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Radiocarbon-dated sediment sequences from the Belgian coastal plain: testing the hypothesis of fluctuating or smooth late-Holocene relative sea-level rise

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Abstract: The late-Holocene deposits of the coastal lowlands bordering the southern North Sea were formed by a renewed expansion of the tidal environment replacing freshwater peats. The sea-level history during the last 2500 years inferred from the post-peat deposits, containing little to no datable organic material, is characterized by two conflicting schools of thought. The first one supports a smooth sea-level curve. The second approach underpins fluctuating sea-level changes. The 'fluctuating' approach also uses facies variations recorded in the post-peat deposits to provide a chronostratigraphical subdivision linked to transgressions and regressions. This paper examines the controlling factors responsible for the facies variations in the post-peat deposits in order to test the approach of the fluctuating sea level. The results of sedimentological investigation coupled with radiocarbon dates of intertidal shells from four shallow outcrops located in nearby sand-filled late-Holocene tidal channels in the Belgian coastal plain are reported. The integration with results from previous work has allowed the reconstruction of the mechanisms and processes of coastal evolution during the late Holocene. The changes in the coastal landscape reflected in the facies variations are caused by multiple factors. The major one is the dynamic nature of the tidal channels responding to changes in accommodation space and sediment budget, and finally, to storm incidence and human impact. The processes of the tidal channel networks, ie, their initiation and evolution, are similar, but the changes happened at different times in different places. The facies variations do not involve rises and falls of sea level, and therefore a fluctuating sea-level rise can not be considered as realistic.

Key words: Tidal back-barrier deposits, facies changes, sedimentary processes, peat and sediment compaction, relative sea-level rise, Northwest European coastal lowlands, Belgian coastal plain, late Holocene.

Introduction

Determining the rate and direction of sea-level change is important to those interested in reconstructing climate change, determining driving mechanisms of coastal change, and to provide a context for the interpretation of coastal archaeology and human behaviour. Two models exist for the interpretation of relative sea-level changes in the southern North Sea during the late Holocene (taken here as approximately the last 2500 years). The first, or 'smoothly rising' model, is based on the careful collection and interpretation of sea-level index points from The Netherlands (van de Plassche, 1982; Kiden, 1995) and Belgium (Denys and Baeteman, 1995). These curves do not show fluctuations in the late Holocene; on the contrary, they suggest a smooth and gradual sea-level rise with an average of 0.7 to 1 mm/yr. The 'smoothly rising' model is important for those interested in the interpretation of coastal evolution since it implies that variations in the late-Holocene facies architecture recorded in coastal deposits throughout the region are not the result of fluctuations in sea level.

In contrast, an alternative school of thought advocates a 'fluctuating' model for sea-level change. Derived from the interpretation of coastal deposits from the same region of the southern North Sea, proponents of this approach use a formal stratigraphical subdivision as a basis for chronostratigraphic correlation and interpretation (eg, van Staalduinen *et al.*, 1979). A characteristic of this approach is the interpretation of facies changes in terms of sea-level fluctuations and associated retrogradational/progradational shifting of the shoreline. Importantly, several studies use these events as the basis for a chronostratigraphical subdivision. For a synthesis of the critical comments the reader is referred to Streif and Zimmermann (1973) for northern Germany, Wheeler and

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Waller (1995) for England, Ebbing *et al.* (2003) for The Netherlands, and Baeteman (1983, 1999, 2005b) for Belgium.

Under the 'fluctuating' model, the traditional stratigraphy for Holocene coastal deposits in western Europe is the subdivision into 'Calais' and 'Dunkerque' transgressions with, for the late Holocene, seven Dunkerque transgressions (D0, DIa, DIb, DII, DIIIa, DIIIb, DIV), each of them separated by a regression. For recent examples, the reader is referred to Behre (2003, 2007), Bungenstock *et al.* (2004), Mauz and Bungenstock (2007) for northern Germany, and Meurisse *et al.* (2005) and Frouin *et al.* (2007) for France. These studies define each transgression/regression phase almost entirely on the basis of their age, with little or no regard to facies variations and the processes responsible for the changes. Behre (2003, 2007) even claims that it is the only stratigraphical system that can be used for supraregional contexts because it is based on dated transgressive and regressive trends, and it has the only terminology that is applicable throughout the southern North Sea. According to this author, a regression corresponds to a sea-level fall occurring synchronously in the entire southern North Sea area and, on this basis, he has reconstructed a strongly fluctuating sea-level curve for the last 3000 years using data collected from north Germany. The author also neglects to explain the cause of such sea-level changes. The rates of absolute sea-level rise and fall, however, are unrealistic. According to the sea-level curve, the rate for the sea-level rise for the DIb, DII and DIIIB transgressions are 84, 50 and 40 cm/100 years, respectively. It is also notable that other authors dismiss the possibility of such short-lived, high-amplitude oscillations in sea-level. For example, Vos and Kiden (2005) argue that sea-level fluctuations of more that 50 cm in a timespan of hundreds of years or shorter are not possible.

Bungenstock *et al.* (2004) and Mauz and Bungenstock (2007) criticize the 'smoothly rising' sea-level curve of the North Sea for the late Holocene, arguing that the apparent continuity of the rise is generated simply by extrapolation from the youngest sea-level index point to the present, a period of time that usually is some 3000 years and represented by the last peat of stratigraphic significance (see below). Their study, based on OSL dating of mud-flat deposits of the East Frisian coast (N Germany), assumes transgressive and regressive phases and associated short-lived (200 to 400 years) sea-level fluctuations with a small absolute rise and fall of less than 1 m of amplitude. Their investigation maintains that the vertical range over which the late-Holocene relative sea level fluctuated was probably not more than *c.* 30 cm. They dismiss the effect of (auto)compaction as relatively small because there is almost no variation in sediment composition. However, they neglect to consider that the mud not only compacts as soon as a fresh sediment accumulates on top, but compaction by simple de-watering continues, although at an exponentially decreasing rate, for thousands of years. Also the deeper Holocene deposits continue to compact. These processes mean that the dated sediments are no longer in their original position, which moreover is impossible to detect because the initial moisture content, which can vary, is not known.

