KONINKRIJK BELGIE

MINISTERIE VAN ECONOMISCHE ZAKEN

Geologische Dienst van België

Jennerstraat 13, 1040 Brussel

QUATERNARY SEA - LEVEL INVESTIGATIONS FROM BELGIUM

A CONTRIBUTION TO IGCP PROJECT 200

Edited by Cecile Baeteman

PROFESSIONAL PAPER 1989/6

Nr 241

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PREFACE

The present special edition of the Professional Paper is a contribution to the International Geological Correlation Programme Project 200 : *Sea-Level correlations and applications*. It attemps to present the state of knowledge through the views of different authors in the field of Quaternary sea-level related research in Belgium.

The volume is to be considered as a first exploration and evaluation of a substantial body of evidence collected on various aspects of sea-level change and the impact on coastal development.

In general relatively little work on coastal and sea-level changes in Belgium was published. However, in the past 10 years much research has taken place in coastal geology. This research was not restricted to present land areas adjacent to the coast or estuary, but also on the continental shelf. The interest and enthusiasm for coastal and sea-level changes among young scientists yielded inspiring research and considerable advances have been made.

Therefore there was a strong need to pull together the knowledge about the subject acquired over the last 10 years. The preparation for this edition began in 1986 with the different papers being contributed at various stages since that time. The contributions are the result of mainly individual efforts of researchers working independently, some of them in the framework of a Ph.D.Thesis. Unfortunately, not all methodologies of and approaches to sea-level investigations carried out in Belgium, are included in the volume, and this for various reasons. Shortcomings are the investigations of the dunes, the archaeology and palaeomagnetism which were undertaken also intensively with respect to sea-level changes.

The primary consideration of the sea-level research in Belgium is to define the history of local or regional sea level,rather then to establish a single graph of the trend of mean sea level or to adjust the latest wiggle in the sea-level curve. The main objectives are especially to identify relevant sea-level indicators and to interpret the processes of coastal landform evolution in relation to sea-level changes.

It is to be hoped that the research activity on sea-level and coastal changes in Belgium will continue steadily and extensively. These papers are offered as a small and modest contribution. May it be a stimulus towards this goal.

At last acknowledgement is expressed to the Belgian Geological Survey for providing the opportunity for the publication of this special issue and for stimulating the Quaternary sea-level research.

November 1989 Cecile Baeteman

CONTENTS

HOLOCENE WATER LEVEL MOVEMENTS IN THE LOWER SCHELDT PERIMARINE AREA

Patrick Kiden (*)

ABSTRACT

Gradient lines and local water level movements in the lower Scheldt river could be reconstructed on the basis of a number of newly collected radiocarbon datings.
Due to the presence of a floodbasin effect in the lower Scheldt river region, local MHW level in the Belgian part of the river was situated below coastal MHW since about 4500 BP. This floodbasin effect controlled the rise of local MHW level up to 500 to 1000 AD. Since then, a marked decrease of the floodbasin effect caused a rapid rise of local MHW level and tidal amplitude along a large part of the lower Scheldt river on Belgian territory. The data also indicate that differential tectonic subsidence of the western and northern Netherlands with respect to the lower Scheldt area can not have been greater than about 1.6 cm/century over the last 5000 years.

1. INTRODUCTION

During the last thirty years, important research on Holocene sea level changes has been going on in the Netherlands. Since the classic work of Jelgersma (1961), sea level studies have been based primarily on the collection and evaluation of a large number of radiocarbon datings. This approach has proved highly successful, although the data set of the SW-Netherlands collected by Jelgersma (1961) has always showed a considerable divergence from the general trend of her own and other sea

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level curves. This problem has hitherto not been solved satisfactorily, although a fourfold explanation was given by van de Plassche (1982).

Probably as a result of these difficulties, the lower Scheldt river area has been somewhat neglected in Dutch sea level research. This may also be due to the fact that reliable water level data for the upstream part of the river Scheldt on Belgian territory have up to now been almost nonexistent.

This paper presents the first results on water level research in the lower Scheldt area in Belgium. An attempt is made to correlate data from Belgium and the SW-Netherlands and to partly explain the problems with the data set from the SW-Netherlands. To facilitate the comparison with the Dutch data, the altitudes will be given relative to the Dutch ordnance datum NAP, which is 2.33 m above the Belgian datum TAW.

2. DATA COLLECTION

Up to now, no accurately levelled paleo-water level data have been published for the lower Scheldt tidal river region on Belgian territory. Therefore, new data had to be collected, consisting mostly of radiocarbon datings of the base of the Holocene peat layer. In a few cases, the original altitude of an already published non-levelled dating could be estimated more or less accurately.

The samples for the newly collected radiocarbon datings were all taken in excavations. In this way, local stratigraphy and possible root contamination could be observed, and an exact altitude determination of the sample could easily be obtained by means of levelling. Moreover, the samples were taken as close to the Late Glacial/early Holocene paleovalley of the river Scheldt as possible, thereby eliminating largely the influence on local peat growth of the slope of the groundwater table towards the river level. As such, it may be assumed that, except for a few cases, these time-depth points actually date river water levels and not local groundwater tables above river level.

Table 1 lists the radiocarbon datings that have been used in this paper to reconstruct part of the Holocene water level evolution in the Scheldt perimarine area. The location of the sampling points has been indicated on Fig. 1.

Most of the samples that have been used consist of fen and/or wood peat, which formed at about average local groundwater level. This coincides approximately with the average local river level. If the area was under tidal influence at the time of peat formation, the peat may be assumed to have formed at about local mean high water (MHW) level (van de Plassche, 1982). In a few cases it was found that peat growth started above local river level. These data will also be discussed below, but they can not be used for the reconstruction of the water level evolution.

The samples from Wintham (index points 1 to 4) and Berendrechtsluis (9 to 12) were taken on the slope of the Late Glacial/early Holocene paleovalley of the river Scheldt. All of them, except no. 4, were situated at the base of the peat layer resting on compaction-free sandy Pleistocene deposits. As the possibility exists that index point no. 4 was affected by compaction of the underlying deposits, it can not be considered a reliable water level indicator.

Index points 11 and 12 from Berendrechtsluis were taken above a pronounced break of slope of the Pleistocene substratum. Peat growth at the location of index point no. 11 took place at about the same time as at sampling point no. 10, although the base of the peat was approximately 0.9 m higher in no. 11.

NO	LOCATION	LAB. NO.	C14-AGE	CAL. AGE		m NAP
1	Wintham 1	IRPA 712	4220± 65	-2915	-2665	-1.83
2	Wintham 2	IRPA 741	5110± 70	-4000	-3790	-2.78
3	Wintham 3	IRPA 740	5550± 75	-4490	-4340	-3.58
4	Wintham 4	IRPA 768	5740± 75	-4722	-4505	-4.40
5	Oosterweel 1	IRPA 713	3890± 65	-2490	-2150	-2.01
6	Oosterweel 2a	IRPA 714a	1840± 55	90	230	-1.06
7	Oosterweel 2b	IRPA 714b	1300± 55	655	810	-1.06
8	Oosterweel 3	IRPA 652	1630± 55	335	510	-1.50
9	Berendrechtsluis 1	IRPA 769	6000± 75	-5034	-4797	-4.24
10	Berendrechtsluis 2	IRPA 770	4480± 70	-3342	-3036	-2.71
11	Berendrechtsluis 3	IRPA 771	4630± 70	-3505	-3343	-1.86
12	Berendrechtsluis 4	IRPA 772	3570± 65	-2031	-1789	-.98
13	Moerzeke-Kastel 1	GrN 5847	4620± 40	-3640	-3155	-2.06
14	Moerzeke-Kastel 2	IRPA 97II	1585± 80	230	610	-1.23
15	Groede	GrN 187	5060±180	-4340	-3390	-3.15
16	Ritthem	GrN 405	5680±120	-4850	-4400	-4.44
17	Waarde	GrN 1112	6330± 85	-5545	-5020	-6.40

Table 1. Radiocarbon datings from the lower Scheldt area which have been used in this paper.

NO: number of index point.

CAL. AGE: range of calibrated age according to Stuiver and Pearson (1986), except for index points 13 to 17, which have been calibrated according to Klein et al. (1982). Negative numbers indicate CAL BC, positive ones are CAL AD.

m NAP: altitude of base of sample in m NAP.

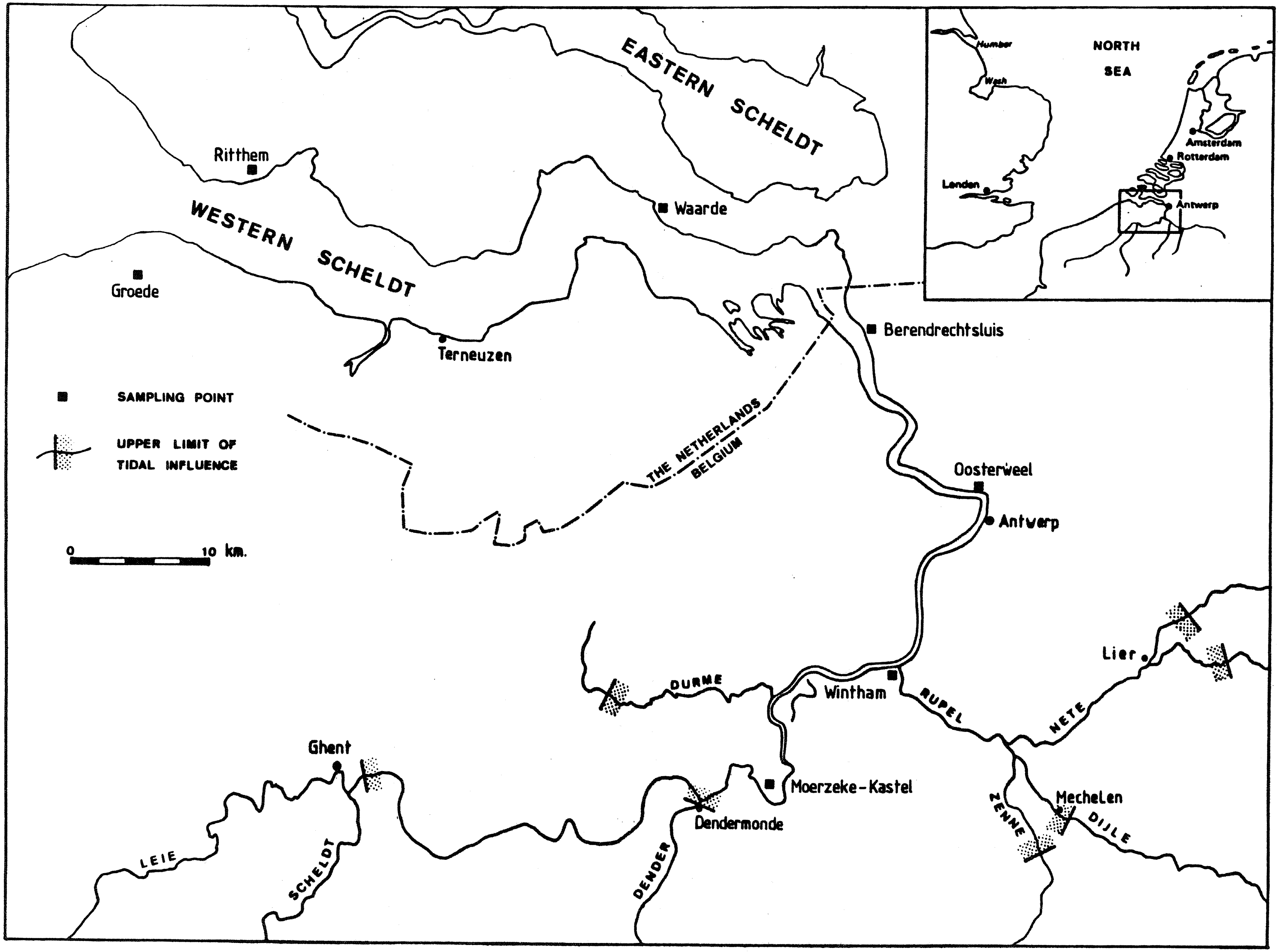

Fig. 1. Study area with location of sampling points for radiocarbon dating.

This can be explained by assuming mesotrophic peat growth due to groundwater seepage on the rather gently sloping sandy surface at no. 11.
This hypothesis is confirmed by pollen analysis of various peat profiles in the former excavation of the Zandvlietsluis, 200 to 300 metres north of the Berendrechtsluis, by Munaut (1967). This author showed that mesotrophic and even oligotrophic peat growth took place at the higher points while at the same time eutrophic fen wood peat was formed in a lower position close to the river Scheldt. Therefore, the lowermost index points 9 and 10 are considered reliable river water level indicators while no. 11 and 12 are not.

The sample Oosterweel 1 (no. 5) was taken at the base of the peat in a small excavation a few kilometres downstream of Antwerp. The small dimensions of this excavation did not permit the reconstruction of the topography of the underlying sandy deposits in the vicinity of the sampling point. The samples Oosterweel 2a and 2b (no. 6 and 7) were taken at the same location as Oosterweel 1 from the top of a peaty clay layer, about 25 cm thick, overlying the peat layer. Both samples are from the same stratigraphic position, but Oosterweel 2b dates the peaty clay itself while Oosterweel 2a dates a piece of wood included in the peaty clay. Oosterweel 2b is considered the most reliable dating, as Oosterweel 2a might represent older (drift)wood. The sample for Oosterweel 3 (no. 8) was collected by T. Oost and K.A.H.W. Leenders in a second excavation, about 100 m northwest of sampling points 5, 6 and 7. It was taken about 30 cm below the top of the peat and therefore is somewhat older than Oosterweel 2b. The exact age of the top of the peat may be assumed to lie in between these two dates, which gives 1450 to 1500 BP.

Index points no. 13 and 14 from Moerzeke-Kastel were collected and published by Verbruggen (1971). Both samples were taken at the same location in an handboring, representing respectively the base and the top of the peat, which overlies sandy Pleistocene deposits. Although no levelling was carried out by Verbruggen, the altitude could be estimated fairly accurately from a detailed altimetric map of the area (Kiden, in press).

The index points 15, 16 and 17 are from Jelgersma (1961) and were collected at the base of the basal peat layer in Zeeland (SW-Netherlands). The data set used by Jelgersma was evaluated a second time by van de Plassche (1982), who showed that from the Zeeland datings only those used in the present paper (15, 16 and 17) could be considered reliable for regional water level reconstruction.

For the period since about 3500 BP, only a few time-depth index points can be derived indirectly from radiocarbon datings and morphological, stratigraphical and historical evidence, resulting however in a somewhat larger margin of error. A detailed description of the procedure, together with the evaluation of a number of paleo-water level indicators is given in Kiden (in press). This will not be repeated here, except for one important time-depth index point, which is discussed below.

The most important water level indicator for the period from 3500 to 1000 BP is the top of the peat in the area of Moerzeke-Kastel. The peat layer thins out on compaction-free Pleistocene deposits at a maximum altitude of about -0.30 m NAP (±0.40 m). The compacted top of the peat has been dated at 1585±80 BP at Moerzeke-Kastel (index point no. 14), which gives a calibrated age of 230 to 610 CAL AD (Klein et al., 1982). This time-depth index point will be used as no. 18 (-0.30±0.40 m NAP, 1585±80 BP).

3. INTERPRETATION

On the basis of the available time-depth index points, the gradient line of the river Scheldt at a given time can be reconstructed. Fig. 2 shows the approximate position of the gradient lines at 5500, 5000, 4500 and 4000 BP. For the period prior to 5500 BP and after 4000 BP too few data are available, except for the upstream area from Moerzeke-Kastel to Wintham, where the local water level evolution could be reconstructed up to the present day. The resulting curve is shown in Fig. 3, together with a smoothed mean sea level (MSL) curve partly based on van de Plassche (1982), and a derived coastal MHW curve.

This approach makes it possible to estimate former river gradients and the presence of floodbasin effects in the tidally influenced part of the river Scheldt. The floodbasin effect (van Veen, 1950; Zonneveld, 1960) results in a relative lowering of the tidal amplitude and local high water level in an upstream direction in a tidal river or estuary. This is mainly caused by friction effects and the presence of storage basins within the estuary, which can absorb part of the flood volume entering the mouth of the estuary.

As can be seen in Fig. 2, the 5500 BP gradient line shows a gentle slope towards the contemporaneous MHW level at sea. It can be assumed that up to that time, no clear floodbasin effect was present in the river Scheldt on Belgian territory. For the period prior to 5500 BP, there are indications that the gradient of the river between Berendrechtsluis (index point 9) and Waarde (point 17) was much steeper, although van de Plassche (1982) noticed the presence of a considerable floodbasin effect west of Waarde. However, the coexistence of a steep river gradient in one reach and a floodbasin effect downstream of that reach is not impossible, as is shown by van de Plassche (1984).

The 5000 BP gradient line seems to be remarkably flat as far upstream as Berendrechtsluis, which indicates that the river slope was at least attenuated by a floodbasin effect further downstream. A similar case has been described in the Rhine-Meuse delta for the period around 6050 BP by van de Plassche (1984). In this respect, it is interesting to note that Minnaert and Verbruggen (1986) found a thin clay layer, dated at about 4900 BP, intercalated in the peat at Doel (on the left bank of the Scheldt, a few kilometres upstream of Berendrechtsluis). Pollen analysis showed a relatively high Chenopodiaceae content of the clay, pointing to the possible presence of brackish water and hence of tides as far upstream as Doel in this period.

At 4500 and also at 4000 BP, a marked floodbasin effect was present in the area upstream of the Dutch-Belgian border. The area of maximum lowering of the MHW level seems to have shifted slightly upstream from 5000 to 4000 BP, although this is difficult to assess due to the insufficient number of time-depth data.

From 4000 to about 1500 BP, very few reliable time-depth data are available. Nevertheless, it will be argued below that an important floodbasin effect has been present in the lower Scheldt area upstream of the Dutch-Belgian border since about 4500 BP up to the Middle Ages.

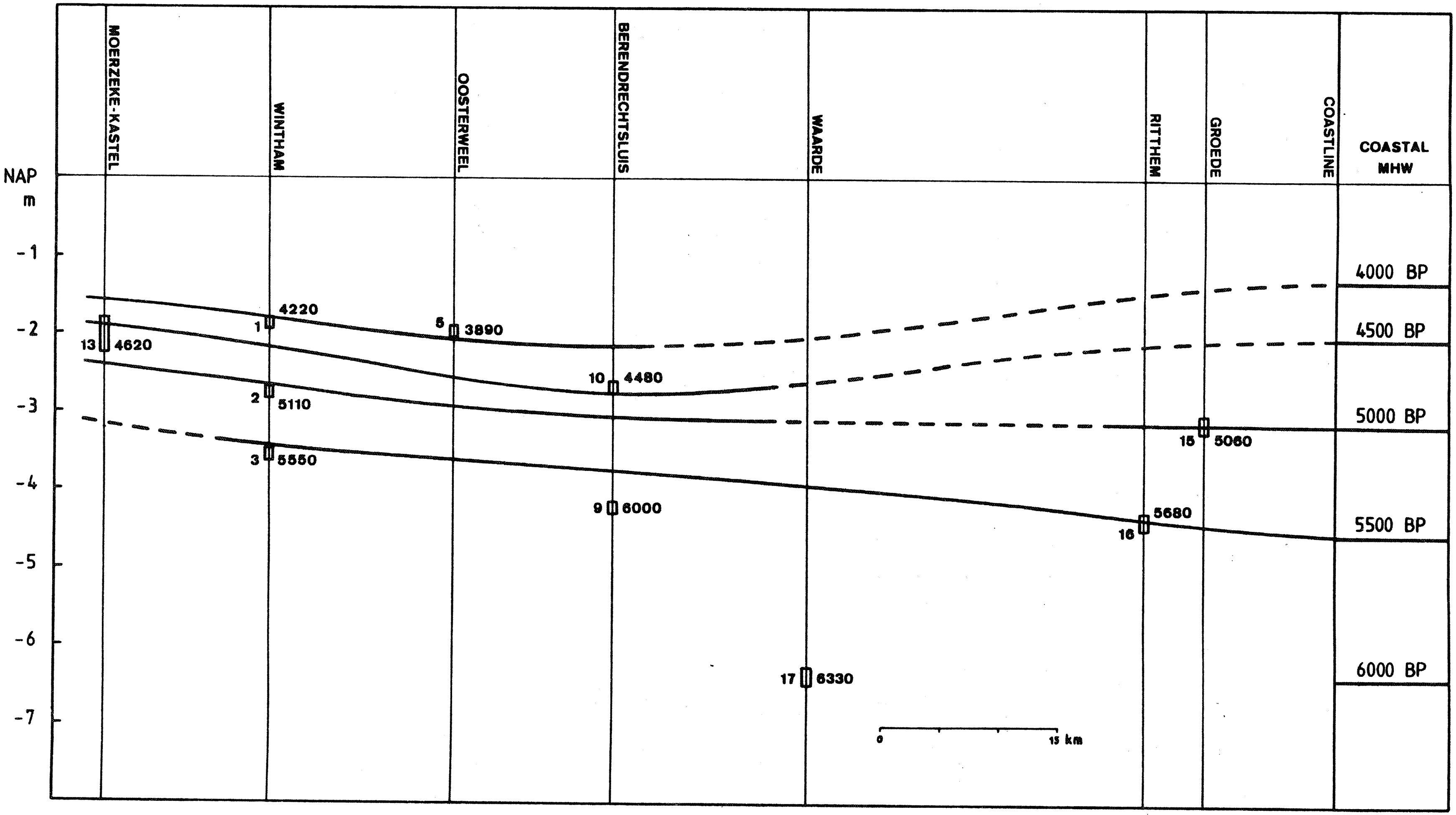

Fig. 2. Reconstruction of gradient lines in the lower Scheldt perimarine area at 5500, 5000, 4500 and 4000 BP.

The general trend of the water level fluctuations in the area from Moerzeke-Kastel to Wintham could be reconstructed from about 5500 BP up to the present day (Fig. 3). For this purpose, the radiocarbon ages were converted into historical years. For the newly collected datings, this was done by M. Van Strydonck from the IRPA, on the basis of the publications of Stuiver and Pearson (1986). For the published datings from Verbruggen (1971) (index points 13 and 14), the calibration tables given by Klein et al. (1982) were used, which resulted in a somewhat larger margin of error.

As can be seen in Fig. 3, local MHW level (curve III) is situated well below coastal MHW level (curve II) from about 3000 CAL BC to about 1700 CAL AD. A more detailed reconstruction of curve III for the period from 2500 CAL BC to about 500 CAL AD can not be given due to the lack of data. Nevertheless it is clear that during this period, but also before 3000 CAL BC already, the rate of rise of local MHW was slower than that of coastal MHW. This can be ascribed to the presence of an important floodbasin effect in the river Scheldt. For the area from Moerzeke-Kastel to Wintham, the floodbasin effect probably reached a maximum around 500 CAL AD, when local MHW level stood about 1 m lower than contemporaneous MHW level at sea.

During the period from 500 to 1000 AD, the rate of rise of local MHW level was increasing, probably as a result of morphological changes affecting the storage capacity of the floodbasin areas in the Scheldt estuary (increased sedimentation, widening and/or displacement of the estuary mouth). Around 1000 AD the start of the first embankments and probably also the formation of the Western Scheldt estuary caused a further increase in the rate of rise of local MHW level. The rapid rise of MHW level along the length of the Scheldt estuary since that date is fairly well documented by historical evidence of storm floods and embankments. The

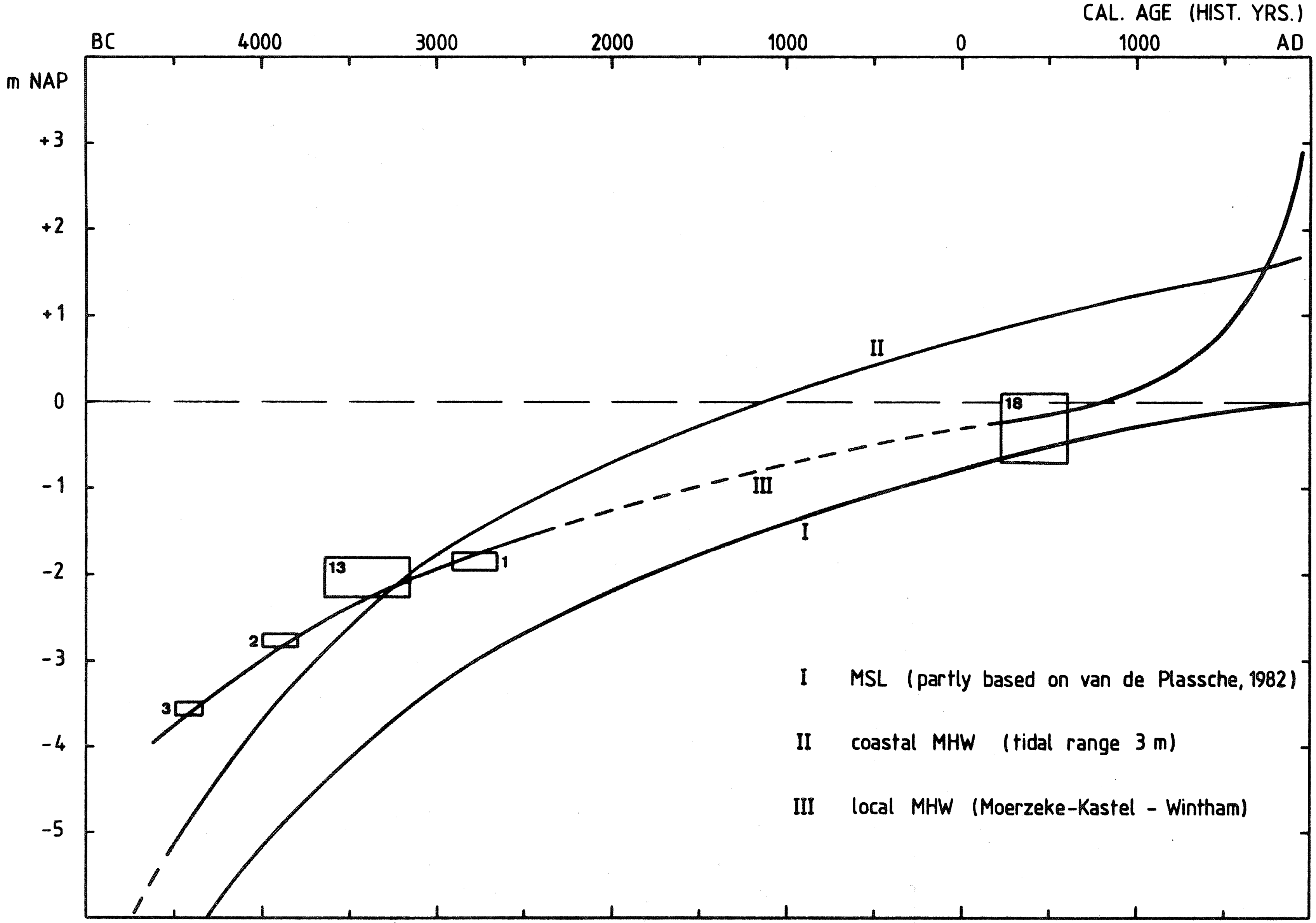

Fig. 3. Local water level evolution in the area from Moerzeke-Kastel to Wintham in relation to MSL and coastal MHW level rise.

reconstruction of local MHW level since about 1100 AD is in fact based on sedimentation levels in various polders of known age, while from 1860 AD onwards tidal observations have been used.
This rapid rise of MHW level and the upstream penetration of tidal action since about 1000 AD may thus be attributed to a decreasing floodbasin effect in the Scheldt estuary. At present, tidal influence reaches as far upstream as Ghent, about 160 km from the coastline.

