A 7000-year record of coastal evolution, Vejers, SW Jutland, Denmark

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The Holocene coastal lowland at Vejers in western Jutland has formed during the last 7000 years. The lowland is composed of a large, NNE-SSW trending spit system associated with minor and only locally developed strandplain or beach ridge systems. The main spit and back-barrier system is bounded to the north and east (inland) by old moraine landscapes (Varde Bakkeø). Most of the coastal system and also large parts of the adjacent moraine landscape is covered by aeolian sand. In this study one of the minor strandplain systems is investigated. This system is developed at the south-western margin of the old moraine landscape at Grovsø, a lake near Vejers. The Holocene sedimentary evolution of this latter system is evaluated on the basis of data from two closely situated cores and Ground-Penetrating Radar (GPR) mapping. Both cores consist of a lowermost unit with marine sediment, a middle unit with lake-aeolian sand and an uppermost unit with aeolian sandplain deposits. Peat layers and peat-rich paleosols are common. These peat-rich horizons are dated by the Accelerator Mass Spectrometry (AMS) radiocarbon technique, while the intervening sand layers are dated by Optically Stimulated Luminescence (OSL). Combined evidence from the sedimentological and chronological studies of the cores and the GPR survey, indicate that the area was first transgressed at about 5100 BC. During the subsequent period (5100–2700 BC) relative sea level rose about 5 me-ters, the strandplain prograded, and small coastal dunes formed. During this progradational event a large strandplain lake formed behind the frontal dune ridge and this lake was filled primarily by aeolian sand. Aeolian sand drift may have been most intense around 3000 BC. This first period of large-scale aeolian activity ended some time before 2300 BC with formation of a peat-rich paleosol. Aeolian activity, however, was soon re-established and resulted in the formation of a large sandplain with small dunes. Aeolian sand movement and accumulation, however, was punctuated by periods of landscape stabilisation and peat-rich paleosol formation. Changes from landscape stabilisation to dune field activity took place at about 2300 BC, 1450 BC, 800 BC, and 650 BC. Aeolian accumulation at the study site terminated at about AD 0, but other evidence indicates renewed aeolian activity in the dune field after AD 300 and between AD 1100 and 1900. The chronology of some of these aeolian activity phases are synchronous with cooling events in the North Atlantic region suggesting that climatic change strongly influenced dune field dynamics.

Keywords: Holocene, coastal evolution, aeolian, climate change

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The coastal lowland at Vejers in SW Jutland (Fig. 1) forms the northernmost part of the Holocene barrier and back-barrier system along the south-eastern and southern part of the North Sea. This part of the North Sea was flooded between 7000 and 6000 BC, and when relative sea-level rise decelerated after 4000–3000 BC the barrier systems began to form (Beets *et al.* 2000). Due to the high-energy wind climate in the southern North Sea region, dune fields have developed on top of most barriers (e.g. Bird, 1990) and in the Vejers area thick accumulations of aeolian sand are present (Clemmensen *et al.* 2001a).

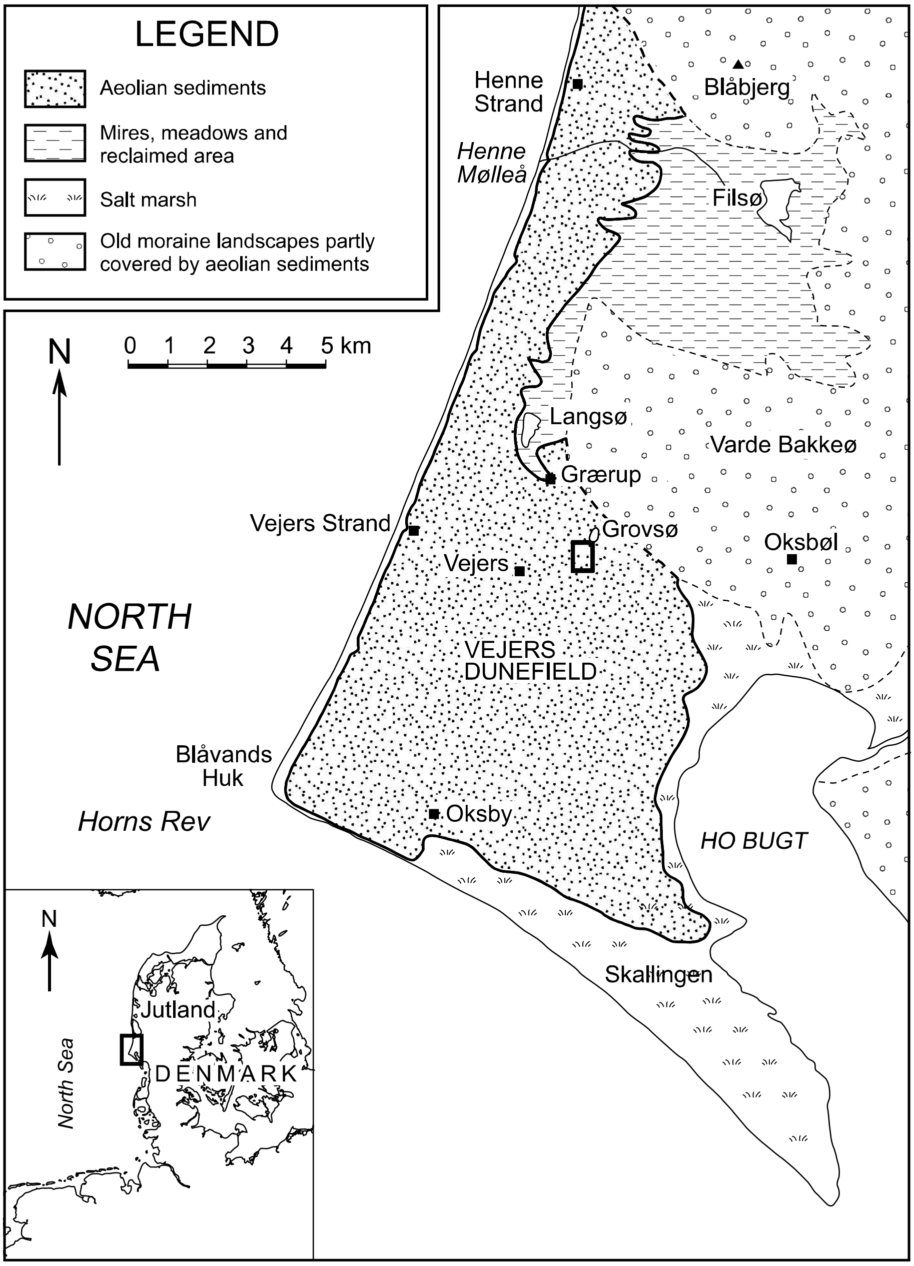
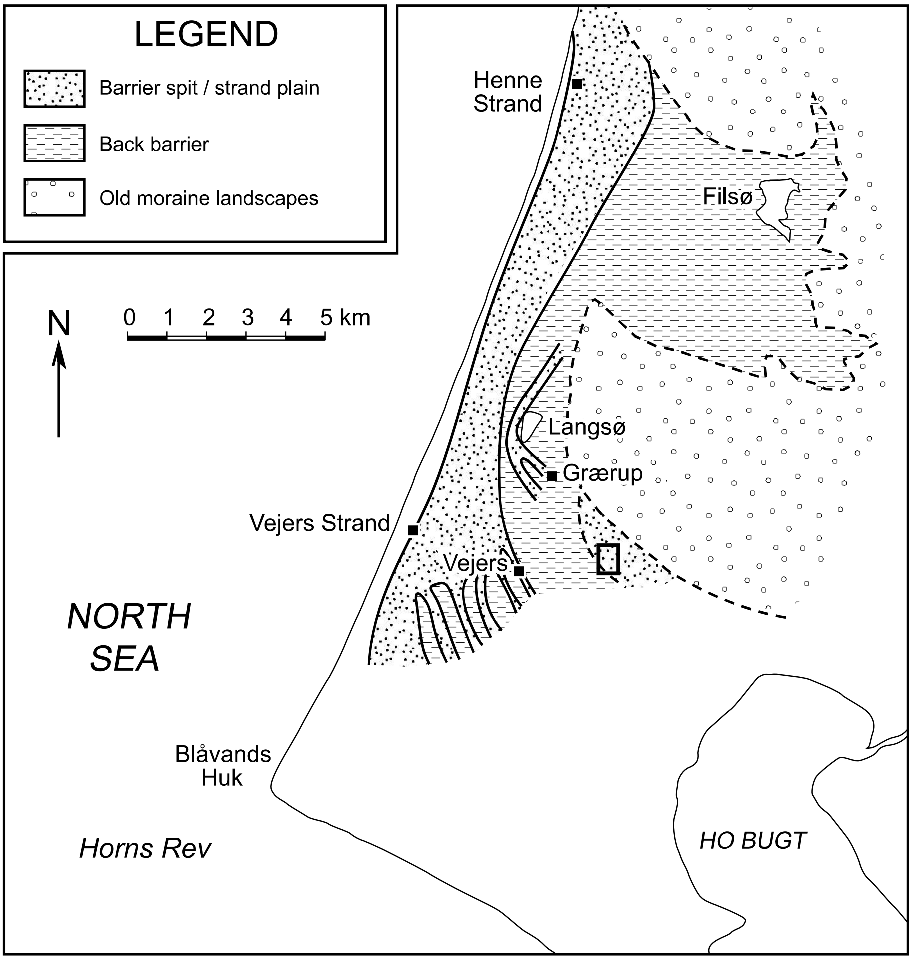


Fig.1. Location of the study area. The Vejers dune field faces the high-energy North Sea and has developed under the influence of strong winds from westerly directions. At the inner margin of the dune field large aeolian sandplains have developed, and up to 14 m of aeolian sediment has accumulated since the post-glacial transgression reached the area at about 5100 BC. This study describes the sedimentary characteristics of the aeolian sandplain and underlying coastal deposits at Grovsø.

The general characteristics of the Mid to Late Holocene coastal lowland at Vejers on the west coast of Jutland (Figs 1, 2) were described by Jessen (1925), but more detailed morphological and sedimentological studies were first carried out by Nielsen *et al.* (1995) and Clemmensen *et al.* (1996). Their descriptions of the Vejers spit system were based on Ground-Penetrating Radar (GRP) mapping, sedimentological studies of trenches and natural outcrops and data from borings. Their reconstruction of the system shows a large spit system linked to an old moraine landscape towards the northeast. Between the spit and the old moraine landscape towards the east (inland) is a low-lying area, the back-barrier system. Locally, and especially at the south-western margin of this old moraine landscape, a minor standplain or beach ridge system has developed. Aeolian sand now covers the complete spit and back-barrier system as well as parts of the adjacent moraine landscape.

More recently a boring was carried out at Grovsø, a lake near the south-western margin of the old moraine landscape (Figs 1, 2), sampling 16 m of Holocene coast-

Fig. 2. Simplified map of the gravel-rich spit prior to the formation of the Vejers dune field. Map based on Clemmensen *et al.* 1996.



al deposits primarily of aeolian origin. The Holocene landscape evolution at this site was reconstructed on the basis of stratigraphical studies of the core succession, Accelerator Mass Spectrometry (AMS) ^{14}C dating and Optically Stimulated Luminescence (OSL) dating of key horizons (Clemmensen *et al.* 2001a). Their study showed that coastal dune formation in this area of initial strandplain formation began somewhere between 5000 BC and 4000 BC. Aeolian sand movement and sand accumulation continued episodically until the end of the 19th century (Clemmensen *et al.* 2001a; Clemmensen & Murray 2006). The study also yielded information on early changes in the shoreline configuration of the area. To support this first study of sequence stratigraphy and landscape evolution at the inner fringe of the coastal system at Grovsø, it was decided to carry out a second boring in the area with improved technique, and to obtain additional samples for AMS and OSL dating.

In this paper we describe the sedimentary characteristics of the strandplain-aeolian system at Grovsø. We interpret the sedimentary environments of the identified sedimentary units, place the two core successions in a stratigraphical and palaeogeographical context by use of GPR data, and reconstruct the local landscape evolution since about 5000 BC on the basis of 11 AMS and 15 OSL dates. Emphasis in the work is on the evolution of the coastal aeolian system.

