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Late Holocene coastal-plain evolution of the Netherlands: the role of natural preconditions in human-induced sea ingressions

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ABSTRACT

This paper demonstrates the decisive role of natural preconditions on the formation of large late Holocene sea ingressions in peaty coastal plains along the North Sea's southern shores. Geological and archaeological evidence shows that these sea ingressions (expansion of new tidal systems) were mainly caused by land subsidence, which occurred due to intensified agricultural use of artificially drained peatlands since the Late Iron Age (250–12 BC). This made the coastal plain sensitive to storm-surge ingression through weak spots, e.g., at the location of existing creeks, in the coastline.

Using The Netherlands as a case study, we show that natural preconditions (i.e., the geological setting at the time of ingression) played a key role in the pacing and extent of tidal area expansion. Ingressive tidal systems eventually reached most far inland in coastal segments with wide peaty back-barrier plains. In contrast, sea ingression formation was hampered in coastal segments with well-developed natural ingression-protecting geomorphic features (e.g., beach-barriers, supratidal levees). Feedback mechanisms, such as additional peat subsidence by loading of sediment imported into the new tidal area, caused further tidal volume increase and created accommodation space for tidal deposits. These combined effects caused irreversible sea ingression over large areas that consequently became unsuitable for habitation for many centuries.

Improved understanding of such sea-ingression mechanisms and their facilitating conditions are essential for the assessment of the sensitivity of many densely populated coastal plains, which experience major human-induced subsidence, eventually leading to coastal plain drowning.

1. Introduction

The Holocene evolution of coastal plains and barrier systems around the world is generally considered to be driven by inherited topography, post-glacial sea-level rise, background subsidence or uplift regime, and sediment distribution by waves, tides, and rivers (e.g., Goodbred and Kuehl, 1999; Giosan et al., 2006; Rossi et al., 2011; Hanebuth et al., 2012; Gao and Collins, 2014; Vos, 2015a). During the last millennia, simultaneously acting natural and human-affected sedimentary processes led to increased coastline progradation in several European deltas resulting from increased sediment load caused by deforestation (e.g., Po and Rhône delta—Stefani and Vincenzi, 2005; Maselli and Trincardi, 2013; Anthony et al., 2014), whereas in other regions human activity (e.g. the construction of dams) caused sediment deficit and subsequent flooding (e.g., Nile delta—Stanley and Warne, 1994; Nile and Ebro deltas—Syvitski et al., 2005; the coastal plain north of the Po delta—Zecchin et al., 2009). Nowadays many densely populated coastal plains experience major human-induced subsidence leading to coastal plain drowning (e.g., Törnqvist et al., 2008; Syvitski et al., 2009). In many northwestern European coastal plains, human-affected developments began relatively early (roughly 2000 years BP), coevally with episodes of transgression (*large-scale landward lateral expansion of back-barrier tidal depositional sedimentary environments*). These late-Holocene transgressions have been documented for the Flemish coast (Baeteman, 2005), the UK Fenlands (Brew et al., 2000), the UK Romney Marsh (Long et al., 2006), Northern Brittany, France (Regnauld et al., 1996), and the Bay of Biscay, France (Clavé et al., 2001). These authors discuss potential natural triggers such as intensified storm

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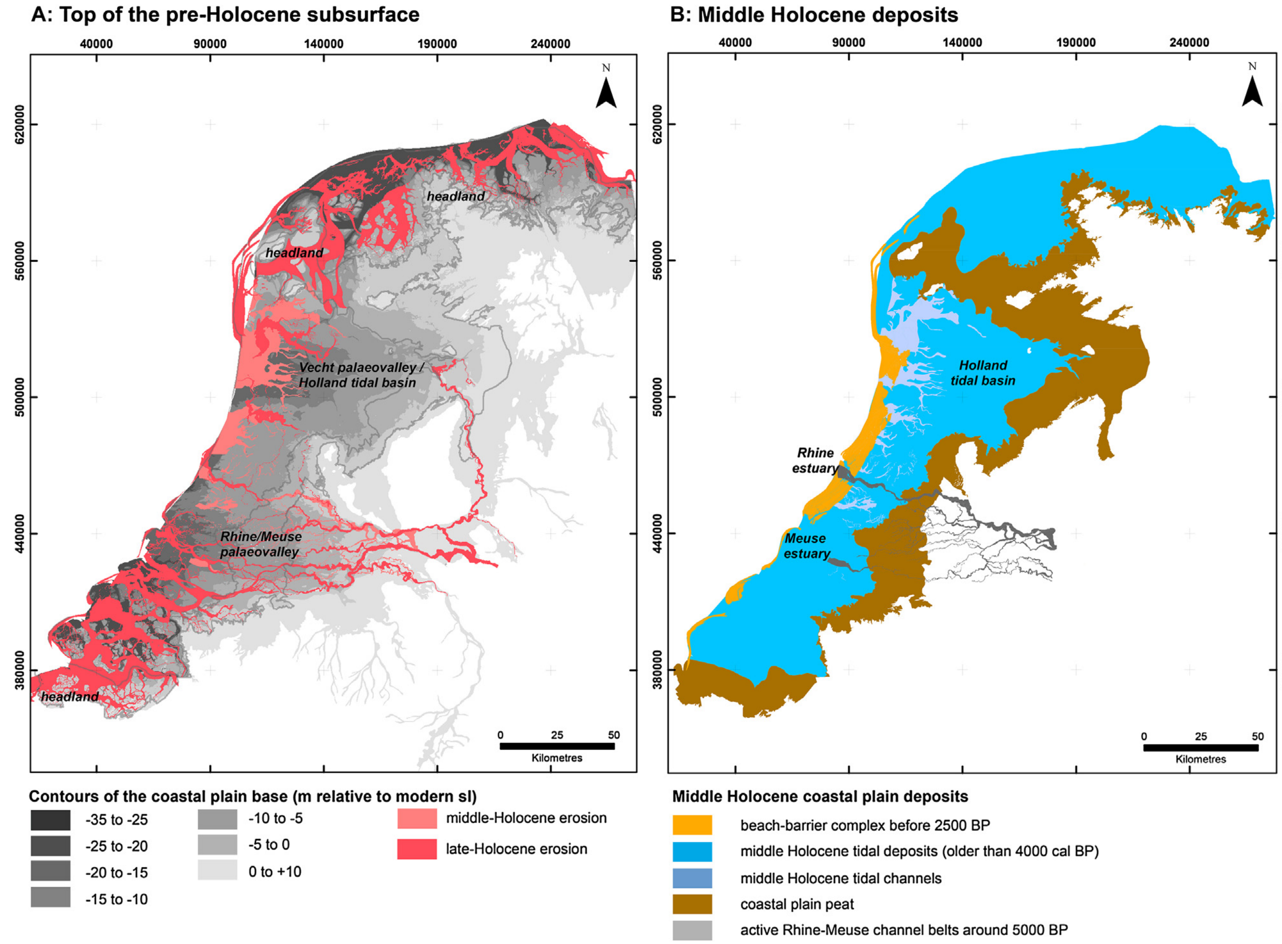

Fig. 1. A: Top of the pre-Holocene subsurface after Vos (2006). The major headlands and the major palaeovalleys mentioned in the text are indicated. The position of late Holocene erosive tidal channels were derived from Pierik et al. (2016), the middle Holocene channels were taken from Cohen et al. (2015). The grey line indicates the extent of the middle Holocene deposits and the coastal peat lands of Fig. 1B.
B: Extent of the middle Holocene deposits and peat area, from Cohen et al. (2015). The Rhine–Meuse channel belts were taken from Cohen et al. (2012).

regimes and facilitating conditions such as decreased sediment availability and human peatland reclamation. Natural preconditions include the geological setting (e.g., coastal plain extent, stratigraphical architecture, sediment delivery) at the time of ingression. These conditions also affect the occurrence and extent of transgression, but the degree to which they facilitated or prevented transgression has hardly been considered.

The coastal plain of The Netherlands is the largest coastal plain of NW Europe and therefore contains a variety of natural preconditions (Figs. 1 and 2 A). Situated in the depocentre of the Southern North Sea, the length of the coastal plain is not bound to inherited valleys, but stretches over hundreds of kilometres along the shore. Also, the width reaches tens of kilometres inland, where – different to many smaller coastal plains – vast peatland areas occur (Figs. 1 B and 2). These peat lands were reclaimed for agricultural use from about 250 BC onwards, causing human-induced subsidence (Borger, 1992; Vos and van Heeringen, 1997; Vos, 2015a). Major sea ingressions (*landward lateral expansion of a new single tidal system*[1]) have been recorded in geological and archaeological studies of these human-occupied coastal plains. The availability of age-specific archaeological artefacts and well-mapped deposits in the coastal plain makes this region very suitable to assess the timing of landscape developments and human activities (e.g., van Liere, 1948; Knol, 1993; Vos and van Heeringen, 1997; Vos and Gerrets, 2005; Vos, 2015a; Pierik et al., 2016).

Peat-surface lowering was triggered by peatland reclamation that involved ditch cutting to drain the topsoil making it suitable for agriculture (Vos, 2015a; Erkens et al., 2016). The start of the reclamations across the coastal plain varies from Late Iron Age to early medieval times (250 BC–AD 1050). The reclamations were wide-spread, and had affected the entire coastal plain by ca. AD 900 (Vos, 2015a). Natural forcings, such as wave and tidal regime show minor variations regionally and are not considered to have varied majorly during the last 4–5 millennia (e.g., van der Molen and de Swart, 2001a,b). Major regional differences exist, however, in the natural preconditions, i.e., geological and geographical setting along the coastal plain. These regional differences also occur in the timing, degree, and impact of sea ingressions. This raises the question how the natural preconditions affected the pacing and final extent of the mainly human-triggered coastal plain changes. Data coverage on the geological situation of the Dutch coastal plain nowadays is quite even (e.g., Vos, 2015a; Pierik et al., 2016). Therefore it provides an ideal area to geographically intercompare the evidence and to assess the mechanisms and controls of late Holocene coastal-plain transgressions.

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[1] A tidal system is a back-barrier tidal environment, connected to open sea trough an inlet channel (containing tidal channels, tidal flats, salt marshes, or lagoons).