4. DISCUSSION

The above mentioned considerations are based on a number of assumptions concerning former tidal range, rate of subsidence and geomorphology in the lower Scheldt region, which will be discussed below.

The reconstruction of the MHW level at sea (Fig. 2 and 3) is based on the MSL curve of van de Plassche (1982), to which half the tidal range has been added. This was taken to be 1.5 m, which seems a rather conservative estimate (tidal range: 3 m). At present, tidal amplitude at the mouth of the Scheldt estuary is almost 4 m. However, tidal range at the mouth was probably smaller at an earlier date, as the mouth of the former Scheldt estuary was situated more to the north during most of the Holocene (Pons et al., 1963; Jelgersma et al, 1979) and tidal range decreases northward along the coastline. If tidal amplitude was larger than 3 m, the floodbasin effect at a given time would have been more pronounced and the river slope less strong than described above.

The implicit assumption is made that tidal range at a given point along the coast has remained constant throughout the period under consideration. It is clear, however, that a northward or southward shift of the mouth of the river Scheldt could have caused a tidal range change in the estuary due to the tidal amplitude gradient along the coastline. Up to now, very little is known about this effect, which depends as much on the geomorphological development of the Scheldt estuary as on tidal range changes through time. Nevertheless, the formation of the Western Scheldt estuary to the south of the Eastern Scheldt seems to have played a major part in the increase of local MHW level and the upstream penetration of tidal action in the river since about 1000 AD.

In this paper, the reconstruction of coastal MSL has been based on the curves presented by van de Plassche (1982), which are valid for the western and northern Netherlands. It is possible, however, that these curves do not represent relative sea level changes in the SW-Netherlands, as this region probably underwent less tectonic subsidence (Jelgersma, 1961). In order to give an approximation of the difference in subsidence between the two regions, the time-depth data from Belgian territory were evaluated. The compaction-free datings that have been used for the reconstruction of former gradient lines (Fig. 2) have been plotted in a time-depth diagram together with the MSL curve of van de Plassche (1982) (Fig. 4).

As can be seen in Fig. 4, the time-depth points from Belgium and from the SW-Netherlands are all situated above the MSL curve of van de Plassche (1982), with the difference in altitude increasing with age. This may be caused by differential tectonic subsidence, but is also likely to be due to an increasing floodbasin effect in the area, resulting in peat formation closer to MSL. Index points 10 and 5, however, are resp. only 0.8 and 0.7 m above contemporaneous MSL.

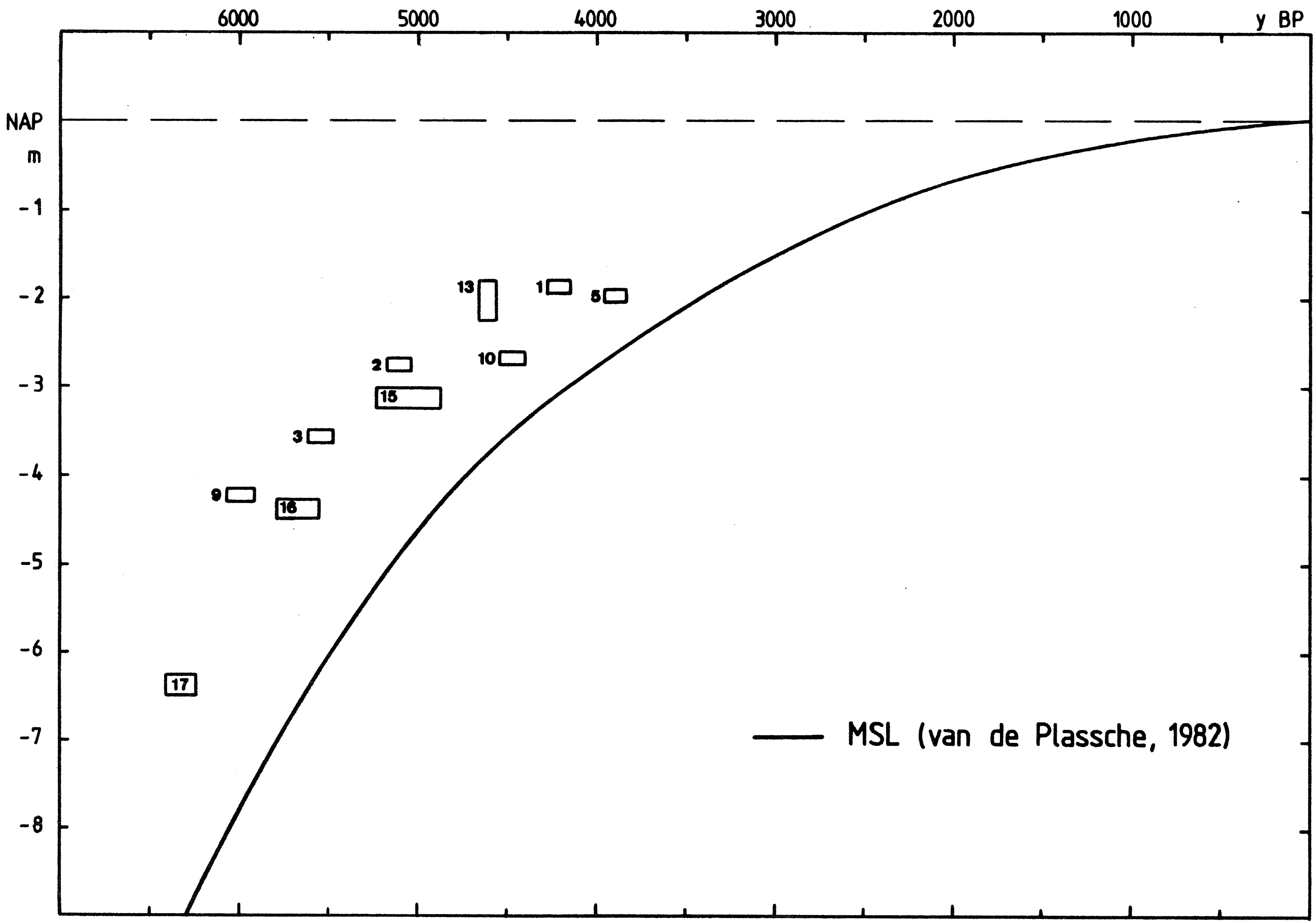

Fig. 4. Time-depth points in the lower Scheldt perimarine area plotted with respect to the MSL curve of van de Plassche (1982).

Therefore, and using the calibrated age of index point 10, the differential tectonic subsidence of the western and northern Netherlands with respect to the lower Scheldt river region can not have been greater than approximately 0.8 m over the last 5000 historical years. Extrapolating linearly up to the present day, this yields a differential tectonic subsidence of about 1.6 cm/century. This agrees well with the results obtained by Mostaert (1985), who calculated a minimum value of 1.5 cm/century over the last 7000 years. A correction for this differential subsidence would result in a more pronounced floodbasin effect and a smaller river slope at a given time than was found in the present study, with a difference that increases with age.

5. CONCLUSIONS

Although up to now reliable water level data from the lower Scheldt river area remain relatively scarce, the model presented above seems to be rather well established. It is likely that refinements of the model will be necessary when more data become available on former coastal tidal amplitudes and differential tectonic subsidence. It may be assumed, however, that these refinements will only affect the exact dating of the events and not the general line of reasoning.

The data indicate the presence of a floodbasin effect in the lower Scheldt river on Belgian territory since about 5000 to 4500 BP. This seems to have controlled the rise of local water level up to 500 to 1000 AD. The rapid rise of local MHW level in the lower Scheldt since about 1000 AD, for which hitherto no satisfactory hydrodynamic explanation could be given, may be attributed to the disappearance of this floodbasin effect.

The present data set for the lower Scheldt area shows an important gap between about 4000 BP to 1500 BP. In the near future, it will be attempted to fill this gap by careful selection of the sampling sites. This will probably enable a further refinement in space and time of the model presented in this paper.

ACKNOWLEDGMENTS

The radiocarbon datings were carried out at the Koninklijk Instituut voor het Kunstpatrimonium (KIKP-IRPA), Brussels, under the direction of Mr. M. Van Strydonck and Ms. M. Dauchot, who I would like to thank here for the smooth cooperation. I want to express my gratitude to Prof. Dr. G. De Moor and my colleagues of the Laboratory of Physical Geography from the State University of Ghent, who provided the necessary support for the research. To F. Mostaert and C. Verbruggen, I am greatly indebted for the interesting discussions on floodbasin and Scheldt related topics. The Instituut tot aanmoediging van het Wetenschappelijk Onderzoek in Nijverheid en Landbouw (IWONL-IRSIA) offered a grant which made this study possible.

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Manuscript received January 1987.

OBSERVATIONS ON THE TRANSITION FROM CALAIS DEPOSITS TO SURFACE PEAT IN THE WESTERN BELGIAN COASTAL PLAIN - RESULTS OF A PALEOENVIRONMENTAL DIATOM STUDY -

Luc Denys (*)

ABSTRACT

The transition from the clastic Calais deposits to the main body of the Holland peat in the western Belgian coastal plain was investigated by means of diatom analyses of seven cores, from close to the present coastline up to the polder border. At all sites the peat was found to rest upon mudflat and saltmarsh deposits and not upon lagoonal sediments, as often supposed. Peat formation in the backswamps was apparently induced by the waterlogging of silted-up areas due to a considerable supply of fresh water, mainly of inland origin.

1. INTRODUCTION

Since 1981 the author has been carrying out diatom analyses on Quaternary sediments from the western Belgian coastal plain. The primary goal of this study is to explore the possibilities of diatom analysis in accurately defining former depositional environments in the area and thus to contribute to our knowledge of its geological history. Emphasis is placed upon the Holocene strata which provide a rather well preserved record of the local land/sea interactions during the past 8000 years.

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Of special interest in this context are the environmental changes taking place at transgressive and regressive overlaps. Firstly by their importance to dynamic geomorphology and ecology and secondly because of the problems encountered in referring the related facies to particular sedimentary environments.

In this paper the regressive overlap between the Atlantic to Subboreal, largely clastic, beds known as Calais deposits and the overlying main body of the Holland Peat member, further referred to as "surface peat", is considered. In particular the general environment prior to this peat formation is focussed upon.

2. GENERAL CONTEXT

In large parts of the western Belgian coastal plain, the Calais deposits (known in the regional literature also as "Atlantische waddenafzettingen" or "sables pissarts") are covered by a generally well developed peat layer, the surface peat. The texture of the Calais deposits, which may contain several thinner peaty layers themselves (the one closest to the surface peat most often yields dates from 6000 to 6400 B.P.), varies strongly from place to place and with depth (in general the sediments are finer towards the polder border and more sandy seawards: Baeteman 1977, 1978, Maréchal 1953, Moormann 1951, Paepe & Baeteman 1979). Just below the surface peat, a very heavy greyish-blue clay is mostly encountered which may attain a thickness of a few dm. This layer, known by the local farmers as "bonk" often contains rhizomes of the common reed (_Phragmites australis_ (Cav.) Trin.), is poor in lime and has reducing characteristics (high content of FeS which oxidises to yellow $Fe_2(SO_4)_3$ when exposed to the air, smell of H_2S, presence of black reduction spots and vivianite). The peat immediately overlying it represents a Phragmitetum - Magnocaricetum, which grew in very shallow (marshy), perhaps slightly brackish conditions (Baeteman & Verbruggen 1979, Stockmans et al. 1948, Stockmans & Vanhoorne 1954). Occasionally, along former watercourses, this reed peat is lacking and a coarse fen-wood or strongly humified peat is found covering the "bonk". The further development of the peat generally tends towards a distinct oligotrophication. There is evidence that it was nevertheless governed, at least partly, by a rising sea-level as it encroached upon the bordering area of Pleistocene coversands, largely developed in minerotrophic conditions and reaches a considerable thickness (Baeteman 1981, Baeteman & Verbruggen 1979, Tavernier & Moormann 1954). The onset of the surface peat formation was not synchronous throughout the area, but ranges from about 5300 to 4000 years B.P. Nowadays the altitude of the peat base lies at approximately -2.2 to +1.8 m TAW.

3. CONCEPTS ON THE ENVIRONMENT LEADING TO PEAT FORMATION

In the literature two main concepts can be found concerning the clayey facies topping the Calais deposits. While both agree on the large-scale events leading to the formation and development of the peat (e.g. the formation of a protective barrier or dune belt, excluding the marine influence from the hinterland, combined with a rising water table), different opinions exist on the precise environment in which the reed peat started growing.

A first concept was introduced by Blanchard (1906), who considered the clay as a marine lagoonal facies, an idea apparently going back as far as the early work of Belpaire (1827). This view was more or less followed by Halet (1931), who interpreted the muddy sediment as a creek or lagoon deposit, and by Paepe (1960). More recently the idea was reintroduced by Baeteman in several contributions. Baeteman &

Verbruggen (1979) and Baeteman (1981), explain the process of peat formation by a simultaneous shallowing of large undeep lagoons and a decreasing marine influence, due to morphological changes of the plain and coastline, leading to the invasion of the lagoons with Phragmites. These lagoons are defined as shallow depressions permanently filled with water of max. 0.1% salinity (to allow for the reed growth) and separated from the open sea by dunes, isles, sand barriers or saltmarshes, but often connected to it by some channel (Baeteman 1981). Curiously, this pronounced subtidal environment is further synonimized with reedmarsh (Baeteman l.c., 1981bis). The subaquatic deposition of the clay is argumented by its lithological characteristics i.c. the fine texture and high organic content, the homogeneous structure, the occurrence of reed rhizomes and reduction spots.

A second opinion was advocated by Cornet (1927) who believed that the clay was of tidal flat origin and that silting up of the mudflats and the shelter provided by a dune belt lead to the growth of the reedmarsh. This idea was taken up by the school of pedologists from Ghent (Moormann 1951, Moormann & Ameryckx 1950, Tavernier 1947, 1948, Tavernier & Moormann 1954, T'Jonck & Moormann 1962) and apparently initially even by Baeteman (1978) when she wrote "...the mudflats and marshes developed into slightly brackish swamps in which reed could start growing...".

In a previous paper (Denys 1985) the paleoenvironments leading to surface peat formation were discussed for a core from Slijpe (Fig. 1, nr. 8) where the base of the peat was dated at 4860$\pm$70 B.P. (IRPA 518). Herein it was concluded that the diatoms from the clayey facies pointed to intertidal rather than to lagoonal conditions. Moreover a succession from high tidal flat to saltmarsh and finally to fen-carr (peat) was proposed as the most probable evolution. The very local nature of these results prevented any consideration of their spatial importance.

4. METHODS

Diatom assemblages from the muddy clay facies below the surface peat were investigated from seven mechanically drilled cores which allowed for the sampling of undisturbed sequences. The location of these cores is indicated in Fig. 1.
Only the transitional part towards the surface peat is considered. The lithology of the sequences is shown in the corresponding figures (2-8). Both transitions with many Phragmites rhizomes in the clay and a reed peat at the top (Pervijze, Wolvenest, Westende, Vliegveld) as sequences topped by a humified peat, containing no recognizable Phragmites remnants (Merkem, Oostkerke, Spoorweg), are represented. In three cores (Oostkerke, Spoorweg and Vliegveld) a thin intercalated peat layer occurs in the clay close to the surface peat (cfr. 2.). These horizons are not to be confused with the "basal peat" (= peat at the base of the Holocene sequence) situated at still greater depth.

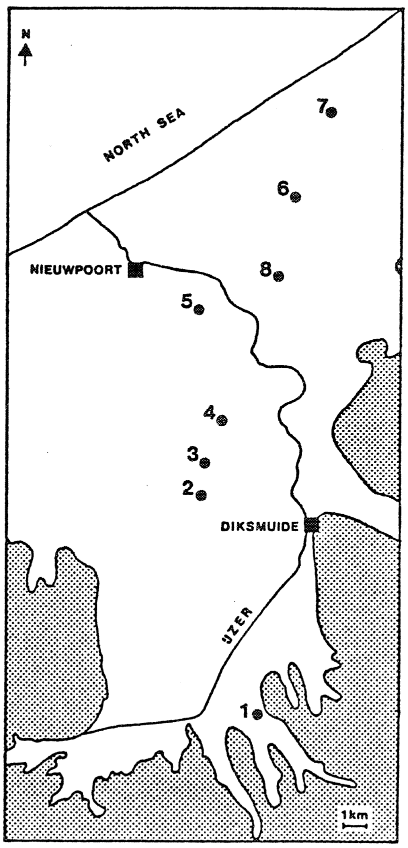

Fig. 1: Map of the study area with indication of the coring sites. Area of outcropping coversands shaded. 1. Merkem, 2. Oostkerke, 3. Pervijze, 4. Spoorweg, 5. Wolvenest, 6. Westende, 7. Vliegveld, 8. Slijpe.

Relative frequencies of the taxa encountered were estimated by counts of at least 400 valves when possible. Sample treatment largely followed the methods outlined by Denys (1985). In this paper only those taxa accounting for the most substantial part of the assemblages are mentioned.

5. RESULTS

5.1. Merkem, core 66W nr. 144 (Fig. 2)

At the base of the clay (645-638 cm) a mixture of brackish and polyhalobous tidal flat diatoms (Opephora marina, Achnanthes delicatula, Cymatosira belgica, Rhaphoneis minutissima) and Fragilaria species (F. pinnata, F. construens var. venter) is found. These Fragilaria's occur in fresh to quite saline, shallow and quiet water and are indicative of a strongly variable water chemistry. Often they are considered as typical pioneers (Haworth 1976) or as indicators of basin isolation (Stabell 1985). In this context, their habitat is expected to have been in the very shallow, rather ephemeral pools and patches of water that cover large parts of badly drained high saltmarshes. Nitzschia frustulum, which is also rather numerous, prefers similar environments. Apparently we are dealing here with a transition from saltmarsh to mudflat.

In the middle, the polyhalobous species Cymatosira belgica and Rhaphoneis minutissima increase strongly, indicating a return to completely intertidal conditions (638-633 cm). The presence of Nitzschia hungarica however points to a sheltered and still elevated position (Brockmann 1950).

The transition to the peat (634-627 cm) shows a luxurious development of Fragilaria construens var. venter, accompanied by F.atomus and F. construens var. subsalina, both diatoms from slightly brackish water (Mölder & Tynni 1970). Here the

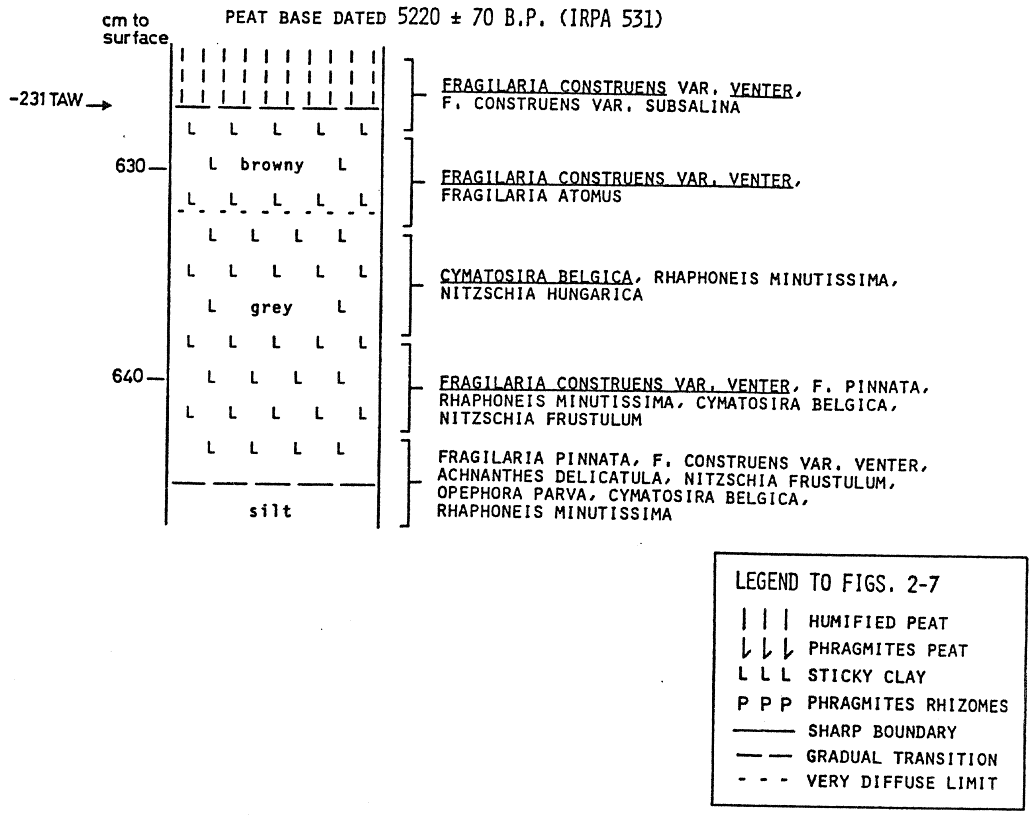

Fig. 2: Lithology and most important diatoms (predominant taxa underlined) of the Calais/peat transition in the core Merkem.

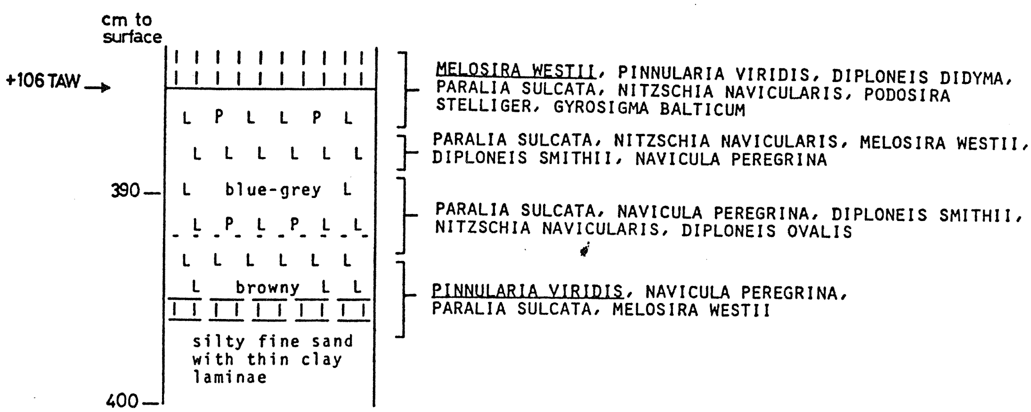

Fig. 3: Lithology and most important diatoms (predominant taxa underlined) of the Calais/peat transition in the core Oostkerke.

silting-up leads to a saltmarsh and finally even to a fen in which a freshwater peat developed.

5.2. Oostkerke, core 51W nr. 135 (Fig. 3)

At some 10 cm below the surface peat a thin peaty intercalation is found. The diatom assemblages from the lower part of the muddy clay which covers it (396-389 cm) are dominated by benthic oligo- and mesohalobous species such as Pinnularia viridis, Diploneis ovalis (both oligohalobous), Navicula peregrina, Nitzschia navicularis and Diploneis smithii (mesohalobous). Melosira westii and Paralia sulcata, benthic-planktonic polyhalobous diatoms, are also common. The environment indicated is a brackish tidal flat near the local MHW level formed by the flooding of an almost fresh-water marsh.

In the blue-grey clay, polyhalobous taxa (Melosira westii, Paralia sulcata, Podosira stelliger) are more prevalent and indicators of slightly brackish conditions (Navicula peregrina, Diploneis ovalis) decrease, pointing to a somewhat increased marine influence (389-384 cm). The environment remains intertidal, though of somewhat higher salinity than before. The appearance of Podosira stelliger and Gyrosigma balticum (384 cm) is most probably related to the final terrestrialisation.

5.3. Pervijze, core 51W nr. 134 (Fig. 4)

Below the peat a layer of silty clay is found (444-413 cm) resting upon a silty fine sand. Both contain very similar, poorly preserved assemblages. Only four taxa are of considerable importance: the brackish Nitzschia navicularis and Diploneis didyma and the allochthonous marine Melosira westii and Podosira stelliger. While Nitzschia navicularis and Melosira westii are dominant throughout, Diploneis didyma retains lower frequencies and Podosira stelliger is rather numerous only from 425 to 414 cm.

In the uppermost few cm the sediment changes to a peaty clay in which Melosira westii prevails, most probably due to its superior preservation qualities. Some indicators of terrestrial conditions (Melosira roeseana and Hantzschia amphioxys) appear here in low numbers.

The sediments appear to have been deposited near the limit between tidal flat and saltmarsh at a well-drained site. Immediately prior to the peat growth rather dry (saltmarsh) conditions probably existed.

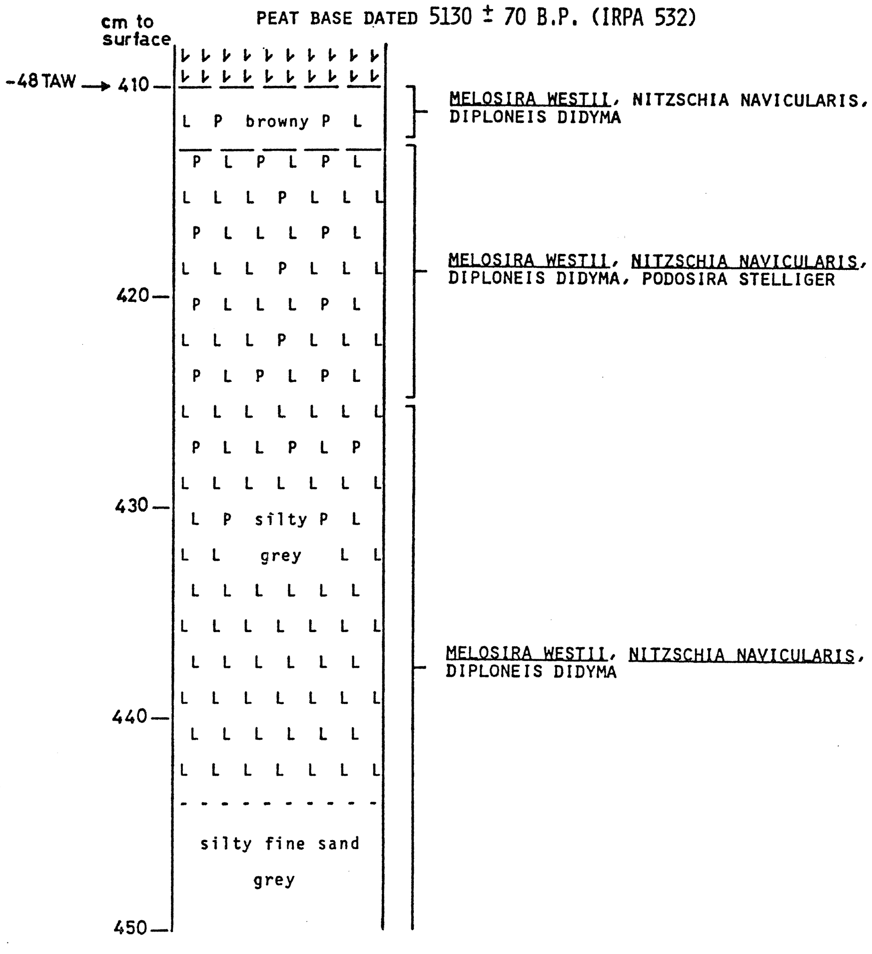

Fig. 4: Lithology and most important diatoms (predominant taxa underlined) of the Calais/peat transition in the core Pervijze.

