper two are of an undulating nature and drape an underlying dune topography.

The coastal dune unit records three or four episodes of aeolian accumulation separated by periods of stabilization and soil formation. Each accumulation episode represents the formation of new coastal dune ridges, the reactivation and growth of existing dune ridges and the modification of existing dunes into actively migrating parabolic dunes. All these processes probably took place between ca. 1550 and 1850 A.D.

Events of increased sand drift and dune formation took place in the Vejers area, and at many coastal regions in Denmark (cf. Brüel, 1918). These aeolian accumulation events are primarily thought to be related to periods of increased storminess although the intensity of individual sand drift events probably was enhanced by human influence.

5. Sand budget

5.1. Sand-moving capacity of the wind

Wind data obtained at Blåvandshuk Fyr were used in this study. Blåvandshuk Fyr is situated at the southwestern corner of the dunefield, and wind data of this station are, therefore, directly applicable for the whole dunefield. The wind data at Blåvandshuk Fyr represent the period 1931–60; data were observed as sixteen compass directions, but reduced to eight compass directions in the work by Frydendahl (1971).

The sand-moving capacity of the wind (drift potentials, DP) is defined by the formula of Fryberger (1979):

$$Q \propto V^2(V - V_t)t \qquad (1)$$

where Q is a proportionate amount of sand drift, V is average wind velocity at 10 m height, V_t is impact threshold wind velocity, and t is the time the wind is blowing, expressed as a percentage in a wind summary.

By using this formula the drift potential is calculated to be 1511 vector units (v.u.), or 106 m^3 (m width)$^{-1}$ yr^{-1} (cf. Fryberger et al., 1984), and the resultant drift potential is calculated to be 780 v.u., or 55 m^3 (m width)$^{-1}$ yr^{-1}, directed east (88°). As documented by Anthonsen et al. (1996), the resultant drift potential along the west coast of Jutland varies considerably with time. At Skagen, for example, the resultant drift potential was high (ca. 700 v.u.) and directed towards the east-southeast in 1887 and 1909, whereas it was considerably smaller (ca. 450 v.u.) and directed towards the northeast between 1924 and 1986. The exact causes of this change in the wind environment are not well known, but it is likely that the change of wind environment reflects a shift in latitude of the depression tracks. In the latter part of the former century Greenland had a cold climate and northwest Europa had cool summers (cf. Dansgaard et al., 1975, Lamb, 1977). It is possible that also previous cold periods and especially those between 1550 and 1850 A.D. in 'the Little Ice Age' were characterized by increased storminess (cf. Hansen, 1957; Lamb, 1977) and the dominance of (summer) winds from the west-northwest. This idea is supported because most stabilized parabolic dunes in the area indicate resulting wind flow from the west-northwest.

5.2. Input of sand

Beach sediments form the main source to the dunefield. The actual rate of aeolian sand transport, however, is much less than the potential rate of sand transport, mainly because of insufficient quantities of dry sand in the beach area. Most storms occur during the winter months when the beaches frequently are wet or moist after rain or flooding. The moisture of the beach sediment significantly reduces the rate of sand transport. Winter rainfall on the Oregon coast reduces seasonal sand transport to 36% of its potential (Hunter et al., 1983). If a similar reduction occurred on the Vejers coast, the actual rate of sand transport would lie around 20 m^3 (m width)$^{-1}$ yr^{-1}.

According to Bartholdy and Pejrup (1994) the present net littoral drift of sand along the coast is between 0.5 and 1.0×10^6 m^3 yr^{-1} southwards. If all this sand is eventually deposited on the shore (length of shoreline along the Vejers dunefield is 22 km) and exposed to wind erosion on the dry part of the beach, the rate of aeolian sand transport would be between 22.7 and 45.5 m^3 (m width)$^{-1}$ yr^{-1}. These figures lie between the actual and potential rates of sand transport.

In periods of increased storminess, rates of aeolian transport must have been higher. Increased storminess would probably also cause an intensified longshore drift and an increased availability of sand on the beaches (cf. Hempel, 1980; Hesp and Thom, 1990). Periods of lowered sea level would also cause an increase in sediment availability in the shore zone (cf. Lamb, 1977; Christiansen et al., 1990). Thus, a combination of increased storminess and a low sea level would result in an increased sand flux into the dunefield.

