16 The Late Holocene Evolution of the Perimarine Part of the River Scheldt

P. KIDEN

Research Fellow IWONL, Laboratory of Physical Geography, Geological Institute, State University of Gent, Krijgslaan 281/S8, B-9000 Gent, Belgium.

ABSTRACT

Following a brief overview of the Holocene evolution of the river Scheldt, the late Holocene evolution of the tidal part of the river is presented in detail. A number of different phases could be distinguished.

At or near to the end of the accumulation of the clayey peat a new non- or microtidal river channel was incised. Due to lateral displacement of the river channel point bars were formed while at approximately the same time peat accumulation came to an end by the deposition of an upper clay layer. At about 900 to 1000 AD, tidal range increased considerably in the study area and the MHW level started to rise continuously up to the present day. At present, tidal range is of the order of 5 m.

A number of palaeo-water level indicators have been critically examined, in order to obtain data on the evolution of local water level and its relation to sea level rise and to the morphological development of the Scheldt estuary.

It is shown that local MHW level around 900 to 1000 AD was about 0.8 m lower than contemporaneous MHW level at sea due to the presence of an important floodbasin effect in the Scheldt tidal river system. The increase of tidal range after this date has been caused by a decrease of the floodbasin effect, partly due to human intervention.

INTRODUCTION

In comparison with the rivers Rhine and Meuse in the Netherlands, relatively little is known about the Holocene evolution of the river Scheldt, which debouches in the Southern Bight of the North Sea about 150 km south of the mouth of Rhine and Meuse (figure 1). So far only limited investigations have been carried out, the results of which are often difficult to correlate. This has led to a number of different views concerning the Holocene development of the river Scheldt.

In this paper emphasis is put on the late Holocene evolution of the perimarine part of the Scheldt estuary (the freshwater tidal zone). During this period the transition took place from peat accumulation to a fluvial morphodynamical regime and then to a freshwater tidal environment. It is obvious that these changes are conditioned by relative sea level changes and are related to the evolution of the mouth of the river Scheldt and of the Scheldt estuary itself. Attention is also paid to local water level changes in the study area and their relation to sea level rise and morphological development of the Scheldt estuary.

Only the first preliminary results can be presented here, as the investigations are still in progress. These results are based mainly on the study of a number of excavations and on 350 handborings, carried out in the period 1982-1985. Moreover, a critical examination was made of the results of about 700 borings which had previously been carried out by other investigators.

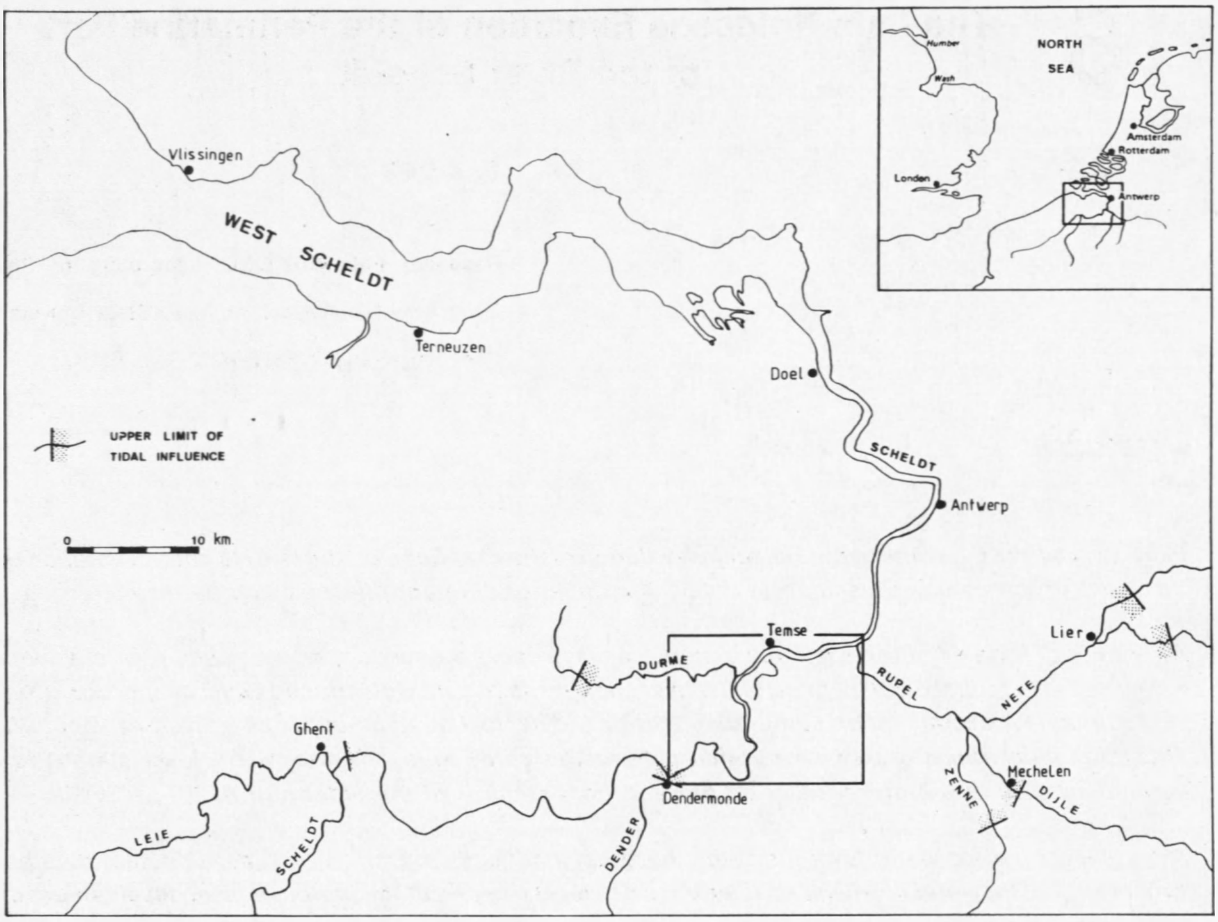

Figure 1 Location map of the study area.

OVERVIEW OF THE HOLOCENE EVOLUTION OF THE RIVER SCHELDT

Although the Pleistocene history of the drainage network in Belgium has since long been a popular subject in national geological research, the Holocene river evolution has been somewhat neglected. A number of facts, however, have been firmly established. A review of the Holocene development of the river Scheldt will be presented here by means of a schematic geological section of the deposits in the Holocene Scheldt valley (figure 2).