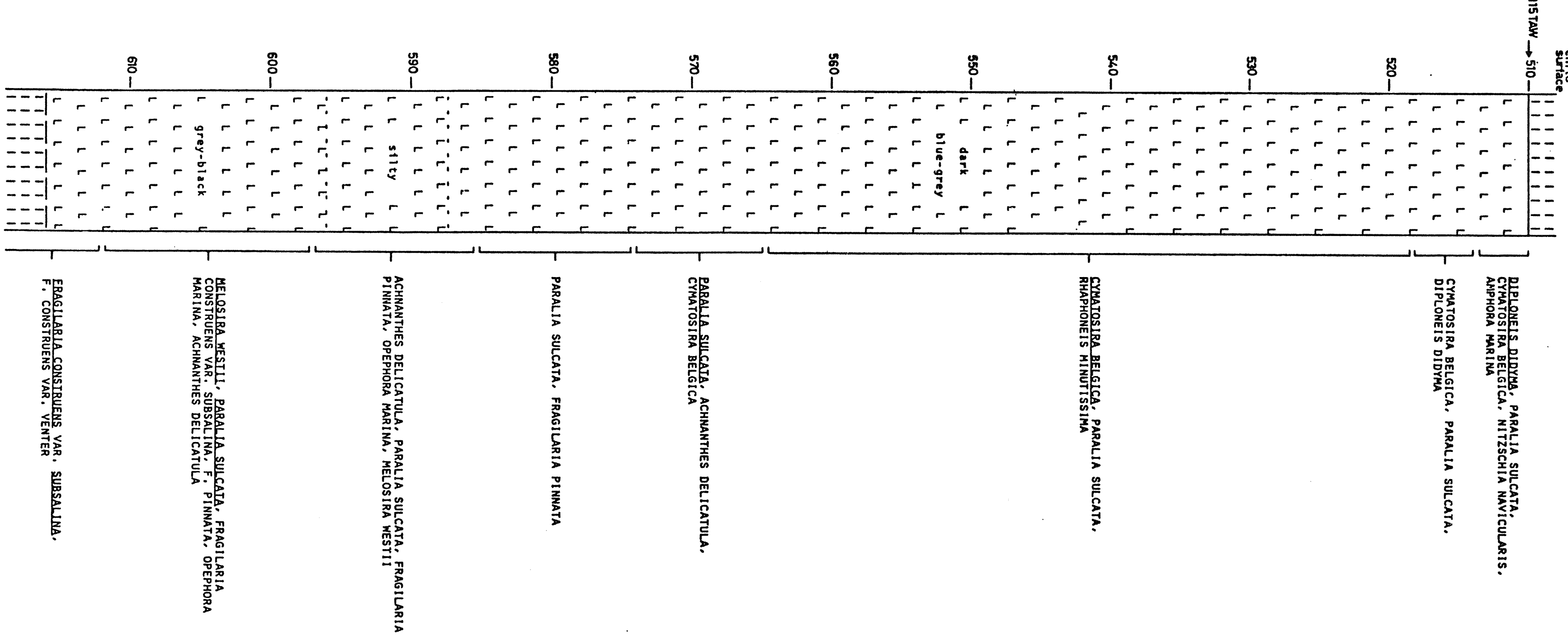

Fig. 5: Lithology and most important diatoms (predominant taxa underlined) of the Calais/peat transition in the core Spoorweg.

5.4. Spoorweg, core 51W nr. 143 (Fig. 5)

The sticky organic clay covering the lower peat layer (616-613 cm) contains a Fragilaria-dominated assemblage from shallow, slightly brackish water which rapidly comes under tidal influence as shown by the increase of Melosira westii and Paralia sulcata (613-597 cm).

Next some small diatoms which grow attached to sediment particles (Achantes delicatula, Opephora marina) become more abundant (597-585 cm). As expected their increase coincides with the more silty facies.

After this marginally more energetic phase, a return to very calm conditions occurs. Paralia sulcata and Fragilaria pinnata are now dominant (585-575 cm). F. pinnata, although generally considered a freshwater diatom, is known to thrive well in brackish water and even at high salinities (Hargraves & Guillard 1974, Hendey 1964, Sundbäck 1983, etc.), as is apparently the case here.

From then on an increase of the polyhalobous Cymatosira belgica and Rhaphoneis minutissima is observed. In most of the upper half of the sequence (575-517 cm) these species dominate and point to almost marine intertidal conditions.
Only in the upper few cm a decrease of the salinity is seen. Firstly by the appearance of the benthic, marine-brackish Diploneis bombus (517-514 cm) and finally by high numbers of the mesohalobous species Diploneis didyma and Nitzschia navicularis (514-510 cm). This again illustrates the silting-up of the flat and the consequent decline of the marine influence.

5.5. Wolvenest, core 36W nr. 156 (Fig. 6)

The slightly silty lower part (415-405 cm) is characterised by tidal flat assemblages composed mainly of polyhalobous taxa

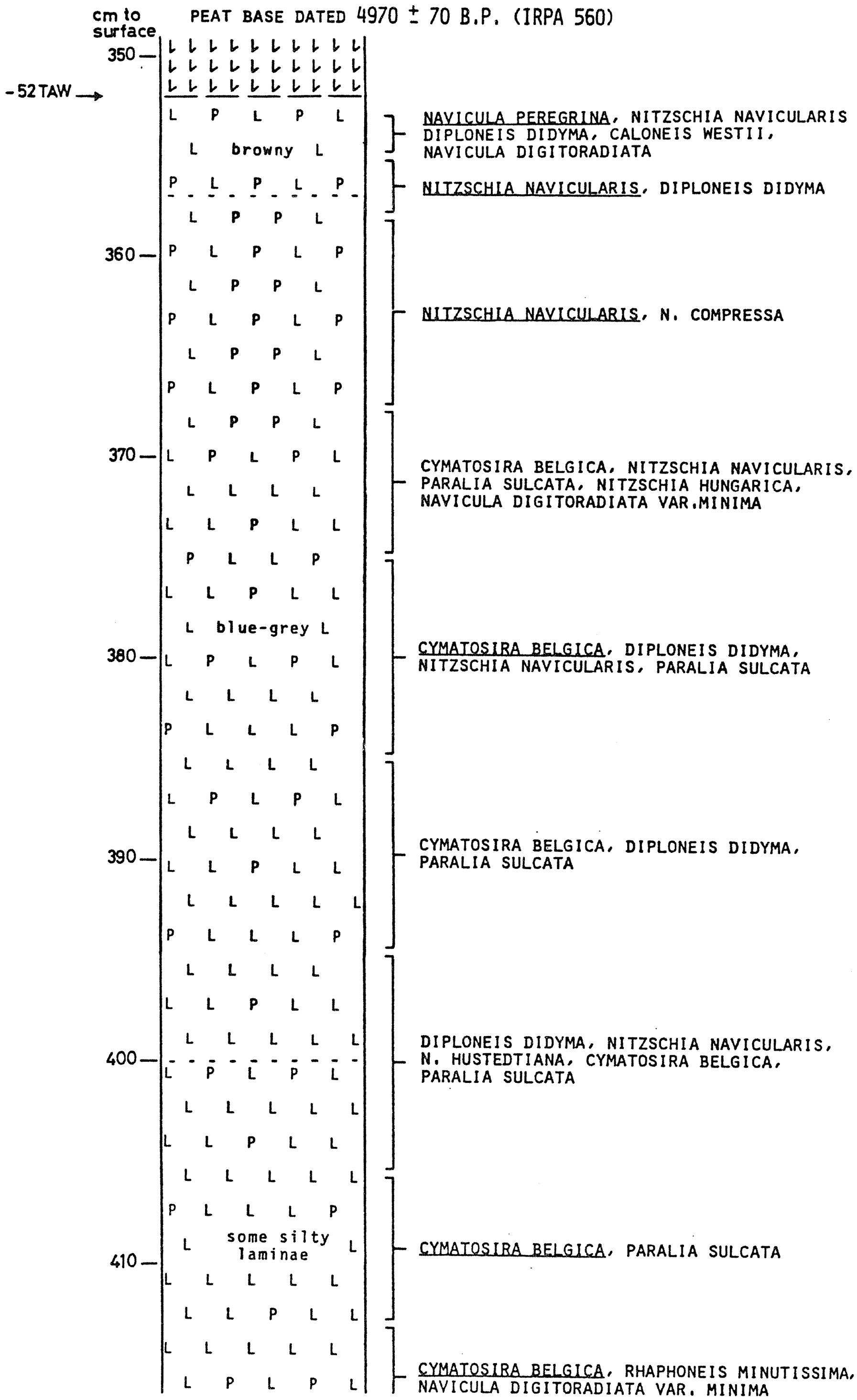

Fig. 6: Lithology and most important diatoms (predominant taxa underlined) of the Calais/peat transition in the core Wolvenest.

and dominated by Cymatosira belgica. Where the sediment changes to a purer clay (405-395 cm), more brackish conditions are revealed by the abundance of Diploneis didyma, Nitzschia navicularis and N. hustedtiana. This last species, classified as mesohalobous and euryhaline by Salah (1952), is not uncommon in the brackish water of estuaries and of saltmarshes in particular (Giffen 1971, Sullivan 1975, 1978).

This brackish character soon becomes less pronounced when Cymatosira belgica regains some of its previous importance (395-368 cm). Nevertheless several of the mesohalobous taxa remain abundant throughout as is the case for Diploneis didyma and Nitzschia navicularis. Higher-up Nitzschia hungarica and Navicula digitoradiata var. minima join in. Simonsen (1962) already stressed the wide salinity tolerance of Nitzschia hungarica, its appearance here nevertheless indicates the beginning of a salinity decline.

Starting from the point at which Phragmites rhizomes become numerous, Nitzschia navicularis becomes the dominant species accompanied by N. compressa and Diploneis didyma (368-354 cm). These diatoms are found on flats and saltmarshes especially on those parts close to the MHW level (Brockmann 1950, Colijn & Koeman 1975, Round 1960). They demonstrate the terminal silting-up. Only at the top Nitzschia navicularis is overwhelmed by Navicula peregrina, a species requiring even lower salinities. Caloneis westii and Navicula digitoradiata, both euryhaline brackish diatoms, also profit from the favourable conditions created immediately prior to the peat formation.

5.6. Westende, core 36E nr. 122 (Fig. 7)

In the lowermost part (510-501 cm) mainly mesohalobous diatoms from the higher tidal flats and from saltmarshes (Diploneis didyma, Nitzschia navicularis, N. compressa) occur. Initially they are associated with species preferring a higher salinity

(<u>Paralia sulcata</u>, <u>Rhaphoneis amphiceros</u>).

Higher-up polyhalobous tidal flat diatoms are predominant (<u>Cymatosira belgica</u>, <u>Paralia sulcata</u>, <u>Rhaphoneis amphiceros</u>), indicating a period of stronger marine influence (501-465 cm).

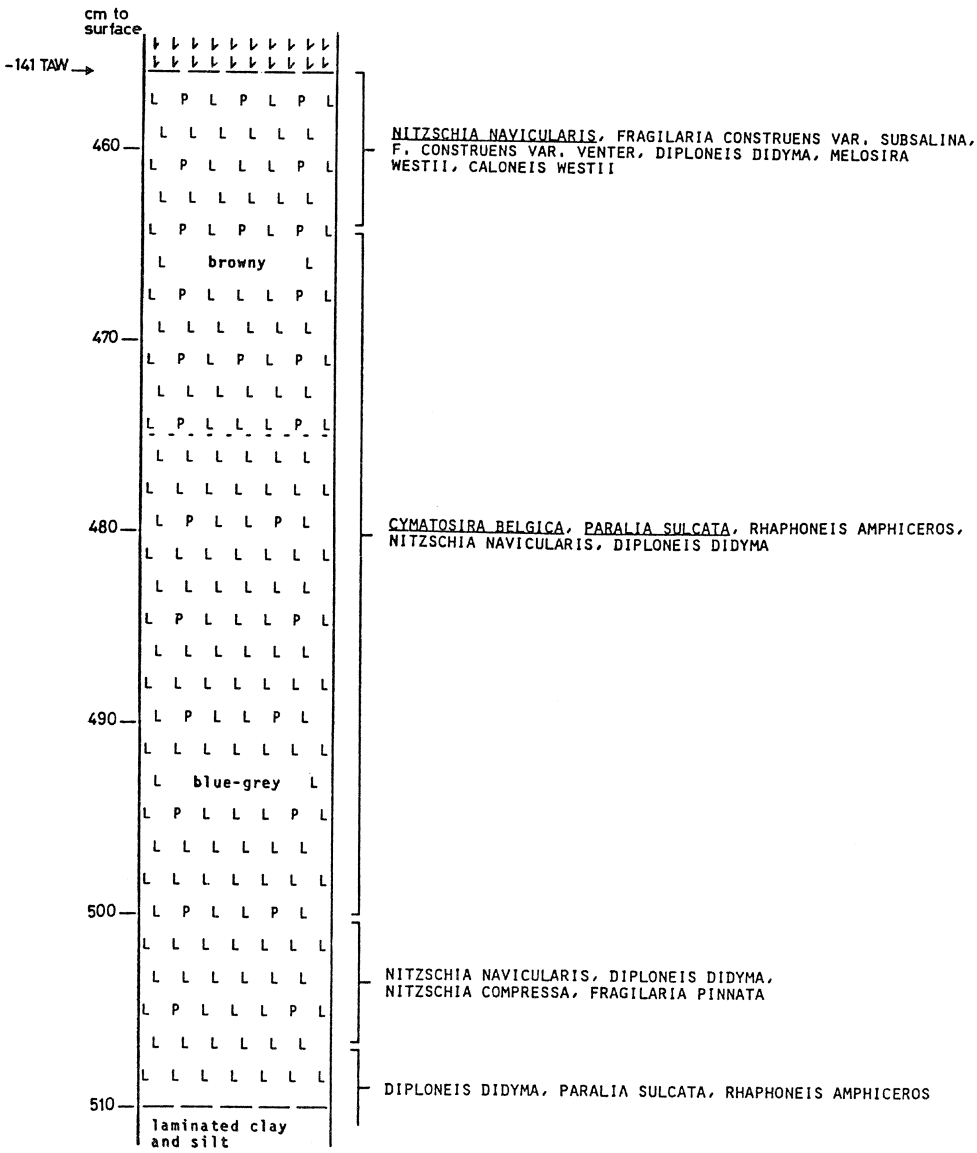

Fig. 7: Lithology and most important diatoms (predominant taxa underlined) of the Calais/peat transition in the core Westende.

Immediately below the surface peat mesohalobous species return (Nitzschia navicularis, Caloneis westii) together with Fragilaria construens var. venter and var. subsalina, pointing to a gradual freshening and the transition from high flat to marsh (465-458 cm).

5.7. Vliegveld, core 21E nr. 221 (Fig. 8)

At the top of the thin peat layer below the clay (575 cm) an assemblage dominated by Fragilaria construens var. subsalina and Cymatosira belgica occurs. Considering their salinity

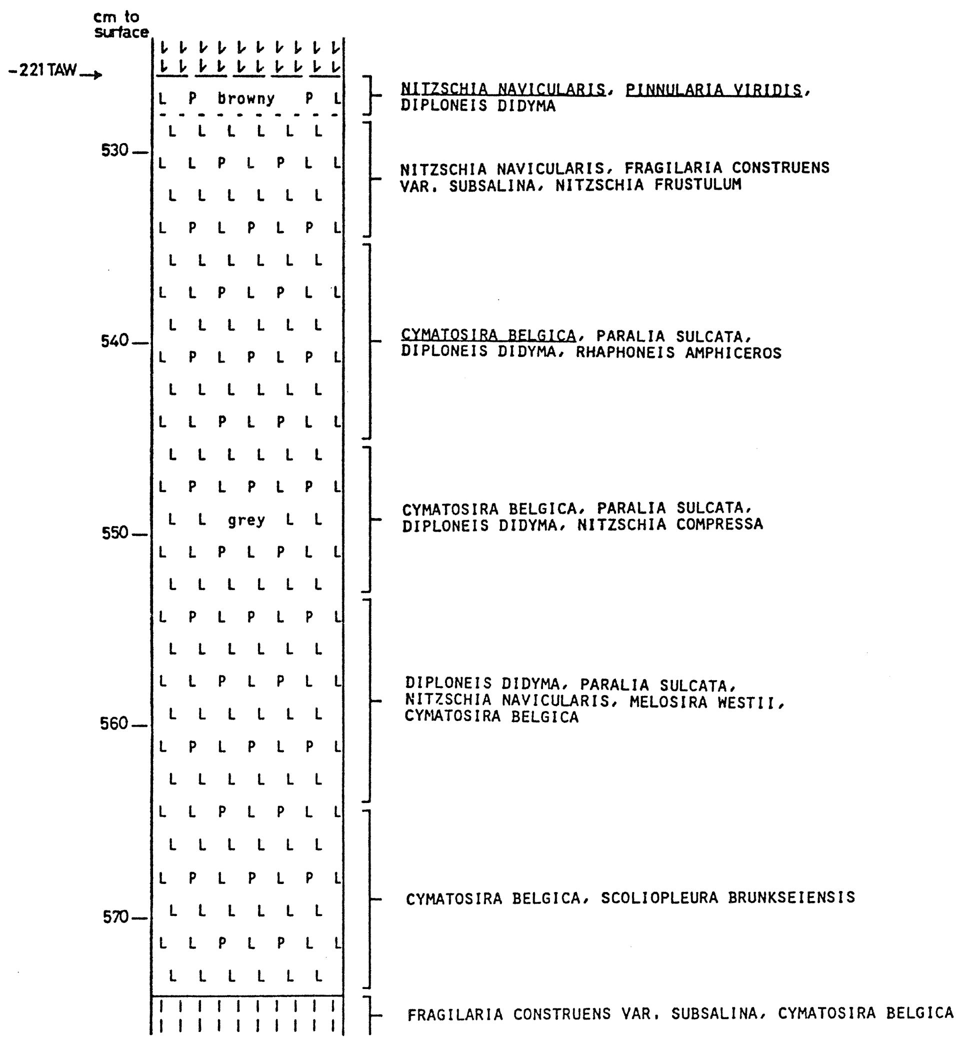

Fig. 8: Lithology and most important diatoms (predominant taxa underlined) of the Calais/peat transition in the core Vliegveld.

requirements, it is improbable that these taxa actually lived together at the same place and time. Rather their valves became mixed when the marsh was covered by the extending mudflat.

Immediately above the peat there is already a codominance of Cymatosira belgica and Scoliopleura brunkseiensis, an epipelic diatom from brackish quiet intertidal habitats (570-565 cm). The rather brackish character of the environment becomes even more explicitly demonstrated by the following assemblage wherein Nitzschia navicularis, Paralia sulcata and Melosira westii are the main species (565-555 cm). This phase of silting-up was only temporary as can be deduced from the gradual return of Cymatosira belgica. Initially the mesohalobous Diploneis didyma and Nitzschia compressa persist, but soon a period of almost marine conditions sets in as reflected by the predominance of Cymatosira belgica (555-535 cm).

Only in the uppermost part of the clay (535-528 cm) a marked salinity decrease takes place, as shown by high numbers of Nitzschia navicularis, Fragilaria construens var. subsalina and Nitzschia frustulum. In the end, at the onset of peat formation (527 cm), the freshwater diatom Pinnularia viridis becomes numerous.

6. DISCUSSION

From the above it becomes clear that quite different sedimentary environments are represented in what at first sight appear to be rather uniform deposits. The sticky clay facies supports diatom associations with salinity requirements ranging from nearly fresh to almost fully marine. As can be expected from marginal environments, diatoms with different ecological requirements are often found intermingled. For example, the polyhalobous Paralia sulcata and Melosira westii which are often surprisingly numerous in assemblages with otherwise mainly mesohalobous diatoms; apparently the chainlike colonies of these tychoplanktonic species tend to accumulate in great numbers near the upper flood limit.

Nevertheless three main environments/assemblages can be recognised:

- the polyhaline mudflat with Cymatosira belgica, Rhaphoneis minutissima, Paralia sulcata and Rhaphoneis amphiceros as characteristic species,
- the mesohaline mudflat and lower saltmarsh typified by Nitzschia navicularis, N. compressa, Diploneis smithii, D. didyma, Caloneis westii and in case of sufficiently low salinity with Diploneis ovalis and Navicula peregrina,
- marshy, brackish to fresh transitional situations initiating or terminating periods of peat formation and characterised by Fragilaria's (especially F. construens var. venter and var. subsalina) and sometimes Pinnularia viridis.

In general the environmental succession below and at the regressive overlaps considered here is one from marine-brackish to brackish mudflat and low marsh to fresh fen. At the transgressive overlaps a reversed sequence is observed. An important role herein was played by an intertidal environment, comparable to the panne-marsh described by Redfield (1972), characterized by a mosaic of bare areas where some shallow water may have stood at low tide and patches with saltmarsh vegetation that fell dry diurnally. The transition from marsh to fen was determined by the supply of large amounts of fresh water, originating mainly from inland drainage. Combined with the upwelling of seepage water of low salinity in certain areas, local precipitation and the limited drainage of the rather flat clay surface this resulted in a rapid waterlogging of the soil once the clastic accretion stopped. This prevented oxidation of the soil and normal marsh maturation but instead induced peat growth at most places. When the water table reached the marsh's surface, a semi-terrestrial peat started to accumulate which increased the storage capacity. Together with the sea-level related general rise of the groundwater table, the growth of various fen vegetations was made possible. Most probably the development of a protective (though not closed) barrier system which reduced the local tidal amplitude and quenched the sea's energy was a major controlling factor, while climatic conditions must have been favourable.

This scenario also explains why the peat has a freshwater character nearly from its very beginning, something which would not be expected from a peat formed in a brackish lagoon. The rather large differences observed from place to place in the time of initiation of peat-growth originate from differences in local sedimentation rates, controlling the flat to marsh transition, perhaps also somewhat from the variable lag period between marsh formation and waterlogging. In Fig. 9 a generalised scheme of the sedimentary situations discussed above is given.

The backswamp succession described above is as plausible as one from from tidal flat to peat bog via truly lagoonal systems. Similar schemes were proposed to explain the formation of Holocene peats in different North Sea countries (for example Great Britain: Chapman 1960; B.R.D.: Menke 1968, Prange 1967; The Netherlands: Bakker & Van Smeerdijk 1982, Bennema 1954). This study suggests that it was the most common mechanism of surface peat formation in the western Belgian coastal plain (or at least in its more eastern part). At some places the peat growth may indeed have started in lagoon-like basins. Apparently these "lagoons" should however be thought of more as limited depressions in the marsh rather than as extensive lakes covering vast parts of the intracoastal area.

Shennan (1986) recently described a "non-lagoon model" for the deposition of the Fen Clay in an open coast situation. Along barrier coasts as well the importance of non-lagoonal environments should be reassessed if precise use of terminology is to be made.

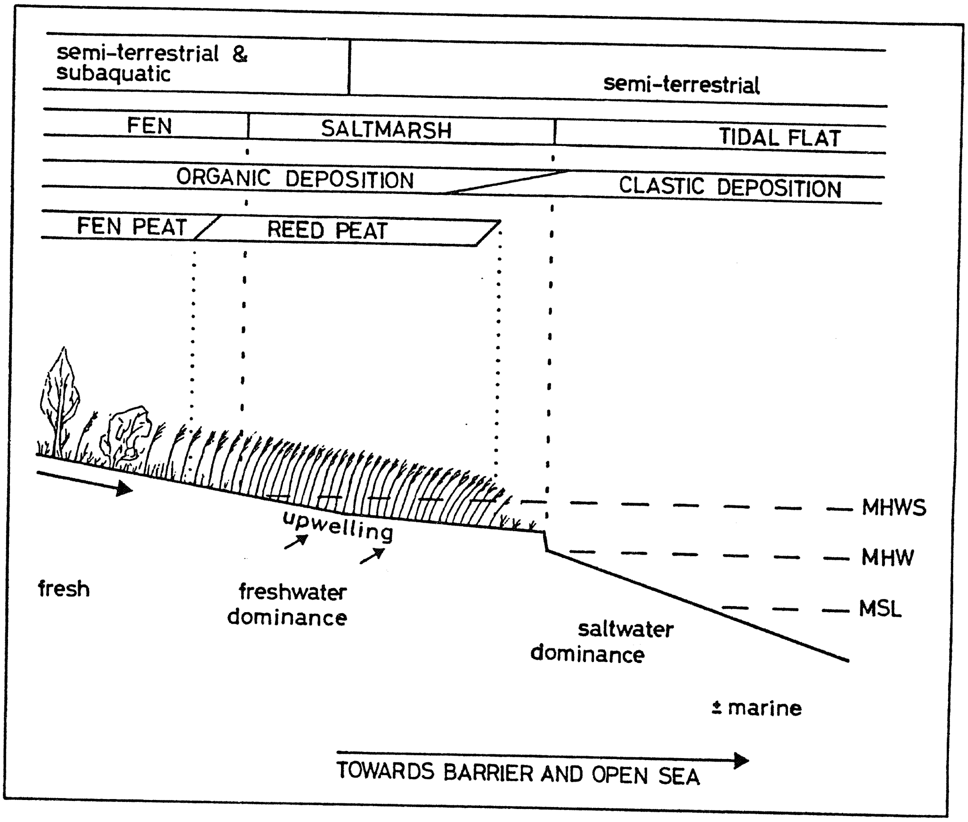

Fig. 9: Sedimentary environments leading to surface peat formation in the western Belgian coastal plain.