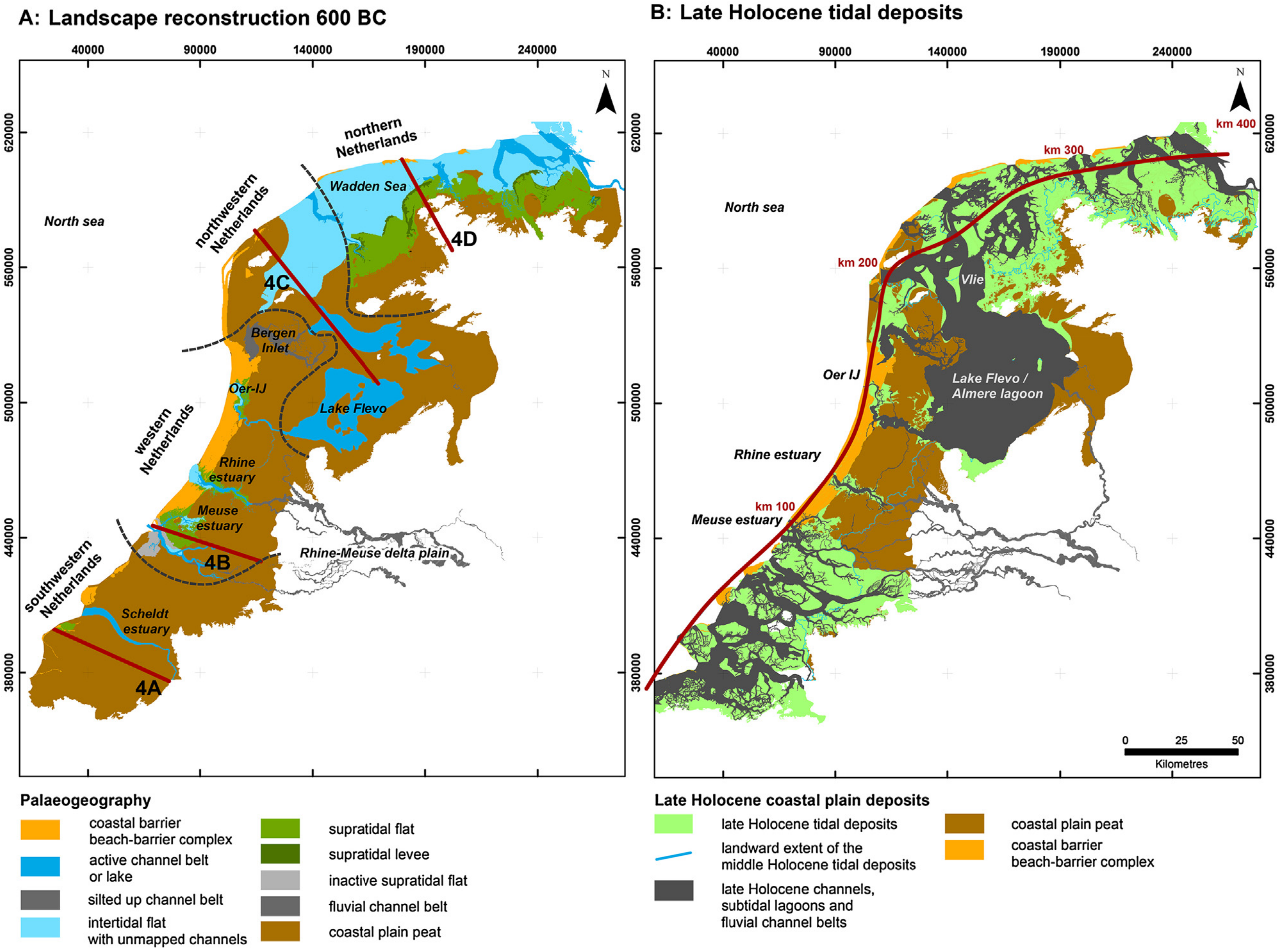

Fig. 2. A: Landscape reconstruction at the onset of late-Holocene transgression (around 600 BC—Pierik et al. (2016)), the according areal percentages can be found in Table 1. The red lines and figure numbers correspond to sections in Fig. 4.
B: Extent of the middle and late Holocene tidal deposits including channels and beach barriers (after Cohen et al., 2015; Pierik et al., 2016). The red line corresponds to the position of the longshore diagram of Fig. 5.

1.1. Aim and approach

This paper aims to (1) reconstruct the natural precondition for late Holocene sea ingressions; (2) reconstruct the late Holocene ingressive development, and (3) identify the mechanisms of ingressive developments in relation to different natural preconditions. The study area reaches from the coastline inland and includes areas with tidal deposits and their flanking peat swamps (Figs. 1 and 2). We focus on the late-Holocene period when peat-surface habitation and coeval ingression development repetitively occurred at multiple locations (Vos, 2015a). We tested the influence of natural preconditions on sea ingression development. Therefore, we compared the divergent ingressive developments between four main segments of Netherlands' coastal plain that faced contrasting late-Holocene developments and had different natural preconditions (i.e., landscape settings around 600 BC before large-scale reclamation—Fig. 2A). The defined segments are:

- The *southwestern* part of The Netherlands (SW-NL). This peat area witnessed major sea ingressions since ca. 2000 cal years BP (e.g. Vos and van Heeringen, 1997);
- The *western part* of The Netherlands (W-NL), with a wide sequence of beach barriers, interrupted by estuaries (Beets et al., 1992). We position its borders in the peat area directly south of the Meuse estuary, and just north of the former Bergen tidal basin. The inland boundary of this segment is positioned at the most inland location of perimarine crevasse splays around the Rhine and Meuse channel belts indicating significant tidal influence, after Cohen et al. (2015);
- The *northwestern* Netherlands (NW-NL), which experienced coastal retrogradation in the west during the late Holocene and at the same time, westward expansion of the Wadden-Sea tidal area in its eastern part. During the late Holocene, the Almere lagoon was connected to the Wadden Sea and it was therefore included in this segment (Zagwijn, 1986);
- The *northern* Netherlands (N-NL) that maintained a barrier-tidal basin coast during the entire Holocene (Vos and Knol, 2015).

To understand the mechanisms controlling the development of tidal systems and sea ingressions, we describe the coastal evolution with its specific local details. In the analysis we used geological information from various national datasets, in which the results of many previous studies have accumulated. The local details unavoidably are selective and incomplete, for further information and additional argumentation we refer to the original publications in the text, for more extensive references see also Vos (2015a) and Pierik et al. (2016). Even in our long-studied research area, the available age control does not allow absolute dating of all stages per individual tidal system. Dates have often been collected at one location per tidal system only, while peat reclamation and tidal channel expansion mainly occurred diachronously. To optimize coastal evolution age control, we used a recently-developed GIS that stores all relevant late-Holocene coastal plain geological mapping, palaeogeography, and local study referencing (Pierik et al., 2016). In the GIS, the extent of coastal plain architectural elements was mapped, a database with dating information was maintained separately. This information was then automatically combined by generating time slice palaeogeographical maps for each desired time slice, which were iteratively improved. This GIS further allowed to reconstruct and quantify the evolution of individual tidal systems and beach barrier segments, as well as intercomparison between the subregions. From the sea ingression cases in the study area, we compiled a conceptual model for their evolution that will be used in the discussion to outline the mechanisms and the influence of natural precondition on sea ingression evolution.

1.2. Geographical setting and boundary conditions

The coastal plain of The Netherlands mainly comprises middle and late Holocene back-barrier deposits, onlapping a modestly sloping pre-Holocene substrate (Jelgersma, 1979; Beets and van der Spek, 2000). The coastline encompasses a chain of beach-barrier complexes and tidal inlet systems (van der Spek and Beets, 1992; Beets and van der Spek, 2000; Vos, 2015a; Fig. 2B). This barrier complex established between 6000 and 5000 BP, after an initial middle-Holocene phase of marked sea-level rise (van de Plassche, 1982; Hijma and Cohen, 2011; Lambeck et al., 2014) and marks the change to coastline stabilization when relative sea-level rise decelerated towards the modest present-day rates. After 3000 cal years BP, the trapping of fine-grained sediment from the Rhine and Meuse in fluvial-deltaic environments (Erkens and Cohen, 2009) and neighbouring peat areas (de Boer and Pons, 1960) started to increase. This clay deposition is understood to result from an increase in sediment supply received from upstream (i.e., external control; Cohen, 2005; Gouw and Erkens, 2007; Hoffmann et al., 2007; Erkens and Cohen, 2009).

Conditions along the coastline are micro to mesotidal (1.5–4 m tidal range), with significant wave energy (cf. Davis and Hayes, 1984; van der Spek and Beets, 1992). The Dutch coast is exposed to predominant westerly winds and wave fields and consequently experiences a dominant north to northeastern oriented net longshore sediment transport (Beets et al., 1992; Beets and van der Spek, 2000). These dominant winds facilitated wave-driven sediment transport from the shallow sea floor to the coast. The tidal range varies from over 3 m in the southwestern and northern Netherlands to 1.5 m in the western Netherlands. Open-sea tidal conditions have remained more or less stable since at least 6000 years BP (van der Molen and de Swart 2001a,b). An increase in wave energy has been reported for the last 6000 years in the western part of The Netherlands attributed to steepening of the shoreface (van Heteren et al., 2011). Gottschalk (1975) documented increased storm activity between AD 1400–1600 in the SW Netherlands based on historical records, whereas Sorrel et al. (2012) reported intervals of increased storminess (relatively high frequency of intense storms) in NW Europe for 1350–450 BC, AD 50–900, and AD 1350–1700 based on sedimentological evidence. Apart from that, hydrodynamical conditions are assumed to have been quite constant over the last millennia (Beets et al., 1992).

Relative sea-level rise (RSLR) over the last 2000 years varied between 0.5 and 1.0 m (van de Plassche, 1982; Roep and Beets, 1988). In these youngest millennia, RSLR was predominantly controlled by tectonic and glacio-isostatic background subsidence (Kiden et al., 2002; Vink et al., 2007; Koster et al., 2016). Small fluctuations of the eustatic sea-level possibly occurred during the late Holocene mainly resulting from steric sea-level movement due to climate fluctuations (van Geel et al., 1996; Behre, 2004, 2007; Gehrels, 2010; van de Plassche et al., 2010; Kopp et al., 2015). Most authors however, do not attribute a major forcing role to this, as they consider the rates of peatland-surface lowering and autogenic tidal and sedimentary changes to be larger (see discussions in Weerts et al., 2005; Bungenstock and Weerts, 2010; Baeteman, 2008; Baeteman et al., 2011; Vis et al., 2015).

2. Middle Holocene coastal plain evolution

This section describes the middle-Holocene evolution of the four coastal segments, determining the preconditions for the late Holocene human-triggered coastal landscape changes. Ages are presented in calibrated years before present (cal year BP), unless otherwise specified.

2.1. Pre-Holocene inherited setting and middle-Holocene transgression

During post-glacial sea-level rise, continental shelves submerged forming vast wetlands where both organic and clastic sedimentation occurred (Shepard, 1932; Jelgersma, 1979; Smith et al., 2011; Bicket and Tizzard, 2015). The Pleistocene and early-Holocene topography that underlies the coastal plain deposits served as substrate for this middle-Holocene transgression (Pons et al., 1963; de Gans and van Gijssel, 1996). This moderately sloping pre-Holocene substrate was intersected by two wide fluvial valleys (~30 km) in the western Netherlands and smaller valleys in other parts of the study area (Fig. 1A). The large valleys provided ample Holocene accommodation space and served as major sediment sinks for coastal and fluvial sediments. These inherited palaeovalleys inundated from about 9500 BP onwards (Hijma and Cohen, 2011; Vos et al., 2015a; Koster et al., 2016). The Rhine–Meuse palaeovalley (Fig. 1A) filled up relatively quickly with estuarine and fluvial sediments (Erkens and Cohen, 2009; Hijma et al., 2009, 2010). The absence of a large river and the subsequent low sediment input caused the Vecht palaeovalley to be transformed into an embayment, where infilling occurred at significantly lower rates (Holland tidal basin, Jelgersma, 1983; van der Spek and Beets 1992; Beets and van der Spek, 2000). Here, subtidal clays were the first preserved clastic marine deposits (Pons and Wiggers, 1959/1960; Pons and Wiggers, 1959; Beets et al., 2003).

