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The Holocene evolution of the barrier and the back-barrier basins of Belgium and the Netherlands as a function of late Weichselian morphology, relative sea-level rise and sediment supply

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Abstract

Flooding of the southern part of the North Sea occurred between 9000 and 8000 BP, when the rate of relative sea-level rise was on the order of 0.7 cm per year for the Dover Strait Region and 1.6 cm per year for the area north of the Frisian Islands, forcing the shoreline to recede rapidly. When relative sea-level rise decelerated after 7000 BP for the Belgian coast and 6000 BP for the central Netherlands coast, sediment supply by the tidal currents balanced the creation of accommodation space in the estuaries and other back-barrier basins. Consequently, the barrier started to stabilize, and the tidal basins and their inlets silted up. Between 5500 and 4500 BP, the Belgian coastal plain changed into a freshwater marsh with peat accumulation, and the same happened 500-1000 years later in the western provinces of the Netherlands. The E-W running barrier/back-barrier system of the Frisian Islands in the northern Netherlands stayed open until today, however, because of lower sediment supply.

The period between 4000 and 2000 BP was relatively quiet due to the strong deceleration of the rate of sea-level rise; peat cushions developed behind the barriers, which were straightened by erosion of the headlands. Major and often catastrophic flooding occurred in the Middle Ages, when the estuaries in the southwestern part of the Netherlands formed.

About 226 (± 15%) × 10^9 m^3 sediment, mostly sand, is stored in the barriers and back-barrier basins of the Netherlands, 70% of which was deposited prior to 5000 BP. About 10% of the stored sediment is estimated to be of alluvial origin. Most of the sediment is derived by the erosion of the Pleistocene basement during recession of the barriers, but tide-induced cross-shore transport from the North Sea forms an additional source for the barriers and back-barriers of the west-facing coast of the Netherlands.

Keywords: coastal plain development, estuaries, Rhine, sediment budget, tidal basins

Introduction

The coastal plains of Belgium and the Netherlands (Figs. 1, 2B) form the southern part of a Holocene barrier/back-barrier system along the southeastern and eastern coast of the North Sea (Fig. 1), bounded by the cliff coast of northern France in the south and the tip of Denmark in the north. At present, a large part of the back-barrier system consists of heavily populated, reclaimed and cultivated tidal flats, estuaries and marshes, situated at or below sea level and protected from flooding by man-made coastal structures and by dunes. Maintenance of this coastal defense remains one of the important Dutch issues. A joint, multidisciplinary project on large-scale and long-term development of the Dutch coast was started for this reason in 1986. It was carried out by a large group of scientists from universities, research institutes and government institutions. Many publications resulted, ranging in scope from the detailed mechanics of sediment-transport processes to the Holocene development of the coast. The present contribution briefly summarizes our present knowledge on the long-term development of the system. We will discuss the morphology of the coastal plain, the hydrodynamics, morphology and sediments of the Southern Bight (Fig. 1), the (variations in) relative sea-level (RSL) rise in the area, and the Holocene succession and his-

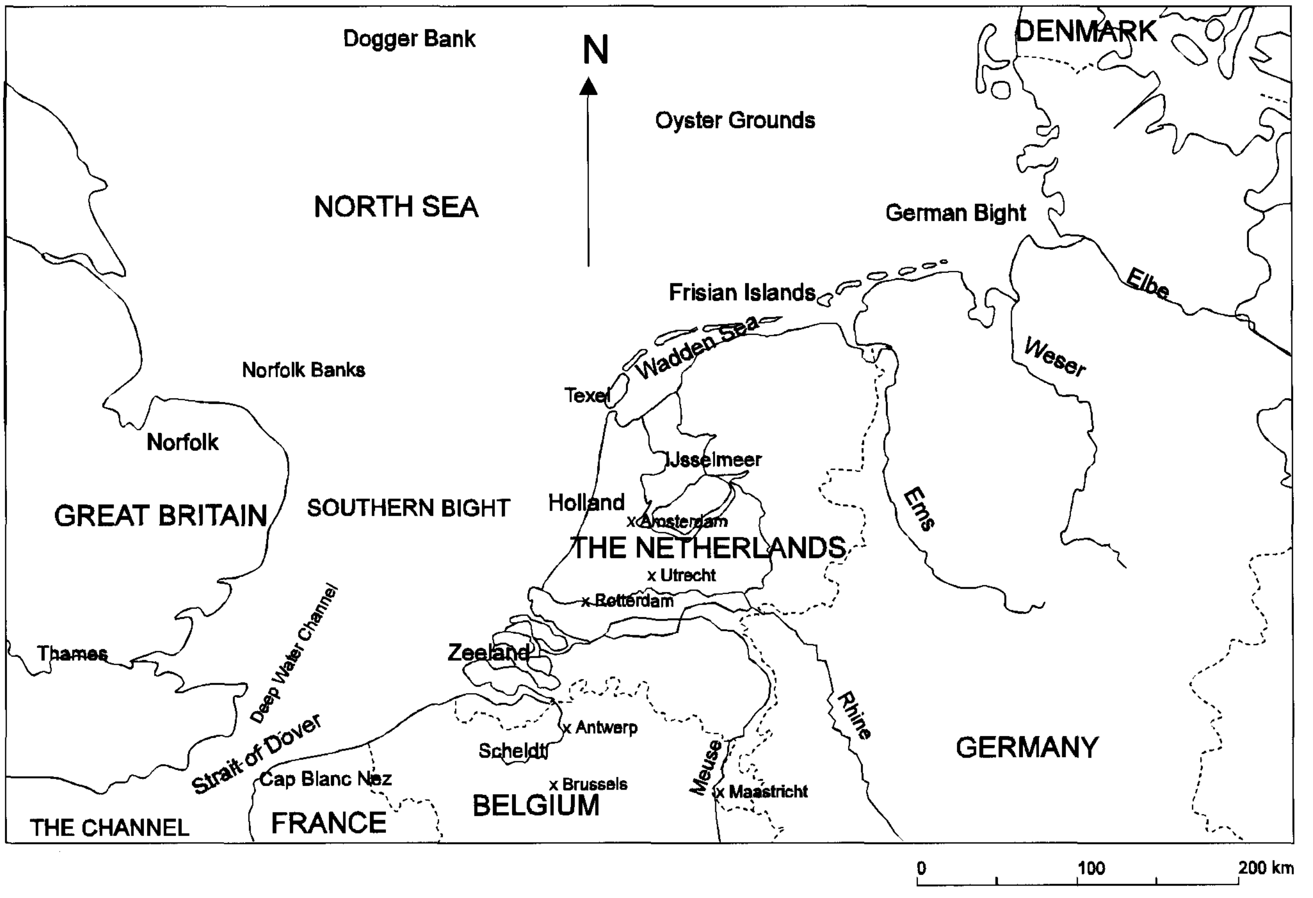

Fig. 1. Map of the study area with the locations of the geographical names used in the text.

tory of the coastal plain. Finally, we present a new, hypothetical model of sediment transport and accumulation in the Southern Bight and its effects on the coastal plain under study.

Morphology of the coastal plain

The Flemish and Dutch coast between Cap Blanc Nez (Fig. 1) in the south and the island of Texel (Fig. 1) in the north has a general SW-NE direction, concave towards the North Sea. North of Texel, the coastline turns to almost E-W. Although the present morphology is strongly modified by tides and waves, first-order features – such as this change in orientation of the coastline – are completely defined by the Pleistocene morphology. The coastal plain consists of a wedge-shaped succession of estuarine deposits, mainly medium- to fine-grained sand and mud, but also some peat. At the present coastline, this wedge varies in thickness from 15 to more than 30 m. The coastal plain reaches its greatest width, about 100 km, in the western Netherlands. Its thickness and width are defined by the westward slope of the pre-transgressional surface and the position of former channels. Slope gradients vary between 1:500 and 1:1000 for the base of the Belgian coastal plain and between 1:2500 and 1:4000 for that of the Netherlands (Fig. 2A).

