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THE QUATERNARY AND TERTIARY GEOLOGY OF THE SOUTHERN BIGHT, NORTH SEA

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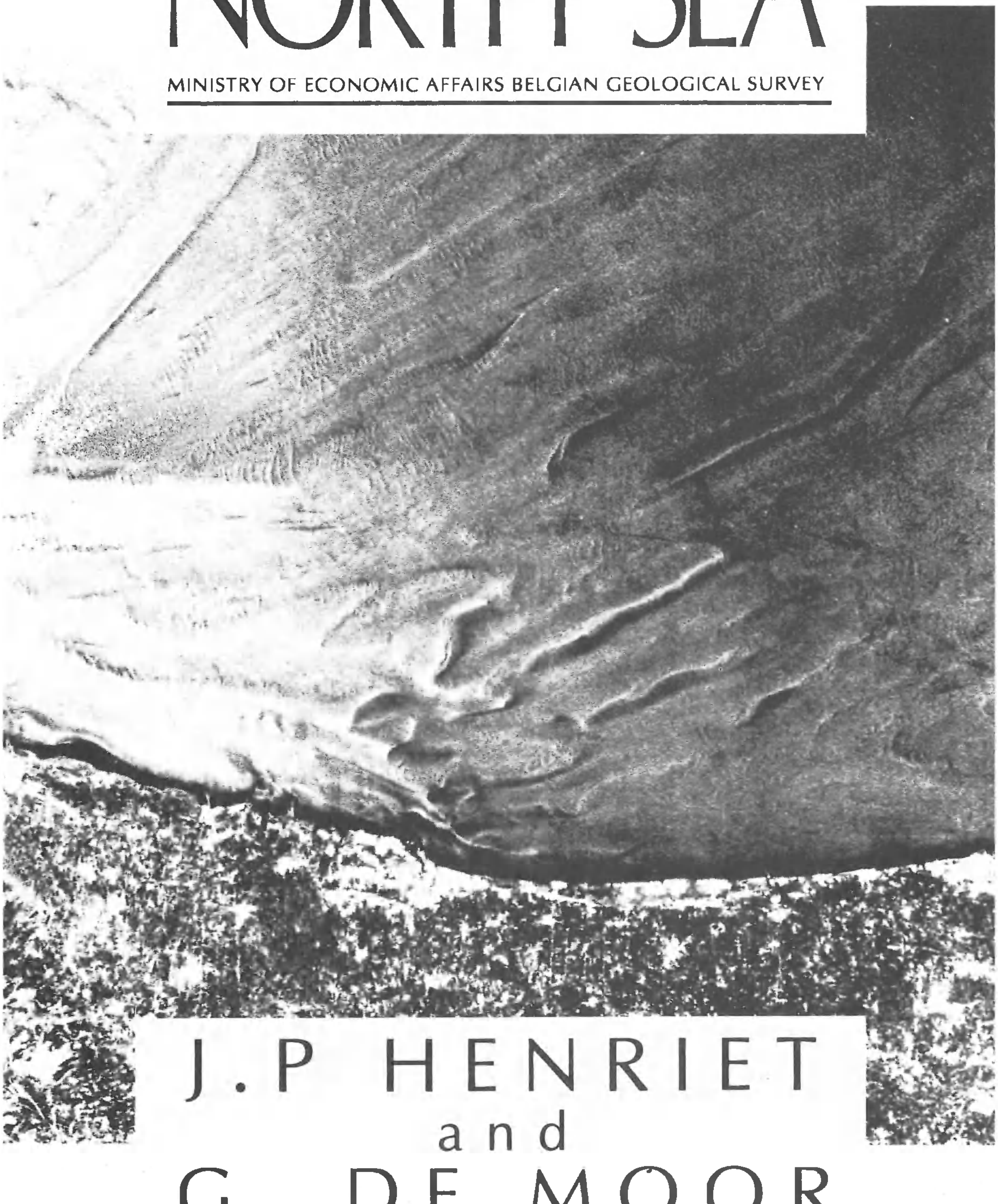

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J.P HENRIET
and
G. DE MOOR
editors

THE QUATERNARY AND TERTIARY GEOLOGY OF THE SOUTHERN BIGHT, NORTH SEA

J.P. HENRIET and G. DE MOOR, editors
M. DE BATIST, assistant editor

Renard Centre of Marine Geology
State University of Ghent, Belgium

A Monography initiated by the International Colloquy on the Quaternary and Tertiary Geology of the Southern Bight, North Sea,
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Contents

Preface

In many aspects, the Flemish Bight or Southern Bight of the North Sea forms a key area for the investigation of the Quaternary and Tertiary geology of the North Sea basin. In this region, gently dipping beds of Tertiary age are directly exposed on the sea bed or but slightly concealed by a thin Quaternary cover. These beds are thus within direct reach of shallow investigation techniques such as bottom sampling and acoustic probing.