The statement by Bungenstock *et al.* (2004) and Mauz and Bungenstock (2007) that the smoothly rising sea-level curve for the late Holocene is an extrapolation from the youngest sea-level index point to present, must be reconsidered. It is true that the post-peat deposits contain little or no organic datable material. Although a very small number, the few reliable sea-level index points for the period between 4500 and 1500 cal. BP have been published in Baeteman (2001, 2004a), and are here presented in a time/depth diagram (Figure 1). The index points were collected in an area where the subsoil consists entirely of sand. Therefore compaction and subsidence due to de-watering caused by the tidal channel incision (see below) or later human activity, can be minimized. With the relatively small number of points, the graph does not show sea-level fluctuations.

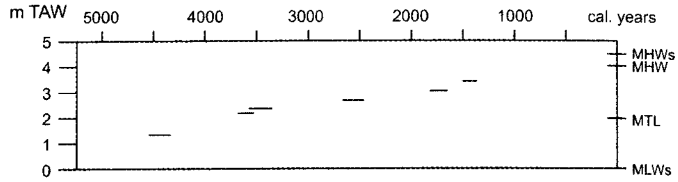

Figure 1 Time/depth diagram of the sea-level index points for the mid and late Holocene from the Belgian coastal plain. The calibrated date ranges are given at the two sigma age range. MHWS, mean high water spring; MHW, mean high water; MTL, mean tidal level; MLWS, mean low water spring; TAW, the Belgian ordinance datum referring to mean low water spring which is about 2 m below mean sea level

The 'fluctuating model' and the (chrono)stratigraphic subdivision have been criticized by many authors (eg, Streif and Zimmermann, 1973; Barckhausen *et al.*, 1977; Baeteman, 1981, 1983, 1991; Van Loon, 1981; Berendsen, 1984; Westerhoff *et al.*, 1987; Denys, 1993, 1999; Wheeler and Waller, 1995; Vos and van Heeringen, 1997; Ervynck *et al.*, 1999; Weerts *et al.*, 2005), whilst a critical review of the sea-level index points used by Behre (2003, 2007) shows that many are not reliable because of uncertain age and height relationships to former sea level (Baeteman, 2007a; Vink *et al.*, 2007).

This short introduction shows two conflicting schools of thought, each developed from the same coastal environments of the southern North Sea Basin, but each interpreting the evidence in very different ways.

This paper will describe in detail the sedimentary characteristics of the late-Holocene deposits in shallow outcrops that were temporarily accessible in the western part of the Belgian coastal plain (Figure 2). The sedimentary investigation, together with radiocarbon dates, is supplemented by data from the large borehole data base obtained during geological mapping, which together enables an integrated reconstruction of the late-Holocene coastal changes in the study area and a rigorous test of the 'fluctuating' sea-level model. The Belgian coastal plain is a stable area with respect to tectonic and isostatic movement (Vink *et al.*, 2007) so that evidence for relative sea-level changes is either a function of sea-surface change or changes in sediment and coastal processes or both. It is argued that the facies data provide strong evidence for a smooth rise in sea level, with changes in coastal evolution controlled by sedimentary processes and variations in sediment supply and accommodation space, and human impact rather than discrete oscillations in sea level.

Study area and previous work

The Belgian coastal plain is an embanked coastal lowland situated at or below Spring Tide level and protected from flooding by dunes and by artificial coastal structures. The coast is tide-dominated with a Spring tidal amplitude of almost 5 m. It is characterized by a straight and closed coastline. The coastal plain is drained by a small river, the IJzer, along which the plain extends further south in the western part.

The late-Holocene deposits of the coastal plain were formed by a renewed expansion of the tidal environment via tidal channels (Baeteman, 2005a). Brackish clastic sediments replaced a peat accumulation. The beginning and end of the peat accumulation are diachronous. It developed at about 6300–5500 cal. BP in the landward part of the plain, and at *c.* 4700 cal. BP in the more seaward areas (Baeteman, 1991, 1999). The peat accumulated almost