The coastline between northern France and the southwestern Netherlands is closed and consists of a narrow barrier (Fig. 2B). Coastline erosion is prominent in the north near to the estuary of the river Scheldt (Fig. 1). Large estuaries of the rivers Rhine (Fig. 1), Meuse (Fig. 1) and Scheldt form the coastal system of the southwestern Netherlands. The inlets of these estuaries are up to 30 m deep. After the catastrophic floods of 1953, most of the inlets have been closed by dikes or sluices; others are protected during extremely high water by mobile storm shields. Barriers at the heads of the islands and peninsulas are narrow in comparison to that of the western coastal area (Holland) of the Netherlands (Figs. 1 and 2), and coastal erosion is prominent. The Holland' coastline is closed over a distance of about 120 km. Except for the southern- and northernmost part of this coastline, the barrier consists of an up to 10-km wide zone of prograded ridges and swales overlain by dunes. The barrier protects a wide coastal plain. The coast is – and has been – eroding significantly in the north and in the south, but is more or less stable in its central part. Eroded sand in the south is transported northward by longshore drift, whereas that in the north dis-

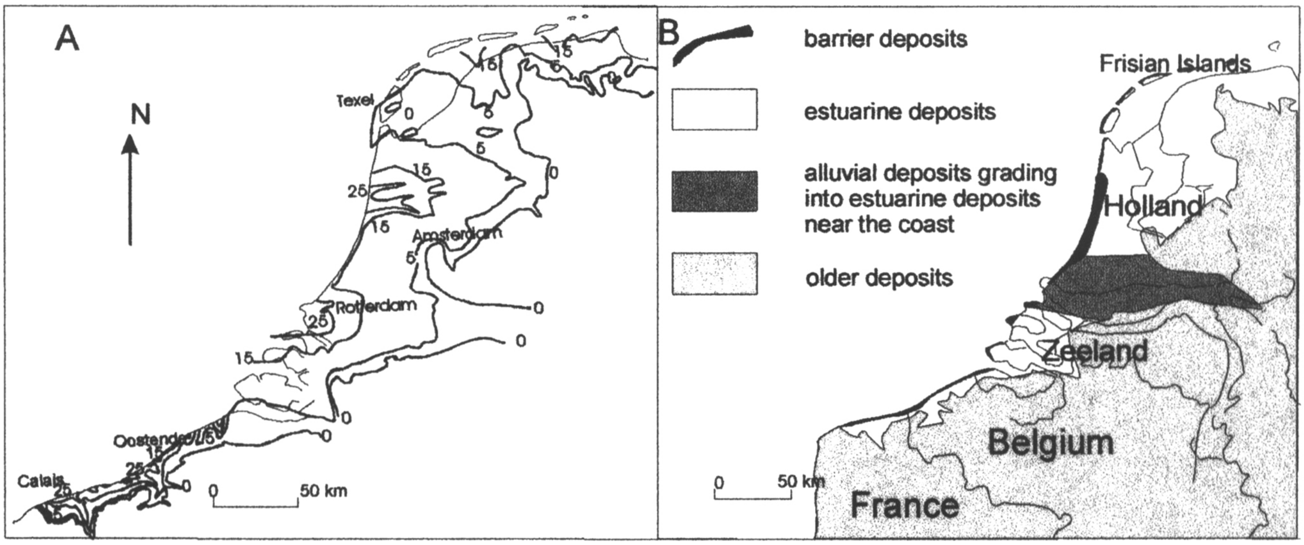

Fig. 2. Geological context.
A: depth to the base of the Holocene deposits (in meters below mean sea level) in the coastal plains of Northern France, Belgium and the Netherlands; after Houthuys et al. (1993) and De Gans & Van Gijssel (1996).
B: main facies associations in the coastal plain of the Netherlands, Belgium and northern France.

appears by way of the inlet between Holland and the island of Texel (Fig. 1) into the Wadden Sea. The Frisian Islands (Fig. 1), separated by tidal inlets 20-30 m deep, form the barrier of the Wadden Sea, a tidal basin separated by dikes from reclaimed land in the south. Overall, the North Sea coastline of the Frisian Islands is receding, but small-scale progradation occurs, although locally and temporarily (Oost & De Boer, 1994; Oost, 1995). Each barrier island along the Wadden Sea consists of dunes on top of sandy shoal and channel deposits (Van der Spek, 1996).

The Southern Bight

The Southern Bight is the shallow, funnel-shaped inland sea enclosed between Great Britain to the west and Belgium and the Netherlands to the east (Fig. 1). It is situated in the temperate climatic zone, characterized by prevailing, commonly strong westerly winds. As shown by Van Straaten (1961), wind and wave climate of the bight are closely correlated. The significant wave height for the central part of the Holland coast is 1.3 m (Kroon, 1990). It increases northward to 1.8 m near the island of Texel (Sha, 1989) and decreases towards the south.

The tidal wave enters the Southern Bight both from the north and from the south (Fig. 3A). The tidal wave from the north splits north-east of Norfolk in a branch moving eastwards towards the Frisian Islands and one moving into the Southern Bight, where it rotates anticlockwise around an amphidromic point between Holland and Norfolk (Fig. 1). It joins in phase the tidal wave coming in through the Strait of Dover from the south, which has a large tidal range but, because of the narrow entrance, consists of a relatively small mass. From Cap Blanc Nez, the tidal range falls rapidly northward to less than 2 m along the southern part of the Dutch coast, and increases again along the northern part. North of the Frisian Islands, the southern and eastern branch of the tidal wave rejoin. The tidal wave in the study area is asymmetric with a shorter flood period; consequently, peak current velocities during flood are higher than during ebb (Dronkers, 1986). The resulting, small residual current causes northward sediment transport along the studied coast (McCave, 1971; Dronkers, 1986). The present northward-directed sand transport at a depth of –20 m along the coast of the Netherlands is estimated to be 10-40 $m^3 \cdot m^{-1} \cdot a^{-1}$, based on process- and behaviour-related model calculations (Van Rijn, 1995).