Due to severe recent, mainly man-made changes of the coastal areas, sites where the natural intergradation from a marine to a peat-forming or lagoonal environment still occurs are becoming exceedingly rare (cfr. Menke 1968, Ranwell 1974). The resulting lack of recent references and the often ambiguous nature of macroscopic criteria makes the interpretation of their fossil counterparts a difficult task. In such cases a closer analysis of the biotic remains, especially of the diatoms, is highly recommended to provide more detailed information.

7. ACKNOWLEDGEMENTS

Without the cooperation of the Belgian Geological Survey, which commissioned the coring, this research would hardly have been possible. Especially Dr. C. Baeteman of the Survey is acknowledged for her interest and active support. The study was carried out at the Leerstoel Algemene Plantkunde, RUCA, which provided laboratory facilities. Part of it was supported by an IWONL grant to the author. Besides by Dr. Baeteman, the manuscript was read by Dr. C. Verbruggen, Dr. L. Beyens and Prof. Dr. D.K. Ferguson. Their suggestions and corrections are much appreciated. The radiocarbon dates were provided by the IRPA, special thanks are due to Mr. M. Van Strijdonck in this respect.

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HOLOCENE SEA-LEVEL INDICATORS IN THE EASTERN PART OF THE BELGIAN COASTAL PLAIN

MOSTAERT Frank *

1. INTRODUCTION

In order to get a detailed insight of the Holocene palaeogeographical evolution, required for sea-level research, a number of temporary excavations and numerous borings have been interpreted in the eastern part of the Belgian coastal plain (fig.1). These investigations made it possible to evaluate the feasibility of the coastal plain sedimentary sequences for classical sea-level research with radiocarbon-dated basal peat samples.
It is argued that the Holocene geological conditions of the eastern part of the coastal plain were not favorite for the application of this classical method. As a result, alternative sea-level indicators had to be searched for. Attention was focused on the study of the palaeogeographical and potential sea-level changes during the last 2500 years.

2. HOLOCENE SEDIMENTARY SEQUENCES AND PALAEOGEOGRAPHICAL EVOLUTION OF THE EASTERN PART OF THE COASTAL PLAIN

The Holocene sequence corresponds to a transgressive prism. The eastern part of the Belgian coastal plain was caracterized by the existence of a barrier coast with intracoastal tidal sedimentation and peat development under rising sea-level conditions. Holocene tidal deposits or peat layers came into existence on a slightly seaward dipping palaeosurface of Weichselian fluvioperiglacial deposits, covered with a thin coversand sheet and with local east-west orientated coversand ridges. This rather flat and high lying Pleistocene substratum with its specific microrelief was extremely favorite for peat development under rising water table conditions related to sea-level rise.

Before 5600 B.P., tidal impact only reached the extreme western part of the eastern coastal plain (Zandvoorde, Bredene, Wenduine). In that area, tidal flat and lagoonal conditions existed. Extensive peat growth started from 5600 B.P. on, wich is obviously earlier than the western part of

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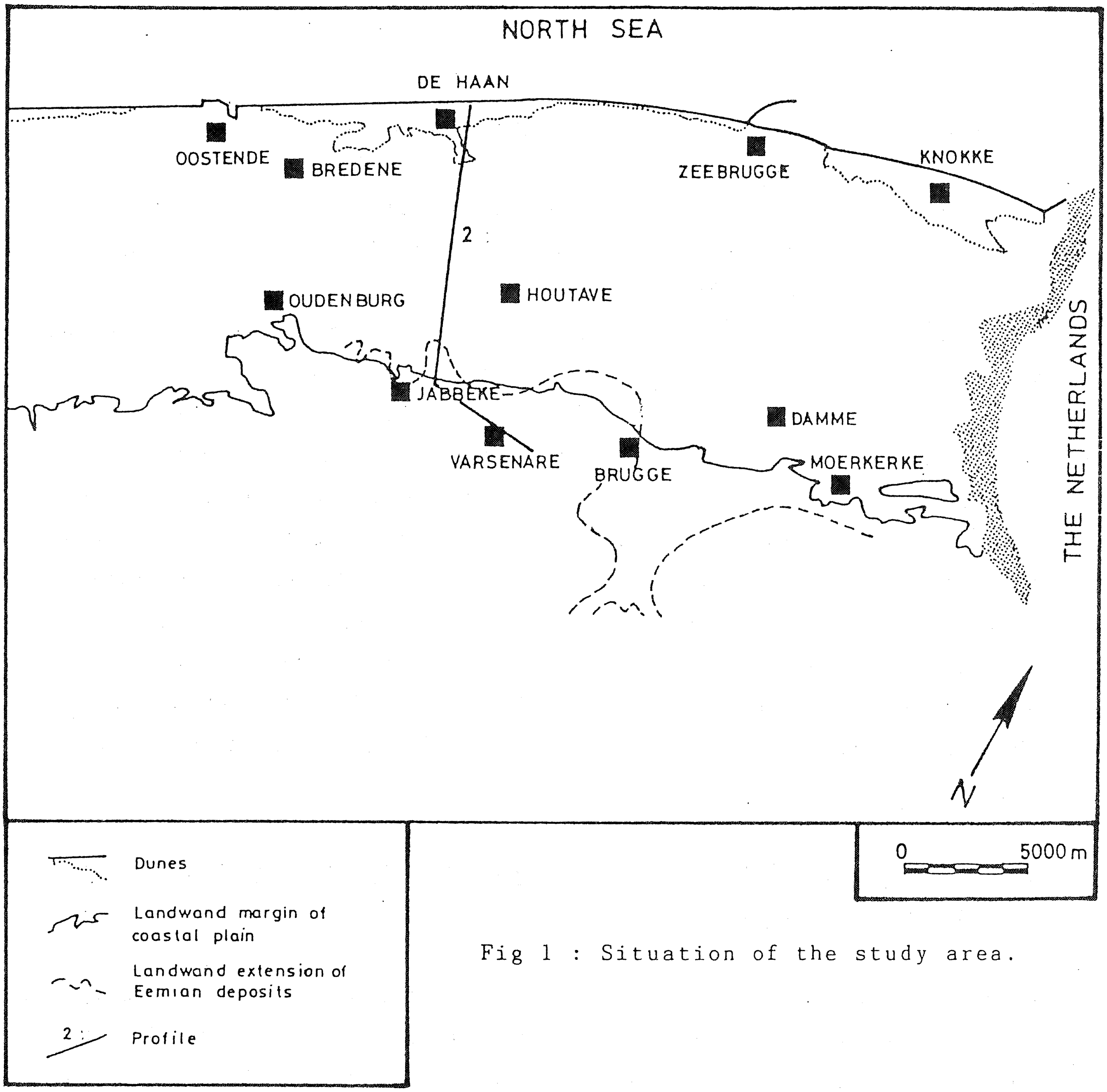

Fig 1 : Situation of the study area.

the coastal plain (BAETEMAN and VERBRUGGEN,1979). While sea-level rose, peat development penetrated further landward, reaching the present-day southern margin of the coastal plain (3000-2500 B.P.). The area was adequately prevented from marine influence by an uninterrupted coastal barrier (5600-2500 B.P.) allowing the development of ombrotrophic sphagnum peat (ALLEMEERSCH, 1986).

The largest part of het eastern coastal plain was prevented from marine influence untill the so-called Dunkerque I transgressive phase (from 2500 B.P.) when the coastal barrier was eroded and tidal gullies and tidal flats developed. Meanwhile, mean sea-level reached at least +0.5 to 1m T.A.W. (*). This Subatlantic inundation occurred on a very flat swamp landscape reaching levels between +2 and +3m T.A.W.

Tidal impact did not disappear from the coastal plain since 2500 B.P. untill medieval reclamation (1000-1700 A.D.). The so-called Roman regression (-100 to +300 A.D.) corresponds with local expansion of the salt marsh area and the reduction of the width and the activity of the tidal gullies. However, salt-water influence still reached the southern margin of the coastal plain (MOSTAERT, 1988). A post-roman enlargement of some tidal gullies could be proved from the sedimentary record, while other tidal gullies simultaneously reduced in space. The so-called Dunkerque II transgression did not correspond with a spectacular enlargement of the area under tidal influence. Historical and archaeological arguments indicate that storm surges often implied important palaeogeographical changes in the tidal flat area.

3. FEASIBILITY STUDY OF THE USE OF RADIOCARBON-DATED PEAT LAYERS AS SEA-LEVEL INDICATORS

At first sight, the alternation of clastic and organic sediments (figure 2) seems to be ideal for sea-level reconstruction based on radiocarbon dates. To evaluate the possibility of the use of 14-C dates of the base of peat layers covering the Pleistocene substratum as sea-level indicators, these peat layers had to be mapped and the base topography of the peat had to be measured. The lithological and hydrogeological characteristics of the substratum have to be taken into account. The area of Houtave and Brugge has been mapped in this way (figure 3).

Apart from very local peat layers, found near Zandvoorde and Bredene, major peat developement was restricted to the period 5600 -2500 B.P., reducing

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(*) T.A.W. : second general levelling, lowest low water level in Ostend = -2.31 m N.A.P.

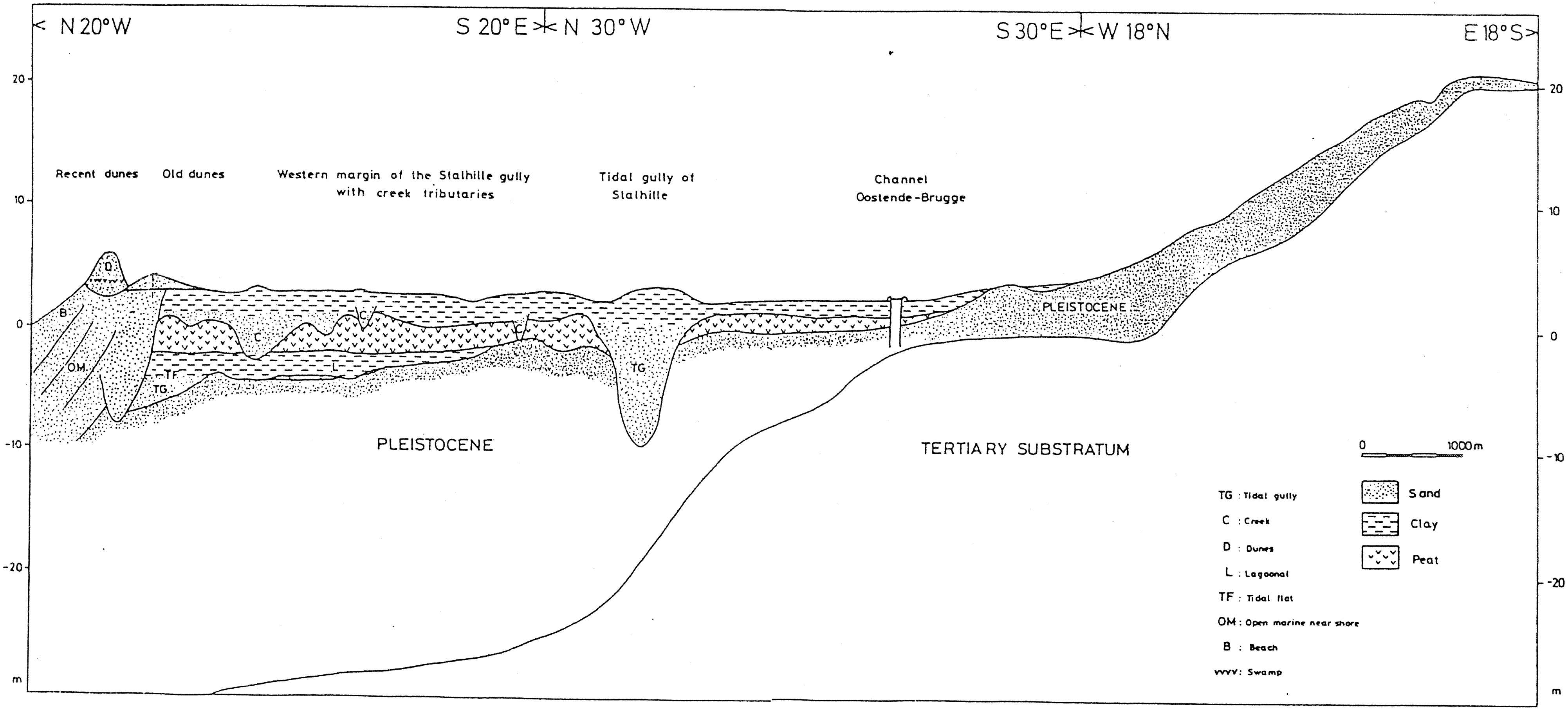

Fig 2 : Holocene sequences.

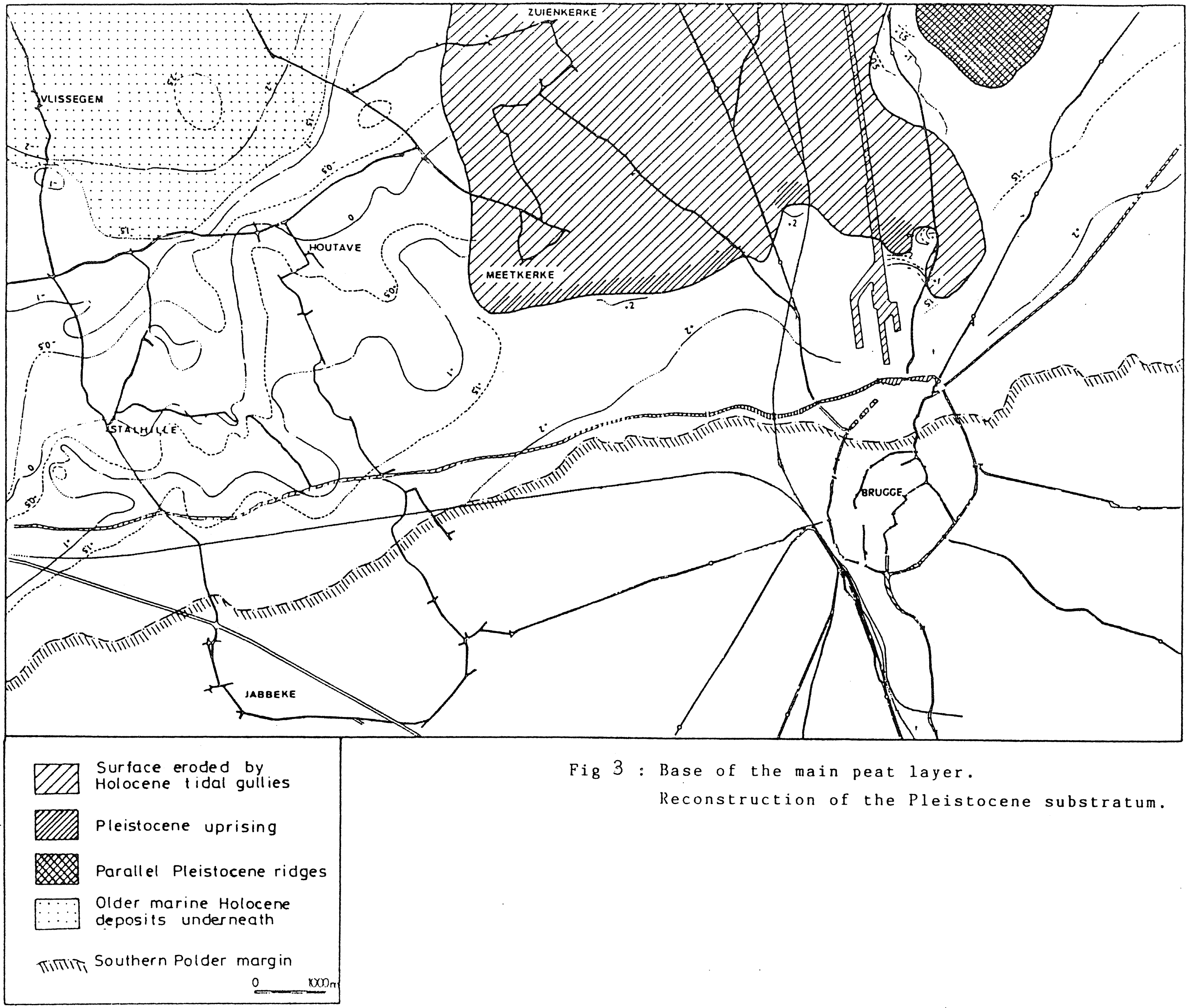

Fig 3 : Base of the main peat layer.
Reconstruction of the Pleistocene substratum.

the sea-level interpretation to this time interval. This restriction implies a reconstructable mean high water level (MHWL) interval between -2 m and +1.5 m T.A.W. Two fairly local marine ingressions with lagoonal sedimentation were detected. These ingressions took place during the main peat development phase and finished at about 5000 B.P. (DAUCHOT-DEHON and HEYLEN, 1969) and at 4380 B.P. (MOSTAERT, 1985), respectively near Uitkerke and Bredene.

An important area to the west of the line Oudenburg-Uitkerke, where peat developed on older clayey lagoonal and salt marsh deposits,can be eliminated for sea-level research due to potential compaction impact on the sea-level indicators (the dotted area on figure 3). The rather large area where peat immediately covered the Pleistocene substratum does not guarantee reliable sea-level indicators. The relationship between the mean high water level and the base of the peat layer is not always clear in the eastern part of the coastal plain due to local palaeohydrogeological factors.
The appearance of Tertiary and Eemian clay layers a few meters below the peat in the landward part of the coastal plain (MOSTAERT and LIBEER, 1988), together with the heightening effect on the ground water table of the cover-sand ridge and the uprising Tertiary substratum of the hinterland (figure 2) mask the sea-level influence on the ground water level.

Interesting sites for sea-level reconstruction, the local Pleistocene uprisings, only exist too far landward where local water table is too much influenced by the hinterland. It must be kept in mind that large parts of the Pleistocene uprisings indicated on the map (figure 3), together with an extensive peat area have been subjected to tidal gully erosion during the Subatlantic inundation, reducing once again the potential sampling area. The seaward trending slope of the Pleistocene substratum is not steep enough and too irregular to exclude the development of eutrophic peat in local depressions, conditioned by local hydrological factors. However, the map (figure 3) indicating the base of the peat allows to select interesting sample sites in the vicinity of Houtave.

As a result of the protecting effect of the coastal barrier since 5600 B.P., the direct sea-level influence on the ground water level might be reduced considerably. The ombrotrophic peat of the eastern coastal plain indicates the absolute absence of salt-water influences. The rather large tidal range along the Belgian coast (4 m) excludes as a matter of fact the co-existence of peat development and tidal influences. Where a small river system, the Reie-Waardamme, entered the coastal swamp near Brugge, eutrophic conditions persisted, but there did not exist any outlet to the open sea during the peat extension (MOSTAERT, 1987). The mere indications for contemporaneous peat development and marine influence was noticed near Bredene and Uitkerke where lagoonal clastic deposits came into existence under strongly tempered tidal influences.

These reflections lead to the conclusion that the eastern part of the Belgian coastal plain is not a favorite site for extensive sea-level research based on peat dates. The few reliable time/depth data available in the eastern coastal plain give following provisional indications. Compared with the mean sea-level curve presented by VAN DE PLASSCHE (1982) for the Netherlands and based on JELGERSMA (1979), the most compaction-free data lay consequently more than 2 m above this curve (MOSTAERT, 1985). The Belgian sea-level indicators reflect the mean high water levels with a tidal range of ± 4 m. There seems to be a decreasing discrepancy in time between the projected mean high water level curve based on a tidal range of 4 m, and the measured depth levels. This might be a result of a decreasing tidal range till 2000 BP or, which is more probable, of a differential megatectonic and isostatic response of the Belgian area in comparison with the Netherlands (a relative subsidence of the Netherlands by 0.15 mm up to 0.35 mm:year can be derived).

4. ALTERNATIVE SEA-LEVEL INDICATORS

The Subatlantic tidal flat sequences allow to collect additional sea-level indicators and information about the tidal range in the tidal flats. A number of sedimentological sea-level indicators will be evaluated. Sedimentological and stratigraphical arguments cannot prove the existence of general sea-level lowering phases during the Holocene. Important palaeogeographical changes, as the appearance or disappearance of swamp conditions, so-called regression or transgression phases can all be explained without the necessity of accepting eustatic sea-level changes. However, local fluctuations of sea-level stands within the tidal flat area might have occurred due to a combination and interaction of factors which can be related with the storage-basin effect and its variability in time. The storage-basin effect is a lowering of the mean high water level within the intracoastal area due to frictional dissipation of energy as the tidal wave moves into the tidal area and by the fact that the intra-coastal area acts as a storage basin (VAN DE PLASSCHE, 1980). Morphological variations of the outlets and changes of the storage basin by different

processes (sedimentation, erosion, open marine impact, eustatic sea-level changes,...) have a net effect on the local high water level and indirectly on the potential sedimentation levels (figure 4).
As the eastern part of the coastal plain is not drained by an important river system, the river gradient effect on the high water level has not to be taken into consideration.

Specific sedimentation levels might be interesting sea-level indicators. The major problem with these sedimentary indicators is their difficulty to be dated. Tidal gully sequences for instance contain datable elements (shells and peat debris, archaeological fragments) which have been reworked making them useless as sea-level datings. The most reliable time indicators for the last 2000 years are archaeological indications and historical arguments such as the well-known reclamation history.
A number of high water level indicators have been measured for the Subatlantic series, such as the height of the landward coastal plain margin, the sedimentation levels of the salt marsh deposits in positions with minimum compaction, the height of fossil tidal levees, the sedimentary transition from beach to aeolian sedimentation.
This regional study, taking into account the embankment history leads to following conclusions. In the eastern part of the Belgian coastal plain high water level , comparable with the present-day one already existed before the first reclamation of the tenth century A.D.
An eastward decrease of the highest tidal sedimentation level along the landward coastal plain margin has been noticed: from 4.3 m T.A.W. in Jabbeke-Houtave to 3.5 m T.A.W. in Moerkerke, which is consequent with an eastward decreasing tidal range comparable with the present-day decrease. Furthermore, the storage-basin effect was most effective before the first embankment in the Zwin gully area, in the extreme eastern part of the coastal plain. This effect seems to be reduced when further reclamation took place.
The comparison of compaction-free sedimention levels with salt marsh sedimentation levels in compaction-sensible area leads to an evaluation of the importance of compaction mounting up to 2 m. Post-embankment compaction sorted most spectacular effects in the landward part of the coastal plain where peat layers occur relatively high,having a thin cover of salt marsh deposits (MOSTAERT, 1986).

A number of temporary excavations at Bredene and to the north of Brugge made it possible to study Subatlantic gully infilling sequences which

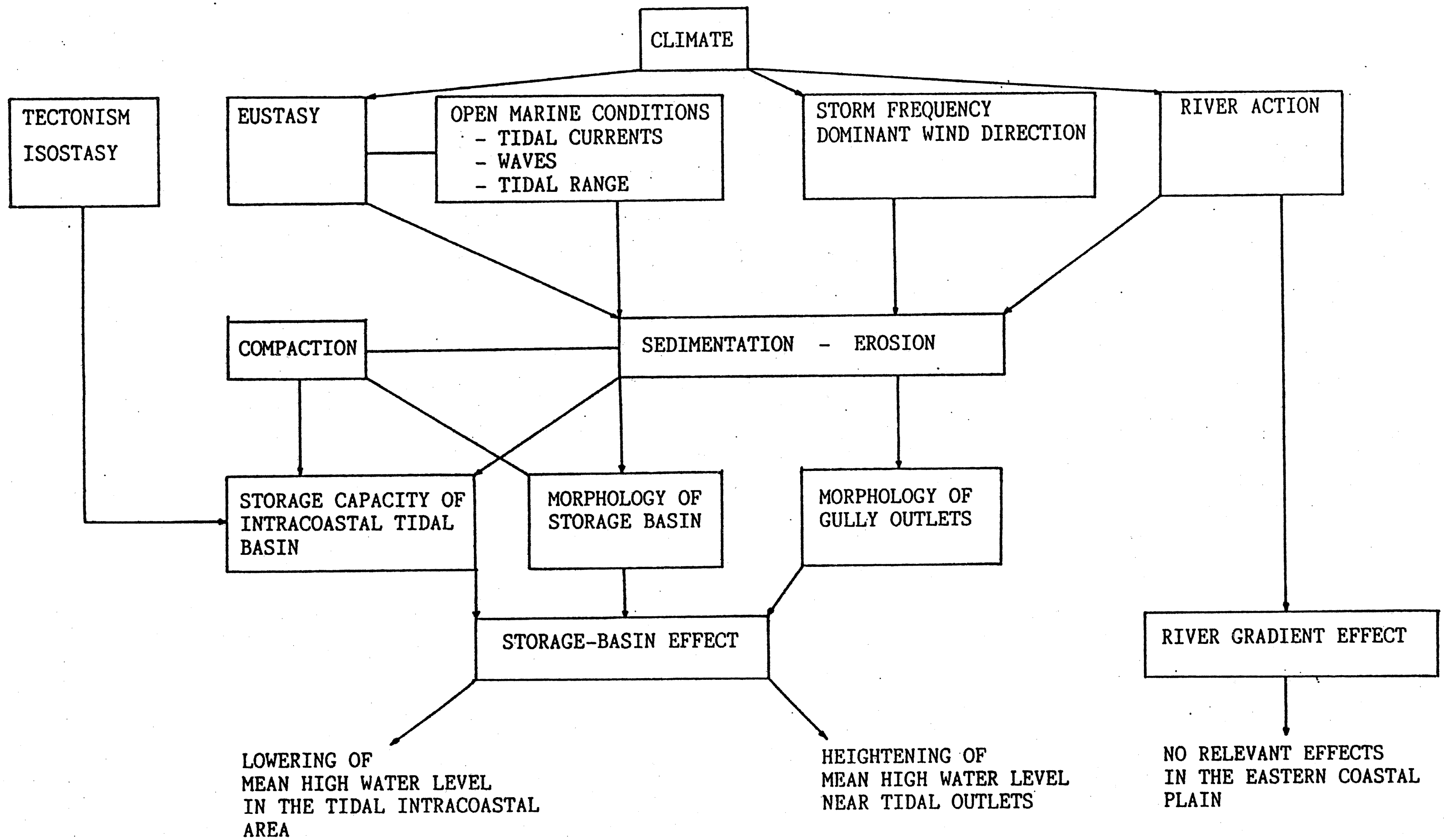

Figure 4: A selection of factors and processes dominating the local Mean High Water Level in intracoastal tidal areas.

came into existence under variable palaeogeographical conditions. Criteria have been searched for to distinguish subtidal sequences from intertidal sedimentary successions. The most important sedimentological indication for subtidal conditions is the occurrence of specific tidal bundles illustrating the neap-spring cycles. These bundles are characterized by eb and flood slack water mud drapes, or peat debris drapes where clay was not available (MOSTAERT, 1985). The highest appearance of these subtidal phenomena is an indication for the lowest low water level. At Bredene (figure 5), subtidal megaripples appear below + 0.75 m T.A.W. and in Brugge subtidal structures have been noticed at + 1.10 m T.A.W.(figure 6). Both the Bredene and Brugge gully infillings came into existence from 400 to 800 A.D. In that period salt marsh deposition took place at levels above + 4 m T.A.W. These observations imply that the local tidal range must have been considerably smaller than the present-day one along the coast. In the intracoastal area a tidal range of $\pm$ 3 m existed with even higher mean sea-level than nowadays (+ 2.7 m instead of + 2.3 m T.A.W.). Now, a mean tidal range of $\pm$ 4 m is measured. The differences between the Brugge and Bredene sequences can be explained by their different palaeogeographical localisation respectively 7 km landward from the coastal barrier islands and near the outlet of the gully.