Currently offshore positioned remains of tidal inlets indicate the position of an old beach barrier around 7000 BP (Rieu et al., 2005; Hijma et al., 2010); formed when decreasing sea-level rise reduced the creation of accommodation space. The back-barrier basins started to fill in with sandy channel sediments and intertidal tidal flat deposits (Pons and Wiggers, 1959/1960; Pons and Wiggers, 1959; de Mulder and Bosch, 1982; Hijma et al., 2009). In the southwestern and northern Netherlands a similar stratigraphical sequence of initial transgressive subtidal clays to sandy subtidal and intertidal deposits developed (de Jong et al., 1960; Vos and van Kesteren, 2000).

The higher elevated parts of the pre-Holocene substrate were situated in the southern part of the southwestern Netherlands and in the northwestern Netherlands (Fig. 1A; de Gans and van Gijssel, 1996). They were headlands during the middle Holocene and acted as sediment sources for coastal evolution (e.g., Beets et al., 1992, 1994; Cleveringa, 2000). They were inundated relatively late during the middle Holocene creating a relatively thin Holocene coastal wedge (generally less than 5 m thick peat and clastic tidal deposits). At present, 65% of the coastal plain area contains middle Holocene tidal deposits (Fig. 1B). The tidal areas were flanked by extensive peatlands on the Pleistocene substrate (~35% of the coastal plain area). The peats contain intercalated clay layers representing distal tidal system deposits (Jelgersma, 1961; Streif, 1978; Allen, 2000; Bertrand and Baeteman, 2005).

2.2. Middle-Holocene turnover to highstand

The decreasing rate of sea-level rise reduced the creation of back-barrier accommodation space from about 6000 BP onwards. During the late Holocene the relative importance of eustatic sea-level rise decreased and land subsidence became more important (Beets and van der Spek, 2000; Cohen, 2005; Hijma and Cohen, 2011; Koster et al., 2016). The continuing sedimentation in the back-barrier area and especially the expansion of the intertidal flats, caused a decrease in tidal storage volume of the tidal basins, to which the tidal inlets adapted by filling in. In the western and southwestern Netherlands this eventually led to their closure and the formation of an elongated uninterrupted beach barrier complex (van Straaten, 1965; Beets et al., 1992; van der Spek et al., 2007; van Heteren et al., 2011). In the back-barrier area of these coastal segments a fining-upward trend is found in the top of the middle Holocene tidal deposits, recording a transition from intertidal to supratidal conditions and a decrease of tidal energy (Pons and Wiggers, 1959/1960; Pons and Wiggers, 1959,Westerhoff et al., 1987; Hijma et al., 2009). In the northern Netherlands, the tidal area never fully silted up, tidal inlets persisted and the chain of barrier islands did not amalgamate into an uninterrupted beach-barrier complex (Beets and van der Spek, 2000).

2.3. Late Holocene mature highstand

The closure of the Holland beach-barrier complex ended tidal dynamics and sedimentation in the largest part of the back-barrier area (Beets et al., 1994), thereby facilitating large-scale peat formation on top of the silted-up tidal systems, especially in the western, southwestern, and northwestern coastal plain segments (Pons et al., 1963; Pons, 1992). At the onset of large-scale peatland reclamation (e.g., between 600 and 250 BC, see Table 1), ~60% of the coastal plain area consisted of peatland, draining to a few river outlets and remaining tidal inlets. At a distance from large rivers and tidal channels, peat bogs had developed that were modestly elevated (2–4 m) above high tide and storm-surge sea levels (Bennema et al., 1952; Pons, 1992; Vos et al., 2015b; Erkens et al., 2016). These bogs acted as local watersheds that are regarded to have stabilised the positions of peat-drainage channels that linked up to the tidal creeks and the sea. In the northern Netherlands the barriers of the Wadden Sea remained interrupted by tidal inlets, whereas along the inland side of the tidal basins supratidal levees developed after 700 BC that facilitated peat areal expansion (Vos and Gerrets, 2005).

The large Holland tidal basin was not completely filled in by the end of the middle Holocene (Bergen Inlet system, Fig. 2A, Pons et al., 1963; van der Spek and Beets, 1992; de Gans and van Gijssel, 1996; van Zijverden, 2017). In the inland distal central part of this basin (over ~40 km away from the coastline), an extensive lake remained ('lake Flevo'), surrounded by peat fens and swamps. Waves eroded and reworked the peat edges of this lagoon into lake detritus that mixed with sand from the locally-eroded outcropping Pleistocene substrate (Pons and Wiggers, 1959/1960; Pons and Wiggers, 1959, van Loon and Wiggers, 1975). After the Bergen Inlet silted up (around 1500–1100 BC, de Mulder and Bosch, 1982; Beets et al., 1996; van Zijverden, 2017), lake drainage occurred via the Oer IJ inlet (Vos et al., 2015b). When drainage shifted northwards to the Wadden Sea by the formation of the Vlie, the Oer IJ inlet silted up as well (around the last century BC), (Fig. 2; Vos et al., 2015b). This marked the beginning of lake Flevos' transformation into the Almere lagoon in late Roman and earliest medieval times. In the western Netherlands, two river outlets were present in the Rhine–Meuse delta (Fig. 2A). The major branch of the Rhine occupied a northernmost route since 6000 BP (Berendsen and Stouthamer, 2000). Because of its long maintained position, a mature fluvial channel belt was present that had steadily supplied sand to the barrier coast, contributing to the beach barrier progradation and the formation of a subaquatic delta. Behind the barriers, dendritic secondary tidal creeks successively formed and silted up along the Rhine estuary, both at its very mouth and at distance inland (Pruissers and de Gans, 1985; Berendsen, 1982; van Dinter, 2013). The smaller river Meuse debouched into a southerly estuary near Rotterdam throughout the Holocene (Hijma et al., 2009), supplying a relatively small amount of sediment to the coast. Here too, secondary tidal creeks branched off the main Meuse estuarine channel.

3. Late-Holocene ingressions

After the relatively stable period characterised by large-scale peat formation, large-scale sea ingressions into the peat area occurred from the Roman period onwards (12 BC–AD 450—Vos, 2015a). Since then, the area containing tidal deposits expanded from 40% to 75% of the coastal plain (Table 1). In this section, we describe the late-Holocene evolution per coastal segment, focussing on the timing of initiation, maturation, and silting up the sea ingressions. This location-specific information, as well as the general trends, are illustrated by Figs. 3–7 .

3.1. Southwestern Netherlands

In the southwestern Netherlands, beach barriers dating to the beginning of the late Holocene have hardly been preserved, due to

Table 1
Areal extent (in 1000 ha) of peat and tidal deposits per segment of coastal plain during the middle Holocene, 600–250 BC and AD 2000. Numbers are given as integers.

	Total coastal plain[a]	Middle Holocene (MH)[b]		600–250 BC		AD 2000
		Tidal deposits	Peat	Tidal deposits	Peat[c]	Tidal deposits
SW	435	282	153	30	405	420
W	551	443	108	134	417	157
Lake Flevo/Almere lagoon[d]	396	140	256			
NW	252	149	103	285	363	558
N	500	382	117	372	128	477
Total	2133	1396	737	822	1312	1612

[a] Derived from the sum of middle Holocene (MH) peat and MH tidal deposits.

[b] Values represent the total area of MH tidal deposits (i.e., not at a specific time step) and are derived from the GIS data shown in Figs. 1 B and 2 A.

[c] Calculated by subtracting tidal deposits area from total coastal plain area.

[d] The NW Netherlands and lake Flevo were separated areas during the middle Holocene and therefore split up in the table. During the late Holocene the Flevo Lake/Almere area connected to the NW coastal segment, and is therefore considered as part of this segment.

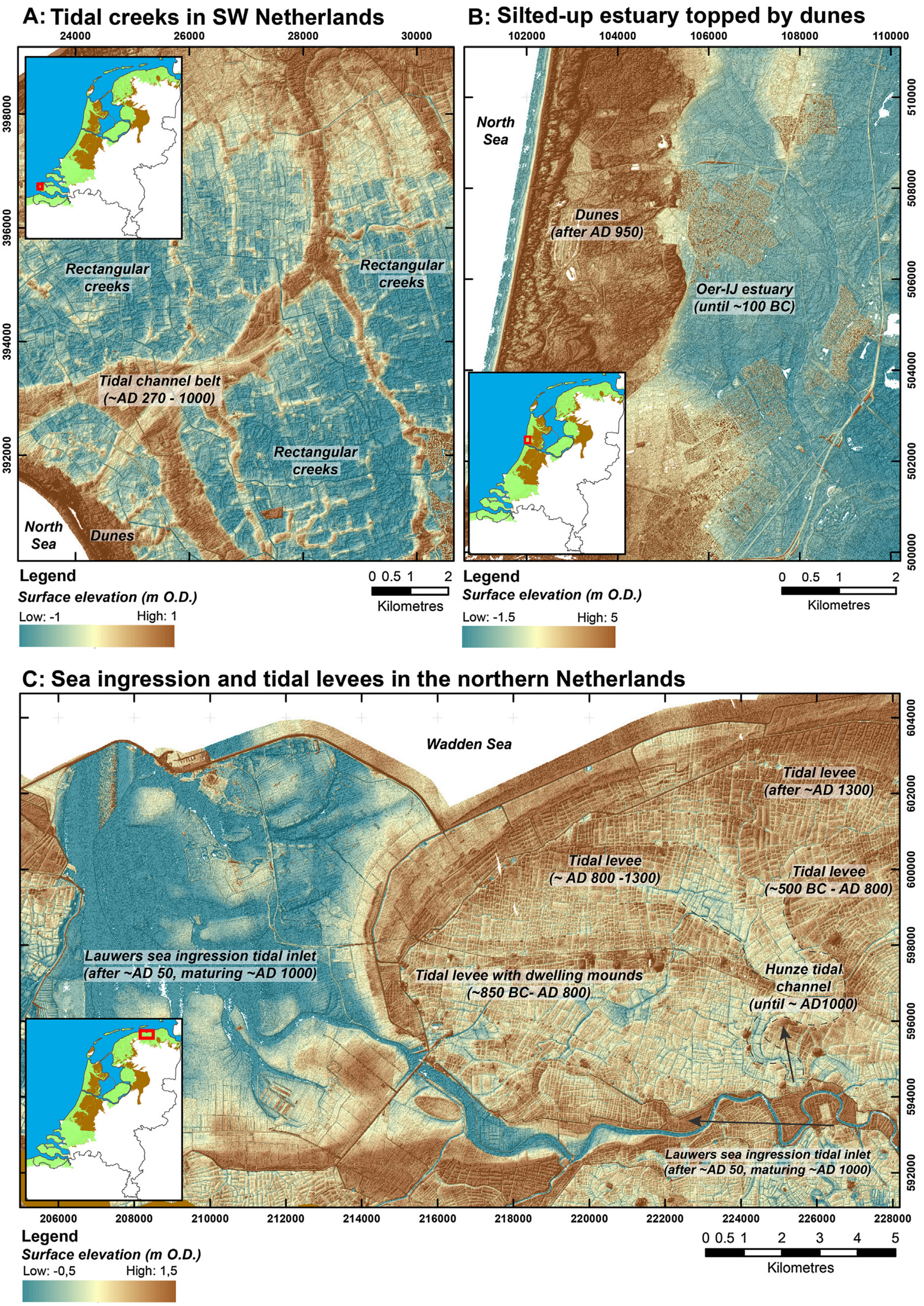

Fig. 3. LIDAR images of three parts of the coastal plain. The small location maps are after Fig. 2B: green: late Holocene tidal deposits, brown: coastal plain peat lands.
A: Rectangular tidal inversion ridges in the former reclaimed peatlands in the SW Netherlands, representing Roman ditches later filled in with sand. Ages are after Vos and van Heeringen (1997).
B: Silted-up estuary western Netherlands covered by parabolic dunes. Ages after Vos et al. (2015b).
C: Multiple generations of supratidal levees protected the coastal plain (ages after Vos and Knol, 2015; Pierik et al., 2016). They were interrupted by small tidal inlets such as the Hunze, which silted up after the Lauwers sea ingression formed during the Early Middle Ages (AD 450–1050).

