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How subsoil morphology and erodibility influence the origin and pattern of late Holocene tidal channels: case studies from the Belgian coastal lowlands

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Abstract

This paper aims at a better understanding of the late Holocene re-entrance of the tidal system in the Belgian coastal plain after a 2–3 ka years long period of peat growth. The re-entrance was associated with the development of deeply incised tidal channels. The initial cause of the re-entrance, still a missing link in the understanding of the late Holocene coastal change, is investigated by focusing on the specific location of the young tidal channels. The research, though based in the western Belgian coastal plain, is relevant to the lowlands of the southern North Sea and English Channel. The investigation combines stratigraphic, radiocarbon and sedimentological data, together with maps showing the morphology of the pre-Holocene surface, the distribution of the Holocene deposits and the erosional surface produced by the late Holocene channels. It appears that the young channels re-occupied the same position as their early and mid Holocene predecessors. The re-entrance of the tidal system began by a removing of the upper part of the older channels which originated from the mainland. Once the channels were cleaned of sediment, tidal waters could re-enter and rework the easily erodable sand of the older channels and Pleistocene deposits.

1. Introduction

The objective of this paper is to better understand the controls on the late Holocene coastal changes in the lowlands of the southern North Sea. In general, the deposits of the late Holocene in the Belgian coastal plain comprise a 1–2 m thick, apparently homogeneous tidal-flat mud. This final infill was formed by a renewed expansion of the tidal environment after a period of 2–3 ka years of almost uninterrupted peat growth. The expansion of the tidal flat environment was associated with the formation of tidal channels which eroded deeply into the mid and early Holocene sediments, and sometimes into the underlying Pleistocene deposits.

Similar late Holocene tidal channels, which also deeply incise into older Holocene sediments at about the same time, have been reported from the lowlands of the southern North Sea and English Channel. Most of them are mentioned occasionally without further comments, or are shown in stratigraphic cross-sections (e.g. Streif, 1972; Baeteman, 1985; Beets et al., 1992; Wheeler and Waller, 1995; de Groot et al., 1996). Few of them have been discussed in a stratigraphical context (van der Spek, 1996; Vos and van Heeringen, 1997; Cleveringa, 2000; Beets et al., 2003). In contrast, erosive features associated with the channels, such as erosion of the upper peat bed and presence of peat inclusions in the overlying tidal flat deposits, are frequently described (e.g. Long and Innes, 1993; Long et al., 1998; Brew et al., 2000). Although numerous sedimentological and hydrodynamic investigations of young infilled channel deposits have been published (e.g. Terwindt, 1981; Roep, 1991; van der Spek, 1995; Allen, 2000a, b; Stupples, 2002;), no research has so far been undertaken

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with the specific intention of examining the processes which control the initiation of the late Holocene tidal channels.

1.1. Late Holocene channel formation

The renewed entrance of the tidal environment has been a subject of discussion since about a decade (Long and Innes, 1993; Beets et al., 1994; van der Spek, 1996; Vos and van Heeringen, 1997; Long et al., 1998; Baeteman, 1999; Baeteman et al., 2002). However, the initial cause and timing of the end of the peat formation and the accompanying marine inundation still remains unclear. Sea-level rise was at an average rate of 0.07 m/ka without a sudden change in that period (Denys and Baeteman, 1995), and changes in the wave climate and tides after ca. 6000 BP, seem to have been limited (Beets et al., 1992; Cleveringa, 2000). A progressive flooding from an expanding tidal basin and associated channel network, together with erosion of the upper levels of the peat, is suggested as explanation for the inundation of the peat. All papers agree that the renewed expansion of the tidal environment happened progressively, marking a landward movement of the marine/brackish conditions together with shoreface erosion and landward shift of the shoreline. Only Beets et al. (1994) and Vos and van Heeringen (1997) propose some suggestions about the initial cause and timing of the re-entrance in the Netherlands. Their ideas have been repeated by Baeteman (1999) and Brew et al. (2000).

According to the hypothesis of Beets et al. (1994), in the period between ca. 7000 and 3000 cal BP, the supply of sediment from marine sources to the tidal basin kept pace with, or exceeded, sea-level rise, leading to sedimentary infilling and coastal progradation. However, after ca. 3000 cal BP, the continued depletion of sediment sources from the offshore led to a deficit in supply relative to sea-level rise. To compensate this, new sources had to be found, and previously deposited Holocene sediments of the shoreface were reworked. This led to shoreface erosion and landward migration of the shoreline and tidal basin. Vos and van Heeringen (1997) explained the inundation of the peat in Zeeland (SW Netherlands) as a result of breaching of the coastal barriers and formation of the drainage pattern on the peat bogs during the Iron Age and Roman Period. The natural process of inundation of the peat surface was enhanced by digging ditches and by the excavation of large quantities of peat for industrial purposes.

For the Belgian coastal plain, the idea of sediment deficit as initial cause for the re-entrance of the tidal system seems less plausible since the offshore area still contains sand in the form of tidal ridges. More importantly, however, is the huge amount of sand that was deposited in the extensive and deep late Holocene tidal channels. It is most unlikely that the shoreface and coastal barrier could have delivered these large volumes of sand. The excavation of large quantities of peat as the cause of the renewed expansion of the tidal environment, as proposed by Vos and van Heeringen (1997), is not plausible neither. In the Belgian plain, peat digging in the Roman Period is recorded (Ervynck et al., 1999), but not in the Iron Age Period, and there is evidence that the re-entrance happened at least before ca. 2300 cal BP. For the Fenlands of East Anglia, England, the idea of barrier breaching is also inappropriate (Brew et al., 2000).

This overview demonstrates that more work is needed to identify the processes that led to the initial cause of the re-entrance of the tidal system. The aim of this paper, therefore, is to examine how these channels initiated by investigating their specific location in relation to the morphology and lithology of the subsoil. The approach is novel in that it combines the use of stratigraphical, radiocarbon and sedimentological data, together with the sequence map of the Holocene deposits and contour map of the Pleistocene subsoil. The investigation does not concern the seaward part of the plain, where the difference between sand of the channels and that of the surrounding sandflats, inlets and barrier deposits is hard to distinguish in borehole records.

2. Study area and data

The western Belgian coastal plain (Fig. 1) is a 15–20 km wide embanked lowland, at an elevation that varies from +2 to +5 m TAW. The Belgian ordnance datum, TAW, refers to mean lowest water spring, which is about 2 m below mean sea level. Holocene coastal deposits are no longer present above about +5 m TAW. The plain is bordered by coastal dunes and crossed by a small river, the IJzer with its tributaries, the Handzame, the Kemmelbeek and the Sint Jansbeek. The coast is characterised by a meso–macrotidal regime, with a mean spring tidal height not exceeding +5 m TAW.

The Holocene deposits of the western coastal plain have been investigated extensively in the past 25 years within the geological mapping programme of the Belgian Geological Survey. The mapping has been carried out mainly using boreholes. The characteristics and ages of the shallow deposits were primarily taken from temporary outcrops. From the mapped area (ca. 450 km^2, Fig. 1), there is a data set of about 100 mechanically drilled cores covering the entire Quaternary sequence, and over 1150 hand-operated gouge augers giving undisturbed cores varying in depth from 2 to 12 m below ground level. About 25 mechanically drilled undisturbed cores penetrated infilled channel deposits, reaching their base and beyond. The hand-operated gouge augers helped to delineate the lateral