The Holocene valley of the Scheldt has been eroded into the mainly sandy sediments of a larger Pleistocene fluvio-periglacial accumulation plain. These sediments have been deposited as the infilling of a complex of Pleistocene valleys which were cut in the Tertiary substratum during periods of Quaternary low sea level stands (DE MOOR, 1963 ; DE MOOR and HEYSE, 1978). The top of the Holocene fluvial sediments in the Scheldt valley reaches +4 to +5 m OP near Ghent and about +1 m OP near the confluence with the river Rupel (figure 1) (OP = Datum, 2.33 m below NAP, the Dutch Ordnance Level). The deepest post-Weichselian river incisions in the Scheldt valley reach -2 to -3 m OP near Ghent and -5 to -6 m OP near the Rupel-Scheldt confluence. This adds up to a maximum thickness of the post-Weichselian deposits of 7 to 8 m. This great thickness, however, is only locally attained in the infillings of the deepest channels. As can be seen in figure 2, the eroded top of the sandy Pleistocene substratum in the Rupel-Scheldt confluence area is mostly situated between 0 and -1 to -2 m OP. In this case, the cover of Holocene sediments is usually 2 to 3 m thick.

The exact age of the above-mentioned post-Weichselian channels is up to now not exactly known and remains a subject of discussion among the investigators. According to VERBRUGGEN (1971) the incision of the deep permanent river courses took place in the beginning of the Late Glacial (about 13000 BP). His conclusion was based on palynological studies and radiocarbon datings of the sedimentary fill of these channels. A ^{14}C-dating of the lower part of the channel fill off a small former upstream tributary of the river Scheldt gave an age of 12655 ± 70 BP (GrN. 6062). DE MOOR and HEYSE (1978) assume that the erosion of the largest meanders of the Scheldt valley took place during the Late Glacial. The deepest incision of the rivers however is supposed to be of early Holocene age and is related to the disappearance of the permafrost layer. The opinion of MIJS *et al.* (1983) is roughly in accordance with that of DE MOOR and HEYSE (1978).

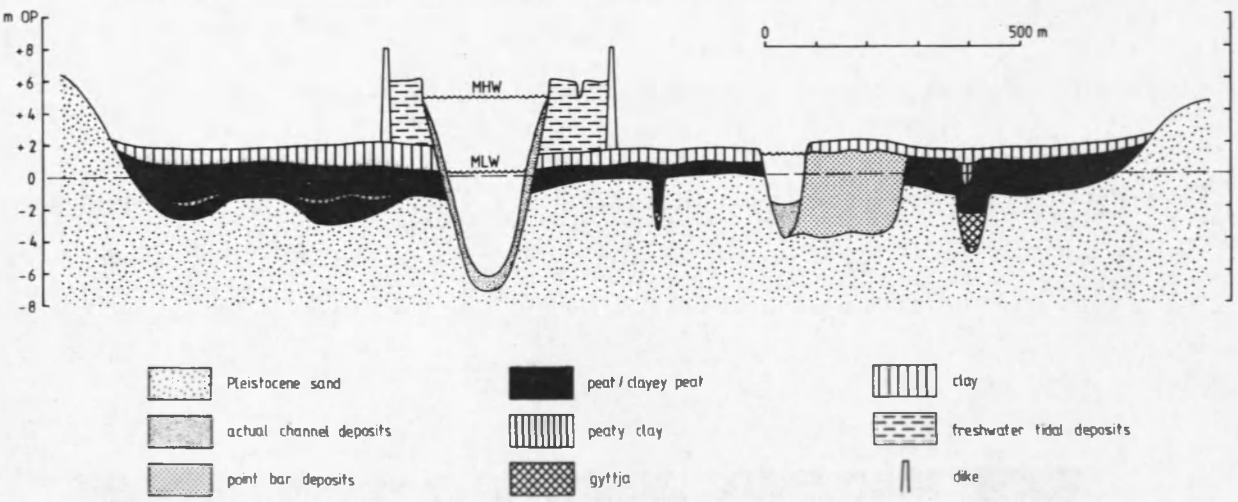

Figure 2 Synthetic geological cross-section of the Holocene deposits of the alluvial plain of the river Scheldt in the study area.

Recent observations in an excavation in the alluvial plain of the Scheldt-Rupel confluence area have yielded some additional data on the late Pleistocene/early Holocene evolution of the river Scheldt. In this excavation, a fining upward sequence of fluvial deposits reaches a level of about +1 m OP. Fossil frost cracks (DYLIK and MAARLEVELD, 1967) with a width of 1 to 5 cm penetrate the upper 2 m of the sequence. They are indicative of the Pleistocene age of these sediments. Thus, the surface of the Pleistocene deposits must have been lowered from at least +1 m OP to about -1 m OP, at present the general altitude of the surface of the Pleistocene sediments in the 2 km wide Holocene alluvial plain of the river Scheldt (figure 2). This period of erosion possibly may be situated in the Late Glacial. In a later phase, vertical erosion became more dominant, fluvial activity became more localized and the incision of the river channels reached a level of -5 to -6 m OP. The question still remains, however, whether this second phase of erosion should be placed in the Late Glacial (VERBRUGGEN, 1971; HUYBRECHTS, 1985) or in the early Holocene (DE MOOR and HEYSE, 1978; MIJS *et al.*, 1983).

Up till now only a few radiocarbon datings have been carried out on the peaty and clayey sediments of the channel fills in the Scheldt valley downstream of Ghent. The oldest dating available gives an age of 8730 ± 70 BP (GrN. 5833) (VERBRUGGEN, 1971). This provides a minimum age for the incision of the channels but does not yield any conclusive evidence on the problem of their Late Glacial or early Holocene formation. In the following paragraphs, their age is assumed to be Late Glacial/early Holocene without any further specification. At the base of the channel fills occurs mostly clay or sandy clay, frequently with small organic remains. On few occasions, a channel lag deposit consisting of coarse sand has been observed. Above these sediments a layer of gyttja can be found, itself in turn overlain by clayey peat. These organic deposits may reach a maximum thickness of 5 to 6 m.

A cover of clayey peat spread over the entire alluvial plain since about 5000 BP (VERBRUGGEN, 1971). The river slowly flowed through this peat area following more or less fixed small gullies. The sedimentary fill of these gullies consists mainly of clay and peaty clay, sometimes with numerous fragments of freshwater molluscs and organic remains. They are often found in zones with a thick peat layer and especially above the deeply incised Late Glacial/early Holocene river channels. These zones were the lowest due to compaction of the underlying gyttja and peat layers and these conditions probably favoured the formation of the gullies. The accumulation of the peat continued during the first part of the Subatlantic and was then brought to an end by the deposition of the upper clay layer (figure 2).