Study area and geological setting

Blåvands Huk and the submarine Horns Rev divide the south-eastern part of the North Sea (the German Bight) into a southern area with tidal influence, and a northern area dominated by wave action (Figs 1, 2). The west coast of Jutland north of Blåvands Huk faces the high-energy, wave-dominated North Sea and is developed as a linear simplified coast. Hilly terrains of Pleistocene deposits facing the sea have been eroded and the resulting material has been

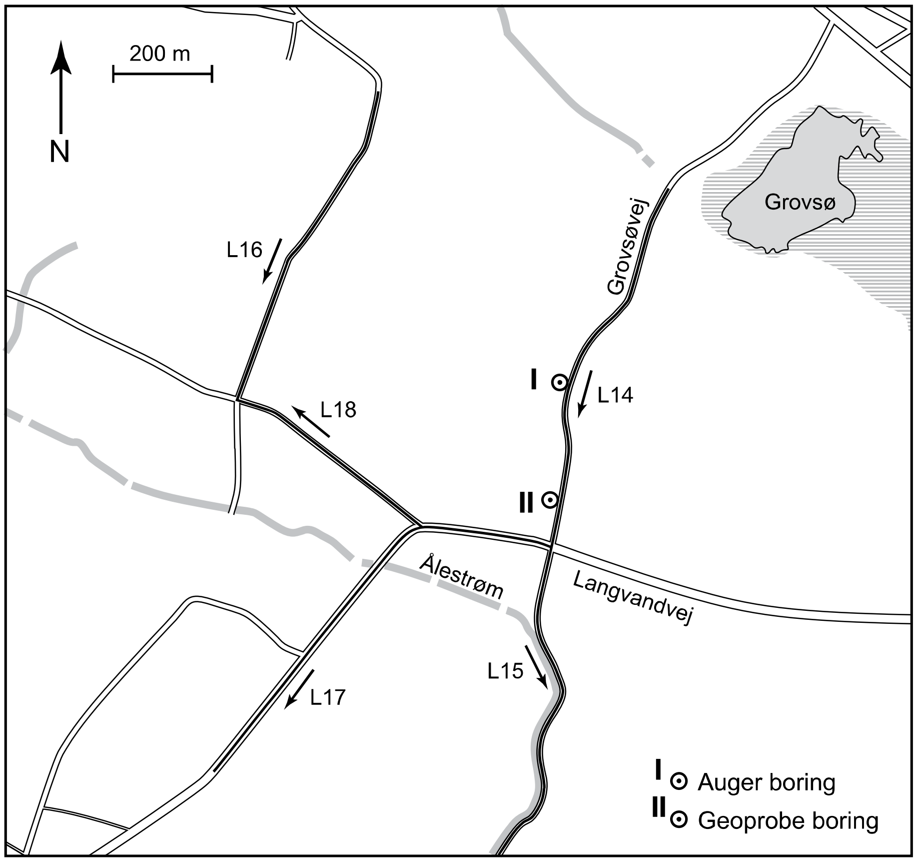


Fig. 3. GPR lines (50 MHz antenna) in the study area near Grovsø, Vejers coastal system, Western Jutland.

transported by littoral currents and accumulated in front of bays in the form of various barrier systems. On the south-western side of the moraine landscape at Varde (Varde Bakkeø) a very large Holocene spit system has developed (Figs 1, 2). It is composed of a northern segment (the Vejers spit) and a southern, still active, spit segment (Skallingen).

The formation of the Vejers spit system was initiated after the mid-Holocene transgression when sea level became sufficiently high to cause intense wave-erosion of the old moraine landscape at Blåbjerg. A strong southward directed littoral drift was established, resulting in southwards growth of the spit system. At about the same time the strandplain system at Grovsø formed. Studies in the Filsø basin indicate that the initial marine transgression took place at the end of pollen zone VI or around 6000 BC (Jonassen 1957). An AMS dating of a transgressive peat at Grovsø gives an age of about 5100 BC for this initial transgression (Clemmensen *et al.* 2001a).

After the initial transgression relative sea level continued to rise, the strong southwards directed littoral drift was increased, and southwards growth of the Vejers spit continued. This growth, possibly in connection with initial aeolian sand movement towards the east or inland, caused a gradual isolation of the Filsø basin from the sea. The transition from lagoon to fresh-water lake took place at the end of pollen zone VII or around 4000 BC according to Jonassen (1957). In the following period the water level in Filsø rose and reached a maximum of 7.2 m a.s.l in the beginning of pollen zone IX or at about 600 BC (Jonassen 1957).

Southwards growth of the gravel-rich Vejers spit system continued until about AD1100–1200, when the coarse-grained sediment supply to the spit system rather suddenly was cut off. After this date only sand was added to the spit, but southwards growth continued and the spit was linked to Blåvands Huk no later than AD1600 (Aagaard *et al.* 1995).

The study site at Grovsø is situated at the innermost margin of the strandplain system that developed at the south-western fringe of the old moraine landscape (Fig. 2). During the early evolution of the Grovsø strandplain system the area was exposed to high-energy wave action, but in time the Grovsø area became more and more sheltered from wave action by the southwards growing Vejers spit. Both systems are now welded together and covered by aeolian sediments.

The Vejers area lies in a cool, temperate climate region with monthly temperatures varying from 0° C in February to 16–17° C in July. Annual precipitation is around 800 mm with highest values in October (data period 1961–90; DMI Technical Report 00-11). The area experiences a high-energy wind climate dominated by onshore (westerly) winds, (Clemmensen *et al.* 1996). The present North Sea shoreline of the Vejers coastal system runs NNE-SSW. The shoreline is wave-dominated and waves have a significant height of about 1.2 m. The spring tidal range is about 1.2 m and storm surges reach 2–2.5 m above mean sea level.

Sedimentary characteristics of the strandplain-aeolian system

Ground-Penetrating Radar (GPR) survey

GPR mapping of the coastal deposits around Grovsø was carried out in three periods: 1987–89, 1991–94 (Clemmensen *et al.* 1996; Clemmensen *et al.* 2001a) and in 2003 (data given here). The GPR survey in 2003 used the Sensors and Software PulseEKKO™ 100 system and includes 5 lines recorded with a 50 MHz antenna (Fig. 3), while 13 lines were recorded with a 100 MHz antenna (Fig. 4). The lines were recorded on existing roads in the area. Most of these roads have altitudes of 10–11 m a.s.l., and topographic variations are insignificant. The 13 lines recorded with 100 MHz antenna are laid out in a grid (Fig. 4) in order to facilitate the reconstruction of 3D topography in selected levels in the coastal system.

The lines were processed using the Sensors and Software PulseEKKO™ version 4.2 processing software. The 100 MHz lines contained only little secondary noise, while the 50 MHz lines were affected by a rather strong secondary signal that made it difficult to isolate the primary signal. In spite of this, the new 50 MHz data are in good agreement with previous data recorded with a 40 MHz antenna (Clemmensen *et al.* 2001a). The ground-water level on the GPR lines lies approximately 1 m under the surface. This interpretation is confirmed by a manual boring on line 1, where the groundwater level was reached at a depth of 1.25 m.

The 100 MHz lines primarily show the sedimentary architecture of the uppermost 10-12 m of the succession, while the 50 MHZ lines also contain information on the architecture of deeper-lying strata. The succession portrayed on the GPR profiles can be divided into a number of genetic units (cf. Clemmensen *et al.* 2001a): A lower marine unit grading upwards into coastal dune deposits capped by a marked peat horizon, a middle lake-aeolian unit bounded upwards by a new pronounced peat or peat-rich paleosol, and an uppermost aeolian sandplain unit with three peat-rich paleosols (Figs 5, 6). Thus in total 5 well-defined peat layers or peat-rich paleosols (here named layer a, b c, d, and e) are identified in the GPR profiles, and these layers are used as stratigraphic marker horizons when establishing a chronological framework for the sedimentary evolution of the study area. The intervening aeolian units are named A, B, C, D, and E (Figs 5, 6).

Core Studies

The lithology and stratigraphy of the Holocene strandplain system at Grovsø can be described in more detail on the basis of information from the two cores (Figs 7, 8). Both core sites were situated on a large aeolian sandplain now stabilized by vegetation. The Grovsø I core site was placed just south of Grovsø where the GPR data suggest that the innermost margin of underlying marine deposits is situated, while the Grovsø II core site was placed 220 m to the SSW where the GPR data indicate the existence of a relatively thick unit of underlying marine deposits (Fig 5).

The Grovsø I core was conducted by an auger in 1998 and reached a depth of 16 m (Clemmensen *et al.* 2001a). Core lithology was described in the field and samples from selected horizons were brought back to the laboratory for grain-size analysis. Samples were taken in the field for AMS dating (complete core succession) and for OSL dating (uppermost 4 m), (Tables 1, 2).

The Grovsø II core was obtained in 2003 by Geoprobe equipment and reaches a depth of 16 m. The Geoprobe coring technique enables good and reliable records of lithology and stratification. Actually two cores were recovered at the Grovsø II core site, with a separation of 1 m. The first core was taken in order to study the lithology of the sedimentary succession and was brought back to the laboratory for more detailed studies. Here a detailed facies log was drawn and samples were taken for grain-size analysis, while peat-rich horizons were sampled for AMS dating (Table 1). In the second Grovsø II core sand was collected for OSL-dating in well-defined stratigrahical levels. These sand samples were collected in light-proof steel tubes. On return to the laboratory, light-exposed material was removed from the ends of the tubes under low-level orange light, and the unexposed sediment treated in the usual manner using HCl, H_2O_2 and HF (heavy

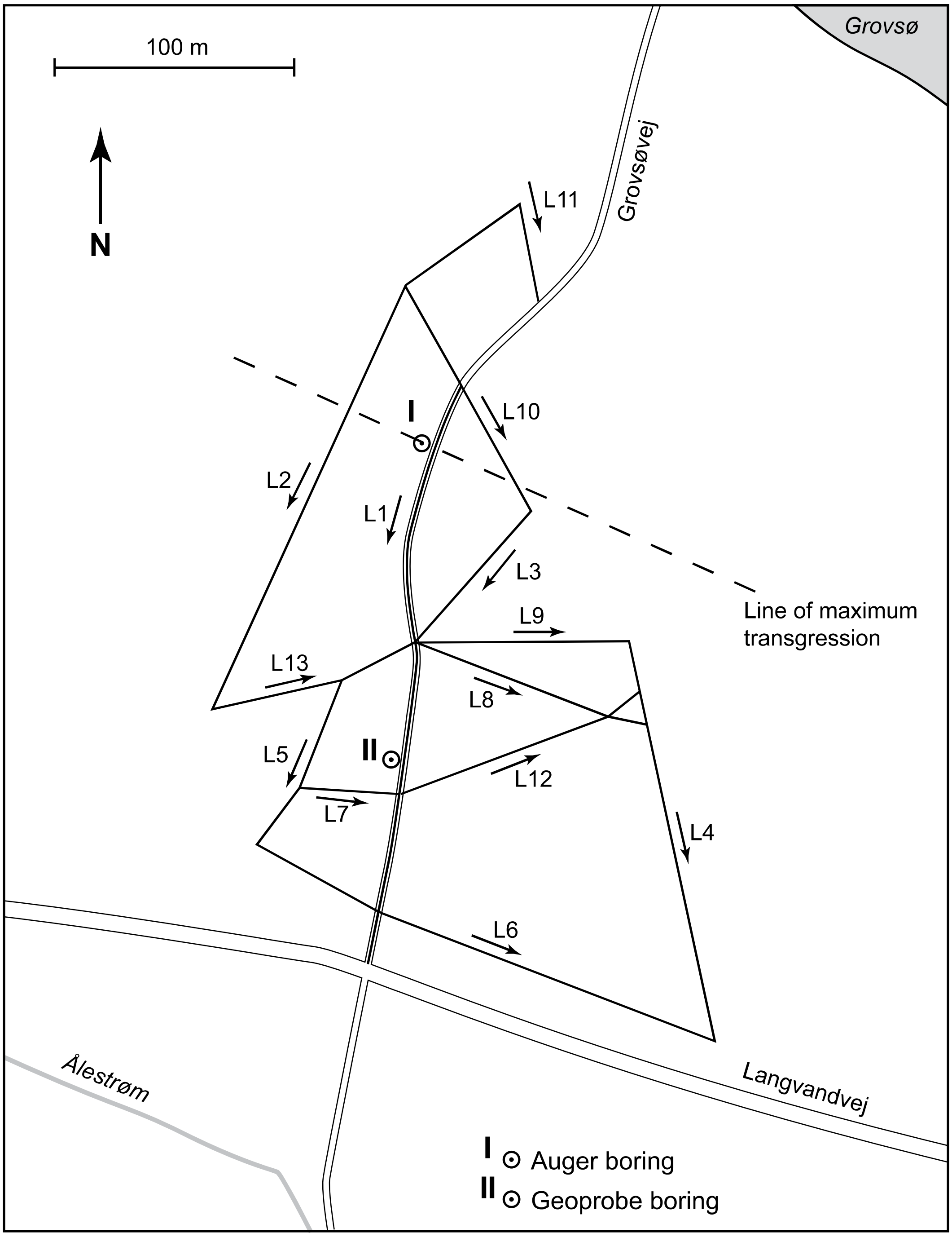


Fig. 4. GPR lines (100 MHZ antenna) in the study area near Grovsø, Vejers coastal system, Western Jutland. Line of maximum transgression of the Littorina Sea is indicated.