In general, tidal currents are strong along the British coast, in the narrow Southern Bight, and along the Frisian Islands, including those in the German Bight. Except for the 'Deep Water Channel' (Fig. 1), an up to 60 m deep depression running from the Norfolk to the Strait of Dover Banks (Figs. 1, 3), the Southern Bight has a water depth of less than 40 m, and is characterised by groups of linear banks or ridges parallel or at a small angle to the main axis of the tidal-current ellipse (Houbolt, 1968; Kenyon et al,. 1981; Johnson et al., 1982; Berné et al., 1994; Collins et al., 1995), and complex fields of superposed subaqueous dunes perpendicular to that axis (McCave, 1971). The ridges may rise more than 30 m

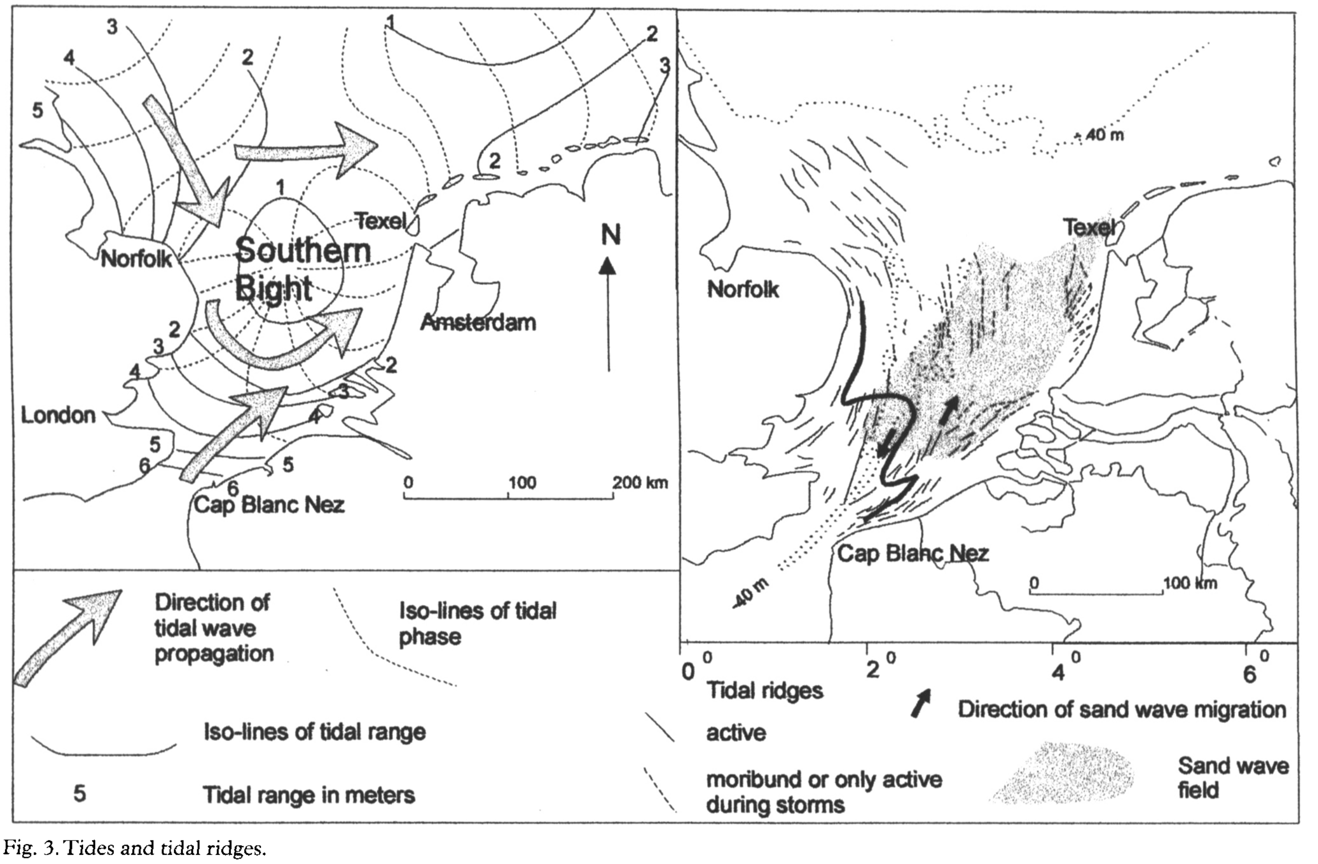

Fig. 3. Tides and tidal ridges.
A: path of the astronomical tide in the Southern Bight (southern North Sea).
B: locations of the tidal ridges and the sand wave field in the Southern Bight; the thick, wavy line between Norfolk and Cap Blanc Nez indicates the approximate boundary between northward and southward transport, as deduced from the sand-wave morphology; modified from McCave (1971) and Houbolt (1968).

above adjacent swales, and can reach to less than 10 m below sea level. They are several tens of kilometres long and 2-5 km wide. Most, if not all, of the Holocene sand in the southern North Sea is reworked from Pleistocene deposits. At present, the rivers surrounding the Southern Bight – as, for instance, Rhine and Meuse – only carry suspension to their mouths. As will be discussed below, it is not very likely that this was much different during most of the Holocene, as bed load as well as part of the suspended load was deposited in the alluvial plain to fill up the space created by the rapidly rising base level (= sea level). Tidal currents may well have transported sand from the Channel via the Strait of Dover into the Southern Bight during the early Holocene, when the southern North Sea was not yet connected to the northern North Sea. Both the orientation of the sand dunes in the Southern Bight (McCave, 1971; Johnson et al., 1982) and tidal modelling (Austin, 1991) show, however, that at present residual currents in the southern part of the Southern Bight promote export of sand to the Channel area (Fig. 3B).

Modelling of tide-induced sediment transport in the North Sea at sea level stands of 0 m (present), –5 m (5000-6000 BP) and –15 m (7000-8000 BP), using a flow model based on the shallow-water equations and a depth-integrated transport model for suspended sediment, shows a dominance of erosion in the southern part of the Southern Bight and deposition to its north. The area of deposition is an E-W to SE-NW running, up to 100 km wide zone which in all three simulations has its southern boundary about halfway the Holland coast (Gerritsen & Berentsen, 1998). The model for the present-day situation is in good agreement with the data on sediment transport and deposition in the Southern Bight. Little is known, however, of the sediment transport in the Southern Bight during most of the Holocene, so that we cannot confirm the results of the modelling at lower sea-level stands. As will be discussed below, our data indicate that not all sediment stored in the Holocene coastal plains of Belgium and the Netherlands comes from erosion of headlands during coastline recession or from an alluvial source. An important part of the sediment is thought to be derived from the Southern Bight, and a provenance mechanism as suggested by the modelling of Gerritsen & Berentsen (1998) would fit well with this concept.

Relative sea-level rise and the flooding of the Southern Bight

Differential glacio- and hydro-isostatic movements are reflected in the three RSL curves for the area, which show a southward decrease in rate of subsidence (Fig. 4A). The northernmost one, for the German Bight and the Oyster Grounds north of the Frisian Islands (Behre et al., 1979; Ludwig et al., 1979, 1981) has the highest rate of RSL rise prior to 7000 BP and merges more or less with the one for the Holland coast (Jelgersma, 1961, 1979; Van de Plassche, 1982; Van de Plassche & Roep, 1989) after that time. The RSL curve for the Belgian coast (Denys & Baeteman, 1995) lies originally above that for the Holland coast but joins it from about 3000 BP onward.

The rate of RSL rise prior to 7000 BP was in the order of 0.7 $cm{\cdot}a^{-1}$ for the Belgian coast and about 1.6 $cm{\cdot}a^{-1}$ for the area north of the Frisian Islands. At this high rate, the Southern Bight flooded rapidly (Fig. 4B). At 8500 BP, the Southern Bight was still isolated from the northern North Sea; at its southeastern side, the sea had reached the Belgian coastal plain (Denys & Baeteman, 1995). Around 8000 BP, the sea invaded the depressions in the coastal plain of the western Netherlands but it was still north of the Frisian Islands. By then, the Southern Bight and the northern North Sea were connected, but the present-day Dogger Bank still formed a large island in between. Considering the slight dip of the pre-transgressional surface, we assume that, at 8500 BP, the size of the Southern Bight was sufficient to produce waves at its eastern shores capable of building a protective barrier behind which a complex of estuaries and tidal basins could develop.