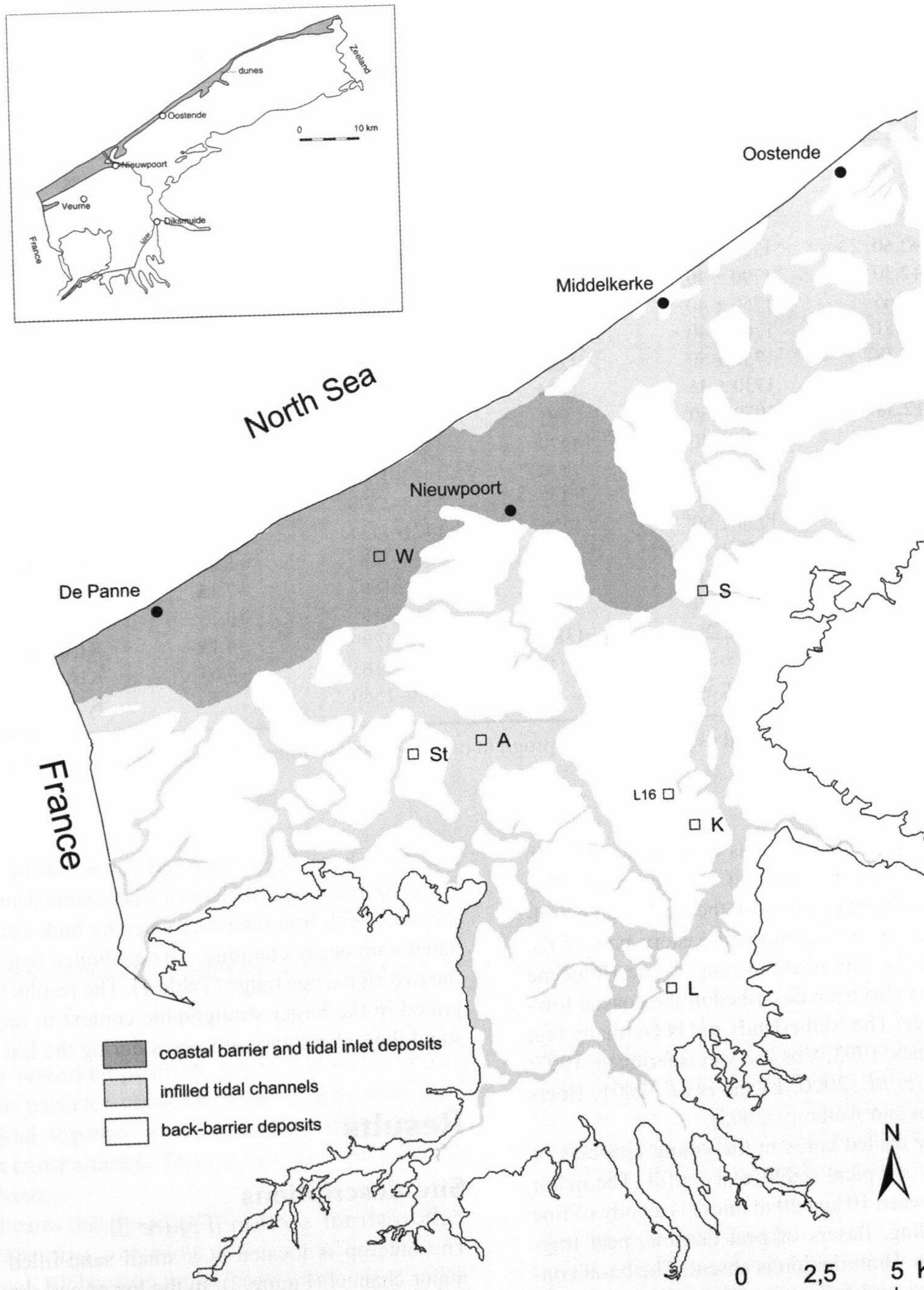

Figure 2 Map of the western part of the Belgian coastal plain showing the location of the late-Holocene sand-filled channels with indication of the outcrops and cores discussed in the text. S, Schorestraat; W, Wulpen; St, Steenkerke; A, Avekapelle

without interuption for a period of 2–3 ka years while the coast was prograding. Precise dates for the end of the peat accumulation are not known because the determination of accurate dates for the end of the peat formation is extremely difficult. A thorough investigation by Waller *et al.* (2006) documents that some erosion and reworking is always likely during the initial stages of inundation. Moreover, the surface of the peat might have been affected by a limited degree of intrusion, oxidation or surface reworking. In the study area only few locations show a gradual transition with the overlying sediments giving reliable dates. The youngest age of all for the top of the peat where no erosion or oxidation is assumed, is 1686–1390 cal. BP (1525 cal. BP, Baeteman *et al.*, 2002), taken from the very landward part of the plain. The other dates from the surface of the peat range from 3370 to 1525 cal. BP, with a cluster between 2250 and 2000 cal. BP. The full data set of the radiocarbon dates of the top of the peat are published in Baeteman *et al.* (2002) and in Baeteman (2005a).

The renewed expansion of the tidal environment marked a profound change in the coastal landscape. Relative sea level at that time was about 2 m lower than at present (Baeteman, 2007b). The expansion was characterized by the formation of tidal channels cutting through the mid- and early-Holocene deposits, and sometimes several metres into the Pleistocene deposits. This happened most often with erosion of the peat. According to the available dates, it is assumed that the re-entrance happened between 2400 and 2000 cal. BP (Baeteman *et al.*, 1999). The radiocarbon dates of peat boulders in the channel fills indicate that they originate from the upper peat bed, even those in the lags at the base of the channels (Baeteman, 2005a). Owing to a lack of age control on the tidal channel incision, a more precise date can not be given. The location of the late-Holocene tidal channels, hereafter called the young channels, was mapped in detail during the systematic geological mapping of the plain with over 50 deep mechanically drilled cores in the sand-filled channels (Figure 2; Baeteman,

Table 1 Radiocarbon dates from the investigated sites

Site	Sample altitude (m TAW)	Age (yr BP)	Calibrated ages (yr BP)	2σ range	δ^{13} C‰	Laboratory number	Dated material
Schorestraat 4	+1.90	1755 ± 25	1289	1388–1214	−2.88	UtC 4355	*Scrobicularia*
Schorestraat 3	+2.00	1730 ± 50	1278	1402–1158	−1.35	UtC 4153	*Cerastoderma*
Schorestraat 2	+2.60	2065 ± 35	1618	1759–1506	−0.88	UtC 4353	*Hydrobia*
Schorestraat 1	+2.60	1730 ± 25	1275	1355–1179	−2.64	UtC 4351	*Scrobicularia*
Wulpen A	+2.30	1790 ± 40	1312	1454–1229	−0.90	UtC 4678	*Cerastoderma*
Wulpen B	+2.65	1750 ± 40	1285	1395–1178	−1.20	UtC 4679	*Cerastoderma*
Wulpen C	+2.51	1705 ± 40	1260	1341–1143	−2.40	UtC 4674	*Scrobicularia*
Wulpen D2	+2.38	1750 ± 50	1285	1415–1163	−1.00	UtC 4681	*Cerastoderma*
Wulpen E	+2.58	1730 ± 35	1274	1367–1168	−1.20	UtC 4682	*Cerastoderma*
Wulpen D1	+2.38	1070 ± 30	981	1007–928		UtC 4745	reed
Voetbalveld 1	+3.14	1740 ± 40	1279	1386–1171		Irpa 1205	*Scrobicularia*
Voetbalveld 2	+2.90	1810 ± 40	1329	1474–1252		Irpa 1206	*Cerastoderma*
Avekapelle	+1.85	1840 ± 50	1372	1513–1273	−4.32	UtC 9429	*Scrobicularia*
Avekapelle	+1.80	1730 ± 60	1278	1421–1131	−5.15	UtC 9430	*Cerastoderma*
Steenkerke	+2.50	1330 ± 50	895	988–728	−5.03	UtC 9428	*Cerastoderma*
Steenkerke	+2.00	2400 ± 50	2021	2188–1867	−7.15	UtC 9417	*Scrobicularia*
Steenkerke	+2.60	995 ± 30	560	644–515	+0.73	Irpa 1279	*Scrobicularia*
Lamp.A42 (L)	+1.80	1850 ± 25	1384	1492–1299	−4.13	KIA 12256	*Scrobicularia*
Kaaskerke (K)	+0.53	1700 ± 25	1260	1329–1165	−4.63	KIA 12248	*Scrobicularia*
L 16	+1.56	1760 ± 30		1740–1560		KIA 32623	peat

Calibration of radiocarbon dates was completed using the CALIB program of Stuiver and Reimer (1993) and Oxcal3.10 and INCAL104 FOR L16.