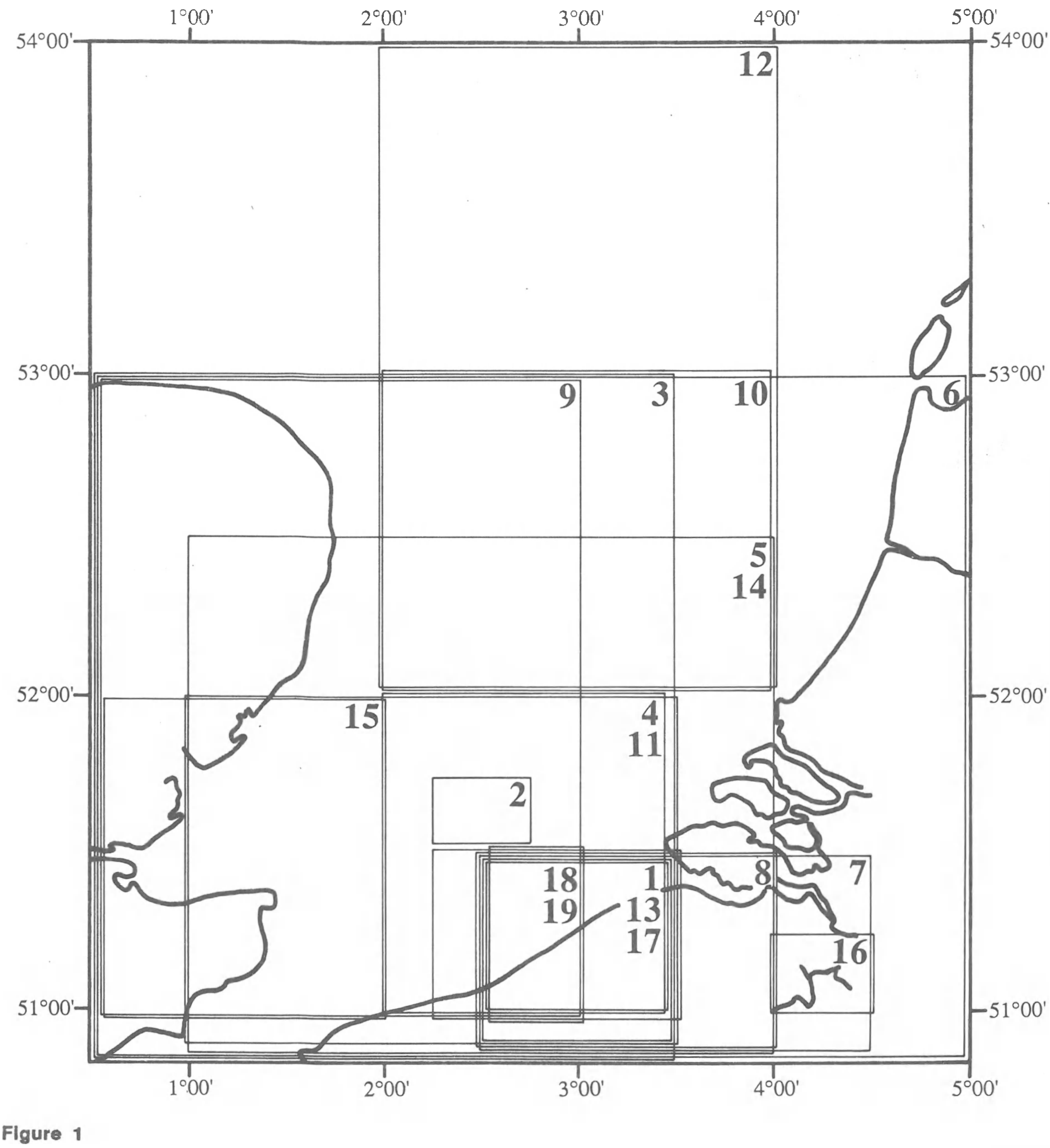
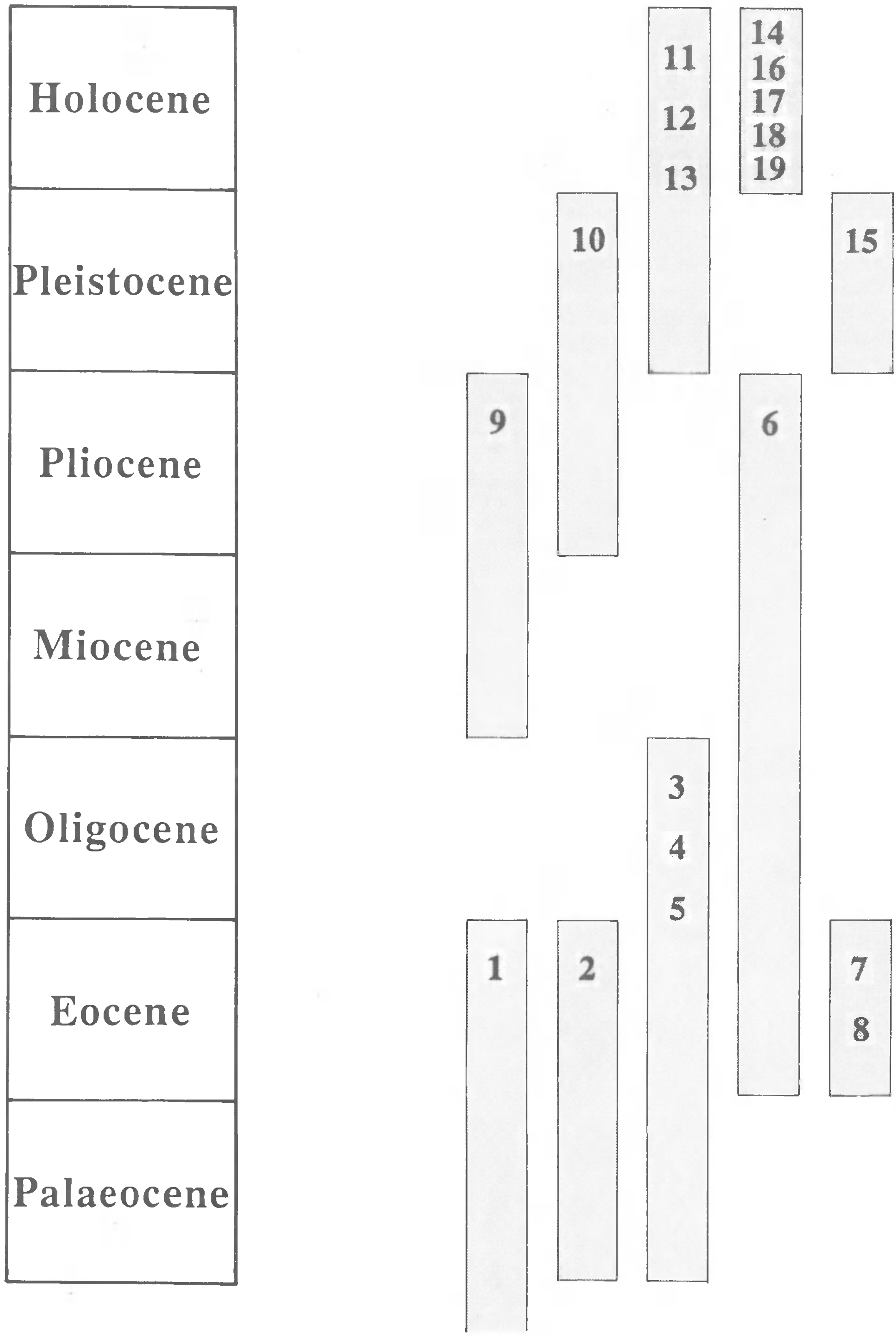


Figure 1



Holocene
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Figure 2

In the past few decades, a number of cruises have shed light on the processes of build-up of Tertiary depositional sequences in response to basin subsidence and global sea level changes, as well as on the occurrence of major or minor breaks in the sedimentary record as a consequence of tectonic events. High-resolution reflection seismic profiling has also unveiled remarkable internal structural features in the main Tertiary clay sequences, which may yield a clue to the early compaction and diagenesis of these sediments.

A systematic mapping of the morphology of the erosion surface at the top of the Tertiary deposits has revealed ancient drainage patterns, dating from glacial sea level lowstand periods. The stratigraphy of the offshore Quaternary deposits has been investigated by seismic profiling and shallow coring. The knowledge thus acquired about the recent geological history of the Flemish Bight no doubt results in a better understanding of the present sediment dynamics in this particular environment.

Results of the investigations of a number of British, Dutch, French and Belgian teams have been presented on a Colloquy, organized in Gent from May 28th to 30th, 1984. A number of papers presented on this meeting have subsequently been collected and edited in the following years, eventually leading to the present monography. These papers cover a wide range of topics, both in space and geological time. The regional coverage of the contributions is illustrated on figure 1, while their geological time span is shown on figure 2.

A synopsis of Palaeogene stratigraphy, supported by high-resolution seismic data and borehole control from both the UK and Belgian sectors and adjacent coastal areas sets the stage for the onset of the Cenozoic record in this region. Attention is also paid to the significant structural deformation patterns, which originated in response to Tertiary tectonic events in Northwest Europe. Follows then a description of the sparse remnants of deposits from the Neogene, which heralds the dramatic change in climatic regime leading to the Quaternary times.

Plio-Pleistocene deposits are best documented in the Dutch and UK sectors. On the Belgian continental sector and in the Thames estuary, the earlier Quaternary history mainly needs to be read from ancient drainage patterns, not yet fully obliterated by marine abrasion. The morphology of major fluvial systems such as those of the Thames and Scheldt rivers also reflects their younger Quaternary evolution. The record of the recentmost sea-level changes from Eemian to Holocene times is well preserved in the Belgian coastal plain. The present-day tide-swept Southern North Sea and Channel form a unique laboratory for the study of the origin and maintenance mechanisms of major sea-bed morphological features like sand banks, as well as for the analysis of associated sediment grades and distribution patterns.

Although a wide range of geological aspects regarding the Southern Bight of the North Sea are dealt with in this volume, there are still a number of outstanding problems. A major question in the Palaeogene record for instance is the exact timing of the opening and closing phases of this important gateway between Boreal and Tethys domains. A related question is to which extent this gateway could possibly have interacted with the evanescent North Atlantic land bridge, formed by the Greenland-Scotland Ridge in Tertiary times, especially in terms of exchange of water masses between the Boreal and the North Atlantic and Tethys marine domains. A particularly intriguing story is also the exact timing and mechanism of the recentmost opening of the Channel, probably in Late Quaternary times. Was it a purely fluvioglacial erosion process, possibly controlled by tectonic accidents, or was there also a significant role played by glacial scouring in earlier glacial times ?

Forthcoming investigations, no doubt also to be carried out by research teams which contributed to the present volume, may shed light on these problems in some near future.