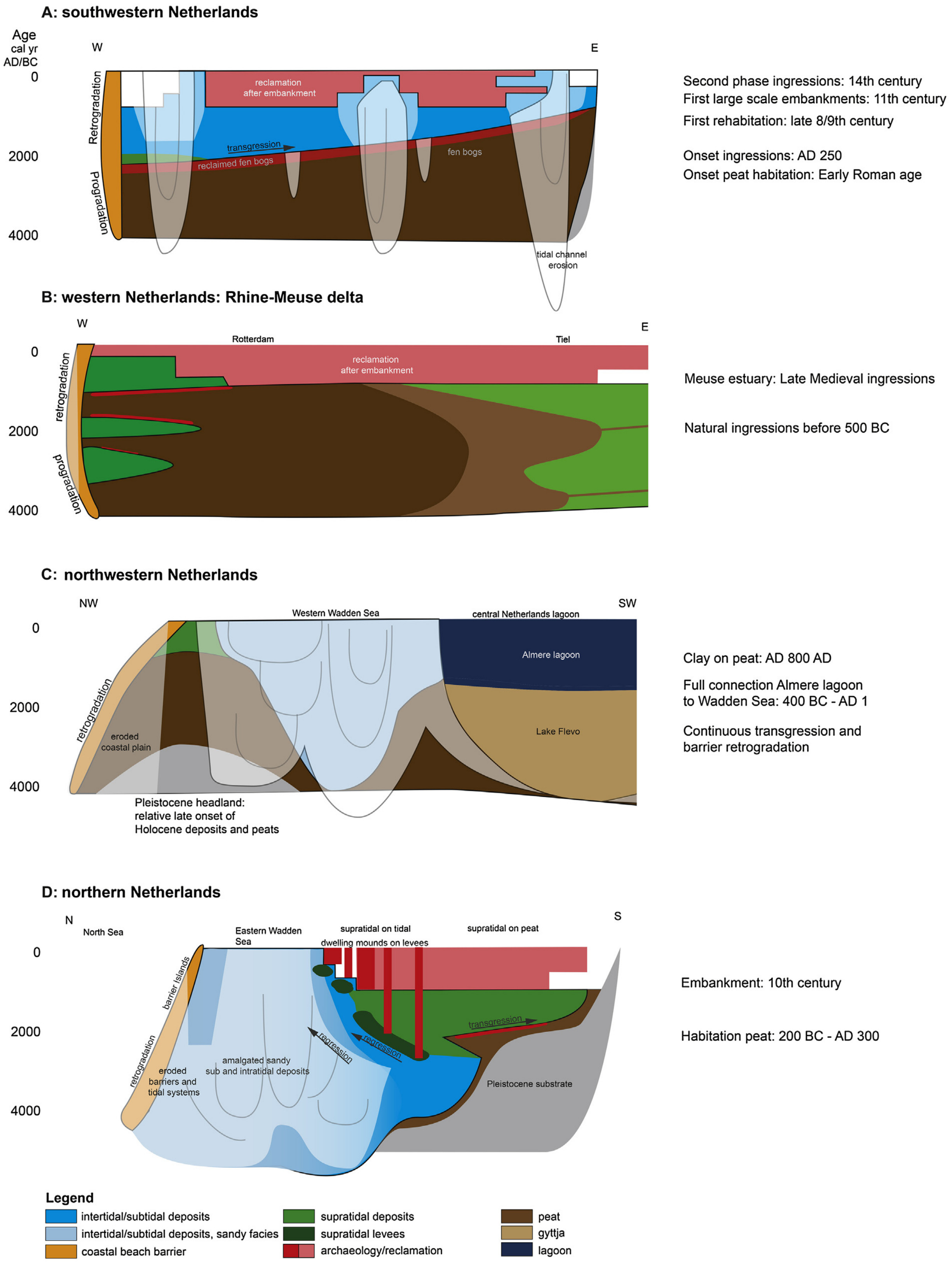

Fig. 4. Time-space cross-sections of coastal segments. Four conceptual cross-shore space-time diagrams illustrating the regional differences of the four coastal plain segments (location Fig. 2A).
A: After cross-section of Vos and van Heeringen (1997: Appendix I, profile B). Reclamation and habitation of the vast peat lands caused flooding and expansion of the tidal area. After several centuries of tidal sedimentation gradual embankment of the area took place.
B: After Hijma and Cohen (2011). Beach barriers projected from The Hague, after Cleveringa (2000); van der Valk (1996a). Intercalated clay layers from secondary tidal systems formed around the Meuse estuary after van Staalduinen (1979) and Vos and Eijskoot (2015), and were inhabited since 600 BC (van Liere, 1950; van Londen, 2006). In the Late Middle Ages large-scale embankment and peat reclamation took place (Borger, 1992; de Bont, 2008) whereas in the areas south of the Meuse sea ingressions occurred.

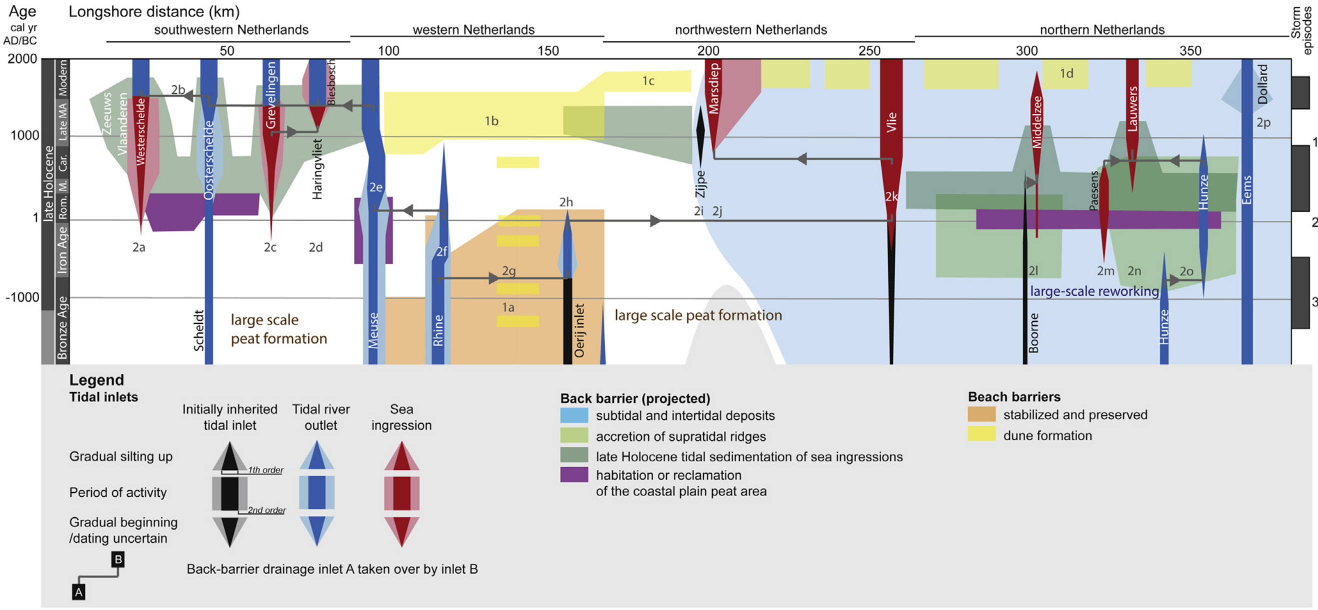

Fig. 5. Longshore time-space section of coastal development. The diagram shows the evolution of tidal inlets in time and space related to presence of rivers and coastal plain habitation. The cross-section is situated along the coastline (Fig. 2B), tidal inlet locations are indicated in Fig. 7. In the northern Netherlands, the line is projected along the intertidal–supratidal transition to visualize back-barrier development. Episodes of increased storm frequencies are taken from Sorrel et al. (2012). We subdivide the tidal inlet systems into three genetic classes after Vos and Knol (2015): (1) inherited tidal inlets formed during early-middle Holocene sea-level rise (black); (2) estuaries (blue); (3) natural or human induced late Holocene sea ingressions. Bright colours indicate the 1st order inlets, which are directly connected to the sea, paler colours indicate the presence of their secondary tributaries. Abbreviations of the archaeological periods on the y-axis: Rom. = Roman period, M. = Merovingian period, Car. = Carolinian period. Numbers refer to the following sources.
Archaeology: Southwestern Netherlands—(Vos and van Heeringen, 1997); Meuse estuary—(van Londen, 2006); northern Netherlands—(Miedema, 1983; Knol, 1993; Gerrets, 2010). ***Beach barriers and dunes***—1a: Beach barriers western Netherlands—(van Straaten, 1965; Beets et al., 1992; van der Valk, 1996a,b; Cleveringa, 2000; Vos et al., 2015b); 1b: Younger Dunes—(Jelgersma et al., 1970; Zagwijn, 1984; Vos et al., 2015b); 1c: Schoorl (1999); 1d: Jelgersma and Ente (1977). ***Tidal inlets***—2a: Onset youngest transgression southwestern Netherlands—(Bennema and van der Meer, 1952; Vos and van Heeringen, 1997); 2b: Avulsion Eastern Scheldt to western Scheldt—(van der Spek, 1997; Vos and van Heeringen, 1997); 2c: Initiation Grevelingen—(Vos, 2015b); 2d: Initiation Haringvliet—(Vos, 2015b); 2e: Extension tidal basin around Meuse estuary—(van Staalduinen, 1979; van Trierum, 1986; Vos and Eijskoot, 2015); 2f: Rhine estuary dynamics—(van Dinter, 2013); 2g: Avulsion Utrechtse Vecht (Rhine tributary) into the Oer IJ (Bos et al., 2009); 2h: Closing Oer-IJ inlet: (Vos et al., 2015b); 2i: Activity Zijpe—(Schoorl, 1999; Vos, 2015b); 2j: initiation Marsdiep—(Ente et al., 1986; Schoorl, 1999; Vos, 2015a); 2k: Vlie—(Ente et al., 1986; Schoorl, 1999; Vos, 2015a); 2l: Boorne and Middelzee—(Cnossen, 1958; van der Spek, 1995; Vos and Gerrets, 2005; Vos and Knol, 2015); 2m: Paesens—(Griede, 1978; Vos and Knol, 2015); 2n: Lauwerszee—(Roeleveld, 1974; Griede, 1978; Vos and Knol, 2015); 2o: Hunze (Roeleveld, 1974; Vos and Knol, 2015); 2p: Dollard—(Homeier, 1977; Behre, 1999; Vos and Knol, 2015). ***Supratidal ridges***—(Roeleveld, 1974; Vos and Gerrets, 2005). Late Holocene transgression—southwestern Netherlands: (Vos and van Heeringen, 1997; Vos, 2015b); northwestern Netherlands—(Westerhoff et al., 1987; Vos, 2015b); northern Netherlands—(Gerrets, 2010; Vos and Knol, 2015).

extensive post-Roman erosion. Therefore, the exact width of the beach barrier complex at the time is unknown, but is assumed to be a few kilometres at maximum (Vos and van Heeringen, 1997). In the back-barrier area, an extensive peatland was situated. In the top of the peat sequence mainly fen peat has been found (indicating formerly raised bogs), which contains abundant evidence of Roman reclamations and habitation. The earliest evidence of peat reclamation in this area is placed during the Late Iron age (Bennema and van der Meer, 1952; Vos and van Heeringen, 1997).