2004b, 2005a). The renewed expansion was also associated with shoreface erosion and a landward shift of the coastline. This resulted in the complete disappearance of the early- and mid-Holocene barrier sequence in the central and eastern part of the Belgian coast (cf. Figure 2). The phenomenon of late-Holocene tidal channel incisions has also been described in the coastal lowlands of England, N France, The Netherlands and N Germany (eg, Streif, 1972; Long and Innes, 1993; Vos and van Heeringen, 1997; Brew *et al.*, 2000; Long *et al.*, 2000; Evans *et al.*, 2001; Beets *et al.*, 2003; Mrani Alaoui and Anthony, 2005).

The deep mechanically drilled cores in the young channels of varying dimensions show a typical sedimentary infill. The major part of the sediments (between 10 and 20 m thick) is a body of fine sand showing cross-bedding, flasers of peat detritus, peat fragments and reworked shells. Bioturbation is absent. The basal contacts are sharp and erosive, and frequently marked by a shell lag or by peat fragments. The upper part, about 2 m thick and called the final fill, consists of interlaminated sand and mud deposited in a low-energy environment. The sand–mud alternations likely resulted from recurring tidal currents. Shells in living position and bioturbated intervals are frequent. Therefore, it is suggested that the deposition of the final fill occurred in a sub- or intertidal flat setting. Radiocarbon dates of shells in living position indicate that the final fill was formed between *c.* 1400 and 1200 cal. BP (Baeteman *et al.*, 2002; Baeteman, 2005a).

In the major part of the plain where the upper peat bed has not been entirely eroded, not all post-peat deposits show the same facies transition. In the most distal reaches of the channels, little or no facies differentiation is discernible vertically in the 1 to 2 m thick clay. Surfaces indicative of breaks in the sedimentation are lacking. This at a first glance implies that the clay was deposited by one single process and at the onset of the tidal inundation. In the proximity of the sand-filled channels, however, facies changes in the post-peat deposits are frequent.

Temporary shallow outcrops in areas near the young sand-filled channels offered the opportunity of investigating the facies variations in the post-peat deposits and of documenting the factors controlling them. Only articulated intertidal shells have been dated and the dates have been corrected for a marine reservoir effect of 400±40. Individual shells as a whole were dated with AMS except the dates with Irpa reference whereby bulk samples of shells were dated with decay counting. The calibrated date ranges are given at the two sigma age range (Table 1). The results have then been integrated in the larger stratigraphic context in order to better understand the sedimentary processes during the last 2000 years.

Results

Site descriptions

Schorestraat section (Figure 3)

The outcrop is located at a small sand-filled channel joining a major channel (Figure 2). In the lower part the outcrop shows the final fill with the typical mud and sand laminations with bioturbations. It is truncated by a thin sand bed in which articulated *Scrobicularia plana* are present, however in a flat position. They have obviously been reworked, but without significant transportation. It is assumed that the shells originate from the underlying deposits where their burrows are still present. This also indicates that there was no deep vertical erosion. The shells have been dated at 1388–1214 cal. BP (1290 cal. BP). The thin sand layer is covered with a 10 cm thick sandy clay with at the top numerous *Cerastoderma edule* dated at 1402–1158 cal. BP (1275 cal. BP). The identical age of the two shell horizons allow us to assume that both originate from the same final fill deposit. The sandy clay is overlain by clay from an intertidal flat that, in view of the oxidized root penetrations, has been colonized by plants implying a transition to a supratidal flat. The latter became again in an intertidal position with deposition of a *c.* 30 cm thick clay layer. Another thin sandy clay bed with an erosional lower boundary covers the clay. Numerous *Hydrobia*, articulated *Scrobicularia* and *Cerastoderma* are concentrated in a chaotic manner. Their age is identical to the shells in the lower sandy clay bed, except for the *Hydrobia*. Dating *Hydrobia* is always problematic because this tidal-flat gastropod is easily reworked and transported and can float in the water for a long time. The shells in the upper horizon