The development of the coastal plains of Belgium and the Netherlands

The Holocene development of the coastal plain is controlled mainly by the morphology of the flooded

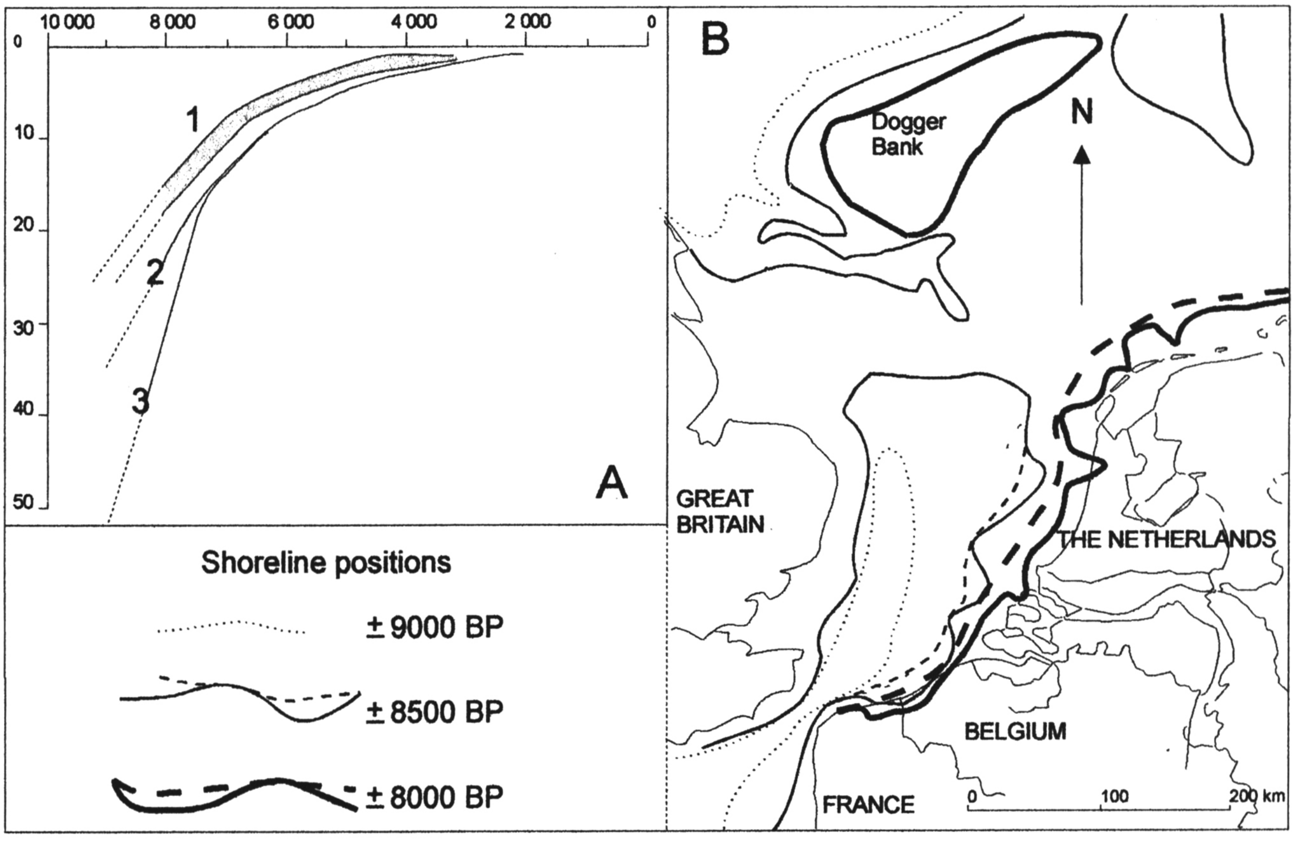

Fig. 4. Sea-level rise and resulting flooding history.

A: time/depth plot of RSL rise for (1) the Belgian coastal plain, after Denys & Baeteman (1995), (2) the Holland coastal plain, after Jelgersma (1961, 1979), Van de Plassche (1982) and Van de Plassche & Roep (1989), and (3) the North Sea north of the Frisian Islands, after Behre et al. (1979) and Ludwig et al. (1979, 1981); the Belgian curve has been plotted as envelope; the basis for drawing this envelope was discussed by Denys & Baeteman (1995)

B: flooding history of the Southern Bight based on the RSL curves of Figure 3A (extrapolated for the 9000 and 8500 reconstruction) and the reconstructed pre-transgressional surface (= reconstructed top of the Early Holocene back-barrier deposits, based on the 1:250 000 sheets of the southern North Sea by the Geological Surveys of the UK, the Netherlands and Belgium, and the unpublished map of the base Holocene of the Dutch sector of the North Sea by Kenneth Rijsdijk; solid lines represent the landward boundary of the back-barrier, dashed lines are the inferred barrier coastlines.

surface, the (decreasing) rate of RSL rise, and the sediment supply.

The morphology of the flooded surface

This aspect defined the position of sinks and sources. Recent work on the Holocene development of the North Sea coastal plain (Baeteman, 1985; Streif, 1988, 1990; Beets et al., 1992; Flemming & Davis, 1994; Oost & De Boer, 1994; Van der Spek, 1996) has shown that the plain was formed by the amalgamation of a large number of relatively small and shallow estuaries, as the rising sea invaded the valleys of the existing drainage pattern and the divides developed into headlands. The landscape prior to flooding, which is now the top of the Pleistocene deposits in the subsurface of the coastal plain, consisted of a drainage pattern of west- and northward-directed streams separated by low divides (Fig. 2A). Except for the Rhine, Meuse and Scheldt rivers, most of these streams provided local drainage.

In contrast to the valleys of these small streams, that of the Rhine/Meuse/Scheldt (the latter two rivers were tributaries of the former during the Early Holocene) was not flooded, as these rivers supplied sufficient sediment to balance the accomodation space created by sea-level rise. Instead, it formed the divide between the Holland tidal basin and a tidal basin at the present site of Zeeland (Fig. 2). As the divides between the valleys were low, they were eventually also flooded, transforming small estuaries into larger tidal basins (in this paper the terms 'estuary', 'tidal basin' and 'back-barrier basin' will be used loosely as synonyms).

The rates of RSL rise and sediment supply

These two parameters defined whether the coast was receding, stable or prograding. All sediment in the back-barrier basins of Belgium and the Netherlands entered from the North Sea; even sediment supplied by the rivers Rhine, Meuse and Scheldt reached the basins via shoreface and inlets (Beets et al., 1992). As will be discussed below, hardly any of the bed load of the rivers reached the sea directly; most was deposited in the alluvial plain. Only after 4000 BP, when the rate of RSL rise at the coast of Holland was reduced to less than 0.2 cm per year, the main branch of the Rhine could build a small delta. The sediments deposited prior to 6000 BP in that part of the alluvial plain offshore of the present coastline were eroded during recession of the coastline and transported by waves and currents into the tidal basins. Consequently, all were filled from the inlets backwards. Sand to fill the basins was derived from the shoreface adjacent to the inlets and from the ebb-tidal deltas. As long as there was accommodation space in the back-barrier basin, the tidal prism set up tidal currents that brought sand and mud to fill up this space. If insufficient sand was supplied to the shoreface by longshore and cross-shore transport to compensate for this sediment loss, the shoreline was forced to recede.

The size of the back-barrier basin, which is a function of the slope of the pre-transgressional surface, and the rate of RSL rise are the main factors defining the accommodation space. At the rates prior to 7000 BP, sediment supply was insufficient to compensate for the accommodation space created by RSL rise. When the rate of RSL rise dropped to 0.25 cm per year after about 7000 BP along the Belgian coast (Denys & Baeteman 1995) and to 0.3 cm per year after about 6000 BP along the Holland coast (Van de Plassche, 1982), sediment supply could, however, gradually catch up with RSL rise along these west-facing coasts. One after the other, the estuaries were filled, resulting in the closure of associated tidal inlets, stabilization of the adjacent coast, and local barrier progradation. Sediment supply to the northern coast of the Netherlands, on the other hand, was insufficient to fill up the estuaries completely, so that this coastal stretch has remained a slowly receding barrier coast until today.