Since the period of peat formation a number of important changes have occurred in the fluvial system of the river Scheldt. They consist of the incision and lateral displacement of a new river channel and the progressive penetration of tidal action in the formerly non-tidal part of the river system. These events will be discussed more fully in the following chapter.

LATE HOLOCENE EVOLUTION OF THE TIDAL PART OF THE RIVER SCHELDT

The preliminary results presented here are based mainly on the investigation of the sediments of the

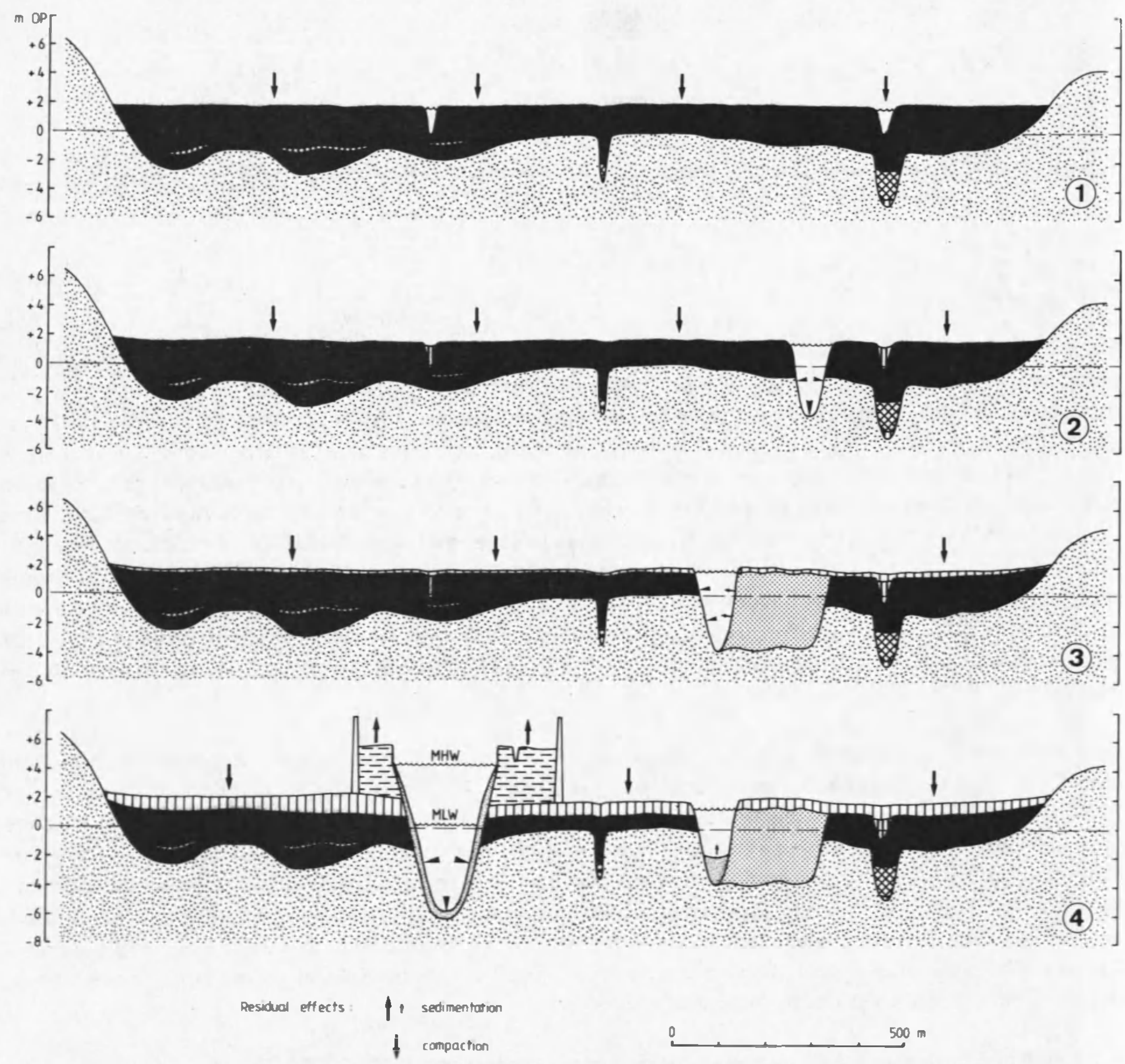

Figure 3 Schematic reconstruction of the Late Holocene evolution of the Scheldt alluvial plain in the study area.

Scheldt alluvial plain in the area between the Dender and Rupel tributaries. The results, however, may probably be generalized over a larger part of the Scheldt fluvial system. The late Holocene development of the river Scheldt will be discussed by means of a series of schematic geological sections representing the main evolutionary phases (figure 3). The numbers of each phase on figure 3 correspond with the numbers of the following sub-chapters.

1. Accumulation of clayey peat

As mentioned before, clayey peat accumulated in the alluvial plain of the river Scheldt since about 5000 BP. Commonly occuring wood fragments seem to indicate wooded conditions. This is confirmed by pollen analysis of the peat sequence, carried out by VERBRUGGEN (1971). At the start of the peat growth, the vegetation consisted of a mixed oak forest (*Quercetum mixtum*). Later on *Alnus, Salix* and *Cyperaceae* became the more dominant elements, indicating wetter environmental conditions. This peat layer levelled off the pre-existing topography of the sandy Late Glacial/early Holocene floodplain, only the highest points of which were not covered with peat. The river Scheldt followed more or less important gullies in the peat landscape which were 1 to 2 m deep and up to 20 m wide.

2. Incision of river channel and end of peat formation

At or near the end of the accumulation of the clayey peat a new river channel was incised into the peat and the underlying sandy Pleistocene deposits. This channel reached -4 to -5 m OP and was 75 to 100 m wide. Presumably, the forementioned small gullies in the peat, active during phase 1, were by now largely filled up with clayey sediments.

The peat layer becomes more clayey towards its top, indicating that the accumulation of organic material continued although more fine grained sediments were deposited during this last phase of peat growth. The clayey peat gradually passes into peaty clay and then into the upper clay layer (phase 3). A few kilometres downstream of Dendermonde (figure 1), the top of the peat layer was ^{14}C -dated at 1585 ± 90 BP (IRPA 97-II) (VERBRUGGEN, 1971).