Around 500 BC, the first small ingressions took place into this peat area along the mouth of the Scheldt estuary (Figs. 4 A, 5, 6 A, 7 A). More large scale deposition of clastic material on the strongly subsided peat began approximately at ~AD 250 and reached further inland at AD 500 (Figs. 4 A, 5, 6 A, 7 C; Vos and van Heeringen, 1997; Vos, 2015b). The wide-spread clay deposits positioned directly on the peat indicate relatively low energy environments during the first stages of inundation (Pons, 1965). The rectangular pattern of many tidal tributaries (preserved as tidal channel inversion ridges—Fig. 3A) indicate the usage of a Late Iron Age or Roman ditch-network that offered a preferential pathway for transgression (Vlam, 1942; Bennema and van der Meer, 1952; Vos and van Heeringen, 1997). With the drowning of the back-barrier area large tidal inlets developed (compare Tables 1 and 2, Figs. 4 A and 6 A) eventually reaching depths up to several 10 s of metres (de Jong et al., 1960). From the moment of drowning onwards, sandy material was deposited in the intertidal areas and channels. This material mainly originated from reworking inherited middle Holocene tidal channel belts as well as from a strong retrogradation of the coast between Walcheren and the adjacent Flemish coast (Ebbing and Laban, 1996; Denys, 2007; Mathys, 2009; Vos, 2015b), leading to scouring of the relatively shallow Pleistocene substrate in this area. The new large tidal channels caused sand to be transported efficiently into the tidal basins. In the 8th century, infilling of the tidal area was advanced to a supratidal level facilitating the first settlements on the silted-up tidal channel belts (Vos and van Heeringen, 1997). Since the 12th century the supratidal area was embanked in successive phases, starting with the then highest elevated

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C: Northwestern Netherlands. The distal lagoonal area was a remnant of the middle Holocene Vecht Embayment and became only part of the northwestern Netherlands coastal segment after ~400 BC when the Vlie connected lake Flevo to the Wadden Sea (Figs. 5 and 7).
D: After space-depth profile (van der Spek, 1996; Fig. 5). Here, beach barriers never closed, amalgamated channel deposits occur in the intertidal area. The expansion of the tidal area in the landward part of this coastal plain occurred diachronously after the Roman period, whereas more seaward supratidal levees accreted (Vos and Gerrets, 2005; Vos and Knol, 2015).

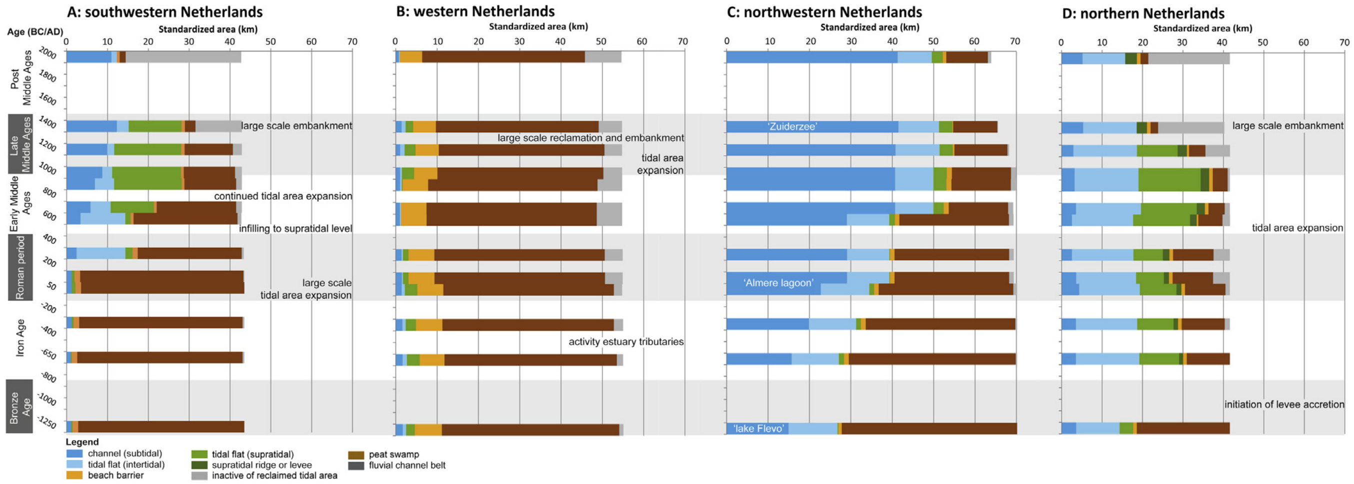

Fig. 6. Areal statistics of sedimentary domains. Regional comparison of the timing and impact of tidal area expansion since 1250 BC. The units were derived from Fig. 7, and their colours correspond to Fig. 7. The areal extent was standardized per km coastal length to correct for the different sizes of the coastal plains by dividing the summed area of the landscape units by longshore coastal section length. Transgression in the northwestern Netherlands mainly comprises the expansion of the Almere lagoon.

supratidal areas at a distance from major channels (Figs. 4 A and 6). The late medieval tidal deposits became more enriched in sand indicating an increase in depositional energy (Pons, 1965), corresponding to the expansion of the tidal area and the increase in tidal inlet size. After the intertidal shoals and smaller channels had silted up to supratidal level they successively could be embanked as well. In the Late Middle Ages (AD 1050–1500), the Scheldt river avulsed to the Western Scheldt as a consequence of

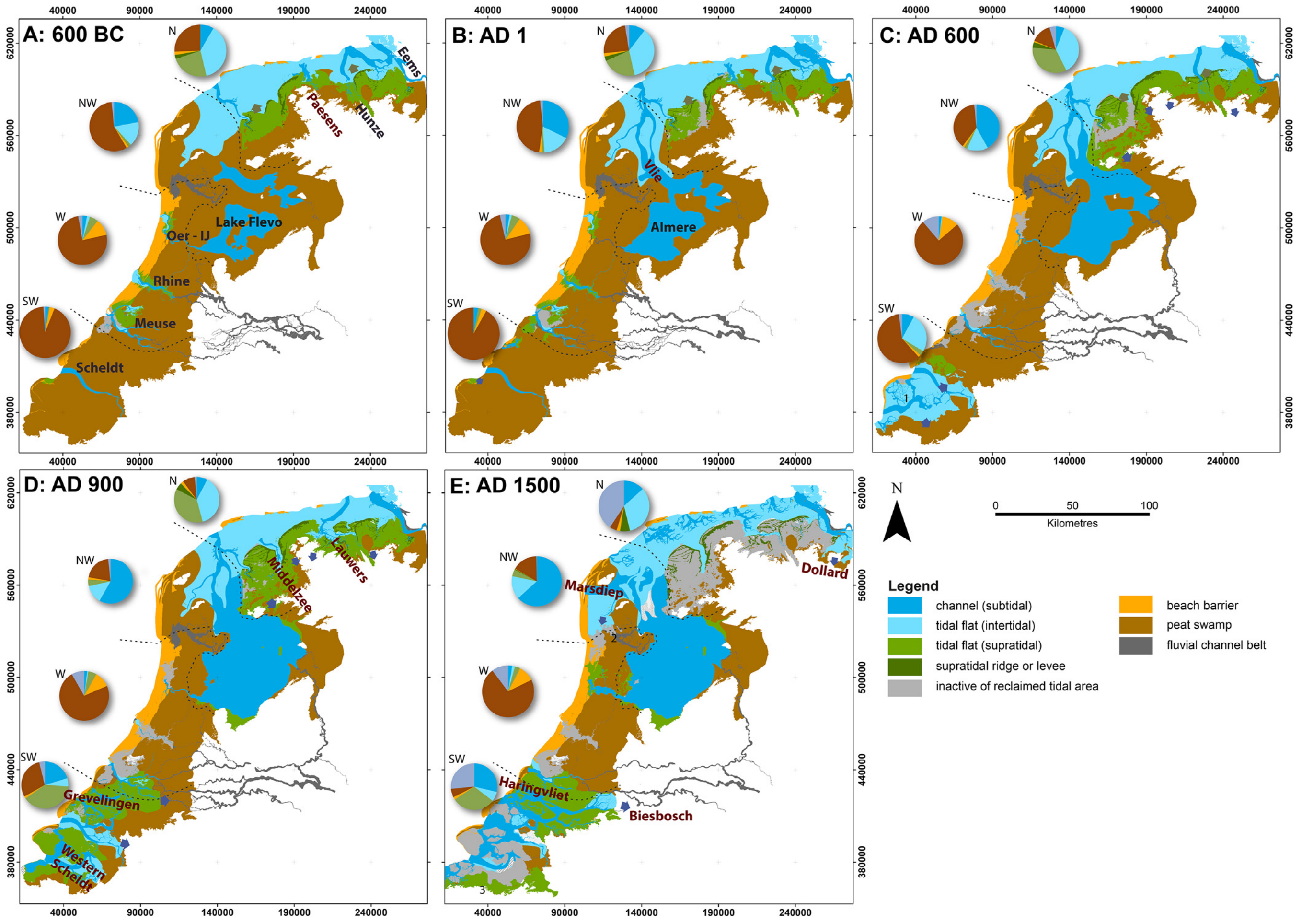

Fig. 7. GIS generated palaeogeography. GIS-derived coastal development since 600 BC (Pierik et al., 2016): blue arrows: tidal areal expansion; green arrows: salt marsh ridges expansion; black names: inherited middle Holocene estuaries and lakes; brown names: newly formed sea ingressions. Pie charts demonstrate relative areal extent of the landscape units per coastal segment, colours correspond to the map legend units. Footnotes: 1—Walcheren, 2—northern part Noord Holland, 3—Zeeuws-Vlaanderen.

Table 2
Increase in tidal channel and total tidal flooded area (channels, intertidal and supratidal flats) in average $1{,}0 \times 10^4\,m^2$ per year, derived from the palaeogeographical reconstructions of Fig. 7.