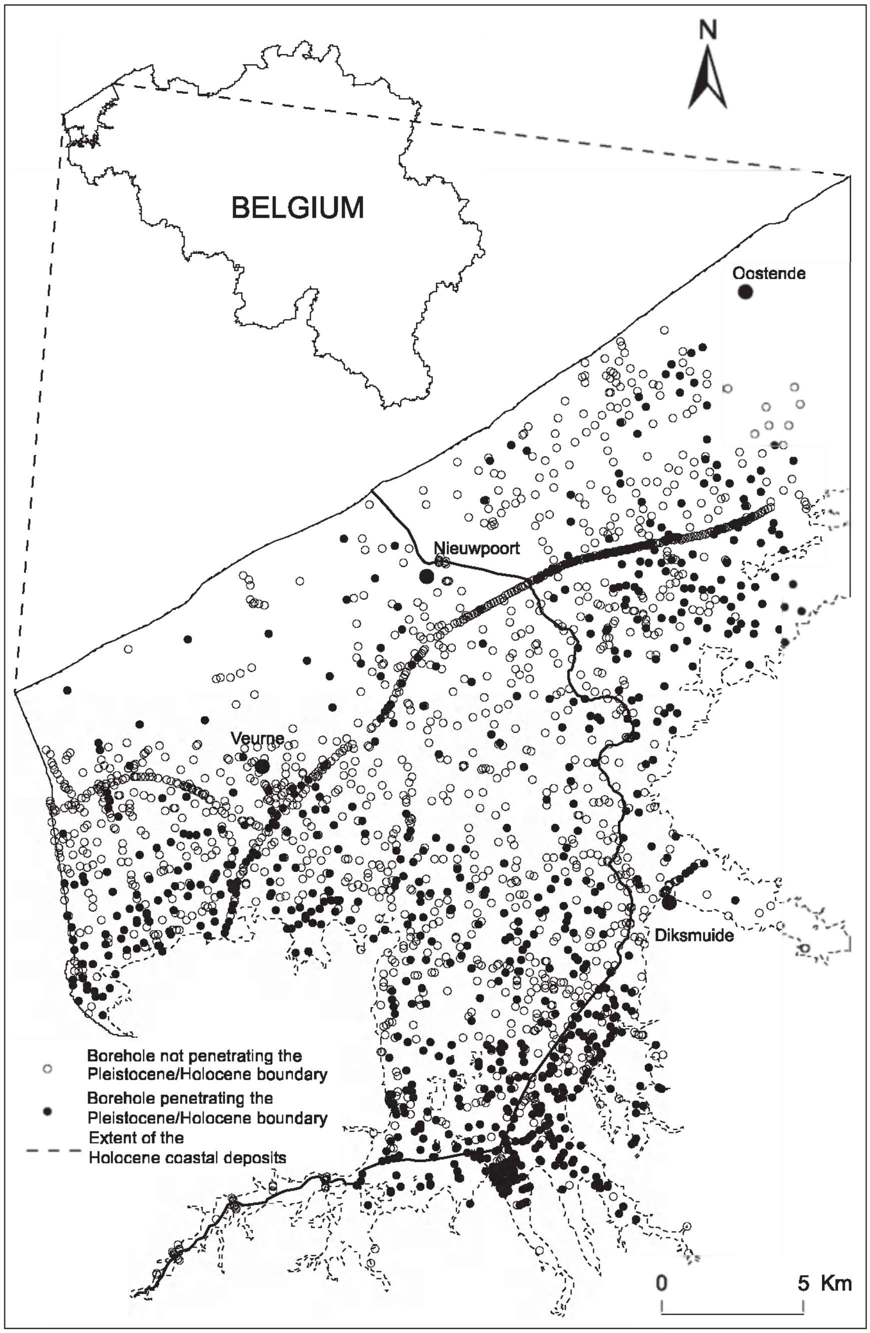

Fig. 1. Map of the western coastal plain of Belgium with location of the boreholes.

extension of the channels. The entire data set forms the basis for the construction of the maps and for the interpretations presented in this paper.

The sedimentary sequences in the cores and outcrops were described and analysed using the same criteria for facies identification, which is on the basis of lithology, sedimentary structures and macrofossils. Diatom analyses of several cores (Denys, 1993) confirmed the field interpretation of the different facies units.

3. Methodology

An understanding of coastal evolution requires more than a detailed microfossil and sedimentological analysis of a point-specific site, since a core shows the record of just one single spot in a complex sedimentary mosaic. The interpretation of single cores and, more particularly, distinct vertical and lateral changes in sedimentary facies, requires the integration of many cores into the entire depositional body of the tidal basin. Therefore, the spatial distribution, or geometry of the various facies must be elaborated in stratigraphic cross-sections whereby every subtle change in facies is of importance (cf. Kraft and Chrzastowski, 1985). Although this paper presents only four cross-sections as a matter of demonstration, the elaboration of the various maps presented and the ideas discussed below, is based on the linkage of numerous cross-sections, some of them previously published (Baeteman, 1985, 1991, 1993, 1999, 2001). To understand the interrelationship of the stratigraphical settings, the following maps will now be compared: the morphology of the pre-Holocene surface, the sequence map of the Holocene deposits, and the morphology of the base of the Holocene deposits (Fig. 2).

3.1. *Morphology of the pre-Holocene surface (Fig. 2)*

An esstential tool to understand the particular distribution of the various facies of the Holocene deposits, is the knowledge of the morphology of the pre-Holocene surface (Beets and van der Spek, 2000; Baeteman and Declercq, 2002). The morphology of the pre-Holocene surface reflects the original topography at the beginning of the transgression. In the study area it is rather easy to recognise the Holocene/Pleistocene boundary. In general, the Upper Pleistocene deposits have a fluvial origin consisting of relatively consolidated non-calcareous clay and fine very silty sand. Moreover, the boundary is represented by the basal peat that formed as the water table rose in advance of the Holocene marine transgression. The basal peat is absent in those areas where late Holocene tidal channels eroded deeply (Fig. 3), and in the seaward part, where in the middle Holocene, tidal scour removed most of the previously deposited Holocene deposits (Baeteman, 1999). Data about the Pleistocene/Holocene boundary is also missing where the Holocene sequence is too thick or sandy, so that the boundary is beyond hand auger reach. The original pre-Holocene surface is therefore reconstructed partly on exact data, and partly on interpretation between areas that were affected by tidal scour and basal peat removal. Because of the many constraints discussed in Baeteman and Declercq (2002) and the specific methodology for such an interpretation, the contour map must be constructed manually and not by geostatistical software. This map (Fig. 2) covers the entire western plain, in contrast with the sequence map (Fig. 4).

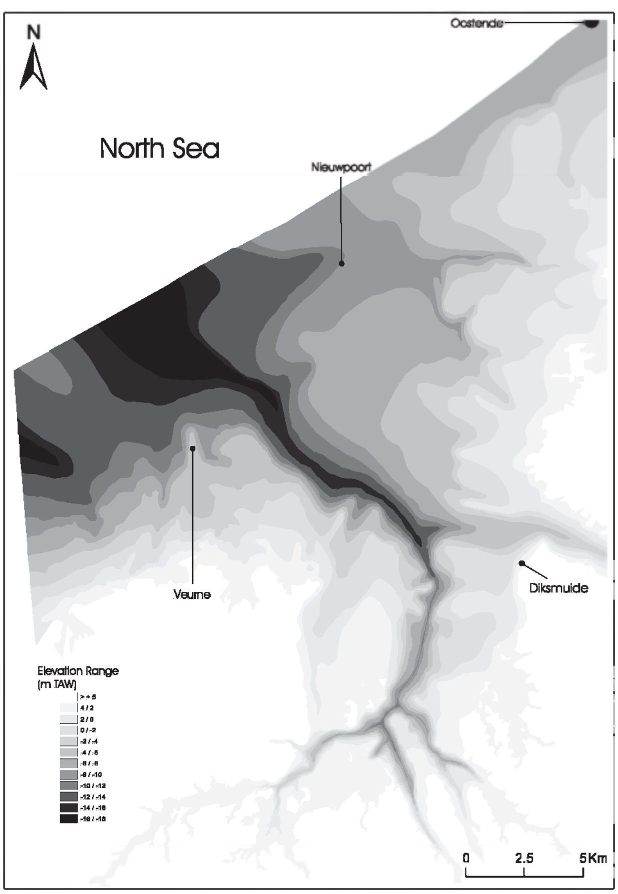

Fig. 2. Morphology of the pre-Holocene surface at a 2 m interval relative to TAW. Due to the vertical exaggeration the map gives a misleading impression of the relief of the drainage pattern. In reality, the valleys are shallow with gentle slopes (after Baeteman and Declercq, 2002).

The landscape prior to marine flooding in general shows a gentle seaward slope. The landscape is mainly characterised by a drainage pattern of small rivers joining in the central part and forming a southeast–-northwest depression. This drainage pattern is interpreted as palaeovalleys, that were important as conduits (in the form of tidal channels) for water and sediment during the Holocene flooding of the area.

3.2. *Sequence map of the Holocene deposits (Fig. 4)*

All the borehole data have been registered in a geographical information system (GIS) using the lithogenetic classification system developed by Streif for mapping purposes (Barckhausen et al., 1977; Streif, 1978, 1998; Baeteman, 1981; Bertrand and Baeteman, 2005). This system allows the construction of a sequence map representing the entire Holocene sequence in the form of profile types. The sequence map only covers the central and landward part of the western plain.

This map shows the general distribution of the various facies. The map reflects the classical picture of the building-up of a tide-dominated coastal plain sequence that developed in the pre-existing palaeovalleys (Bertrand and Baeteman, 2005). Along the limit of the plain and in a restricted part in the south, the X1