Jean-Pierre HENRIET, Guy DE MOOR and Marc DE BATIST

Renard Centre of Marine Geology
State University of Gent, Belgium

1 A geological profile along the Belgian coast

P. LAGA (1) and N. VANDENBERGHE (2)

1. Belgian Geological Survey, Jennerstraat 13, B-1040 Brussel, Belgium.
2. Department Geology-Geography, Catholic University of Leuven, Redingenstraat 16, B-3000 Leuven, Belgium.

INTRODUCTION

The drilling of three new wells along the Belgian coast by the Geological Survey was certainly not because a good geometrical lithoframe for the area would have been lacking. HALET (1921) and GULINCK (1970) produced well known sections along the coast. The data available to these authors however were provided by samples from older water flush wells and hence precise and detailed lithological descriptions were rare as were the samples suitable for modern biostratigraphical analyses. Besides geotechnical-, exploration- and research activities offshore the Belgian coast would benefit from more modern geological and geophysical reference data along the coast. Three well sites were selected, at Den Haan (22W/276), Knokke (11E/138) and Oostduinkerke (35E/142). The three wells are drilled till the Caledonian basement of the Brabant Massif. The Knokke well has been entirely cored ; both other wells were flushwells. All three wells had a classical well log program (resistivity, spontaneous potential, natural radioactivity and caliper) ; additionally in the Knokke well density and acoustic transit time measurements were done.

THE MAIN RESULTS FROM THE THREE INDIVIDUAL WELLS

a. The Knokke well

A full report about the Knokke well is available (LAGA and VANDENBERGHE, in press) with contributions dealing with lithology, mineralogy, biostratigraphy and well geophysical data.

The Knokke well section starts at the top with 30 m mainly sandy Pleistocene, estuarine tidal deposits belonging to the Flemish Valley river system.

Between 30 and 71.50 m depth the top of the Tertiary is made up by the lower two clay units of the Kallo Formation. The lower one with a glauconitic rich base being the Asse clay ; both clay units are separated by the first sandy layer of the Kallo Formation complex.

A thin fine sand unit between 71.5 and 74 m contains *Nummulites wemmelensis* and is therefore identified as the Wemmel sands. The 74-79 m section is badly cored because of the alternation of fine sands and cemented calcareous sandstones. At 73 m *Callista proxima*, diagnostic for the Brussels sands is present together with numerous rounded *Nummulites laevigatus*. Therefore underlying the Wemmel sands are probably the Brussels sands and not the Lede sands as regional geological knowledge would expect. However the idea of a widespread sheet of Brussels sands almost completely eroded before the deposition of Lede and Wemmel sands was put forward already in 1954 by GULINCK and HACQUAERT while a more recent study in the nearby Zeebrugge harbour area by DEPRET and WILLEMS (1983) confirmed the lack of the Lede sediments and the presence of *Nummulites laevigatus* which according to these authors could belong to sediments comparable either to the Den Hoorn (Upper Panisel) or to the Brussels sands.

The 79 m - 105 m interval of fine clayey glauconitic sands with numerous shells and calcareous sandstones corresponds to the Upper Panisel facies. Between this typical Upper Panisel facies and the clayey Lower

Panisel facies between 124 and 133 m a glauconitic sand with stone layers occurs. The basal Panisel clay (Merelbeke clay) is only two metres thick between 133 and 135 m.

The Ieper Formation extends from 144 m down to 288 m. The overlying interval 135-144 m is a laminated clay with fine sand horizons and even cemented sandstones and is now called the Egem Formation. The Ieper Formation consists of homogeneous greengrey heavy clay with some rare silty spots or thin silt laminae. Several horizons of brecciated Ieper clay exist. Although some of the fragments have some rounding, most of the breccia fragments are still angular and have barely moved. At least in some instances brecciation is associated with small faults, probably indicating gliding. The angular shape of the clay fragments and the fact that the fragments have not been compressed suggests brecciation at an already well compacted stage, at least some tens of metres burial depth or even more.

The breccia are related to observations by HENRIET and MARECHAL (1982) who have interpreted, from sparkerseismic surveys, diapirs and other deformations in the Ieper clay offshore. It is suggested that this common occurrence of Ieper clay deformation in the Belgian North Sea in contrast to almost undeformed Ieper clay on land is due to unloading as the Channel was eroded.

Practically the whole Landen Formation between 288-311 m has to be associated with continental Upper Landen facies. The upper 11 m are quartz sands with peat debris. The 297-308 m interval with silts, fine sands, heavy clay and shells has a typical Sparnacien type appearance. The lower part of the Landen Formation consists of coloured sands.

The Upper Cretaceous chalk sequence is very homogeneous. It is a fine grained white chalk, containing mollusc fragments. Several indurated horizons occur and the chalk as a whole becomes more endurated towards the base. Calcareous nannofossils show the sequence to be mainly Campanian to Lower Maastrichtian at the very top. The base of the Upper Cretaceous is more differentiated. It contains small black phosphatic pebbles, cemented glauconite sandstone and fish remnants. Its age is Santonian.

The top of the Brabant Massif is reached at 432 m and consists of a moderately dipping jointed slate of the Lower Revinium Oisquercq slates (LEGRAND, 1968).

b. The Den Haan and Oostduinkerke wells

The Tertiary in both wells only consists of Ieper clays and the Landen Formation.

The detailed lithological variations within the Ieper clays can be correlated through the natural radioactivity logs. Based mainly on resistivity, the Landen Formation in the two wells can be divided into an upper continental facies with a variable mainly sandy lithology and a lower marine facies with a rather homogeneous fine grained lithology.

The Landen Formation in the Den Haan well occurs between 193 and 223 m and in the Oostduinkerke well between 138.5 and 180 m.

The chalk mass in Oostduinkerke between 180 and 264.45 m is apparently less permeable than the chalk in Den Haan between 223 and 297 m, based on the separation between long and short spaced resistivity readings.

The biostratigraphic work by LOUWYE (in LAGA and VANDENBERGHE, op.cit.) based on *Dinophyceae* attributes the almost entire section to the Lower Campanian except for the 295-299 m part which is attributed to the Santonian. The main chalk mass in the Oostduinkerke well is associated with the Upper and Lower Turonian. The lower two metres (262.4 m - 264.45 m) is a pale conglomerate with Caledonian fragments in a limy and glauconitic cement, and containing fossil fragments. It is thought to be equivalent with the Cenomanian base conglomerate (Serrasin de Bellignies of MARLIERE 1954, p. 422-425).

The top of the Brabant Massif in the Den Haan well at 297 m is a Revinian quartzic slate, the top of which is altered to a reddish clay. Total depth at Den Haan is 321 m. The top of the Brabant Massif in the Oostduinkerke well consists of slightly altered quartz slates, with red clay fillings and a tectonic breccia.