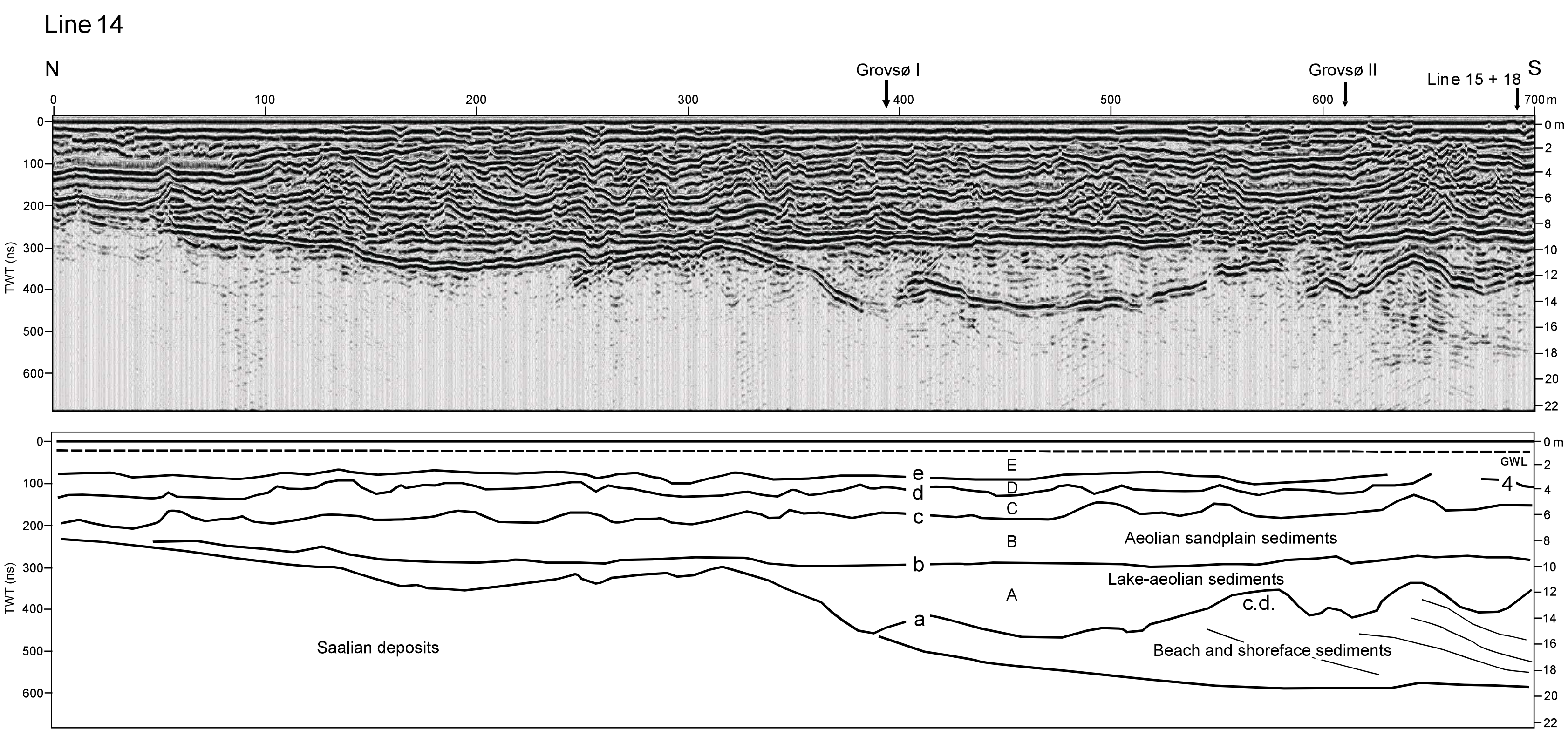


Fig. 5. Stratigraphical architecture of the coastal deposits in study area based on GPR mapping (line 14) with 50 MHZ antenna (see also Clemmensen *et al.* 2001a). Peats and peaty paleosols (a–e) and aeolian units (A–E) are indicated. The lowermost beach and shoreface sediments are overlain by low coastal dunes (c.d.). Present surface in 10.5 m a.s.l.

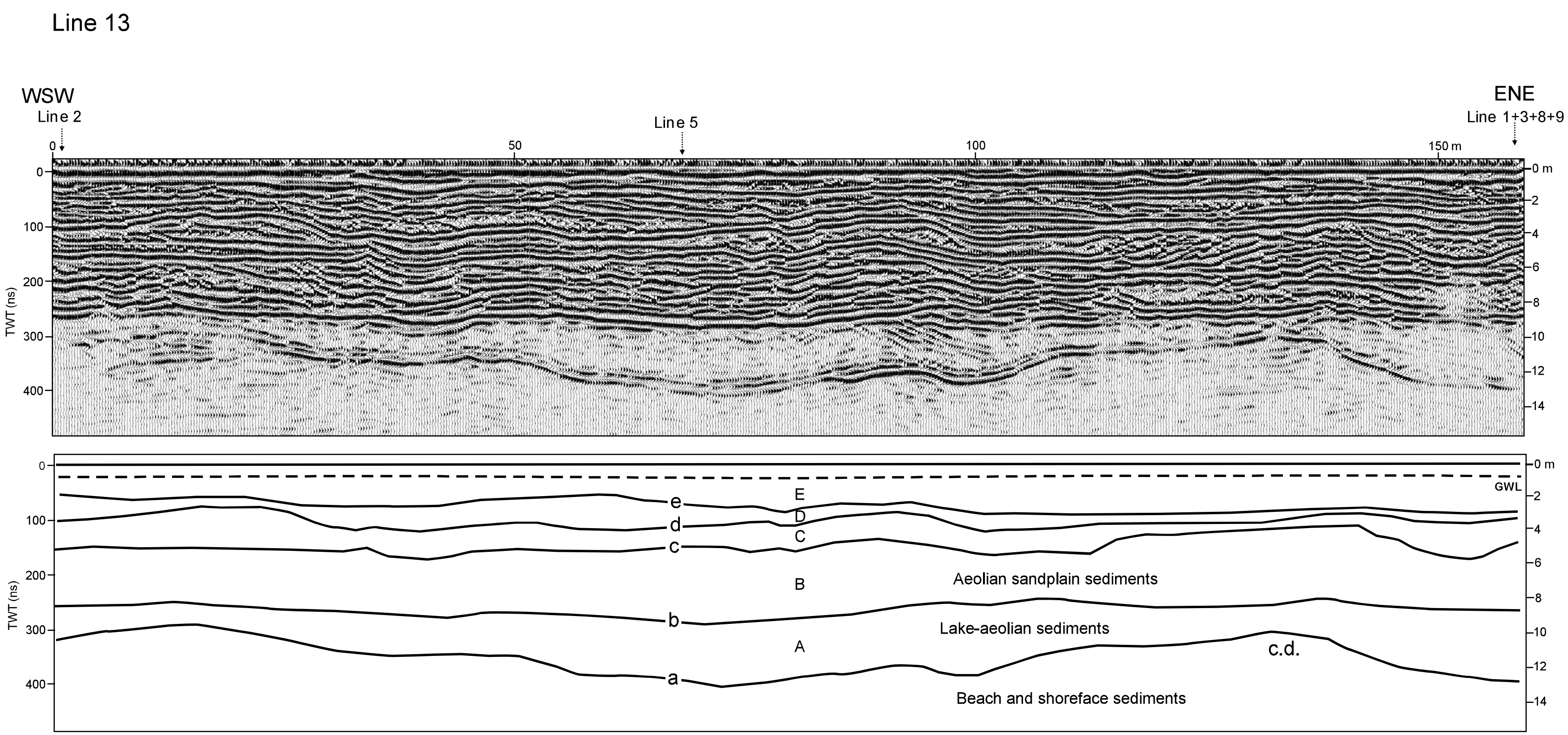


Fig. 6. Stratigraphical architecture of the coastal deposits in the study area based on GPR mapping (line 13) with 100 MHz antenna. Peats and peaty paleosols (a–e) and aeolian units (A–E) are indicated. The lowermost beach and shoreface sediments are overlain by coastal dunes (c.d.). Present surface in 10.5 m a.s.l.

Table 1. AMS ^{14}C dates of organic rich horizons, Grovsø I and Grovsø II cores, Vejers coastal system, Denmark.

Sample no.	Sample unit	^{14}C age (yr BP)	Calibrated age* (1 sigma range)	Mid point of calibrated age interval, yr
Grovsø I				
AAR-4876	Rootlet horizon, 1.95 m	2255 ± 50	BC 390 – 210	300 BC
AAR-4598	Peaty palaeosol (e), 2.6 m	2480 ± 50	BC 770 – 410	590 BC
AAR-4599	Peaty palaeosol (d), 3.2 m	2495 ± 35	BC 765 – 520	643 BC
AAR-4600	Peaty palaeosol (e2), 5.1 m	3075 ± 50	BC 1410 – 1260	1335 BC
AAR-4601	Peaty palaeosol (c1), 6.0 m	3290 ± 45	BC 1680 – 1520	1600 BC
AAR-4602	Peaty palaeosol (b), upper part – 7.9 m	3850 ± 45	BC 2405 – 2205	2305 BC
AAR-4603	Peaty palaeosol (b), lower part – 8.1 m	3890 ± 50	BC 2470 – 2290	2380 BC
AAR-4604	Peat (a), 11.7 m	5185 ± 50	BC 4040 – 3960	4000 BC
AAR-4605	Transgressive peat, 13.2 m	6145 ± 60	BC 5230 – 4960	5095 BC
Grovsø II				
AAR-8668	Peaty palaeosol (e), 3.5 m	2510 ± 75	BC 800 – 520	660 BC
AAR-8669	Peaty palaeosol (d), 4.0 m	2631 ± 41	BC 829 – 792	811 BC
AAR-8670	Peaty palaeosol (c), 6.0 m	3173 ± 45	BC 1500 – 1405	1453 BC
AAR-8671	Peaty palaeosol (b), 8.0 m	3569 ± 42	BC 1980 – 1870	1925 BC
AAR-8672	Peaty palaeosol (b), 8.5 m	3845 ± 50	BC 2360 – 2200	2280 BC
AAR-8673	Peat (a), 11.5 m	4143 ± 49	BC 2870 – 2620	2700 BC

*Calibrated years in calendar years have been obtained from the calibration tables in Stuiver et al. (1998)

Table 2. OSL dates of sand in the Grovsø I and Grovsø II cores. Vejers coastal system, Denmark

Sample	Description	Depth, m	Age, yr	Dose, Gy	(n)	Dose rate,Gy/ka	w.c. %
Grovsø I							
994702	Aeolian sand (unit E)	1.25	250 BC ± 150	1.74 ± 0.03	16	0.77 ± 0.04	17
994703	Aeolian sand (unit E)	2.4	270 BC ± 190	1.66 ± 0.10	31	0.73 ± 0.04	18
994704	Aeolian sand (unit C)	3.55	290 BC ± 150	1.72 ± 0.03	29	0.75 ± 0.04	21
Grovsø II							
034709	Aeolian sand (unit E)	3.0	AD 110 ± 120	1.41 ± 0.05	18	0.75 ± 0.04	27
034710	Aeolian sand (unit D)	3.9	780 BC ± 200	1.68 ± 0.06	18	0.60 ± 0.03	29
034711	Aeolian sand (unit C)	5.0	1500 BC ± 300	2.21 ± 0.06	18	0.63 ± 0.04	28
034712	Aeolian sand (unit B)	6.9	1400 BC ± 200	2.16 ± 0.04	17	0.64 ± 0.04	27
034713	Lake-aeolian sand (unit A)	9.3	3100 BC ± 400	2.60 ± 0.09	18	0.51 ± 0.03	28
034714	Lake-aeolian sand (unit A)	11.2	3000 BC ± 500	2.71 ± 0.13	18	0.54 ± 0.04	28
034715	Coastal dune/beach sand	12.1	2700 BC ± 500	2.87 ± 0.12	18	0.61 ± 0.06	28
034716	Shoreface sand	15.6	4600 BC ± 500	4.80 ± 0.16	18	0.73 ± 0.04	28

Notes: (n) is the number of aliquots used to determine the Dose; water contents (w.c.%) for Vejers dune field and Grovsø I are observed values, Grovsø II are saturated. A 4% uncertainty is assumed on all water content values. Dose rate includes a comsic ray component based on Prescott and Hutton (1994).