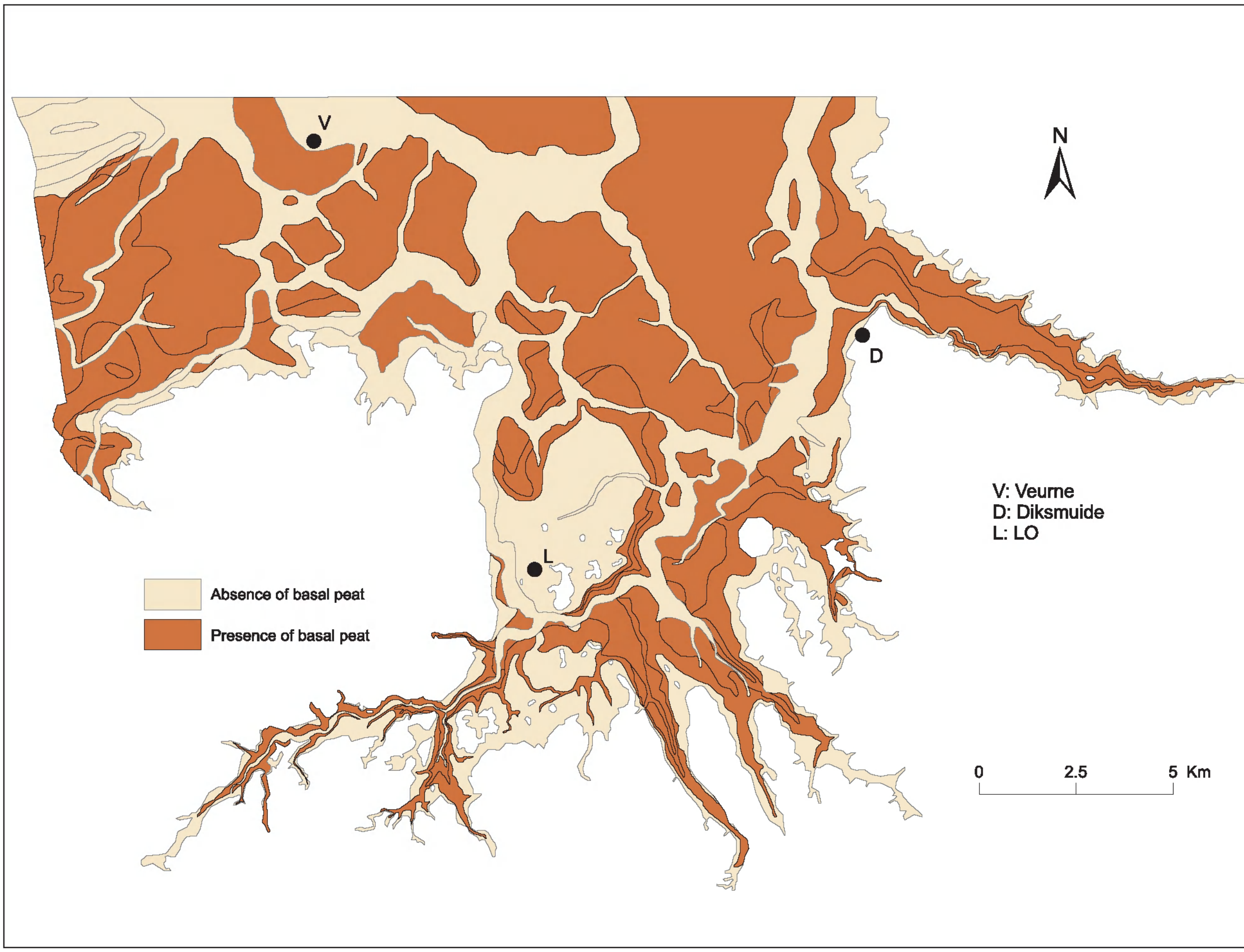

Fig. 3. Distribution of the basal peat in the study area.

type occurs, i.e. only a thin cover of (late) Holocene mud is present. This results from the morphology of the Pleistocene subsoil. When the latter is at an elevation higher than about +2.5 m TAW, basal peat did not develop (Baeteman, 1999). From the limit of the plain towards the centre of the map, a typical lateral succession of the following profile types is seen: X1, X2, Z3 and Y2. This particular distribution results from the presence of the palaeovalleys of the IJzer and Handzame rivers. The depressions of the pre-Holocene surface were flooded by the early Holocene transgression which led to the continued accumulation of mud in the tidal flats and sand in the channels. After ca. 7800 cal BP, intercalated peat beds developed that alternated with mud. These areas are represented by the Y2-profile type. Where the Pleistocene surface gradually rises, all the peat beds merge to form a sequence composed entirely of peat (with a thin clastic cover), represented by the Z3-type. A special profile type (X11) has been introduced to demonstrate the distribution of the late Holocene infilled channels. Only the major channels are represented, since the mapping of all the minor channels requires an even denser boring grid. The map shows that the channels developed a widely spread network that extended far landwards.

3.3. Morphology of the base of the Holocene deposits (Fig. 5)

This contour map depicts the elevation of the base of the Holocene deposits (Fig. 5). The map was constructed partly on exact data, and partly on interpreted data considering the effects of late Holocene erosion, which removed major parts of the Pleistocene deposits, and in many places, all of them until the stiff and compacted Eocene clay. These places are reflected in the over deepened areas. The basal depth of the channels reach −23 m TAW, but usually varies between –15 and –20, −10 and –15 and –6 and –8 m TAW in, respectively, the seaward, central and landward part of the study area. The wide extent of the channel deposits (shown on the sequence map, Fig. 4) and the deep incision is apparent for the large tidal channels.

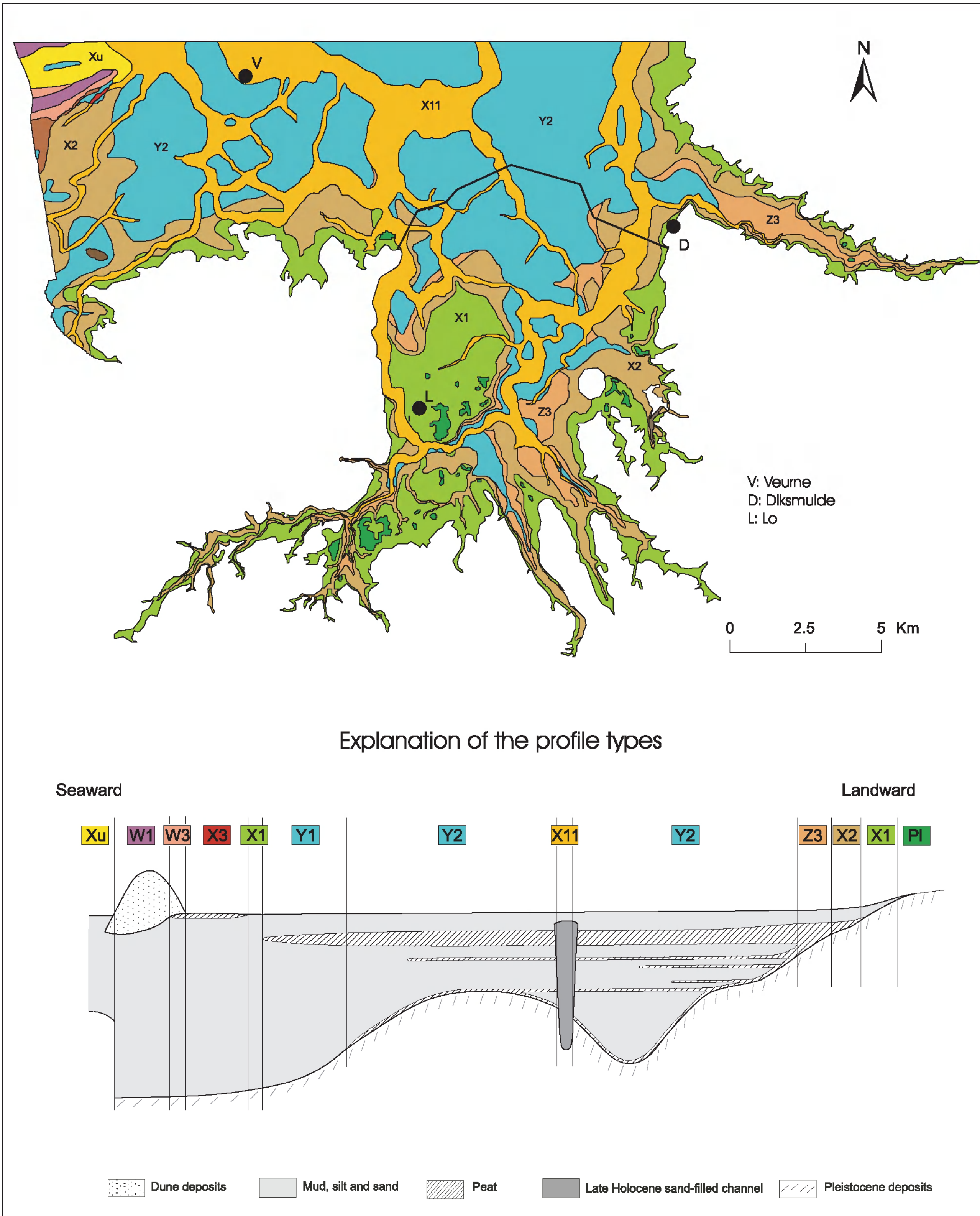

Fig. 4. General sequence map of the central and landward portion of the western coastal plain and explanation of the profile types. The location of cross-section 4 (Fig. 11) is indicated on the map.

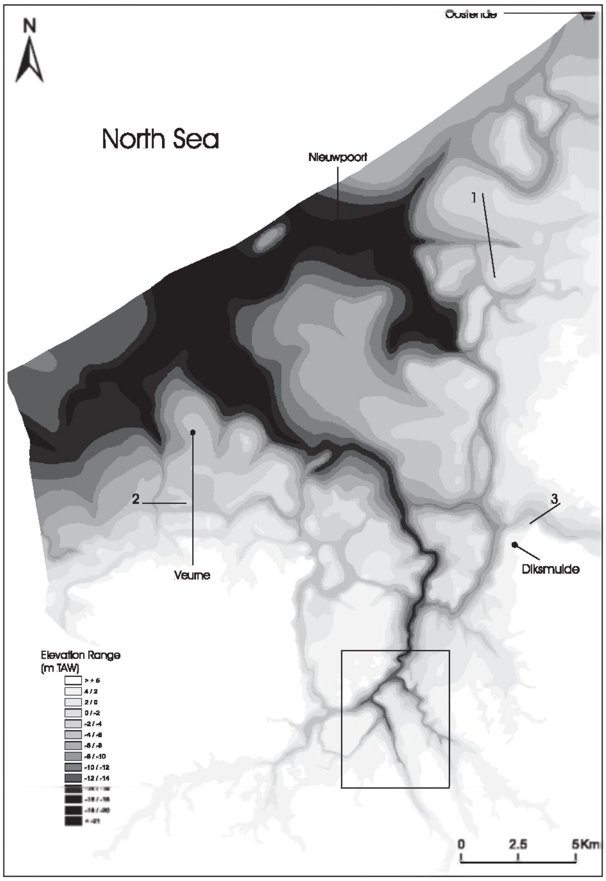

Fig. 5. Isohypse map of the base of the Holocene deposits at a 2 m interval relative to TAW. The lines with numbers 1–3 indicate the location of the geological cross-sections, the inset shows the location of Fig. 12.