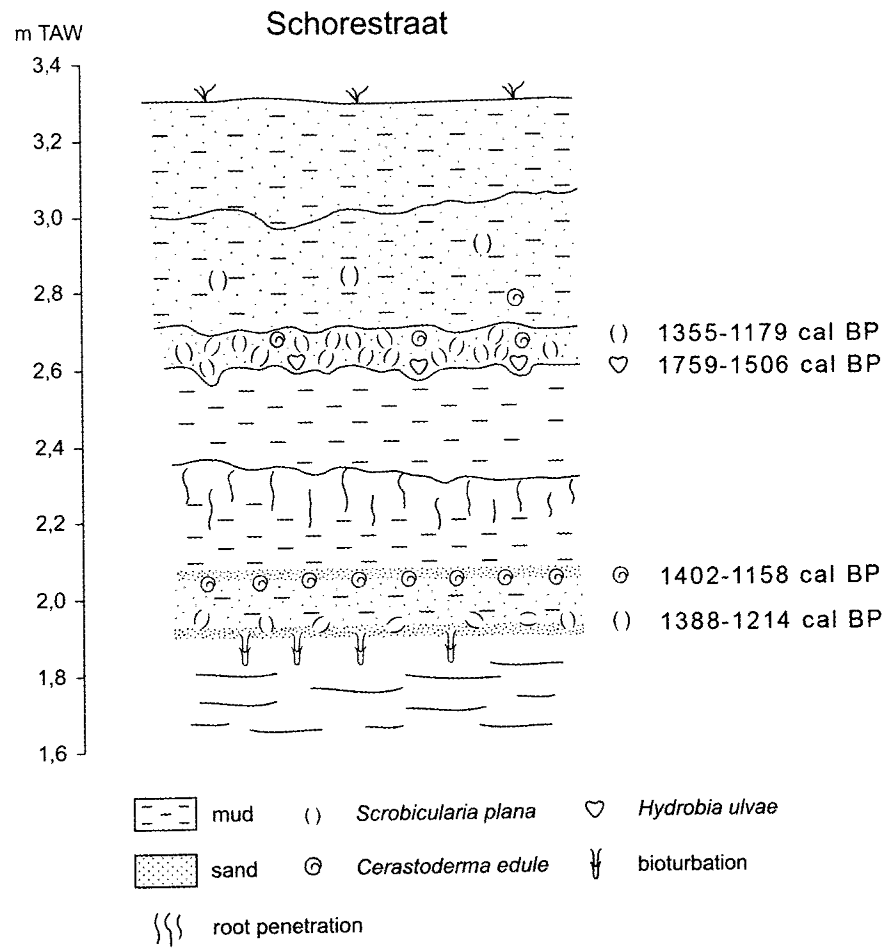

Figure 3 Stratigraphic section of the Schorestraat site. See Figure 2 for location (S)

are obviously in a reworked position, but no selective sorting or removal of fine particles occurred. This suggests that they have been reworked together with their host sediments, but without significant transportation. This unusual lag is probably the result of a storm event.

This section indicates that the final fill of the channel has been reworked twice. The first reworking most probably happened shortly after 1275 cal. BP because the shells originate from the top of the underlying sediments where the burrows are still present, which excludes significant vertical erosion. There is no evidence available for the time of the second reworking. The development of a supratidal surface implying prolonged subaerial exposure prior to the resumption of deposition indicates that it happened much later.

Wulpen section (Figure 4)

The Wulpen section is located at a major sand-filled channel within the reach of the tidal inlet (Figure 2). Therefore, the upper part of the channel fill does not show the typical interlaminated sand and mud because of higher energy conditions prevailing in the inlet. Two outcrops at a distance of about 500 m apart, Wulpen and Voetbalveld, show a quite different stratigraphy.

In Wulpen, the channel sand is overlain by soft mud with irregular sand laminae, numerous *Cerastoderma*, peat fragments and reed fragments. The latter have been transported because they are in a flat position. The *Cerastoderma* have been dated at 1454–1229 and 1415–1163 cal. BP (1312 and 1285 cal. BP). A reed fragment gave an age of 1007–928 cal. BP (981 cal. BP) and the *Scrobicularia* at the top 1341–1143 cal. BP (1260 cal. BP). The soft mud is erosively overlain with a 10 cm thick clayey sand, in turn covered with a 15 cm thick fine sand without any lamination and containing reed fragments and articulated *Cerastoderma*. The shells have been dated at 1367–1168 cal. BP (1274 cal. BP) and 1395–1178 cal. BP (1285 cal. BP). It is obvious that the 50 cm thick deposit overlying the channel sand results from erosion and reworking. It is covered with laminated fine sand and mud.

Voetbalveld (Figure 4)

At this site, the sandy final fill is erosively overlain by a soft mud but at a slightly higher position than in Wulpen. However, the numerous burrows in the top of the sand exclude deep vertical erosion. The soft mud contains irregular sand laminae, rounded peat fragments and numerous articulated *Cerastoderma* in a chaotic position at the top. The shells have been dated at 1474–1252 cal. BP (1329 cal. BP). The mud, which most probably represents a lag deposit, is erosively overlain with silty sand that is strongly burrowed, indicative of sedimentation at low-energy conditions. A concentration of *Scrobicularia* in their living position was found at the top of the sand and dated at 1386–1171 cal. BP (1279 cal. BP). The burrowed silty sand is overlain by laminated fine sand and mud. The age of the *Scrobicularia* indicates that here the soft mud was deposited before *c.* 1300 cal. BP as opposed to the Wulpen site where the mud was deposited some 300 years later. These two sites also document that the final fill has been reworked twice. The first reworking happened around *c.* 1300 cal. BP; the second one later than 1000 cal. BP as documented by the age of the reed.

A comparison between both sections indicates that the sequence with the burrowed silty sand observed in the Voetbalveld section is missing in the Wulpen section where it has probably been eroded. Therefore, it is assumed that the deposition of the overlying laminated sand and mud in Voetbalveld coincides with the erosion