It is interesting to note that a rather similar course of events has occurred also in other areas during the Subatlantic. BERENDSEN (1984) observed important changes in the river system of the western part of the Netherlands around the beginning of the Christian era. On the opposite side of the North Sea, in the valley of the river Yare in East Anglia (England), the main river channel was enlarged and incised in the peat around 2500 BP. Its large cross-sectional area and steep gradient seem to indicate a higher freshwater discharge at that time than at present, which also served to maintain freshwater conditions in the valley before an upper estuarine clay layer was deposited on top of the peat (COLES and FUNNELL, 1981).

3. Formation of point bars and deposition of upper clay layer

Due to lateral displacement of the river channel, point bars have been formed on the convex banks while the peat and the underlying sandy deposits have been eroded on the concave banks of the river. The point bars are mostly 200 to 400 m wide and up to 5 m thick (figures 3 and 4). They consist of sand with thin clay laminae which increase in thickness and frequency upwards, passing gradually into sand-clay bedding and eventually into clay or sandy clay with sand laminae. The point bar sediments contain small plant debris and overlie the Pleistocene deposits, from which they are often difficult to distinguish due to their sandy nature. The formation of the point bars must have taken place after the main phase of peat accumulation, as the sediments do not show any interfingering with the peat layer.

Important lateral channel migration has also taken place in the lower course of the river Dender, a tributary which joins the river Scheldt at Dendermonde (figure 1). Here, point bars were formed possibly as late as the 14th-15th century AD, as is shown by the dating of archaeological finds from the point bar deposits by PIETERS (Personal communication, 1985).

A thin clay layer has been deposited over the whole of the alluvial plain of the river Scheldt by overbank flow during floods. This upper clay caps the point bars and marks the end of the accumulation of the peat. Taking into consideration the above mentioned ^{14}C -date of 1585 ± 90 BP for the top of the peat, we can assume that this clay layer has been deposited during the last 1,500 years. It must already have been deposited for the greatest part, however, before the start of the embankments in the period 1100-1300 AD.

Since the end of the peat accumulation, compaction of the peat layer has caused a slight lowering of the surface of the alluvial plain. The point bars however have not been affected by this process as the peat had been eroded in these zones and replaced by sandy deposits. As a result of this differential compaction, the point bars at present stand out in the landscape as low ridges.

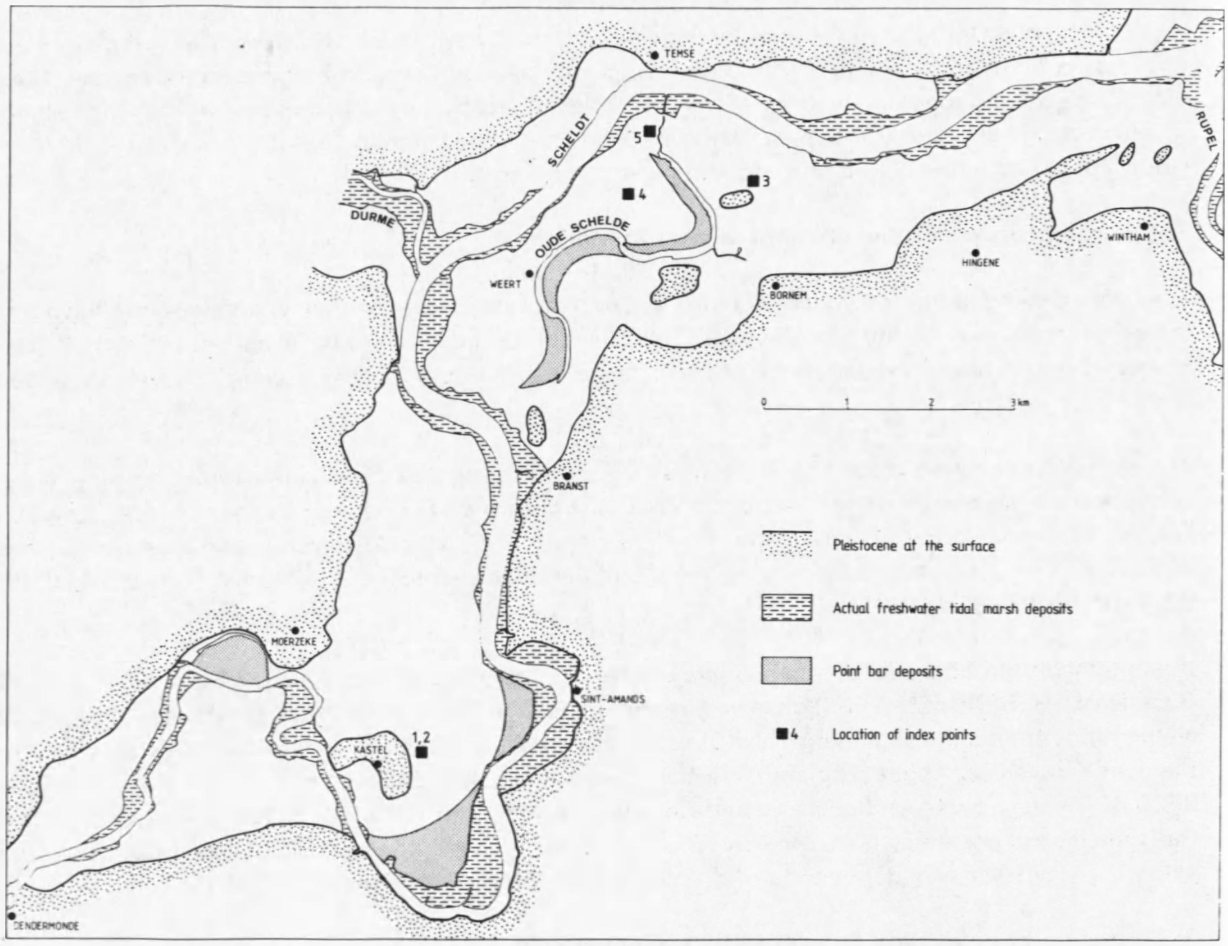

Figure 4 Map of the study area showing late Holocene point bar and freshwater tidal marsh deposits and location of MHW index points in the alluvial plain of the river Scheldt.