4. Results and interpretation

4.1. Lithology and sedimentary characteristics of the late Holocene infilled channels

The full thickness of the late Holocene channel deposits varies between 5 and 25 m. Their lower boundary is always abrupt and erosive. Two different phases of deposition can be distinguished: the major lower part that comprises the bulk of the channel fill, and the overlying part, here called the final fill of the channel (Fig. 6). However, the latter is only present in the smaller channels while in the larger channels, the deposits with the characteristics of the lower part, extend to the surface (see Fig. 11).

The overall texture of the lower part of the infilled channels consists of fine sand to silty fine sand with mainly an evenly laminated bedding, which alternates with zones comprising small-ripple bedding where the flasers consists of a concentration of peat detritus. Also present are erosional unconformaties and chaotic sequences, usually 5–15 cm thick, which contain numerous mud pebbles, peat boulders and disarticulated bivalve shells embedded in fine sand. All shells originate from tidal environments. Several channel lags with mainly articulated *Cerastoderma* shells and mud pebbles could occur at different depths in the same fill. Most probably, they reflect a complex of smaller stacked channels or periodic reactivations of the channel during overall infilling. The lowermost channellag, containing mainly peat fragments, mud pebbles and few shell fragments, often does not occur at the very bottom of the channel, but frequently 5–10 cm above it. Bioturbation is completely absent throughout the sequence of the lower part. The deposits of these channel fills contain numerous peat fragments which vary in size from millimeters to decimeters. The fills located in the more landward area frequently contain parts of trees broken up by thin sand laminae. Some fills also show slumped blocks of tidal-flat mud, in which a thin peat bed in situ may be preserved. The absence of mud in the channel deposits indicates that tidal flow was important. Few cores at the side of the channels show point bar deposits. They are characterised by inclined tidal bedding which consists of millimetre- and centimetre-interlayered silty sand/mud with laminae of concentrated peat detritus and centimetre-thick mud layers. All these characteristics clearly indicate a major vertical downcutting event followed by relatively rapid subtidal deposition with erosion and sedimentation alternating.

The final fill of the channels (about the upper 2 m) often consists of a rapid alternation of cm-thick mud and fine sand beds in a flaser and wavy bedding, more or less strongly burrowed by *Scrobicularia* and *Cerastoderma*. Coarsening-fining upward microsequences, attributed to neap–spring tide variations, are well developed (Terwindt, 1981). These characteristics imply rather low-energy conditions with a relative low rate of sedimentation. Some final fills show longitudinal cross-bedding with mud drapes, characteristic of pointbar or shoal deposits of migrating channels. In some places, this sequence overlies the (partly) eroded peat bed. The final fill in turn also shows shallow erosion and/or reworking in some cases (Baeteman et al., 2002).

From the sediment description above, it is suggested that the development of the late Holocene channels starts with a vertical down cutting followed by relatively rapid subtidal infilling. The final fill is characterised by lateral sedimentation due to migration.

4.2. Chronology of the down-cutting channels

The channels described above post-date the period of main peat formation (between ca. 6000 and 3830/1525 cal BP, Baeteman, 1985, 1991, 1999; Baeteman et al., 2002). Reworked peat in the channel deposits have been dated at different locations and depths. Their ages

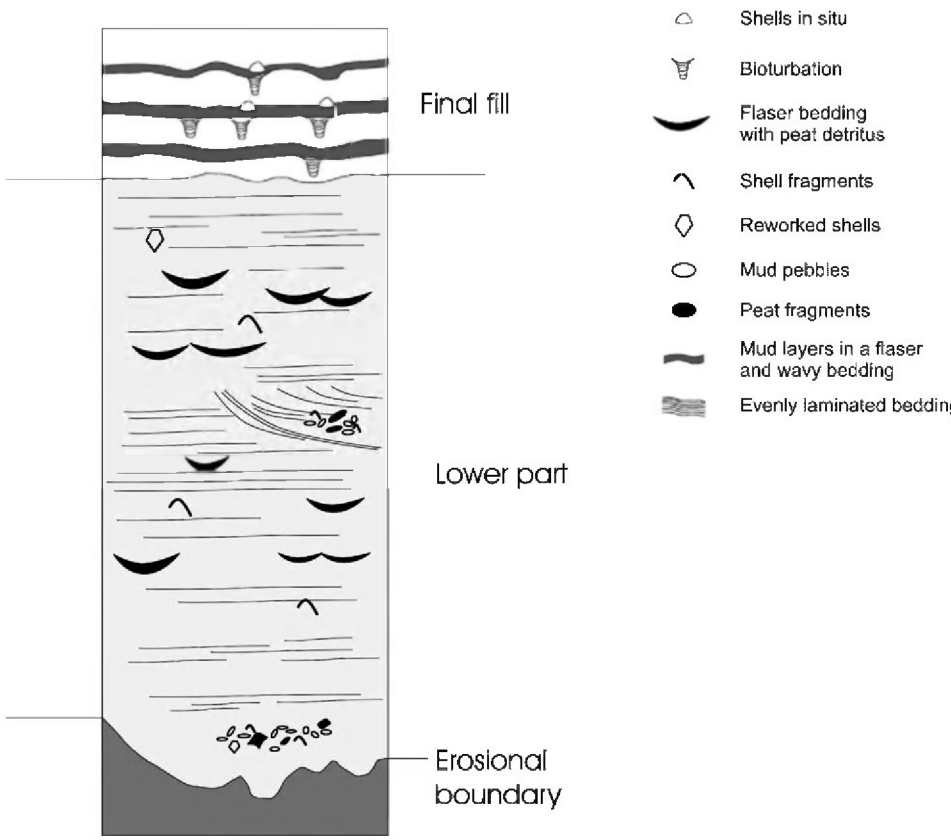

Fig. 6. Schematic sedimentary log of the late Holocene channel fill sequence.

(see Table 1 and Fig. 7) demonstrates that the peat is reworked from the upper peat bed.

Only two reworked *Cerastoderma* shells collected at the same elevation in a lag deposit of channels cutting through the peat are dated at ca. 2278 and 1775 cal BP (see Table 2). Although the channels are at short distance from each other, and the lags are at broadly the same level, there is a significant difference in age. The reworked shells indicate that a nearby mudflat was stripped. However, the depth from which the deposits/shells were removed is unknown. Deeper removal will give an older age than shallow erosion. The age of the shells indicates that the incision happened prior to at least ca. 2280 cal BP.

Shells in living position (*Scrobicularia* and *Cerastoderma*) as well as reworked ones from lags have been dated from the final infill of the channels (Table 2 and Fig. 7). The age of the in situ shells vary between ca. 2085 and 1190 cal BP, with a cluster of dates between about 1400 and 1200 cal BP (Baeteman et al., 2002). The extremely young date of 560 cal BP is an exception, which comes from *Scrobicularia* at the bottom of a shallow mud-filled creek representing the very end of all channel activity. The reworked shells in the final fill give an age between 2285 and 895 cal BP. From the dates available for the end of the peat (Baeteman et al., 2002), it was suggested that as from around ca. 3400 cal BP, the peat growth was directly affected by the main channels. The younger dates of the end of the peat mark a landward movement of tidal conditions associated with the shift of the channel network.

The available ^{14}C dates of the reworked shells in the main body of the channel fill are not sufficient to date the initial formation of the channel. The shells in the final fill, however, indicate that the late Holocene channels developed at least before ca. 2280 cal BP. It should be mentioned that the shells come from different channel systems which probably experienced the same process, but maybe at a different time.

4.3. *The geometry of the infilled channel deposits and their surrounding setting*

To understand the geological setting of the late Holocene channels, three detailed sections crossing a major channel are presented and a larger-scale section demonstrating the infill of the palaeovalley and down-cutted channels. It is true, the construction of a cross-section involves a certain measure of subjectivity. However, the lack of stratigraphic data in areas between the boreholes is interpreted on the basis of borehole records in the nearby area. The cross-sections have been

Table 1
Radiocarbon dates of reworked peat in infilled channel deposits

	Site	Sample depth (m)	Sample altitude (m TAW)	Age (^{14}C years BP)	Calibrated ages (yr BP) 2σ range	Laboratory number
1	Kapelhof	7.60	−3.42	2680±60	2975–2600	IRPA 540
2	Avekapelle	9.90	−5.72	3890±70	4580–4130	IRPA 539
3	Wolleboom	7.00	−3.00	3360±60	3820–3500	IRPA 557
4	Mechelhof	9.95	−4.75	2870±60	3205–2810	IRPA 565
5	Langeleed	10.60	−6.60	4880±70	5825–5340	IRPA 679

The location of the sites is shown in Fig. 7. Dates are calibrated using the calibration program of Stuiver and Reimer (1993).