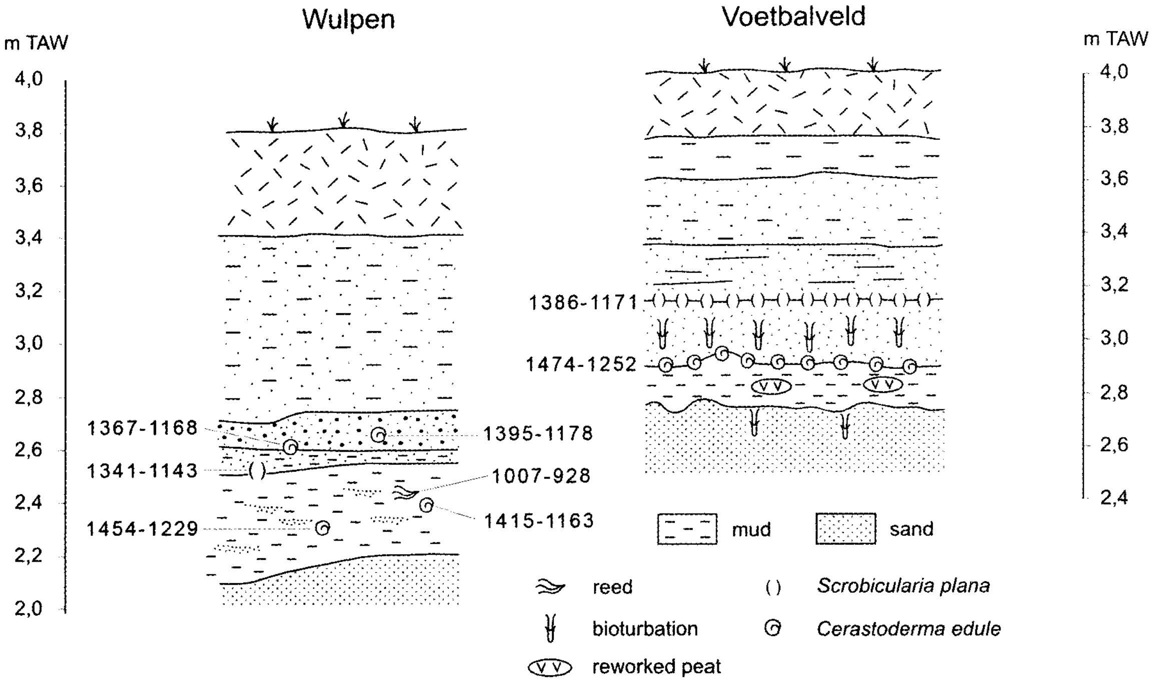

Figure 4 Stratigraphic section of the Wulpen and Voetbalveld sites. See Figure 2 for location (W)

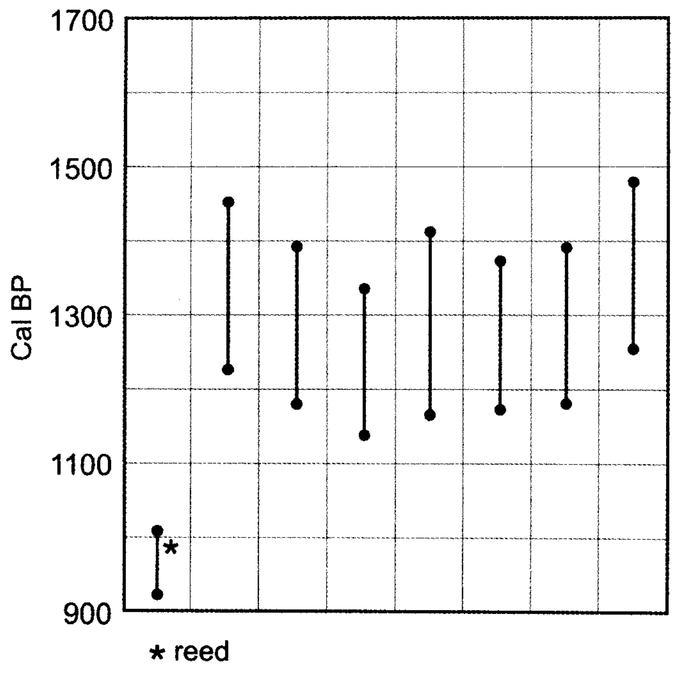

Figure 5 Diagram with the ages of the shells and reed from the Wulpen and Voetbalveld sites. The calibrated date ranges are given at the two sigma age range (from Van Strydonck and Baeteman, 1998)

in the Wulpen site. This example shows that the same sedimentary process does not necessary result in a similar sediment succession, even at short distances in the same sedimentary environment. The diagram (Figure 5) with the ages of the shells from both sites, shows that statistically the shells can have the same age. The identical ages of the reworked and the *in situ* shells also indicate that the changes happened rapidly. Therefore, the different stratigraphies in both sites can not be attributed to a sea-level fluctuation. This example also documents that radiocarbon dates of shells (or microfossil analyses) must be interpreted with regard to their sediment context, and that their age may not necessarily date their host sediment.

Avekapelle (Figure 2)

This site shows the final fill of the same major channel as the one at the Wulpen sites. Here the typical mud and sand laminations with numerous burrows representing the final fill, are present. The reworked mud deposit is not present. *Scrobicularia* and *Cerastoderma* in their living positions have been dated at 1513–1273 and 1421–1131 cal. BP (1372 and 1278 cal. BP) which is an identical age to the reworked shells of the Wulpen sites. Therefore it is assumed that the shells and mud in the Wulpen sites originate from the reworking of the final fill of this major channel. At the Avekapelle location, however, the final fill contains small and rounded peat fragments which are not typical for the final fill. This indicates that during the time of the formation of the final fill previously eroded peat fragments in the channel sand were again reworked. This can be explained by the process of lateral migration of the channel with shallow vertical erosion once the channel was almost filled. At this location the sand-filled channel is unusually wide (cf. Figure 2) which suggests lateral erosion in the period following the initial incision and infill.

The Steenkerke site (Figure 6)

The Steenkerke site is located in a small channel which joins the major channel of the Wulpen and Avekapelle sites (Figure 2). Here the upper peat bed has not been entirely eroded and different phases of sedimentation could be observed. The peat is partly eroded by a shallow channel filled with mud and sand containing a lot of peat detritus. The channel fill laterally changes into a thin bed of mud. *Scrobicularia* in their living position at the top indicate an intertidal position. The shells have been dated at 2188–1867 cal. BP (2021 cal. BP). This deposit probably represents the initial tidal inundation of the peat. The shallow channel was then incised by a larger channel that became sand filled. The whole is erosively overlain with a layer of fine sand which is not very thick, but shows a wide lateral extension. The sand also covers the peat, but does not show deep erosion. Locally, the longitudinal bedding with mud drapes documents a point-bar deposit produced by the lateral migration of a shallow channel through meandering. Articulated *Cerastoderma* at the base of the sand have been dated at 988–728 cal. BP (895 cal. BP); this is more