4. Penetration of tidal action

According to a.o. SNACKEN (1964) and MIJS *et al.* (1983) tidal influence progressively penetrated in the river Scheldt downstream of Antwerp from about 1100 AD onwards. This conclusion is based mainly on historical data, e.g. periods of embankment. On the basis of palynological investigations and radiocarbon datings, however, VERBRUGGEN (1979) and MINNAERT and VERBRUGGEN (1986) assume that already in Roman times tidal influence was present in the river at Doel (downstream of Antwerp, figure 1). As will be shown below this conclusion does not necessarily hold true for the river Scheldt upstream of Antwerp. In this area the morphological and chronological evidence points towards a more recent penetration of tidal action than is assumed by the last mentioned authors. Tidal influence seems to have reached the study area definitively around 1000 AD, and from then onwards tidal range has increased from about 0 m to its present-day value of about 5 m, which is approximately 1.3 m more than the coastal tidal range measured at the mouth of the Scheldt estuary (CLAESSENS and BELMANS, 1984).

This considerable and rather rapid increase of tidal range and mean high water (MHW) level has had a number of important morphological consequences. Dikes were constructed to counter the threat of flooding to the low lying areas of the Scheldt alluvial plain. This was not completely successful, however, as the embanked polders were flooded several times, often in a catastrophic way, during storm surges. Behind the dike-breaches spill sediments have been deposited and deep more or less circular depressions (Dutch: wiel) have been formed. After a prolonged period of inundation the flooded area developed into a mature brackish or freshwater tidal flat and marsh landscape, with a drainage pattern of tidal gullies and channels diverging from the former dike-breaches. These processes have not only played a major, if not dominant role in the landscape genesis downstream of Antwerp but have also been effective in the freshwater tidal area upstream of Antwerp.

On the outside of the dikes, tidal flats and marshes have been built up concurrently with the rising MHW level (figure 4). At present, these tidal deposits reach a maximum thickness of 4 to 5 m and sedimentation still continues at a rate of approximately 10 mm/y. In the stream channel itself, ebb- and flood-oriented bedforms have been formed. More or less distinct ebb- and flood channels (VAN VEEN, 1950) can now be found as far upstream as Temse.

The increase of tidal range has probably caused the cut-off of the Oude Schelde, a nowadays abandoned meandering river course of the non- or microtidal Scheldt in the vicinity of Temse. Although the exact date of this cut-off is not known, it must have occurred not long before 1241 AD. The channel of the Oude Schelde has been partly silted up and dammed at both ends in 1320-1322 (figure 4) (MEES, 1913). Presumably the present-day tide-influenced river Scheldt follows partly the course of the former Durme tributary (SNACKEN, 1964; MIJS *et al.*, 1983). Due to tidal action, lateral and vertical erosion was intensified, resulting in the deepening and widening of the river channel. The Oude Schelde was not more than 4 to 5 m deep and 75 to 100 m wide, while the tide-influenced river Scheldt at present reaches a depth of 6 to 9 m below mean low water (MLW) and is 150 m wide at MLW and 250 m at MHW (figure 3).

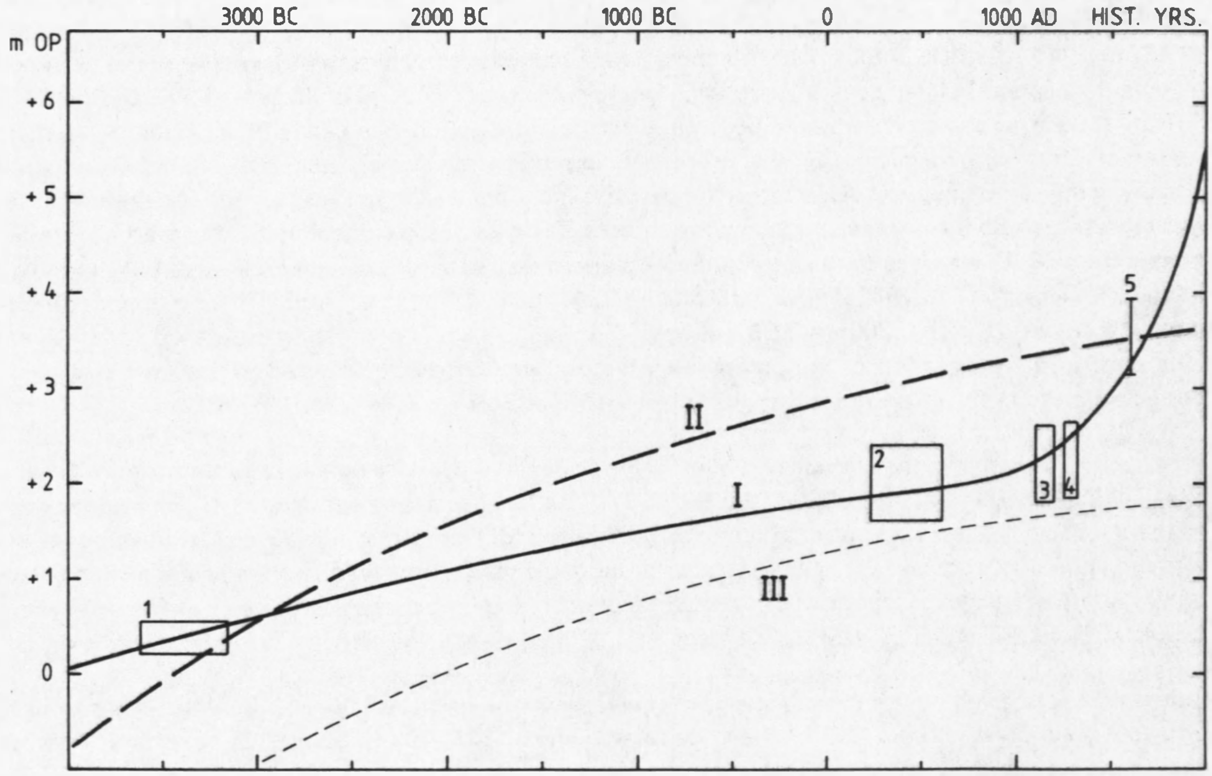

Figure 5 Time-depth diagram with error boxes of MHW index points and local and coastal water level curves. Curve I represents local MHW level, while curves II and III respectively indicate coastal MHW and MSL. A full explanation is given in the text.

THE LATE HOLOCENE FLOODBASIN EFFECT IN THE SCHELDT TIDAL RIVER SYSTEM

In the course of the present study, a number of palaeo-water level indicators have been critically examined in order to obtain data on the evolution of local water level in the perimarine part of the river Scheldt. A number of MHW indicators from the Scheldt alluvial plain between the Dender and Rupel tributaries have been plotted in a time-depth diagram (figure 5). Two main curves were drawn : one representing the evolution of local MHW during the last 4,500 years (curve I) and the other indicating the rise of coastal MHW at the mouth of the Scheldt estuary during the same period (curve II). The reliability of the time-depth data, the method of curve construction and the interpretation of the results are discussed below.