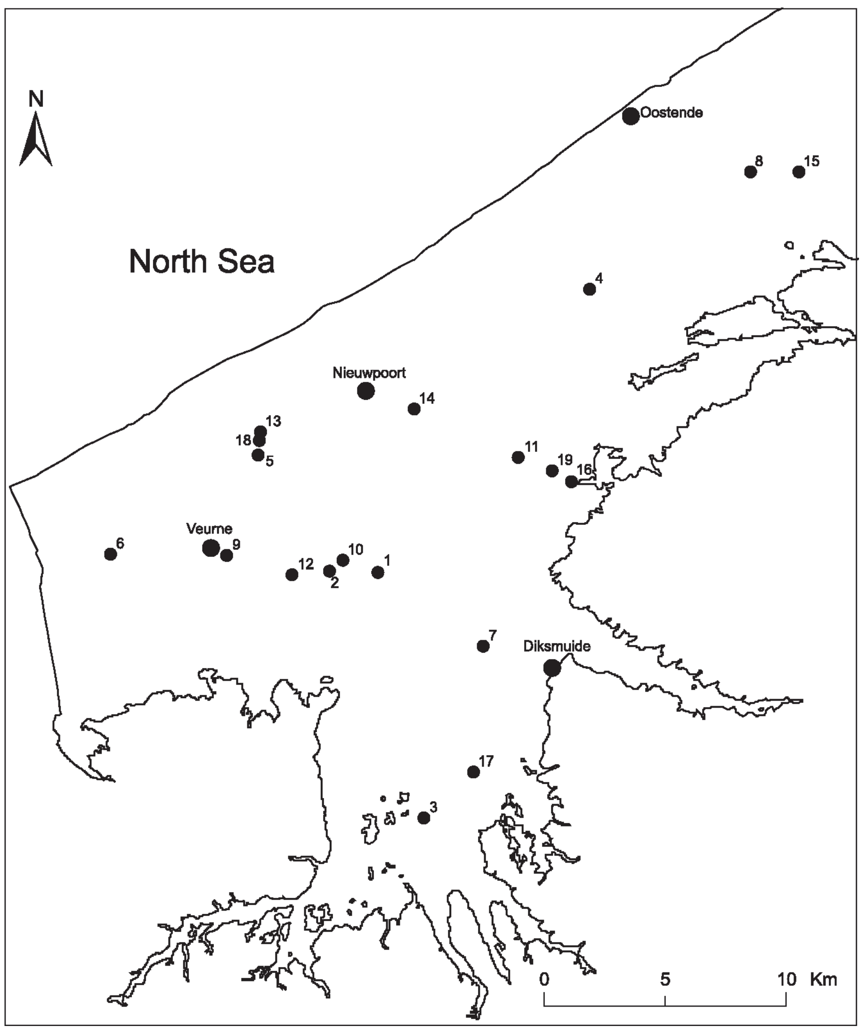

Fig. 7. Position of locations where peat and shells were sampled. The numbers refer to the number of the site in Table 1 and 2.

selected in different areas where sandflat deposits are absent, so all sand in the subsoil is from a tidal-channel origin. The cross-sections are made with a strong vertical exaggeration in order to show the rapid facies changes in the vertical sequence.

Cross-section 1 (Fig. 8) crosses a west–east tidal channel, east of the city of Nieuwpoort (cf. Fig. 5). None of the cores reaches the base of the channel sand. Boreholes 807 and 540, and 562 and 531 show two down-cutted channels. At the edges, the channel eroded the peat partly (boreholes 533, 539, BC203). The presence of the main peat bed (although excavated partly) overlying sand in borehole 806, indicates that another generation of channel is present. The latter is situated in a depression formed in the Pleistocene deposits (Fig. 2). The evidence for that is given in the

Table 2
Radiocarbon dates of the shells in infilled channel deposits

	Site	Dated material	Sample altitude (m TAW)	Age (^{14}C years BP)	Calibrated age BP	2σ range	Laboratory number
Reworked shells in lag deposits							
6	Kromfort 1	*Cerastoderma*	+1.60	2575±30	2278	2332–2119	UtC 5538
6	Kromfort 6	*Cerastoderma*	+1.60	2180±40	1775	1886–1615	UtC 5385
Shells in living position in final fill							
7	Kaaskerke	*Scrobicularia*	+0.53	1700±25	1260	1329–1165	KIA 12248
8	Zandvoorde	*Scrobicularia*	+0.90	2330±35	1937	2068–1823	KIA 13561
9	Veurne	*Scrobicularia*	+1.60	2450±60	2084	2290–1920	UtC 3939
10	Avekapelle	*Cerastoderma*	+1.80	1730±60	1278	1421–1131	UtC 9430
10	Avekapelle	*Scrobicularia*	+1.85	1840±50	1372	1513–1273	UtC 9429
11	Schorestraat 4	*Scrobicularia*	+1.90	1755±25	1289	1388–1214	UtC 4355
12	Steenkerke	*Scrobicularia*	+2.00	2400±50	2021	2188–1867	UtC 9417
12	Steenkerke	*Scrobicularia*	+2.60	995±30	560	644–514	Irpa 1279
13	Voetbalveld 2	*Cerastoderma*	+2.90	1810±40	1329	1474–1252	Irpa 1206
14	Nieuwpoort	*Scrobicularia*	+3.00	1635±30	1189	1275–1077	UtC 4349
13	Voetbalveld 1	*Scrobicularia*	+3.14	1740±40	1279	1386–1171	Irpa 1205
15	Plassendale	*Cerastoderma*	+3.20	2090±40	1688	1797–1534	KIA 12067
15	Plassendale	*Cerastoderma*	+3.40	2150±40	1748	1865–1597	KIA 12066
16	Lekebek	*Scrobicularia*	+3.80	2340±60	1946	2130–1797	UtC 4148
Reworked shells in final fill							
8	Zandvoorde	Scrobicularia	+0.90	2575±30	2285	2335–2126	KIA 13563
17	Lamp. A42	Scrobicularia	+1.80	1850±25	1384	1492–1299	KIA 12256
18	Wulpen E	Cerastoderma	+2.38	1730±35	1274	1367–1168	UtC 4682
11	Schorestraat 2	Hydrobia	+2.60	2065±35	1618	1759–1506	UtC 4353
8	Zandvoorde	Cerastoderma	+2.90	1950±40	1548	1628–1358	KIA 12065
11	Schorestraat 1	Scrobicularia	+2.60	1730±25	1275	1355–1179	UtC 4351
19	Wale 1	Cerastoderma	+3.40	1805±30	1325	1450–1254	UtC 4673
12	Steenkerke	Cerastoderma	+2.50	1330±50	895	988–728	UtC 9428

The location of the sites is shown in Fig. 7. Dates are corrected for a marine reservoir effect of 400±40, and calibrated using the calibration program of Stuiver and Reimer (1993).

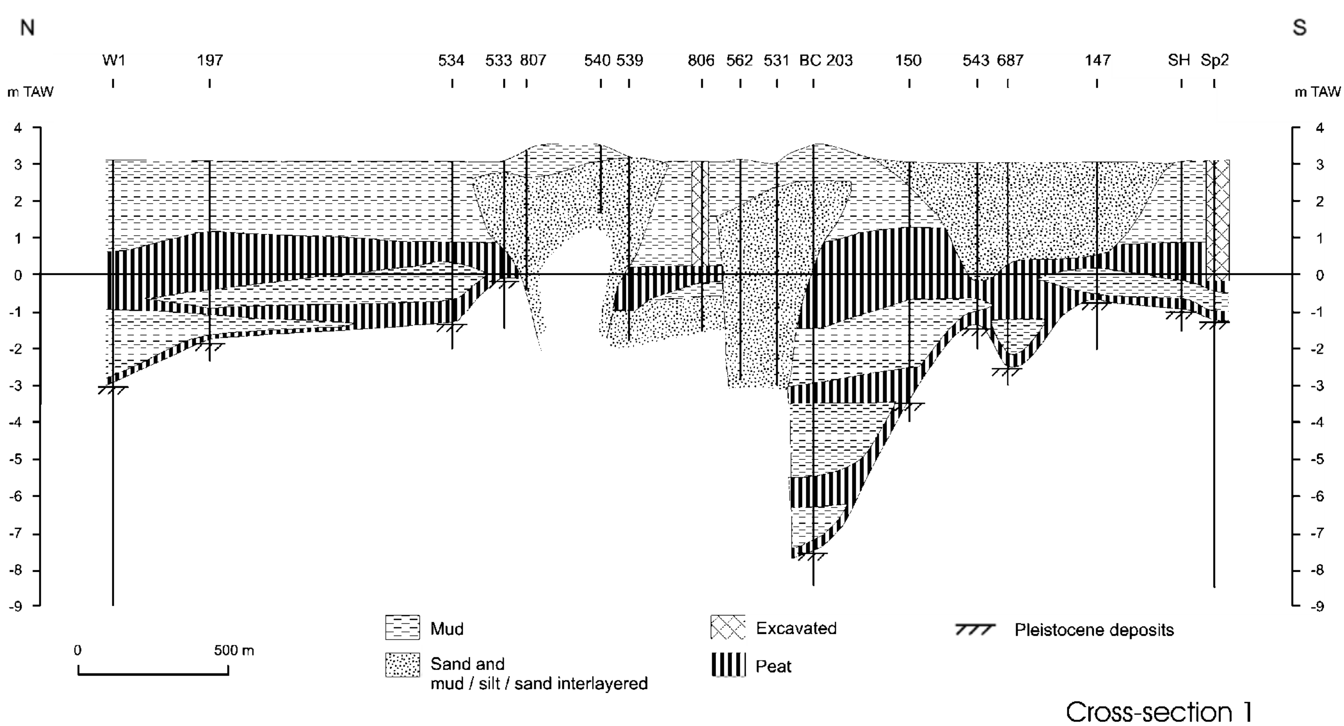

Fig. 8. Cross-section 1 through the palaeovalley east of Nieuwpoort showing a late Holocene tidal channel incision and a portion of the mud and peat fill indicative for the existence of a depression in the pre-Holocene surface. See Fig. 5 for location.