Apart from the static inventory of sedimentation levels as sea-level indicators attention was focused on a detailed palaeogeographical and stratigraphical reconstruction fitted in the existing chronostratigraphical framework. A number of sedimentation phases have been differenciated in the tidal sequences of the last 2500 years. These phases have no direct relation with the three so-called Dunkerque transgressions. Although similar sedimentary sequences occur along the major tidal gullies, no relevant evidence for their simultaneous development could be gathered. The original swamp gradually changed into a lagoonal environment often related to a long period of non-deposition (BAETEMAN, CLEVERINGA and VERBRUGGEN,1981). Gradually tidal gullies enlarged and a tidal flat came into existence characterized by an extensive salt marsh area. This gully enlargement phase was followed by a decrease of the tidal impact implying tidal gully reduction and salt marsh extension. Most sequences indicate a tidal gully reactivation phase followed by a renewed salt marsh expansion. The final tidal gully infilling phase can be accelerated by human impact: the reclamation induced an artificial reduction of the width of the storage-basin reducing the erosive potential of the gullies.

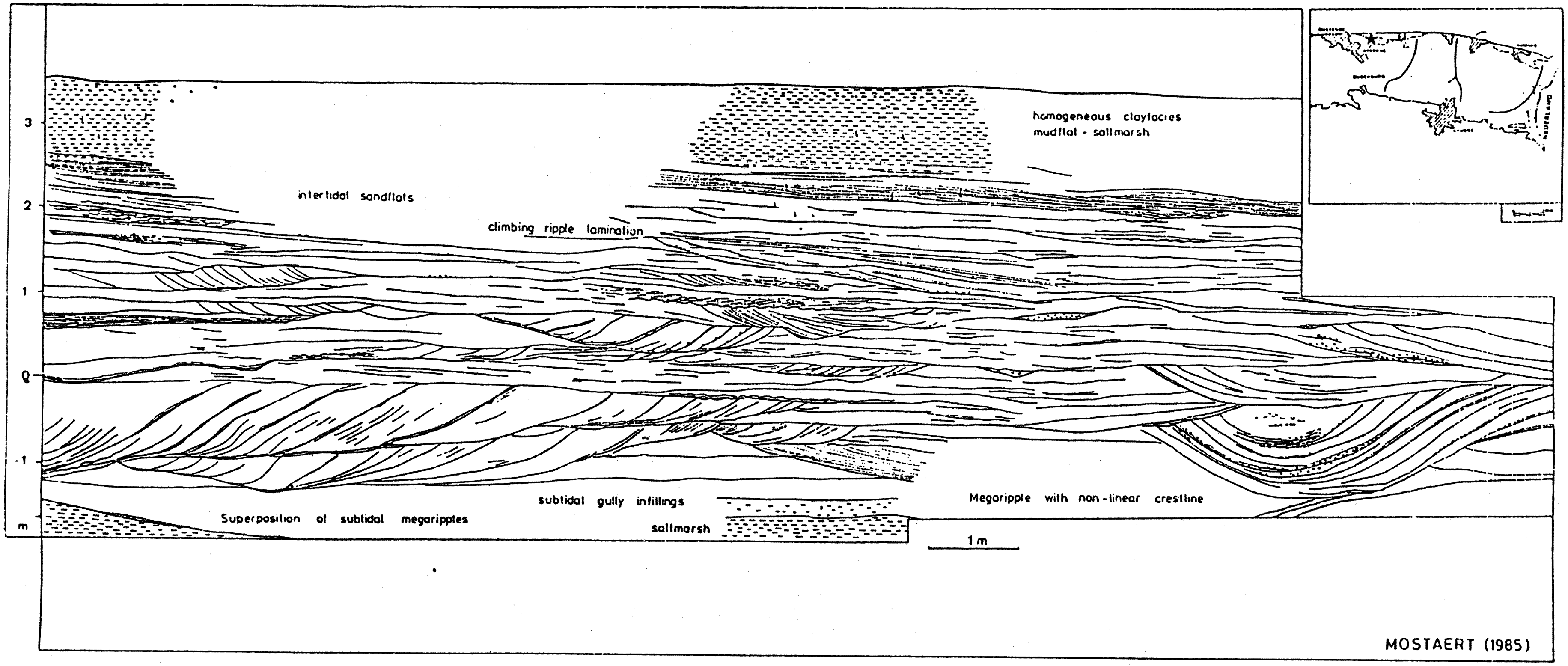

Fig. 5: Holocene Dunkerquian tidal gully sequence: Bredene

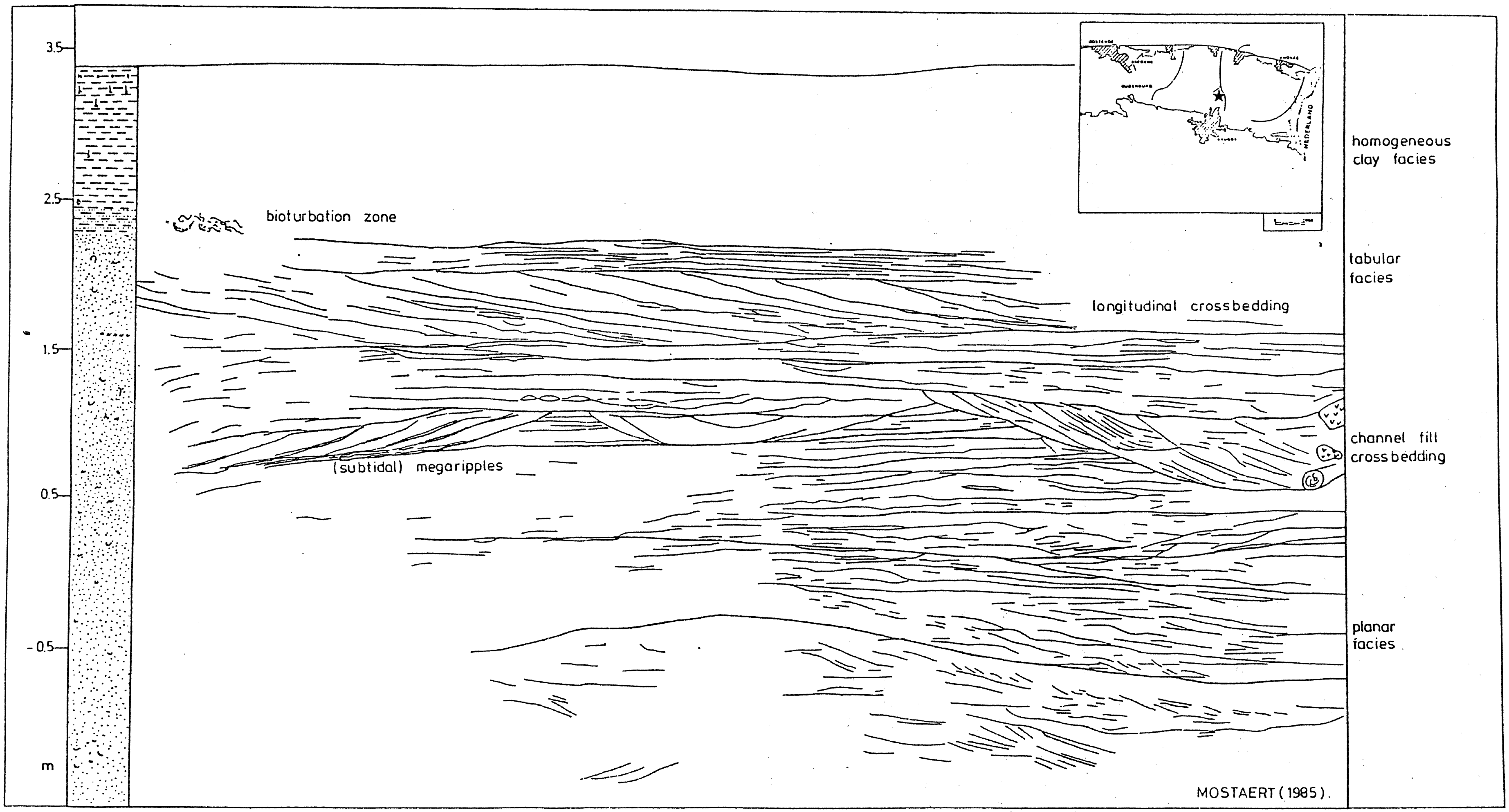

Fig. 6 : Holocene Dunkerquian tidal gully sequence : Brugge

5. CONCLUSIONS

Holocene sea-level research in the eastern Belgian coastal plain is complicated by local hydrogeological factors and by special palaeogeographical conditions, with an apparently well-developed coastal barrier system during peat growth and the specific palaeomorphology of the peat substratum. Sedimentary sequences include high water and low water level indicators and give an indication of important tidal range reduction in the intracoastal tidal area of the eastern coastal plain.

A multidisciplinary approach to the sea-level problem, concentrated on a selection of specific sites with the integration of archaeological research, may elucidate the detailed palaeogeographical evolution of the coastal plain. Some important questions wait for an answer such as the problem if the eastward shifting increase of areal importance of tidal gully infillings in the area considered is an indication for the longer duration of tidal activity in the gullies in eastward direction. Are there other factors which control this phenomenon?
The knowledge of the palaeogeographical evolution is a primary condition for an objective insight of the sea-level indicators and the factors influencing them.

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RADIOCARBON DATES ON PEAT FROM THE HOLOCENE COASTAL DEPOSITS IN WEST BELGIUM.

Cecile BAETEMAN* & Mark VAN STRIJDONCK**

INTRODUCTION

The objective of this paper is to present the radiocarbon dates that came available at the occasion of the systematic mapping of the Holocene deposits in the western coastal plain of Belgium (fig.1). The mapping was undertaken by means of hand- and power-driven boreholes, the latter yielding suitable and sufficient material for age determinations on peatlayers.

The power-driven boreholes are not regularly spread as they only form a complement to the hand-augerholes. Hence the radiocarbon dates do not cover the entire plain, neighter all the significant changes in processes and sea-level tendencies. Therefore the radiocarbon dates will not be treated statistically. They serve a basis for reconstruction of coastal environments in space and time, which is a first essential step to be carried out before the evaluation of sea-level index points and the construction of a time-altitude graph.

The radiocarbon dates discussed in this paper are not calibrated. It never has been a common practise to calibrate radiocarbon dates in geological studies for two reasons. Geological phenomena are not always datable in the rigorous meaning of the word, because often the investigated material obtains its carbon from different reservoirs. In many cases

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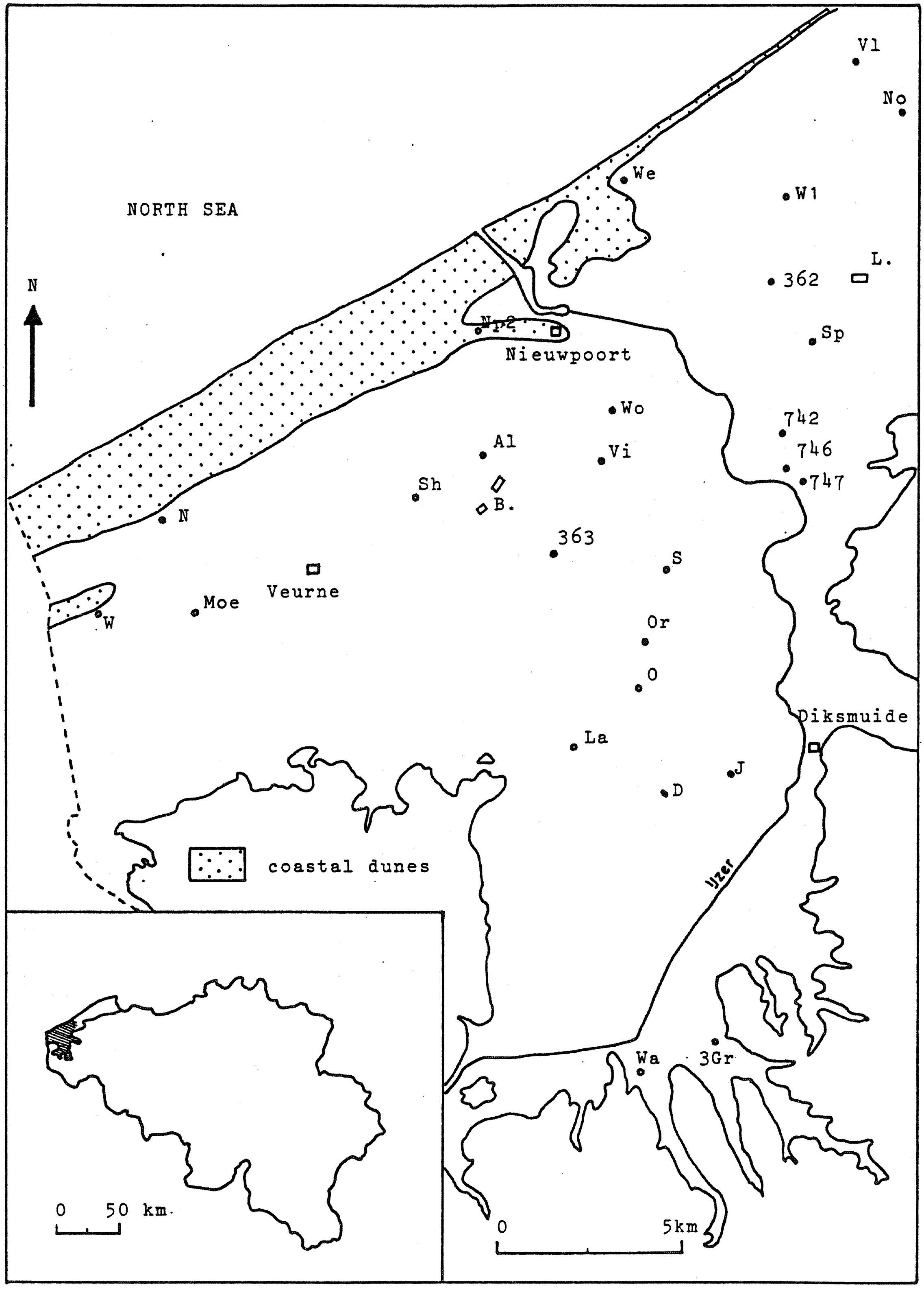


Fig. 1. Western coastal plain of Belgium with location of the 14-C datings

L. : archaeological excavation Leffinge

B. : outcrop Booitshoeke

the relation between the radiocarbon age and the real time is not so important, as long as the relative age is known. The second reason is a pure practical one. The calibration curve goes only back to about 7200 y BC. However not calibrating the radiocarbon dates includes an error if conclusions are to be drawn in terms of absolute time and in terms of duration. The possibility of clustering radiocarbon dates on random peat growth, due to the wiggles in the calibration curve, is to be considered (Geyh, 1980, De Jong, 1981) (fig.2).

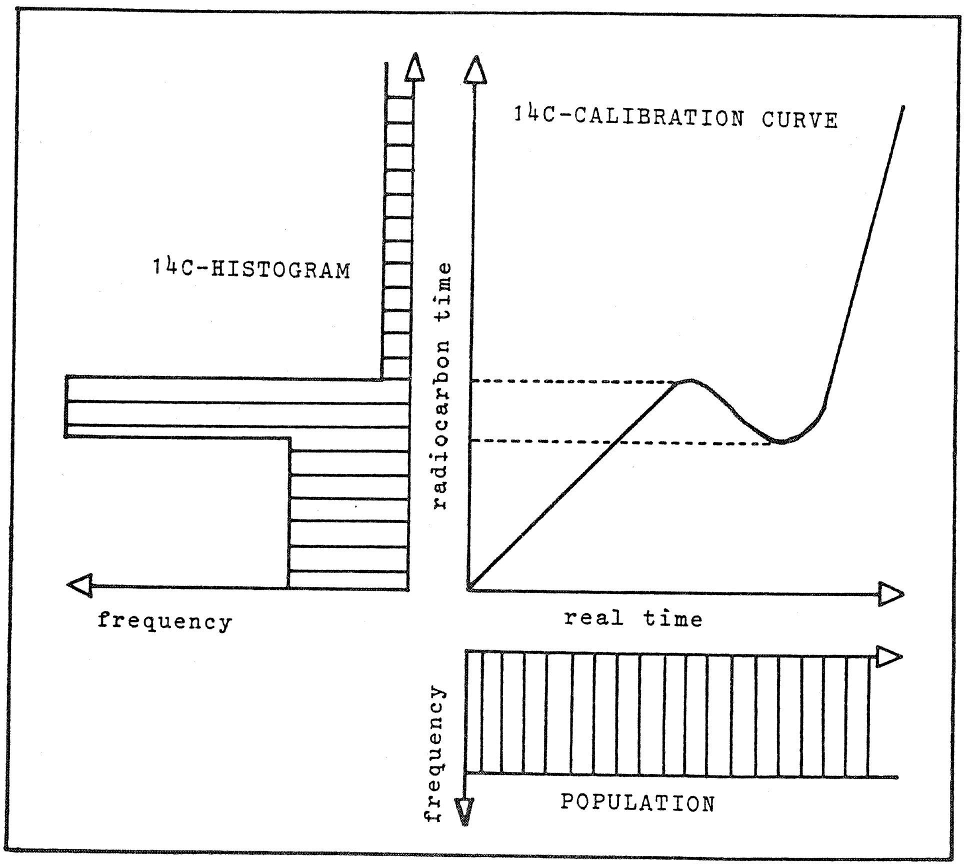


fig. 2. Apparant clustering of samples in radiocarbon time scale due to a wiggle in the calibration curve

Stratigraphy of the Holocene coastal deposits

The coastal deposits represent the major infilling of the area under marine, freshwater and terrestrial conditions during the Holocene. The deposits reach their greatest thickness of about 30 m in the seaward region and wedge out toward the Pleistocene hinterland.

These unconsolidated coastal deposits are characterised by lateral zonation. In the seaward region, only marine and brackish clastic sediments are present overlying a basal peatlayer in some places. In the central part of the plain, the deposits consist in general of an alternation of brackish-marine sediments and peatlayers. Toward the Pleistocene hinterland, the deposits are formed by only a basal peatlayer overlain by a cover of clastic brackish-marine sediments, while at the border of the outcropping Pleistocene area, the cover of brackish-marine sediments form the entire Holocene sequence.

Such lateral zonation, which is typical for the coastal plains of the southern North Sea, led to the development of a lithological classification of coastal deposits based on the vertical succession and lateral interfingering of clastic sediments and peat. The classification consists of complexes and sequences (Barckausen *et al.*,1977 & Streif, 1978). This lithological classification has been applied for the Belgian coastal plain deposits since 1981 (Baeteman,1981a, 1981b, 1987 and Mostaert,1985).

In the seaward region the deposits belong to the **clastic complex** bearing one sequence, *viz.* the clastic sequence, possibly underlain by the basal peatlayer, represented as organic basal sequence. In the central part of the plain, labelled as transition zone, the deposits, characterised by clastic sediments and intercalated peatlayers, are grouped into the **interfingering complex** with the following sequences as further subdivision: lower clastic, splitting up, upper clastic and possibly organic basal sequence (fig.3).

The organic basal sequence on the one hand and the splitting up sequence, bearing one or more peatlayers, on the other hand offer the possibility for the development of a geochronology, linking the lithological classification.

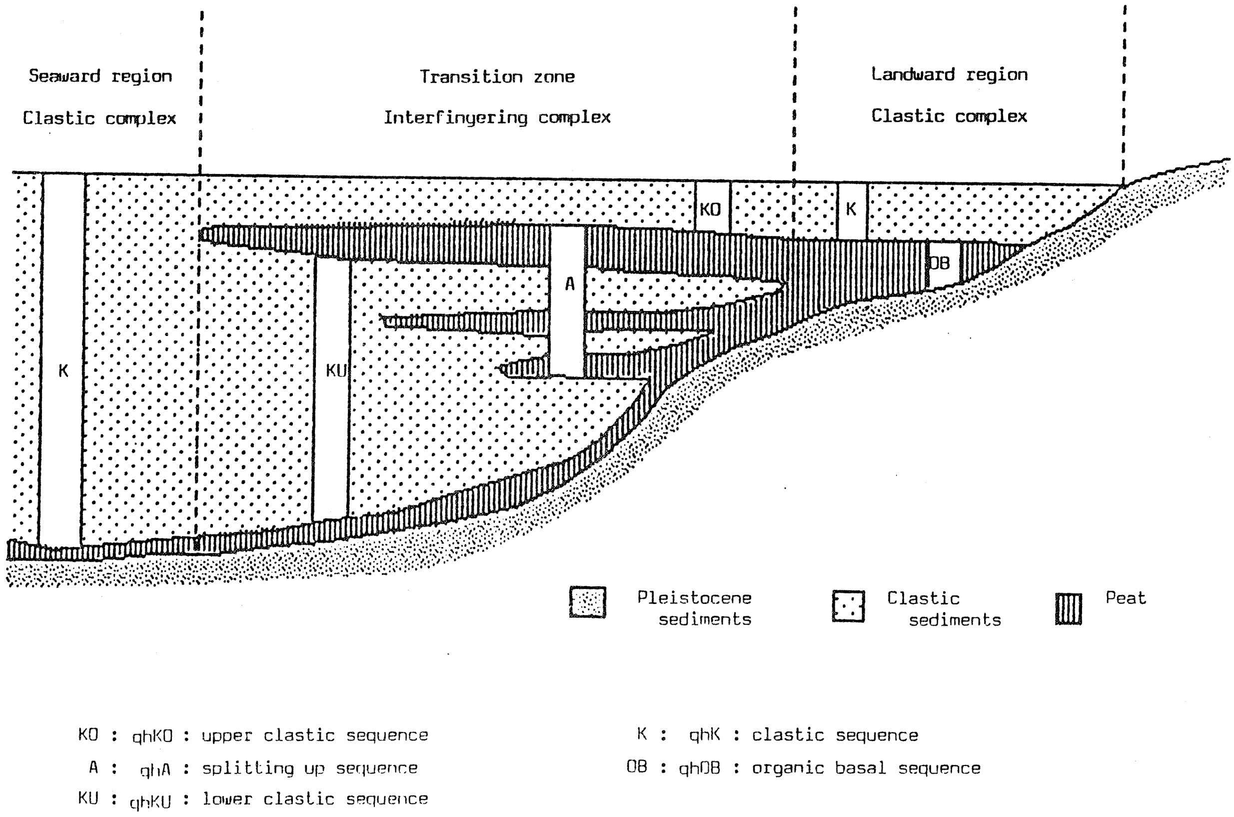


Fig. 3. Schematic cross-section of the Holocene deposits with indication of the complexes and sequences

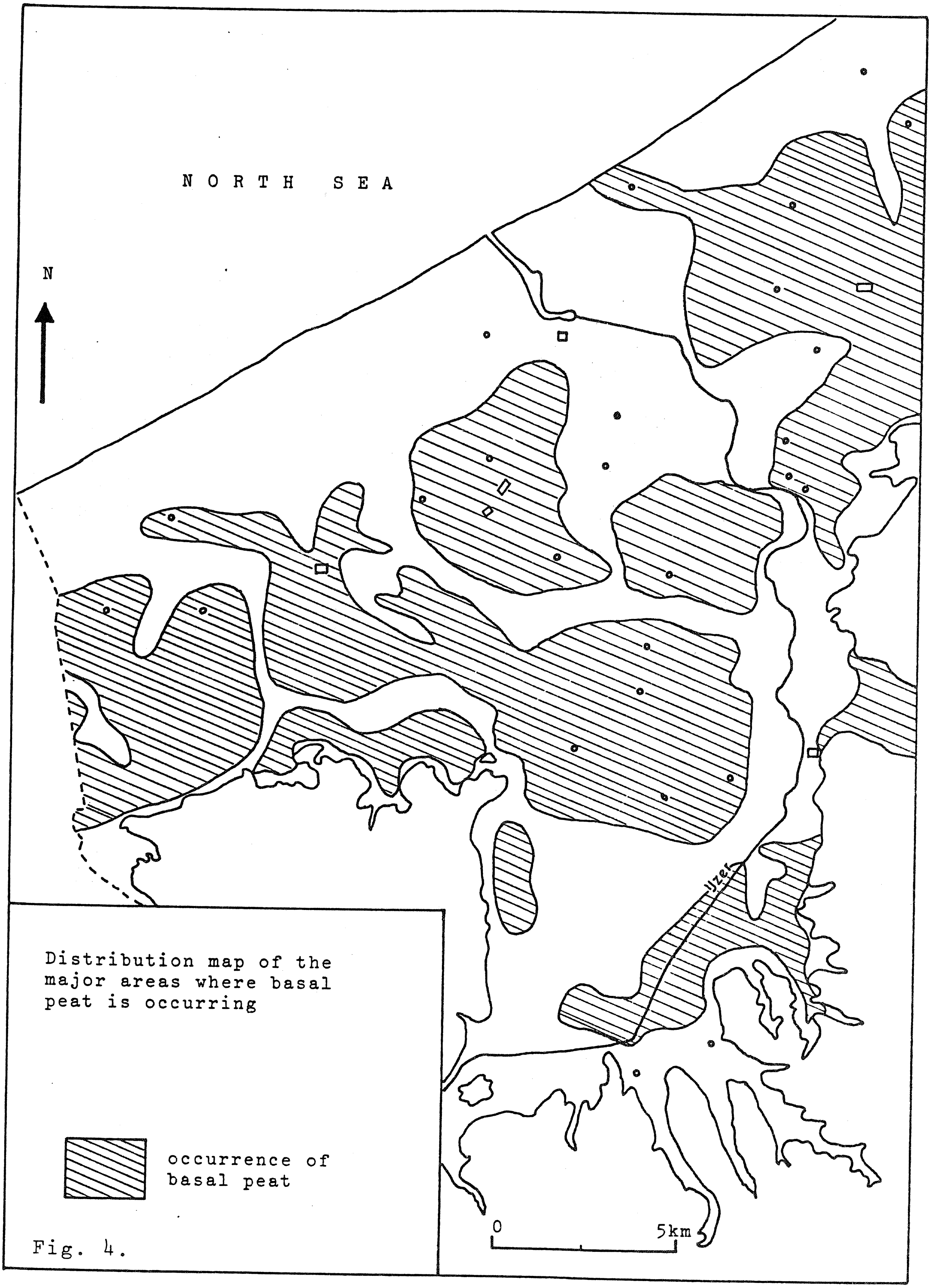
PRESENTATION OF DATA

Organic basal sequence

The organic basal sequence is formed by the basal peat. The term **basal peat** is used to indicate the peatlayer occurring at the base of the marine Holocene sequence. The question wether that peat is attributed to a rise of the groundwater table resulting either directly or indirectly from the rise of the sea level, will not be discussed here as such a relationship can only be concluded on base of palynological investigations, combined with a very detailed survey of the subsurface topography. These investigations have not been carried out in the framework of the systematic mapping.