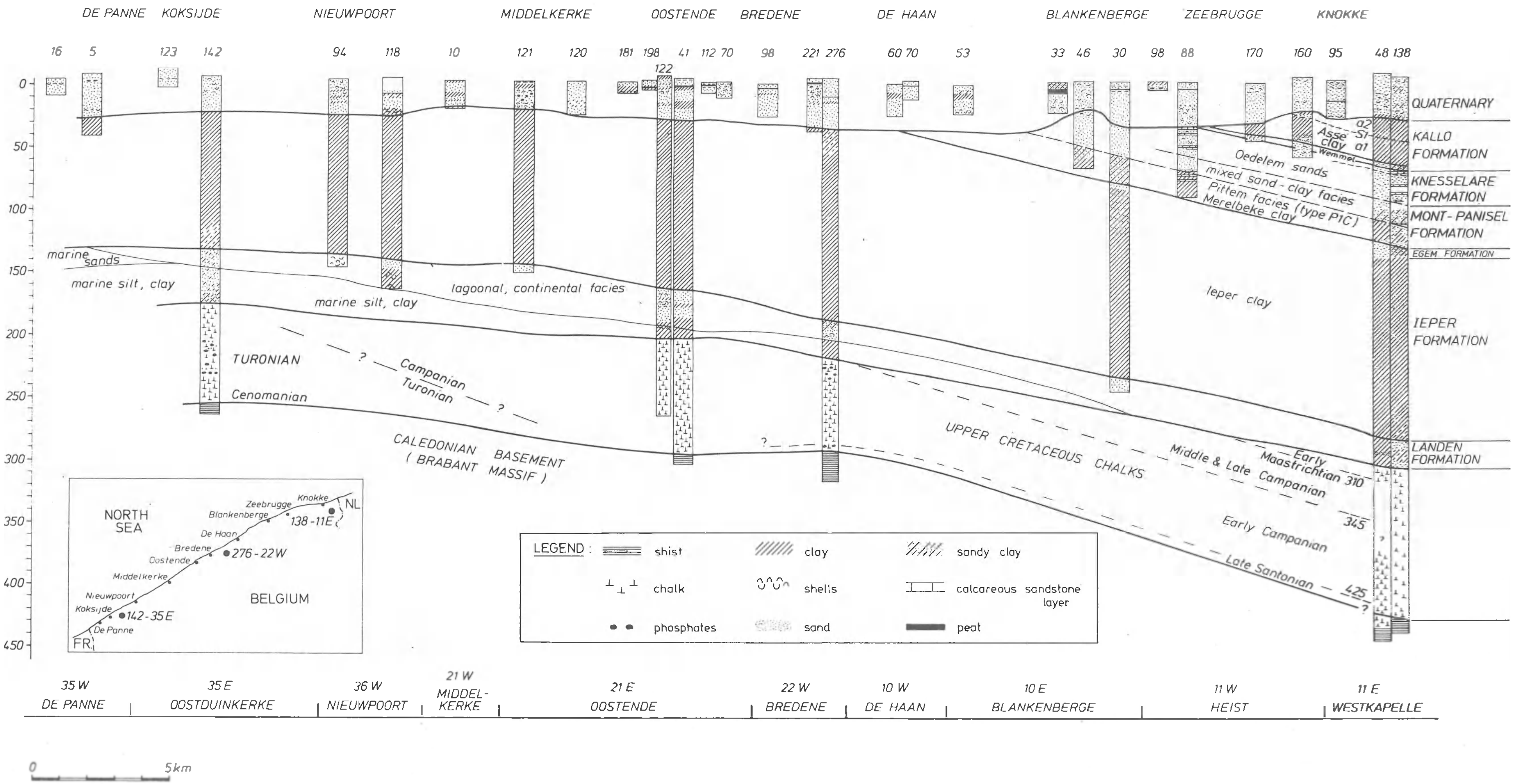

Figure 1 Cross section along the Belgian coast.

According to LEGRAND (1968) these rocks should belong to the Salmian. Total depth at Oostduinkerke is 270.3 m.

THE CORRELATION OF THE THREE WELLS AND A CROSS SECTION ALONG THE COAST

The profile represented on figure 1 incorporates the three wells discussed above together with other well data available in the files of the Geological Survey.

The boundaries of the Tertiary formations and facies and also the top of the Brabant Massif can be traced without too much difficulty.

Within the Upper Cretaceous, the Turonian of the west coast (Oostduinkerke) has to wedge out under the Santonian-Campanian probably between Den Haan and Oostende (see also LEGRAND, 1968). The northern limit of the Turonian sea was formed by a relative high in the Brabant Massif (MARLIERE, 1954, p. 420). After the Turonian the area north of Den Haan has started to subside relative to the southern area where Turonian had been deposited before.

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2 A northeast trending structural deformation zone near North Hinder

J.P. HENRIET (1), M. DE BATIST (1)(2), B. D'OLIER (3) and J.P. AUFFRET (4)

1. Renard Centre of Marine Geology, Geological Institute, State University of Gent, Krijgslaan 281/S8, B-9000 Gent, Belgium.
2. Research Assistant, National Fund for Scientific Research, Belgium.
3. Department of Geology, City of London Polytechnic, Walburgh House, Bigland Street, London E1 2NG, United Kingdom.
4. Laboratoire de Géologie Marine, Université de Caen, Département de Géologie, Esplanade de la Paix 14, F-14032 Caen CEDEX, France.

ABSTRACT

A northeast trending sequence of structural deformations east of North Hinder on the Belgian continental shelf and adjacent areas seems to be the surface expression of deeper faults, cutting across the whole width of the London-Brabant Massif in the axial zone of the eastern Channel. These fractures have probably been reactivated in a wrench-fault style in Tertiary times.

INTRODUCTION

Detailed and systematic high-resolution seismic investigations carried out in recent years in the Southern Bight of the North Sea have shed some new light on the tectonic setting of the offshore extension of the London-Brabant Massif. A classical view is that this Palaeozoic basement high behaved throughout post-Hercynian times as a rigid unit, unaffected by the major Mesozoic deformation phases which left their scars along its periphery.

Some doubts may however be cast on this model, particularly in view of the recent seismic observation of a sequence of troughs and faults in the Mesozoic and Early Cenozoic cover of the Brabant Massif, trending in a northeastern direction a few kilometres east of the North Hinder sand bank. The fact that this sequence apparently has eluded the scarce former seismic observations in this region may probably be attributed to its coincidence with the Channel traffic separation lanes, a traditionally hostile environment for seismic activities.

THE NORTH HINDER STRUCTURE

The seismic-stratigraphic subcrop map of the Palaeogene substratum (figure 1) reveals a N50°E-trending alignment of three major synclinal depressions, roughly circular in shape, affecting the Eocene sequences over a total distance of at least 25 to 30 km.