As will be discussed below, the rate of sediment supply to the back-barrier basins during recession of the coast was high. In the relatively small back-barrier basins of Belgium, intertidal flat and supratidal saltmarsh deposits are therefore common, despite the high rate of RSL rise prior to 7000 BP (Baeteman, 1985, Baeteman et al., 1999). Vegetation levels and thin peat layers are intercalated between the estuarine deposits, indicating that short periods of estuarine sedimentation alternated with even shorter episodes with freshwater conditions and that, locally and temporarily, sediment supply outran the creation of accomodation space by RSL rise (Baeteman et al., 1999). Between about 5500 and 4500 BP, when the rate of RSL rise decreased to less than 0.1 cm per year, and the barrier stabilized, the entire coastal plain changed into a freshwater marsh with peat formation.

The development of the coastal plain of the Netherlands, although similar, differs in a number of respects, such as timing. In the Zeeland area, the early development of the estuary is similar to that of the Belgian coastal plain, but for the absence of peat intercalations. Between 5000 and 4500 BP, this coastal plain silted up because of the drop in the rate of RSL rise (Fig. 5), and the barrier stabilized. During the next centuries, the tidal inlets closed and the entire

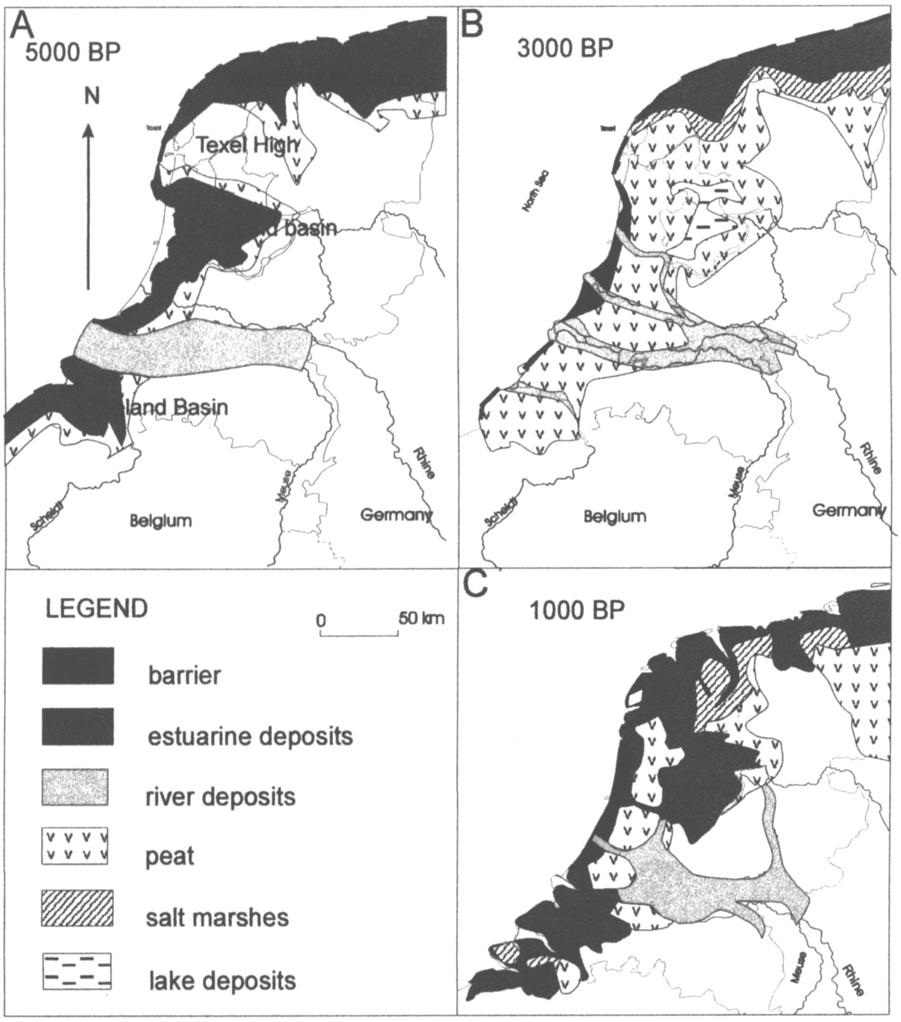

Fig. 5. Simplified paleogeography of the coastal plain of the Netherlands at approx. 5000 BP, approx. 3000 BP and approx. 1000 BP; modified from Zagwijn (1986).

area changed into a peat marsh (Vos & Van Heeringen, 1997). This marsh soon developed into an oligotrophic peat cushion, which remained intact until about 2000 BP.

The Zeeland tidal basin is separated from that of Holland (Figs. 2 and 5) by a wide E-W belt of deposits of the Rhine and Meuse rivers, which occupied this location in the central Netherlands between the Weichselian Pleniglacial and about 4000 BP (Törnqvist, 1993; Weerts & Berendsen, 1995; De Groot & De Gans, 1996). After 4000 BP, a more northerly branch became the main Rhine discharge (Figs. 5B and 6).

The Holland tidal basin is shallow and narrow at its southern end but deepens and widens towards the north, where it occupies the valley formed by the confluence of a number of local westward-draining streams (Figs. 2A and 5A). In the north, it is bounded by the Texel High, the former divide between the westward and northward drainage. This divide formed a headland throughout most of the Holocene development. In combination with one of the major inlets of the Holland tidal basin, just south of this headland, it became an efficient barrier for longshore transport from the SSW-NNE running Holland coast to the E-W-running Frisian Island coast (Fig. 5A and 5B). The southern part of this basin silted up after 6000 BP and the oldest preserved barrier dates from about 5500 BP (Van der Valk, 1996a). Because of its larger volume, filling of the northern part of this tidal basin took much more time. Here, the barrier shifted landward until about 4400 BP (Beets et al., 1996) and the last tidal inlet closed at about 3300 BP (Roep & Van Regteren Altena, 1988). In the meantime, the barrier of the Holland tidal basin started to prograde, indicating that the supply of sediment in the south outran the capacity of the northward longshore transport (Fig. 6). Sedimentation in the Holland tidal basin prior to 6000 BP was largely subtidal, as sediment supply lagged behind the creation of accommodation space by RSL rise. Intertidal shoals could only be maintained near the barrier and along the sediment-supplying channels (Van der Spek & Beets, 1992). When the rate of RSL decreased after 6000 BP, the subtidal areas in the western part of the tidal basin were rapidly filled in. Simultaneously, the distal,