1. Available time-depth data in the study area

Index point 1 is a radiocarbon dating with a conventional ^{14}C -age of 4620 ± 40 BP (GrN. 5847), collected by VERBRUGGEN (1971) at Kastel (figure 4). The calibrated age according to KLEIN *et al.* (1982) is 3640-3155 BC. In the original publication of VERBRUGGEN, only the sampling depth below the topographic surface is given. Nevertheless, an accurate estimate of the altitude of the sample relative to the OP-datum could be made on the basis of a detailed height-point map of the area, made available by Agrotechnic N.V. Therefore, the vertical margin of error is considered to be small (± 0.15 m). The sample was taken at the base of the Subboreal-Subatlantic peat layer overlying the sandy Pleistocene subsoil and does not need to be corrected for compaction of underlying deposits. As the dated sample consists of fen wood peat, it is assumed to have formed at about local MHW level. The altitude of the MHW level indicator is thus the same as that of the sample : +0.35 to +0.40 ± 0.15 m OP = +0.20 - +0.55 m OP.

Index point 2 is a radiocarbon dating with a ^{14}C-age of 1585 ± 90 BP (IRPA 9711), collected by VERBRUGGEN (1971) at the same location as index point 1. The calibrated age according to KLEIN *et al.* (1982) is 230-610 AD. The sample was taken at the top of the Subboreal-Subatlantic clayey peat layer and consists of fen wood peat. The altitude relative to the OP-datum, estimated in the same way as for index point 1, is +1.18 to +1.24 ± 0.15 m or +1.21 ± 0.18 m OP. These figures have to be corrected for the compaction of the underlying clayey peat, which has a thickness of 0.85 m. The compaction of peaty sediments can amount to as much as 85 to 90 % of the original thickness (BENNEMA *et al.*, 1954; VAN DE PLASSCHE, 1980). MOSTAERT (1985) found average compaction values of 40 to 60 % for the upper peat layer in the eastern Belgian coastal plain. An average compaction of 50 % is also mentioned by BEHRE and STREIF (1980). In the study area the peat thins out on compaction-free sandy Pleistocene deposits at a maximum level of about +2 m OP. This represents approximately the highest level of peat growth in the vicinity of the sampling point. If the top of the peat has more or less the same age in the entire alluvial plain, this represents also the original altitude of the sample. The calculated compaction factor for the underlying peat is then 48 %, which is in the right order of magnitude. An additional uncertainty of 0.20 m must be added here as vertical margin of error in this estimation process. The total vertical margin of error for index point 2 is then 0.18 m + 0.20 m = 0.38 m (above-mentioned error in estimating the altitude of the sample plus error in the estimation of the compaction factor), which may be rounded to 0.40 m. This gives a corrected altitude for index point 2 of +2.00 ± 0.40 m OP or +1.60 - +2.40 m OP.

Index point 3 represents the altitude of the top of the upper clay layer of a polder at Bornem (figure 4), which has been diked around 1100-1200 AD (MEES, 1937). The clay has been deposited as a backswamp deposit at about local MHW level or even slightly higher. It thins out on compaction-free Pleistocene deposits at an average altitude of about +2.2 m OP. Due to anthropogenic reworking and disturbance of the surface layer, however, the accurate estimation of this altitude is difficult and introduces a large vertical margin of error of ± 0.4 m. Thus, for index point 3 with an age of 1100-1200 AD, the altitude is +1.80 - +2.60 m OP.

Index point 4 is given by the top of the upper clay layer in the polder of Weert (figure 4). This polder has probably been diked around 1250 AD and certainly before 1320 AD (MEES, 1913). A location has been chosen where a thin clay layer (0.35 m) rests on a peat layer with a thickness of 0.5 m, which in turn overlies compaction-free sandy Pleistocene deposits. The top of the clay layer reaches an altitude of 1.60 ± 0.2 m OP. A compaction factor of 50 % has been assumed for the peat, which gives a lost peat thickness of 0.5 m. An original thickness of 0.5 m has been assumed for the half ripened upper clay layer, which is equivalent to a compaction factor of 30 % (lost thickness 0.15 m). The vertical margin of error must be increased by 0.2 m, which adds up to a total margin of error of 0.4 m. The original altitude of index point 4 can now be calculated : $1.60 + 0.50 + 0.15 = 2.25 \pm 0.4$ m OP or +1.85 - +2.65 m OP.

Index point 5 represents the top of the upper clay layer of a small polder which was diked in 1604 AD (MEES, 1937). This polder is situated near the former northeast (downstream) confluence of the river Scheldt and the Oude Schelde (figure 4). The upper clay layer actually consists partly of freshwater tidal deposits resting on channel fill and point bar sediments of the Oude Schelde. A location has been chosen where the clay layer has a thickness of 1.1 m and overlies the sandy point bar deposits, which have undergone only negligible compaction and rest directly on compaction-free Pleistocene sediments. The top of the clay layer reaches an altitude of $+2.80 \pm 0.2$ m OP. A compaction factor of 40 % has been assumed, which gives an original altitude for the top of the clay of $+3.53 \pm 0.2$ m OP. An additional 0.2 m is added to the vertical margin of error as a result of the uncertainty involved in this correction for compaction. The final result for index point 5 is then $+3.53 \pm 0.4$ m OP or +3.13 - +3.93 m OP. It must be noted, however, that this might indicate the former spring high water level as the clay has probably been deposited above the MHW level as a freshwater tidal marsh sediment.

2. Method of curve construction

A smooth curve (curve I) has been drawn through the available time-depth boxes. From 1862 AD onwards, the curve has been constructed on the basis of MHW data from tidal observations (STESSELS, 1865; CLAESSENS and BELMANS, 1984), which have not been indicated on figure 5 as curve I passes right through them.

The calibrated mean sea level curve established by VAN DE PLASSCHE (1982) has been taken as a basis for the construction of the MHW curve at the mouth of the Scheldt estuary (curve II). VAN DE PLASSCHE considers his curve to be representative only for the northern and western Netherlands coastal area. It will be shown below, however, that this does not seriously affect the results of the present study. The slightly fluctuating MSL curve, which had been established by VAN DE PLASSCHE for the period from about 4750 BC to 750 BC, has been smoothed out and plotted from about 4000 BC to 750 BC in figure 5 (curve III). It has been connected with the MSL time-depth point for 1862-1863 AD derived from tidal observations by STESSELS (1865).