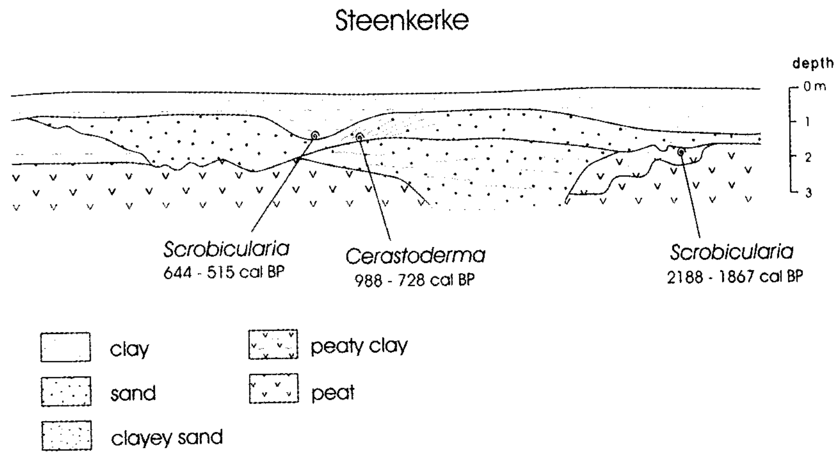

Figure 6 Stratigraphic section of the Steenkerke site. See Figure 2 for location (St)

than 1000 years after the initial inundation of the peat. It is obvious that the concentration of the *Cerastoderma* at the base of the channel fill are in a reworked position, but since there is no deep erosion, it is assumed that the shells originate from the intertidal flat adjacent to the migrating channel and that the shells have had a limited history of reworking and are only slightly older than the host deposit. Therefore the deposition of the sand most probably happened at about 900 cal. BP. A mud-filled creek represents the ultimate silting up of the channel. *Scrobicularia* in living position at the base of it have been dated at 644–515 cal. BP (560 cal. BP). This indicates that the area evolved into a supratidal flat in late Mediaeval times.

The Steenkerke outcrop provides the evidence that the incision of the tidal channels did not happen simultaneously with the initial inundation of the peat. This location changed to an intertidal position about 1000 years after the initial inundation.

The stratigraphy at more distal reaches of the young channels

The post-peat deposits in areas at more distal reaches of the young channels have been investigated in boreholes. In order to understand the chronological relation between the infill of the young channels and the deposition of the clay, shells have been sampled near the top of the peat in two cores (Figure 7). The cores K and L are located in a more landward position of the plain at a distance of 1.5 km and 750 m from a young channel, respectively (Figure 2). Although the shells were not in living position, it was expected that their age would give a limiting date for the onset of the post-peat sedimentation and for the end of the peat growth. The *Scrobicularia* in core K and the *Cerastoderma* in L gave an age of 1329–1165 cal. BP (1260 cal. BP) and 1492–1299 cal. BP (1384 cal. BP), respectively. These young ages are similar to those from the Wulpen and Avekapelle sites. In borehole L16 located nearby core K and at a distance of 400 m from a small tributary channel, the top of the peat showing a gradual transition was dated at 1740–1560 cal. BP (Figures 2 and 7). At about 50 cm above the peat a sandy mud with peat fragments and a concentration of *Hydrobia* indicates deposition at slightly higher energy conditions in a later phase.

These dates suggest that at the location of the K and L cores the peat was not covered by deposits during the initial formation of the young channels (between about 2400–2000 cal. BP, Baeteman, 2005a), but occurred during the period the final fill was formed, and at least about 250 years after the end of the peat growth in that area. The sediments deposited during the initial infill of the channel are apparently missing here. Foraminifera

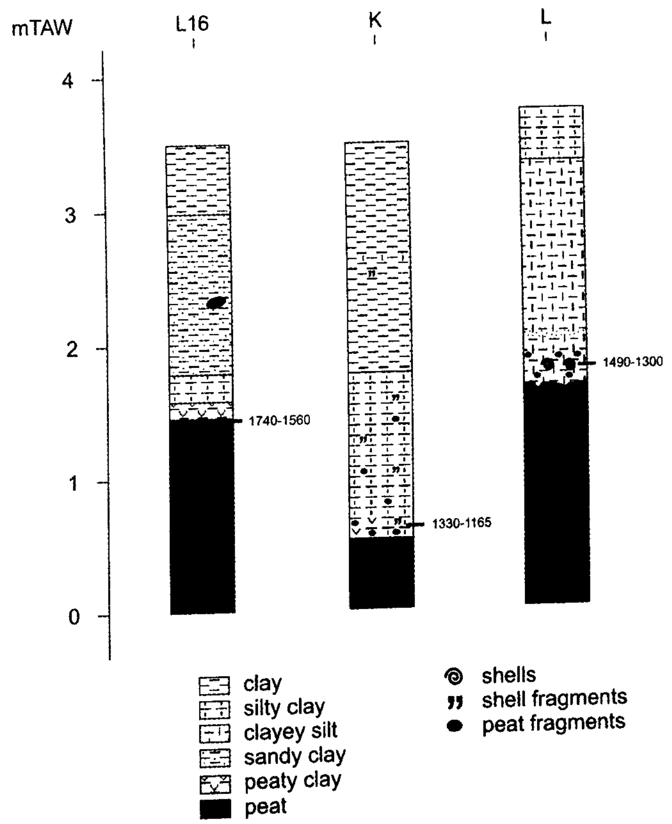

Figure 7 Stratigraphy of the post-peat deposits in cores at more distal reaches of the sand-filled channels showing the position of the dated samples. See Figure 2 for location

analyses indicated that the deposits overlying the peaty clay in core K are characterized by tidal sedimentation with strong open marine influence (Baeteman *et al.*, 2002). Since no evidence of erosion has been observed in the post-peat deposits in the two cores, it is assumed that during the initial infill of the channels, no sediments have been deposited at the K and L locations.

Discussion

Mechanisms and processes during the late-Holocene coastal evolution

The study of the shallow outcrops in the proximity of sand-filled tidal channels has shown the complexity of the post-peat deposits.

The complexity is apparent from the facies changes and their lateral relations, together with the ages of the shells. The major evidence observed in the outcrop and core examples can be summarized as follows: (1) the initial inundation of the peat was at first not associated with the deep cutting of the channels; (2) multiple reworking of the final fill of the channels occurred without deep vertical erosion; (3) some of the changes in the coastal landscape happened rapidly; (4) lateral migration of the channels took place once they were almost filled; (5) a time and sediment gap is found in the areas at distal reaches of the channels.