5.3. Sand content in the dunefield

Georadar and morphological analysis indicate that the Vejers dunefield contains ca. 800×10^6 m^3 sand. The Lower unit contains ca. 250×10^6 m^3 sand and the Upper unit contains ca. 550×10^6 m^3 sand. Assuming that the total sand volume was deposited since ca. 300 A.D., an average long-term aeolian sand transport rate of ca. 21 m^3 (m width)$^{-1}$ yr^{-1} results, and assuming that the sand in the Upper unit was deposited since 1000 A.D. we get a sand transport rate of 25 m^3 (m width)$^{-1}$ yr^{-1}. These figures are close to the estimated modern actual rate of sand transport in the area. The figures also compare favourably with modern rate of sand transport in the Alexandria coastal dunefield, South Africa, which is 31 m^3 (m width)$^{-1}$ yr^{-1} (Illenberger and Rust, 1988), and in the Oregon coastal dunefield, which is 34 m^3 (m width)$^{-1}$ yr^{-1} (Hunter et al., 1983). Assuming that the formation of the Upper unit of the Vejers dunefield was restricted to the period 1550–1850 A.D. an average rate of sand transport of ca. 83 m^3 (m width)$^{-1}$ results. Such a short-term rate of sand transport is very high, but not impossible during a period of increased storminess and a high supply of beach sand.

The accumulation of sand in the Vejers dunefield was very rapid, and an average long-term accumulation rate of ca. 4 mm yr^{-1} (accumulation of all sediments in 1700 yr) and a short-term accumulation rate of ca. 14 mm yr^{-1} (accumulation of Upper unit in 300 yr) can be calculated. These figures are very high when compared to the Alexandria coastal dunefield where the average rate of deposition is 1.5 mm yr^{-1} (Illenberger and Rust, 1988).

6. Dunefield dynamics and conclusions

The aeolian sediments at Vejers formed in two main time intervals (Fig. 12). The first phase of dune formation occurred after 300 A.D. and ended around 1000 A.D., when widespread peat formation took place. The second phase of dune formation occurred between 1550 and 1850 A.D. as documented by historical sources. This period was relatively cold ('the Little Ice Age') and characterized by a sea level lowstand (cf. Christiansen et al., 1990; Tanner, 1993). The lowering of wave base made more sand available for landward transport (by marine processes), the shoreline prograded and wide areas were exposed for aeolian erosion. The eroded material was transported inland by strong winds and deposited in the Vejers dunefield. Because of frequent erosion by storm surges, strong winds and/or overgrazing, the foredune ridges were transformed into transgressive dunes (cf. Hesp and Thom, 1990). The first main phase of dune formation in the area seems to have occurred during a period of relatively low sea level, but during this phase only low dune forms developed. A sea level lowstand around 800 A.D. has been documented by Shennan (1992) and Tanner (1993). The climate was probably cold and stormy.

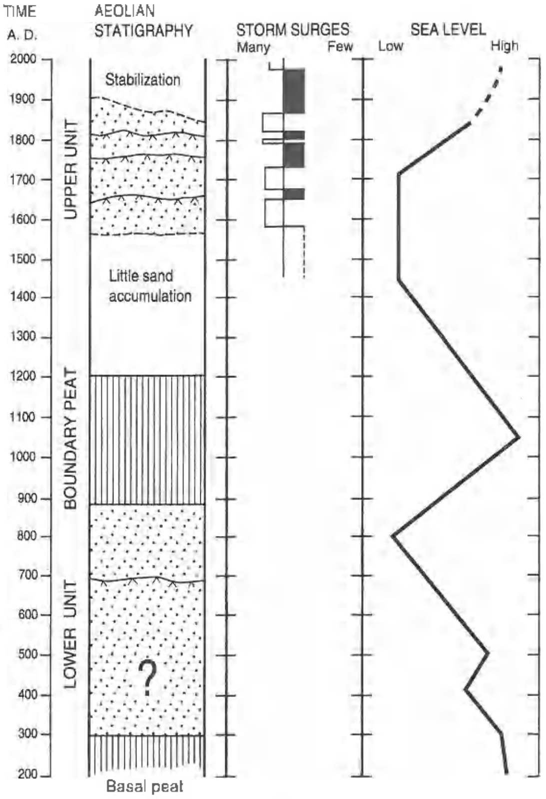

Fig. 12. Summary scheme showing aeolian stratigraphy in the Vejers dunefield in relation to sea level and storminess (frequency of storm surges). Sand drift and dune formation in the Vejers area took place in two main phases separated by a period of peat formation around 1000 A.D. The occurrence of several soils indicate episodic growth of the dunefield. The sea level curve is based on data in Tanner (1993). The curve on storm surge frequency is based on data in Gram-Jensen (1991).