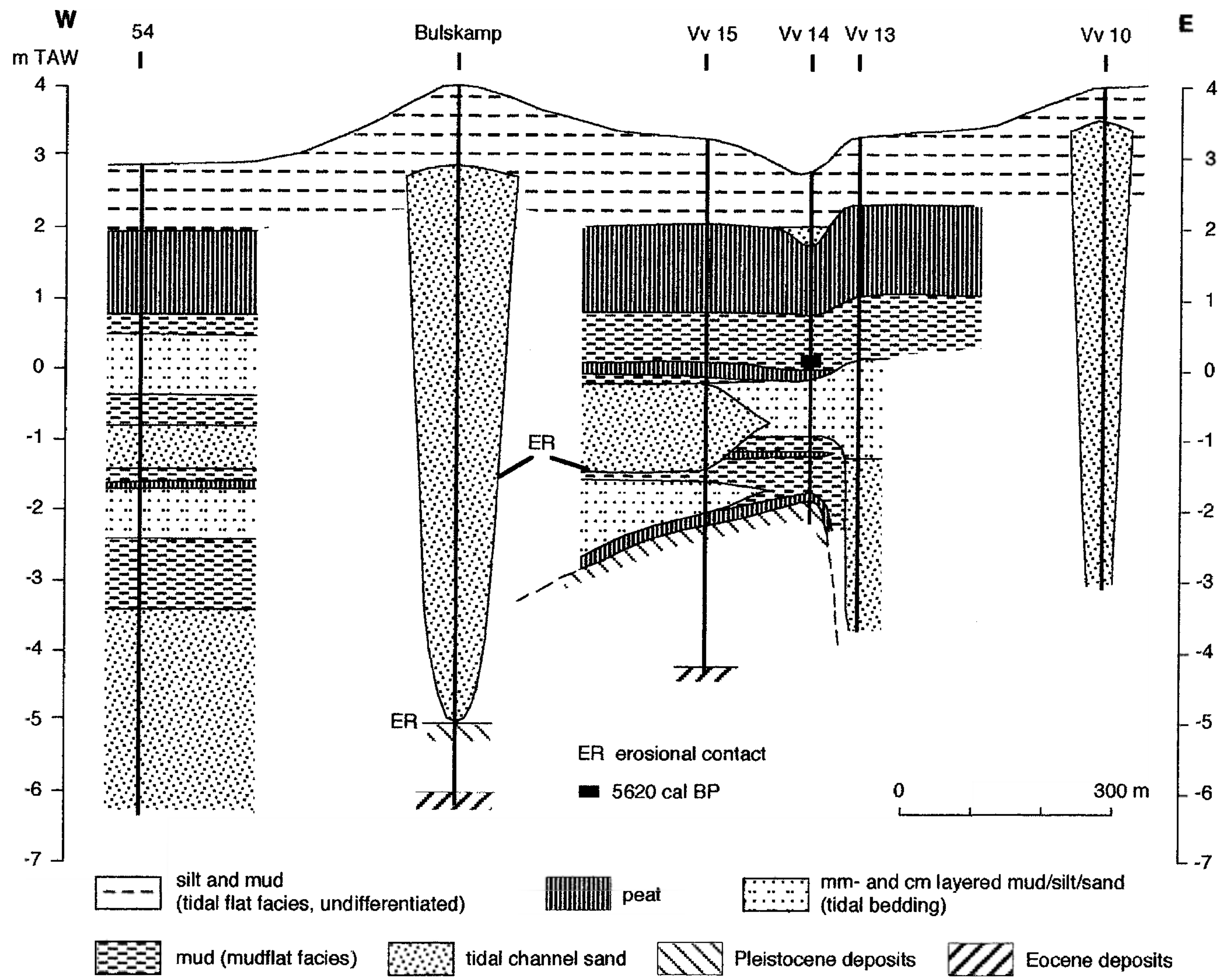

Fig. 9. Cross-section 2 south of Veurne showing two down-cutted tidal channels and evidence for older channels in the boreholes nearby. See Fig. 5 for location.

adjacent core BC203 showing a 11 m thick complete Holocene sequence with mud and peat beds. The presence of basal peat indicates that the pre-Holocene surface is not erosional.

Cross-section 2 (Fig. 9), about 1.5 km long, is located south of Veurne and crosses a north–south running tidal channel. The section shows two peat-cutting channels (boreholes Bulskamp and Vv10). The Bulskamp channel reaches the Pleistocene deposits, but part of it has been eroded as suggested by the top Pleistocene in boreholes Vv15 and Vv 14.

The boreholes adjacent to the down-cutted channels show a complex stratigraphy. In boreholes 54 and Vv13, the lower part consists of tidal-channel sand. The basal peat (in boreholes Vv15 and Vv14) and channel sand (in boreholes 54 and Vv13) are overlain with a sequence wherein intercalated peat beds are interlayered with sand and thinly laminated sand/silt/mud. The latter are tidal-bedding deposits which develop subtidally in point-bars or at the outer edge of a tidal channel, or which reflect the silting-up phase of a channel. The sand is the result of overbank deposits from a tidal channel. The sequence between –2.5 and +0.5 m shows evidence that another channel generation was active adjacent to the location of the late Holocene one. A *Scrobicularia* in living position at 0 m was dated at 5620 cal BP (5753–5543 cal BP) indicating that the tidal channel was active well before 5600 cal BP.

Cross-section 3 (Fig. 10) crosses the Handzame valley north of Diksmuide (Fig. 5). Today, this valley still acts as a freshwater drainage. The cross-section clearly shows a major depression in the Pleistocene subsoil reaching –4 m TAW over a distance of about 600 m. Half of the depression is filled with peat with basal peat at the bottom and, in the lower part, two flooding phases which originate from a tidal channel nearby. The lower

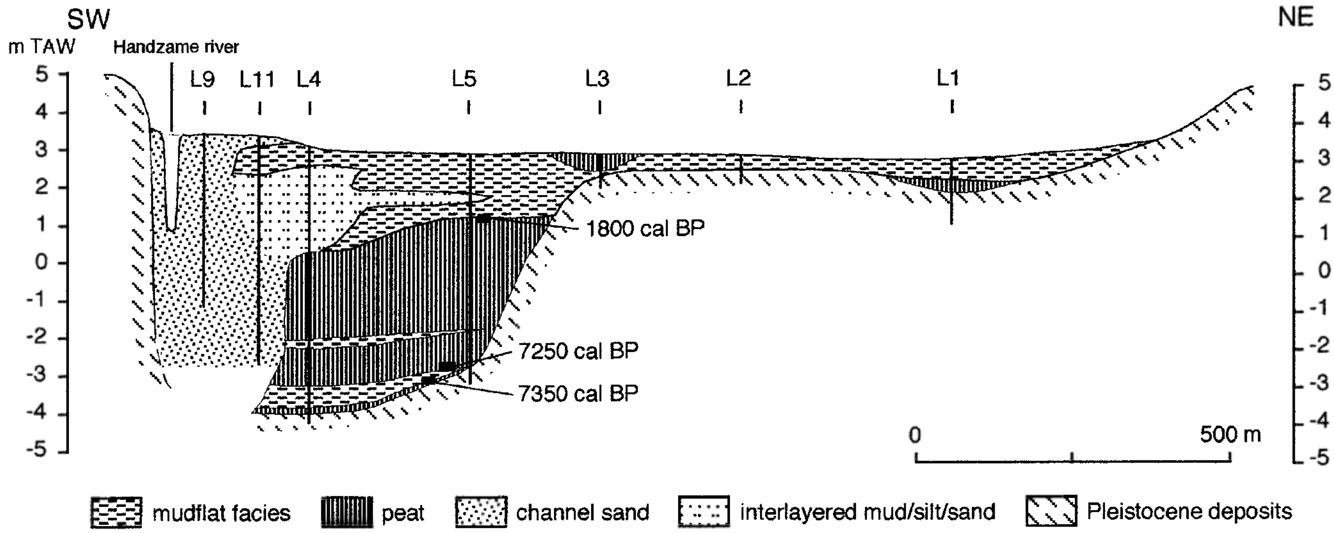

Fig. 10. Cross-section 3 north of Diksmuide showing a late Holocene channel incision into the peat and mud fill of a palaeovalley. See Fig. 5 for location.

one is dated at 7350–7250 cal BP. The major part the peaty infill has been eroded by a late Holocene tidal channel. The latter is filled with fine sand. The base of it was not recovered. At one side of the channel, the sand grades into interlayered mud/silt/sand displaying a tidal bedding, probably representing a pointbar. This part belongs to the final fill of the channel when it migrated laterally. The thin lateral extension suggests local overbank flooding.

The cross-section demonstrates that the late Holocene channel incised into the fill of a pre-Holocene palaeovalley. The flooding phases in the lower part prove that nearby a tidal channel must have been present, at least in the middle Holocene. However, no deposits of this channel have been found. The incision happened before 1800 cal BP, which is the age of the end of the peat growth.

Cross-section 4 (Figs 11 and 4) shows the palaeovalley of the IJzer in which an early Holocene tidal channel developed after a period of mudflat deposition covering the basal peat dated at ca. 9000 cal BP (Baeteman, 1999). After a silting up phase with a short-lived peat growth between ca. 7800 and 7585 cal BP, the channel became active again, at about the same position. At the site of borehole Oostkerke, the channel remained open until ca. 5470 cal BP, although silting up, and there followed a short-lived episode of peat growth (not drawn in the section), that occurred at ca. 5825 cal BP. This is the date that the thick peat bed in general started to develop across much of the Belgian coastal plain. The section shows several peat-cutting channels. Those without borehole, are inferred from the sequence map, as is the width of the channels. Only one down-cutted channel re-occupies the same position as the palaeovalley. The two large channels located in the very west and east, are incised in an area close to the outcropping Pleistocene deposits consisting of fine silty sand. Whether or not the channels are confined to a palaeochannel is not known, because all the possible evidence has been reworked. However, it is more a question of being lucky rather than the rule to encounter the channel sand of the early and mid Holocene fill of a palaeovalley by means of boreholes. In many cases, only the sand of the down-cutted channel is found.

