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Invited review

Holocene evolution of tidal systems in The Netherlands: Effects of rivers, coastal boundary conditions, eco-engineering species, inherited relief and human interference

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ABSTRACT

Estuaries and tidal embayments are partly enclosed coastal bodies of water with a free connection to the open sea at their tidal inlet and with minimal (tidal embayments) or substantial fluvial input (estuaries). Their tidal inlets can only remain open over multiple centuries to millennia when (1) the formation of accommodation space exceeds infilling or (2) the inlet system is in dynamic equilibrium (i.e., sediment input equals output). Numerical modeling studies often suggest that estuaries and tidal embayments can develop toward a dynamic equilibrium under constant boundary conditions and consequently remain open over centuries to millennia, whereas in the Holocene sedimentary record many estuaries and tidal embayments are observed to have filled up and closed off. This raises the questions whether and how tidal inlets can remain open over long timescales (centuries to millennia), and what the effects are of river inflow and sediment supply. Here we compare the long-term evolution of contrasting tidal systems along the Dutch coastal plain to empirically identify the most important factors that control their long-term evolution. We study tidal systems along the Dutch coast because of (1) high data density, (2) abundant well-preserved and well-described estuaries and tidal embayments with contrasting boundary conditions and morphodynamic evolution and (3) their low-sloping setting with soft boundaries. This makes contrasting estuarine dimensions and development largely dependent on initial conditions, boundary conditions and internal biogeomorphological processes.

In the Middle Holocene, Dutch estuaries and tidal embayments were mainly formed by rapid relative sea-level rise. In the late Holocene, they were predominantly the result of natural and human-induced subsidence in coastal plain peatlands. Tidal inlets connected to rivers (estuaries) persisted and attained dynamic large-scale equilibrium while tidal embayments without or with a marginal fluvial inflow were unstable and closed off under abundant sediment supply. Estuaries probably attained a quasi-stable configuration wherein sediment input equaled export due to river-enhanced ebb flow, until fluvial influx was cut off by upstream avulsion causing transition to an embayment and system closure. Long-term net import of sediment from the sea into Dutch tidal embayments is favored by strong, flood-dominated, tidal asymmetry along the Dutch coast, the shallow sand-rich floor of the North Sea, erosion of inherited coastal promontories, and the abundance of mud in the coastal area supplied by the Rhine and Meuse rivers. While sandy tidal embayments without fluvial feeders and with fixed boundaries may obtain dynamic equilibrium and remain open over long timescales, we hypothesize that an abundance of mud and eco-engineering species often culminates in continuous embayment filling with fine sediment and the growth of intertidal and supratidal areas, eventually resulting in closure of the embayment.

1. Introduction

Estuaries and tidal embayments (or lagoons) are ubiquitous in coastal areas worldwide. Here, we define them as semi-enclosed coastal bodies of water with a free connection to the open sea at their tidal inlet, and with freshwater input ranging from no to marginal in what we here call tidal embayments, to substantial in what we here call estuaries. These systems typically include multiple of the following morphological elements; ebb-tidal delta, flood-tidal delta, shoals, sand bars, channels, tidal flats, salt marshes to freshwater marshes and

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natural levees. Estuaries and tidal embayments form on inherited topography and substrate under rising relative sea level, fluvial and coastal sediment inputs by tidal currents, waves, density-driven circulation, and biogeomorphological processes (e.g., Pritchard, 1967; Boyd et al., 1992; Dalrymple et al., 1992; Perillo, 1995; Gingras et al., 1999; Van Ledden et al., 2004; De Swart and Zimmerman, 2009; Wang et al., 2012; Seminara et al., 2012; Coco et al., 2013; Bouma et al., 2014).

Estuaries and tidal embayments are important for many reasons and have many functions. They exhibit a high species density and diversity and harbor highly productive natural habitats (e.g., Howard and Frey, 1975; Gingras et al., 1999; Beck et al., 2001). At the same time they are of pivotal societal importance for agriculture, fishing, shipping, ports, and urban building (e.g., Savenije, 2005). Sustainable exploitation of estuaries and tidal embayments is, however, continuously threatened by accelerating sea-level rise, changing river inflow and increasing human interference with sedimentary processes and ecology (e.g., Craft et al., 2008; Syvitski et al., 2009; Bouma et al., 2014; Yang et al., 2015). How these threats will influence the future evolution of estuaries and tidal embayments around the world is largely unknown. Reconstructing the impact of past climatic changes on estuaries and tidal embayments may reveal how these systems will respond to future environmental changes. Moreover, in areas where humans have interfered with the natural landscape for centuries to millennia such reconstructions may also help to unravel the relative effects of natural versus human-induced forcings on the long-term evolution of tidal systems.

Most present-day estuaries and tidal embayments formed early in the middle Holocene (8500–7500 yr BP) when sea level rose rapidly, continental shelfs were flooded and shifting coast lines were approaching their current positions (Stanley et al., 1994; Smith et al., 2011; Hijma et al., 2012). Pre-Holocene topography determined the initial planform of many estuaries in that low-lying glacial and river valleys were inundated first (e.g., Dalrymple et al., 1992; Hori and Saito, 2007; Rossi et al., 2011; Tanabe et al., 2015; Wetzel et al., 2017). The subsequent evolution of estuaries and tidal embayments is mainly determined by the balance between the formation and infilling of accommodation space (e.g., Nichols, 1989). Here, we study how different paths of development arise depending on this balance. Estuaries and tidal embayments are very efficient sediment traps, with sediment potentially being delivered from rivers and the sea (e.g., Boyd et al., 1992; Metcalfe et al., 2000; Atwater et al., 2001; Dalrymple and Choi, 2007). Over their lifetime, many of the forcings affecting estuaries and tidal embayments change and many estuaries have therefore been constantly adjusting to changing boundary conditions over time. Some estuaries and tidal embayments that formed during the early to middle Holocene transgression still persist today, whereas others have completely filled in (e.g., Dalrymple et al., 1992; Martinius and Van den Berg, 2011). This raises the question if, and under which conditions, estuaries and tidal embayments can be in large-scale equilibrium with balanced sediment input and output and consequently remain open over centuries to millennia, and what determines the timescales of adaptation to changing boundary conditions.

Current understanding and predictive capabilities of long-term and large-scale estuary and tidal embayment development have critical gaps regarding long-term effects of mud accretion, eco-engineering species and externally imposed geometry of the drowned valley or sedimentary coastal plain, size of rivers feeding freshwater and sediments, sea-level change and human interference. Past ecological work, numerical morphological modeling and geological reconstructions remain poorly linked and have not yet resulted in a comprehensive conceptual model from which to proceed with long-term numerical modeling. Physics-based models convincingly reproduced hypsometry and channel-shoal patterns in sandy estuaries and tidal embayments from tidal range, river inflow, and sand input and output conditions (Lanzoni and Seminara, 2002; Hibma et al., 2004; Townend, 2012; Van der Wegen and Roelvink, 2012; Savenije, 2015; Braat et al., 2017), indicating that the fundamental processes of water and sand motion largely control the morphology of channels and bars. However, these models must assume a planform shape with fixed perimeters and a given characteristic length of exponential widening (Savenije, 2015; Dronkers, 2017) lest the banks continue to erode (Van der Wegen et al., 2008). Recent modeling that includes fluvial mud supply shows that mud can stabilize the otherwise erosive estuarine banks (Braat et al., 2017) in a similar manner as river floodplains with mud and vegetation stabilize river channels (Kleinhans, 2010). Yet, assuming a fixed planform shape ignores dynamically evolving boundary conditions and possibly large effects of biogeomorphological interactions, particularly at the system margins (e.g., Temmerman et al., 2007; Kirwan and Megonigal, 2013). Eco-engineering species can significantly change their environment at the landscape scale (Jones et al., 1994) as is well-known for rivers (e.g., Kleinhans, 2010; van Asselen et al., 2017), but at present large-scale effects on tidal system development remain poorly understood. Mapping shows that tidal system dimensions and development in many cases largely depend on pre-Holocene surface (e.g., Boyd et al., 1992, 2006) while many other cases initiated largely independently from inherited relief, and as a result of storm-surge ingressions or river floods (van der Spek, 1995; van de Plassche et al., 2006; Vos, 2015; Pierik et al., 2017). Fossil biota, often abundant in the sedimentary record (e.g., Gingras et al., 1999; Vos and Van Kesteren, 2000; Dashtgard, 2011; Gingras and MacEachern, 2012), imply significant eco-engineering effects that often remain unquantified or unmentioned in reconstructions. Reconstructions of tidal-system evolution compiled from geological data are hardly informed and constrained by physics-based modeling and may therefore be biased toward particular conceptual models or present-day analog systems. Clearly, there is a wide gap between the detail of processes and sedimentary products qualitatively postulated in palaeogeographical reconstructions, in which it is difficult to specifically identify biophysical interactions, and the idealized numerical models applied to millennial timescales that usually ignore biogeomorphological processes and have overly simplistic initial and boundary conditions.

The objective of this paper is to identify the main processes that determine the evolution of estuaries and tidal embayments on timescales of centuries to millennia. In particular, we study effects of river discharge, sediment supply by riverine and coastal processes, and eco-engineering species in combination with mud trapping, on the long-term evolution of estuaries and tidal embayments. We then address the question whether and how estuaries and tidal embayments can develop into a state of long-term equilibrium. To this end, we combine results of numerous palaeogeographical reconstructions from literature, which have been based on a vast amount of data on Holocene deposits of estuaries and tidal embayments in The Netherlands. The presence of multiple (> 20) tidal systems with different initial settings and boundary conditions gives us some degree of control, allowing isolation of important processes (a few examples of studied tidal systems are given in Figs. 1 and 2).

Below we first develop the main research questions of the paper on the basis of literature. Then we summarize current understanding of the Holocene evolution of the Dutch coast and synthesize the long-term evolution of estuaries and tidal embayments and the effects of multiple boundary conditions thereon. Based on this framework of data, reconstruction and generalization we then infer trends and causes, and formulate generic hypotheses and open questions.

1.1. Problem definition

1.1.1. Long-term dynamic equilibrium in estuaries and tidal embayments?

An estuary or a tidal embayment can be considered to be in dynamic equilibrium if the overall planform shape and size and hypsometry of an estuary or tidal embayment undergo little or no change due to a balance between the input and output of marine and riverine sediment under the prevailing boundary conditions (e.g., Pethick, 1994; Dyer, 1997; Pye and Blott, 2014). Such a dynamic equilibrium may be maintained

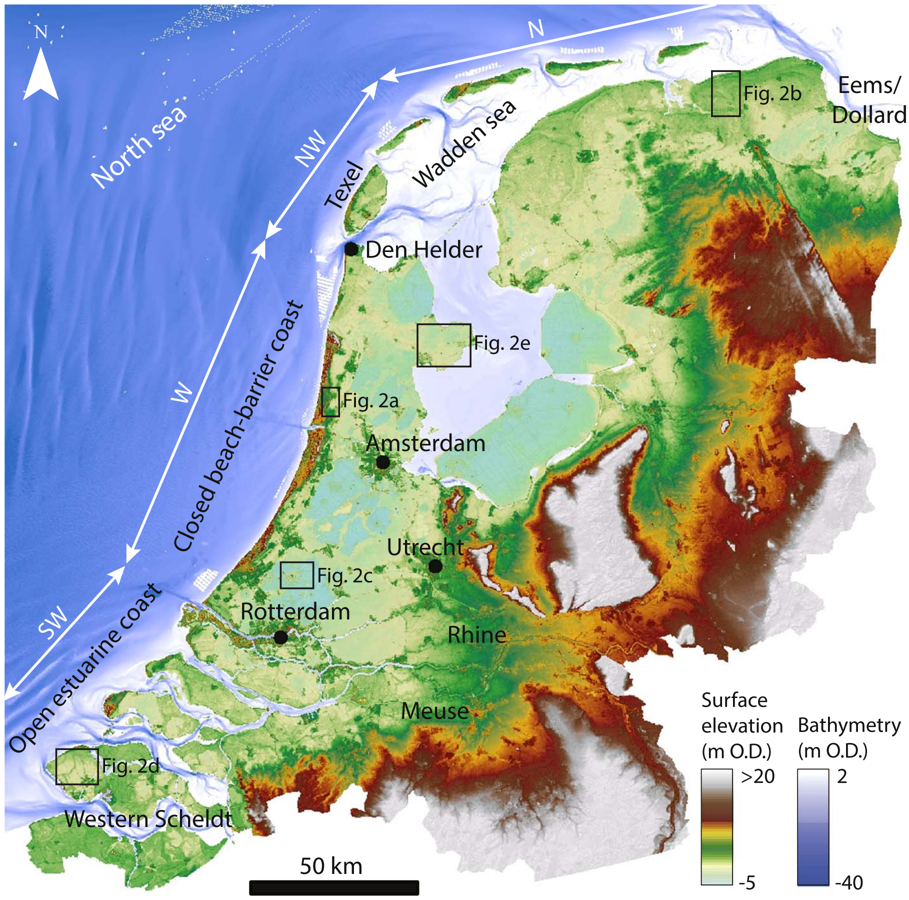


Fig. 1. Present-day elevation map of The Netherlands. SW, W, NW and N stand for the southwestern, western, northwestern and northern parts of The Netherlands.
Source: AHN (Actueel Hoogtebestand Nederland: digital elevation data (Rijkswaterstaat-AGI, 2005))

when an estuary or tidal embayment alternates between flood-dominant and ebb-dominant conditions resulting in no net sediment import or export when averaged over a given timescale (Pethick, 1994; Dyer, 1997). Here, dynamic means that bars and channels may form, migrate, amalgamate, disappear and so on, as long as the basin as a whole has balanced sediment import and export. Numerical models (e.g., Lanzoni and Seminara, 2002; Todeschini et al., 2008; De Swart and Zimmerman, 2009; Hsu et al., 2013), analytical models (e.g., Seminara et al., 2010; Toffolon and Lanzoni, 2010) and the few laboratory experiments that exist today (e.g., Tambroni et al., 2005; Kleinhans et al., 2012, 2015) suggest that both estuaries and tidal embayments can indeed obtain a stable long-term morphodynamic equilibrium configuration and remain open over time. However, with the exception of the laboratory experiments the perimeters of these systems were all fixed, and the mud and biology that affect natural tidal systems were absent.

How is such a dynamic equilibrium maintained on long timescales according to these models? Relatively deep tidal embayments and estuaries with mostly subtidal bed levels are efficient sediment traps and tend to be net infilling as long as ample sediment is available. Fluvial sediments settle in bay-head deltas (predominantly sand), in the lower energy deeper parts of the basin, and on tidal flats and salt marshes (mainly mud). In relatively deep basins, sand and mud are imported from the sea as a result of tidal asymmetry. Due to the larger water depth during flood compared to ebb the tidal wave crest propagates at a higher velocity than the wave trough, which causes higher inshore than offshore-directed tidal peak flow velocity. As the relation between flow velocity and sediment transport is nonlinear, this leads to net import of sediment, which can be represented by the ratio of velocity amplitudes of the flood and ebb phase (Friedrichs, 2011).

In shallower sedimentary basins extensive shoals and intertidal areas emerge (e.g., Leuven et al., 2017). As a basin fills in, this may cause a shift from flood-dominated toward equilibrium or to ebb-dominated conditions due to feedbacks between hydrodynamics, sand transport and basin hypsometry (e.g., Boon and Byrne, 1981; Dronkers, 1986; Friedrichs and Aubrey, 1988; Wang et al., 2002; Kang and Jun, 2003; Fortunato and Oliveira, 2005; Moore et al., 2009; Friedrichs, 2010). Extensive intertidal areas cause the maximum flood velocities to occur at a later stage in the flood period and at a relatively high water level, during which the shoals and intertidal areas are flooded. Maximum ebb velocities occur at lower water levels when the water is concentrated in channels. As such, the ebb current has a much lower width-to-depth ratio with less bed friction and a higher flow velocity than the flood current. The ratio of flood to ebb sand transport thus decreases with increasing intertidal area. Sedimentary infilling with sand continues until a dynamic equilibrium is reached at which there is no tidal asymmetry and the estuary is in dynamic equilibrium (e.g., Boon and Byrne, 1981; Dronkers, 1986; Friedrichs and Aubrey, 1988; Dronkers, 1998).