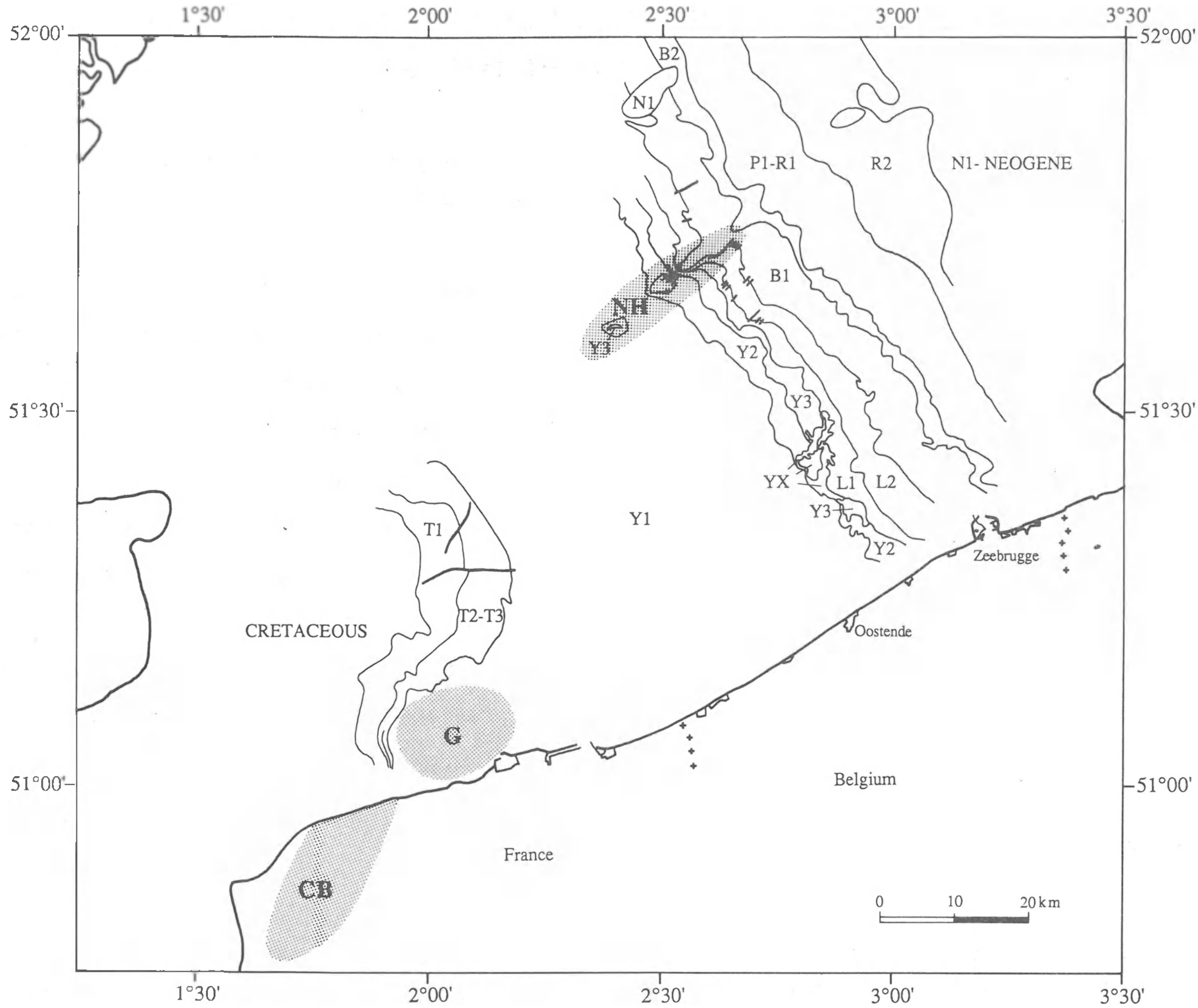

Figure 1. Seismic-stratigraphic solid map of the Southern Bight Palaeogene deposits (HENRIET *et al.*, this volume), with localization of structural anomalies.
NH = North Hinder structures. G = Gravelines structure. CB = Calais Basin.

The southwestern depression stands out as an isolated feature, with an observed amplitude of some 150 m at its deepest point. It is bordered to the east by a fault, throwing about 50 to 60 m.

The central and northwestern depressions seem to be more or less structurally connected, forming a complex, elongated and asymmetric synclinal trough. Its amplitude decreases towards the northeast to a few tens of meters.

A pattern of smaller-scale deformations under the form of dense block-faulting is superimposed on the main folds, a phenomenon which is particularly prominent in the central depression. This might find its origin in a reactivation and amplification of pre-existing clay-tectonic deformations (HENRIET *et al.*, 1988).

On some cross-sections the synclinal depressions observed in the London or Ieper Clay seem to develop vertically out of a system of major faulted an tilted blocks in the underlying Palaeocene deposits (figure 2). Locally, such faults can be traced through the Cretaceous chalk cover down to the top of the Palaeozoic basement. The exact configuration and trend of these basement faults could not be ascertained yet.

A remarkable observation in the surface depressions is a weakly defined east-west elongation, making an angle of 30° to 45° with the dominant trend of the alignment, thus creating an "en échelon"-pattern.

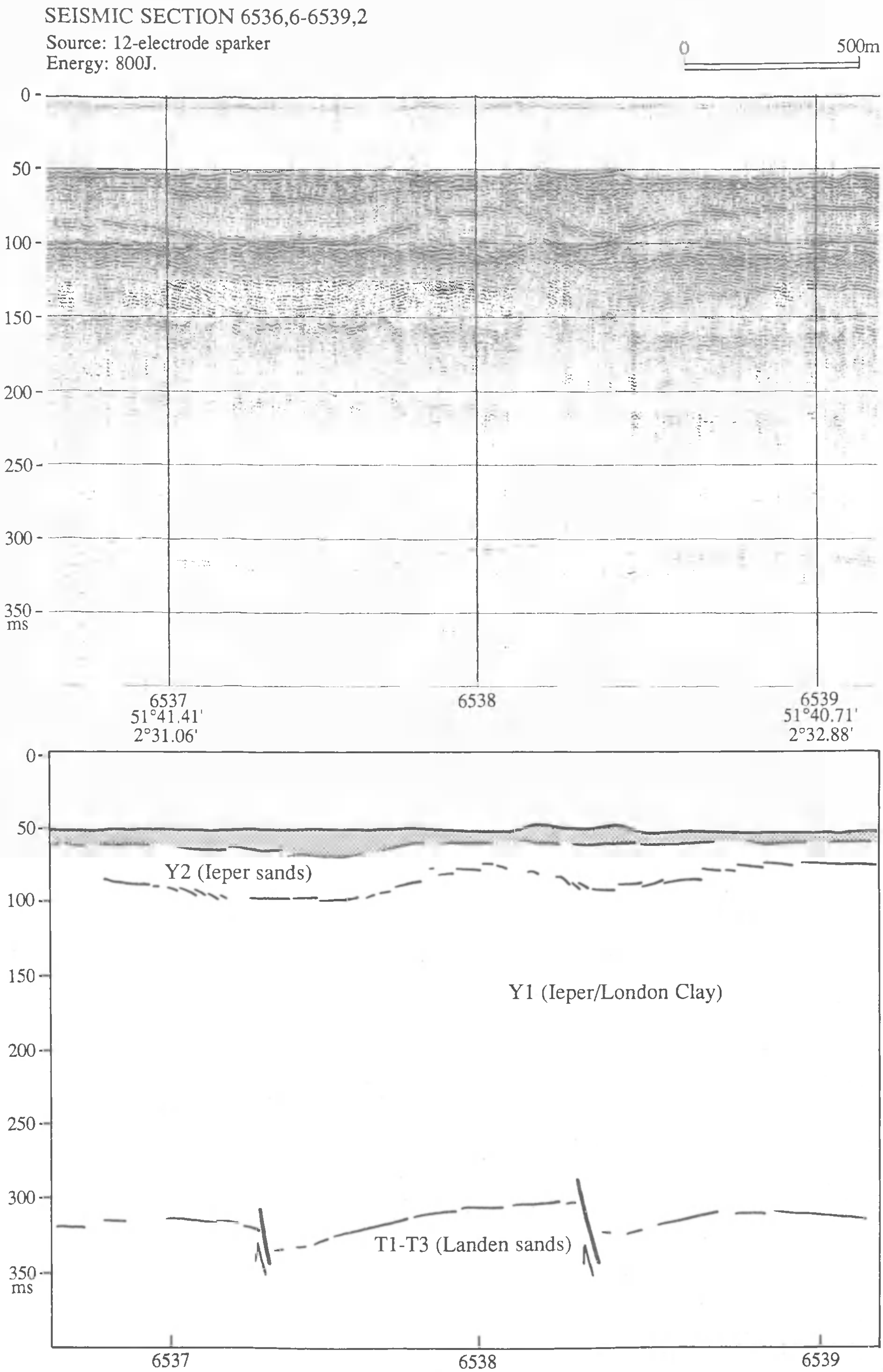

Figure 2 Analog-recorded watergun section and interpreted line-drawing showing the relation between Cretaceous-Palaeocene block-faulting and overlying synclinal deformations in Eocene deposits.
Vertical scales in ms are two-way time.

THE GRAVELINES STRUCTURE

Further south, off Calais, another major deformation structure can be observed : a northeast trending trough affecting Mesozoic and Cenozoic beds. It is best developed off Gravelines, where it shows as a deep, nearly circular depression with an amplitude of about 120 m. The southward extension of this feature in the onshore region is traced down to the Palaeozoic basement, as shown by the isobath map (figure 3) of the top of the Brabant Massif (LEGRAND, 1968).