Basal peat development may have resulted under various conditions, such as regular inundations by run-off water in valleys, or in isolated pre-existing depressions and in poorly drained areas. On the other hand, basal peat certainly can originate as a result of the rising groundwater and coastal freshwater table, caused by the rising sea level. The peat formation is then related to the occurrence of diffuse seepage of groundwater in a belt parallel to the coast as well as the level of the local freshwater table. As a result of the rising of the sea level, this belt shifts progressively inland and upward in the course of time (Roeleveld, 1974). In order to point out the origin of the peat formation, Lange & Menke (1967) proposed restricting the term basis peat only where a causal relationship can be proved between peat formation and sea-level rise (*fide* Streif, 1982).

In the western coastal plain the basal peat is present (fig.4), although in the Belgian literature this always was very much open to doubt (Baeteman,1983). The problem why the



Distribution map of the major areas where basal peat is occurring

Fig. 4.

basal peat never was considered as such is due to the fact that its stratigraphical position and chronology was not well understood. Basal peat was assumed to be of only Preboreal or Boreal age and it ought to occur at greater depths. It should be mentioned that only very few and moreover very surficial data were available at the time these conclusions were put forward. Since 1981, the presence of a basal peat in the Belgian coastal plain and its age have been discussed (Baeteman,1981, Baeteman,1985, Mostaert,1985).

In the western part of the coastal plain the basal peat generally occurs as a well developed peatlayer, although strongly compressed, consisting of mainly *Phragmites* and *Carex*. In most instances the thickness ranges between 15 and 30 cm, but at some locations it is restricted to a podzol in the top of the Pleistocene sands. Toward the border of the outcropping Pleistocene deposits, the basal peat can reach a thickness of more than 1 m and is mainly composed of wood. In these locations, along the outcropping Pleistocene deposits, the overlying Holocene sequence, reaching a maximum thickness of about 2 m, consists of only a basal peat covered by clastic sediments without intercalated peatlayers (fig.3). Hence the basal peat has always been confused with the uppermost intercalated peatlayer (see below) which indeed is occurring at about the same altitude. However from its stratigraphical position, *viz.* at the base of the Holocene sequence, the peatlayer must be regarded as basal peat.

In some areas of the plain, the basal peat is lacking due to erosional incision. These incisions are observed even far inland in the coastal plain, and are the result of tidal channels which eroded several metres into the Pleistocene deposits. In general the basal peat is also lacking in the very seaward area. In this zone, even the Pleistocene deposits are almost completely eroded and in some locations incisions into the Eocene deposits are observed.

Basal peat datings

In general the top of the basal peat was sampled only when an uninterrupted gradual transition from non-marine to brackish and marine environment was observed. Hence the radiocarbon dates reflect the age of the initial transgressive overlap at each locality. When erosion at the top was suspected, only the base was sampled. On the other hand, when the basal peat was not well developed, the entire peatlayer had to be sampled, yielding a mean radiocarbon age.

The radiocarbon dates of the basal peat are represented on profiles delineating the topography of the Pleistocene subsurface to give a better presentation of the relationship between the dates and the topography of the Pleistocene subsurface. (fig.5 & fig.6). Two profiles (1 & 2) across the coastal plain from the mainland towards the present day shoreline in a NNW-SSE and a N-S direction, respectively, are shown. Profiles 3 & 4 are located in the seaward area, following an almost W-E and WSW- ENE direction, respectively. The complete data of the radiocarbon dates are gathered in Table 1.

The deepest sediments recorded, *viz.* -16.97 m to -16.64 m (borehole Sh, profile 3), were observed in the seaward area in a pre-existing depression of the Pleistocene subsurface and dated at 9940 ± 110 BP (base) and 8440 ± 130 BP (top). Still in the same depression in the seaward area, but on a higher level (-11.34 m, borehole Al), a date of 8250 ± 95 BP (mean) was obtained. In the landward extension of the same depression, the base of the basal peat (-15.60 m, borehole O, profile 1 and 2) is dated at 8170 ± 90 BP. At about the same altitude the top of a basal peat (-15.17 m, borehole W, profile 4), although located in the seaward

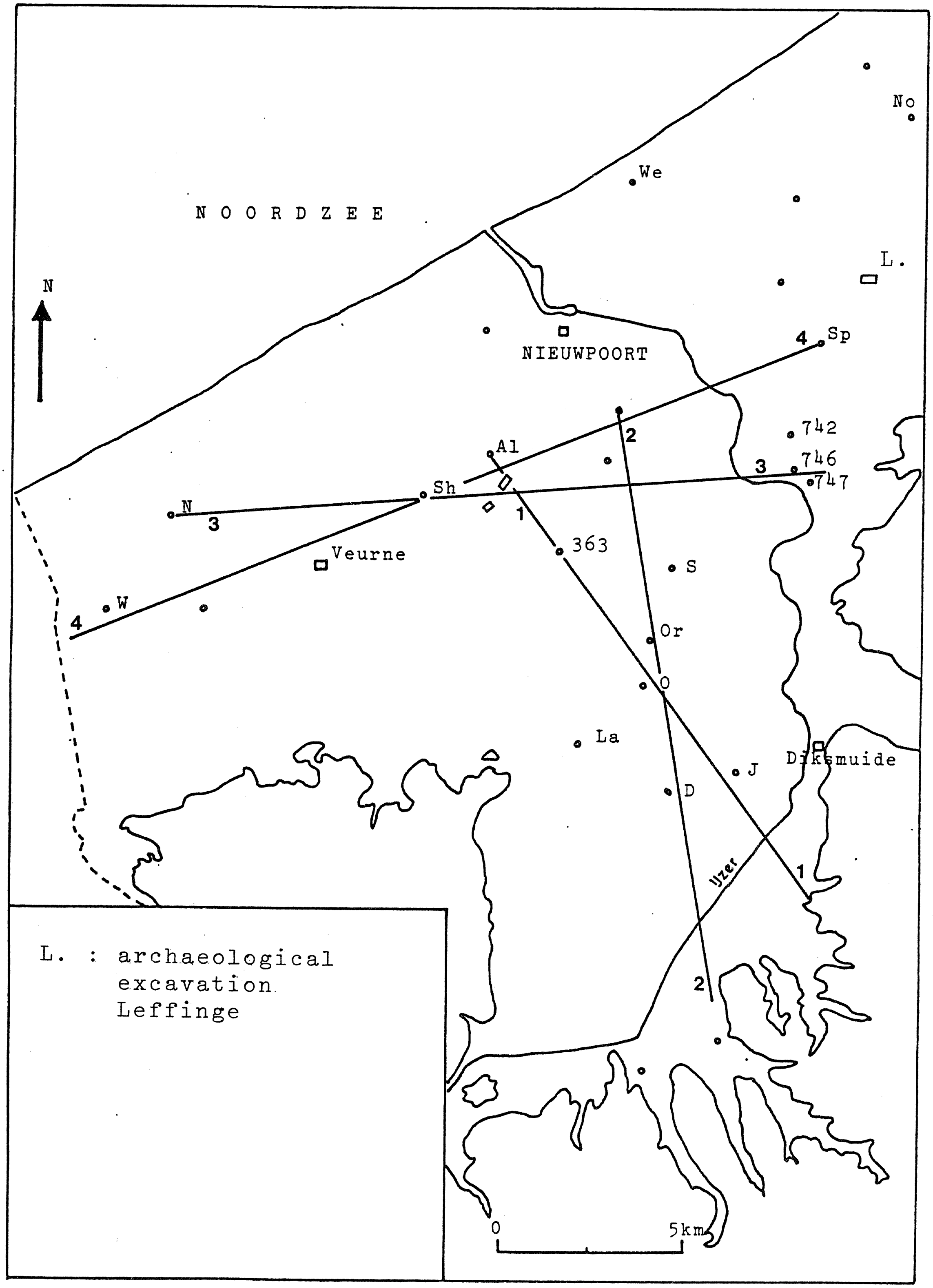


Fig. 5. Location map of the basal peat datings and profiles

area, revealed a similar age: 8120 ± 100 BP.

In the very western part of the coastal area a remarkable situation was observed (fig.5 & 6, profile 3, borehole N). A clayey peat, dated at 7620 ± 90 BP (base) is occurring at -13.71 to -13.75 m overlying a non-marine fine sand from which the humic content is decreasing with depth. Such a gradual transition is the most common sequence for basal peat overlying Pleistocene sands. However, slightly deeper, at -14.05 to -14.27 m another humic to peaty sand with wood remains at the top, also showing gradual upper and lower boundaries, was found overlying Pleistocene marine sands. The top of it was dated at 9190 ± 185 BP. Such a situation was never encountered until now. It is though that the sediments overlying the deepest peatlayer, which is to be considered as basal peat, did not accumulate due to the invading sea or due to a high groundwater level, but most probably are the result of dry conditions prevailing after the formation of the peatlayer, during which eolian sediments were deposited. However, only palynological investigations can be conclusive about this.

A second group of dates was collected from the area more landward where the Pleistocene subsurface is at a higher position (fig.5) :

7155 ± 270 BP (mean)	at -7.00 m	(borehole 363)
7230 ± 85 BP (base)	at -5.20 to -5.23 m	(borehole Or)
7110 ± 90 BP (top)	at -5.13 to -5.17 m	(borehole Or)
6870 ± 80 BP (base)	at -5.07 to -5.10 m	(borehole D)
6680 ± 80 BP (top)	at -4.90 to -4.93 m	(borehole D)
6665 ± 60 BP (top)	at -3.67 to -3.86 m	(borehole S)

This series of radiocarbon ages shows an age gradient in relation to the depth of the Pleistocene subsurface, except for the basal peat at borehole 363 which has the same age range as the dates from borehole Or, but occurs at a much

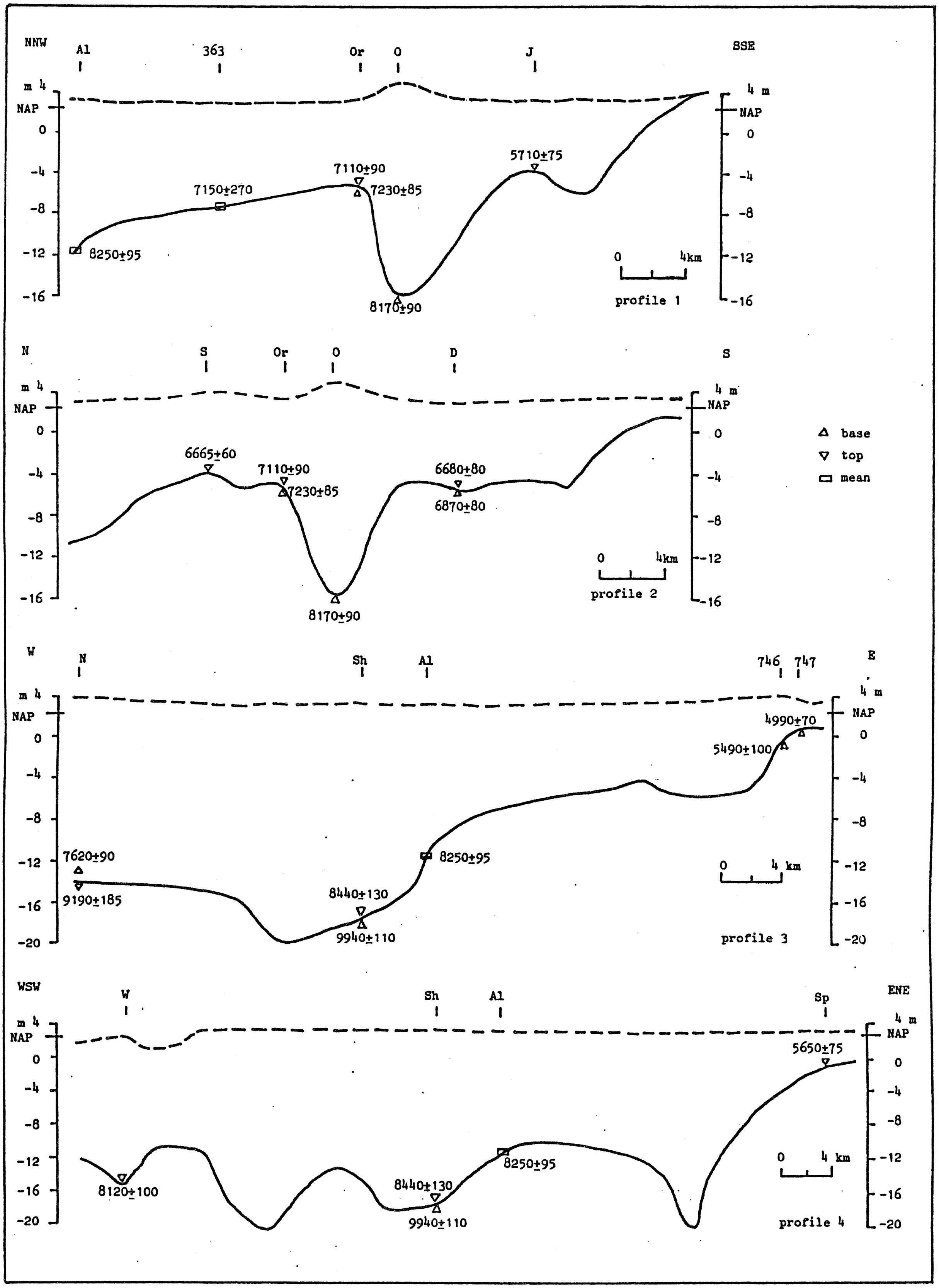


Fig. 6. Topography of the Pleistocene subsurface with indication of basal peat datings

deeper position. However, the sample from 363 is the whole of the layer and it has a large standard error.

Such a relatively high altitude for the Pleistocene subsurface is not only observed in the landward areas, but also in the very northeastern part of the surveyed area. Here only one site was dated (borehole We, fig 5), but revealed an apparant age inversion, viz. base : 6780 ± 80 BP at -5.14 to -5.19 m ; top : 7160 ± 85 BP at -5.03 to -508 m. Probably a stagnation in the peat growth, resulting in oxidation and mouldering, can be at the origin of the apparant age inversion.

Finally a last group of dates was collected from the area almost adjacent to the landward border of the coastal plain :

5970 ± 120 BP (base) at -2.10 to -2.25 m (borehole 742, fig.5)

5490 ± 100 BP (base) at -0.56 m (borehole 746)

4990 ± 70 BP (base) at +0.90 m (borehole 747)

5710 ± 75 BP (top) at -2.58 to -2.70 m (borehole J)

5650 ± 75 BP (top) at -1.04 m (borehole Sp)

5310 ± 190 BP (base) at -1.20 m (borehole La, fig.5)

5100 ± 140 BP (top) at -0.90 m (borehole La)

Here as well, the dates show a rather consistent age gradient with depth, although in this region it becomes critical to compare base with top from different loci as the basal peats are rather well developed (20 cm for borehole Sp and 80 cm for borehole J). Such a thick development of basal peat at higher altitudes implies a smaller degree of compaction on the one hand (the duration of compaction is shorter and the overburden is smaller), and on the other hand the peat growth prevailed for a much longer period in these areas where the Pleistocene subsurface is relatively high. Such a situation was observed in the very northeastern part of the area (borehole No, fig.5) where a basal peat of

2.5 m could develop continuously without being interupted by brackisch-marine sedimentation for about 3500 radiocarbon years. This basal peat revealed an age of : base: 5770 ± 100 BP at -2.30 to -2.37 m ; top: 2220 ± 55 BP at +0.21 to +0.17 m.

A similar situation was observed at the occasion of an archaeological excavation, called *Leffinge* , (fig.5) where the basal peat developed on an elevation of the Pleistocene subsurface. In that specific area, no intercalated peat layers are occurring, but the basal peat is found at almost the same level as the uppermost intercalated peat layer (see below). A series of dates from the basal peat in *Leffinge* revealed the following ages (previously published in Baeteman & Verbruggen, 1979, Baeteman, 1981a and Baeteman *et al.*, 1981) :

base :	5190 ± 140 BP at +1.00 m (wood)		(ANTW 105)
	4630 ± 140 BP at +1.20 m (wood)		(ANTW 102)
	4465 ± 220 BP at +1.80 m (peat)		(IRPA 282)
top at +2.20 m:	2960 ± 50 BP		(Hv 8800)
	3140 ± 165 BP		(IRPA 283)
	3340 ± 185 BP		(IRPA 337)
	3520 ± 60 BP		(ANTW 227)
	3225 ± 160 BP	(wood)	(IRPA 338)

The start of the peat growth in *Leffinge* most probably is in relation with the altitude of the Pleistocene subsurface, but its further evolution and end are to be compared to the uppermost intercalated peatlayer.

The splitting up sequence

The splitting up sequence is defined as a sedimentary succession between the bottom of the lowermost intercalated peat layer and the top of the uppermost one (fig.3). Thus it consists of peat layers (only one in a special case) as well as clastic sediments which lie between the intercalated peatlayers (Streif,1978).

Far too often the alternation of peat and clastic sediments has been regarded as an alternation of regressions and transgressions and hence the chronology of the peatlayers has acquired a chronostratigraphic association purely based on the subdivision in transgressions-regressions, which moreover is serving a basis for regional correlation and interpretation of climatic changes whenever it is fitting. Furthermore the transgressions and regressions were considered as directly reflecting vertical changes in sea level. But they are in no way synonymous with a rise and fall in sea level.

The ambiguity of the meanings of transgression and regression and their significance for sea-level changes have been thoroughly discussed by several authors, especially for the sake of interpreting sea-level related data and establishing regional correlations (Shennan, 1982a, 1982b, 1983, 1986, 1987, Shennan *et al.*,1983; Streif, 1979, 1982, ; Ludwig,*et al.*, 1981; Baeteman, 1981a, 1981b, 1987; Tooley 1982; Haggart,1988).

It is true, the alternating clastic and biogenic layers can be used to infer sea-level movements, but they also reflect recent earth movements, climatic changes, coastal processes and changes in sediment origin and supply from drainage basins and the continental shelf (Tooley, 1982). It never can be repeated enough, as clearly stated by Streif (1982) and Kraft & Chrzastowski (1985), that the vertical changes of the sea level are only one component among a great variety of factors which influence the development of a retreating or prograding coastline.

To avoid any further misinterpretation and inconsistencies about the sense of transgression and regression, the terms **transgressive overlap** and **regressive overlap** have been proposed by Shennan (1982a

& 1982b) and promoted by Tooley (1982) to use widely as descriptive terms in which no process is implied. This was suggested after the initial introduction of the term regressive overlap in the coastal Holocene by Streif (1979) who gave the following description : "a regressive overlap occurs simultaneously with a reduction of the marine influence on the sedimentary environment, and there is an accretion process with increasing biogenic production".

Intercalated peat datings

The radiocarbon dates available from the power-driven boreholes are represented on cross-sections 1, 2 & 4(fig.7, 8, 9 & 10), which form a more elaborated picture of the profiles 1,2 & 4.Indeed sequences of events must be compared spatially, including altitude, and through time. The entire Quaternary sequence occurring in the plain is represented in order to provide information in understanding the infilling and erosional processes of the coastal plain during the Holocene. The Pleistocene deposits, however are not differentiated yet. Anyhow it is clear that the topography of the Pleistocene and underlying Eocene deposits, both showing a substantial relief, influenced to a large extent the Holocene infilling.

The cross-sections demonstrate clearly the occurrence of the splitting up sequence in the central part of the plain where several peatlayers are intercalated in the tidal flat sediments. Toward the south, the splitting up sequence only consists of one peatlayer which moreover finally merges with the basal peat. On the other hand, the seaward region is characterised by less , and especially thinner peatlayers.

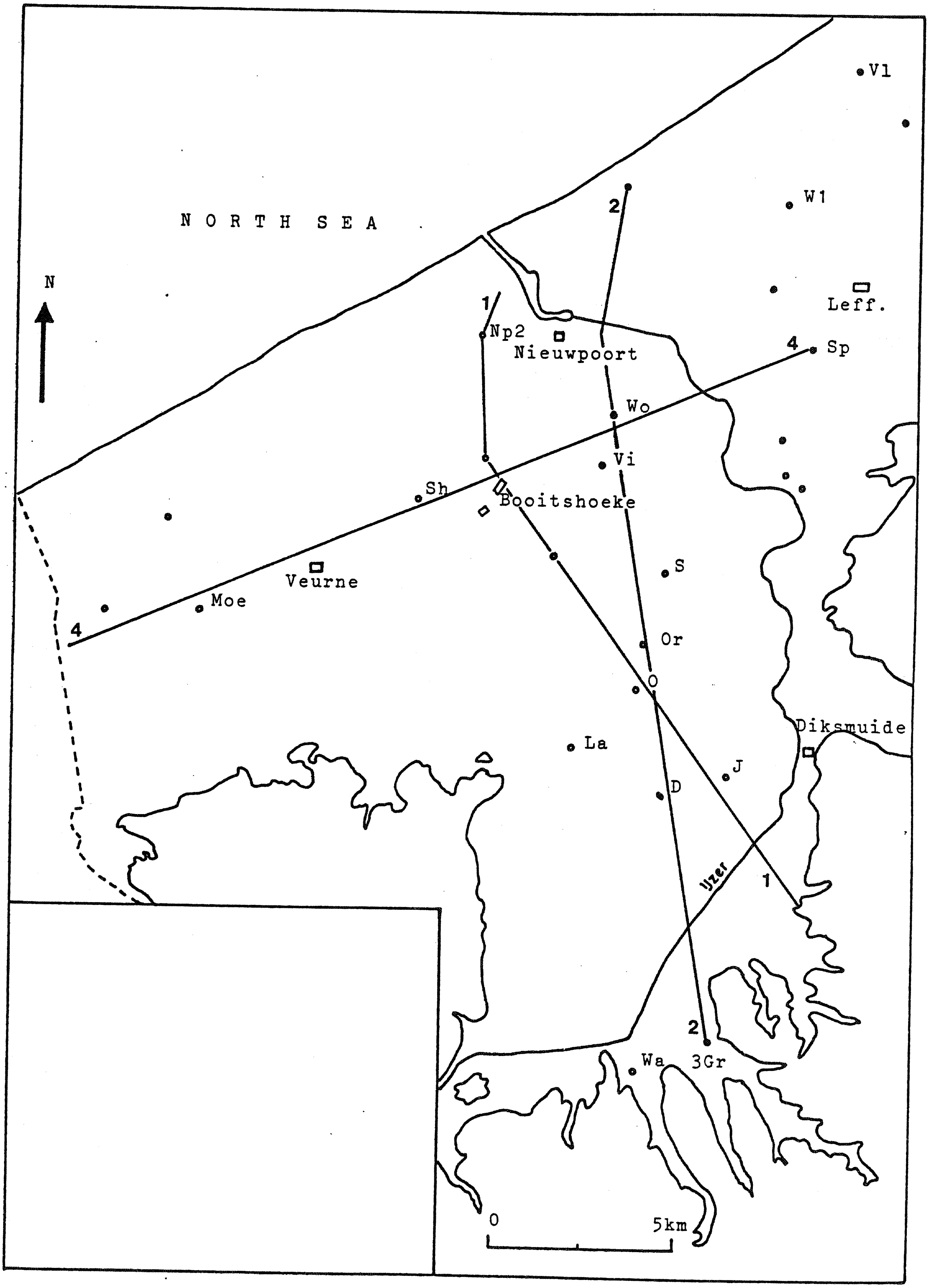


Fig. 7. Location map of the intercalated peat datings and cross-sections

The oldest and deepest known intercalated peatlayers were observed in depressions (cross-section 2), one of them was formed in the Pleistocene subsurface before the initial Holocene marine influence (Baeteman,1985).

The southernmost depression (borehole 3 Gr, cross-section 1 & 2, fig.7) is very much related to the fluvial system from the outcropping Pleistocene area. In the southern part, the coastal plain extends rather far southward in the valleys of the Pleistocene area. The sedimentary sequence of borehole 3 Gr, reaching a thichness of about 13 m, is remarkable in comparison with the rest of the plain. Preliminary palynological investigations revealed that the entire sedimentary seqence belongs to the Holocene (personal communication C. Verbruggen), resting directly on the Eocene clay. The lowermost part, under the peat layer, consists of fluvial sediments, *viz.* non-calcareous clay and silt with numerous plant remains, root penetrations and vivianite spots. It is covered by a clayey peat in which freshwater gastropods were observed. This peat occurs at an altitude of -6.24 to -6.85 m and the base of it revealed an age of 7030 ± 85 BP. The peat is then covered by tidal flat sediments reflecting the incoming marine influence in that area as far south. In these tidal flat sediments, another two intercalated peat layers were found; the deeper one at -3.98 to -4.28 m, the base of which revealed an age of 6500 ± 95 BP. The upper one will be discussed later.

In the depression of the central part of the plain, which has a quite different evolution than the southern depression, the deepest intercalated peatlayer was observed at a slightly lower altitude (-7.35 to -7.46 m, borehole O). The radiocarbon age from the base however, is similar : 7000 ± 80 BP ; the top was dated at 6750 ± 80 BP. It can be supposed that this deeper intercalated peat layer merges with the basal peat outside the depression (cross-section 2).