These models thus suggest that there are hydromorphologic feedbacks on the evolution of sandy tidal embayments, resulting in sediment input in basins with high accommodation space and no net sediment import and dynamic equilibrium in basins with low accommodation space. However, the analysis so far ignores that the long-term evolution of most estuaries and tidal embayments is also controlled by the dynamics of muddy sediment and biological processes, which may alter the hydrodynamic feedback mechanism discussed above.

1.1.2. Substantial effects of mud and eco-engineering species on the evolution of estuaries and tidal embayments?

The transport of mud from the sea into tidal systems is mainly caused by tidal dispersion, estuarine circulation, tidal asymmetry and

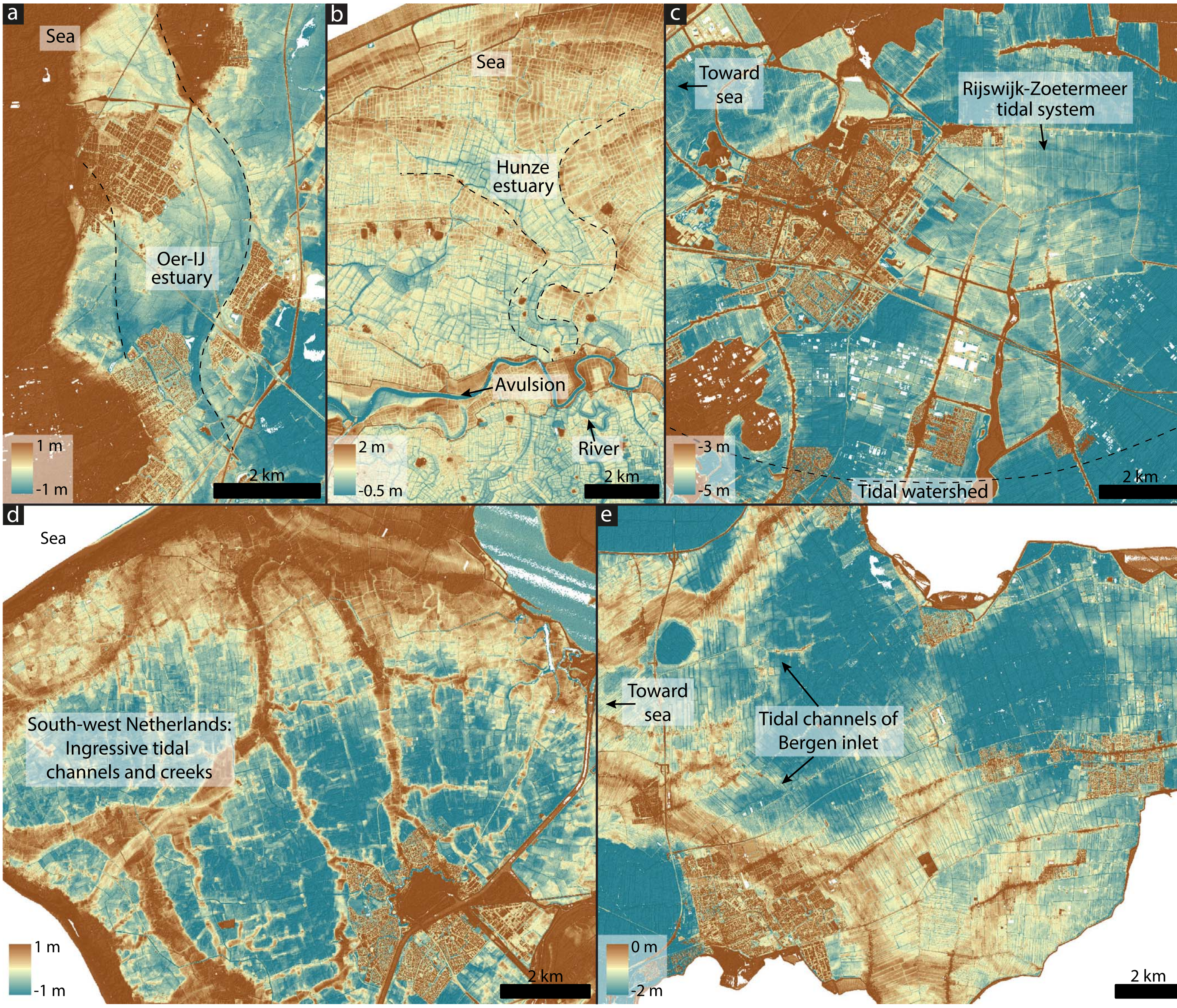


Fig. 2. Remnants of former estuaries, tidal inlets, tidal channels and tidal creeks visible in the present-day surface relief of The Netherlands. High elevated areas are mainly sandy remnants of creeks and tidal levees, lower areas are more subsided peat/clayey tidal floodbasin areas. See Fig. 1 for geographical location of the panels. (a) The Oer-IJ estuary in the west. (b) The Hunze estuary in the north. (c) The Rijswijk-Zoetermeer tidal system in the west. (d) Remnants of tidal channels and creeks in the southwest, formed by sea ingressions that eventually led to the formation of the Western Scheldt and Eastern Scheldt estuaries. (e) Remnants of tidal channels of the Bergen inlet (also referred to as Hauwert inlet) in the northwest. Elevation in m O.D., north is up in all panels.
Source: AHN. Period of activity of the systems is depicted in Fig. 10

scour and lag effects (e.g., Dronkers, 1986, 2005; Friedrichs, 2011), while mud may also be supplied by rivers. Dispersion leads to the transport of suspended sediment from areas of high concentration to areas of low concentration. In tidal systems with substantial freshwater inflow, a density-induced pressure gradient may generate a landward directed flow component near the estuary bottom and a seaward directed flow component near the surface; a flow pattern referred to as estuarine circulation. Settling of sediment particles into the lower part of the water column is followed by their landward transport through estuarine circulation, up to the limit of sea-water intrusion where they collect to form a turbidity maximum (e.g., Meade, 1972; Biggs et al., 1983; Wolanski et al., 1995; Uncles et al., 2006; Manning et al., 2010). Tidal asymmetry affects transport of fines in a similar way as transport of sandy sediment; a higher flood velocity results in landward transport and vice versa. But deposition of fines also depends on the slack time: a longer slack period after flood (high-water slack) compared to the slack period after ebb (low-water slack) leads to more settling of fines after flood and net landward transport. Conversely, a longer slack period after ebb will result in net seaward transport. High-water slack in interaction with bathymetry further favors landward transport and trapping of fines in estuaries, even in the absence of asymmetry (e.g., Visser, 1980; Dalrymple et al., 1991; Pritchard and Hogg, 2003). Around high-water slack a great amount of fine sediment moves over tidal flats (shoals), where the current velocity is low, while around low-water slack sediment particles are moving through channels, where velocities are much higher (e.g., Pritchard and Hogg, 2003). As a result, the period of low flow velocity is longer during high-water slack and more mud is able to settle. The average water depth during high-water slack is also lower as the flow is then closer to the lateral system boundaries and on the shoals, so that settling sediments reach the bottom faster. The minimum flow velocity for eroding mud is higher than the flow velocity for sedimentation (scour lag) (e.g., Sundborg, 1956; Van Straaten and Kuenen, 1957; Amos and Tee, 1989). As a result, more mud is typically deposited and remains on the seabed after high-water slack than after low-water slack resulting in net landward transport of mud (e.g., Dronkers, 2005). Finally, during stormy conditions on the shoreface, wave action may prevent deposition of fine sediment, whereas fines do settle in the sheltered environment of a tidal embayment or estuary leading to the import of mud into tidal embayments and estuaries (e.g., Dronkers, 2005).

There are thus multiple mechanisms that may cause ongoing accretion of intertidal and supratidal areas in estuaries and tidal embayments by import of marine sediments, which may counteract the tendency to evolve toward a morphodynamic equilibrium between hydrodynamics and sand transport predicted by numerical models (e.g., Boon and Byrne, 1981; Dronkers, 1986; Friedrichs and Aubrey, 1988; Dronkers, 1998). Evidence for this has been found in multiple estuaries around the world. For example, at a mudflat in the Eems-Dollard estuary (The Netherlands) with ebb-dominant tidal currents, Dyer et al. (2000) measured net sediment import. Similarly, numerical modeling of the Blyth estuary (UK) suggests that the hydrodynamics of this estuary are ebb dominant (French et al., 2005), whereas geomorphological evidence clearly demonstrates net sediment accretion within the estuary in the past 80 yr, mainly on the tidal flats and salt marshes (Pye and Blott, 2014). The Dovey estuary in Wales is currently ebb dominant at its mouth but flood dominant along the head of the estuary (Brown and Davies, 2010) and on the tidal flats (Braat et al., 2017). Likewise, the Dee estuary exhibits weak ebb dominance in its main channel whereas its banks and tidal flats are flood dominant and accreting (Moore et al., 2009). The Western Scheldt estuary is exporting sand and importing mud (Dam et al., 2016), and van der Spek (1997) showed that a reduction in intertidal area (rich in mud and animal and plant species) in the Western Scheldt over the last four centuries led to a strong reduction of flood dominance but not to ebb dominance, although there is net sand export and net mud import. Field data and detailed numerical modeling show that the net sediment transport direction in tidal inlets may vary spatially over an inlet. For example, the tidal inlet between the island of Texel and the Dutch mainland is hydrodynamically ebb-dominated on average over its cross-section, but flood dominated along its southern shore and ebb dominated along the northern shore (Elias, 2006). Moreover, during flood, sand is deposited in places where ebb transport cannot take it back offshore, resulting in net sediment import. This net sediment import is further enhanced during storms, when sediment is suspended from the ebb-tidal delta and subsequently imported into the Wadden Sea basin (Elias, 2006). These observations suggest that invoking the mechanism of tidal asymmetry is perhaps biased by a one-dimensional and single-grain size view on tidal systems. In reality, estuaries and tidal embayments are complex three-dimensional systems wherein the path of development depends on a complex set of interacting boundary conditions.

Based on the differences between sand and mud transport, Van den Berg et al. (1996) suggest that there are two stages of morphological development during the infilling of tidal embayments and estuaries along the Dutch coast (Fig. 3). In the first stage tidal asymmetry results in ample sand transport into the estuary until shoals and other intertidal areas have grown large enough to cancel or significantly diminish the flood-dominated tidal asymmetry. By this time, the channel and shoal morphology approaches a dynamic quasi-equilibrium and the system enters the second stage. During the second stage, continued infilling with mud causes accretion of intertidal shoals to supratidal levels, thus reducing the total intertidal shoal area, as also shown in numerical modeling (Braat et al., 2017). This then partially restores flood dominant conditions, causing renewed net sand accumulation in the basin, followed again by further growth of supratidal areas. This feedback loop leads to ongoing filling of the tidal embayment or estuary rather than equilibrium. This is exacerbated by the reduction of tidal prism (the volume of water that enters a tidal system at flood tide), changing the balance between prism and longshore sand-transport. In a final stage of closure, the prism-littoral drift ratio determines inlet stability

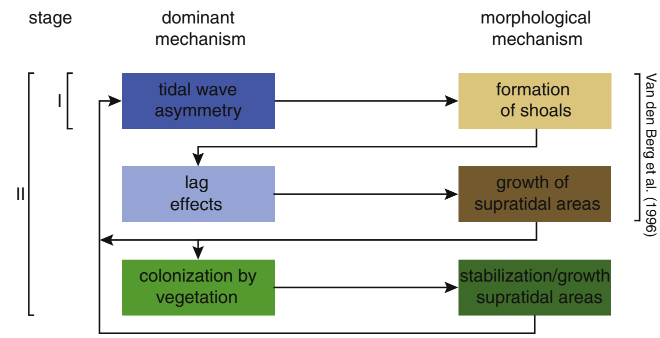


Fig. 3. Hypothetical evolution of tidal embayments and estuaries, extended from Van den Berg et al. (1996). (I) Stage 1: in a relatively deep tidal embayment with sand supply, tidal wave asymmetry leads to a predominantly sandy embayment infilling. This causes formation and growth of shoals until channel and shoal morphology are in dynamic equilibrium (cf. Friedrichs and Aubrey, 1988). (II) Stage 2: infilling of mud by settling and scour lag and ecological effects, mainly vegetation, lead to growth of supratidal areas, resulting in a reduction of intertidal shoal area, thereby re-enhancing tidal wave asymmetry effects and sand import. This feedback mechanism progressively reduces the tidally active area of tidal embayments and estuaries and may result in closure of tidal embayments.

and an inlet will close when tidal flushing becomes inefficient relative to the longshore sediment supply (e.g., Bruun and Gerritsen, 1960).

The conceptual model by Van den Berg et al. (1996) suggest that estuaries and tidal embayments typically fill up when sand is available, until intertidal areas are formed. When the boundaries are fixed and when there is only sand this may lead to equilibrium. But in the presence of mud the net sedimentation on intertidal and supratidal areas co-determines whether an estuary or tidal embayment remains open or fills up further. The hydromorphic feedbacks potentially lead to complete filling and closure of the tidal embayment or estuary. However, before this conceptual model can be used for accurate prediction of tidal system evolution and model design, a number of processes and feedbacks controlling the evolution of intertidal and supratidal areas need to be addressed.

To begin with, the biogeomorphodynamics of intertidal and supratidal areas are controlled by a large range of processes that may either promote accretion or erosion under different conditions. Tidal asymmetry, scour and settling lags and dispersion cause landward sediment transport and accretion of tidal flats during calm conditions when wave energy is low (e.g., Dronkers, 1986; Dyer, 1995; Dronkers, 2005; Friedrichs, 2011). On the other hand, during stormy conditions, sediments in intertidal areas are brought into suspension by waves. Wave heights tend to be highest around high water, so more sediment is suspended around slack after flood than slack after ebb, leading to seaward net transport (Friedrichs, 2011). In such cases the wind induced resuspension leads to an export of fine sediment. Landward tide-induced sediment transport thus results in tidal flat accretion during low-wave conditions, while more energetic waves may result in erosion of tidal flats (Allen and Duffy, 1998; Ridderinkhof et al., 2000; Janssen-Stelder, 2000; Yang et al., 2008, 2003; Friedrichs, 2011; Zhou et al., 2015). However, at the open coast the strong wave action during storms leads to ample sediment entrainment, and therefore to inflow of sediment-rich water into the tidal system potentially leading to net sediment input regardless of the internal sediment dynamics in the tidal system (e.g., Dronkers, 2005). Whether calm weather or storm conditions dominate net accretion of mud on the long term is generally unknown. Rainfall disrupts tidal flat sediment more around low water, and as a result the flood tide following a rainfall event tends to move this sediment landward (e.g., Tolhurst et al., 2006). Yet, rain-induced runoff or bank collapse might cause erosion of intertidal and supratidal areas (e.g., Tolhurst et al., 2006; Kleinhans et al., 2009; Choi et al., 2013). Small-scale gullies within tidal flats may then export sediment

(Dyer, 1995; Mariotti and Fagherazzi, 2011; Ralston et al., 2013), although on relatively broad flats cross-shore currents may be more important for sediment dispersal (Le Hir et al., 2000).