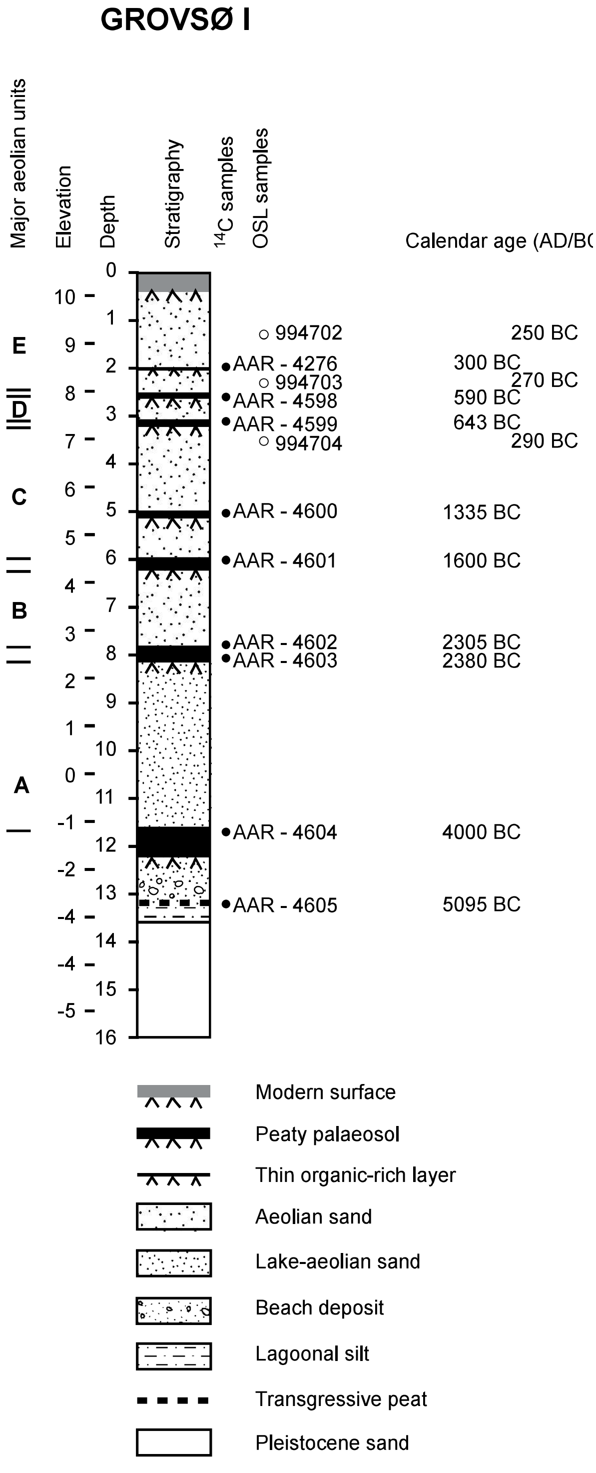


Fig. 7. The Grovsø I (auger) core with aeolian units A–E and the stratigraphical position of ^{14}C and OSL samples (see also Figs 5 and 6).

liquids were not used) to separate 180–250 mm quartz grains. Quartz purity was checked by the absence of signal during infra-red stimulation. Equivalent doses were measured using the SAR procedure (Murray and Wintle, 2000) on Risø TL/OSL readers (Bøtter-Jensen *et al.* 2000) fitted with integral $^{90}Sr/^{90}Y$ beta sources delivering accurately known dose rates of ~0.1 Gy.s^{-1}. Our mesaurement protocol used a preheat at 200°C for 10 s, and a cut heat to 160°C. These measurement conditions give rise to negligible thermal transfer (data not shown). To confirm that our protocol was able to measure a known dose administered in the laboratory before any thermal treatment of the sample, 21 previously unused aliquots were bleached twice at room temperature using 40 s of blue stimulation, with a 10 ks pause between the two stimulations. The mean ratio of observed to given dose was 0.96±0.01, which is considered satisfactory.

Dose rates were calculated from measured radionuclide concentrations (Murray *et al.*, 1987) and the conversion factors given in Olley *et al.* (1996). The effects of water content and grain size were taken into account as described by Aitken (1985), and the cosmic ray contribution calculated from Prescott and Hutton (1994). The luminescence data are summarized in Table 2.

The Holocene succession at the Grovsø I site unconformably overlies older (presumably Middle Pleistocene) sand deposits, separated by a thin lag deposit at the contact. The basal 0.2 m of the Holocene succession is composed of presumed lagoonal silt and peat, whereas the remaining part of the succession is composed of sand intercalated with peat-rich layers (Fig. 7). By comparison with the GPR data, the sand in the core is thought to represent three different aeolian facies: coastal dune sand overlying beach sand (13.2–12.2 m), lake-aeolian sand (11.6–8.1 m), and aeolian sandplain deposits (7.9–0 m), (Fig. 7). Well-developed peats or peaty paleosols separate the sand units, and the sandplain unit contains internal paleosols.

The presumed aeolian sand-sheet and coastal dune deposits are composed of very well sorted fine-grained sand, thus supporting the given interpretation. Also the lake-aeolian unit is composed of clay-free, fine-grained sand with mean values (D_{50}) between 0.215 and 0.240 mm (Table 3). The dispersion (D_{90}–D_{10}) is as low as 0.150–0.190 mm indicating that the sediment is very well sorted (cf. Saye & Pye 2006). These grain-size parameters are almost identical to those from modern dune sediments in the area (Table 3) suggesting that most sand in the lake-aeolian unit was of aeolian origin. The grain-size parameters of shore sand from the modern lake Grovsø, which has a high influx of aeolian sand, are identical to those in the lake-aeolian unit (Table 3).

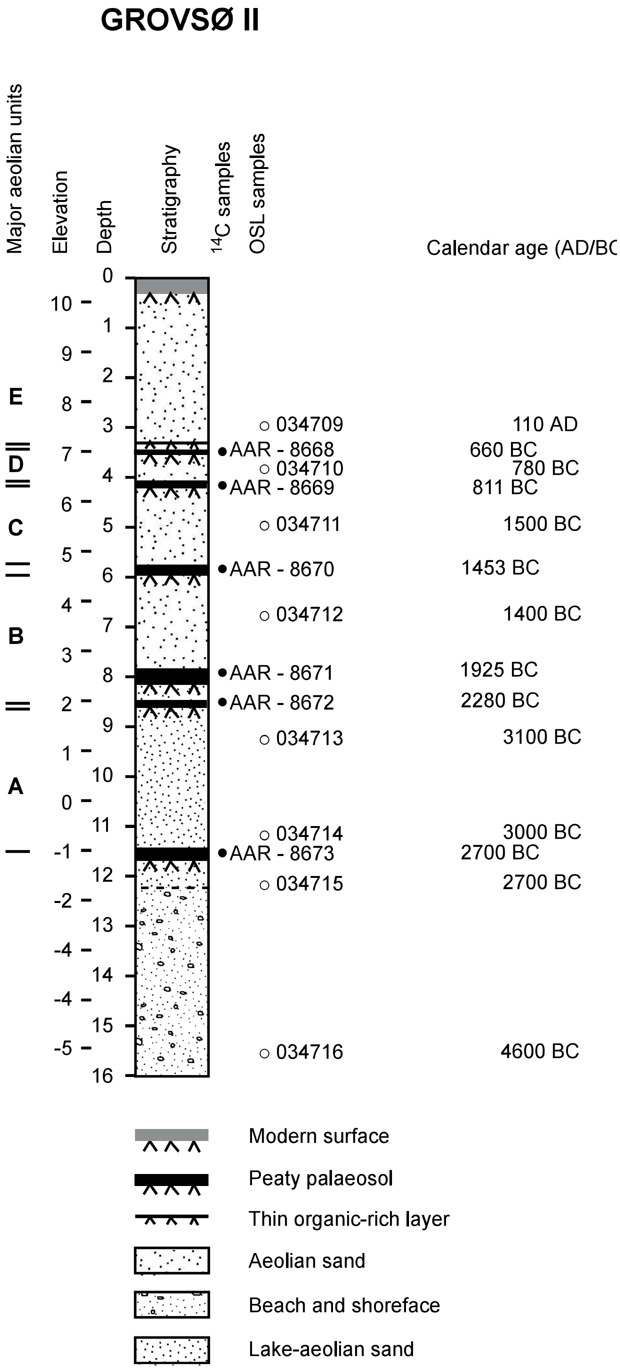


Fig. 8. The Grovsø II (Geoprobe) core with aeolian units and the stratigraphical position of ^{14}C and OSL samples (see also Figs 5 and 6).

The Grovsø II core does not reach the underlying Pleistocene deposits. The core succession is composed of Holocene sand intercalated with peats or peat-rich paleosols (Fig. 8). By comparison with the GPR data, the sand in the core is interpreted to represent the following facies: coastal dune sand overlying marine sand (16.0–11.6 m), lake-aeolian sand (11.4–8.6 m), and aeolian sandplain deposits (7.9–0 m). Well-developed peats or peaty paleosols separate the sand units, and the sandplain unit contains internal paleosols. Sand in the aeolian units has grain-size characteristics similar to those in the Grovsø I core, while the marine sediment at the base of the core is slightly more coarse-grained.

The five marker horizons observed in the GPR profiles were interpreted as peat-rich horizons, an interpretation confirmed by the core study. The lowermost peat marker (marker layer a) which caps the coastal dune deposits, occurs in 12.2–11.6 m depth in the Grovsø I core and in 11.7–11.4 m depth in the Grovsø II core. The next peat-rich paleosol (marker layer b) which separates the lake-aeolian unit from the overlying sandplain unit, occurs in 8.1–7.9 m depth in the Grovsø I core. In the Grovsø II core there are two closely associated peat layers, one at 8.6–8.5 m depth and one at 8.1–7.9 m depth. The next three peat-rich horizons (marker layers c, d, and e) are all situated in the upper aeolian sandplain unit. Marker layer c is composed of two closely associated peaty layers (at 6.1 m and 5.1 m depth) in the Grovsø I core, but only of one peaty layer (at 5.9 m depth) in the Grovsø II core. Marker layer d occurs in 3.2 m depth in the Grovsø I core and in 4.2 m depth in the Grovsø II core, and marker layer e occurs in 2.6 m depth in The Grovsø I core and in 3.5 m depth in the Grovsø II core.

Thus core studies support the stratigraphical subdivision made on the basis of the GPR data. Five aeolian units are identified on top of the basal beach-coastal dune deposits: The lake-aeolian unit (unit A), a lowermost aeolian sandplain unit (unit B), and three aeolian sandplain units (units C, D and E). In the following the age and formation of the upermost five aeolian units will be discussed on the basis of AMS dating of the peaty layers, supplemented by OSL dating of the sand units.