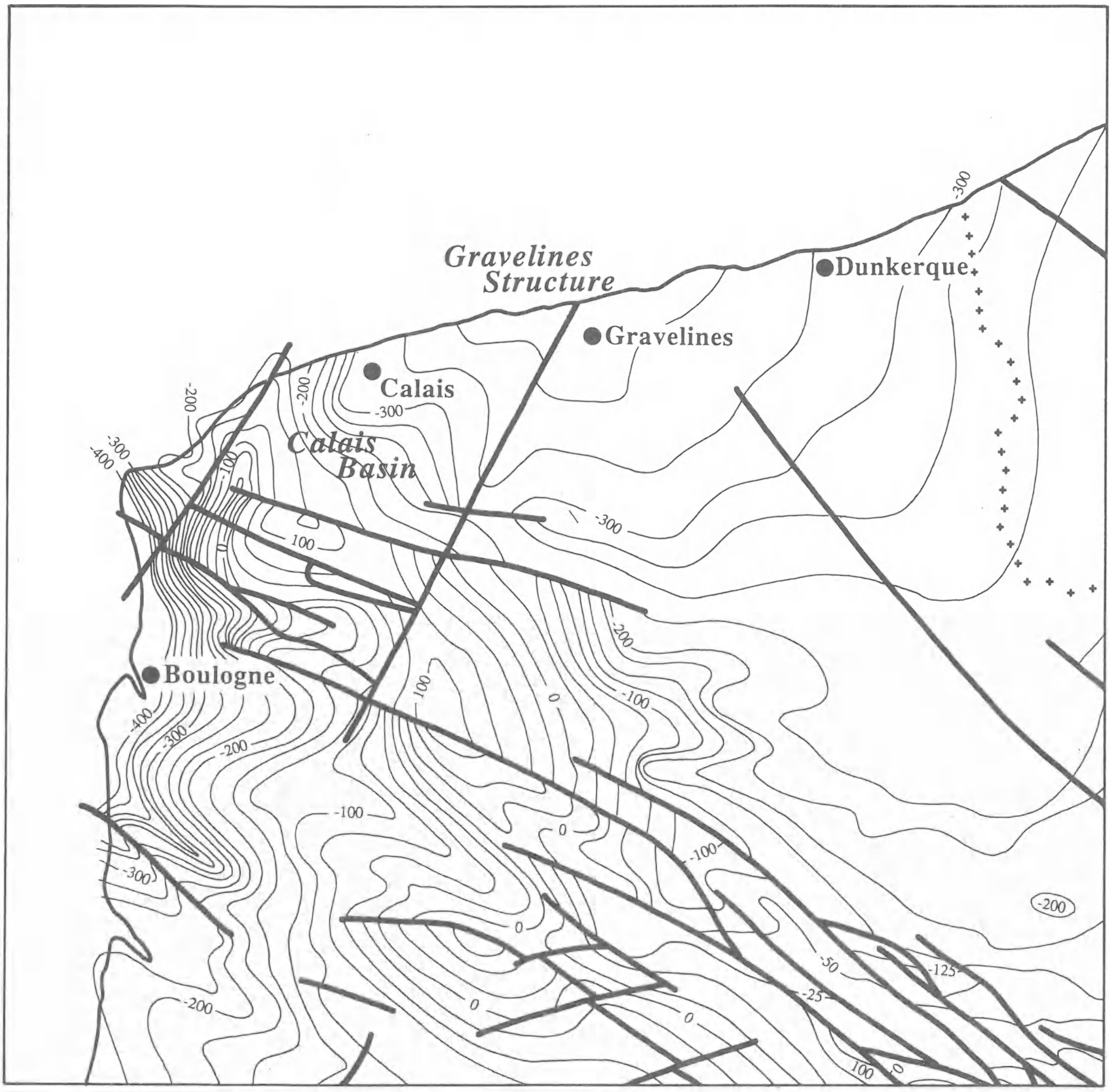

Figure 3 Isohypse map of the top of the Palaeozoic basement in the Calais region (LEGRAND, 1968).

The Gravelines depression is indeed located on top of the supposed offshore extension of the Hercynian Calais Basin, a riftlike structure shown on the geological map (figure 4) of the top of the Palaeozoic basement, compiled by a group of French oil companies (C.F.P.(M) *et al.*, 1965).

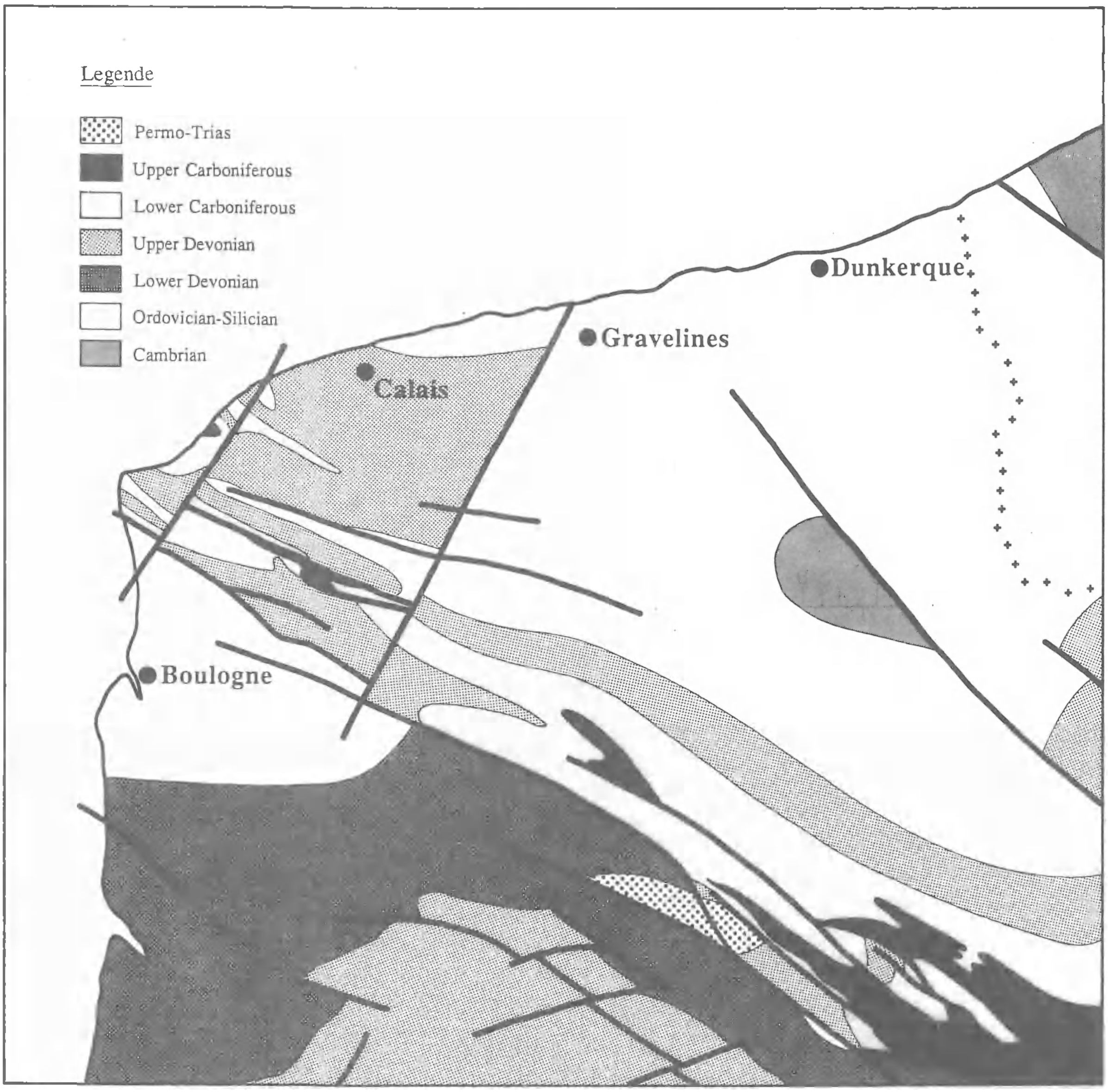

Figure 4 Geological map of the top of the Palaeozoic basement in the Calais region (LEGRAND, 1968 ; C.F.M.(P) *et al.*, 1965).