The long-term evolution of the Vejers dunefield seems primarily related to sea level. The composition of the aeolian sediments clearly indicates, however, that in both time intervals periods of dunefield growth alternated with periods of dunefield stabilization and formation of organic-rich horizons. Therefore, periods of high sand supply and a rapid accumulation rate alternated with periods of less sand supply and a slow or negligible accumulation rate. A maximum of four periods of dunefield accumulation have been identified during the final phase (1550–1850 A.D.) of dune formation, but unfortunately none of these accumulation periods have been dated exactly. These smaller-scale events of dune activity may be related to short time intervals of increased storminess during 'the Little Ice Age', whereas periods of dune stabilization occurred during periods of reduced storminess. Because few data on wind characteristics exist in this time interval, data on storm surge frequency along the Danish west coast are here used as an indicator of storminess. These data indicate that four periods of exceptional storminess occurred during 'the Little Ice Age' (1590–1650 A.D., 1660–1730 A.D., 1790–1800 A.D., and 1820–1870 A.D., Fig. 12). Judged from the number of storm surges, storminess was particularly high in the first two periods, which matches well with historical records on large-scale sand drift in these years.

The overall aeolian stratigraphy at Vejers corresponds well with the stratigraphy of other areas in Denmark. Christiansen et al. (1990) mention three main phases of sand drift and dune formation in northwestern Jutland. The latter two phases match well with the two main phases of dune formation observed in the Vejers dunefield. In Denmark large-scale sand drift and dune formation first took place during 'the Little Ice Age'. Previous phases of sand drift only resulted in the formation of low dunes. A similar pattern has been observed on the Sefton coast in northwest England (Pye and Neal, 1993). Here, the final and most important phase of dune formation took place between ca. 1200 and ca. 1850 A.D. In the Netherlands an older dune complex (the Older Dunes) and a younger dune complex (the Younger Dunes) are recognized (e.g. Klijn, 1990). Deposition of the older dune complex started as early as 2800 B.C. and continued up to Roman times (Klijn, 1990). Deposition of the younger dune complex took place between ca. 1000 A.D. and ca. 1850 A.D., and three periods of dune activity are described.

Pye and Neal (1993) explain the formation of active dunes in the Middle Ages as a result of exceptional storminess and rapid coastal erosion. Klijn (1990) also relates the formation of the Younger Dunes in the Middle Ages to increased coastal erosion caused by a higher storm surge frequency. Klijn (1990) suggests that these periods of increased storm surge frequency took place during rising sea level.

In the present study dune formation took place on a coastline that was prograding. The long-term evolution of the Vejers dunefield is mainly related to sea level with main phases of dune formation occurring during late Holocene sea level lowstands, or possibly during early transgressive phases. The short-term depositional dynamics of the dunefield were primarily controlled by variations in the characteristics of the wind. Episodes of transgressive dune formation were probably caused by a combination of factors including overgrazing, storm surge erosion, and blow-out formation.

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References

Anthonsen, K.L., Clemmensen, L.B. and Jensen, H.J., 1996. Evolution of a dune from crescentic to parabolic form in response to short-term climatic changes: Råbjerg Mile, Skagen Odde, Denmark. In: N. Lancaster (Editor), Response of Aeolian Processes to Global Change. Geomorphology, 17: 63–77, this issue.

Bartholdy, J. and Pejrup, M., 1994. Holocene evolution of the Danish Wadden Sea. Senckenbergiana Marit., 24: 187–209.

Breed, C.S. and Grow, T., 1979. Morphology and distribution of dunes in dunefields observed by remote sensing. In: E.D. McKee (Editor), A Study of Global Dunefields. U.S. Geol. Surv., Prof. Pap., 1052: 253–302.

Brüel, J., 1918. Klitterne i Vestjylland og på Bornholm. Nordisk Forlag, København, 133 pp.

Christiansen, Ch., Dalsgaard, K., Møller, J.T. and Bowman, D., 1990. Coastal dunes in Denmark. Chronology in relation to sea level. In: Th. W.M. Bakker, P.D. Jungerius and J.A. Klijn (Editors), Dunes of the European Coasts. Catena Suppl., 18: 61–70.

Clemmensen, L.B., 1986. Storm-generated eolian sand shadows and their sedimentary structures, Vejers Strand, Denmark. J. Sediment. Petrol., 56: 520–527.

Dansgaard, W., Johnsen, S.J., Reh, N., Gundestrup, N., Clausen, H.B. and Hammer. C.U., 1975. Climatic changes, Norsemen and modern man. Nature, 255: 24–28.

Fryberger, S.G., 1979. Dune forms and wind regime. In: E.D. McKee (Editor), A Study of Global Dunefields. U.S. Geol. Surv., Prof. Pap., 1052: 137–169.

Fryberger, S.G., Al-Sari, A.M., Clisham, T.J., Rizvi, S.A.R. and Al-Hinai, K.G., 1984. Wind sedimentation in the Jafurah sand sea, Saudi Arabia. Sedimentology, 35: 21–41.