Chronology

The age of the Holocene succession in the study area is well constrained by AMS and OSL dating (Tables 1, 2; Figs 7, 8). The chronology of the two cores comprising the oldest aeolian deposits is given by 15 AMS and 11

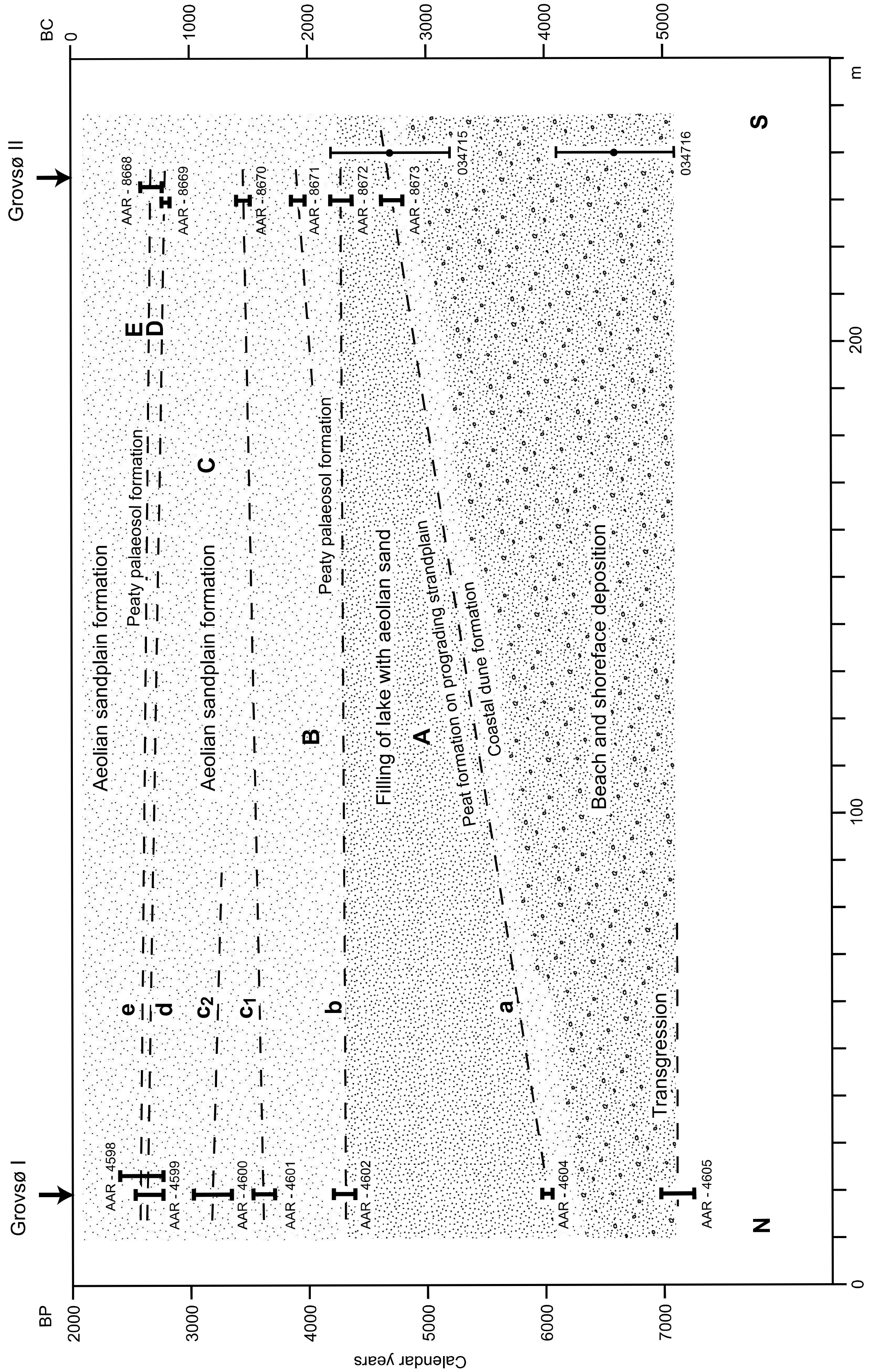


Fig. 9. Sedimentary evolution of the investigated coastal system at Grovsø between 5000 BC and 0 BC.

OSL dates. The chronology of the youngest aeolian sediments, which comprises the now stabilized dune deposits, is given by Clemmensen *et al.* (1996) and Clemmensen & Murray (2006).

The AMS dates as given in ^{14}C years (BP) years, as well as in calibrated years, with one standard deviation (Table 1). In the text only calibrated years are used (BC/AD) and for the sake of simplicity the age of a certain layer is given as the mid-point of the calibrated age interval.

The accuracy of the AMS dates is high, with most calibrated ages having age intervals (one standard deviation) that vary less than ± 5% from their mid-point values. Only samples having ^{14}C ages between 2750 and 2450 BP yield calibrated age intervals that vary by 10-12 % from their mid-point values; this is due to the existence of the well-known plateau in the radiocarbon calibration curve in this time interval (van Geel *et al.* 1996).

Uncertainties on the OSL ages lie between 6 and 11%. OSL ages are absolute, in that they do not rely on a calibration curve, and the uncertainties are dominated by experimental uncertainties in the measurements of radionuclide concentrations and equivalent dose. Uncertainties in water content contribute ~ 3% on average.

The AMS dates are given in calibrated (BC/AD) years and are generally based on organic-rich material from the uppermost part of a sampled paleosol, thereby recording the final phase of stabilization immediately before the soil (landscape) was buried in sand (Table 1). In one sample material for dating was taken from the top and the base of a relatively thick, peaty paleosol (samples AAR-4602 and AAR-4603). The two dates suggest that soil formation and landscape stabilization in this case was of very short duration. Assuming that this was the general pattern, it would imply that aeolian sand movement was the dominant scenario during the evolution of the sandplain. This suggestion, however, has to be supported by more systematic AMS dating of the peaty paleosols.

The AMS dates from these paeosols form the main framework on which the description of the chronology of the sedimentary evolution of the coastal system has been based. Correlation between paleosols in the two cores indicates that, generally speaking, landscape stabilization and soil formation during the evolution of the aeolian sandplain system were isochronous events (Fig. 9). The AMS data also indicate that the formation of the peat on top of the initial coastal dunes was time-transgressive (Fig. 9).

The OSL samples were taken in order to obtain supplementary information on the timing and duration of the aeolian activity phases. The OSL dates are also expressed in years BC/AD, (Table 2). In the Grovsø I core we only have OSL dates from the uppermost 4 m. Here the OSL dates in general terms support the chronology given by the AMS dating. However, due to the limited number of OSL dates it is not possible to estimate the duration of the individual aeolian activity phases. In the Grovsø II core there are OSL dates from all aeolian units, but only unit A (lake-aeolian sand unit) is dated both at the base and at the top. Also here there is general agreement between the chronology established by the AMS dates and that given by the OSL dates, but the number of OSL dates is still insufficient to provide an estimate of the duration of the aeolian activity phases.

Sedimentary evolution of the strandplain-aeolian system

This discussion of sedimentary evolution between 5000 BC and AD 0 is based on the new GPR data supplemented by previous 40 MHz GPR data, the borings and the AMS and OSL dates (Fig. 9). Prior to the mid-Holocene transgression most of the area formed an undulatory landscape originally developed during the Saale glaciation and later modified by a number of terrestrial processes. The seaward part of this landscape

Table 3. Grain-size parameters

Sample	Mean (D_{50})	Dispersion (D_{90}-D_{10})
Lake-aeolian sediment, 8.8 m, Grovsø I	0.230 mm	0.190 mm
Lake-aeolian sediment, 9.5 m, Grovsø I	0.240 mm	0.190 mm
Lake-aeolian sediment, 11.5 m, Grovsø I	0.215 mm	0.150 mm
Sediment from modern Lake Grovsø	0.210 mm	0.125 mm
Sediment from modern frontal dunes*	0.201 – 0.376 mm	0.156 – 0.336 mm

*Data in Saye & Pye (2006).

was levelled by ravinement proceses during the first post-glacial transgression.

The age of the mid-Holocene transgression in south-western Jutland is dated to about 5100 BC by the transgressive peat seen near the base of the Holocene succession in the Grovsø I core (Fig. 7). Judging from core data the relative sea level was approximately 3 m below present sea level at this stage. The dated transgression level constitutes a new fix-point in the southern North Sea region during a time when the relative sea-level rise decelerated and many coasts were transformed into stabilized or prograding barrier systems (Beets *et al.* 2000). Relative sea level rose in the study area after the initial transgression, but due to high rates of sediment supply a strandplain formed and began to prograde towards the SSW. The middle part of these prograding beach and shoreface deposits is dated to about 4600 BC (sample 034716), while backshore/coastal dune deposits at the top of this prograding unit have an age of about 2700 BC (sample 034715). The sedimentary architecture of the coastal deposits suggests that the relative sea level rose to an altitude of about +2 m (Clemmesen *et al.* 2001a).

A thick peat formed more or less continuously behind the frontal dune ridge during this progradational event, and the associated AMS dates clearly indicate the time-transgressive nature of peat formation, with ages of 4000 BC and 2700 BC (samples AAR-4604 and AAR-8673).

The initial shoreline was probably linear, trending NW–SE and flanked by a chain of relatively small coastal dunes. Coastal dune formation probably occurred concurrent with coastal progradation, and one OSL date (sample 34715) from the Grovsø II core yields an age of 4700 years (2700 BC) for these oldest dune deposits in the area. Coastal progradation was accompanied by the formation of a number of strandplain lakes. These lakes may occasionally have been flooded by sea-water, but the textural composition of their sediment fill suggests that most sand was of aeolian origin (unit A). Aeolian influx to the lake may have started soon after the shoreline was established, i.e. shortly after 4000 BC and sand influx could have continued for almost 1700 years. Two OSL dates (samples 034714 and 034713) from the Grovsø II core suggest a particularly strong influx of sand at about 3000 BC, but clearly more dates are needed to obtain a more detailed understanding of the chronology of this aeolian activity phase.

The next step in the landscape evolution involved formation of an undulatory aeolian sandplain. Aeolian accumulation on this sandplain, however, was episodic as indicated by the development of peat-rich paleosols (layers b, c, d, and e). Paleosol formation records a period of landscape stabilization. Based on the ages of these paleosols major stabilization periods are recognized at about 2300 BC samples AAR-4602 and AAR-8672), 1600–1450 BC (samples AAR-4601 and AAR-8670), 800 BC (samples AAR-4599 and AAR-8689), and 650 BC (samples AAR-4598 and AAR-8668). The latter two events, however, are difficult to separate chronologically, and other stabilization events are complex, being composed of two paleosols (e.g. paleosols c1 and c2) with different ages.

The paleosols separate sheetlike to somewhat undulatory aeolian sandplain deposits (units B, C, D, and E). These units are also recognizable on the GPR data recorded outside the present study area, suggesting that the aeolian activity events modified large parts of the Vejers dune field. The aeolian units record phases of intensified aeolian activity. At present it is not possible to say with certainty the duration of these aeolian activity events with certainty. But it is probably safe to state that aeolian sand movement was the cause of soil burial, and it is therefore inferred that the termination of soil formation is synchronous with the onset of sand movement. Knowing that once initiated, aeolian sand movement is hard to stop in high-energy wind climates, it may well be that most of the time interval between two successive periods of soil formation was characterized by intensive sand movement. This conclusion is supported by the immature nature of the paleosols.

The repeated shifts from landscape stabilization to aeolian sand movement mark important events. From our dates it is suggested that such events took place at about 2300 BC, 1450 BC, 800 BC, and 650 BC (using dates from the Grovsø II core to define these events chronologically; mid-points of calibrated age intervals rounded off to the nearest 50-year-interval). In previous studies of coastal dune dynamics on the west coast of Jutland it has been suggested that these abrupt changes in landscape evolution primarily reflected climatic change and in particular variation in summer storminess (Clemmensen *et al.* 2001a; 2001b). During late spring or early summer, or during periods when dune sand was most dry, any increase in storminess would increase the chance of large-scale dune activity and sand movement (Clemmensen & Murray 2006).