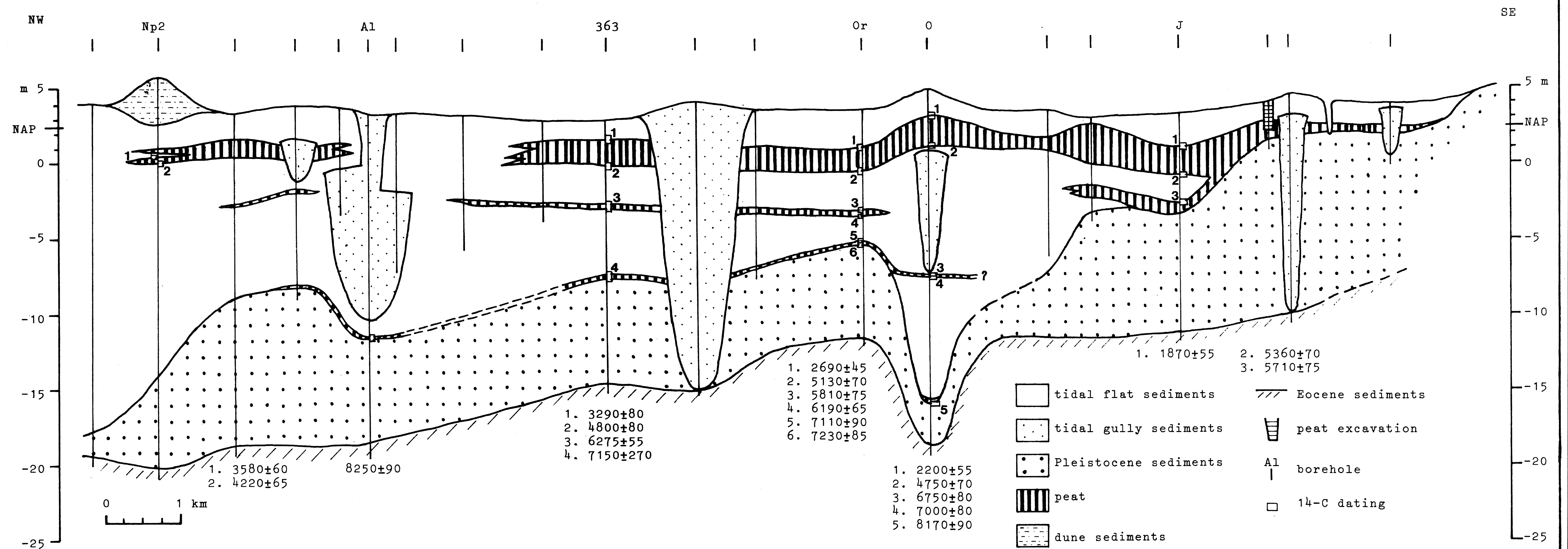


Fig. 8. Cross-section n°1

The deeper intercalated peat horizons from the depressions on the one hand, and the basal peats on top of the higher Pleistocene subsurface on the other hand, are all overlain by sand, silt and clay from the tidal flat sediments.

As from an altitude of about -3.5 m until about +3.0 m, the sequence then consists of an alternation of peatlayers and tidal flat sediments. Until now, 4 different peatlayers were observed,
but not always regularly at the same altitude or with the same extension and thickness. However, one peat horizon of this series of intercalated peats shows a rather regular extension and is most often found at an altitude ranging between -2.5 and -3.0 m (cross-section 1,2 & 4). This peatlayer was sampled at four locations (fig.7) :

borehole Wo : base : 6420 ± 80 BP at -2.73 to -2.77 m
top : 6200 ± 80 BP at -2.63 to -2.67 m
wood in top : 6160 ± 80 BP at -2,63 m
borehole S : base : 6375 ± 60 BP at -2.51 to -2.65 m
borehole Or : base : 6190 ± 65 BP at -3.33 to -3.38 m
top : 5810 ± 75 BP at -2.95 to -3.01 m
borehole D : top : 5550 ± 75 BP at -2.54 to -2.59 m

These radiocarbon dates agree rather well with two previously published dates from most probably the same peatlayer (Baeteman & Verbruggen, 1979 and Baeteman, 1981a);

borehole 363 : mean : 6275 ± 55 BP at -2.60 m
borehole 362 : base : 6015 ± 65 BP at -2.00 m

It should be noticed that the peat in borehole 362 is occurring on a higher elevation than generally observed.

Comparing the dates of this intercalated peat, it seems that the peat growth occurred slightly earlier in the seaward than in the landward part of the plain, although the number of data is far too insufficient to be conclusive on that.

The uppermost peatlayer of the series of intercalated peats is the most extended and thickest one. Its thickness almost reaches 1 to nearly 2 m and it is situated in general between the altitudes of ca -0.5 and +1 m. In the Belgian literature it is usually referred to as *surface peat* . This peat layer was sampled at several locations (cross-sections 1, 2 & 4; fig.7) :

borehole	**top** (in y BP)	**base** (in y BP)
Wo	2710 ± 60	4970 ± 70
Vi	_____	5160 ± 70
S	1185 ± 40	4920 ± 55
Or	2230 ± 40	5130 ± 70
	2690 ± 45 *	
O	2200 ± 55	4750 ± 70
3Gr	1750 ± 55	5220 ± 70
Np2	_____	4220 ± 65
J	1870 ± 55	5360 ± 70
Sc	_____	4540 ± 65
Wa	1610 ± 55	_____
We1	_____	5125 ± 55
Vl	2580 ± 60	4700 ± 70
Sp2	_____	4860 ± 70
Moe	_____	4830 ± 70

*The top of the peat in Or was sampled in two different boreholes at a distance of only few metres, and yet revealed a significant difference in age of nearly 500 radiocarbon ages. This demonstrates once more that data from boreholes always have to be evaluated critically.

The radiocarbon dates from the base of this uppermost intercalated peat layer show an age range of 5360 ± 70 BP to 4220 ± 65 BP. However the greatest number of dates reveals an age in a much smaller range, i.e. 4700 BP - 5220 BP in which moreover a concentration of dates is observed between resp. 4700 BP - 4970 BP and 5130 - 5160 BP.

The two youngest dates happen to come from locations in the seaward area (borehole Sh & Np2). On the other hand, the older dates were obtained from the peat occurring in the very landward part of the plain, rather close to the outcropping Pleistocene deposits (borehole J & 3Gr, cross-section 1 & 2, fig.7).

The comparison of the age of the base of the peatlayer with previously published dates shows a rather distict similarity. The base of the peat in the Booitshoeke-Avekapelle series (Baeteman & Verbruggen, 1979), locations which are situated in a relatively seaward part of the plain, was dated at 4770 ± 215 BP and 4800 ± 80 BP. A younger date of 4295 ± 195 BP was rejected as it was suspected to be younged by modern rootlet contamination. In that paper, a comparison was done with age determinations obtained from the peat at *Leffinge*, sampled at the occasion of an archaeological excavation located close to the outcropping Pleistocene area, revealing ages of 4465 ± 220 BP, 4630 ± 140 BP and 5190 ± 140 BP for an oak trunk below the peat. However, the peatlayer in question is in fact a basal peat, the onset of which has a completely different origin compared to an intercalated peat. Hence such a comparison is not relevant.

Still in literature, a much younger date of 4640 ± 65 BP was obtained from the peat in Lampernisse (La,fig.7), (Baeteman,1981a), while in 1985 and 1987, Baeteman stated that the peat growth started to develop at about 5000 yBP and 5300 yBP, respectively, based on dates from boreholes J and Or, discussed in this paper.

In very restricted zones of the plain, it is seen that the uppermost intercalated peat is split by a very thin clay intercalation in its upper part. At two locations

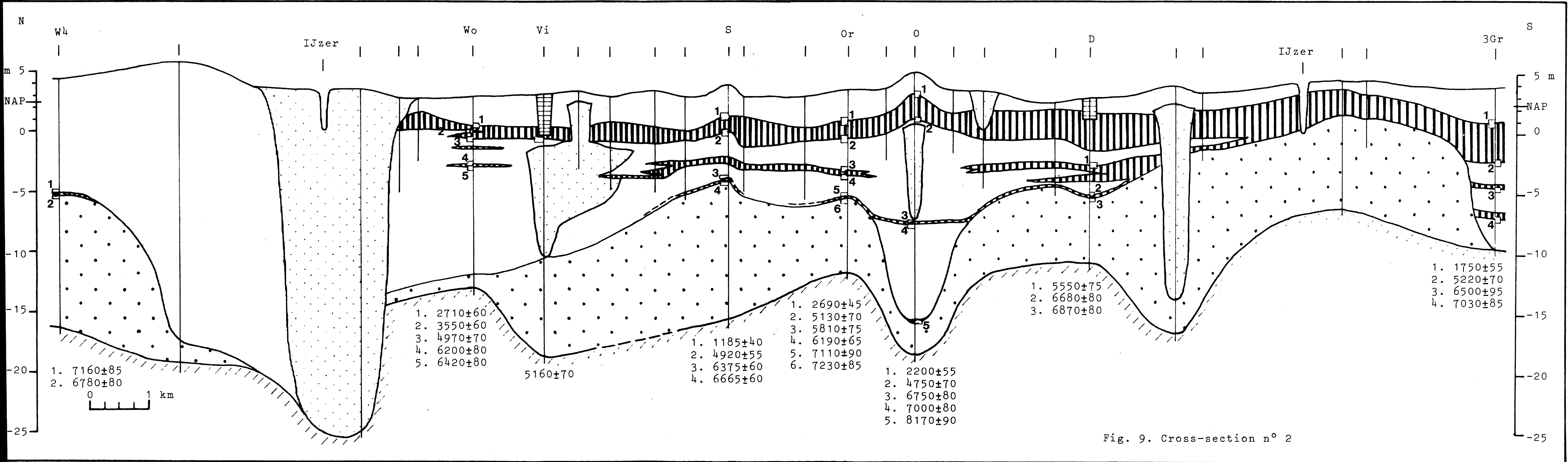


Fig. 9. Cross-section n° 2

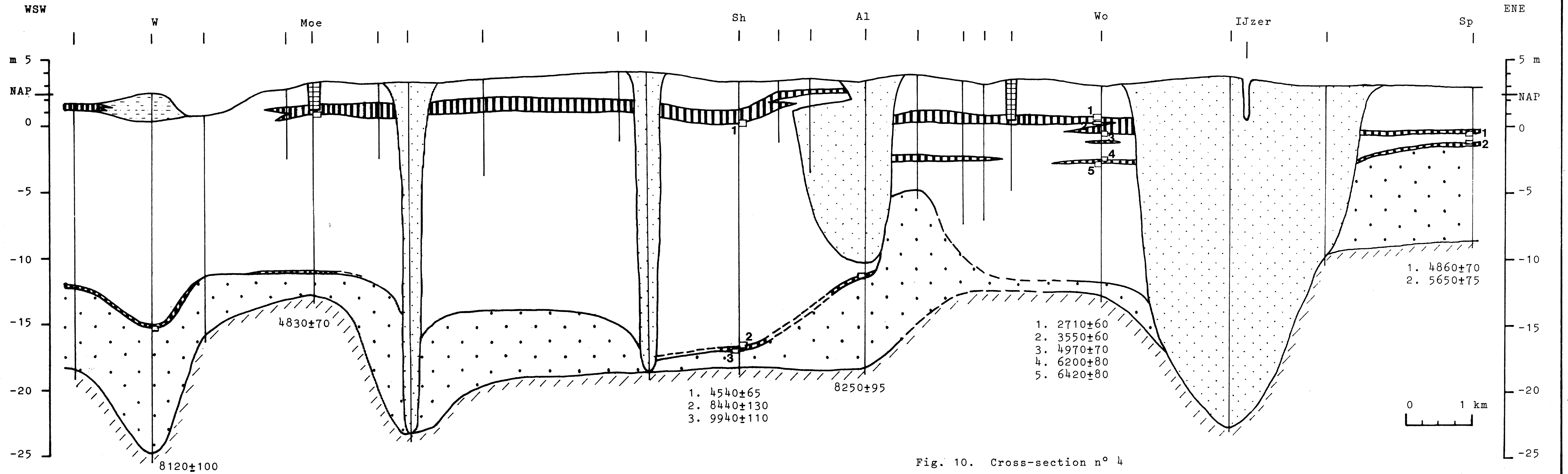


Fig. 10. Cross-section n° 4

showing such a situation, the transgressive overlap was sampled (respectively base and top in boreholes Np2 & Wo, cross-section 1 & 2), revealing a similar radiocarbon age : 3580 ± 60 BP and 3550 ± 60 BP, resp. (fig.11).

These ages are not in agreement with the ages of a similar situation located in the area of Booitshoeke-Avekapelle (southwest of Np2), where a thin intercalating clay layer in the uppermost peat was observed. Palynological investigations of the peat where the clay is not recorded, revealed significant wetter conditions in the peat growth which were correlated with the transgressive overlap. The top of the second peat horizon and the beginning of the wetter conditions were dated at : 4260 ± 210 BP; 3965 ± 190 BP and 4240 ± 190 BP (Baeteman & Verbruggen, 1979), which is about 400 to 700 radiocarbon years older than the situation in Np2 and Wo and most probably representing a different transgressive overlap recorded in the peat sequence.

The sequence of transgressive overlap in the upper intercalated peat from the Np2 and Wo boreholes was completed with dates from the uppermost peat horizon, sampled at the occasion of a temporary outcrop in Wulpen (Wu). Beginning and end of the regressive overlap were dated at 3490 ± 60 BP and 2970 ± 60 BP, respectively (fig.11).

In the northeastern zone of the surveyed area, another (new) sequence of the uppermost intercalated peatlayer was seen at location Vl (fig.7). The peatlayer, reaching a thickness of 1.34 m is underlain by a few-cm thick reedswamp mud, followed by another 36-cm thick peat horizon. The top and base of the underlying peat horizon was dated at 4820 ± 60 BP and 5540 ± 55 BP, respectively, indicating that between 4820 ± 70 BP and 4700 ± 70 BP, at location Vl, a brief period of clastic sedimentation took place and a transgressive overlap is recorded in the lower part of the uppermost intercalated peat sequence (fig.11).

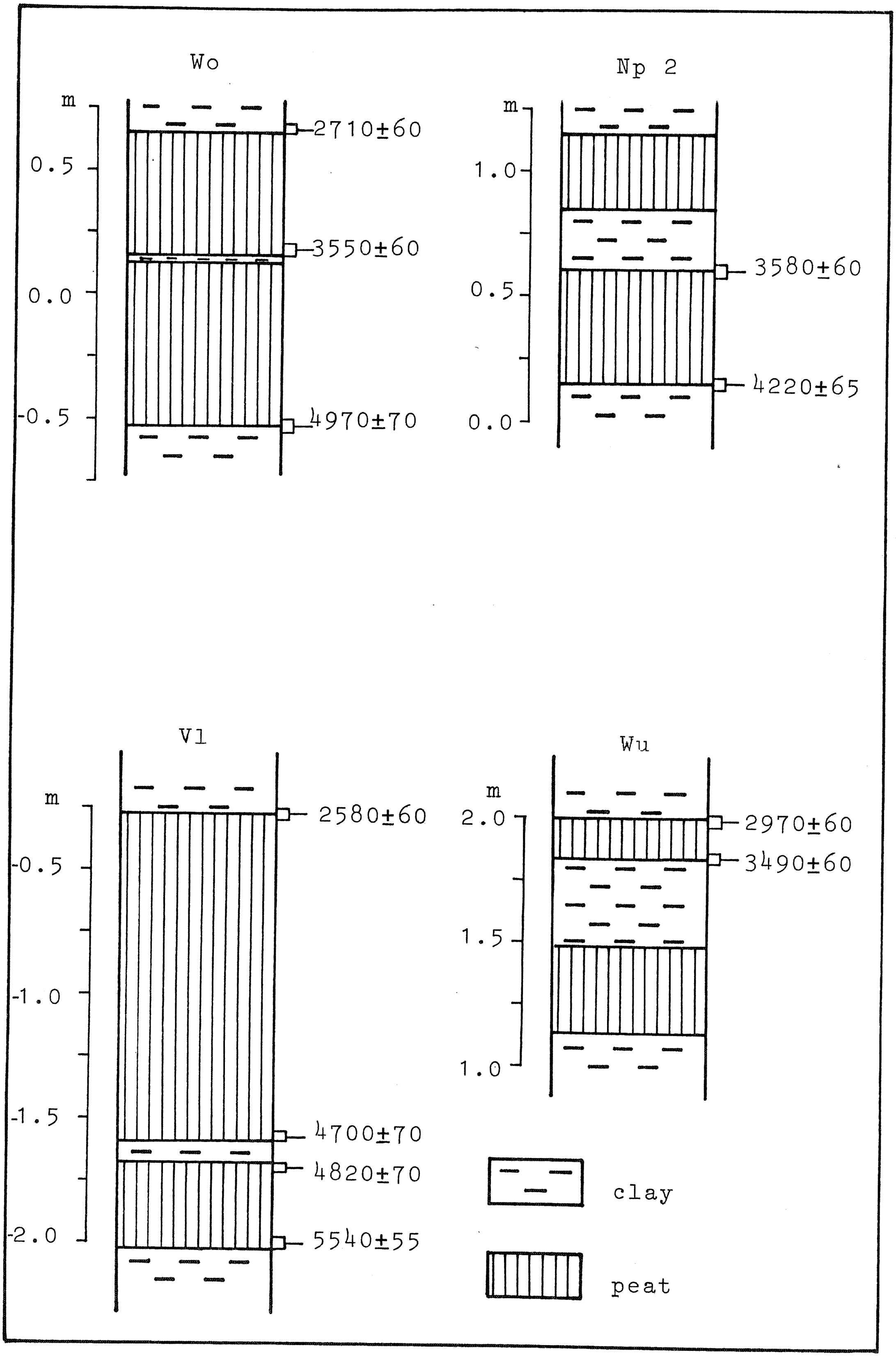


fig. 11. Datings of the transgressive overlaps in the uppermost peat sequence (in y BP)

The comparison of the radiocarbon ages from the top of the uppermost peat layer with previously published data shows smaller correspondence. In general it was stated that the end of the peat growth is to be situated at about 3000 - 3300 yBP and in some cases, more landward, at about 2000 yBP (Baeteman, 1981a, Baeteman, 1985, Baeteman, 1987b). Such a statement in fact is somewhat too general, because the radiocarbon dates from the top of the peat (which also here was sampled only when no erosional phenomena were observed) were grouped into two series : one ranging between 3000 and 3300 yBP and the second one ranging between 2000 and 2300 yBP (Baeteman & Verbruggen, 1979 and Baeteman,*et al.* 1981)), hence the preliminary conclusion was that after a general halt of the peat growth at about 3200 yBP, locally there was a resume until 2000 yBP.

The comparison of the previously published data and the data presented in this paper can corroborate this finding. However, the peat growth which lasted until about 2000 yBP is not to be considered any longer as a local situation, but on the contrary, most probably as the more general situation. Besides, the new results presented in this paper revealed that in the landward part of the plain, the peat growth came to an end in an even later period *viz.* between 1600 and 1900 yBP. The very young age (1185 ± 40 yBP) at borehole S forms a stricking exception, and the date is suspected to be younged by modern rootlet contamination.

FINAL CONSIDERATIONS

This paper presents a first stage of exploration of radiocarbon dates from the western coastal plain of Belgium. From the data available it is possible to draw a general overview of the chronology of sediment deposition in the plain during the Holocene. It is evident that further adjustment to the chronolgy is expected, as more data will become available.

The oldest known onset of the marine sedimentation, observed at an altitude of -16.60 m, occurred at 8440±130 BP. As from then a tidal flat, and more particularly sandflats and associated tidal gullies started to develop, characterised by a continuous deposition of clastic sediments. At the same time, the marine influence shifted landward and upward causing the end of the basal peat growth. The evidence available supports the assumption that a rather close correlation exists between the beginning of the peat formation and the altitude of the Pleistocene subsurface, and hence the sea-level height. The initial infilling of the residual valleys in the very southern part of the plain, however, occurred independently on the contemporary sea-level height and is therefore to be considered as a local phenomenon bearing its own development and evolution.

As from 7000 BP a significant change in the general tendency took place. In the residual valleys, previously infilled with sandflat deposits and fluvial deposits in the southern areas, peat growth started while in the rest of the (contemporary) plain, mudflats and salt marshes developed. As from 6400 BP, however, general peat growth is observed over nearly the entire (contemporary) plain. From the evidence available it seems that the peat growth started slightly earlier in the seaward areas.

This peatlayer represents the onset of what is usually called the typical cyclic formation of coastal deposits where peat repeatedly came into being alternating with the deposition of tidal flat sediments. As from the beginning of the period of cyclic formation, tidal flat sediments were deposited far south, evidently only in these zones where the Pleistocene subsurface is not at a too high elevation. The cyclic formation with the intercalated peatlayers, indicating temporary regressive tendencies, generally came to an end in the time interval of 2700 - 2200 BP, and in more landward areas, between 1900 and 1600 BP.

At present there are still too few datings available to permit proper statistical treatment. As the deposits of a coastal plain are mainly characterised by very frequent facies changes, due to the general upward and landward movement of the transgressive coastline and associated sedimentary environments in space and time, there is no single site at which all the overlaps are recorded. It is sure that still a lot of transgressive and regressive overlaps have not been sampled yet. Besides, it was put forward (Shennan, 1982) that a minimum of 40 dates per 1000 years is necessary to yield reliable, non random, results.

Therefore, only a denser, and in particular a regular spatial distribution of dated samples, combined with palynological investigations, will permit to establish a time pattern of regressive and transgressive overlaps and subsequently the periods of positive and negative tendencies of sea-level movements.

Table 1 ***Radiocarbon dates form basal peat and intercalated peat layers from the Western coastal plain.***

Bore-hole	Altitude (m T.A.W.)	14-C years BP	Calibrated ages	Lab. nr.
Sh	-16.92 to -16.97	9940 ± 110		IRPA 680
Sh	-16.64 to -16.67	8440 ± 130		IRPA 681
Sh	0.05 to 0	4540 ± 65	3363-3102 cal BC	IRPA 682
Al	-11.27 to -11.34	8250 ± 95		IRPA 566
O	-15.60 to -15.64	8170 ± 90		IRPA 734
O	- 7.44 to - 7.48	7000 ± 80	5972-5744 cal BC	IRPA 536
O	- 7.34 to - 7.37	6750 ± 80	5716-5552 cal BC	IRPA 535
O	+ 1.08 to + 1.05	4750 ± 70	3637-3380 cal BC	IRPA 868
O	+ 3.15 to + 3.10	2200 ± 55	377-190 cal BC	IRPA 867
W	-15.17 to -15.25	8120 ±100		IRPA 616
N	-14.05 to -14.12	9190 ±185		IRPA 677
N	-13.73 to -13.75	7620 ± 90	6558-6409 cal BC	IRPA 678
Or	- 5.20 to - 5.23	7230 ± 85	6131-5986 cal BC	IRPA 533
Or	- 5.13 to - 5.17	7110 ± 90	6080-5845 cal BC	IRPA 534
Or	- 2.95 to - 3.01	5810 ± 75	4783-4586 cal BC	IRPA 612
Or	- 0.46 to - 0.49	5130 ± 70	3999-3818 cal BC	IRPA 532
Or	+ 2.01 to + 1.94	2230 ± 40	384-206 cal BC	IRPA 847
Or	+ 1.10 to + 1.05	2690 ± 45	899-810 cal BC	IRPA 832
D	- 5.07 to - 5.10	6870 ± 80	5813-5641 cal BC	IRPA 542
D	- 4.90 to - 4.93	6680 ± 80	5639-5489 cal BC	IRPA 541
D	- 2.54 to - 2.59	5550 ± 75	4467-4346 cal BC	IRPA 613
S	+ 1.33 to + 1.26	1185 ± 40	717-959 cal AD	IRPA 826
S	+ 0.56 to + 0.49	4920 ± 55	3781-3649 cal BC	IRPA 848
S	- 2.51 to - 2.65	6375 ± 60	5378-5210 cal BC	IRPA 871
S	- 3.67 to - 3.86	6665 ± 60	5632-5489 cal BC	IRPA 927
We	- 5.14 to - 5.19	6780 ± 80	5732-5573 cal BC	IRPA 615
We	- 5.03 to - 5.08	7160 ± 85	6093-5965 cal BC	IRPA 614
J	- 2.58 to - 2.70	5710 ± 75	4713-4468 cal BC	IRPA 617

J	- 0.80 to - 0.82	5360 ± 70	4341-4049 cal BC	IRPA 538
J	+ 1.13 to + 1.10	1870 ± 55	74-218 cal AD	IRPA 537
Sp	- 1.04 to - 1.09	5650 ± 75	4656-4399 cal BC	IRPA 519
Sp	- 0.29 to - 0.39	4860 ± 70	3774-3538 cal BC	IRPA 518
No	- 2.30 to - 2.37	5770 ±100	4780-4510 cal BC	IRPA 729
No	+ 0.21 to + 0.17	2220 ± 55	386-197 cal BC	IRPA 730
3 Gr	- 6.75 to - 6.85	7030 ± 85	5985-5759 cal BC	IRPA 520
3 Gr	- 4.29 to - 4.35	6500 ± 95	5500-5340 cal BC	IRPA 515
3 Gr	- 2.16 to - 2.22	5220 ± 70	4218-3982 cal BC	IRPA 531
3 Gr	+ 1.07 to + 1.03	1750 ± 55	223-346 cal AD	IRPA 521
Wo	- 2.73 to - 2.77	6420 ± 80	5476-5244 cal BC	IRPA 561
Wo	- 2.63 to - 2.67	6200 ± 80	5238-5059 cal BC	IRPA 559
Wo	- 0.49 to - 0.53	4970 ± 70	3931-3696 cal BC	IRPA 560
Wo	+ 0.18 to + 0.15	3550 ± 60	2011-1782 cal BC	IRPA 860
Wo	+ 0.68 to + 0.63	2710 ± 60	916-812 cal BC	IRPA 859
Vi	- 0.40 to - 0.45	5160 ± 70	4036-3824 cal BC	IRPA 562
Np 2	+ 0.19 to + 0.12	4220 ± 65	2913-2700 cal BC	IRPA 726
Np 2	+ 0.61 to + 0.57	3580 ± 60	2032-1883 cal BC	IRPA 727
We 1	- 0.75 to - 0.80	5125 ± 55	3994-3820 cal BC	IRPA 846
Moe	+ 0.95 to + 0.90	4830 ± 70	3698-3525 cal BC	IRPA 564
Wu	+ 1.83	3490 ± 60	1897-1743 cal BC	IRPA 527
Wu	+ 2.00	2970 ± 60	1310-1099 cal BC	IRPA 528
742	- 2.10 to - 2.25	5970 ±120	5048-4770 cal BC	IRPA 725
		6700 ±125	5711-5480 cal BC	IRPA 725B
746	- 0.46 to - 0.56	5490 ±100	4439-4290 cal BC	IRPA 722
747	+ 0.99 to + 0.89	4990 ± 70	3939-3701 cal BC	IRPA 723
Wa	+ 2.09 to + 2.04	1610 ± 55	393-533 cal AD	IRPA 872
V1	- 1.91 to - 1.96	5540 ± 55	4460-4348 cal BC	IRPA 924
V1	- 1.64 to - 1.68	4820 ± 70	3696-3524 cal BC	IRPA 866
V1	- 1.52 to - 1.56	4700 ± 70	3620-3371 cal BC	IRPA 865
V1	- 0.23 to - 0.26	2580 ± 60	809-769 cal BC	IRPA 512

363	- 7.00	7170 ±275		HV 8797
363	- 7.20	6580 ±1730	7155 ±270 BP (mean)	HV 8798
363	+ 1.75	3335 ±170		IRPA 336

363	+ 1.75	3290 ± 80		HV 8793
363	+ 0.10	4800 ± 80		HV 8794
363	- 2.50	6340 ±110		HV 8795
363	- 2.70	6245 ± 70	6275 ±55 BP	HV 8796
362	- 2.00	6015 ± 65	(mean)	HV 8799

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PALAEOGEOGRAPHICAL RECONSTRUCTION OF THE OFFSHORE AREA OFF THE BELGIAN COAST - ACOUSTIC INVESTIGATIONS

Stanislas WARTEL

INTRODUCTION

Since 1986 the subbottom of the continental shelf off the Belgian coast has been surveyed. An area extending from Nieuwpoort to Zeebrugge and some 14 Km seaward has been investigated. The study aims to contribute to the reconstruction of the palaeoenvironment during the Quaternary, and more specifically during the Holocene, in this area. Indeed, on the basis of the seaward extension of the uppermost intercalated peat layer, as was observed by Baeteman (1981) it can logically be assumed that beach and dune formations of Quaternary age occur on some parts of the continental shelf off the Belgian coast. The shoreward extension of thick peat deposits is one of the indicators for a more northwestward located older coastline.