In addition to mud erosion and sedimentation processes the expansion of intertidal and supratidal areas (tidal marshes) may be substantially affected by eco-engineering species, in particular those that significantly change their environment at the landscape scale (Jones et al., 1994; Bouma et al., 2014). Eco-engineering species may either act as sediment stabilizers (e.g., vegetation, diatoms and oysters) that promote accretion (e.g., Reise, 2002; Corenblit et al., 2007; Bouma et al., 2013; Gurnell, 2014) or as sediment destabilizers (mainly bioturbating species, such as snails, shells, worms and shrimps) that promote erosion (e.g., Widdows and Brinsley, 2002; Reise, 2002; Le Hir et al., 2007; Kristensen et al., 2012). The net long-term effect of these species on the tidal system scale will depend on the species distribution and density as well as on recovery after sustained environmental disturbance by currents, storm- and ship-generated waves and winter temperature (e.g., Frostick and McCave, 1979; Paterson, 1989; Orson et al., 1992; Andersen et al., 2005; Friedrichs et al., 2008). The aggregated effect of these eco-engineering species on the century to millennia evolution of estuaries and tidal embayments, however, remains largely unknown.

We hypothesize that, in general, the aggregated effect of species favors sediment accretion in estuaries on the century to millennium scale, and therefore we mainly focus on the potential effects of sediment stabilizers, especially vegetation, on the long-term evolution of tidal systems. Vegetational effects have been identified to play a key role in salt-marsh evolution and have therefore been commonly incorporated in salt-marsh evolution models (e.g., Allen, 2000; Fagherazzi et al., 2012; Zhou et al., 2016). When intertidal flats become high enough relative to mean water level they can be colonized by vegetation (e.g., Temmerman et al., 2007; Fagherazzi et al., 2012; Bartholdy, 2012). Colonization of tidal flats by laterally expanding vegetation is widely reported from tidal areas across the world (e.g., Callaway and Josselyn, 1992; Sanchez et al., 2001). When clastic sediment supply is sufficient, salt marsh vegetation can accumulate extensive amounts of fine-grained sediment resulting in the formation of stable salt marsh plateaus (e.g., Gleason et al., 1979; Allen, 1989; Morris et al., 2002; van Proosdij et al., 2006; Yang et al., 2008; Li and Yang, 2009). Vegetation minimizes erosion of intertidal and supratidal areas by decreasing wave and current energy and roots that enhance the stability of the subsurface, thereby favoring salt marsh accretion and expansion. Moreover, salt marshes may partly accrete by direct deposition of organic sediments (e.g., Turner et al., 2000; Neubauer, 2008; Langley et al., 2009; van Maanen et al., 2015). The rates of accretion of both clastic and organics are positively related to plant biomass (e.g., Gleason et al., 1979; Dijkema et al., 1990; Morris et al., 2002; Li and Yang, 2009), which is controlled in part by the elevation of the marsh platform (Morris et al., 2002; Li and Yang, 2009), and in part by species richness (Ford et al., 2016). There is thus a strong feedback between the elevation of salt marshes and marsh vegetation density (Fagherazzi et al., 2012).

In terms of the conceptual model for long-term estuary evolution by Van den Berg et al. (1996), progressive estuary infilling by continuous expansion of supratidal areas may not only be caused by morphodynamic conditions favoring continuous mud accretion but also substantially enhanced by biomorphological feedbacks with vegetation and possibly other biota (Fig. 3). We test this concept by analysis of the long-term evolution of Dutch estuaries and tidal embayments as a result of the aggregated effects of the aforementioned biomorphological processes and feedbacks. Before we do so, we first briefly discuss potential effects of river input on the stability and evolution of estuaries and tidal embayments.

1.1.3. Long-term effects of rivers on the stability of estuaries and tidal embayments

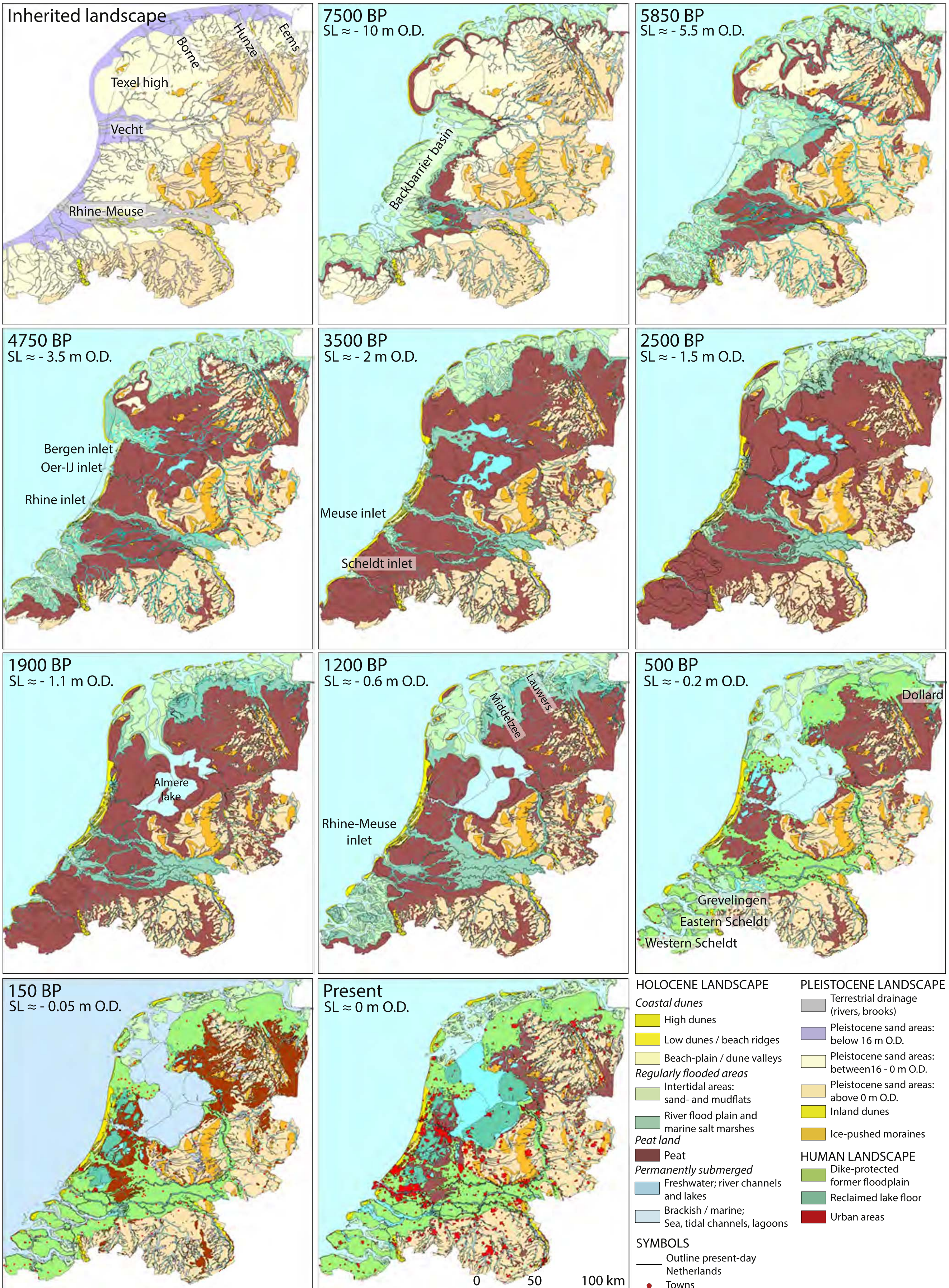
River inflow interacts with tides and partly determines morphological development and equilibrium dimensions of the estuary (Reynolds, 1889; Townend, 2012; Savenije, 2015). River influxes directly affect estuarine morphodynamics by supplying freshwater, sand and mud. The unidirectional freshwater inflow enhances ebb currents, and interacts in other ways with tides (e.g., Guo et al., 2014). As such, rivers substantially influence the evolution of estuaries on short to long timescales. Over short timescales, river discharge dampens incoming tidal waves by increasing tidal friction (e.g., Savenije, 2005), may induce stratification and density currents (e.g., Chant et al., 2011; Gong et al., 2014) and constrains landward saltwater intrusion by enlarging ebb currents (e.g., Horrevoets et al., 2004). River-enhanced ebb currents provide a major mechanism for exporting estuarine sediments seaward (e.g., Garel et al., 2009). River floods may cause short- to medium-term (i.e., year to decade) cycles of morphological change, generally characterized by estuary erosion during floods and post-flood recovery (e.g., Cooper, 2002, 1993; Karunarathna, 2011; Nittrouer et al., 2012). Locally, river floods provide large amounts of sediment to estuaries (Dalrymple et al., 2015; Jablonski and Dalrymple, 2016). These short-term effects have been studied relatively well (e.g. Savenije, 2005). The long-term (i.e., centuries or more) effects of river influxes on estuarine evolution, on the other hand, are poorly understood. Some direct effects were recently investigated numerically (e.g., Guo et al., 2014; Bolla Pittaluga et al., 2015; Guo et al., 2016), but these models have a strongly simplified and imposed estuary shape, thereby generally neglecting effects of intertidal areas, tidal flats, fine-sediment dynamics, vegetation and discharge and sediment supply variations over daily to centennial timescales. How rivers interact with estuarine processes and how they influence the long-term evolution and stability (dynamic equilibrium) of tidal systems is thus poorly understood. Here, we compare the long-term evolution of Dutch tidal systems with and without substantial fluvial input, to determine the aggregated effect of river input on estuary development.

2. Controls on the contrasting Holocene evolution of tidal systems along the Dutch coast

In this section we summarize the Holocene evolution of the Dutch coastal plain and its estuaries and tidal embayments. We first briefly describe the geographical setting of the present-day coast of The Netherlands. Then we depict the Holocene boundary conditions that controlled the evolution of the Dutch coast: pre-transgressive topography, sea-level rise, tidal range, wave height and fluvial sediment supply. Then the sources of sediment that made up the coastal plain are presented. This information is used to describe the development of The Netherlands' coastal system as a whole over the Holocene. Finally, we compare and summarize the contrasting paths of evolution of the main estuaries and tidal embayments, which are the combined results of inherited topography, fluvial and coastal boundary conditions, sand and mud transport and deposition and eco-engineering species, and identify the major controls on their evolution.

2.1. Geographical setting

The present-day coastal plain of The Netherlands consists of barriers and back-barrier deposits on top of a low-sloping shelf (Fig. 1) (e.g., Beets and van der Spek, 2000; Vos, 2015). The coast of The Netherlands and its tidal systems are not bound by bedrock. The Dutch tidal systems are therefore comparable to amongst others the modern tidal systems along the US Gulf Coast and the US southeast coast (Georgia) (e.g., DeLaune et al., 1987; Dame et al., 2000; Anthony et al., 2009), but differ from many other traditionally studied modern estuaries and tidal embayments such as along the Canadian and US East Coast (e.g., Dalrymple et al., 1994), the French Atlantic Coast (e.g., Allen and Posamentier, 1993; Chaumillon et al., 2010), south-eastern Australia (e.g., Roy, 1994), Portugal (e.g., Vis et al., 2016) and the UK (e.g., Pye and Blott, 2014). The formations that underlie and border the coastal



(caption on next page)

Fig. 4. Palaeogeographical maps of the Holocene development of the Dutch coastal and deltaic plain. All ages are in calibrated yr BP. Animation in online supplement.
Source: Adapted from Vos (2015).

wedge largely consist of unconsolidated sand, locally with gravel, silt, clay and peat; predominantly deposited in the last two glacial cycles in various periglacial terrestrial environments (Busschers et al., 2007; Cohen et al., 2014; Peeters et al., 2015), at low topographical gradients of 0.15–0.20 m/km. Furthermore, the sand-rich North Sea off the Dutch coast is relatively shallow and subject to tidal and storm wave reworking, thus providing a large source of sediment to the Dutch coast (e.g., Beets et al., 1992; Cleveringa, 2000). As a result, an extensive coastal plain developed during the Holocene in The Netherlands, including extensive supratidal areas and peat bogs in former tidal areas (e.g., Beets et al., 1992; Vos, 2015). Rapid post-glacial rise in sea level led to flooding of the wide, low-sloping and unconsolidated substrate creating ample accommodation space for developing a coastal plain with multiple estuaries and tidal embayments. These estuaries and tidal embayments formed, filled in and closed off or persisted over time (e.g., Beets and van der Spek, 2000; Vos, 2015; Pierik et al., 2017). Moreover, many avulsions occurred in the rivers bifurcating across this coastal plain during the Holocene (Berendsen and Stouthamer, 2000; Stouthamer et al., 2011; Cohen et al., 2012), thereby affecting the tidal systems along the Dutch west coast. The deposits of the former estuaries, tidal embayments and rivers are well-preserved in the subsurface (Figs. 1, 2), and the spatio-temporal development of many of these systems as well as their boundary conditions have been reconstructed in great detail (e.g., Pons and Wiggers, 1959; Jelgersma, 1961; Pons et al., 1963; Roeleveld, 1974; De Mulder and Bosch, 1982; Beets et al., 1992; Stouthamer and Berendsen, 2001; Hijma and Cohen, 2011; Cohen et al., 2012; Vos, 2015; Pierik et al., 2016, 2017).

The Holocene tidal systems in The Netherlands were mostly self-formed in the sense that their planform geometry, dimensions and development depended mainly on initial conditions, boundary conditions and biogeomorphological processes rather than inherited valley shape or other hard-rock constraints. This allows us to infer to some degree how contrasting development depended on each of these factors.

From the landward side, several rivers debouch into the North Sea. The Rhine (yearly-averaged discharge $\sim$2300 $m^3\ s^{-1}$) and Meuse (yearly-averaged discharge $\sim$230 $m^3\ s^{-1}$) rivers presently debouch into the North Sea just west of the city of Rotterdam. The present-day Scheldt river (yearly-averaged discharge $\sim$100 $m^3\ s^{-1}$) flows from Belgium through the Western Scheldt estuary into the North Sea in the south of The Netherlands. North of the river Rhine no major rivers currently debouch into the North Sea (Fig. 1). At the Dutch-German border the Eems river flows into the Dollard estuary and Wadden Sea with a yearly-averaged discharge of $\sim$90 $m^3\ s^{-1}$ (Van Maren et al., 2016).

The present-day Dutch coast has three sectors with distinct morphologies (e.g., Beets et al., 1992; Van der Molen and de Swart, 2001b) (Fig. 1). The first is an open estuarine coast in the southwest, which has been largely closed off by dams or sluices. This is the region between the Belgian border and the present mouth of the river Rhine and Meuse. The second is a closed beach-barrier/strand-plain in the western Netherlands between the mouth of the Rhine and Meuse rivers and the Wadden Sea. The third sector is an open barrier/back-barrier coast: the Wadden Sea in the northwest to north. The differences in the sectors can be attributed to varying paths of evolution, discussed later in more detail.

2.2. Holocene initial and boundary conditions

The Dutch coast is exposed to prevailing strong westerly winds and dominant north to east-going longshore sediment transport (Beets and van der Spek, 2000). Hydrodynamic conditions are mixed, with a micro- to mesotidal regime (0.5–4 m tidal range) and significant storm wave energy but little swell (Beets et al., 1992). The mean tidal range varies from 3.4 m at the mouth of the Western Scheldt in the southwest, to 1.4 m at Den Helder, and increases to 2.6 m at the mouth of the Eems estuary (see Fig. 1 for locations). The present-day annually averaged significant wave height along the western part of The Netherlands is $\sim$1.3 m (Kroon, 1990), and increases northward to 1.8 m near the island of Texel (Sha, 1989). The offshore significant wave height during 10-year storms is up to 9–10 m and dissipates on beaches and coastal dunes (Ruessink et al., 1998).