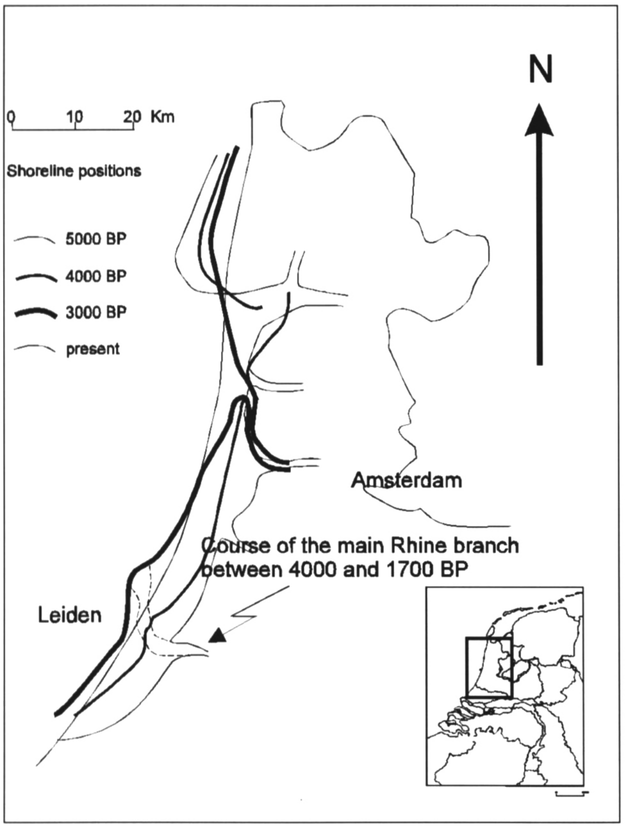

Fig. 6. Coastline positions of the Holland barrier. Timelines of the receding Texel High coast are inferred (after Beets et al., 1992, 1996); see inset for location.

eastern part of the tidal basin became isolated and changed into a major lake, the precursor of the present IJsselmeer (Figs. 1 and 5). After closure of all the tidal inlets, the silted part of the tidal basin changed into a peat swamp (Fig. 5B). Progradation of the barrier coast between the headlands of the alluvial plain in the south and the Texel High in the north continued until after 2000 BP. In the Middle Ages, erosion and recession of this part of the coast started (Jelgersma et al., 1970), mainly because of large-scale inland-directed sanddrift (Van der Valk, 1996b).

The estuaries of the northern Netherlands formed at about 7000 BP when the sea invaded the valleys of two northward discharging local streams in what is now the eastern part of the Dutch Wadden Sea (Fig. 2A). The western part of the Wadden Sea formed much later, around 2000 BP, when the Texel High was flooded (Fig. 5C). The main difference in the development of the northern estuaries in comparison to that of the tidal basins of Holland and Zeeland is that they have not been filled, but remained open until today. The prevailing explanation relates this fact to a difference in sediment supply, governed by the position of the outlets of the Rhine and other major rivers along the west-facing coast. As will be discussed below, the supply by the rivers is too small, however, to account for this difference. Alternatively – and more likely – this variability in sediment supply reflects the hydrodynamics and morphology of the Southern Bight.

Throughout most of the Subboreal (5000-2900 BP), the coastal plain of Belgium and the Netherlands was a freshwater marsh with peat cushions spreading all over the landscape (Zagwijn, 1986), except for its northward-draining tidal basin. The only open-marine influence occurred along the outlets of the rivers. The RSL rise continued, however (although at a much slower rate than before), so that this situation gradually changed. Estuarine conditions spread along the river outlets and caused the drainage and resultant collapse of the peat cushions, creating potential and real tidal prisms. In the Middle Ages, this process culminated in extensive flooding of once habitable areas. Flooding of the Belgian coastal plain and the Zeeland area started shortly after 200 AD and transformed the peat areas into wide estuaries with channels, mud flats and salt marshes. Most of the interfluve areas between the channels in the Zeeland area were silted up to high-tide level by 1000 BP, and people started to construct dikes to protect these newly formed islands. In the northern and northwestern part of the Netherlands, the Texel High was flooded and the Wadden Sea became connected to the large lake in the centre of the country (Fig. 5C). At the same time, extensive peat areas behind the coastal barrier of Holland changed into tidal mud flats. By then, political organization was such that protective measures could be taken to protect the people and their land.

Sediment volume in the coastal plain of the Netherlands

We have quantified the amount of sand, mud and peat in the subsurface of the Dutch coastal plain by subdividing the outcrop area into 16 sub-areas, for each of which the amount of sand, mud and peat was calculated on the basis of one or more profiles (cf. Van der Spek, 1995). It was possible, by using different combinations of profiles, to constrain the error range to ± 15%, which holds for all the figures given below.

The total amount of sediment stored in the Holocene coastal plain of the Netherlands is about $226 \cdot 10^9$ m^3, almost 70% of which is sand (Table 1). The amount of peat is small, as much has either disappeared by erosion, oxidation and exploitation, or has been strongly reduced in volume because of compaction. The Wadden Sea tidal basin contains slightly more sediment than the Holland and Zeeland (Table

Table 1. Total sediment volume of the Holocene deposits in the coastal plain of the Netherlands.

material	volume	relative percentage
sand	$153 \cdot 10^9$ m^3	68 %
mud	$62 \cdot 10^9$ m^3	27 %
peat	$11 \cdot 10^9$ m^3	5 %
total	*$226 \cdot 10^9$ m^3*	*100%*

Table 2. Sediment volumes (in 10^9 m^3) in the three main tidal basins of the Netherlands.

area	sand	mud	peat	total	percentage
Wadden Sea	64	19	3	86	38 %
Holland	44	27	4	75	33 %
Zeeland	45	16	4	65	29 %
total	*153*	*62*	*11*	*226*	*100 %*

2) basins. When calculated over the length of the shoreline (Table 3), however, sediment supply to the Holland and Zeeland basins was considerably higher, particularly when considering that the Wadden Sea basin stayed open and received sediment throughout the Holocene, whereas the others were closed during a large part of the Subboreal and Subatlantic.

We tried to quantify the variations in sediment supply through time, but because of insufficient ^{14}C data to construct isochrons in the sediment succession, we could make only a rough guess based on the assumption that sedimentation in the back-barrier basin would follow RSL rise. This assumption is justified for most of the period after 5000 BP when the rate of RSL rise was less than 3 cm per year. Prior to 5000 BP, the sedimentation rate in the Holland tidal basin lagged behind RSL rise, requiring a correction which we based on the existing ^{14}C record (see for details Van der Spek, 1995). As expected, sediment supply was high during the Atlantic, and decreased during the Subboreal and during the Subatlantic (Table 4). This decrease primarily reflects the decrease in the rate at which new accomodation space was created, governed in turn by falling rates of RSL rise.

Not all sediment accumulation is related to the creation of accommodation space in the back-barrier basins, however, as reflected by the Subboreal progradation of the Holland coast (Fig. 6), which involved $9\text{-}10 \cdot 10^9$ m^3 of sand (Beets et al., 1992). Nevertheless, our numbers show a strong overall decrease in time, to almost zero at present, of the sediment-supply rate to the Dutch coastal plain. In order to explain this decrease, we need to investigate the sediment sources.

Table 3. Sediment volumes of the Holocene deposits in the coastal plains of the Wadden Sea, Holland and Zeeland areas (in 10^3 m^3 per meter coastline).

area	length of the coast-line (km)	sand	mud	peat	total of clastics
Wadden Sea	161 km	402	119	17	521
Holland	100 km	440	269	43	709
Zeeland	96 km	472	162	38	634

Table 4. Accumulated sediment volumes and rates of accumulation per time interval.

time interval	age (14C years BP)	volume (10^9 m^3)	accumulation rate (10^6 m$^3 \cdot$ a^{-1})
Atlantic	8000-5000	136 (60%)	41
Subboreal	5000-2900	67 (30 %)	27
Subatlantic	2900-recent	23 (10%)	7

Sediment sources

All clastic sediments in the coastal plain of Belgium and the Netherlands derive from only three sources: an alluvial source (the Rhine, Meuse and Scheldt rivers), the Pleistocene basement eroded during recession of the shoreline, and the North Sea (Van Straaten, 1965). The individual contribution of the different sources is difficult to quantify on the basis of grainsize and mineral composition as the Pleistocene basement of the North Sea in front of the coasts of Belgium, Zeeland and Holland consists of low-stand Rhine deposits or coversands derived from the latter. As shown by Eisma (1968), the heavy mineral association of the coastal sands along the Holland coast has a strong Rhine signature, but whether the sands are derived by longshore drift from the outlets of the river or by cross-shore transport from reworked Pleistocene Rhine deposits on the North Sea floor cannot be established.