	600–300 BC	300 BC–AD 50	AD 50–300	AD 300–600	AD 600–900	AD 900–1200	AD 1200–1500
SW NL tidal area increase	13	18	466	0[a]	416[a]	0[b]	40[b]
SW NL channel increase	7	0	38	28	117	102[b]	78[b]
N NL tidal area increase	0	0	0	230	104	–[b]	6[b]
N NL channel increase	0	−20	0	0	27	–[b]	–[b]

[a] The pacing of tidal areal increase in the southwestern Netherlands after AD 300 is uncertain.
[b] land loss compensated by embankments, only new losses are shown.

continuous backward ingression of the latter (van der Spek, 1997; Vos, 2015b). While in Late Middle Ages, new areas were embanked, new land losses also took place. These late medieval ingressions especially affected the more inland part of the coastal plain in the subsided peat area of the northern part of the southwestern Netherlands, as well as the southernmost edge of this segment. The confinement and planform smoothing of the tidal area by dikes caused high water levels to increase, posing more flooding risks to adjacent embanked areas (van der Spek, 1997). Additionally, the dikes were generally not well maintained when the disasters took place (Zonneveld, 1960; Verbraeck and Bisschops, 1971; Kleinhans et al., 2010; Missiaen et al., 2016 Fig. 7E).

3.2. Western Netherlands (Holland)

Unlike the other segments, no large sea ingression occurred in the western Netherlands during the late Holocene, mainly because of the presence of a wide beach-barrier complex. This barrier complex was dissected by three estuaries (Meuse, Rhine, Oer IJ; Fig. 7A), from which secondary tributaries branched out into the flanking peat area. In the Meuse estuary, the levees and supratidal areas of the active tributaries were inhabited since the Iron Age (van Liere, 1950; van Londen, 2006). After Roman times most of these secondary tidal systems silted up and were overgrown by peat (Fig. 4B; van Trierum, 1986; Beets et al., 1994; Dijkstra, 2011; Vos and Eijskoot, 2015). Around 2000 BP the Oer IJ and Rhine estuaries silted up and beach-barrier progradation stopped along this coastal segment. Waves reworked the seaward bending beach ridges near the former Rhine estuary, leading to local coastline retrogradation (Fig. 4C; van Straaten, 1965; Cleveringa, 2000). Coastal retrogradation has been attributed to a decrease in sediment supply from the shallow offshore area caused by depletion of the offshore sand volume as a result of continued wave transport towards the coastal barriers during the millennia before (Roep, 1984; Beets et al., 1994; van Heteren et al., 2011). Additionally, it coincided with avulsion of the Rhine towards the Meuse estuary after the Roman period (Berendsen and Stouthamer, 2000; Stouthamer and Berendsen 2001), resulting in a loss of direct fluvial sand supply to the beach barrier complex around the former Rhine estuary. Afterwards, the Meuse estuary received more discharge and sediment load (e.g., Erkens and Cohen, 2009; Hijma et al., 2009). It is quite possible that these sediments contributed to the silting up of the tidal areas of the late-medieval sea ingressions in the northern part of southwestern Netherlands. Iron Age and Roman habitation was mainly confined to the barriers, the tidal levees, and possibly also to the edges of the peatlands. From the Middle Ages onwards, large scale peat reclamation took place (Borger, 1992; Vos et al., 2015b).

3.3. Northwestern Netherlands

The present coastal barrier in this segment is composed of relatively narrow mainly post-medieval beach ridges. Behind the barrier system, back-barrier peat is covered by clastic deposits of medieval age (Zagwijn, 1986; Vos, 2015a), whereas tidal deposits from the Iron Age and Roman period have not been found. Due to continuous erosion, the timing of developments in this area is less known. This erosion is documented by historical sources that describe a medieval and post-medieval retrograding trend of the coastline and the expansion of tidal inlets (Schoorl, 1999). Coastal plain erosion and tidal deposition in this area not only occurred directly as a result of ingressions from the North Sea. On the eastern side, expansion of the Wadden Sea tidal flats in the peat area took place (Eisma and Wolff, 1980). Similar to the other coastal sections, this medieval tidal area expansion (probably since AD 800) has been registered in the northern part of Noord-Holland as a clay cover on peat and is associated with peat reclamation (Fig. 5; Vos et al., 2015c). In a final phase of tidal area expansion, during the Late Middle Ages, the Marsdiep inlet was formed (Figs. 5 and 7), which since then drained a large part of western Wadden Sea (Ente et al., 1986; Schoorl, 1999). Meanwhile, waves continuously eroded the peaty shores, causing the lagoonal area to expand (Figs. 4 C and 7). The remaining peatland surrounding the lagoon became covered with clay during the Middle Ages (Veenenbos, 1950; Pons and Wiggers, 1959/1960; Pons and Wiggers, 1959, Westerhoff et al., 1987; van den Biggelaar et al., 2014).

3.4. Northern Netherlands

In the northern Netherlands, barrier islands alternated with tidal inlets that each drained large intertidal areas of the Wadden Sea (Figs. 2 and 7). The occurrence of back-barrier tidal-channel deposits underneath the barrier islands (Oost, 1995; van der Spek, 1996; Vos and van Kesteren, 2000), and remnants of tidal channels in the current offshore realm (Beets et al., 1994) demonstrate that the barrier-island coastline retrograded during the last millennia. This retrogradation provided sand for the back-barrier intertidal area. Along the most inland part of the intertidal area, supratidal levees accreted between 700 BC and AD 500 (Roeleveld, 1974; Griede, 1978; Vos and Gerrets, 2005; Figs. 3 C, 4 D, 5 and 6). Their clayey texture made them relatively resistant to lateral erosion by tidal channels and they probably also hampered the drainage of the surrounding subsiding peatland (Knol, 1993). Between ~AD 300 and AD 600 the inland peatlands became regionally covered by a transgressive clay layer (Tables 1 and 2, Figs. 6 and 7C; Veenenbos, 1949; Veenenbos and Schuylenborgh, 1951; Roeleveld, 1974; Knol, 1993). This clay deposition is attributed to peat-surface subsidence mainly resulting from reclamation activities and peat mining during the Late Iron Age (250–12 BC) and the Roman period (Fig. 4D; Griede, 1978; de Groot et al., 1987; Gerrets, 2010; de Langen et al., 2013). The deposits are associated with newly-formed ingression tidal inlets such as the Lauwers and Middelzee. Their size and depth expanded, caused by the increased tidal volume initiated by subsidence of the peat area (Knol, 1993; van der Spek, 1995; Vos and Knol, 2015). In the tidal sediment sequences, this expansion is observed as a gradual upward increase in sand content (Pons, 1965). The new ingressions reached their maximum extent approximately AD 1000, after which they were embanked in different phases (van der Spek, 1995). It is presumed that they initiated in the Roman period (Vos and Knol, 2015) taking over the drainage of local small channels from the previous generation of tidal inlets (Fig. 5). Compared to regions in the southwestern Netherlands, the late Holocene transgressed area in the northern Netherlands was small and therefore the number of new ingressive systems was smaller as well. New tidal depositional areas connected to these inlets were also smaller than those in other coastal sections and furthermore better protected from the sea by the supratidal levees (Fig. 5–7). Similar to the southwestern Netherlands, a second generation of extensive land losses in the embanked former tidal areas occurred. The most dramatic example is the land loss owing to the 15th–16th century ingression along the river Ems known as the 'Dollard' (Homeier, 1977; Behre, 1999; Vos and Knol, 2015).

4. Discussion

4.1. Evolution of sea ingression tidal systems: mechanisms, timing, and pacing

From the series of examples encountered in The Netherlands, it is observed that quite a large coastal plain area was flooded instantaneously by major storms, but that the tidal channels took a few centuries to reach their equilibrium dimensions before the tidal system finally silted up (Figs. 4, 5, 7). To better understand sea ingression evolution we summarised the facilitating controls, triggering mechanisms, and feedbacks, derived from the cases in our study area, into a conceptual model showing the evolution of a typical late Holocene tidal system (Fig. 8). In our conceptual model we distinguish the following key phases:

4.1.1. Phase 0—pre-triggering phase

Before sea ingression took place, natural or anthropogenic developments caused subsidence in the coastal plain subside, making it susceptible to sea ingression (phase 0 in Fig. 8). At least for the southwestern and northern Netherlands we consider human-induced subsidence of the peatland to be the prerequisite for sea ingressions in our study area, based on the timing of initial tidal sediment deposition after peatland occupation and the unprecedented scale of the sea ingressions (Pons, 1992; Vos and van Heeringen, 1997; Ervynck et al., 1999; Vos, 2015a). This subsidence ranged from several decimetres to up to several metres and resulted from oxidation and compaction caused by groundwater table lowering (Erkens et al., 2016). In a later stage, subsidence was further aided by peat digging for salt mining (e.g., Griede, 1978; Jongepier et al., 2011). The large-scale surface lowering was not immediately compensated for by sedimentation and therefore caused irreversible drowning and transgression.

Natural factors have also been postulated for sea ingression initiation. An increased discharge of rivers would have caused widening or deepening of estuaries, facilitating sea ingressions (e.g., Pons, 1992; Baeteman, 2005). In our study area, this would