During the Last Glacial Maximum (LGM; $\sim$22 ka BP; all ages in this paper are in calibrated years before present) the rivers of the southern North Sea merged in the area off the SW Netherlands' coast (Hijma et al., 2012) and joined to continue as the Channel River to the Atlantic Ocean into a low-stand estuary well south of the Strait of Dover (e.g., Toucanne et al., 2009). The rivers predominantly had a braided planform and formed wide periglacial palaeovalleys in The Netherlands, of which the Rhine-Meuse and Vecht valleys were the largest (Fig. 4: Inherited landscape, $\sim$11,000 yr BP) (Pons, 1954; Cohen et al., 2002; Busschers et al., 2007; Cohen et al., 2014; Koster et al., 2016). These two major valleys were typically incised 2–5 m and were 40–60 km wide (Koster et al., 2016). The glacial to interglacial transition toward the end of the Pleistocene resulted in a climate-driven change of river style from braided in the Late Pleniglacial ($\sim$15 ka BP) to meandering in the early Holocene (Pons, 1957; Berendsen et al., 1995; Hijma and Cohen, 2011). Low-stand incision of the rivers in The Netherlands has been modest only, and consequently valley relief on the drowning early Holocene coast was gentle with erodible boundaries (Hijma et al., 2012).

Sea-level rise was relatively fast in the Early and Middle Holocene, because of the combined effects of absolute sea-level rise and accelerated subsidence from forebulge collapse (Vink et al., 2007; Koster et al., 2016). From 10 to 8 ka BP sea level rose from −34 to −13.5 m O.D. in the southern North Sea, at an average rate of 1 m per 100 yr (Fig. 5a). All transgressive tidal systems listed in this paper formed after this period, as the tidal landscapes of the Early Holocene existed offshore from the modern coastline. From 8.45 to 8.25 ka BP, around the start of the middle Holocene, sea-level rise temporally accelerated (e.g., Hijma and Cohen, 2010; Törnqvist and Hijma, 2012). After $\sim$8.25 ka BP sea-level rise progressively slowed down from 0.55 m per 100 yr between 8 and 7 ka BP to 0.05 m per 100 yr in the last 2 ka BP. Relative sea-level rise during the last 8 ka was largest along the northern coastal sector of The Netherlands ($\sim$20 m) and decreased toward the south ($\sim$12.5 m) due to southward declining subsidence by forebulge collapse (Jelgersma, 1979; Ludwig et al., 1981; Denys and Baeteman, 1995; Kiden, 1995; Kiden et al., 2002; Van de Plassche et al., 2010; Hijma and Cohen, 2010).

The changing sea level in the shallow North Sea led to changing tidal currents and wave heights due to bed friction and depth-limitation. The Holocene evolution of the significant wave height and tidal range off the Dutch coast were reconstructed by Franken (1987), Van der Molen and Van Dijck (2000), Van der Molen and de Swart (2001b) and Van der Molen and de Swart (2001a). These reconstructions were based on numerical modeling using the present-day bathymetry as a starting position, and are considered to capture the overall trends while near the coast the evolution may have been more complicated. Mean significant wave height along the Dutch coast increased from $\sim$0.6 m around 7.5 ka BP to the present-day value exceeding 1 m along the entire coast (Fig. 5c) (Van der Molen and de Swart, 2001b). At water depths $>$ 20 m, the wave-induced sand transport mode changed from dominantly suspended transport before 6 ka BP to dominantly bedload transport thereafter due to the increasing water depth in the North Sea by sea-level rise (Van der Molen and de Swart, 2001b). Waves probably

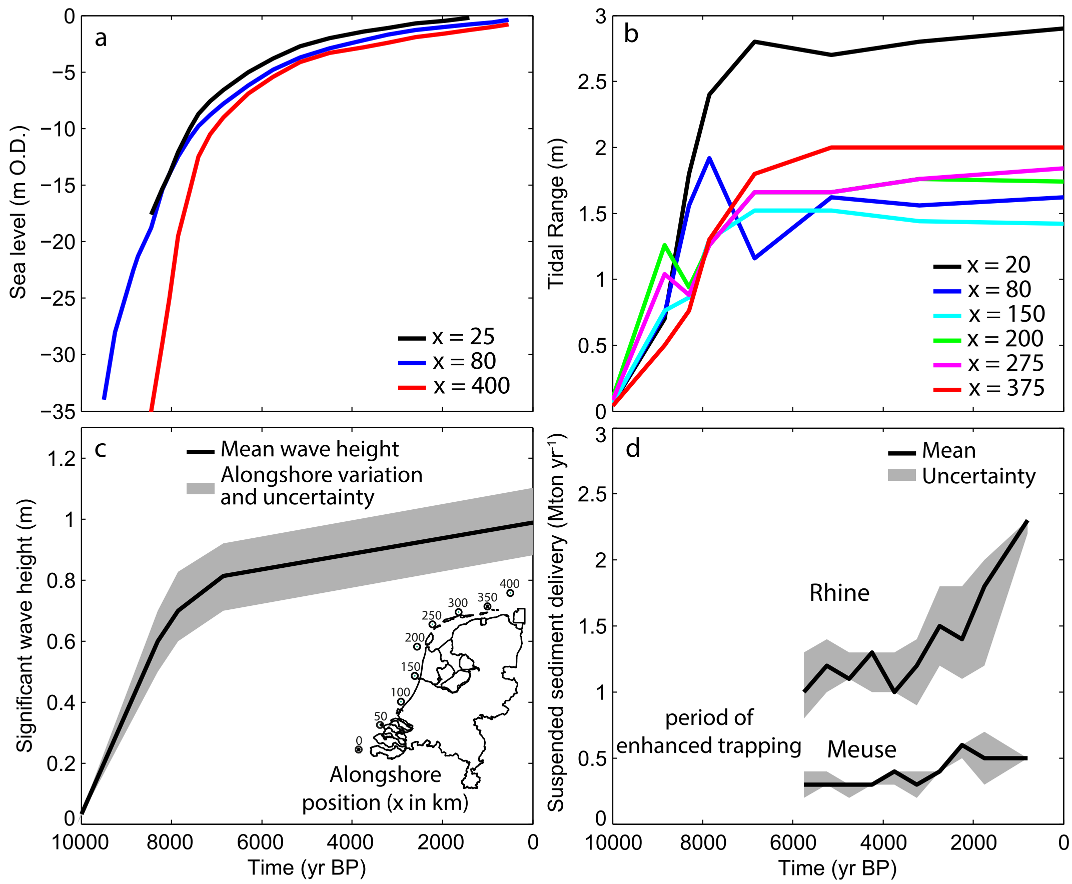


Fig. 5. Holocene environmental boundary conditions. (a) Relative sea level (after Jelgersma, 1979; Ludwig et al., 1981; Kiden, 1995; Van de Plassche et al., 2010; Hijma and Cohen, 2010). See inset map in c for alongshore locations (x in km). (b) Tidal range (after Van der Molen and de Swart, 2001a). (c) Mean significant wave height (after Van der Molen and de Swart, 2001b). (d) Trapped part of suspended sediment supplied to the fluvial area of Rhine-Meuse delta (between 50 and 150 km) (after Erkens, 2009). Trapping rates were relatively high until 6000 yr BP due to enhanced trapping by the high sea-level rise and are therefore not plotted here. Fluvial sediment delivery to other parts of the coastline was orders of magnitude smaller.

only had a substantial contribution to the large-scale net sand transport before 6 ka BP when the North Sea was shallow, mostly by stirring up bed sediment which enhanced current-driven transport (Van der Molen and Van Dijck, 2000; Kleinhans and Grasmeijer, 2006).

Until approximately 9 ka BP there was no connection between the northern and southern parts of the North Sea, and consequently tidal amplitudes off the Dutch coast were low (Fig. 5b) (Van der Molen and Van Dijck, 2000; Van der Molen and de Swart, 2001a; Van der Molen, 2002; Uehara et al., 2006). As sea level increased, the Doggerbank land bridge between Great Britain and The Netherlands was drowned and the North Sea basin became deeper, which resulted in an increase in tidal and current amplitudes until 6 ka BP. Since 6 ka BP tidal conditions have been relatively stable. At 7.5 ka BP, tidal currents were oriented toward the Dutch coast, which likely resulted in transport of substantial volumes of sand from the bottom of the North Sea to the Dutch coast by tidal asymmetry. After sea level rose further, the tide was able to propagate closer to the shore and the main tidal current became redirected in an alongshore direction (Hijma et al., 2010). This resulted in a north-eastward alongshore net sand transport direction, although a coastward component remained present in southwest Netherlands that diminished to very small or slightly negative values after 6 ka BP (Van der Molen and Van Dijck, 2000; Van der Molen and de Swart, 2001a). In the last 7500 yr tides have dominated net sand transport off the Dutch coast (Van der Molen and Van Dijck, 2000; Van der Molen, 2002) by a strong flood-dominated tidal asymmetry on the shallow North Sea floor off the southwestern to northwestern parts of The Netherlands (Dronkers, 1986; van der Spek, 1994; van Dijk and Kleinhans, 2005).

2.3. Sediment sources contributing to coastal plain formation

Large amounts of sand were transported from the sandy, shallow, North Sea bed toward the coast of the southwest and western parts of the Dutch coast until approximately 6 ka BP, after which the net sand supply from the sea decreased. Van der Molen and Van Dijck (2000) estimate that the tidal sand transport to the coast of southwest Netherlands decreased from 65 m^3 m^{-1} yr^{-1} (volume per linear meter of coastline per year) at 7.5 ka BP to -15 m^3 m^{-1} yr^{-1} at present and for the coast of western Netherlands from 120 m^3 m^{-1} yr^{-1} at 7.5 ka BP to 10 m^3 m^{-1} yr^{-1} at present, in general agreement with field measurements (van Dijk and Kleinhans, 2005; Kleinhans and Grasmeijer, 2006). The decrease in net sand transport from the North Sea floor to the Dutch coastal plain is mainly caused by a decrease in accommodation space over time, and thus less efficient sediment trapping, resulting from progressive infilling of the coastal backbarrier basins (e.g., Beets et al., 1992, 1994). The Wadden Sea coast (north-western and northern Netherlands) exchanged less sand ($\sim$5 m^3 m^{-1}) with the North Sea throughout the Holocene, mainly because of the greater depth of the North Sea off the Wadden Sea (Van der Molen and Van Dijck, 2000). This difference in sediment availability partly explains the difference between the closed western and open northern sectors of the Dutch coast. Moreover, large amounts of sediments were transported to the western Netherlands by the Rhine and Meuse rivers.

Sediment delivery from the rivers also changed during the Holocene. Due to the very low gradient of the lower part of the Rhine-Meuse system and predominantly fine-grained deposits, the channels were characterized by low lateral mobility, large-scale crevassing and a

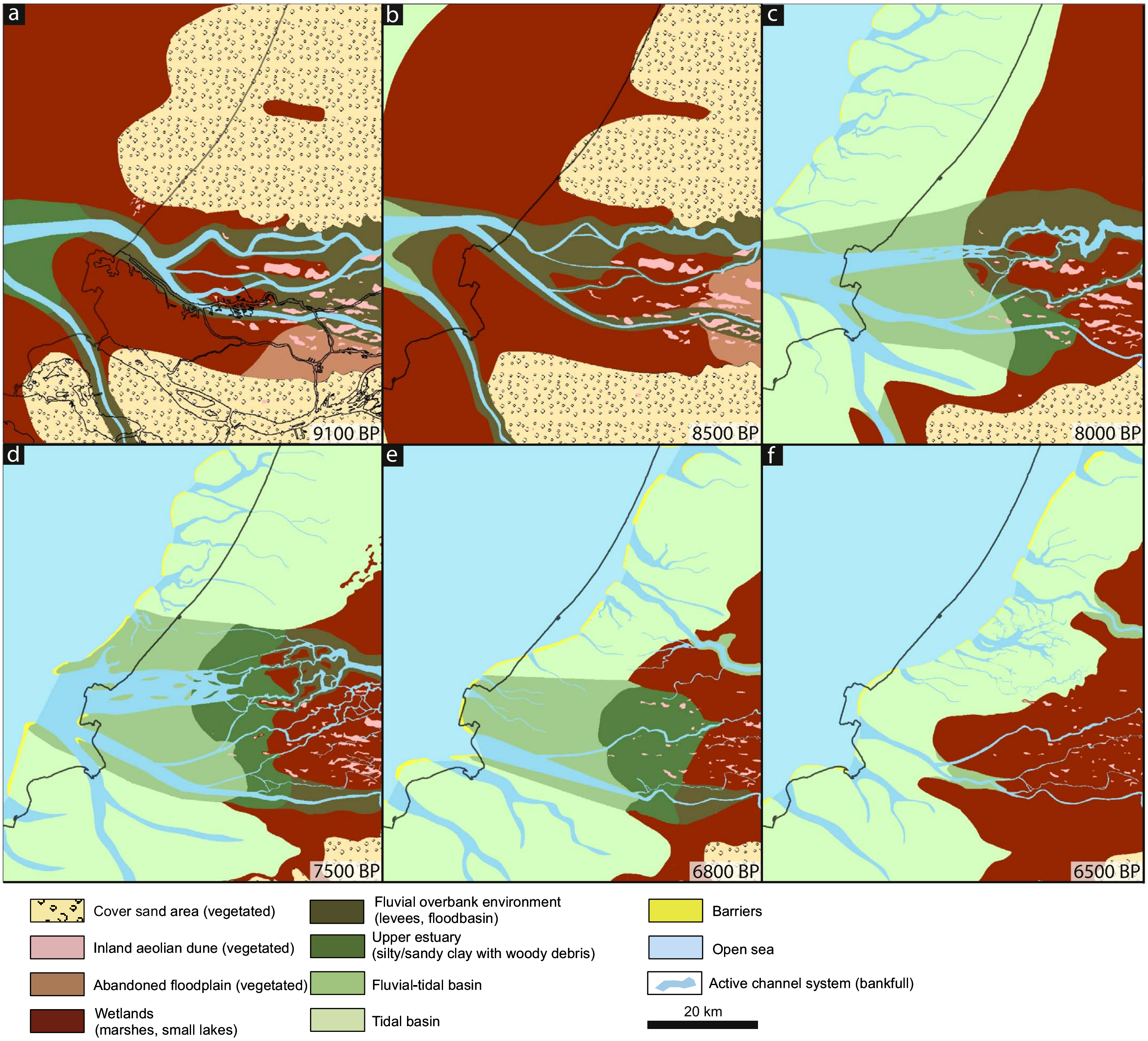


Fig. 6. Detailed palaeogeographic reconstruction of the Rhine and Meuse estuary between 9100 yr BP and 6300 BP, showing the early transgressive development of the Rhine-Meuse estuary (from Hijma and Cohen, 2011). During the middle Holocene this tidal system evolved from a bay-like estuary constrained by the Pleistocene palaeovalley of the Rhine and Meuse to a laterally extensive backbarrier basin.

high avulsion frequency throughout the Holocene (e.g., Tornqvist, 1993; Berendsen and Stouthamer, 2000; Stouthamer and Berendsen, 2001). This repeatedly changed the spatial distribution of fluvial sediment in the western section of The Netherlands over the Holocene. Between highstand and embankment, the Old Rhine was the most dominant river branch transporting sediments to the coast. The other rivers ran into the peat area or Meuse estuary, transporting relatively small amounts of sediments to the coastline. At present, the Rhine and Meuse rivers transport suspended sediment to the North Sea estuaries and harbors due to extensive river embankment (Beets and van der Spek, 2000; Hudson et al., 2008). Intra-Holocene climate change is considered to have had a limited effect on the discharge of the Rhine and Meuse Rivers (Ward et al., 2008; Erkens, 2009). Discharge variation of the Rhine, for example, is estimated to have been within 10% during the Holocene (Stouthamer et al., 2011). On the other hand, Erkens (2009) showed that early Holocene suspended sediment trapping on the fluvial plain was 1.5 times larger than middle Holocene trapping, while human-induced land use changes in the hinterland, after some delay, caused the fluvial mud transport rates to progressively increase again since 3000 ka BP, despite a slight decrease in trapping efficiency (Erkens, 2009) (Fig. 5d). This suggests that mud concentrations in the estuaries connected to rivers were high in the early Holocene, lower in the middle Holocene and higher again in the late Holocene. This may have affected the rate of coastal plain infilling but this is unknown.