5. Discussion

The cross-sections discussed above demonstrate that the late Holocene tidal channels incised into the fill of a palaeovalley. At three of the four sites considered, it is clear that the late Holocene tidal channels re-occupied the same location as their early and mid Holocene predecessors. The comparison of the morphology of the pre-Holocene surface, and of the base of the Holocene deposits, shows that the palaeovalleys have been over deepened in the late Holocene. Their fill of easily erodable sand were clearly favourable locations.

The opposite is also true. One of the palaeovalleys in the very south of the study area is filled with a 16 m thick sequence of peat and gyttja (Baeteman, 1993, 1999). No traces of channel sand has been found despite a very dense boring grid (see Fig. 1). The simplified profile type map of that area (Fig. 12, profile type 3) shows that the channel fill of that particular valley escaped from later erosion and no late Holocene tidal channel developed.

The channels where no evidence of re-occupation was found (cf. Fig. 11), incised in areas located close to the present-day landward limit of the plain. At the time of incision, these areas were characterised by a thin peat

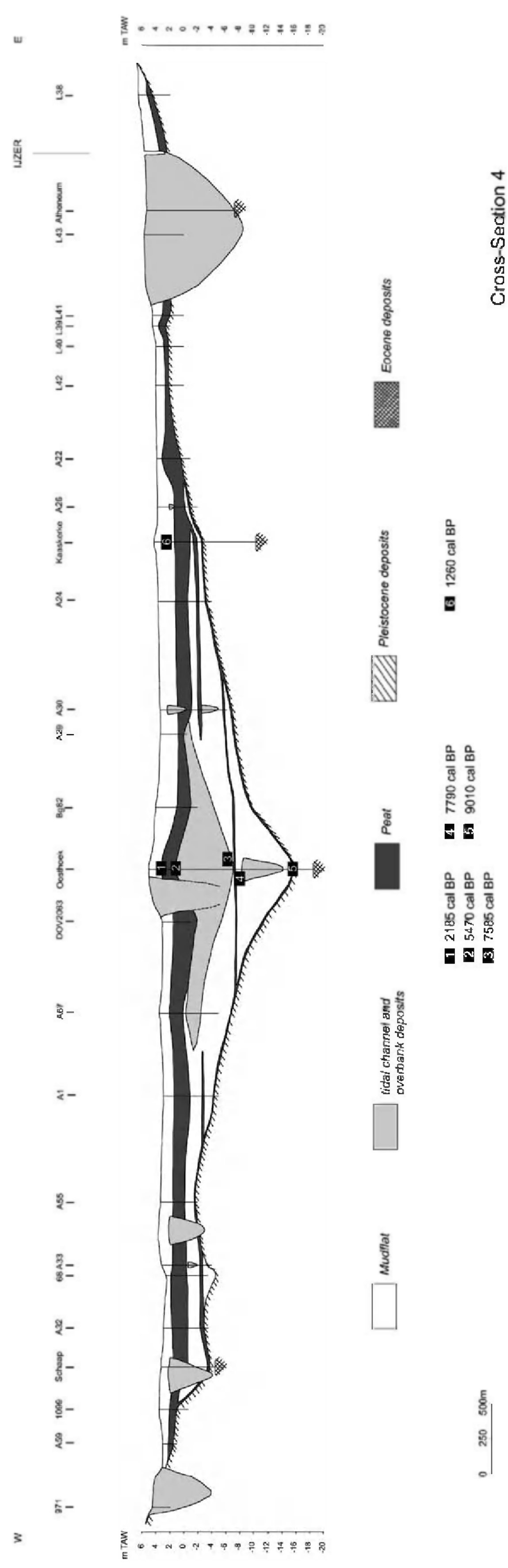

Fig. 11. Cross-section 4 crossing the palaeovalley of the IJzer (borehole Oosthoek) where an early and mid Holocene tidal channel developed. The western and eastern parts of the section demonstrate a major down-cutted late Holocene channel. See Fig. 4 for location.

bed (X2 on the sequence map) covering relatively high-lying Pleistocene deposits. According to the Quaternary map (Bogemans and Baeteman, 2003), the pre-Holocene deposits here consist of fluvial fine sand and silt, also easily erodable.

The re-occupation of the late Holocene channels in the same location as the older channels can explain the cause of the initial entrance of the tidal waters. During the 2–3 ka year period of mid Holocene peat accumulation, the coastal plain was a vast freshwater swamp beyond the reach of almost all tidal activity. In the neighbourhood of the city of Oostende, peat developed beyond the present-day coastline indicating coastal progradation in that period. The major part of the peat was groundwater dependent and its accumulation could keep pace with the sea-level rise which was at an average rate of 0.7 m/ka in that period (Denys and Baeteman, 1995). Seawards, the freshwater marsh was bordered by salt marshes, mudflats, sand flats and a coastal barrier. The latter environments were crossed by tidal channels feeding the tidal environment in relation to the sea-level rise in the 2–3 ka long period. These channels, however, also crossed the freshwater marsh. This has been documented by traces of brackish water in the thick peat bed, which indicates that the freshwater swamp has been influenced by the flooding from a tidal channel (Denys, 1993). These traces were found in the most landward parts of the plain suggesting that the channel network which developed during the early and mid Holocene, still existed. The later date for the beginning of peat growth in the Oostkerke borehole (cf. Fig. 11) is additional evidence. However, in the freshwater marsh, the channels mainly served as freshwater drainage for the terrestrial run-off. Without drainage, the peat would rapidly be inundated and drowned. The main body of the remaining channels (outside the peat area) has never been found in boreholes but, from the discussion above, it is assumed that such channels must have been present. Actually, the sand from the older channels cannot be found any more, because their sand has been reworked during the late Holocene incision.

The drainage channels in the freshwater marsh were almost completely silted up during the period of mid Holocene peat accumulation. We know this because, had they not, they would have drained and dewatered the peat bog which would then have been unable to accumulate for such a long period. The cross-sections show that the level of the sedimentation surface in the channels while they were still tidally active, is relatively high and at about the same level on which the peat started to develop.

Because of the high silted up position of these channels, tidal waters were not able to penetrate landwards. Possible high-energy events most probably only affected the tidal flat and the freshwater swamp in the seaward parts. The only way that the tidal water

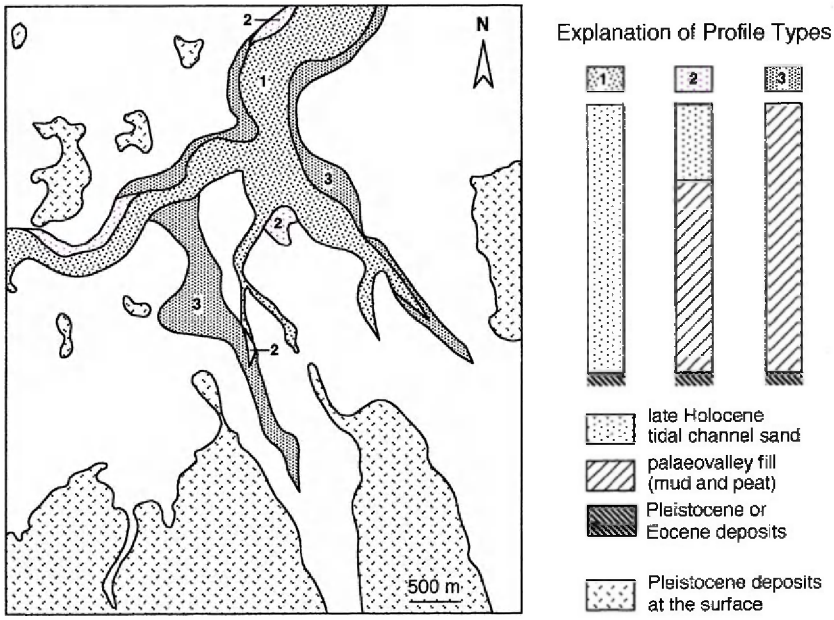

Fig. 12. Simplified sequence map of the southern part of the study area showing that a palaeovalley, filled with peat and gyttja without evidence of an older tidal channel, has not been affected by late Holocene erosion. The part of the map which is left blank, consists of a thin cover of late Holocene clay with or without basal peat. See Fig. 5 for location (after Baeteman, 1999).

could enter the remaining channels, is the creation of space for it. Therefore, it is suggested that the early and mid Holocene pre-peat channels have been "cleaned", at least the upper part of the infill. It is unlikely that the cleaning process originated from the sea. This would have required a very high water level during which water stored in the plain reached far above normal high-water level. In contrast, erosion in the channels only happens during falling water levels, when huge amounts of water is discharged. The conditions for a higher than normal discharge occurs when a storm from the northwest coincides with the flood phase (Baeteman et al., 1999). However, it is inconceivable that one major ebb-surge storm scoured all the channels, unless one must assume a series of successive storms in a short timespan. Moreover, an ebb-surge storm produces shifts in the position of the channel and crevasse splays (Cleveringa, 2000). For the lower part of the late Holocene channel fill, no evidence of lateral shift of the channel has been found so far. On the contrary, their course remained largely confined to the position of the older channels. The effect of a heavy storm in the freshwater marsh which happened probably before ca. 2400 cal BP has been observed but it was restricted to small, short-lived crevasse channels only (Baeteman et al., 1999).