Using the above evidence and the data from previous work, it is possible to document the mechanisms and processes that caused the facies variations, and the evolution of the changing coastal landscape during the last 2000 years. The renewed tidal expansion entered the area via the mid-Holocene tidal channels, which were meandering and landward branching across the peat bog to almost the landward edge of the plain. The channels were almost completely silted up and the major ones served as drainage for the peat bog and the higher-lying hinterland. As suggested by Baeteman (2005a), an excessive run-off from the hinterland caused by an abrupt climate change around *c.* 2800 cal. BP, involving an increase in precipitation, most probably combined with tree cutting in the hinterland during the Iron Age period, eroded the upper part of the mid-Holocene channels. Tree cutting in this period has not yet been documented in the study area, but is most likely by analogy with the observations of Smyth and Jennings (1990). Once the channels were superficially 'cleaned', tidal waters could enter again. At first the tidal waters inundated the peat in the areas adjacent to the channels. This happened with modest erosion of the peat along the channel banks. Moreover, in the areas adjacent to the channels seawater killed the peat growth which consequently lost its potential of retaining its water resulting in compaction of the bog (in the sense of a reduction in volume) adjacent to the channel. Both causes provided accommodation space for the intertidal sediments. Interlaminated mud and sand with peat detritus, indicative of tidal deposition, was formed. The *Scrobicularia* at the top of the intertidal deposits in the Steenkerke site indicate that the initial inundation of the peat happened before *c.* 2020 cal. BP. Although the erosion was modest, it had far-reaching consequences. Little by little, it caused drainage and de-watering of the peat bog with compaction and collapse of it in the vicinity of the channels. This resulted in a gradual increase of the tidal prism to which the channels adapted by enlarging their cross-section. Because the peat has a great erosion resistance (Allen, 2000), the channels scoured deeply into the easily erodable sand of their predecessors (the sand-filled mid-Holocene channels) because the young channels reoccupied the same location (Baeteman, 2005a). However, tidal currents in the channels detached large masses of peat that slumped down and eventually became buried in the channel fill. During this high-energy phase, the channel network enlarged erosively further responding to the continuously increasing tidal prism. Progressively, larger parts of the peat bog were attained and collapsed because of de-watering. This eventually contributed to the expansion of an increasing dense channel network in the major part of the plain.

The time and sediment gap suggests that some areas experienced a long period of slow or no sediment deposition. The sites where the peat collapsed, most probably became subtidal with a minimum of sediment deposition. It is assumed that this is the result of a negative sediment balance. All available sediments were used to fill the deeply scoured tidal channels. Moreover, during the 2–3 ka years of uninterupted peat growth, sea level has risen by about 2 m. This, together with the collapse of the peat bog, implies that a huge accommodation space first had to be filled before a state of dynamic equilibrium was reached between the channel cross-section, tidal prism, sediment supply and sea level. A portion of the necessary sediment came from the early- and mid-Holocene channel fills. However, since such a large volume was needed, the seaward area also supplied sediment. This explains the landward migration of the shoreline due to erosion of the tidal deltas and shoreface (Beets *et al.*, 1994).

Once the dynamic equilibrium between the controlling factors was attained, the sediment surface reached an intertidal position. The youngest ages of all for the top of the peat indicate that the peat growth came to an end and was replaced by a tidal flat in the entire plain between about 1600 and 1500 cal. BP. While the major part of the plain changed into a tidal flat and silted up, the final fill of the channels developed under low-energy conditions, which happened in the period between *c.* 1400 and 1200 cal. BP. Because of the silting up of the area and the very weak relative sea-level rise, accommodation space was no longer created. Therefore, the channels started to migrate laterally through meandering (cf. van der Spek and Beets, 1992). This caused shallow erosion and reworking of the final fill and the adjacent intertidal flats. The available dates indicate that this reworking of the final fill did not happen simultaneously in the various outcrops. The dates of the sites Voetbalveld and Schorestraat are quite similar, ie, shortly before and slightly after *c.* 1275 cal. BP, respectively. The reworking in the Wulpen section, however, happened after 1000 cal. BP and at about 900 cal. BP in Steenkerke. Other examples most probably would give other dates.

Finally, the channels and tidal inlets silted up to a large extent, but yet not completely, and the channel networks contracted. Archaeological and historical evidence document that from the tenth century AD on, the major part of the plain consisted of salt marshes evolving into salt meadows, which were progressively embanked as well as some of the channels that had become sufficiently shallow (Ervynck *et al.*, 1999; Loveluck and Tys, 2006). Consequently, the reduced cross-sectional area of the channels, because of the sediments and/or the dikes, could no longer accommodate a large volume of tidal water. When storms occurred, the storm-induced currents during the return flow after the storm set-up, eroded and reworked the final fill again. Historical data document that storms were frequent in the eleventh and twelfth centuries AD (Vos and van Heeringen, 1997). The effect of storms in terms of facies variation, however, was restricted to the channels and nearby areas. Moreover, not every channel reacted in a similar way because of the interplay of local factors that control their dynamics, and therefore, the facies changes and the time periods are not identical. This implies that sea-level fluctuations were not responsible for the facies variations recorded in the coastal deposits.

Conclusions

The indicative meaning of facies changes in the late-Holocene tidal back-barrier deposits of the Northwest European coastal lowlands has seldom been considered in terms of processes and mechanisms, but rather on tendencies of sea-level movements or sea-level fluctuations. Some exceptions exist, such as the work by Hoffmann (1998), Vos (1999), Evans *et al.* (2001) and Beets *et al.* (2003). However, facies changes should not necessarily be associated with transgressions and regressions or sea-level fluctuations. This paper shows that the facies variations reflecting changes in the coastal landscape in the area studied were caused by sedimentary processes whereby tidal channels played a major role. The record of the changes does not involve rise and fall of sea level.

This paper also documents that the interpretation of the lithological changes in the post-peat sediments requires an integration of the observations and age determinations in a larger stratigraphic context taking into consideration the dynamic nature of the tidal channels and the prevailing conditions about sediment supply and accommodation space. Lithological variations are not caused by simply one influential factor, but result from the interplay of all contolling factors. Therefore, the facies changes and time period of the changes are not similar over the entire region. The processes of the channel networks, ie, their initiation and evolution, are similar, but the changes happened at different times in different places because of the progressive expansion of the channel network and the non-uniform compaction of the peat and mud.

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