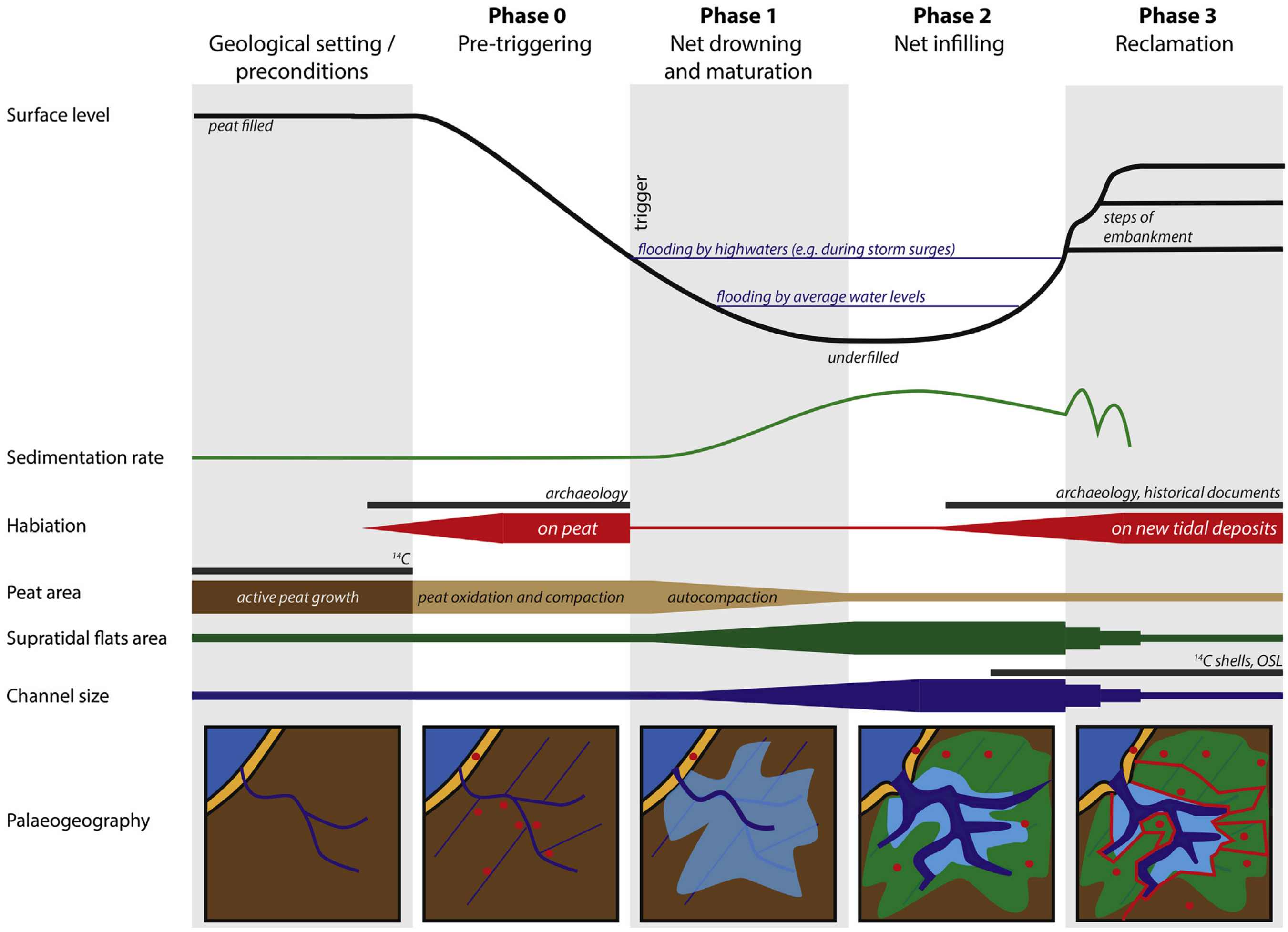

Fig. 8. Infographic on the development of a sea ingression, for new ingressions in reclaimed peat land. For further explanation see text.

apply to the Meuse estuary that received more river discharge due to the avulsion of Rhine channel belts towards the Meuse estuary. Here, however, the opposite is observed: increased river discharge coincided with large scale peat formation around the estuary rather than sea ingression (Section 3.2). Moreover, most sea ingressions in our study area occurred in absence of these major rivers or even at new positions along the coast (Fig. 5) making it unlikely that this played a major role for the sea ingressions considered in this study. Depletion of offshore sediment sources where coastal barriers were narrow made the coastal plain more prone to back-barrier tidal area expansion. This mechanism has been discussed for the Flemish coast—Baeteman (2005), UK Fenlands—Brew et al. (2000), UK Romney Marsh—Long et al. (2006) as well as for the Holland coast (Section 3.2). This could also have played a role in the northwestern Netherlands, although the ingressions seem to follow periods of reclamation as well (Vos, 2015a; Section 3.3). The extent and timing of the reclamation and the sea ingression in this segment remains to be established more accurately to assess the role of this facilitating natural precondition (Vos, 2015a).

4.1.2. Phase 1—net drowning and maturation

Ingression triggering: The subsiding peat area provided storage capacity, initially for storm water, later for diurnal high tide water, and in some cases even for permanent water. This large-scale flooding took place after the peatlands became connected to tidal or fluvial systems, most likely triggered by storms or spring tides (start of phase 1). Several studies attribute the increased influence of marine conditions to an increase in regional storm intensity (e.g., Regnauld et al., 1996—Britanny, France, Long et al., 1998—Humber estuary, UK; Clavé et al., 2001—Gironde estuary, France). As demonstrated in Fig. 5, documented NW European episodes of increased storminess (after Sorrel et al., 2012) coincide with two phases of sea ingressions in our study area (1900–1050 BP and 600–250 BP), whereas a third episode of enhanced storminess (3300–2400 BP) hardly had an effect on coastal plain development. This suggests that the facilitating conditions of large scale peat subsidence in enlarging back-barrier storage capacity after 2000 BP was an essential prerequisite for the sea ingressions. The tidal storage generated by this facilitating condition was necessary to cause water flow through existing weak spots along the coastline (e.g., small channels, lower spots in barriers or levees) to develop into new large tidal inlets at the observed scale (Fig. 5). Single storms or episodes of enhanced storminess may have accelerated sea ingression or triggered the beginning of sea ingression, but they are unlikely to be the main cause behind the sea ingressions.

Sea ingression maturation: The creation of accommodation space in the back-barrier area provided an increase in tidal volume to which the size of the sea ingression tidal channels adapted proportionally during phase 1 and 2. For empirical relations considering channel dimensions in relation to tidal discharge, we refer to O'Brien (1931, 1969), Jarrett (1976), and van der Spek (1995). In our reconstructions, we observe that the areal extent of the drowning part of the coastal plain is proportional to the size of the tidal channel belts (Tables 2 and 3, Fig. 7). During this stage, an erodible sandy substrate allowed fast adaptation to the new back-barrier tidal area, whereas in more resistant clay and peat channel adaptation took longer. The gradual maturation is reflected in the sedimentary record by the upward increasing sand content of clays in the southwestern and northern Netherlands indicating an increase in energy.

Initial infilling and autocompaction feedback: The growing accommodation space has been filled with sediments supplied through the expanding tidal channels. The weight of clastic deposits caused compaction of the underlying peat by loading resulting in additional subsidence. This positive feedback created a further increase of tidal volume and accommodation space. Its effect is proportional to the thickness of peat deposits at the time of inundation (e.g., Allen, 1999; van Asselen et al., 2011). The thickness of the peat therefore controls the amount of accommodation space resulting from sediment loading and hence the timing of the shift from net drowning to the net infilling of the tidal system (phase 1 to 2).

Flanking area feedbacks: Another positive feedback includes the collapse of flanking unreclaimed peat areas caused by lowering of the groundwater table in the adjacent area after ingression. The areal extent of this feedback and its pacing is unknown, but it may have played a role in further expanding ingressions. Furthermore, in response to the expanding ingressions, peat-land habitation can shift towards more inland positions, inducing new local land subsidence, which further enlarges the ingression prone area. Developing a detailed chronology of drowning in the northern part of the southwestern Netherlands and in the northwestern Netherlands may help further understanding of this mechanism.

4.1.3. Phase 2—net infilling

In the study area, tidal basins had a tendency to fill in as a result of tidal asymmetry and additional feedbacks (e.g., scour lag, settling lag—van Straaten and Kuenen, 1958; van den Berg et al., 1996). The timing of the shift from phase 1 to 2 is determined by sediment availability, but also by accommodation space and its feedbacks as described above (auto-compaction and flanking area feedbacks), which delay the net infilling. We expect sediment

Table 3
Relative contribution of natural preconditions on flooding, the formation of sea ingressions and infilling of the coastal plain.

Natural preconditions	Enhance LH flooding or sea ingression initiation	Enhance far inland LH transgression?	Enhance LH infilling?	Remarks/feedbacks	Example
Low sloping Pleistocene substrate (result: wide peat filled plain)	0	++	0	Shallow peat, but far inland	NW-NL, N-NL
Thick peat	No initial effect	No initial effect	–	Subsidence by loading ++	SW-NL, W-NL
Small MH beach barriers	+	0	0	Enhance MH peat growth, and thus LH flooding potential	SW-NL, NW-NL
Wide MH beach barriers	−	−	+	Reworked beach barrier material for sea ingressions infilling	W-NL
Supratidal levees	−	−	0		N-NL
MH presence of large rivers	−	0	0	Clastics flanking estuaries	W-NL (Meuse, Rhine)
LH presence of rivers	?	?	+	Estuary remains open, sediment source	W-NL (Meuse)

MH = middle Holocene, LH = late Holocene, + positive effect, − negative effect, 0 no effect.

import to be proportional to channel size, i.e., the sediment will be most efficiently distributed over the tidal system when the size of the channels is adapted to the tidal volume of the tidal system.

4.1.4. Phase 3—reclamation by embankment

After sediment accretion had caused the tidal area to be sufficiently elevated, the area could be inhabited, reclaimed or successively embanked. Despite the increased supply of suspended sediment, not all late Holocene created accommodation space has been filled in and embanked today (e.g., Westerschelde, Oosterschelde, Dollard, Eastern Wadden Sea).

4.2. Natural preconditions controlling the pacing and extent of sea ingressions

Several natural preconditions control the development of sea ingressions and the resulting new tidal areas. We use the phases of Fig. 8 to outline the relative contribution of natural preconditions, which are also summarised in Table 3.

Coastal-plain width: A large coastal peat area results in a relatively large potential for the creation of tidal storage and accommodation space during phase 0, which will eventually develop in a large transgressed area during phase 1 (Table 3). This explains the large extent of the transgressed area in the southwestern part (30 km per km coastal length—Fig. 6) compared to the northern Netherlands (10 km per km coastal length) Fig. 5. When the areal extent of drowned peat is large, more sediment is required to push the tidal area from phase 1 to phase 2 (Fig. 8). Furthermore, within the considered coastal segments, the slope of the pre-Holocene substrate of the most inland part of the coastal plain controlled the inland extent of the late Holocene tidal deposits during phase 1 (Fig. 9). When late-Holocene gradual RSLR and peat collapse took place, the presence of a flat and shallow Pleistocene surface (i.e., former headlands), topped by a thin middle Holocene peat layer facilitated the sea ingressions to penetrate deeper inland compared to the middle Holocene tidal areas (Figs. 2 B and 9). This is observed in the northwestern Netherlands and on a smaller scale in the southern part of the southwestern Netherlands (in the areas indicated as *headland* in Fig. 1A) and in the peat area of the northern Netherlands. The underfilled Flevo lakes in the coastal plain in the northwestern part have never entirely been filled up by peat, making them sensitive to a different natural transgression mechanism: peat erosion by wave activity. This facilitated large scale late-Holocene expansion of the lagoon (Almere) and the growth of the proportionally large Vlie channel (Fig. 7).

Protecting elements: Three types of protecting geomorphological features have counteracted coastal ingression of human-reclaimed land: beach barriers, supratidal levees, and inherited tidal tributaries. In the aftermath of storm surges, such elements delayed or prevented the formation of a permanent tidal channel connecting flooded subsidence-prone land to the sea (phase 0 towards 1 in Fig. 8). They also delayed lateral expansion of the newly forming tidal channels (during phase 1 and 2). In the western Netherlands, the wide beach-barrier complex topped by coastal dunes prevented large-scale flooding and the formation of new tidal inlets. Peat-filled back-barrier areas in the southwestern and northwestern Netherlands had less wide beach barriers and were more sensitive to flooding and subsequent formation of sea ingressions. Other protecting elements were the inherited secondary tidal systems along the long-active estuaries (Meuse, Old Rhine, Oer IJ; Fig. 5). In the millennia before reclamation, these systems had formed a more elevated levee-like topography around the estuaries inhibiting sea ingressions (Pons, 1992). The long presence of tides in these rivers caused floodwater to be relatively high on a regular basis, probably allowing the natural levees to build up relatively high (several decimetres). In addition to this effect, this already naturally auto compacted ribbon zone was less prone to subsidence compared to peatlands and therefore less sensitive to ingressive erosion. These elements probably became especially important from AD 1000 onwards, when the flanking peatlands were reclaimed at large scale around the Rhine and Meuse estuary. In the northern Netherlands supratidal clayey levees formed a line of protection. This probably delayed channel formation, causing the tidal channels to mature relatively late compared to the southwestern Netherlands (Fig. 5). The position and elevation of protecting elements did not only influence the pacing of sea ingressions, but also the location of the late Holocene sea ingression tidal inlets. They mainly formed at new positions relative to their precursors (e.g., Lauwers, Middelzee, Westerschelde Fig. 5), i.e., at weak locations in the protecting elements, possibly taking advantage of already-existing small creeks.