The Calais Basin has a pronounced asymmetric character, with a major downward slip displacement along its eastern border, the Ardres-Gravelines fault. This northeast trending fault separates the comprehensive Devonian to Lower Carboniferous stratigraphic sequence of the rift basin from the Silurian rocks of the London-Brabant Massif.

FRAMING THE NORTH HINDER STRUCTURE IN SPACE AND TIME

Considering the apparent alignment - possibly with some offset - of a number of structural anomalies, ranging from the Palaeozoic Calais Basin with its northeast trending border faults, the depression off Gravelines and the cluster of deformations near North Hinder, it is tempting to speculate that all these features belong to one major, northeast trending deformation zone.

The deformations of the cover sediments of the London-Brabant Massif should be younger than Eocene. A weak unconformity at the base of the Bartonian deposits observed in a syncline in the eastern part of the Belgian continental shelf suggests a first pulse of tectonic activity. The main tectonic phase however cannot be more accurately described than ranging between Oligocene and Present in age, due to the fact that all deformations are observed within an area of outcropping Eocene deposits, discontinuously covered by undeformed Quaternary sediments. A fair guess for the main deformation period might be the Miocene phase of peak Alpine compressive stresses, also experienced in other places of the Northwest European continental crust.

At this level it might be useful to mention that the hypothesis of a fracture zone cutting through the London-Brabant Massif has already been advanced by COLBEAUX *et al.* (1980), who assumed a possible fault control of the Pleistocene opening of Dover Strait. Also the apparent offset of geological units on both sides of Dover Strait has been cited by various authors as an argument for axial wrench-faulting (WALLACE, 1968 ; DESTOMBES and SHEPHARD-THORN, 1972 ; SHEPHARD-THORN *et al.*, 1972).

The observation or induction of faults cutting across the London-Brabant Massif and displaying activities possibly ranging from Late Palaeozoic times to Present should not be surprising. It would indeed be hard to explain how for instance the Eocene cover deposits near North Hinder could have been deformed under Late Cenozoic tectonic stresses, known as relatively moderate in these regions, if they had not been located on top of older scars in the otherwise rigid Palaeozoic basement. It should be noted here that most of the Tertiary deformations in the Channel and southern North Sea are considered as reactivations of zones of crustal weakness (ZIEGLER, 1981).

It is thus postulated that the North Hinder deformation zone forms the surface expression of deep-seated faults, possibly of Late Palaeozoic age, which have been reactivated at least in Late Cenozoic times.

The presence of deep, inherited faults cutting across the London-Brabant Massif are not only corroborated by the rift structure of the Calais Basin, but possibly also by the analogy with other transversal basement ridges in the North Sea transected by - often more developed - rift structures.

The "en échelon"-pattern of the slightly elongated depressions suggests that the Tertiary stress fields probably reactivated the deep faults underneath the North Hinder anomalies in a dextral wrench-fault mode. A more detailed structural analysis will however be required for assessing the behaviour of these faults under stress fields, which may have considerably changed both in amplitude and orientation throughout Cenozoic times (BERGERAT and GEYSSANT, 1980).

CONCLUSIONS

Reflection seismic investigations carried out in the northern part of the Belgian continental shelf have revealed a northeast trending sequence of structural deformations, affecting the Mesozoic and Cenozoic overburden of the London-Brabant Massif.

Although further detailed reflection seismic investigations and structural analyses will be required for defining the exact tectonic style and regional context of these deformations, there is little doubt that these features are the superficial expression of deeper scars, cutting across the London-Brabant Massif in the axial zone of the eastern Channel.

ACKNOWLEDGEMENTS

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3 The Palaeogene of the UK southern North Sea, 50°N-53°N : a survey of stratigraphical data

R.W.O'B. KNOX

British Geological Survey, Nicker Hill, Keyworth, Nottingham NG12 5GG, United Kingdom.

ABSTRACT

The Palaeogene sediments of the Flemish Bight are largely concealed beneath a cover of Neogene and Quaternary sediments, but their overall character can be determined by reference to seismic records and to sequences from adjacent onshore and offshore areas. The Cretaceous Chalk is overlain by a more or less uninterrupted mudstone sequence of Late Palaeocene to Late Eocene age. Sands are found in Woolwich Beds equivalents (late Palaeocene / early Eocene), and are thought to occur in Harwich Member equivalents (early Eocene) in the south of the area, where the tuffaceous mudstone facies passes southwards into near-shore sands. Sands may similarly be expected to appear in the upper part of the Lower Eocene sequence as it is traced southwards (Bagshot Sands equivalents).

The Palaeocene and Eocene beds underwent uplift, folding and erosion before renewed sedimentation in the early Oligocene resulted in the deposition of thin and condensed glauconitic clays over much of the area. Subsequent erosion in Middle Oligocene to early Pleistocene times has removed the Oligocene sediments from all but three small structural basins.

INTRODUCTION

This paper is concerned with the Palaeogene of the Flemish Bight (or 'Southern Bight', as used by HOUBOLT, 1968). For the UK sector at least, there is very little direct geological information concerning the central and southern parts of the area, but there is a considerable amount of information available in the form of borehole and outcrop data from the northern and western margins. By using the information from these marginal areas, it is possible to establish both east-west and north-south facies and thickness trends; these trends can then be projected to the south and east.

Much of the UK sector has been mapped on a scale of 1:250,000 by the BRITISH GEOLOGICAL SURVEY (1985 ; in press). Using data from both published and unpublished maps, generalized Palaeogene outcrops are shown in figure 1. In the preparation of the BGS maps, the Palaeocene/Eocene boundary was drawn at the base of the *Wetzeliella astra* dinoflagellate cyst zone, following COSTA *et al.* (1978). On this basis, the London Clay Harwich Member (KING, 1981), with its abundant volcanic ash layers, forms the uppermost part of the Palaeocene sequence. However, a recent correlation based on widespread early Palaeogene ashes (KNOX, 1984) indicates that the base of the NP10 nannoplankton zone (widely taken as marking the base of the Eocene) occurs below the Harwich Member and hence within the *Apectodinium hyperacanthum* zone. This revised boundary is not, however, readily mappable, whereas the top of the Harwich Member is marked by a prominent seismic reflector. Accordingly the BGS map boundary has been retained in this paper (figure 1 and elsewhere) but the sediments above and below are referred to as Eocene and Palaeocene / early Eocene respectively.

The stratigraphical information presented in this paper has been obtained from five main sources (see figure 1 for locations) :