The age of some of the soil-sand transitions observed here are apparently simultaneous with regional cooling events in the North Atlantic region (Bond *et al.* 1997; Bond *et al.* 2001). Keeping in mind the dating uncertainties, the 2300 and 800 BC events recognised in this study are possibly identical to the events at 4200 and 2800 BP in the Bond scheme. During these Bond events ice-bearing water from the north of Iceland was carried far southwards in the North Atlantic; apparently related to periods of reduced sun activ-

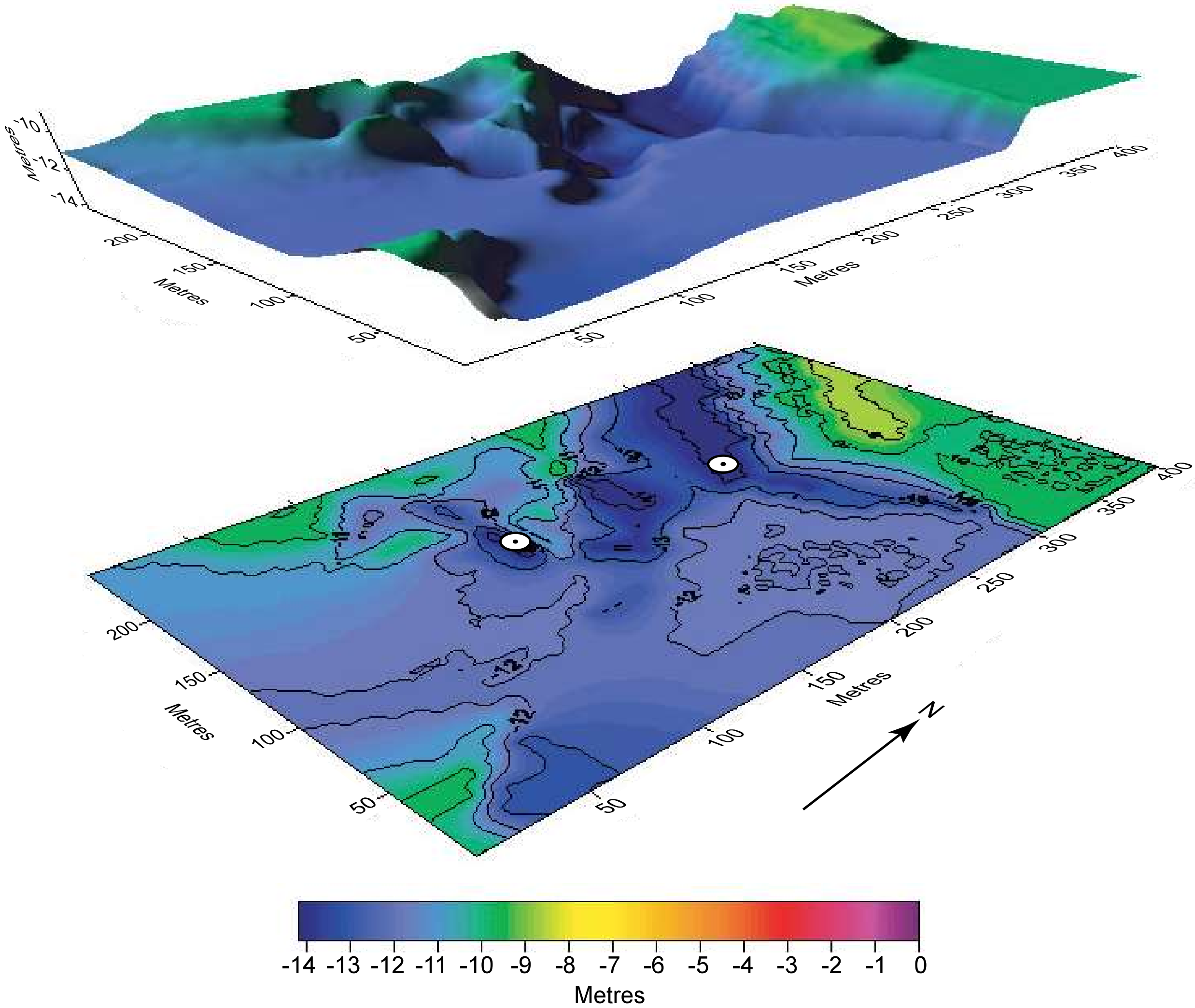


Fig. 10. 3D topography of the coastal landscape, 4000–2700 BC. The maps show from north to south: the relatively high-lying old moraine landscape (green), strandplain lakes (dark blue), and one or two discontious dune ridges (green and light blue). Positions of the two borings are marked. The vertical scale in metres refers to depth below present surface; this surface is an aeolian sand plain that lies 10-11 m a.s.l.

ity. Such marked changes in the oceanography in the North Atlantic and in particular in the distribution of sea ice would probably have influenced the position of the preferred storm tracks in the North Atlantic region and thereby also controlled (summer) storminess in Jutland. Any naturally occurring aeolian sand movement would have been intensified in periods of increased removal of the dune vegetation.

The ages of these marked changes in landscape evolution in the coastal dune field at Vejers can also be compared with the ages of the classic recurrence surfaces in Swedish peat bogs (Granlund 1932). These surfaces, dated by pollen to 2300 BC, 1200 BC, 600 BC, AD 550, and AD 1250, are thought to record marked changes from warm and dry to cold and humid conditions (Granlund 1932). Aaby (1976) studied raised bogs in Denmark and found several climatic shifts between 3500 BC and AD 1500. One of the best defined climatic shifts towards wetter conditions took place around 600 BC.

A detailed record of the climate in this period has been established in Dutch peat bogs (van Geel *et al.* 1996). The nature of these peat deposits indicates that the climate changed from warm and dry to cold and humid between 800 BC and 500 BC. The climatic change was stepwise and characterised by an initial climatic deterioration at about 800 BC followed by climatic improvement at about 600 BC, succeeded by a new climatic deterioration shortly after. This evolution in climate can probably also be recognised in the Vejers dune field. A first phase of aeolian sand movement started around 800 BC (climatic deterioration and increased storminess) and was followed by a brief period of stabilization at about 650 BC (climatic improvement and reduced storminess). Finally a second and longer lasting phase of aeolian sand movement was initiated just after 650 BC (renewed climatic cooling and increased storminess).

Landscape evolution after AD 0 is not documented by the present data, but has been described by Clemmensen *et al.* (1996), Clemmensen *et al.* (2001a) and Clemmensen & Murray (2006). There was probably a brief stabilization phase at about AD 0–300. Sand movement and dune formation dominated after around AD 300–500, but around AD 950–1100 much of the landscape was again stabilized as documented by the formation of an extensive peat-rich paleosol. Much of the uppermost sand in the system was probably remobilized during the final event of aeolian activity (AD 1100–1900). This final phase of sand movement, which was enhanced after AD 1650, apparently took place during cold and stormy intervals during the Little Ice Age and had catastrophic implications in large parts of the area. Much farmland was covered by

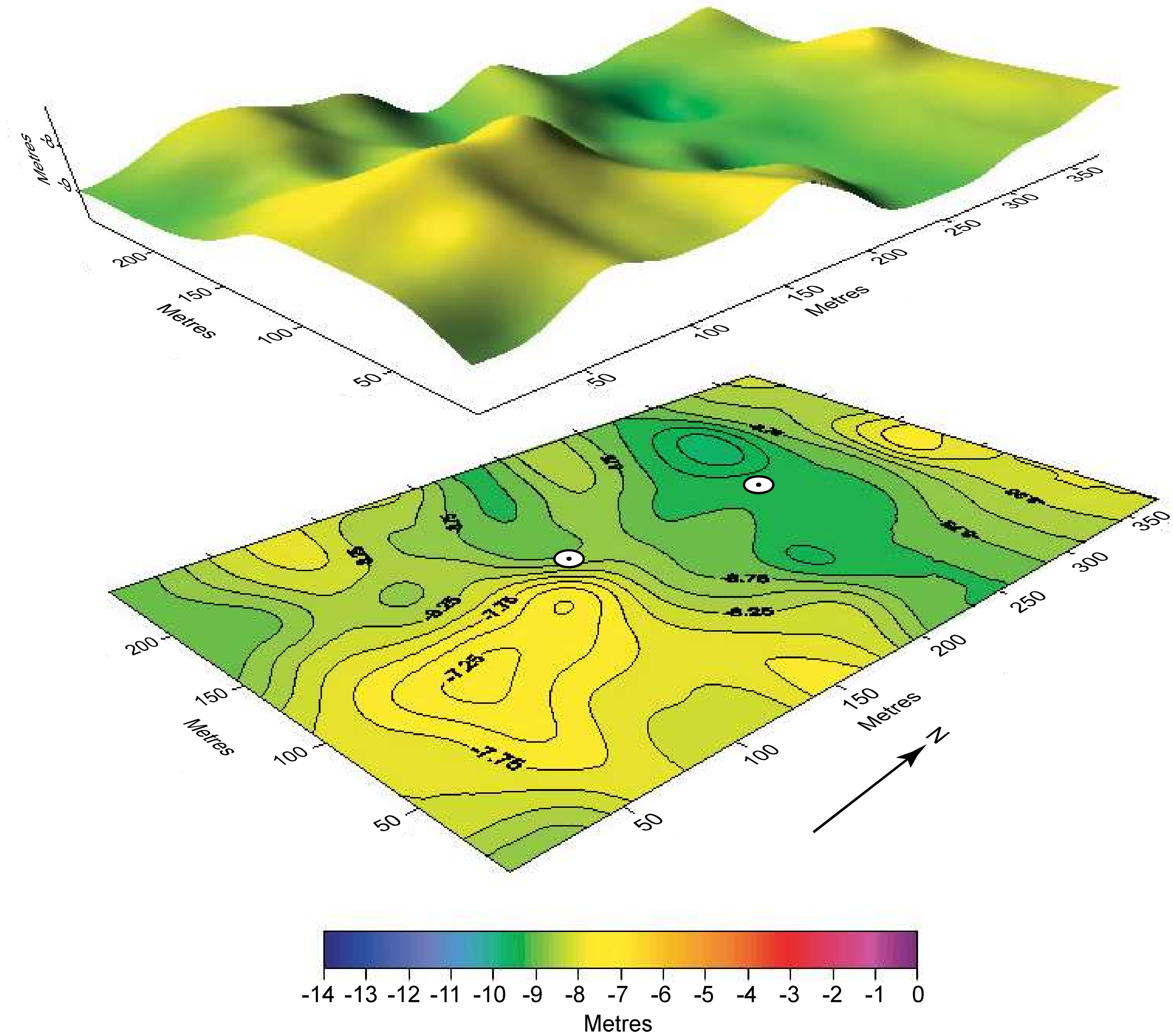


Fig. 11. 3D topography of the coastal landscape at about 2300 BC. The maps show an undulatory aeolian sand plain with low dunes; the dune forms (yellow) have no preferred orientation. There is a relatively low-lying area (green) to the north, which is a remnant of the previous strandplain lake.

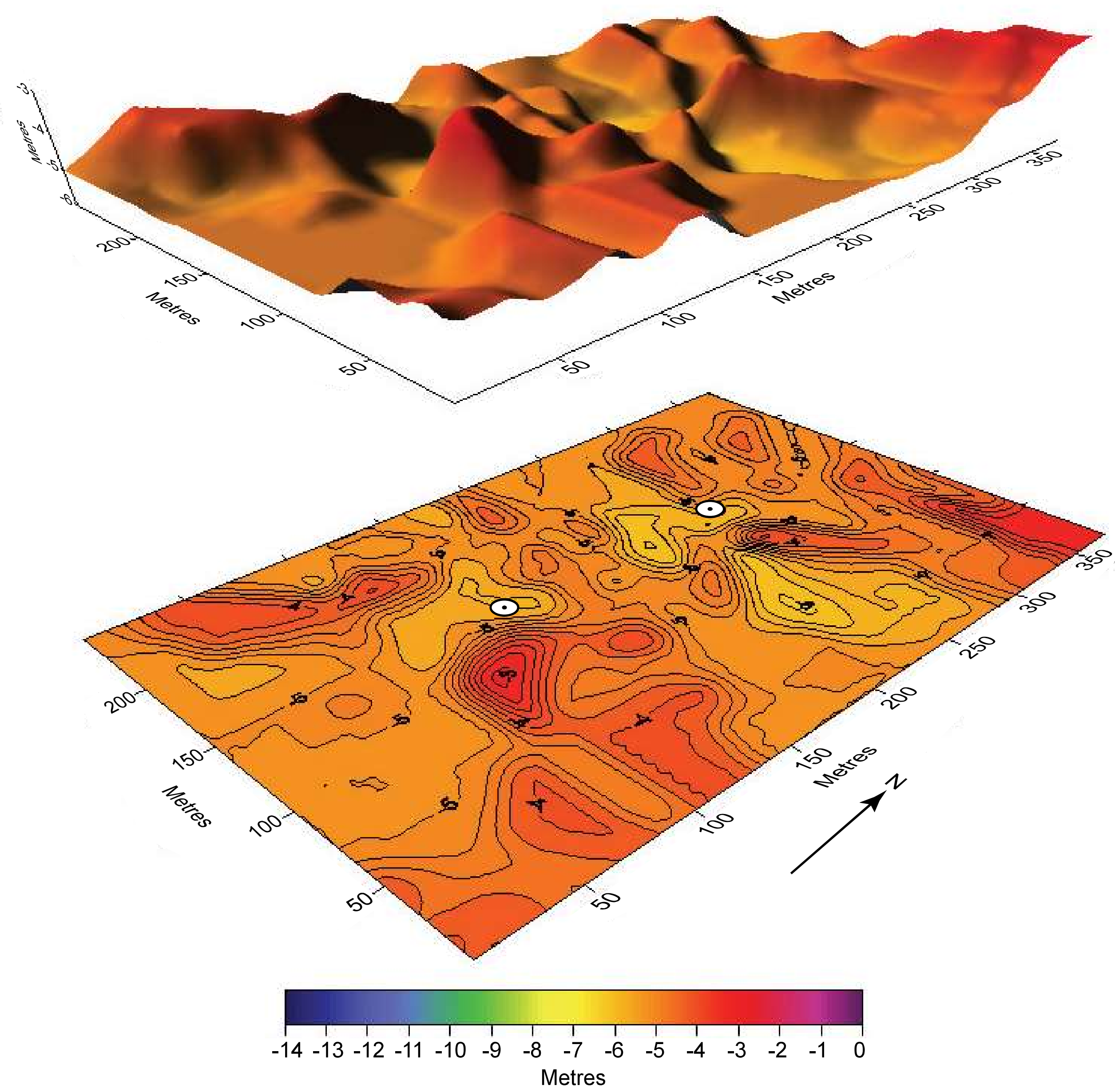


Fig. 12. 3D topography of coastal landscape at about 1450 BC. The maps show an undulatory aeolian sand plain with low dunes; the dune forms (orange and red) have no preferred orientation.

drifting sand or migrating dunes, and settlements had to be abandoned (Brüel 1918). In AD 1792 measures were taken to control the sand movement; however they were not particularly successful at first, probably because of the frequent storms during most of the 19th century (Clemmensen & Murray 2006). Many dunes in the area were still active around AD 1820, but the dunes were finally stabilized at end of the 19th century.