In the Late Middle Ages (AD 1050–1500), the first Roman to early medieval generations of sea ingressions that formed relatively close to the sea had silted up, to a level that made them suitable for human activities. They also formed protecting elements for the remaining inland peatlands. This could not prevent a second late medieval generation of ingressions taking place further inland in already embanked areas (Biesbosch, Dollard, Zeeuws Vlaanderen; Figs. 5 and 7E). In contrast to the earlier ingressions, the position of the dikes in the landscape controlled the ingression extent and modulated the tidal amplitude (Section 3.1), whereas their state of maintenance controlled the sensitivity for failure.

Sediment delivery to the back-barrier area was determined by overall sediment supply and by the proximity of the inundated area to main paths of tidal flow and wave activity. During the natural preconditional phase, sediment supply facilitated the presence of protecting elements as described above. When the ingression took place, it co-controlled the transition from net drowning to net infilling (transition from phase 1 to 2) and finally reclamation (phase 3). After the creation of additional accommodation space by autocompaction during phase 1 the balance shifted towards net infilling when sediment supply is abundant. In our study area, sources of sand mainly included reworked material from pre-Holocene headlands and the seafloor (Beets and van der Spek, 2000), consisting of overstepped back-barrier deposits (Hijma et al., 2010). Suspended material was mainly imported through the tidal inlets from the sea, where it was originally supplied from river mouths. As a result of deforestation in the upstream catchments, the amount of suspended load trapped in the Rhine–Meuse delta gradually increased since 500 BC, up to 3 Mton/year by AD 1000, which is twice the amount compared to the middle Holocene (Erkens and Cohen, 2009). We presume that the amounts of clays and silts transported to the sea and consequently available for sedimentation in the coastal plain must have increased as well, most likely causing relatively faster silting up of the new tidal systems. Around AD 1400 the avulsed Rhine system directly supplied sediments to the southwestern Netherlands (Biesbosch; Kleinhans et al., 2010), accelerating the rate at which the young tidal system could fill in (8–10 mm/a proximal to the river) and lost peatland could be reclaimed as fresh supra-tidal surface (Fig. 4). Finally, the development of many new tidal channels favoured efficient sediment distribution over large parts of the coastal plain during phase 2 and 3 (in the southwestern Netherlands). In contrast, coastal parallel-transgressed areas (clay on peat areas in the northern Netherlands) drained by a comparatively smaller number of channels, hampered sediment distribution (especially sands) over the distal part of the coastal plain.

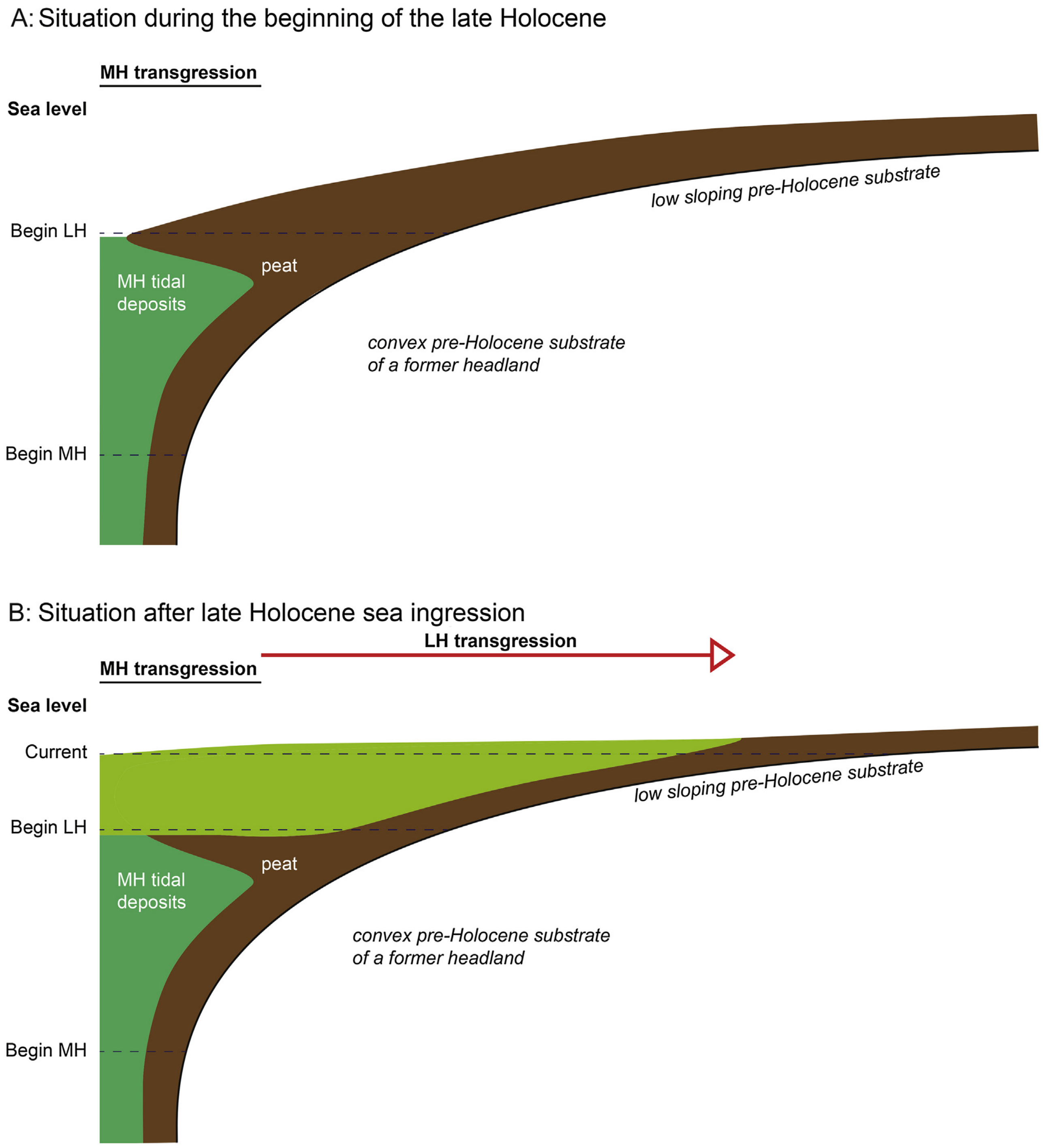

Fig. 9. Infographic on the development of a sea ingression on a convex sloping pre-Holocene substrate. This situation represents pre-Holocene headlands indicated in Fig. 1A where the sea ingression can reach relatively far during the late Holocene compared to the middle Holocene tidal systems. This is a result of the low sloping shallow part of the pre-Holocene substrate.

4.3. Outlook

Within our study area, several topics deserve additional research. A first topic is the non-linear response of flanking unreclaimed peat collapse feedback after initial ingression. Continued dating effort in geoarchaeological contexts will show how far the entire peat area is affected by possibly rather local human-induced subsidence. Other unresolved issues are the role of a potential storm-frequency and intensity increase in triggering sea ingressions, and the complex interplay between tidal, storm and river-discharge conditions controlling sea ingressions around larger rivers. For this, further refining of storm-event records around the North Sea for the entire late Holocene, flood records from rivers, and refining the timing of river branches and tidal inlet maturing will provide new insights. The mechanisms derived from the palaeogeographical reconstructions may be tested using numerical models.

The mechanisms and suitable preconditions for transgression reconstructed for the Dutch coastal plain can be used to study drowning coastal plains in other areas in the world. In these areas boundary conditions may be different or the substrate could consist of another subsidence-prone lithology (e.g., unconsolidated clays). Successful maturation of ingressions could be related to either lack of sediment supply (e.g., Stanley and Warne, 1993 and Zecchin et al., 2009) or creation of additional accommodation space (Törnqvist et al., 2008; this study). Despite these possible differences, these drowning coastal plains have in common that a rapid human-induced transgression results in sea ingressions with comparable feedbacks. This pushes the system into a long state of drowning and land loss until sedimentation compensates for this. Considering the role of natural preconditions in these mechanisms as in our approach can help to assess the differential impact of coastal plain subsidence and drowning.

5. Conclusions

Large-scale coastal plain subsidence causes a major impact, both on the landscape and on habitation. We demonstrate this impact for The Netherlands' coastal plain, where mainly human-induced sea ingressions took place since the late Roman period continuing in the Middle Ages, on a remarkably large scale. Unlike the middle-Holocene transgression, these late-Holocene ingressions occurred in a peat filled back-barrier area and were not forced by rapid sea-level rise. In our study we identified the role natural preconditions (i.e., geological setting) in the coastal plain changes caused by human-induced subsidence area for different coastal sections with varying natural preconditions. The contrasting developments in coastal plain evolution in the different coastal sections are used to derive generic mechanisms of the initiation and maturing stages of sea ingressions:

- Wide coastal plains filled with subsidence-prone peat are most sensitive to surface lowering which facilitated ingression after reclamation. The formation of such wide coastal plains is facilitated by a low sloping pre-Holocene substrate. Cutting of ditches during reclamation lowers groundwater tables, which causes peat oxidation and compaction. The resulting surface lowering rate outpaces by far the rate of regional eustatic sea-level rise and is therefore considered a much more important factor for sea ingression. Late Holocene tidal-system ingression can reach tens of kilometres far inland when all controlling factors add up.
- Coastal plain subsidence yields an increase of tidal storage in the back-barrier area. This can cause weak points in the coastline (e.g., creeks, lower parts of barriers or levees) to grow into new tidal channels, that grow proportionally to the back-barrier drainage area. Additional feedbacks such as subsidence by sediment loading lead to extra tidal volume increase and accommodation space for tidal sediments. These combined effects result in irreversible sea ingression over large areas that become unsuitable for habitation.
- In some segments of the coast, certain protecting elements (barriers, supratidal levees) in the landscape control the location of sea ingressions that can develop in their weak spots. Moreover, protecting elements can delay or even prevent channel formation between the sea and the subsiding back-barrier area. Peat areas around estuaries are also less susceptible because they are protected by a zone of erosion resistant clayey deposits from estuary tributaries that coevally formed during peat accumulation.
- Many coastal segments only partially fill in after sea ingression. Filling in the newly formed tidal area to a level that is suitable for habitation takes several centuries. The time required for this development is controlled by the extent of the ingressed area, sediment supply, and the strength of the peat area degradation or subsidence feedbacks. Large drowned areas with thick peat sequences, at distal positions to sediment supply take several centuries longer to fill in.
- When the oldest generation of ingressions is filled in with tidal deposits, these areas are less prone to new ingressions and become suitable for habitation again. Further inland however, vulnerable peatlands can still face younger generations of sea ingressions. In contrast to the earlier ingressions their extent is limited by the location of dikes rather than by natural feedback mechanisms.

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