In the Holocene coastal plain approximately 226×10^9 m^3 of sediment is stored (sand, 65–70%; clay, 25–30%; peat, 5%) (Beets and van der Spek, 2000). Beets and van der Spek (2000) estimate that about 60% was trapped prior to 5 ka BP, 30% between 5 ka and 2.9 ka BP and 10% since then. All clastic sediments in the coastal plain of The Netherlands were derived from three sources: a fluvial source (e.g., Rhine, Meuse and Scheldt rivers), the Pleistocene basement eroded during recession of the shoreline, and the North Sea floor (both predominantly consisting of older fluvial deposits) (Beets and van der Spek, 2000). Van der Spek and Beets (1992) estimate that the sedimentary succession of the Holocene backbarrier deposits in the western Netherlands, formed during the middle Holocene, has an average mud content of 50–60% and an average sand content of 40–50%, highlighting the importance of mud on the evolution of these tidal systems.

2.4. Holocene evolution of the Dutch coastal plain

2.4.1. Middle Holocene transgression-formed tidal systems

The sea approached the present-day Dutch coastline ~8.5 ka BP, around the start of the middle Holocene. Due to the rapid postglacial sea-level rise low-stand valleys were inundated by the sea and large wetlands were formed (Fig. 4: 7500 yr BP) (e.g., Pons et al., 1963; Beets et al., 1992; Vos, 2015). Lowest lying valleys were inundated first, but as the valley shoulder topography stood just a few meters higher, soon after a major part of the Dutch coast became submerged. Along the modern coastline only the higher-lying Pleistocene cape of Texel (Fig. 4: Inherited topography) withstood sea-level rise for a few thousands of years longer. At the beginning of the transgressive stage, creation of accommodation space by the rapid sea-level rise of up to 2 m per 100 yr outpaced the relatively large input of sediment from marine and fluvial sources (e.g., Van der Molen and Van Dijck, 2000; Beets and van der Spek, 2000). Although this led to a net increase in accommodation space, the large sediment availability caused especially high rates of sediment deposition in the tidal areas in the southwest to northwest Netherlands. Marine aggradation in the low-lying fluvial palaeovalleys started around 8500 yr BP, and large estuaries and tidal embayments formed in these valleys (van der Spek and Beets, 1992; Beets et al., 2003; Hijma and Cohen, 2011; Vos, 2015).

From 8500 to 7500 yr BP a large estuarine system existed in the Rhine-Meuse palaeovalley with substantial fluvial input from both the Rhine and Meuse rivers (Fig. 6) (Hijma and Cohen, 2011). Simultaneously, an extensive tidal embayment formed in the Vecht palaeovalley without substantial fluvial input with diminishing river inflow from relatively small local rivers. Both systems were ~30–40 km wide and valley shoulder topography was approximately 5–10 m high. Smaller estuaries formed in the Scheldt in the southern Netherlands (Kiden, 1995) and in the relatively deep and narrow palaeovalleys of the Eems, Hunze and Borne valleys in the northern Netherlands (5–10 m deep, ~10 km wide) (Behre, 2003; Vos and Knol, 2015). The size of these systems during this period was largely determined by the size of the palaeovalleys wherein they formed.

From 7500 until 6000 yr BP the Rhine-Meuse estuary silted up, and because of several avulsions in northward direction the main freshwater and sediment supply of the Rhine was lost from the former Rhine-Meuse estuary (e.g., Hijma et al., 2009). Similarly, in the southwestern Netherlands the Scheldt River abandoned its former estuary to form a more southward-located new estuary. Due to progressive sediment trapping in these systems their size decreased over time. Along large parts of the Dutch coast beach barriers developed that migrated inland until ~6 ka BP. The formation of new tidal inlets between these barrier islands resulted in an increasing number, smaller-sized tidal systems around 6000 yr BP than during the initial transgression from approximately 8500 to 7500 yr BP. Moreover, their locations were generally no longer directly linked to the former palaeovalleys. This development was most pronounced in the southwestern to northwestern Netherlands, whereas occupation of accommodation space proceeded at lower rates in the northwestern to northern Netherlands so that systems decreased in size more slowly.

The rapid sea-level rise from 7500 to 6000 yr BP caused sand imported to the tidal inlets in between the beach barriers to deposit adjacent to the tidal channels (van der Spek and Beets, 1992). This resulted in the formation of levees along these channels, which were initially submerged but later became dry during low tides (e.g., Beets et al., 2003). Due to the sheltering effect of these levees, relatively tranquil lagoonal conditions prevailed behind the levees resulting in deposition of large quantities of muds (e.g., Westerhoff et al., 1987; Van der Spek and Beets, 1992; Beets et al., 2003). The enhanced mud trapping may have affected benthic animal species density, which, in turn, may have affected the evolution of the tidal systems but this hypothesis has not been tested so far.

Around 6 ka BP marine and fluvial sedimentation kept up with or exceeded sea-level rise, which had declined to ~0.3 m per 100 yr around that time (Fig. 4: 6000 yr BP). This led to silting up of the backbarrier area, which resulted in closure of the coast in the southwestern and western parts of The Netherlands by amalgamating beach barriers (e.g., Beets et al., 1992; Cleveringa, 2000; Van Heteren et al., 2011), because reduced tidal flushing could no longer keep open many of the inlets. Plant and animal species were abundant in these basins (e.g., Bakker and Van Smeerdijk, 1982; Vos and Van Kesteren, 2000) and an increasing vegetation density is recorded in supratidal salt marsh sediments compared to intertidal flat sediments (Bakker and Van Smeerdijk, 1982), again suggesting a significant role of eco-engineering species on tidal system evolution. Silting up of the basin happened in two-phases, as recognizable in the sedimentary succession of these basins (Fig. 7) (e.g., De Mulder and Bosch, 1982; Vos and Van Kesteren, 2000; Hijma et al., 2009; Vos, 2015). Initially, basins were dominantly filled by sand and mud, progressively reducing tidal prism and channel volumes. Then in the last phase, intertidal to supratidal deposition of mud, promoted by the emergence and expansion of vegetation, dominated. This was followed by full closure of the tidal embayments as inferred from a thick layer of freshwater peat on top of this sequence (e.g., De Mulder and Bosch, 1982; Vos and Van Kesteren, 2000; Hijma et al., 2009). Infilling rates during this last phase of muddy to peaty deposition were locally high, as evidenced by the abundance of in-viva reed remnants found in sedimentary successions showing that centimeters to decimeters of mud accumulated within the lifetime of the reed. These observations highlight the importance of biota, in this case reeds, in trapping fine sediments and closure of the tidal systems.

The filling and closure of the large inherited palaeobasins of the Vecht and the Rhine-Meuse, evolved differently as a result of differences in sediment supply. While abundant fluvial sediment input filled up the landward part of the Rhine-Meuse basin (Hijma et al., 2009), the landward part of the Vecht basin only filled in modestly (van der Spek and Beets, 1992). The marine processes were only able to import sandy sediments up to a few tens of kilometers from the tidal inlet of the Vecht basin so that peat growth dominated further inland (Bergen inlet; Fig. 4: 4750 yr BP). Continuous infilling at the seaward part of the Vecht basin resulted in gradual inlet closure around ~3000 yr BP, which is relatively late compared to other inlets along the western coast. In the more landwards area a lake remained surrounded by peat lands.

The closing of the beach barriers of the southwestern to northwestern part of the Dutch coast largely ended the marine influence in the back-barrier area (Fig. 4: 5000–3500 yr BP). Silting-up of tidal channels and creeks led to a considerable decline of drainage and an increasingly freshwater environment in the back-barrier area (Pons, 1992; Vos, 2015). As a result, peat that was initially only present along the margins of the back-barrier area was able to expand into the former tidal areas. This large-scale peat formation was a self-enhancing process; the more peat was formed, the more the drainage decreased, leading to further peat growth (Vos, 2015). Moreover, as the peat raised mires, the input of siliciclastic sediment was reduced or excluded entirely. Around 5 ka BP large parts of the western coast, and around 3.5 ka BP the southwestern coast, of The Netherlands were closed and the hinterland covered with peat (Fig. 4: 5000–2500 yr BP) (Vos and Van Heeringen, 1997; Cleveringa, 2000). Notably, only where large rivers flowed out into the sea the tidal systems remained open, such as the estuary of the Meuse near Rotterdam and the estuary of the Rhine near Leiden, whereas tidal embayments without river input in the western Netherlands closed. During this period most tidal systems along

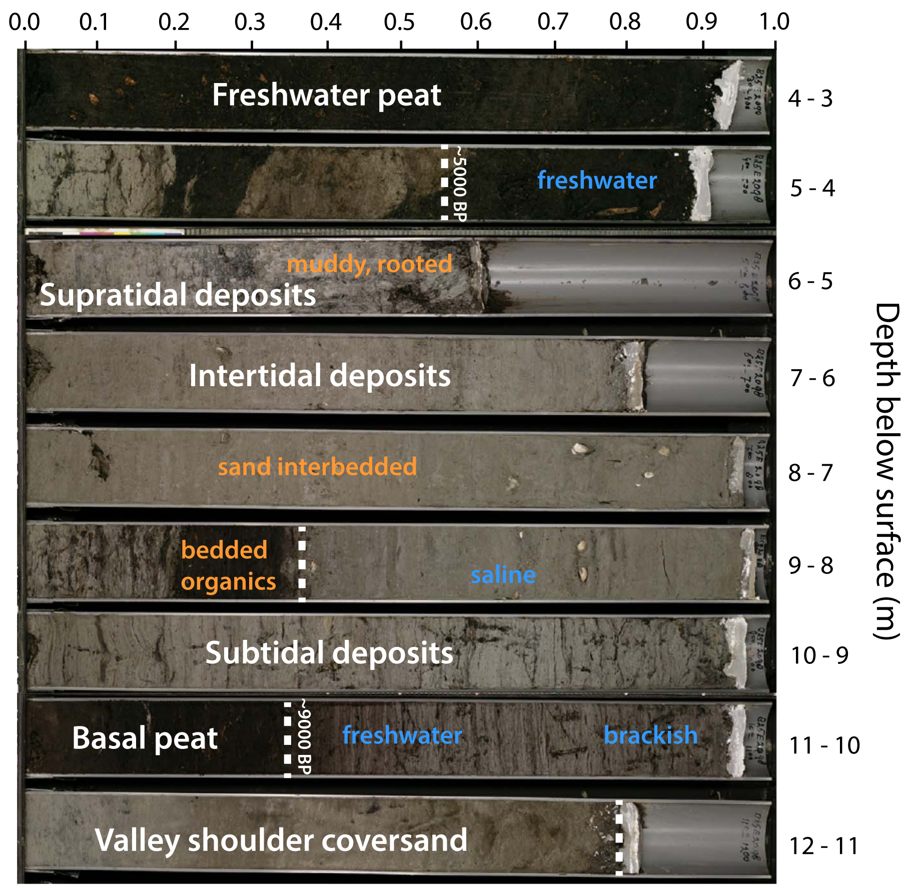


Fig. 7. Core photograph with a typical Holocene sedimentary succession of the backbarrier area in western Netherlands. The vertical succession is related to the spatio-temporal succession shown in 1a and 1b in Fig. 9 from 7500 to 3500 yr BP, at longshore distance of 130 km in Fig. 10. On top of sandy Pleistocene deposits, a basal peat layer formed due to groundwater level rise induced by sea-level rise. As sea level rose further and transgression continued the basal peat became covered by dominantly clayey subtidal deposits formed in a lagoonal environment. These deposits became covered by sandy to clayey deposits formed in a more energetic, brackish, environment. As the tidal embayments progressively shallowed the deposits became increasingly enriched in clay, bioturbated (intertidal flats) and rooted (supratidal flats) showing that the final stage of embayment filling depended on mud import. After filling of the backbarrier area and closure of the coastline, freshwater peat formed on top of the intertidal and supratidal deposits. Core surface at −1.25 m O.D., which is a few meters above the 3500 yr BP sea level.
Source: TNO - Geological Survey of The Netherlands; core B25E2098.

the western Dutch coastline had probably transformed from extensive embayments to more elongated and narrow estuaries. In contrast, the transgressive development of the coastline in the northern Netherlands had continued and the coast remained open. The tidal embayments of the Boorne, Hunze and Fivel rivers (see Fig. 4: Inherited landscape, for their location) reached their maximum extent around 2750 BP (Vos and Van Kesteren, 2000; Vos and Gerrets, 2005; Vos and Knol, 2015). The reason for the northern coastline remaining open is still debated. Possible explanations are a higher glacio-tectonic subsidence, less fluvial sediment input, less marine sediment input due to less tidal asymmetry and a deeper North Sea floor compared to the western Netherlands, the recent formation of the western part of the Wadden Sea after 3000 yr BP, the relatively large wave heights within the basin and on the coast due to the long fetch from the north-west or a any combination of those factors (e.g., van der Spek, 1994; Vos, 2015). The backbarrier basin that was present in west Netherlands was muddier than the present-day Wadden Sea, because of reclamation of the muddy supratidal areas of the Wadden Sea over the past few centuries (Flemming and Nyandwi, 1994). Therefore, these backbarrier systems probably hosted different species assemblages and density. As such, the role of species on the evolution of these systems may have differed.

The marine influence along the southwestern and western parts of the coastal plain was lowest between 5000 and 3000 yr BP. From closure of the coast until ~2000 yr BP the beach ridges and dunes in southwestern to northwestern Netherlands gradually expanded seaward forming a ~10 km wide beach barrier complex (Van der Valk, 1996a; Cleveringa, 2000; Pierik et al., 2017), and openings in this part of the coastline were the river mouths of the Rhine, Meuse, Scheldt and Oer-IJ and the final stages of the Bergen inlet (Fig. 4: 3500–1900 yr BP). The peat areas behind these coastal barriers expanded and enlarged further. In the northern Netherlands salt marshes and coastal freshwater peat environments both expanded seaward resulting in regression of the backbarrier area, similar to the coastal regression that occurred in the southwestern and western Netherlands one to two millennia before despite that the northern coastline remained open. The formation of extensive peat areas behind the coastal barriers provided a substrate that was prone to subsidence, which later resulted in Late Holocene human-induced coastal ingressions as described below.

2.4.2. Late Holocene human-induced ingression-formed tidal systems

After ~2000 yr BP the coastline of the southwestern Netherlands was being eroded and small tidal inlets formed in the barrier system (Vos and Van Heeringen, 1997; Vos, 2015; Pierik et al., 2017). Erosion replaced accretion around that time as a result of a subsiding backbarrier area that was compensated by reworking of coastal sediments, mainly sediments formerly trapped in beach barriers. The subsiding former backbarrier terrain was caused by oxidation and volume loss of the drained, widely present, peat lands, which caused land-surface lowering and triggered regular floodings. This human-induced creation of accommodation space, helped by feedback effects discussed below, was irreversible and led to large coastal ingressions in the southwestern and northern parts of the Dutch coast (Fig. 4: 1200 yr BP; Fig. 8e–h) (Vos and Van Heeringen, 1997; Vos and Knol, 2015; Pierik et al., 2017). Silting-up of the tidal embayments in the northern Netherlands continued and vast salt marshes developed.