The MHW curve has been obtained from this MSL curve by adding the half of the tidal range (the tidal amplitude). This was 1.8 m at Vlissingen for the period 1862-1863 (STESSELS, 1865). Tidal range at the mouth of the Scheldt estuary, however, has probably not been constant through time. This was not caused by a change of tidal amplitude at sea but by a displacement of the mouth of the Scheldt estuary itself. It is generally accepted that the inlet of the Scheldt estuary was situated more to the north during the greatest part of the Holocene and that it shifted towards its present position during the last 2,000 years (PONS *et al.*, 1963). As tidal range along the coastline decreases in a northward direction, it must have been smaller in front of the former Scheldt estuary than it is at present. Therefore, for the construction of curve II, a coastal tidal amplitude has been chosen which gradually decreases in time from its present value of 1.8 m to a value of 1.5 m at 1000 BC and remains constant at 1.5 m before that date. It must be noted that these assumptions are only valid if tidal range at a given point along the coast has been constant during the last 6,000 years. Moreover, abstraction has been made of any abrupt change of tidal range due to a sudden displacement of the mouth of the Scheldt estuary or the opening up of a new tidal inlet.

3. Interpretation

As can be seen in figure 5, index points 2, 3 and 4 plot well below the contemporaneous coastal MHW level (curve II). Index points 1 and 5, on the other hand, take up a position approximately on the coastal MHW curve. As a result, the local MHW curve (curve I) plots below the coastal MHW curve from about 3000 BC to about 1700 AD. Before 3000 BC, local MHW was higher than coastal MHW level and the two curves seem to diverge slowly. At about 900 to 1000 AD, local MHW started to rise faster than coastal MHW, catching up with it at about 1700 AD and overtaking it to reach its present level, which lies approximately 1.5 m above coastal MHW. These changes of local MHW level are clearly distinct from the general trend of coastal MHW rise, especially during the last 2,000 years. The differences between curve I and II can be attributed to a combination of morphological and hydrodynamical effects which may affect the local water level in an estuary or a tidal river.

In a normal river, the water level rises in an upstream direction. This is also the case in a tidal river or an estuary, where high and low water levels tend to be raised in an upstream direction due to the influx of river water. This effect is comparable to the backwater effect known in hydraulic engineering (CHOW, 1959) and is called river-gradient effect (LOUWE KOOIJMANS, 1974).

Tidal range may increase in a landward direction due to a narrowing of the estuary. This causes the MHW level to be raised in an upstream direction. On the other hand, tidal range and MHW level may be reduced in a landward direction as a result of energy dissipation of the incoming tidal wave due to boundary friction (IPPEN and HARLEMAN, 1966). Furthermore, reduction of tidal range in an upstream direction may also be caused by the so-called floodbasin effect (VAN VEEN, 1950; ZONNEVELD, 1960). In this case, the incident tidal wave enters a basin with a large storage capacity after passing through a narrow inlet. The water volume entering through this inlet around the time of high water is not sufficiently large to allow the basin to fill completely before the water level falls again. As a result, the MHW level in this basin is lower than the coastal MHW level. This basin could also be an estuary with a narrow mouth and a large storage capacity in the form of numerous diverging intertidal channels and creeks. As it is in most cases impossible to distinguish between the floodbasin and friction effect, the latter will be included in the former if no explicit difference between the two is made.

From figure 5, it can be concluded that local tidal amplitude around 1000 AD was significantly lower than at sea. If the upper limit of the time-depth boxes 3 and 4 is considered to represent local MHW level, it must have been at least 0.8 m lower than coastal MHW. This can be attributed to the presence of an important floodbasin effect in the Scheldt estuary at the time. This floodbasin effect may have been even greater because the vertical margin of error of index points 3 and 4 is large (0.4 m). The storage capacity was probably increased by a large intertidal flat area downstream of the study area prior to the start of embankments.

The rapid rise of MHW level since about 900 to 1000 AD has probably been caused partly by a reduction of the storage capacity due to the start of the embankments along the Scheldt estuary and partly by a widening of the mouth of the Scheldt estuary due to more intense tidal erosion. During the last 100 years, the rise has been even more rapid than before. This has been caused by continuous dredging operations in the navigation channels which have been deepened and widened, thereby reducing the frictional energy loss of the tidal motion.

Prior to about 3000 BC, local MHW level was continuously higher than coastal MHW, which may be attributed to the gradient effect in the study area. Tidal influence had not yet been able to penetrate up to the study area at that time.

4. Discussion

It must be noted that the position of curve I between coastal MHW and coastal MSL can be explained in two different ways. The first possibility is that the low local MHW level is caused by an upstream reduction of tidal range, with tidal influence nevertheless still reaching the study area. A second possible explanation could be that tidal range had already been reduced to zero downstream of the study area by an even more important floodbasin effect. The position of local MHW above MSL would then have been caused by a river-gradient effect. In this case, tidal influence would have been absent from the study area. In view of the data available at present, it seems difficult to distinguish between the two above-mentioned hypotheses.

The coastal MSL curve (curve III) is based on the MSL curve established by VAN DE PLASSCHE (1982). He considers this MSL curve to be representative only for the northern and western Netherlands due to a different amount of tectonic subsidence in other areas. Any eustatic MSL curve would probably lie above the MSL curve of VAN DE PLASSCHE and diverge from it further back in time. As the study area is located more to the south and probably underwent less subsidence, this could affect the conclusions drawn above. This could especially alter the older part of curve II and III, thereby rising the contemporaneous MHW level relative to the position of index point 1. In this case, the period with a marked floodbasin effect would be extended further back in time.

The above mentioned considerations, however, cannot significantly change the conclusions concerning the presence of an important floodbasin effect in the study area around 1000 AD. In fact, correcting for an eustatic MHW curve would only enlarge the value obtained for the floodbasin effect. Thus, the general hypothesis will probably remain valid although the exact value of the floodbasin effect could still need to be adjusted when more data will become available.

CONCLUSIONS

A floodbasin effect has probably been present in the river Scheldt since about 3000 BC. In the study area, it reached its maximum value of approximately 0.8 m as recent as 900 to 1000 AD. It has been possible to trace the end phase of dissappearance of this floodbasin effect, which has been influenced by the morphological development of the Scheldt estuary itself as well as by human intervention.