At present, the rivers bring little or no sand to their outlets. The mean discharge of the Rhine at the Dutch/German border is 2300 m$^3 \cdot$s^{-1}, stemming from both rain and snow. Maximum discharges are up to 12,000 m$^3 \cdot$s^{-1}. It is estimated that the total amount of bed and suspended load passing the German/ Dutch border is in the order of $0.5 \cdot 10^6$ m$^3 \cdot$a^{-1} and $2 \cdot 10^6$ m$^3 \cdot$a^{-1}, respectively, based on measurements of sediment transport carried out over the last fifteen years in the main branch of the Rhine (Ten Brinke, 1998). Most of the supplied bedload is dredged, to keep the river navigable. The mean discharge of the Meuse north of the town of Maastricht near the Belgian/

Dutch border is 230 $m^3 \cdot s^{-1}$, and that of the river Scheldt is even less, so that the contribution of these rivers to the present-day sediment supply is negligible. We postulate that low fluvial sediment supply by the rivers characterized much of the Holocene, based on the amount of bed load deposited by the main branch of the river Rhine in the interval between approx. 4000 BP (corresponding with approx. 2500 BC: Stuiver & Reimer, 1993) and 1700 BP (200 AD), and the evolution of channel morphology during the Holocene (Törnqvist, 1993; Törnqvist et al. 1993; De Groot & De Gans, 1996). No Holocene delta deposits have been preserved, except for a small protrusion in the otherwise smooth barrier and swale morphology of the prograding barrier sequence around the outlet of the largely Subboreal Rhine branch near the town of Leiden (Figs. 5B and 6). Over a period of about 2700 year, this was the main branch of the Rhine (Törnqvist, 1993; De Groot & De Gans, 1996).

During this period, the river built an about 2-km wide, 10-m deep and 70-km long sand-rich meander belt, which contains about $1.4 \cdot 10^9$ m^3 of sand. By extrapolation of the barrier morphology offshore, a modest delta can be reconstructed which contained about $0.5 \cdot 10^9$ m^3 of sand if we place its base at a mean depth of -15 m NAP (Beets et al., 1992). The reconstructed delta formed part of the prograding Holland barrier coast. Model calculations by Zitman (in Beets et al., 1992) show that between approx. 5000 and 3000 BP about 10^9 m^3 of sand was supplied to this coast by northward-directed longshore transport. As the delta was incorporated in this longshore transport cell, it changed the gradients in longshore transport so that river sand removed by longshore transport towards the north was replenished from the south. Only after abandonment of the channel, longshore transport smoothed the coast: at present, the protrusion can only be recognized in the fossil barrier and swale morphology. Reconstruction of the delta gives us therefore a rough estimate of the sand transport by the main Rhine branch between approx. 2500 BC to 200 AD: ± $1.9 \cdot 10^9$ m^3 in about 2700 calender years, which is approx. $0.7 \cdot 10^6$ m^3 annually.

Sediment supply by the rivers during the Holocene depends on the discharge (which is a function of climate and vegetation), on the channel gradient and on the erodability of the rocks in the catchment area. Holocene climatic variations (Lamb, 1982; Bond et al., 1997) certainly influenced the rainfall, as is evident from, for instance, lake levels in the Alps (Joos, 1982) and changes in the vegetation in peat bogs (Van Geel et al,. 1996). Consequently, discharge of the rivers may have been greater in the past. On the other hand, the catchment areas were characterized by a dense vegetation reducing the runoff and inhibiting erosion. None of these factors can be quantified as yet, but, as the life time of the Rhine branch discussed above is about twice the duration of the cycles of climatic fluctuations as established by Bond et al. (1997), we assume that the variations in sediment transport because of variations in discharge are incorporated in our calculation of the total sediment output.

The decreasing gradient due to base-level rise (= RSL rise) finds expression in the succession of channel morphologies in the alluvial and coastal plain (Törnqvist, 1993; Weerts & Berendsen, 1995; De Groot & De Gans, 1996; Makaske, 1998). The Holocene succession of channel types consists of meandering channels in the lower part overlain by anastomosing channels which pass upwards in the present-day channel morphology. The transition from meandering to anastomosing in the area between Rotterdam and Utrecht (Fig. 1) took place at about 7000 BP and is the result of the rapidly rising base-level, which forced the river to dump all or a large part of its bedload in the alluvial plain to keep pace with RSL rise. Törnqvist et al.(1993) showed that the anastomising channels graded into the meandering type upstream, indicating that the two different channel types are not the result of changes in discharge and sediment transport. This lateral change in channel morphology and overall alluvial facies also suggest that the meandering channels in the basal part of the Holocene succession in the present coastal plain were once connected to anastomosing channels active at a much lower base level when the coast was situated in the present North Sea. Beets et al. (1992) assumed that the sand deposited in those channels was incorporated in the receding shoreline and eventually redistributed by longshore transport.

The above considerations give little support to major differences in discharge and sediment transport of the rivers during the Holocene, and support our view that discharge and sediment load have not changed considerably since the Early Atlantic. For this reason we consider the alluvial contribution to the Holocene sand budget to be in the order of 10^6 m^3 per year. The present annual discharge of suspended mud by Rhine and Meuse was estimated by Terwindt (1977) at$1.5 \cdot 10^6$ metric ton, which is comparable to the sand supply. Even if we assume that all mud and sand supplied by the rivers during the Holocene was deposited in the coastal plain, the contribution of the rivers hardly surpasses 10% of the $226 \cdot 10^9$ m^3 of sediment stored in the coastal-plain succession of the Netherlands. Hence, 90% of all sediment stored in the coastal plain derived from the two other sources, i.e. the Pleistocene basement and the North Sea.

There are two possibilities to distinguish roughly between the contributions of these two sources. The first is by using the sediment budget of the prograding Holland barrier sequence; the second is by comparing the sediment supply to the north-facing Wadden Sea back-barrier basins to that of the west-facing Holland basin. One should realize that both represent very rough estimates and are only meant to give some approximation for the contribution of the various sources to the sand budget. As outlined by Beets et al. (1992), about 70% of the $9\text{-}10 \cdot 10^9$ m^3 of sand stored in the prograding Holland barrier sequence could be accounted for by the erosion of headlands (approx. $3 \cdot 10^9$ m^3), by the small Rhine delta discussed above ($0.5 \cdot 10^9$ m^3) and by the incorporation of ebb-tidal deltas of former inlets ($3.5 \cdot 10^9$ m^3). They assume that the remaining "3 billion m^3 must have come from reworking of Pleistocene sands in the southern North Sea and from landward transport by shore-normal processes". Progradation took place between about 5000 (corresponding with approx. 5700 calender years BP) and 2000 BP over a coastline of about 80 km long, which gives a sediment supply from this North Sea source of about 10 $m^3 \cdot m^{-1} \cdot a^{-1}$. This is a minimum value as most of the mud from this source was not incorporated in the barrier sequence.