Coastal ingression was enhanced by the following factors (Vos and Van Heeringen, 1997; Vos and Knol, 2015; Vos, 2015; Pierik et al., 2017): (1) man-made drainage channels and ditches enabled the sea to penetrate far into the peatland at high tide and storm surges; (2) tidal clay deposition on top of the peat as well as the formation of new tidal channels that enhanced peat drainage increasing peat compaction, subsidence and thus accommodation space; (3) peat that gets undercut and laterally eroded vanishes and thereby directly increases the tidal prism (because the bank material contains a high water fraction itself

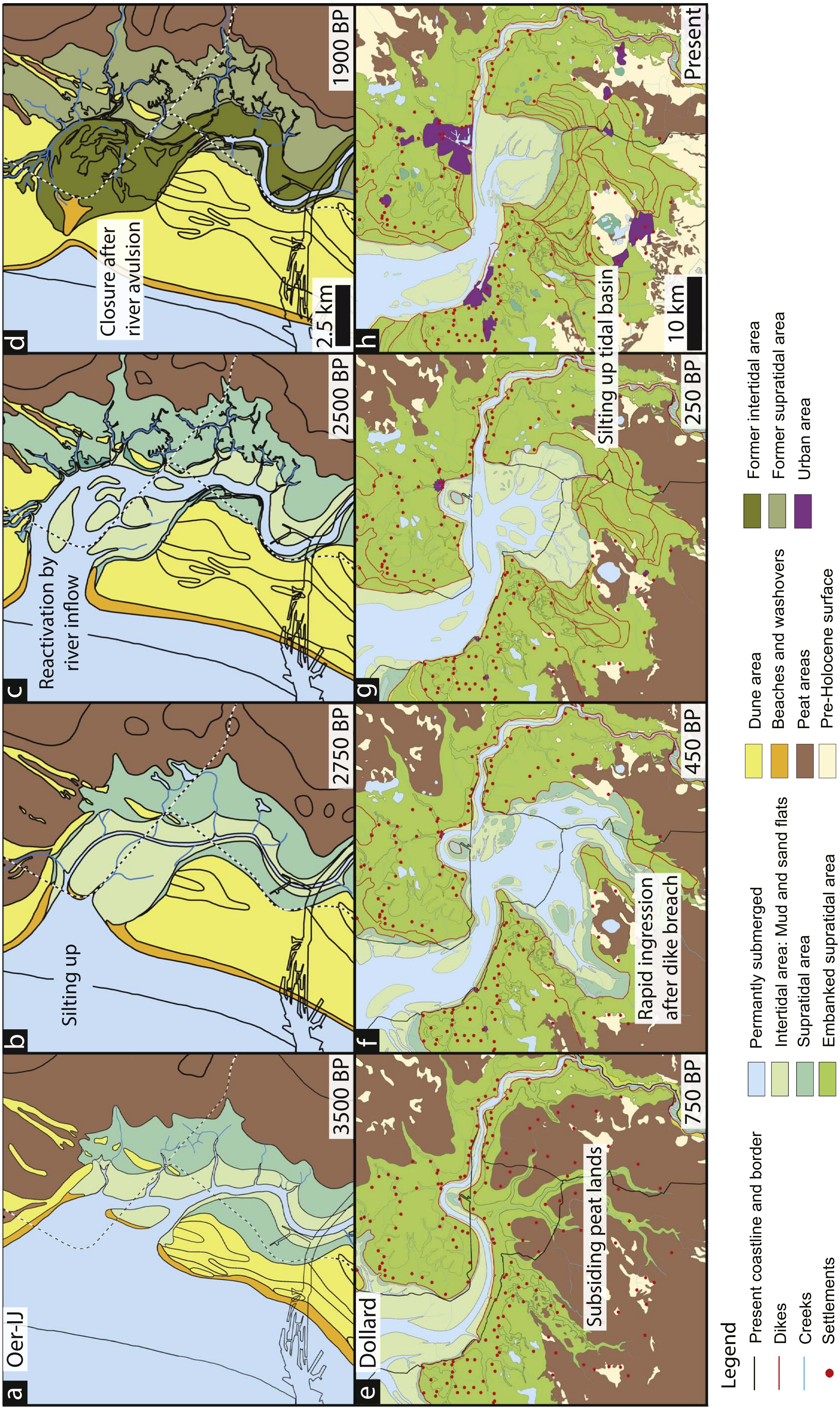


Fig. 8. Detailed palaeogeographic reconstructions of the Oer-IJ and Dollard systems. (a–d) The middle Holocene Oer-IJ estuary, which silted up and closed after upstream river avulsion (from Vos et al., 2015). (e–h) The Dollard ingression formed in the late Holocene, occupying subsided peat land (from Vos and Bungenstock, 2013).

and because organics decay when entering brackish estuarine waters); (4) an increase in accommodation space increases tidal prism, which led to enlargement of tidal channels both in cross-section and extent, further increasing accommodation space by peat erosion and compaction by clastic deposition on the peat and enhanced drainage. Through these positive feedback mechanisms the sea ingressions reached their maximum extent around 1200 BP after several centuries of inundation. Around this time the sediment supply to these systems surpassed the combined effects of subsidence and sea-level rise, partly because most of the peat had fully subsided (Pierik et al., 2017). Large parts of the ingressions in southwest and north Netherlands subsequently silted up to supratidal level, became habitable again, and were finally embanked and stabilized (Fig. 4: 1200 yr BP–present). No ingressions occurred in the western part of The Netherlands, predominantly as a result of the protective effect of the wide barrier complex that had developed here (Pierik et al., 2017).

In short, the Holocene Dutch estuaries and tidal embayments formed in two classes of initial conditions: (1) middle Holocene systems formed by sea-level driven transgression, which can be subdivided in (a) large, underfilled bay-like tidal systems constrained by pre-Holocene palaeovalley dimensions and (b) smaller systems that formed after silting up of the drowned paleovalley systems and barrier formation; and (2) ingressions mainly driven by peat-subsidence during marine highstand in the late Holocene (Figs. 9 and 10). The locations of the middle Holocene transgressional systems were probably largely determined by the topography of the inherited Pleistocene palaeovalleys. The location and size of late Holocene ingressive systems were determined by the formation of accommodation space by man-made and natural subsidence of peat areas, which indirectly depends on the evolution of the middle Holocene systems.

2.5. Contrasts in long-term evolution of Dutch estuaries and tidal embayments

The spatio-temporal positions of the main estuaries and tidal embayments along the Dutch coastline during the Holocene are summarized in Fig. 10. At the beginning of the transgressive phase, estuaries and tidal embayments formed in Pleistocene palaeovalley depressions (Fig. 6). These systems gradually silted up, a barrier coast developed, new estuaries and tidal embayments formed, and many tidal embayments eventually closed (e.g., Uitgeest inlet and Rijswijk-Zoetermeer inlet, Fig. 2c). However, estuaries with fluvial feeders filled up the accommodation space created by relative sea-level rise until they transformed into smaller throughflow systems in morphodynamic equilibrium, wherein fluvial sediment input was exported toward the North Sea (e.g. the Rhine and Eems Rivers, Fig. 10). By 3000 yr BP the remaining inlets were almost exclusively estuaries (e.g., Scheldt, Meuse, Rhine, Oer-IJ and Eems systems, Fig. 10). The systems formed by late Holocene sea ingressions were all silting up after initial enlargement (Figs. 8e–h, 9). Some systems have already completely closed and are embanked, such as the Middelzee (Griede, 1978; Knol, 1993; van der Spek, 1995; Vos and Knol, 2014; Vos, 2015), whereas others, such as the Western Scheldt, Oosterschelde and Grevelingen in the south, the Dollard in the north (and e.g., the Jade Busen in Germany), are still present and only partly silted up for lack of sediment or due to dredging activities (Vos and Zeiler, 2008; Wiersma et al., 2009; Vos and Knol, 2015; Pierik et al., 2017).

Upstream river avulsion frequently caused rivers to change their outflow location into the sea (Fig. 10). This conveniently provides us with a certain degree of control over the effects of fluvial feeders on tidal systems. River abandonment in all cases resulted in closure of the former estuaries, now tidal embayments, whereas avulsion into a former tidal embayment caused the new estuaries, former tidal embayments, to persist and remain open (Fig. 10). Examples are the avulsion of the Meuse around 4700 yr BP (Hijma et al., 2009; Cohen et al., 2012), and the avulsion of the Rhine that changed its course toward a new tidal inlet near Leiden by two successive avulsions following each other around ~7000 and ~6300 yr BP (Hijma et al., 2009; Hijma and Cohen, 2011; Cohen et al., 2012). This tidal inlet remained open for over 4000 yr and formed a large and narrow estuary near Leiden, until upstream river avulsion redirected the Rhine mouth south toward its former position, resulting in closure of the former estuary (Van der Valk, 1992; Vos, 2015). The Oer-IJ tidal embayment was nearly filled and closed around 3000 yr BP, but avulsion of a Rhine river branch into the Oer-IJ (Bos et al., 2009) kept the system open and enlarged it. Subsequently the system closed around 2000 yr BP when the fluvial input ceased due to upstream river avulsion, and formation of the Vlie tidal inlet that connected the Almere inland lake fed by upstream rivers with the North Sea (Figs. 2a and 8a–d) (Vos, 2015, 2008; Vos et al., 2015). The Hunze estuary persisted far into the late Holocene, although it shifted around 3000 yr BP (Fig. 10) (Roeleveld, 1974; Vos and Groenendijk, 2005). However, it rapidly closed off around 1500 yr BP following river avulsion toward the Lauwers tidal embayment that formed a few hundred years before by sea ingression (Fig. 2b).

To summarize, tidal inlets typically only remain open for long time periods given abundant sediment supply when there is a river flowing into the embayment (Fig. 11), despite the relatively small river discharge compared to tidal prism in many systems (e.g., $<1\%$ in Western Scheldt (Wang et al., 2002)). In deep and largely unfilled embayments there is net aggradation from sediment input from the sea and river in estuaries, and from the sea in tidal embayments. As such, both deep and largely unfilled estuaries and tidal embayments gradually silt up, resulting in the formation of intertidal and supratidal areas. Once an estuary has been filled up to near-equilibrium planform and hypsometry, a narrow throughflow system may develop in the presence of a river. In contrast, tidal embayments in the absence of a river are unstable environments that silt up over time until they close given abundant sediment supply. These tidal embayments are thus net sediment importing systems throughout their lifetime.

3. Discussion

3.1. Effects of geological constraints and inherited topography

Inherited pre-Holocene topography initially strongly determined the position, size and shape of the Middle Holocene estuaries and tidal embayments, essentially forming drowned bays in the former palaeovalleys. When these progressively filled in throughout the middle Holocene, an equilibrium situation developed either filling them in completely or keeping them open during sea-level highstand. The dimensions of the tidal systems present during highstand did not depend on the inherited topography, because of coastal aggradation above the pre-Holocene palaeovalleys crests, the strongly modified shape of the Dutch coastline by Holocene coastal processes (e.g., Beets et al., 2003) and because major changes occurred in the Rhine-Meuse delta river network (Berendsen and Stouthamer, 2000; Cohen et al., 2012). The Meuse and Eems systems are the only systems that have essentially retained their position since the start of the transgression, whereas many other river outlets have left their initial inherited position (e.g., Rhine, Scheldt, Hunze; Fig. 10).

Sea ingressions during sea-level highstand formed in areas where natural and human-induced subsidence of peatbogs created accommodation space. The actual occurrence of a sudden ingression thus became a disaster waiting to happen, and historical records indeed show that these ingressions were initiated during large storms (e.g., Gottschalk, 1975). Following initiation, many of these new estuaries and tidal embayments extended to their largest size within one to two centuries. The maximum size of the ingressions was limited by the distribution of peatlands and their reclamation history, which was, in turn, partly determined by Pleistocene topography (Vos, 2015; Pierik et al., 2017).

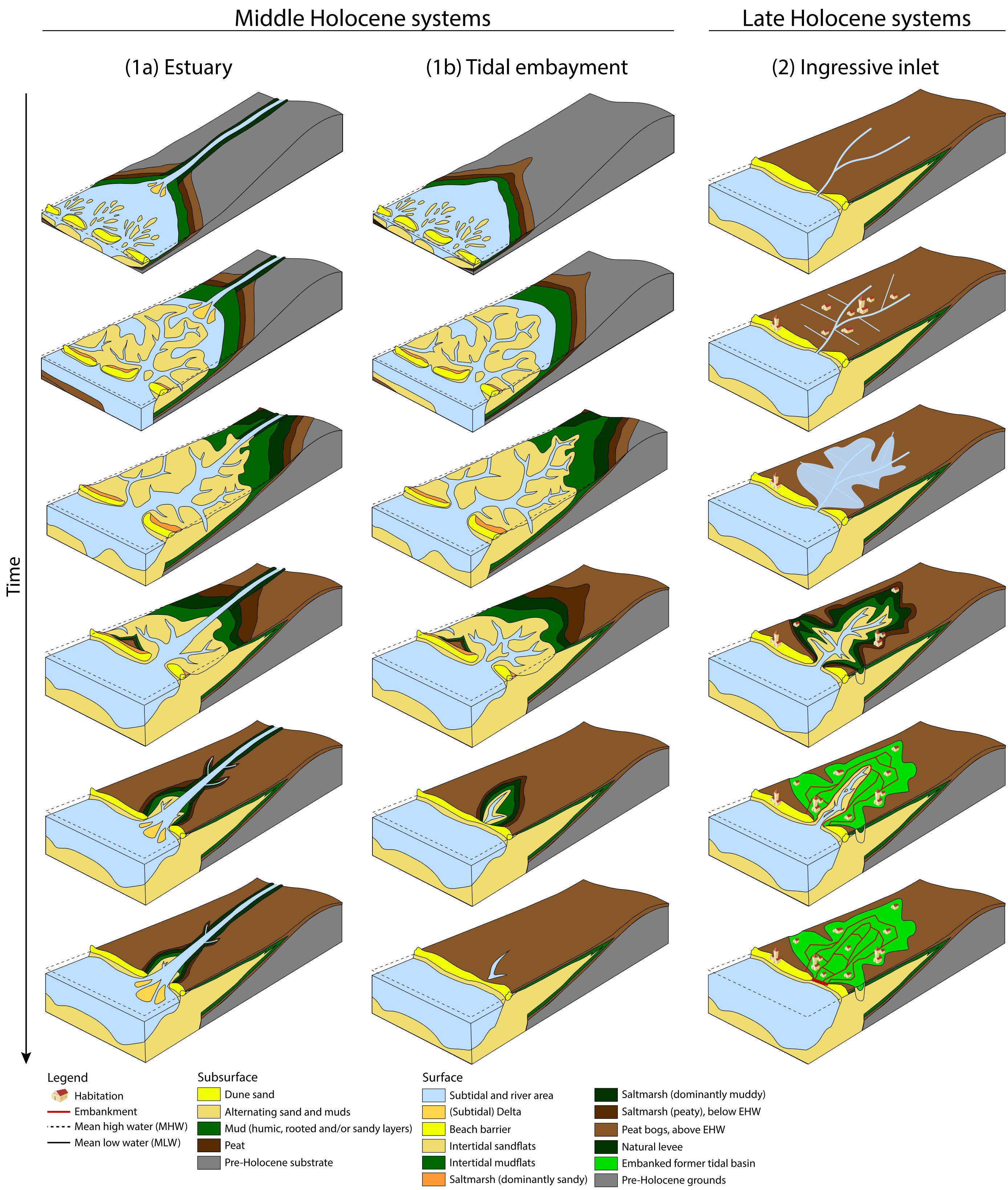


Fig. 9. Schematic block diagrams depicting the evolution of estuaries (1a) and tidal embayments (1b) formed during rapid sea-level rise in the middle Holocene (transgressional systems), and (2) Late Holocene sea ingressions mainly driven by natural and human-induced soil subsidence during marine highstand. Examples of 1a are loosely based on the Meuse (top) and Rhine estuaries (middle to bottom, at alongshore positions 80 and 120 km, see Fig. 5), 1b is loosely based on the tidal system in the Vecht palaeovalley (Bergen inlet at 180 km) and 2 is loosely based on the Western Scheldt (10 km), Middelzee (300 km) and Dollard (400 km) systems. The tidal embayment evolution is loosely based on Vos and Van Kesteren (2000). Late Holocene system evolution is loosely based on Pierik et al. (2017). Animation in online supplement.