Because tidal waters reached the landward limit of the plain from the beginning of the re-entrance (Baeteman et al., 2002), it is more likely that the cleaning process came from the mainland and was associated with an environmental change. As discussed above, the late Holocene channels re-occupied the position of the early and mid Holocene channels, which in turn occupied the palaeovalleys present in the Pleistocene subsoil. These palaeovalleys actually are the continuation of the small valleys in the mainland draining a basin with elevations varying in general between +10 and +20 m TAW with some higher areas up to +40 m TAW in the southeast.

A hypothesis of excessive run-off from the mainland is proposed as mechanism of erosion of the upper part of the infill of the remaining channels. The excessive run-off concentrated in the valleys of the mainland and continued in the remaining channels scouring the upper part of the easily erodable sand. The excessive run-off could have been caused by a climatic change like an increase in precipitation. In NW Europe, a climate change at the Bronze Age–Iron Age transition around 850 cal BC (2800 cal BP) has been detected (van Geel et al., 1998; van Geel and Berglund, 2000). The abrupt climate change from relatively warm and continental to cooler and wetter conditions resulted in a considerable increase of effective precipitation and a decline of temperature after the climate shift. It is still to be investigated whether also antropogenetic activity, such as tree cutting during the Iron Age period, is to be considered as additional cause of the excessive run-off. However, no data are available on this matter from the study area.

Once the channels were sufficiently cleaned, tidal waters could enter again and began to rework the easily erodable sand of the older channels. The peat bog in the vicinity of the channels was thereby affected. Erosion and collapse at the banks of the channels also happened whereby blocks of peat and mud slumped into the channels. This produced dewatering of the peat with

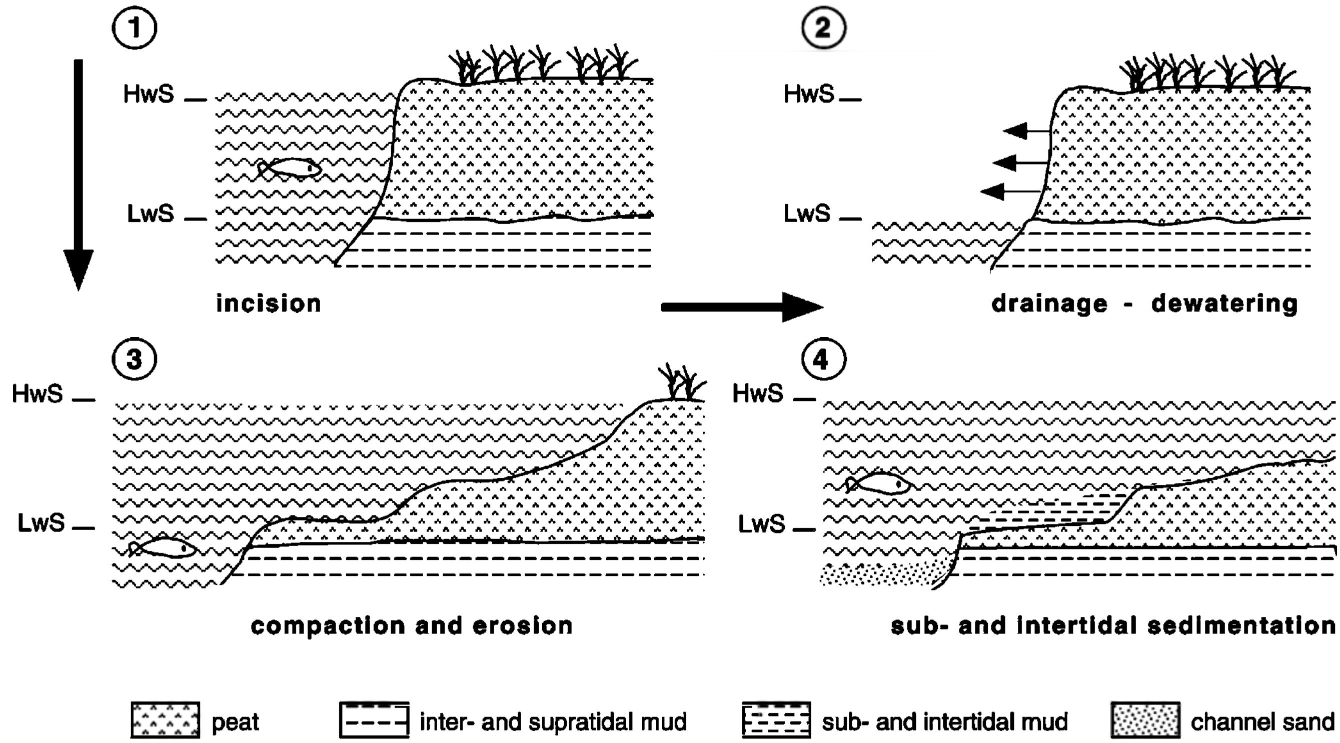

Fig. 13. Schematic presentation of the mechanism of incision of a channel resulting in dewatering of the peat with consequently compaction and lowering of its surface causing an increase in the tidal prism of the channel.

consequently compaction and lowering of the surface in the areas bordering the channel (Fig. 13). This resulted in an increase of the tidal prism of the channel which adopted its size by vertical scour (Eysink, 1990; Cleveringa, 2000). The underlying sandy Pleistocene deposits were eroded and reworked during this phase of channel incision. Meanwhile, new tributaries from the main channels developed because of the increase of the tidal prism. Therefore more peat areas subsided, and the channel network progressively enlarged and prograded landward.

The hiatus of about 500–900 years between site inundation and deposition of the overlying (post-peat) sediments suggests that the peat area bordering the channels came into a subtidal position with minimal sedimentation. At first, all the available sediment was used to fill the deepened channels. The sedimentary characteristics point to a rather rapid infill, although no dates are available for the infill of the initial incised channels. From the available dates, it is assumed that the vertical down-cutting and rapid infill happened somewhere around 2300 cal BP. After the rapid infill, a period of low-energy conditions prevailed until about 1400 to 1200 cal BP. This period is represented by the final fill whereby the channel cross-section, sediment supply, and tidal prism were in equilibrium. By that time, the sedimentation surface in turn reached an equilibrium with the sea level and the plain changed into an inter- and supratidal setting. The channels started to migrate laterally reworking the deposits of the salt-marshes and mudflats that developed in the seaward areas as well as the upper parts of the channel fills and the peat. Foraminiferal assemblages show that in this phase, open marine sediments were brought in, pointing to shoreface erosion (Baeteman et al., 2002). This most probably happened on the occasion of storm surges or catastrophic flows. This suggests that the shoreface erosion, as put forward by Beets et al. (1994) for the initial cause of the re-entrance of the tidal system, happened in a later phase and is not the initial cause of it.

6. Conclusion

The combind stratigraphic, radiocarbon and sedimentological data, together with maps showing the morphology of the pre-Holocene surface, the morphology of the erosional base and distribution the Holocene deposits provided the basis for a better understanding of late Holocene coastal change in the Belgian coastal plain. The investigation reveals the basic control for the specific location of the incised late Holocene tidal channels. Knowledge about the origin of the specific location in turn, provide additional information about the initial cause of the re-entrance of the tidal system, the missing link in the understanding of a major change in the coastal evolution in the late Holocene.

In the study area, the pre-Holocene surface began to be inundated at about 9400 cal BP. Early Holocene tidal channels developed in the palaeovalleys without any evidence of erosion, as indicated by the almost ubiquitous presence of the basal peat. The channel and mudflat deposits conformably overlie the basal peat. There was no need to erode, because sufficient accommodation space, created by the rapid sea-level rise, was available. With the sea-level rise, the plain continued to aggradate in an inter- and supratidal setting, whereby the mid Holocene channels followed the same course as their early Holocene predecessors. From about 6000 cal BP, tidal sedimentation was gradually replaced by peat development which lasted for about 2–3 ka years. During that period, the tidal channels were still present, but they were silted up. They served as drainage of the freshwater marsh. The environmental conditions that caused the end of the peat are not known yet.

This investigation suggests that the late Holocene re-entrance of the tidal system was not caused by vertical scouring of the channels from the beginning. This is in contrast with the generally proposed idea about the re-entrance. Rather, it was preceded by a cleaning of the upper part of the sediments in the existing early and mid Holocene channels. In view of the re-occupation of the late Holocene channels in the same location as their predecessors, which in turn is a continuation of the mainland drainage, it is assumed that the cleaning process originated from the mainland. It most probably was caused by excessive run-off or floods. That a similar situation of re-entrance of the tidal system with down-cutting channels at about the same time is observed in the lowlands of NW Europe, suggests that the driving mechanism behind the environmental change reflects a regional process. The abrupt climate change around 2800 cal BP, followed by an increase in precipitation, could have been at the origin of the excessive run-off. Although the climate change was synchronous, the re-entrance of the tidal system was not. However, the exact dating of the re-entrance still remains uncertain. From the available data it is suggested that it happened around 2300 cal BP. It is critical to give an age for the initial down cutting of the channels. One should bear in mind that the different channel systems experienced the same process, but not necessary at the same time because the channels also react to local situations. Future work on the chronology of the late Holocene channels would benefit from the dating of the sand itself by optically stimulated luminescense. This investigation furthermore demonstrated that also deeper stratigraphic data must be considered for the explanation of young features.

Finally, the investigation highlighted the vulnerable areas in the present-day coastal plain. The observation that the late Holocene channels re-occupied the same position as their predecessors, suggests that the sand-filled channels with easily erodible sediments will similarly be affected in the case of possible future catastrophic flows and/or a possible future sea-level rise, in particular when dikes and dunes along the shore are not sufficiently high and/or resistant which is the case for the major part of the Belgian coast.

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