Provided that all sediment of the Wadden Sea tidal basin and barrier was supplied by erosion of the basement during recession of the barrier, and assuming that sediment supply by erosion during recession of the barrier is independent of the orientation of the coastline and was roughly the same per meter coastline for the Wadden Sea tidal basin and the Holland tidal basin, it is possible to estimate the difference in sediment supply per time unit, which gives us the supply rate of the North Sea source. The total amount of sand and mud in the Wadden Sea tidal basins is $521 \cdot 10^3$ m^3 per meter coastline (Table 3). When progradation of the Holland coast ended at about 2000 BP, and no sediment was added anymore to the Holland coast and basin, about 90% of $521 \cdot 10^3$ $m^3 \cdot m^{-1}$ coastline (= $469 \cdot 10^3$ $m^3 \cdot m^{-1}$) was present in the Wadden Sea tidal basin; the clastic sediments present in the Holland tidal basin and barrier amount to $709 \cdot 10^3$ $m^3 \cdot m^{-1}$ (Table 3). Provided that both areas received the same amount of sediment by erosion during recession of the coast, the North Sea source supplied $709 \cdot 10^3\ m^3 \cdot m^{-1} - 469 \cdot 10^3\ m^3 \cdot m^{-1} = 240 \cdot 10^3$ $m^3 \cdot m^{-1}$ to the Holland tidal basin. If we assume that this North Sea source operated from about 8000 BP (approx. 8700 calender years BP) onwards and supplied sediment to the Holland coast until 2000 BP, the supply rate would be in the order of an average of 30-40 $m^3 \cdot m^{-1} \cdot a^{-1}$. As discussed earlier, the sediments deposited by the rivers in the alluvial plain are reworked by coastal recession and are therefore not treated as an independent source in these calculations.

The calculated rate of sediment supply by the North Sea source to the west-facing coasts during the Holocene is of the same magnitude as the modern northward-directed longshore transport at the -20-m depth contour along the coast of Holland, which amounts to 10-40 $m^3 \cdot m^{-1} \cdot a^{-1}$ (Van Rijn, 1995). According to Van Rijn (1995), hardly any of this sand reaches the present coastline. Modelling of the sediment transport in the North Sea at the present and lower sea-level stands by Gerritsen & Berentsen (1998) has shown that there is no basic difference between the pattern of erosion and deposition at present and in the past, suggesting that changes in the coastline configuration are responsible for this change in sediment dispersal.

The most important changes in coastal morphology were the transition from a more or less open coast to a closed coast in the Late Atlantic and Early Subboreal, and the gradual steepening of the west-facing shoreface of Holland in the Subatlantic, in particular during and after the Middle Ages (Van der Valk, 1996a). During the Atlantic, the sediment flux probably had an important east-west component, from the Southern Bight into the estuaries, as the back-barrier basins provided accomodation space and tidal prism. Those of Belgium were the first to silt up because of their size and the relatively early decrease of the rate of RSL rise. Given the fact that there was little or no progradation of the coastlines of Belgium and Zeeland after their closure, it is tempting to assume that the strong tidal currents on the coast-parallel Flemish and Zeeland Ridges prohibited cross-shore transport, but that part of the northward flux of sand was bent towards the Holland coast. The onset of progradation of the coast of Holland in the late Atlantic (approx. 5500 BP) marks the beginning of the change from an open coast with a large number of inlets to the closed coast of today (Pons et al., 1963; Beets et al., 1992). Ebb-tidal deltas of degrading and closed inlets, modified by waves and tidal currents into shoreface-connected ridges, could have been the stepping stone for sand from the North Sea to enter the upper shoreface, where waves could transport it shorewards. Incorporation of these modified ebb-tidal deltas into the prograding barrier sequence during the course of the Subboreal and Early Subatlantic would have cut off this cross-shore pathway for the Southern Bight sand flux. This could eventually have led to shoreface erosion in historical times (Jelgersma et al., 1970; Van der Valk, 1996a,b).

In contrast to the SE-NW-running shoreline of Holland, no such extra sediment source was available to the northern Wadden Sea tidal basin. All the sand of this estuary came from erosion of the Pleistocene deposits during ongoing barrier recession. The Texel High formed a headland that remained an obstacle for longshore transport of sand from the Holland coast to the Wadden Sea coast almost up to Medieval times.

Summary and conclusions

The following seven main conclusions can be drawn.

Rapid flooding of the southern part of the North Sea, reflecting high rates of RSL rise, took place between 9000 and 8000 BP. Differences in the rate of RSL rise are due to differential glacio- and hydro-isostatic movement. Ages of estuarine sediments indicate that the sea invaded the depressions in the Belgian coastal plain around 8500 BP and those of the Netherlands at about 8000 BP.

The morphology of the pre-transgressional surface defined the initial position of estuaries and headlands.

The balance between the creation of accommodation space and the sediment supply defined whether the early barriers receded or stabilized. In a tide-dominated back-barrier basin with shoals and channels, the tidal prism, and consequently the transport capacity is directly affected by changes in accommodation space. As all or most sand brought into the back-barrier basin derived from the fronting barrier, an increase in accommodation space meant erosion and recession of the barrier.

The Belgian barrier and back-barrier basin stabilized after 7000 BP when the local rate of RSL rise had fallen to 0.25 $cm \cdot a^{-1}$; between 5500 and 4500 BP, it changed into a freshwater marsh with peat accumulation. The Zeeland and Holland barrier systems followed slightly later; the oldest barrier in the southern part of the Holland coasts dates from 5500, whereas the last tidal inlet closed at about 3500 BP. Stabilization of the barrier and silting of the tidal basins occurred because of the decreasing rate of RSL rise, causing sediment supply to surpass the creation of accommodation space. In contrast to these barrier systems, the Wadden Sea in the northern part of the Netherlands has stayed open until today, indicating that sediment supply has been insufficient to fill up the tidal basin.

Between about 5000 and 2000 BP, the central part of the coast of Holland prograded almost 10 km. Sand for the progradation was provided by the reworking of ebb-tidal deltas of former inlets, by the erosion of the adjacent headlands, by shore-normal transport from the North Sea, and by a small alluvial source.

A total of $226 \cdot 10^9$ m^3 (± 15%) of sediment is stored in the Holocene coastal plain of the Netherlands. It is estimated that about 60% was deposited prior to 5000 BP, 30% between 5000 and 2900 BP (Subboreal) and 10% since then. In other words, sediment supply decreases with the decreasing sea-level rise. This reflects, among others, the decrease in accommodation space.

Although we have little control on the amount of sediment supplied by the Rhine and Meuse rivers, the best calculations available to date show that the contribution of the rivers to the total amount of sediment of the coastal plain is of the order of 10%. This implies that 90% of the sediment derived from erosion of Pleistocene deposits during recession of the barriers and from the North Sea by tide-induced shore-normal currents. Most sediment comes from the first source. Modelling of tide-induced sediment transport in the southern North Sea at sea-levels of 0, -5 and – 15 m by Gerritsen and Berentsen (1998) shows that eroded sediment in the southern part of the Southern Bight has been transported northward throughout most of the Holocene. We suggest that part of this sediment flux found its way into the west-facing back-barriers basins of Belgium, Zeeland and Holland, and is responsible for the early silting up of these basins. Simple calculations, based on the sand budget of the prograding barrier of Holland, and on a comparison of the tidal basins of Holland and the Wadden Sea suggests that the North Sea source supplied 10-40 $m^3 \cdot m^{-1} \cdot a^{-1}$ to the Holland coast. Modelling shows that sand from this latter source does not reach the tidal basins of the Wadden Sea coast, which might explain why these tidal basins still exist.

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