Fryberger, S.G., Krystinik, L.F. and Schenk, C.J., 1990. Tidally flooded back-barrier dunefield, Guerrero Negro area, Baja California, Mexico. Sedimentology, 37: 23–43.

Frydendahl, K., 1971. Danmarks Klima, I. Vind. Det Danske Meterorologiske Institut, Klimatologiske Meddelelser, 1, 184 pp.

Gram-Jensen, I., 1991. Stormfloder. Danish Meteorologcal Institute, Scientific Report, 91-1, 121 pp.

Hansen, V., 1957. Sandflugten i Thy og dens indflydelse på kulturlandskabet. Geogr. Tidsskr., 56: 69–92.

Hempel, L., 1980. Zur Genese von Dürengenerationen an Flachküsten. Beobachtungen auf den Nordseeinseln Wangerooge und Spiekeroog. Z. Geomorphol., N.F., 24: 428–447.

Hesp, P.A. and Thom, B.G., 1990. Geomorphology and evolution of active transgressive dunefields. In: K. Nordstrom, N. Psuty and B. Carter (Editors), Coastal Dunes, Form and Process. John Wiley, New York, pp. 253–288.

Hunter. R.E., Richmond, B.M. and Alpha, T.R., 1983. Storm-controlled oblique dunes of the Oregon Coast. Bull. Geol. Soc. Am., 94: 1450–1465.

Illenberger, W.G. and Rust, I.C., 1988. A sand budget for the Alexandria coastal dunefield, South Africa. Sedimentology, 35: 513–521.

Jessen, A., 1925. Beskrivelse til Geologisk Kort over Danmark. Kortbladet Blaavandshuk, Danm. Geol. Unders., I rk., 16, 76 pp.

Jonassen, H., 1957. Bidrag til Filsøegnens naturhistorie. Medd. Dansk Geol. Foren., 13: 192–205.

Klijn, J.A., 1990. The younger dunes in the Netherlands; chronology and causation. In: Th.W.M. Bakker, P.D. Jungerius and J.A. Klijn (Editors), Dunes of the European Coasts. Catena Suppl., 18: 89–100.

Kristensen, L. and Frydendahl, K., 1991. Danmarks vindklima fra 1870 til nutiden. Havforskning fra Miljøstyrelsen, 2, 57 pp.

Krog, H., 1979. Late Pleistocene and Holocene shorelines in western Denmark. In: E. Oele, R.T.E. Schuttelhelm and A.J. Wiggers (Editors), The Quaternary History of the North Sea. Acta Univ. Ups., Uppsala, pp. 75–83.

Kuhlman, H., 1968. Æolisk geomorfologi (klitmorfologi). Københavns Universitet, Geografisk Centralinstitut, 36 pp.

Lamb, H.H., 1977. Climate: Present, Past and Future, Vol. 2. Climatic History and the Future. Methuen, London, 835 pp.

Nielsen, S.T., Andreasen, F. and Clemmensen, L.B., 1995. The middle and late Holocene barrier spit system at Vejers, Denmark: structure and development. Bull. Geol. Soc. Denm., 42: 105–119.

Pontoppidan, E., 1769. Den danske altas eller Kongeriget Danmark. Tomus 5, København.

Pye, K., 1993. Late Quaternary development of coastal parabolic megadune complexes in northeastern Australia. In: K. Pye and N. Lancaster (Editors), Aeolian Sediments. Ancient and Modern. Spec. Publ. Int. Assoc. Sedimentol., 16: 23–44.

Pye, K. and Neal, A., 1993. Late Holocene dune formation on the Sefton Coast, northwest England. In: K. Pye (Editor), The Dynamics and Environmental Context of Aeolian Sedimentary Systems. Geol. Soc. Spec. Publ., 72: 201–217.

Ribe Amtskommune, Miljøministeriet, Danmarks Geologiske Undersøgelse and Skov og Naturstyrelsen, 1989. Redegørelse om muligheden for indvinding af ral og sand i Vejers-området, 54 pp.

Shennan, I., 1992. Holocene sea-level changes in the North Sea region. In: M.J. Tooley and I. Shennan (Editors), Sea-Level Changes. Basic Blackwell, pp. 109–151.

Stuiver, M. and Pearson, G.W., 1986. High-precision calibration of the radiocarbon time scale A.D. 1950–500 B.C. Radiocarbon, 13: 969–979.

Tanner, W.F., 1993. An 8000-year record of sea-level change from grain-size parameters: data from beach ridges in Denmark. Holocene, 3: 220–231.

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