In conclusion, major events of aeolian sand movement was initiated at about 3000 BC, 2300 BC, 1450 BC, 800 BC, 650 BC, AD 500, AD 1100, and AD 1650. The majority of these periods of increased storminess in western Jutland apparently coincide with storm events in Halland, SW Sweden (de Jong *et al.* 2006) suggesting that dune field dynamics at Vejers record regional-scale climatic change.

3D topography of the coastal landscape

This reconstruction of 3D topography for selected intervals in the landscape evolution is based on the 100 MHz GPR data. In the GPR profiles five paleosols are identified (layers a, b, c, d, and e). The top surface of the paleosol layers form key surfaces from which it is possible to reconstruct the 3D topography of the coastal system at well-defined time intervals. The topography of the paleosols has been generated using the interpolation software Surfer 8 from Golden Software, Inc. Interpolation between data points extrapolated from the interpretation of the GPR profiles was done using the geostatistical gridding method „point kriging". The interpolation routine is set to interpolate and smooth out data points in a grid with 80 × 100 grid lines perpendicular to each other. The kriging algorithm estimates the values of the data points at the 8000 grid nodes, and thereby attempts to express the trends suggested in the data so that, for example, high points might be connected along a ridge rather than isolated by bull's-eye type contours.

Before looking at the topographical maps, however, it is necessary to make some comments on how an aeolian sediment body is created. The creation of an aeolian sedimentary record has three phases: (1) dune

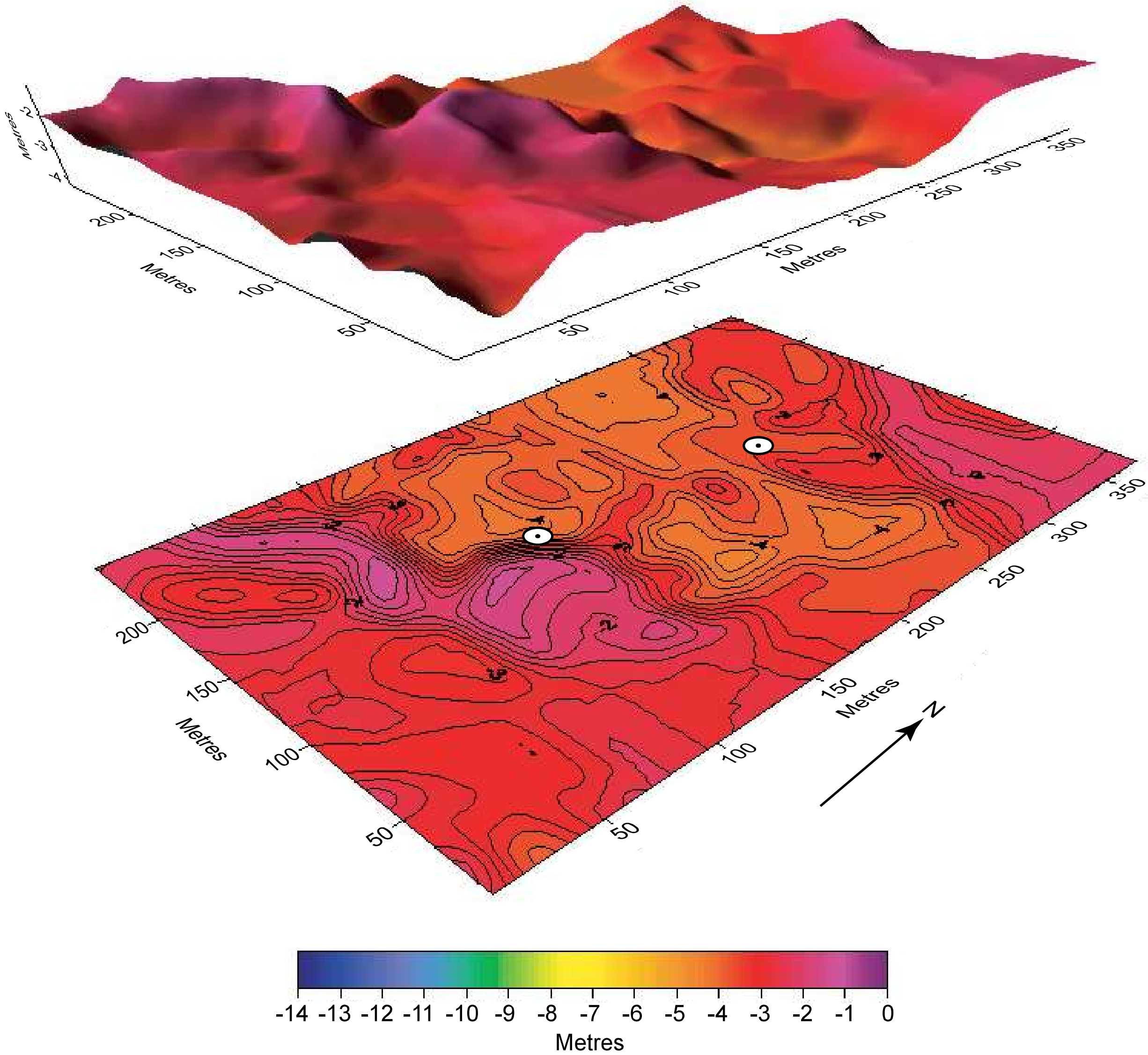


Fig. 13. 3D topography of the coastal landscape at about 800 BC. The maps show part of a low parabolic dune with arms (red and purple) enclosing a deflation plain (orange). The parabolic dune is seemingly open towards the west.

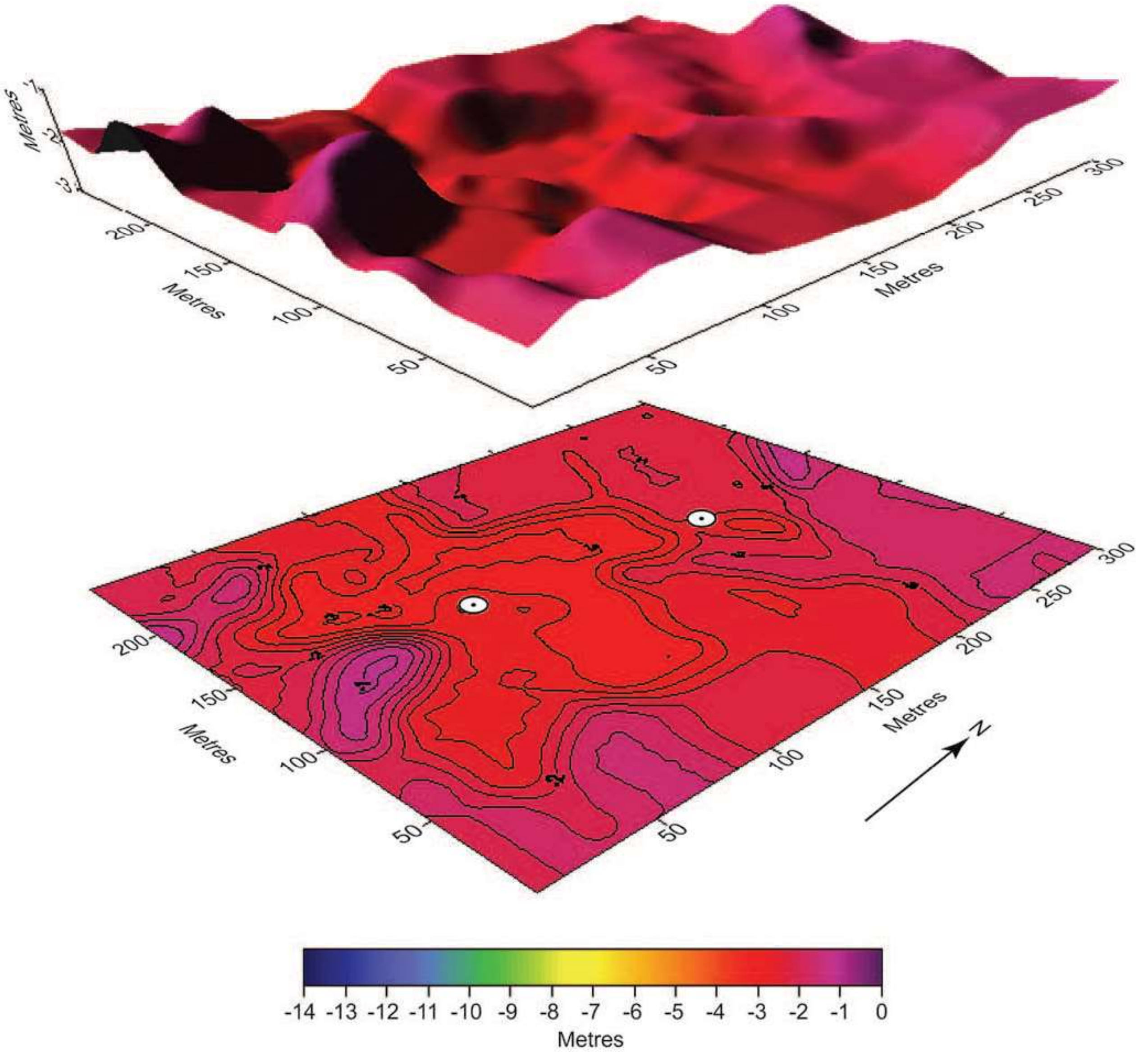


Fig. 14. 3D topography of the coastal landscape at about 650 BC. The maps show an undulatory aeolian sandplain with low dunes; the dune forms (purple) have no preferred orientation.

field/sandplain formation, (2) accumulation of aeolian sand, and (3) preservation of the sand (Kocurek 1999; Clemmensen *et al.* 2001c; Pedersen & Clemmensen 2005). To construct a dune field, accumulate the sand, and preserve it during subsequent events, requires a period of intensified sand influx, a period of rising ground-water table, and a period with a continued high ground-water table and/or the establishment of protective vegetation. Even if all these conditions are fulfilled we cannot be certain that it is the original dune field/sandplain topography we see below the paleosol. Part of the original topography may have been removed. Such erosive events e.g. resulting in blow out features, could take place during brief periods of falling ground water or during periods of intense storms. Keeping these uncertainties in mind, the evolution of the coastal system is described and discussed below in five time intervals.

4000–2700 BC

Landscape evolution in this period was characterized by coastal progradation towards the SSW. A linear coast with coastal dunes developed. A reconstruction of the topography of the coastal landscape during this time interval is given in Fig. 10, and shows three major topographic elements: A relatively high-lying moraine landscape situated towards the north is flanked by a system of disconnected strandplain lakes, which in turn are lined by one or two discontinuous dune ridges to the south. The strandplain lakes may have been flooded occasionally by sea water. The frontal dune ridges reach heights of about 4 m, which is much less than the modern frontal dunes, which have heights up to 10 m. The dune ridges and strandplain lakes trend NW-SE or parallel to the morphological trend of the landward-lying moraine landscape. This suggests that the shoreline in this period was trending NW–SE.

2300 BC

Landscape evolution between 2700 and 2300 BC was apparently characterized by continued (inland directed) aeolian sand movement. A reconstruction of the topography of the coastal landscape around 2300 BC when the aeolian topography was stabilized by vegetation is given in Fig 11, and shows a low-relief sandplain with low dune forms. The dunes have heights around 2–3 m. It is still possible to identify the former depression (strandplain lakes now almost completely filled in with aeolian sand; unit A) behind the frontal dune ridges. The dunes show no apparent orientation but the open sea must have been situated towards the southwest.