The subbottom survey also inted to supply information for the interpretation of a network of borings recovered on the continental shelf, offshore the Belgian coast, during a vast drilling project of the Belgian Geological Survey.

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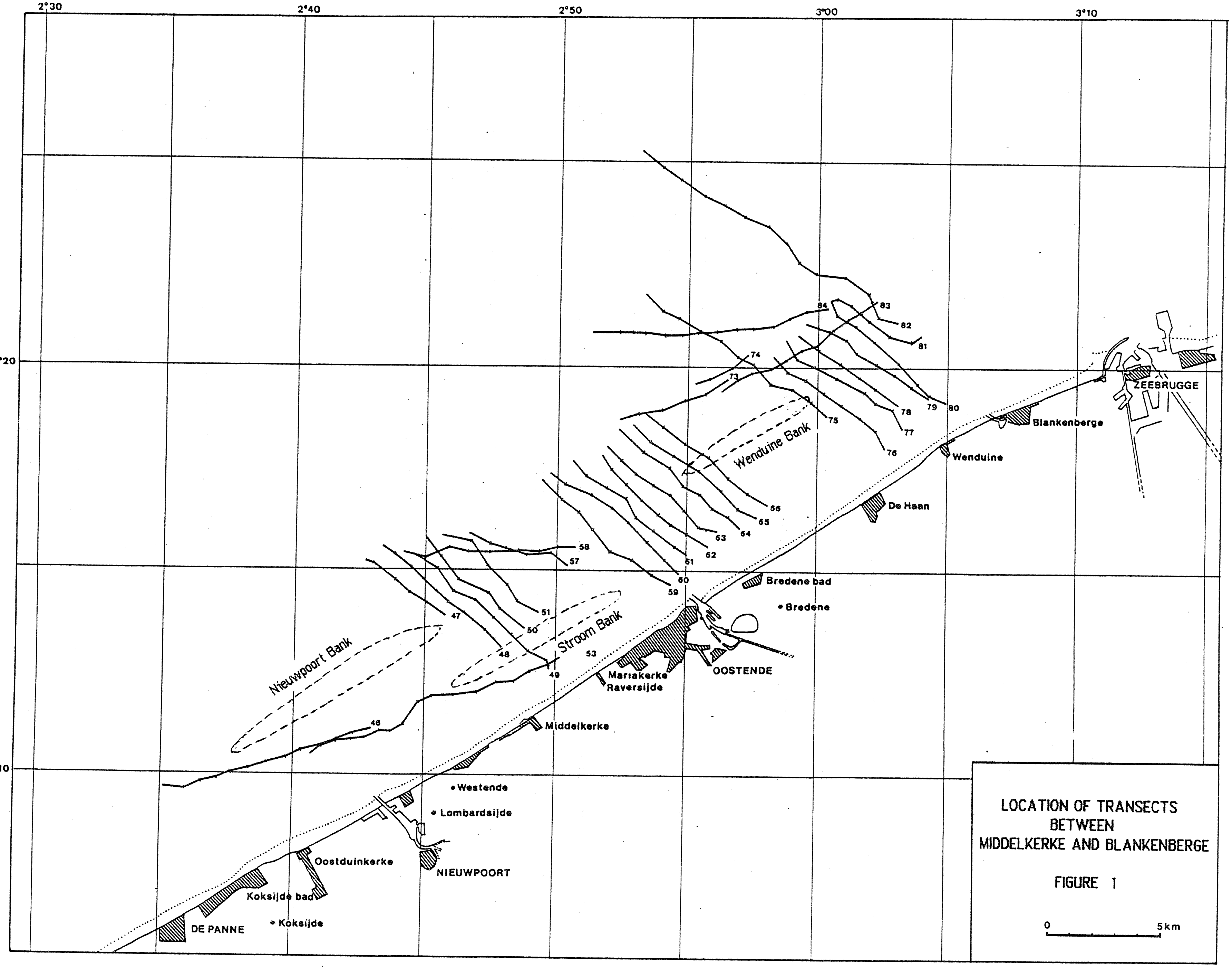


LOCATION OF TRANSECTS
BETWEEN
MIDDELKERKE AND BLANKENBERGE

FIGURE 1

K.B.I.N. M.P.

METHODS

70 transects, most perpendicular to the coastal sand bars, in total over 350 km in length, were investigated on board of the Belgian oceanographic ship "BELGICA". An area of 515 Km^2 has been covered. The subbottom profiling was performed using a ORE 3.5 kHz subbottom-profiler and the data were recorded on an EPC 2000 graphic recorder. Part of the data were also stored on a NAGRA IV professional tape recorder for subsequent laboratory analyses.

Subsurface sediments, up to 4m in depth, were sampled in PVC tubes, using a pneumatic vibrocorer. The bulk density of these sediments was measured on the unopened PVC tubes using a gammadensitometry technique. A 1 mCi Ba133 radioactive source was used for this purpose, and the continuous density profile was recorded on a YT-recorder. Radiographs of the cores were obtained using a 80 KeV X-ray tube at 5mA.

ACOUSTIC INVESTIGATIONS

The internal structures of the Nieuwpoort Bank (figure 1) have been described in a previous publication (De Maeyer et all, 1985). A landward migration of the bank, over the underlying older deposits was postulated. Underneath the bank 3 major horizons can be recognized acoustically. A boring, 4m in depth, near the front of the bank (figure 5 : core 83A04) confirmed this observation. The lowermost horizon consists of the uppermost part of a stiff clay layer, the Ypresian Clay (Eocene) as was confirmed by the microfossil assemblages (J. De Coninck, unpublished data). It has a bulk density larger than 2. The homogeneous character of this clay is demonstrated by the very uniform bulk-density distribution in this section of the core (figure 2: -235cm and deeper). The top of this Eocene clay produces a relatively strong acoustic reflection, easily recognized on the acoustic records and thus produces a relatively good tracer in this area. At several places small gullies are incised in the top of the clay. Also larger gullies occur; some of them are incised beyond the range of the acoustic signal used so the deepest part of the incision was not always recorded. The gullies are younger than the Eocene clay and are pre-Holocene or Holocene in age. In relation to this reference cqn be made of the AveKapelle tidal-gully observed near Nieuwpoort (Baeteman, 1985) and having a pre-Holocene age.

The clay is overlain by a 1 m thick layer extending from -240 to -140 cm below the sea floor. The lower part of this layer is composed of sand and gravels (mostly mud pebbles probably eroded from the underlying clay) (figure 2, -200 to -235 cm) passing upward into a 60 cm thick shell layer (-140 to -200 cm). The bulk density ranges between 1.9 and 2.4, except for a small mud lens at -152 cm wich has a density of 1.6. These densities are of the same magnitude as the

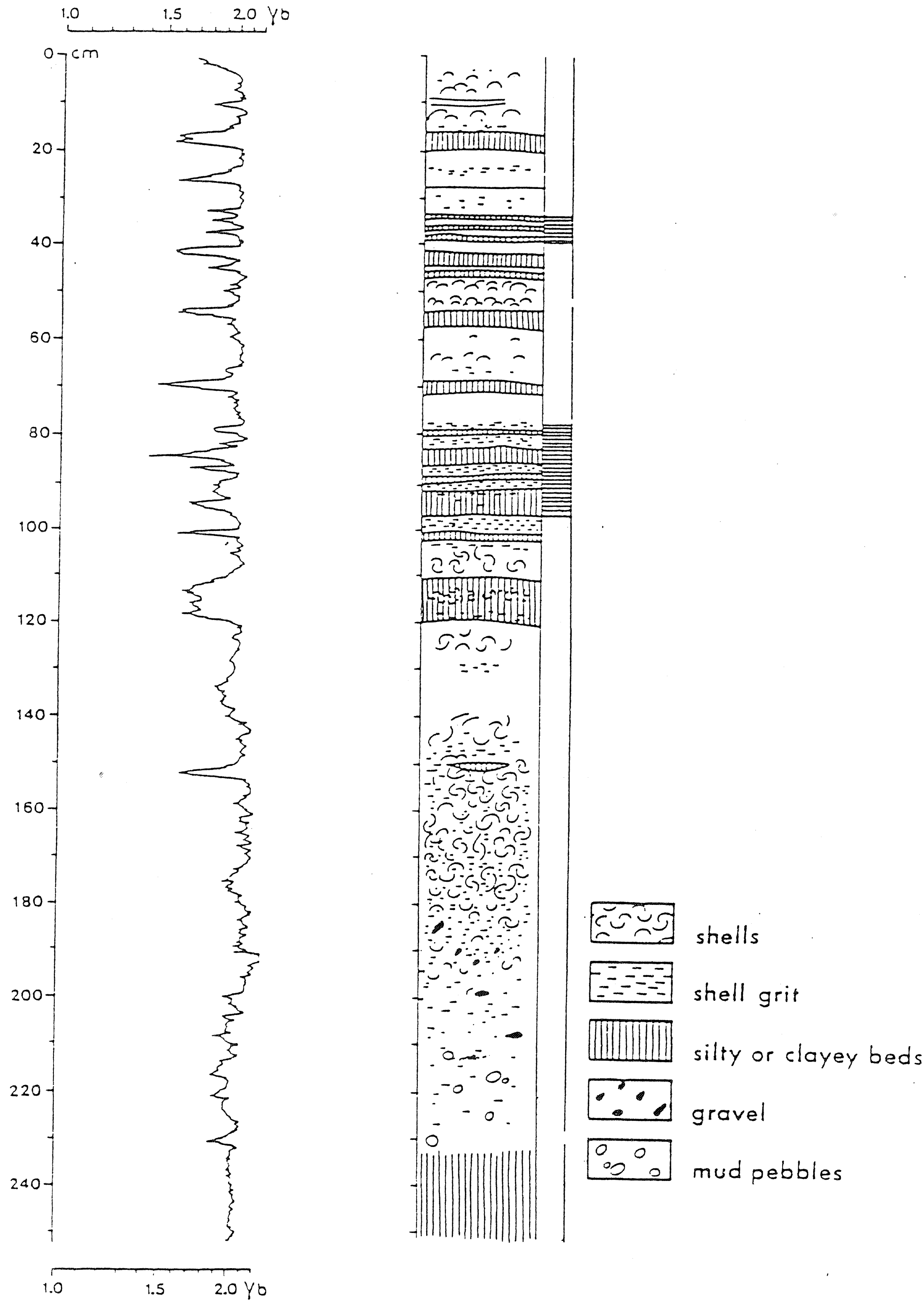


FIGURE 2 : Bulk density distribution and lithological description of boring 83A04. Depths are given below the top of the core

underlying Eocene clay. However, they are higher than similar modern deposits occurring on the continental shelf and having a density which is generally lower than 1.8. This indicates that the sand, gravel and shell layer was subjected to a considerable compaction either due to a sediment cover which must have been thicker than the actual one (only 1 m), or due to the weight of the actual sediment cover but acting over a very long time span. However, at this time no data are available for confirmation.
The density fluctuations observed in this layer result from differences in the relative abundance of shells.

The shells found in this layer are comparable to the so-called Angulus Pygmaeus association. This association was studied in the Zeeland Banken by Laban et al. (1981), who attributed an Atlantic to Subboreal age to it. From this it can be postulated that the sand gravel and shell layer underneath the Nieuwpoort Bank is comparable in age to similar deposits underneath the Zeeland Banken.
The occurrence of gravels and the random orientation of the shells furthermore indicate that these sediments were deposited in a highly energetic environment, presumably the nearshore or the beach.
This horizon occurs at depths between -16 and -17m near the Nieuwpoort Bank and can be followed acoustically in the direction of the shore where it is occurring at a higher altitude (-11 m).

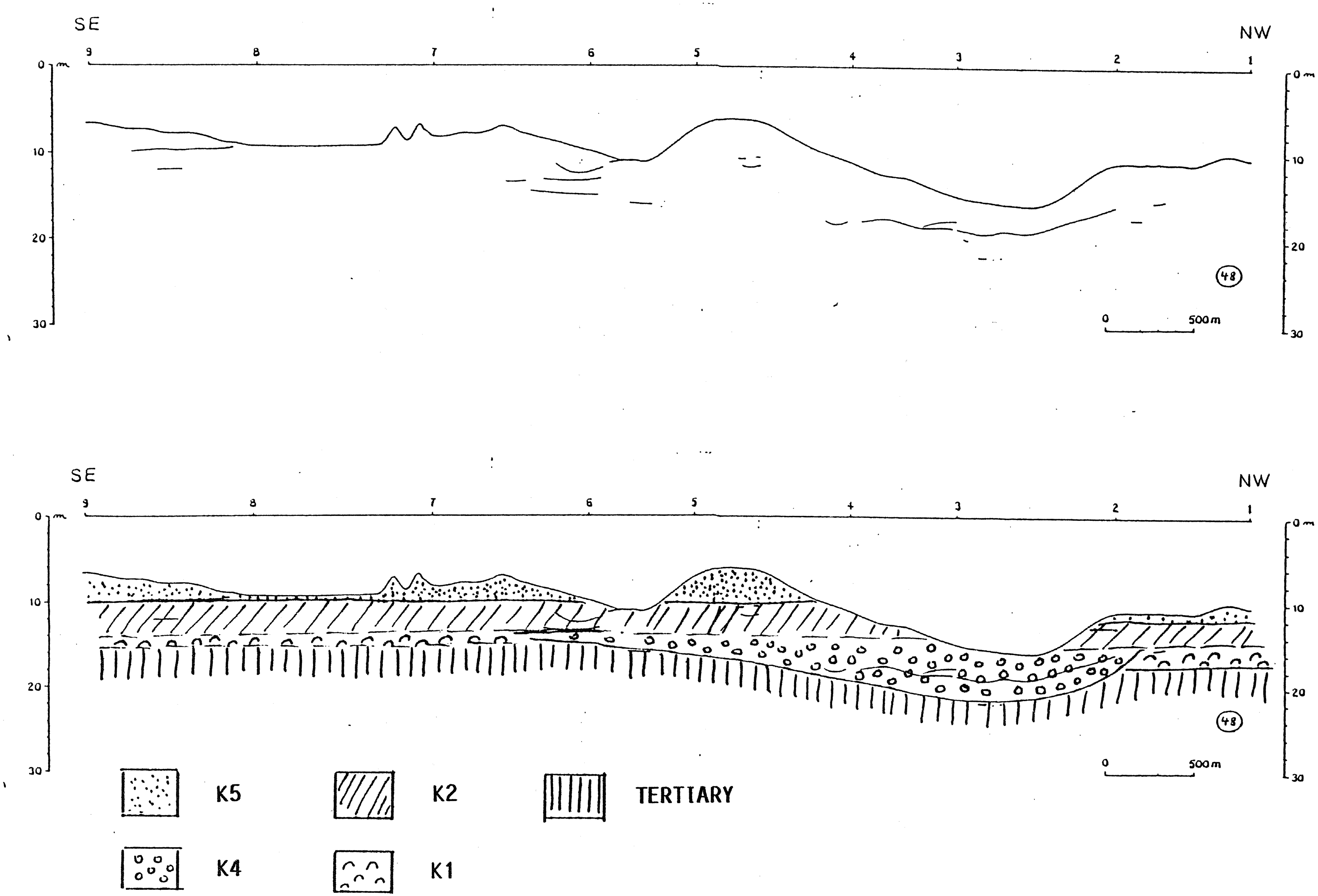
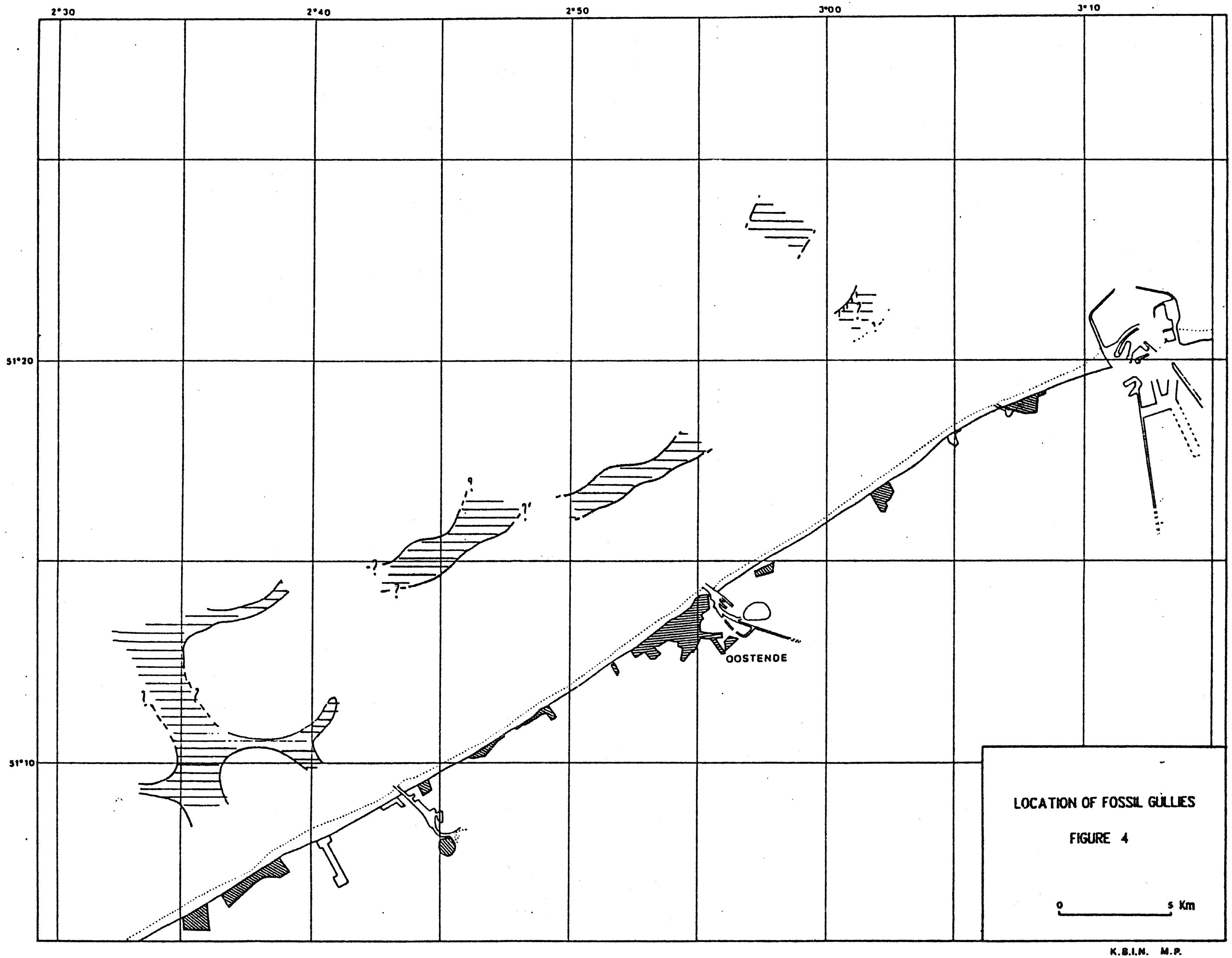


FIGURE 3 : Subbottom profile (A) of transect 48 (see figure 1) and a tentavie interpretation (B). of this profile. See text for explanation of symbols

The sand, gravel and shell horizon is overlain by deposits with somewhat lower densities (1.7 à 1.8) and showing an alternation of sand and clay beds or laminae with occasionally some shells, all deposited with their concave sides upward. The general sedimentological aspect of this horizon suggests deposition in an intertidal environment. This is furthermore emphasized by the occurrence of several relatively small gullies.

A similar succession of horizons can be recognized in most of the subbottom profiles. It extends in a northeastern direction. Figure 3 shows a subbottom structure as was observed in some 20 transects in the area between Middelkerke and Wenduine (figure 1). Between positions 2 to 4 a large gully (figure 3 : horizon K4), eroding deeply incised the underlying Tertiary deposits, can be recognized. The Tertiary deposits are overlain by two layers showing some similarities with layers observed in the Nieuwpoort Bank area. In the uppermost horizon (K2) a small gully can be seen near position 6. No gullying is seen in the lowermost horizon (K1). Since at this time no correlation between the acoustic results and borings has been made any comparison with the Nieuwpoort Bank area remains uncertain.

Besides small gullies also larger gullies occur in horizon K2. Although the acoustical reflection records are not always such that the uppermost level of these gullies can be recognized clearly, it seems that most of the gullies



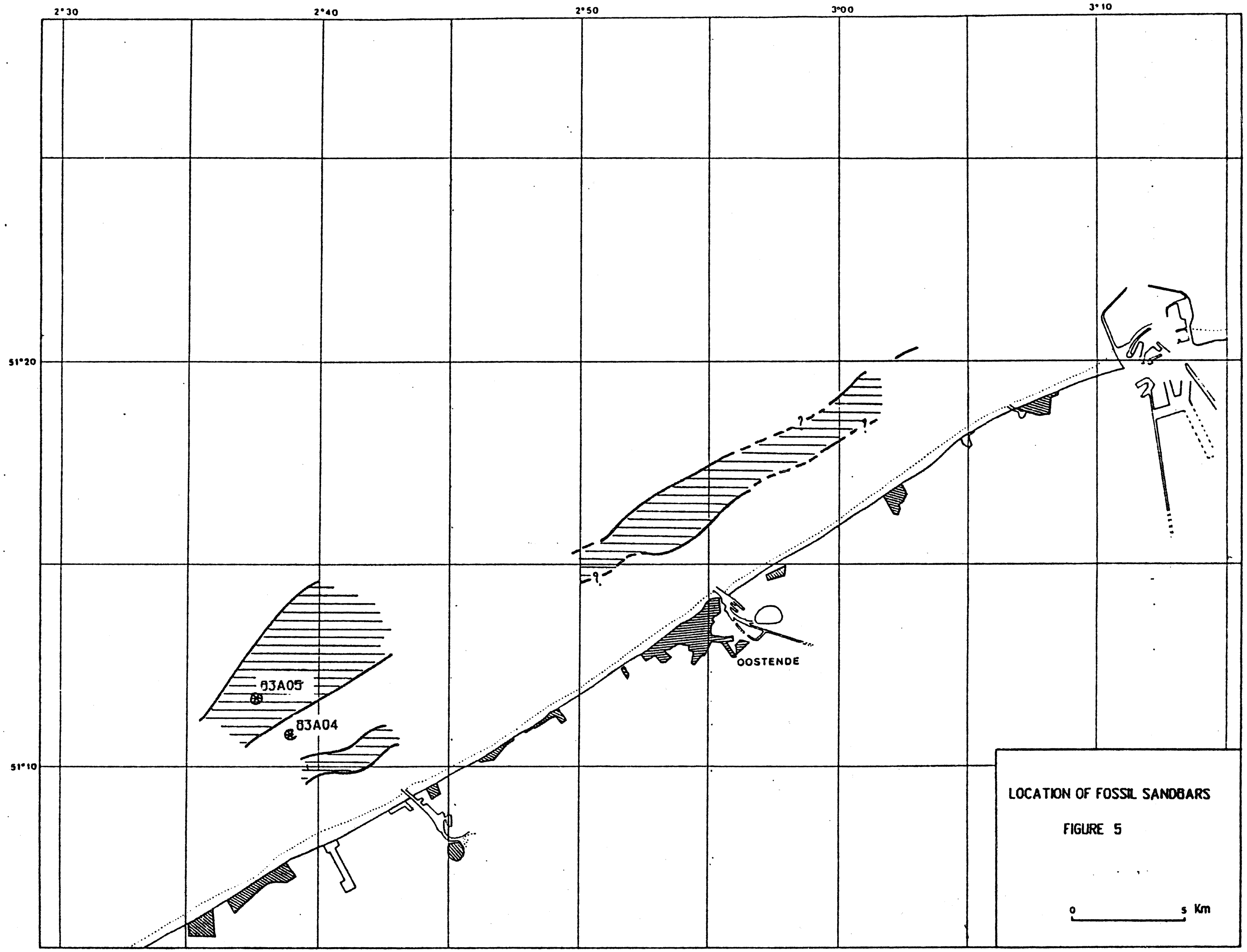
LOCATION OF FOSSIL GULLIES

FIGURE 4

originate in horizon K2. This observation may confirm the previously postulated tidal environment for this environment. Furthermore these larger gullies were easily recognized in adjacent transects and a mapping wass elaborated. Figure 4 shows their areal distribution. However, it must be emphasized that this map does not imply yet any stratigraphic relationship between the gullies mutually.

Apparantly one of these gullies correlates well with a gully observed in the coastal plain area. The nortwestward oriented gully off Nieuwpoort lies in the seaward extension of the Avekapelle tidal gully observed by Baeteman (1981, 1985) and is comparable in size.

In several subbottom profiles it was seen that horizon K2 is overlain by a larger sediment body in which interior acoustic reflections are deficient. The areal distribution of these sediment bodies is shown in figure 5. They were mapped resting on their seaward and, or landward dipping edges. Grain size analyses of a core sampled in the sediment body north of the Nieuwpoort Bank (figure 5 : core 83A05) reveals a relatively uniform grain-size distribution composed mainly of medium-sized sand and having a median diameter ranging between .170 and .2mm. A minor fraction of 1% coarser than .4mm also occurs. No data are available yet as to the age of these sediments. It is logical to assume that they are younger than the underlying deposits of horizon K2, and that they are older than the recent deposits building up the Nieuwpoort Bank and



LOCATION OF FOSSIL SANDBARS

FIGURE 5

K.B.I.N. M.P.

the Wenduine Bank (figure 3 : horizon K5) and covering them partly.

ACKNOWLEDGEMENTS

This study was part of the Offshore Programme from, and was granted by, the Quaternary department of the Belgian Geological Survey (contract n° KWA/86/3.1.2.b).

The author is very indebted to Dr J. De Coninck (RUG) for the determination of microfossils in the Ypresian Clay and to Dr J. Bakeljau (KBIN) for the determination of shells. He is also indebted to C. Baeteman for critical reading of the manuscript.

Finally he also acknowledge the Management Unit of the Mathematical Model North Sea for the use of the oceanographic vessel BELGICA and the crew of the BELGICA for their appreciated cooperation.

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