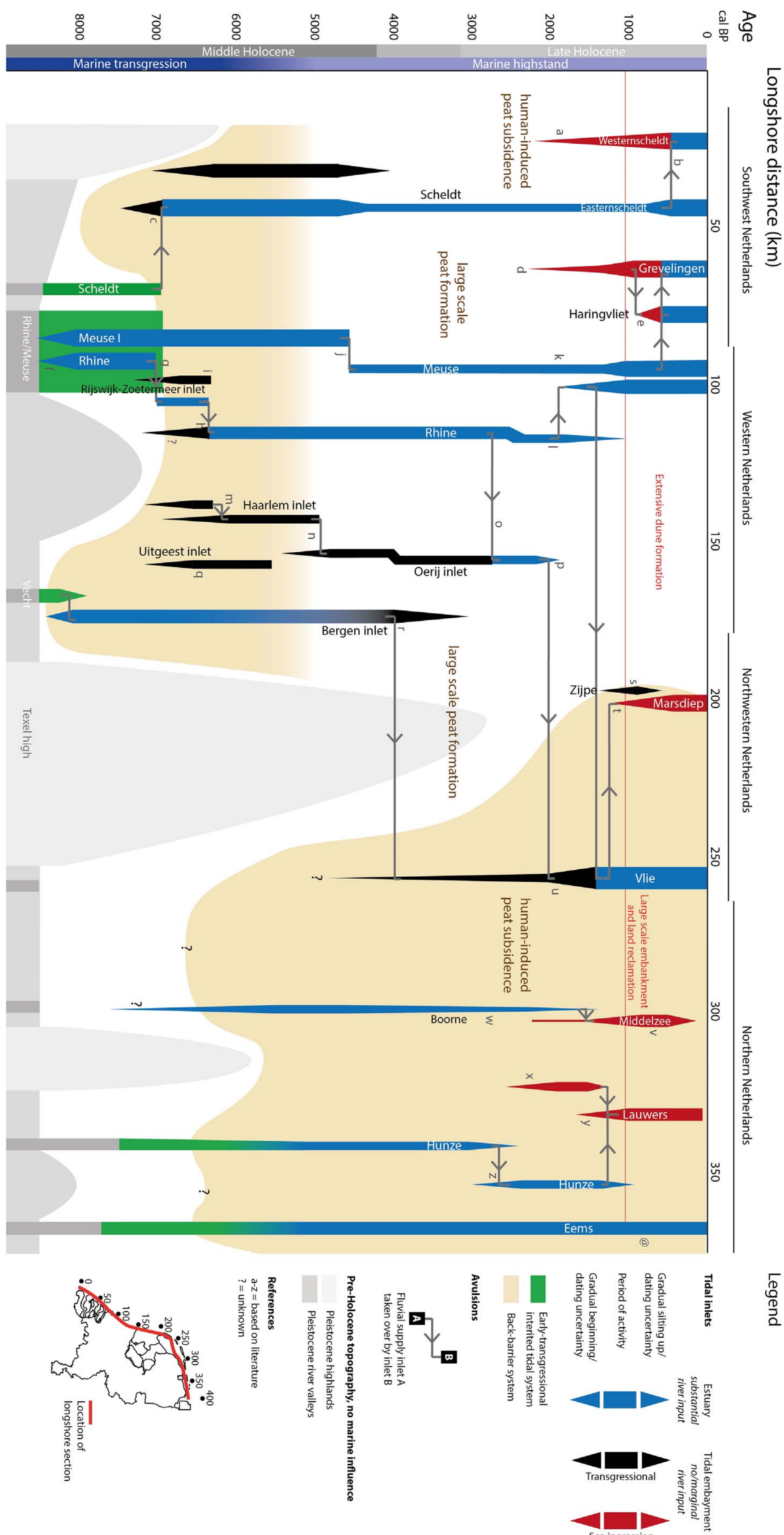


Fig. 10. Tidal system development in time and space, extended from Pierik et al. (2017). Map shows location of the cross-section along the coastline (see Fig. 5 for boundary conditions). In northern Netherlands the line is projected along the intertidal-supratidal transition to visualise back-barrier development. Arrows indicate upstream avulsion causing fluvial supply to shift toward a different tidal inlet. Sources and references given in Appendix A.

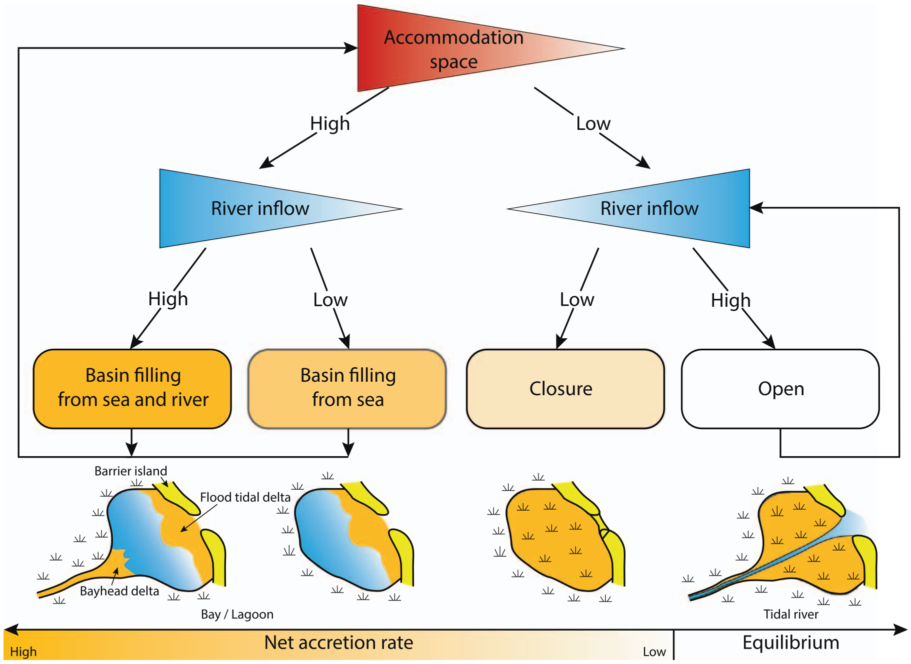


Fig. 11. Generalized development pathways of Holocene tidal systems along the Dutch coast with typical boundary conditions of sea-level rise leading to accommodation space, tidal conditions, wave climate and significant sediment supply from the sea. Long-term equilibrium, and thus persistence and stability of tidal inlets only occurred in systems with upstream freshwater inflow. When accommodation space is large and tidal embayments are relatively deep systems fill in both with and without river inflow, for which littoral sediment supply is a necessary condition. When infilling progresses and intertidal areas emerge accommodation space becomes low, after which a system will progressively silt up and close in absence of river inflow, or will develop into a relatively narrow estuary that persists when river inflow is substantial.

The availability of vast amounts of unconsolidated pre-Holocene sands and muds was important for the Holocene evolution of the Dutch coast. The formation of barrier islands and subsequent closure of the coast was possible by the availability of large amounts of sandy late Pleistocene and early Holocene river deposits on the shallow North Sea floor, as well as from erosion of capes of older Pleistocene deposits (Beets et al., 1992). On the other hand, the Wadden Sea tidal embayment has not been filled and closed, because of a limited sediment input compared to creation of accommodation space.

Owing to the relatively shallow Pleistocene valleys and complete drowning of the coast, the evolution of estuaries and tidal embayments along the Dutch coast was less strongly influenced by inherited topography than those of typical incised-valley systems. For example, along the North Sea basin the estuaries along the south-eastern coast of Great Britain are of a different geomorphological nature to almost all the ones on the Flemish-Dutch side (e.g., Devoy, 1979; Baeteman, 2005; Cohen et al., 2014). Many British estuaries transformed from flooded valley into bar-built estuary, and back, several times over multiple glacial-interglacial cycles. Sediments of older stages have been preserved as terraces along the valley margins (e.g., Roe and Preece, 2011). This is because there is no major subsidence, and because river avulsions are far less common in such incised valleys compared to broad low-sloping shelf settings such as that of the Dutch coastal plain. Thus, palaeoestuarine and palaeovalley situations alternated at positions fixed by inerodible valley walls over repeated eustatic transgression-cycles on the English side, whereas shallow and erodible valleys and self-formed entirely alluvial estuaries formed and moved with the coastline and with littoral drift at The Netherlands' side (Cohen et al., 2014, 2017). Due to the impossibility of upstream avulsion, incised valley estuaries are likely to receive fluvial input over long timescales and therefore are more likely to remain open even when ample mud is supplied and eco-engineering species are present.

The Holocene evolution of the Dutch tidal systems as reconstructed here points to the need for more realistic initial conditions in long-term morphological modeling. The earlier transgressive systems formed over long periods mainly by multiple sedimentary processes in the barrier, in the basin or floodplain and in the channel and its margins, while erosion of the inherited valley margins was limited; accommodation space was initially created by drowning of palaeovalleys. On the other hand, the Late-Holocene ingressive systems formed first and foremost by large-scale erosional processes following initiation, and later filled in by the aforementioned sedimentary processes. The sudden initiation of the ingressive-type systems elucidates a difference in characteristic time-scale of the erosional and sedimentary processes that is not understood yet. The first long-term models by Van der Wegen et al. (2008) proceeded from a small initial channel which allowed modeling of the system boundaries by a simple erosional process. In contrast, there are no models yet for estuaries shaped by net sedimentation, which is the natural situation for many estuaries and tidal embayments worldwide. Modeling tidal systems formed by net sedimentation is much more challenging because it requires incorporation of a number of biogeomorphological processes combined with mud dynamics. Much of the present-day understanding of tidal-system morphodynamics is based on currently present tidal systems, and this is often applied to older systems (e.g., the present-day Wadden Sea as analog to the formerly present back-barrier basins in the western and southern Netherlands). However, these systems are often fundamentally different, the middle Holocene back-barrier basin in the western Netherlands was for example richer in mud and its channels were laterally far less mobile compared to the present-day Wadden Sea. Coastal boundary conditions of former systems may thus be fundamentally different from those of present-day systems.

3.2. Long-term stability of estuaries and tidal embayments

The tendency of tidal embayments and estuaries to fill in by mud sedimentation and vegetation in their final filling stage raises the question whether long-term equilibrium of the large-scale planform shape and size is possible at all. The data show that along the Dutch coast only tidal systems with significant fluvial input continued to exist after sea-level rise ceased. Tidal embayments without or with a marginal fluvial inflow were unstable and eventually closed off (Figs. 9–11), while the estuaries have filled their accommodation space over time until a system formed that effectively bypassed sediment and attained dynamic equilibrium (a delta cf. the definition for coastal systems of Dalrymple et al., 1992). Examples of such systems are the Old Rhine, Meuse and Eems-Dollard estuaries. In contrast, systems that turned into tidal embayments after upstream avulsion of their feeding river generally closed within a few hundred years, although closure rates depend on the balance between sediment input rate and accommodation (Fig. 10). These observations suggest that a dynamic equilibrium can be obtained in estuaries whereas tidal embayments are unstable and thus ephemeral systems along the Dutch coast. This agrees with the finding that tidal embayments along a shallow sandy coast, such as the western coast of The Netherlands, cannot be stable since sediment from the coast will always be imported into the embayments if available (Dalrymple et al., 1992; van der Spek, 1994). As outlined in the introduction, these observations contradict findings in many numerical models (e.g., Lanzoni and Seminara, 2002; Todeschini et al., 2008; De Swart and Zimmerman, 2009; Hsu et al., 2013) and analytical models (e.g., Seminara et al., 2010; Toffolon and Lanzoni, 2010) that suggest that estuaries and tidal embayments can both obtain long-term morphodynamic equilibrium and persist over time (see Section 1.1.1). This contradiction may be attributed to the choice of seaward boundary condition for sediment availability. In the presence of enough sand and mud to build barrier systems that eventually define estuaries and tidal embayments, there is enough sediment to be imported into the embayments and this cannot be neglected in the model, unlike in other systems where sand is not, or no longer, available in the littoral zone. Moreover, a system may have sufficient wave energy relative to tidal energy (related to tidal prism), to construct a barrier that is robust enough to survive post-storm conditions and permanently close off a tidal system. This explanation is further supported by laboratory experiments that show a long-term equilibrium precisely because their seaward boundaries do not allow sediment import from the sea or barrier formation (e.g., Tambroni et al., 2005; Kleinhans et al., 2012, 2015). On top of this, biogeomorphological interactions and mud dynamics are an important, but often overlooked, factor in the long-term evolution of tidal systems (e.g., Lessa and Masselink, 1995).

So under which conditions can estuaries and tidal embayments obtain long-term dynamic equilibrium? In relatively large and deep, largely subtidal, embayments the hydrodynamic conditions lead to efficient trapping of sediment from the sea and rivers. As a result, such systems are typically net infilling until sufficient amounts of intertidal and supratidal areas emerge. The evolution of the intertidal and supratidal areas then determines whether tidal embayments remain open or close off, and is thus key to the long-term evolution of tidal systems. The long-term evolution of intertidal and supratidal areas depends, however, on a complex and interacting set of processes operating through time. The combined effects of these different processes and feedbacks either favoring tidal flat accretion or erosion are complicated and hard to predict (e.g., Pethick, 1980; Friedrichs, 2011), which also renders prediction of the long-term evolution of tidal embayments and estuaries difficult. The processes that are currently included in numerical and analytical models predict development toward dynamic equilibrium, but we hypothesize that implementation of sound biophysical descriptions of intertidal and supratidal flat dynamics, appropriate boundary conditions relating to the supply of sediment from both the river and the sea, and/or sufficient wave energy to create a barrier that closes the mouth of an embayment, may result in progressive basin closure.

Based on the above review, we propose a conceptual model portrayed as a causal loop diagram denoting the interaction between the most important processes and feedbacks controlling the evolution of tidal systems (Fig. 12). We will use this model in future work to guide both palaeogeographical interpretations of former estuaries and tidal embayments as well as design of numerical and laboratory models for long-term simulations. In a relatively large and deep tidal system, such as typical for many transgressive estuaries in the beginning of the middle Holocene, a flood-dominant tidal asymmetry develops on the continental shelf and/or within the embayment. This results in net import of sand and mud by tidal asymmetry and scour and lag effects, respectively (e.g., Dronkers, 2005). Moreover, estuarine circulation further promotes mud trapping in the estuary. As input of sand and mud continues the tidal system will progressively accrete and intertidal areas will eventually emerge within the system. The emergence of intertidal areas invokes a negative, balancing, feedback loop by decreasing the flood asymmetry and hence reducing sand import (e.g., Friedrichs and Aubrey, 1988; Friedrichs, 2010). Estuarine circulation is hampered in shallow estuaries, decreasing mud import from the sea. Nevertheless, mud import by scour and lag effects continues (e.g., Van den Berg et al., 1996) as well as by the fluvial feeder system if present, promoting the growth and expansion of intertidal and supratidal areas. The growth of intertidal areas is further promoted by the colonization of intertidal areas by vegetation and potentially other sediment stabilizing species such as reef builders and diatoms depending on the salinity gradient (e.g., Gleason et al., 1979; Allen, 1989; Morris et al., 2002; Yang et al., 2008; Li and Yang, 2009). This is done by enhancing trapping efficiency of mud, increasing resistance to erosion and damping wave energy (e.g., Yang et al., 2008; Fagherazzi et al., 2012). As intertidal areas grow and expand parts may transform into supratidal areas, which grow further and stabilize due to eco-engineering species. The formation of supratidal areas partially restores flood asymmetry in the embayment (e.g., Van den Berg et al., 1996). This is because it results in a larger mean water depth in the embayment during flood, thereby enhancing flood-dominated asymmetry and allowing sand import until a new equilibrium is established between intertidal area and tidal asymmetry. This feedback loop, which extends the loop of Van den Berg et al. (1996) with effects of eco-engineering species, may eventually result in closure of tidal embayments as observed along the Dutch coast. This entails a reinterpretation of the van der Spek (1994) conjecture: closure is an effect of mud and eco-engineering species and the availability of sand is another necessary condition rather than the main one.