1450 BC

Landscape evolution between 2300 and 1450 BC was characterized by a second phase of aeolian activity and continued accumulation on the aeolian sandplain (unit B). A reconstruction of the topography at about 1450 BC when the aeolian topography was stabilized by vegetation is given in Fig. 12, and shows an undulatory aeolian sandplain with a number of poorly defined dunes. The dunes reach heights up to 2.5 m, but show no apparent orientation.

800 BC

Landscape evolution between 1450 and 800 BC was characterized by a third phase of aeolian activity and continued accumulation on the aeolian sandplain (unit C). A reconstruction of the aeolian topography at about 800 BC when it was stabilized by vegetation is given in Fig. 13, and shows an undulatory sandplain with what appears to be a partly preserved parabolic dune with dune arms to up to 2.5 m high. The orientation of the parabolic dune suggests winds from the west.

650 BC

Landscape evolution between 800 and 650 BC was characterized by a fourth phase of aeolian activity and continued accumulation on the aeolian sandplain (unit D). A reconstruction of the aeolian topography at about 650 BC when the landscape was stabilized by vegetation is given in Fig. 14, and shows an undulatory sandplain. Dune forms are poorly defined and have heights less than 2 m. Aeolian accumulation continued soon after 650 BC resulting in the formation of unit E. This unit is capped by the present vegetation surface in the area.

Accumulation history

The maps indicate that up to 14 m of aeolian sand has accumulated in the study area between 4000 BC and AD 0. Between 2300 BC, when aeolian sandplain formation was initiated, and AD 0, up to 9 m of aeolian sand has accumulated. This yield aeolian accumulation rates around 4 mm/yr. Comparable aeolian systems in Thy, NW Jutland, also started to accumulate sand in the Mid Holocene (Clemmensen *et al.* 2001c; Pedersen & Clemmensen 2005). Here typical aeolian accumulation rates are around 2 mm/yr.

Accumulation was controlled by a rising ground-water level. The modern ground-water level in the study area lies at about 9.5–10 m or approximately 1 m below the surface of the aeolian sand plain. This rise in ground-water level may primarily have been influenced by a long-term change towards more humid conditions, in combination with coastal progradation (Clemmensen *et al.* 2001a).

Conclusions

1. This study describes the sedimentary characteristics of a Holocene strandplain system covered by aeolian sand (Vejers, SW Jutland,) and reconstructs the sedimentary evolution of the system within a well-constrained time frame.
2. The Holocene sedimentary evolution of the system is evaluated on the basis of data from two closely situated cores supplemented by Ground-Penetrating Radar (GPR) mapping.
3. The cores consist of a lowermost unit of marine (or lagoonal) and coastal dune sediment, a middle unit of lake-aeolian sand and an uppermost unit of aeolian sandplain sediment. Peat layers and peat-rich

paleosols are common and divide the succession into genetic units.
4. The peat-rich layers are dated by AMS radiocarbon technique (15 dates), while the marine and aeolian sand deposits are dated by OSL luminescence technique (11 dates).
5. The new data indicate that the area was first transgressed around 5100 BC. During the subsequent period (5100–2700 BC) relative sea-level rose up to 5 meters, the strandplain prograded towards the SW and small coastal dunes formed. A number of strandplain lakes formed behind the frontal dunes and the lakes were filled in with sand of presumed aeolian origin. After a brief period of landscape stabilization aeolian activity was re-established at about 2300 BC and a large aeolian sandplain with low dune forms developed. Aeolian sand influx and accumulation on the sandplain was punctuated by periods of landscape stabilization and paleosol formation. Onset of aeolian activity phases are dated to 1450 BC, 800 BC, and 650 BC. Aeolian activity at the core sites ceased at about AD 0, but elsewhere in the dune system there was aeolian activity after about 500 AD and between 1100 and 1900 AD.
6. The GPR data enables a reconstruction of 3D topography of the study area (300 × 300 m) in five time intervals. The reconstructions suggest that the shoreline initially was trending NW-SE but after 2300 BC, data only reveal the topography of aeolian deposits and contain no useful evidence on shoreline orientation. The maps indicate the existence of low dune forms (typical height up to 2–3 m) on the aeolian sandplain.
7. The onset of some of these aeolian activity phases is synchronous with major cooling events in the North Atlantic region, suggesting that climate change was an important control on dune field dynamics.

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Dansk sammendrag

Det holocæne marine forland ved Vejers i det sydlige Vestjylland er opbygget af et stort oddesystem, samt mindre lokale strandslettesystemer. Oddesystemet er udbygget fra Varde Bakkeø ved Blåbjerg og strækker sig omkring 22 km sydpå til Blåvands Huk. Oddesystemet samt store dele af det nærliggende bakkeølandskab er i dag dækket af æoliske aflejringer. Langs den sydvestlige rand af bakkeølandskabet findes et mindre strandslettesystem. Den sedimentære arkitektur af dette kystsystem beskrives ud fra boredata og georadarkortlægning. Systemets holocæne udviklingshistorie placeres i en kronologisk ramme på baggrund af en række ^{14}C dateringer af tørveholdige horisonter og OSL dateringer af marine eller æoliske sandlag. De nye data viser, at den første post-glaciale transgression nåede området omkring 5100 BC. I den efterfølgende periode (5100–2700 BC) steg det relative havniveau med omkring 5 meter, strandsletten byggedes ud mod SV og der blev dannet flere generationer af små kystklitter. En relativ dyb strandsø dannedes bag kystklitterne og søen blev efterhånden udfyldt med sand med en tekstur der tyder på, at hovedparten af dette sediment var vindtransporteret. Denne første fase af æolisk aktivitet blev afbrudt omkring 2300 BC og både klitter og den sandfyldte standsø dækkedes af tørveholdige aflejringer. Sandflugten blev dog hurtigt genetableret og det vindaflejrede sand opbyggede nu en æolisk sandslette med mindre klitformer. Sandflugten fortsatte med få mindre afbrydelser indtil omkring 0 AD i det undersøgte område. I perioder med reduceret sandflugt blev hovedparten af det æolisk landskab stabiliseret af plantevækst og der udvikledes tørverige jordbundshorisonter. Skiftene fra et stabiliseret æolisk landskab til et æolisk landskab præget af sandflugt og klitdannelsen skete omkring 1450 BC, 800 BC og 650 BC. Andre steder i klitsystemet er der tegn på lignende skift i landskabsudvikling omkring 500 AD og 1100 AD. Hovedparten af disse skift i den æolisk landskabsudvikling ved Vejers skete samtidig med at klimaet i det Nordatlantiske områder blev køligere. Meget tyder således på, at klimaændinger havde afgørende indflydelse på den holocæne udvikling af det æoliske landskab ved Vejers.

References

Aaby, B. 1976: Cyclic climatic variations in climate over the past 5500 years reflected in raised bogs. Nature 263, 281–284.

Aagaard, T., Nielsen, N. & Nielsen, J. 1995: Skallingen – Origin and Evolution of a Barrier Spit. Meddelelser fra Skalling-Laboratoriet, XXXV, 1–85.

Aitken, M.J. 1985: Thermoluminescence dating. Academic Press, London.

Beets, D.J. & Spek, A.J. van der 2000: The Holocene evolution of the barrier and back-barrier basins of Belgium and the Netherlands as a function of late Weichselian morphology,

relative sea-level rise and sediment supply. Geologie en Mijnbouw/Netherlands Journal of Geosciences 79, 3–16.
Bird, E.C.F. 1990: Classification of European dune coasts. In: Bakker, Th.W., Jungerius, P.D. & Klijn, J.A. (eds): Dunes of the European Coasts. Catena (Suppl. 18), 15–24.
Bond, G., Cheseby, M. Lotti, R., Almasi, P., deMenocal, P., Priore, P., Cullen, H., Hajdas, I. & Bonani, G. 1997: A Pervasive Millennial-Scale Cycle in North Atlantic Holocene and Glacial Climates. Science 278, 1257–1266.
Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M.N., Showers, W., Hoffmann, S., Lotti-Bond, R., Hajdas, I. & Bonani, G. 2001: Persistent Solar Influence on North Atlantic Climate During the Holocene. Science 294, 2130–2136.
Brüel, J. 1918: Klitterne i Vestjylland og på Bornholm. Gyldendalske Boghandel, 133 pp.
Bøtter-Jensen, L., Bulur, E., Duller, G.A.T. & Murray, A.S. 2000: Advances in luminescence instrument systems. Radiation Measurements 32, 523-528.
Clemmensen, L.B. & Murray, A. 2006: The termination of the last major phase of aeolian sand movement, coastal dunefields, Denmark. Earth Surface Processes and Landforms 31, 795–808.
Clemmensen, L.B., Andreasen, F., Heinemeier, J. & Murray, A. 2001a: A Holocene coastal aeolian system, Vejers, Denmark: landscape evolution and sequence stratigraphy. Terra Nova 13, 129–134.
Clemmensen, L.B., Andreasen, F., Nielsen, S.T. & Sten, E. 1996: The late Holocene coastal dunefield at Vejers, Denmark: characteristics, sand budget and depositional dynamics. Geomorphology 17, 79–98.
Clemmensen, L.B., Murray, A.S., Bech, J.-H. & Clausen, A. 2001b: Large-scale aeolian sand movement on the west coast of Jutland, Denmark in late Subboreal to early Subatlantic time – a record of climate change or cultural impact? Geologiska Föreningens Förh. 123, 193–203.
Clemmensen, L.B., Pye, K., Murray, A. & Heinemeier, J. 2001c: Sedimentology, stratigraphy and landscape evolution of a Holocene coastal dune system, Lodbjerg, NW Jutland, Denmark. Sedimentology 48, 3–27.
De Jong, R., Björck, S., Björkman, L. & Clemmensen, L.B. 2006: Storminess variation during the last 6500 years as reconstructed from an ombotrophic peat bog in Halland, SW Sweden. Journal of Quaternary Science. DOI: 10.1002/jps. 1011.
Granlund, E. 1932: De svenska högmossarnas geologi. Deres bildningsbetingelser, utvecklingshistoria, och utbretning jämte sambandet mellan högmossebildning og försumpning. Sveriges Geologiska Undersökning C373.
Jessen, A. 1925: Beskrivelse af Geologisk Kort over Danmark. Kortbladet Blaavandshuk, Danm. Geol. Unders., I rk., 16, 76 pp.
Jonassen, H. 1957: Bidrag til Filsøegnens naturhistorie. Medd. Dansk Geologisk Forening 13, 192–205.
Kocurek, G. 1999: The aeolian rock record (Yes, Virginia, it exists, but it really is rather special to create one). In: Goudie, A. & Livingstone, I. (eds): Aeolian Environments, Sediments and Landforms, 239–259.
Mertz, E.L. 1924: Oversigt over de sen- og postglaciale Niveauforandringer i Danmark. D.G.U. II 41, pp.
Murray, A.S., Marten, R., Johnson, A. & Martin, P. 1987: Analysis for naturally occurring radionuclides at environmental concentrations by gamma spectrometry. Journal of Radioanalytical and Nuclear Chemistry Articles 115, 263–288.
Murray, A.S. & Wintle, A.G. 2000: Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. Radiation Measurements 32, 57–73.
Nielsen, S.T., Clemmensen, L.B. & Andreasen, F. 1996: The middle and late Holocene barrier spit system at Vejers, Denmark: structure and development. Bulletin of the Geological Society of Denmark 42, 105–119.
Olley, J.M., Murray, A.S. & Roberts, R.G. 1996: The effects of disequilibria in the uranium and thorium decay chains on burial dose rates in fluvial sediements. Quaternary Science Reviews 15, 751–760.
Pedersen, G.K. & Clemmensen, L.B. 2005: Unveiling past aeolian landscapes: A ground-penetrating radar survey of a Holocene coastal dunefield system, Thy, Denmark. Sedimentary Geology 177, 57–86.
Prescott, J.R. & Hutton, J.T. 1994: Cosmic ray contributions to dose rates for luminescence and ESR dating: large depths and long-term time variations. Radiation Measurements 23, 497–500.
Saye, S.E. & Pye, K. 2006: Variations in chemical composition and particle size of dune sediments along the west coast of Jutland, Denmark. Sedimentary Geology 183, 217–242.
Stuiver, et al. 1998: INTCAL98 Radiocarbon Age Calibration, 24000–0 cal. BP. Radiocarbon 40, 1041–1083.
van Geel, B., Buurman, J. & Waterbolk, H.T. 1996: Archaeological and palaeoecological indications of an abrupt climate change in the Netherlands, and evidence for climatic teleconnections around 2650 BP. Journal of Quaternary Science 11, 451–460.