Weak tidal action has possibly been present in the study area during the last 2,000 years. A new non- or microtidal river channel was incised in the peat and due to lateral migration of this river channel small but distinct point bars were formed. Around 400 AD peat growth was brought to an end by the deposition of the upper clay layer. At about 900 to 1000 AD, local MHW level in the study area started to rise rapidly as a result of a considerable decrease of the floodbasin effect. Tidal range increased and tidal action penetrated further into the river system. This caused the cut-off of the Oude Schelde, a nowadays abandoned branch of the formerly non- or weakly tidal river Scheldt. At present, the rise of the MHW level in the Scheldt estuary is largely caused by human activity, especially by the continuous dredging operations.

ACKNOWLEDGEMENTS

The author wishes to thank Prof. Dr. G. DE MOOR and the co-workers of the Laboratory of Physical Geography for their support throughout the research project. I am grateful to the IWONL for offering a grant which has made this investigation possible.

REFERENCES

BEHRE, K.-E. and STREIF, H., 1980. Kriterien zu Meeresspiegel- und darauf bezogene Grundwasserabsenkungen. - *Eiszeitalter u. Gegenwart* 30, 153-160.

BENNEMA, J., GEUZE, E.C.W.A., SMITS, H. and WIGGERS, A.J., 1954. Soil compaction in relation to quaternary movements of sea-level and subsidence of the land, especially in the Netherlands. *Geol. Mijnbouw, Nw. Serie* 16, 173-178.

BERENDSEN, H.J.A., 1984. The evolution of the fluvial area in the western part of the Netherlands from 1000-1300 AD. In : BERENDSEN, H.J.A. and ZAGWIJN, W.H. (eds.) Geological changes in the western Netherlands during the period 1000-1300 AD. *Geol. Mijnbouw* 63, 231-240.

CHOW, VEN TE, 1959. *Open-channel hydraulics*. McGraw-Hill, New York, 680 pp.

CLAESSENS, J. and BELMANS, H., 1984. Overzicht van de tijwaarnemingen in het Zeescheldebekken gedurende het decenium 1971-1980. *Extr. Tijdschr. Openb. Werken Belg.* 1984 (3), 96 pp.

COLES, B.P.L. and FUNNELL, B.M., 1981. Holocene palaeo-environments of Broadland, England. *Spec. Publ. Int. Ass. Sediment.* 5, 123-131.

DE MOOR, G., 1963. Bijdrage tot de kennis van de fysische landschapsvorming in Binnen-Vlaanderen. *Tijdschr. Belg. Ver. Aardr. Studies* 32, 329-443.

DE MOOR, G. and HEYSE, I., 1978. De morfologische evolutie van de Vlaamse vallei. *De Aardrijkskunde* 4, 343-375.

DYLIK, J. and MAARLEVELD, G.C., 1967. Frost cracks, frost fissures and related polygons : a summary of the literature of the past decade. *Meded. Geol. Sticht.* 18, 7-21.

HUYBRECHTS, W., 1985. *Morfologische evolutie van de riviervlakte van de Mark (Geraardsbergen) tijdens de laatste 20.000 jaar.* Thesis, Vrije Universiteit Brussel, 250 pp.

IPPEN , A.T. and HARLEMAN, D.R.F., 1966. Tidal dynamics in estuaries. In : IPPEN, A.T. (ed.) Estuary and coastline hydrodynamics. *Eng. Soc. Monographs*, McGraw-Hill, New York, 493-545.

KLEIN, J., LERMAN, J.C., DAMON, P.E. and RALPH, E.K., 1982. Calibration of radiocarbon dates. *Radiocarbon* 24, 103-150.

LOUWE KOOIJMANS, L.P., 1974. The Rhine/Meuse delta. Four studies on its prehistoric occupation and Holocene geology. *Analecta Praehistorica Leidensia* 7, 1-421.

MEES, J., 1913. La formation du Vieil Escaut. *Ann. Congr. Hist. Arch.*, Gand.

MEES, J., 1937. De economische ontwikkeling van Bornem. *Eigen Schoon en de Brabander* 20, 4-23.

MIJS, M., GULLENTOPS, F., JANSSENS, P., WUYTACK, M.J. and STINISSEN, H., 1983. De holocene evolutie van de alluviale vlakte van de Beneden-Schelde. *Tijdschr. Belg. Ver. Aardr. Studies* 52, 7-33.

MINNAERT, G. and VERBRUGGEN, C., 1986. Palynologisch onderzoek van een veenprofiel uit het Doeldok te Doel. *Bijdragen Archeol. Dienst Waasland* 1, 201-208.

MOSTAERT, F., 1985. *Bijdrage tot de kennis van de kwartairgeologie van de oostelijke kustvlakte op basis van sedimentologisch en lithostratigrafisch onderzoek.* Thesis, Rijksuniversiteit Gent, 587 pp.

PONS, L.J., JELGERSMA, S., WIGGERS, A.J. and DE JONG, J.D., 1963. Evolution of the Netherlands coastal area during the Holocene. *Verh. K. Ned. Geol. Mijnbouwkd. Genoot., Geol. Serie* 21-2, 197-208.

SNACKEN, F., 1964. De ontwikkeling van het Scheldepolderlandschap. In : Verslagboek 4de Int. Havencongres Antwerpen 22-27 juni 1964, 485-490.

STESSELS, M., 1865. *Description hydrographique de l'Escaut.* 27 pp.

VAN DE PLASSCHE, O., 1980. Compaction and other sources of error in obtaining sea-level data ; some results and consequences. *Eiszeitalter u. Gegenwart* 30, 161-170.

VAN DE PLASSCHE, O., 1982. Sea level change and water-level movements in the Netherlands during the Holocene. *Meded. Rijks Geol. Dienst* 36(1), 1-93.

VAN VEEN, J., 1950. Eb- en vloedschaarsystemen in de Nederlandse getijwateren. *Tijdschr. Kon. Ned. Aardr. Gen., Tweede Reeks* 67, 303-325.

VERBRUGGEN, C., 1971. *Postglaciale landschapsgeschiedenis van zandig Vlaanderen.* Thesis, Rijksuniversiteit Gent, 440 pp.

VERBRUGGEN, C., 1979. Paleoecologische en palynologische benadering van enkele bekende historisch-geografische problemen in Vlaanderen. Bronnen voor de historische geografie van België. *Handelingen Coll. Brussel* 25-27-4-1979, 487-497.

ZONNEVELD, I.S., 1960. De Brabantse Biesbosch. Een studie van bodem en vegetatie van een zoetwatergetijdendelta. *Meded. Stichting Bodemkartering, Bodemk. Stud.* 4, 396 pp.