However, the diagram (Fig. 12) also shows that there are conditions under which tidal embayments can remain open over long timescales. For example, as long as the formation of accommodation space exceeds filling a system will remain open. Moreover, if waves in tidal embayments are sufficiently strong to inhibit mud accretion on intertidal areas expansion and growth of intertidal areas is hampered (e.g., Ridderinkhof et al., 2000; Adlam, 2014; Friedrichs, 2011). As a result, the growth of vegetation, the main stabilizer, is suppressed and intertidal areas are unable to grow into supratidal areas. The intertidal areas and hydrodynamics, governing sand transport in the basin, then remain in dynamic equilibrium and the tidal embayment remains open. It is, however, not yet clear under which conditions this situation can arise but clearly it requires sufficient fetch to develop significant waves (e.g. Fagherazzi and Wiberg, 2009; Sassi et al., 2015), suggesting that this may be valid for larger tidal embayments but not for smaller ones, which implies that a threshold may exist below which such systems 'catastrophically' shift to another stable state, namely closure. Another important effect of embayment size is that it determines tidal prism at the mouth of the system: a small system with a small tidal flux at the mouth will thus be relatively easily closed by barrier growth, and vice versa.

Alternatively, effects of initial and offshore boundary conditions may keep a tidal embayment open over long timescales. As long as there is ample sediment available for transport on the continental shelf or from river input, sand and mud accretion may continue within the tidal embayment. Vice versa, when sediment is scarce it may limit sand and mud accretion so that intertidal and supratidal areas are unable to grow and the embayment remains open. Additionally, tidal embayments remain open when the creation of accommodation space (by sea-level rise and subsidence) exceeds or equals sediment input. However, these situations are entirely determined by the boundary conditions rather than internal processes such as local wave generation. Finally, the ratio between embayment size, which determines tidal prism, and offshore wave conditions, which determine the strength of littoral drift, regulates the point at which the mouth of an embayment is closed by littoral drift. Again this leads to the general question under which conditions long-term development and equilibrium is mostly driven by boundary conditions or by internal processes and feedbacks.

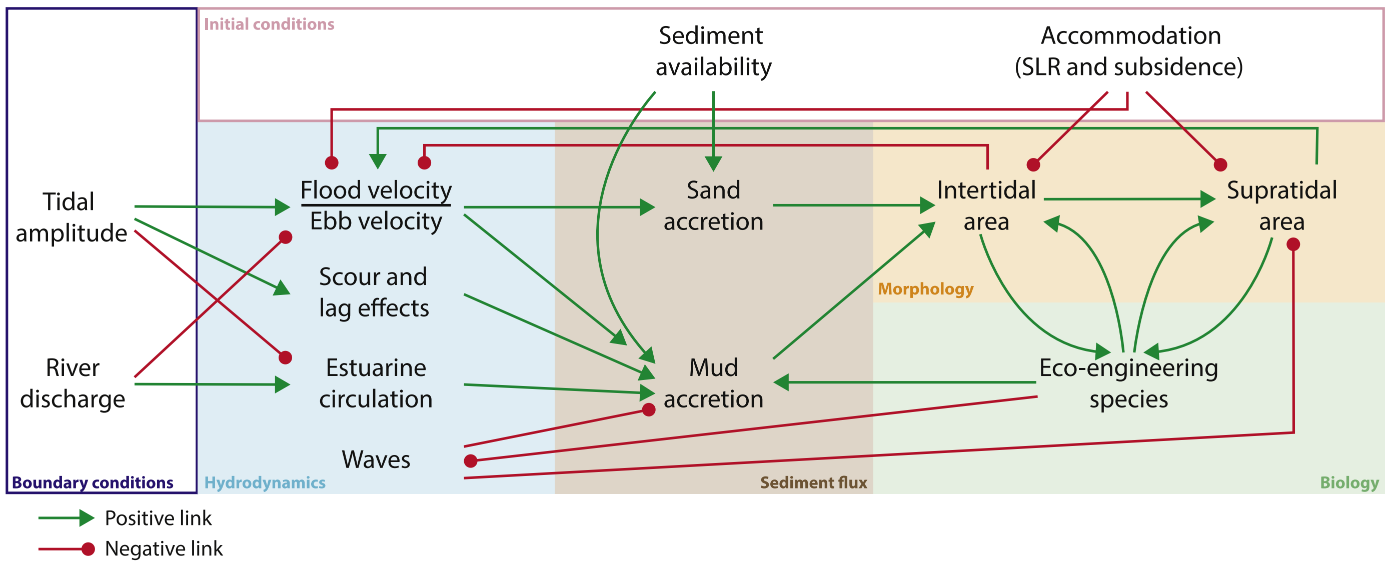


Fig. 12. Causal loops of main processes and feedbacks determining the long-term evolution of estuaries and tidal embayments. See text for possible paths of evolution in the absence and presence of river discharge and of eco-engineering species.

3.3. Effects of river discharge and sediment supply

The above described processes and feedbacks operate in both tidal embayments and estuaries. Additionally, for estuaries the dynamics and evolution are influenced by the presence of rivers. As depicted in Fig. 12, river inflow enhances ebb currents, which may balance potential sediment import from flood-dominated tidal asymmetry and lag effects. This hampers the growth of intertidal and supratidal areas, keeping the system open. Here, effects of river inflow and sediment supply on tidal systems are described in more detail.

The data described here shows that along the Dutch coast, fluvial inflow is needed to retain a stable and open inlet over multiple centuries to millennia. Similarly, Heap et al. (2004) show that Australian estuaries obtain a relatively stable geomorphology once the central basin of an estuary is largely filled and a river has established a direct connection with the ocean. The tendency of large rivers to split into multiple branches in deltaic areas, such as the Rhine in the Dutch coastal plain, helps to retain multiple open tidal systems at the river mouths for relatively long periods, at least as long as each fluvial branch remains active. We attribute the necessity of fluvial input for long-term tidal system survival to river discharge enhancing ebb-directed residual currents, thereby counteracting the net import of coastal sediment so that the riverine sediment input is effectively flushed seaward. For example, river discharge may lead to ebb-dominated sediment transport despite a flood-dominated asymmetry in the tide (e.g., Brown and Davies, 2010; Guo et al., 2014). This agrees well with one-dimensional idealized models that predict the formation of a final equilibrium configuration wherein fluvial sediment is flushed through the estuary toward open sea (e.g., Guo et al., 2014; Canestrelli et al., 2014; Bolla Pittaluga et al., 2015). However, these models neglect the presence of intertidal and supratidal areas, scour and setting lag effects, wave effects and biological effects. Moreover, they assume uniform river discharge thereby neglecting temporal variations in fluvial water and sediment discharge, as well as neglecting temporal variations in the tidal amplitude. Furthermore, river discharge and associated river floods are known to cause critical short- to medium-term (i.e., year-to-decade) estuarine morphological changes (Cooper, 2002, 1993; Karunarathna, 2011). Likewise, Dalrymple et al. (2015) show that conditions within an estuary can alternate between river-dominated and tide-dominated on a seasonal basis as a result of seasonally varying input of fluvial input.

Some authors suggested that there is a critical discharge below which the influence of river discharge on the estuarine hydrodynamics and morphodynamics can be neglected (e.g., Powell et al., 2006; Sassi et al., 2012; Guo et al., 2014). Powell et al. (2006) suggest that fluvial effects can be neglected when the ratio between tidal mean discharge and river discharge exceeds a value of ~20, although this value likely varies between estuaries (Sassi et al., 2012). However, for estuarine evolution over long timescales the presence of a threshold value ignores the law of flow continuity: river discharge of any magnitude needs to be evacuated into the sea except in the unlikely cases in temperate climates where this flux is countered by groundwater escape or evaporation. As such, estuaries with only a very small fluvial contribution (ratio between tidal mean discharge and river discharge > 20) may remain open over time, at the very least with a mouth large enough to evacuate the river discharge. They may initially silt up and change their planform until river discharge is non-negligible, thereby transforming into an equilibrium system that effectively bypasses sediment. This is illustrated by for example the Hunze system (Figs. 2b, 10), which is fed by a very small river that dewaters peatlands in the hinterland. Despite the minor fluvial component in the flow, the Hunze estuary remained open until the upstream river avulsed and fluvial input ceased.

The above discussion raises the question whether simplistic boundary conditions such as used in idealized models are useful for predicting the evolution of tidal systems over their lifetime. The relatively good correspondence between predicted and observed long-term equilibrium of balanced bed sediment input and output in idealized one-dimensional models suggests that qualitatively variations in boundary conditions are of minor importance for the general, long-term, evolution of estuaries. This works as long as the boundaries of the estuary can be considered fixed. However, even under constant sea level, frequent floods are a necessary factor in building up levees and floodplains and infrequent high-magnitude storm events or extreme spring tide may flatten shoals in wider systems such as the Wadden Sea, curb the growth of salt marsh and close or open inlets. It is yet unclear which aspects of large-scale estuarine morphology and planform shape depend on what range of the frequency-magnitude spectrum.

4. Conclusions

We reviewed (1) the main processes that determine the long-term (centennial to millennial timescale) evolution of estuaries and tidal embayments, (2) whether and how estuaries and tidal embayments are able to persist over time, and (3) the effects of river discharge on the long-term evolution of estuaries. To this end we employed published results of idealized numerical models and the vast amount of data on Holocene deposits of estuaries and tidal embayments in The Netherlands to compare the long-term evolution of estuaries and tidal embayments. The forged connections between geological, physical and biogeomorphological literature led to a number of unanswered questions and novel insight regarding the possibility of large-scale equilibrium size and shape of estuaries and tidal embayments in relation to fluvial and coastal boundary conditions, inherited landscape conditions and effects of eco-engineering species.

Many estuaries with substantial fluvial input and tidal embayments with no to marginal fluvial input have formed and closed off or persisted along the coast of The Netherlands. In general, these estuaries and tidal embayments were formed during (1) the middle Holocene driven by rapid relative sea-level rise, these systems can be subdivided in (a) large tidal systems somewhat constrained by pre-Holocene palaeovalley dimensions and (b) smaller systems that formed after silting up of these paleovalleys and the formation of a barrier coast and (2) the late Holocene driven by natural and human-induced soil subsidence in the peat bogs.

Along the Dutch coast, tidal inlets of estuaries persisted for major parts of the Holocene owing to the river connection (thousands to tens of thousands of years). In contrast, tidal embayments with marginal or no fluvial inflow were ever importing sediment from the sea and ultimately closed off. Tidal embayments are thus unstable, temporarily existing, systems that gradually silt up on timescales of centuries to millennia. Along the Dutch coast, net sediment import from the sea into tidal embayments and estuaries is favored by strong, flood-dominated, tidal asymmetry, the shallow sand-rich floor of the North Sea and the abundance of mud in the coastal area supplied by the Rhine and Meuse rivers.

These observations contrast results of numerical models, analytical models and laboratory experiments that suggest that tidal embayments can obtain a long-term equilibrium configuration and remain open. This is probably mainly because these models generally ignore that the long-term evolution of most tidal embayments is also controlled by the dynamics of muddy sediment and biological processes: abundance of mud and vegetation often culminates in continuous fine sediment import and growth of intertidal and supratidal areas.

This paper qualitatively demonstrates how geological reconstructions of the long-term evolution of tidal systems help to unravel how internal processes and externally imposed conditions determine tidal system evolution. Such reconstructions may further be used to guide comprehensive model design as well as for model validation and calibration for quantification of the relative importance of internal and external factors. In future modeling of long-term evolution of estuaries and tidal embayments, sensitivity to changing boundary conditions and inherited estuary or tidal embayment configuration needs to be determined. Moreover, we show that the effects of mud and vegetation are key to the long-term evolution of tidal systems, and that these need to be taken into account for successful numerical and landscape experiments of the long-term evolution of estuaries and tidal embayments.

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Appendix A. Fig. 10 sources

(a) Youngest transgression SW Netherlands (Bennema and Van der Meer, 1952; Vos and Van Heeringen, 1997);
(b) Avulsion Easternscheldt to Westernscheldt (van der Spek, 1997; Vos and Van Heeringen, 1997);
(c) Avulsion Scheldt river (Hijma and Cohen, 2011; Vos et al., 2011; Hijma et al., 2012);
(d) Initiation Grevelingen (Vos and Zeiler, 2008);
(e) Initiation Haringvliet (Vos and Zeiler, 2008);
(f) Formation Holocene Rhine and Meuse channel belts (De Gans and De Groot, 1996; Hijma et al., 2009; Hijma and Cohen, 2011; Cohen et al., 2012);
(g) Northward avulsion Rhine estuary I (Hijma et al., 2009; Hijma and Cohen, 2011; Cohen et al., 2012);
(h) Northward avulsion Rhine estuary II (Hijma et al., 2009; Hijma and Cohen, 2011; Cohen et al., 2012);
(i) Rijswijk-Zoetermeer tidal embayment (Van der Valk, 1992; Cleveringa, 2000; Hijma et al., 2009; Hijma and Cohen, 2011);
(j) Northward avulsion Meuse estuary (Hijma et al., 2009; Cohen et al., 2012);
(k) Extension tidal embayment around the Meuse estuary (Van Staalduinen, 1979; Van Trierum, 1986; Vos and Zeiler, 2008);
(l) Rhine estuary (Van Dinter, 2013);
(m) Haarlem inlet (Van der Valk, 1996a,b)
(n) Initiation Oer-IJ (Zagwijn, 1971; De Mulder and Bosch, 1982; Westerhoff et al., 1987; Van der Valk, 1996a,b; Beets et al., 2003; Vos et al., 2015);
(o) Avulsion Rhine branch into Oer-IJ (Bos et al., 2009; Cohen et al., 2012);
(p) Silting up of the Oer-IJ estuary (Vos, 2008; Vos, 2015; Vos et al., 2015);
(q) Uitgeest inlet (De Mulder and Bosch, 1982; Westerhoff et al., 1987; Donselaar and Geel, 2007);
(r) Bergen inlet (De Mulder and Bosch, 1982; Westerhoff et al., 1987; Roep and van Regteren Altena, 1988; van der Spek and Beets, 1992; Beets et al., 1996; Van Zijverden, 2013; van Zijverden, 2017);
(s) Zijpe inlet (Schoorl, 1999);
(t) Marsdiep (Ente et al., 1986; Schoorl, 1999);
(u) Vlie (Ente et al., 1986; Schoorl, 1999);
(v) Middelzee (Cnossen, 1958; Ter Wee, 1976; De Groot et al., 1987; Knol, 1993; van der Spek, 1995; Vos and Baardman, 1999; Vos and Gerrets, 2005);
(w) Boorne (Cnossen, 1958; Ter Wee, 1976; De Groot et al., 1987; Knol, 1993; van der Spek, 1995; Vos and Baardman, 1999; Vos and Gerrets, 2005);
(x) Paessens (Griede, 1978; Knol, 1993; Vos and Knol, 2014);
(y) Lauwerszee (Roeleveld, 1974; Griede, 1978; Knol, 1993; Oost and De Boer, 1994; Oost, 1995; Groenendijk and Vos, 2002; Vos and Groenendijk, 2005);
(z) Hunze (Roeleveld, 1974; Groenendijk and Vos, 2002; Vos and Groenendijk, 2005);
(@) Eems/Dollard (Roeleveld, 1974; Homeier, 1977; Vos and Bungenstock, 2013; Vos and Knol, 2015; Vos, 2015).

Appendix B. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.earscirev.2017.10.006.

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