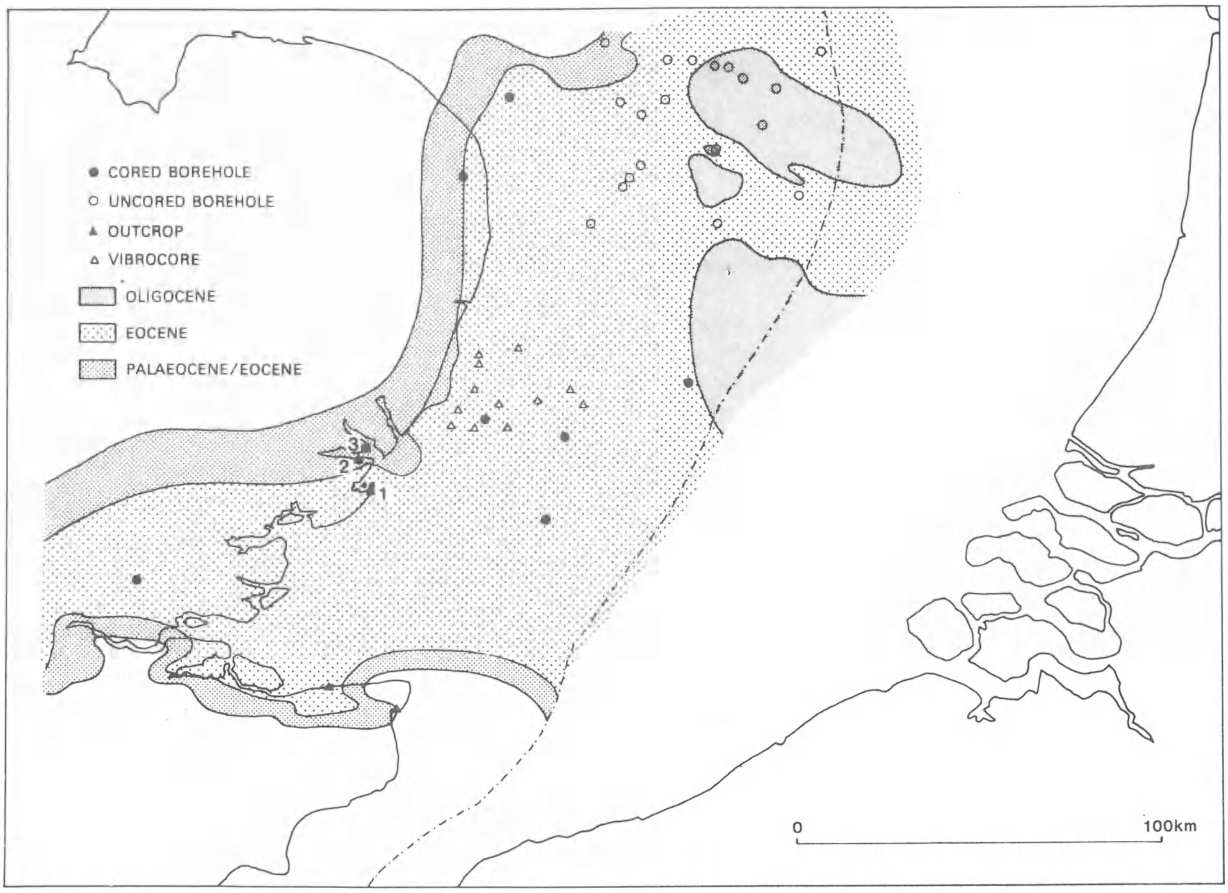

Figure 1 Generalized Palaeogene outcrop map. The top of the 'Palaeocene / Eocene' unit corresponds to the top of the Harwich Member.
Numbered localities are : 1 = Walton-on-the-Naze. 2 = Wrabness. 3 = Shotley Gate. Other localities mentioned in the text are identified in figures 2 to 8.

1. Deep offshore gas exploration boreholes. No cores have been taken in the Tertiary beds, but a relatively detailed sequence can be established from the examination of cuttings samples and downhole geophysical logs.

2. Shallow offshore boreholes put down for site investigation or as part of the BGS mapping programme.

3. Shallow onshore boreholes put down for the BGS mapping programme.

4. Offshore vibrocores (BGS).

5. Onshore outcrops (solid squares in figure 1).

As the outcrops show (figure 1), the Palaeogene strata are disposed in a shallow complex syncline, whose axis runs southwards through the Flemish Bight, and then turns westwards into the London Basin. There is no single cored section through the complete stratigraphical sequence, but it is possible to construct a reasonably accurate composite sequence for the northeastern part of the area, as shown in figure 2. This sequence consists almost entirely of mudstone, but it is possible to distinguish the Palaeocene / Eocene, Eocene and Oligocene divisions on the basis of mudstone lithology and geophysical log character.

The sequence begins with Late Palaeocene sediments, which rest directly on Maastrichtian chalk. Early Palaeocene sediments are absent, but the local occurrence of reworked Danian fossils indicates that Danian limestones formerly extended into the area. Palaeocene and early Eocene sedimentation took place in response to a series of regressions and transgressions, leading to a rather complex stratigraphy, as described in the next section. By contrast, later Eocene sedimentation was more or less continuous, but seismic records suggest that a slight angular unconformity exists between the Eocene and the Oligocene.

The main stratigraphical units are examined in more detail below. References to dinoflagellate cyst zones are based on identifications by R. HARLAND (BGS).

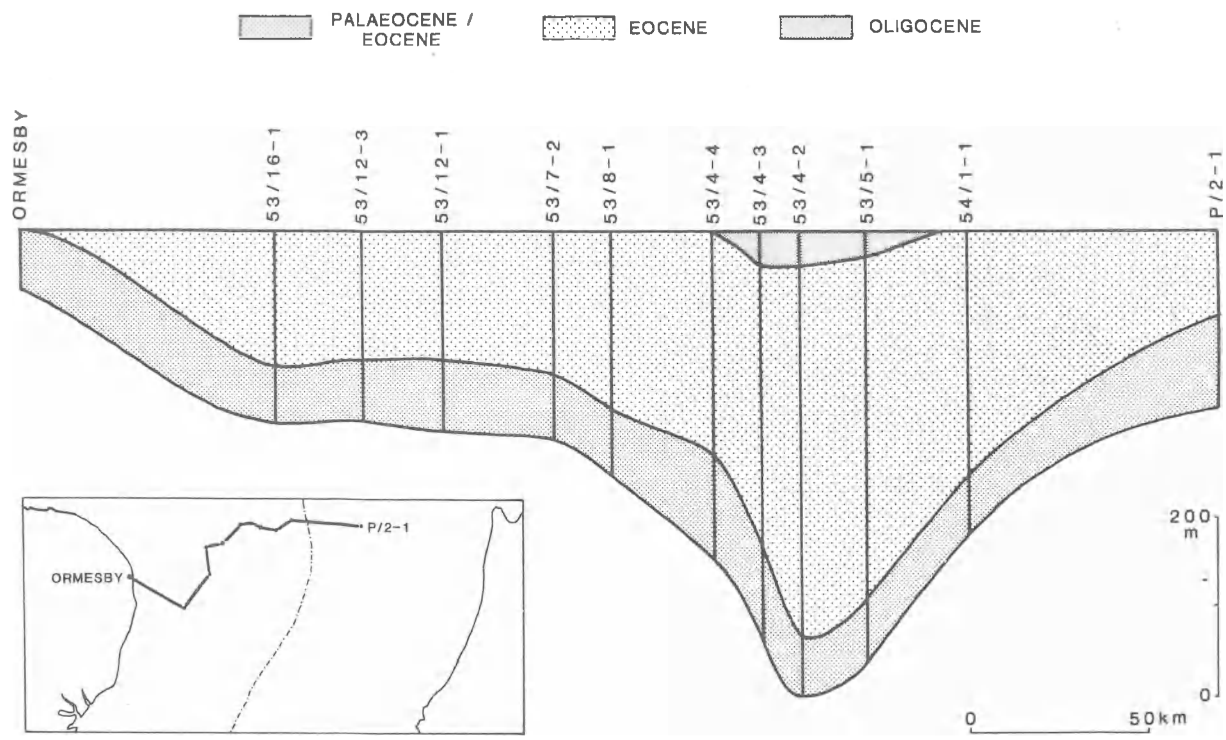

Figure 2 Palaeogene sequence across the northern margin of the Flemish Bight area. Divisions as in figure 1.

PALAEOCENE/EOCENE

Along the northern margin of the area, the Palaeocene/Eocene sequence is dominated by mudstone. Lithological details are best provided by the cored Ormesby Borehole (COX *et al.*, 1985) (figure 3). Here the sequence consists of three units :

1. A lower unit (late Palaeocene) of grey-green mudstones with an abundant microfauna. The beds are equivalent to the Lista Formation (DEEGAN and SCULL, 1977) of the central North Sea area, or Units C2 and C3 of KNOX *et al.* (1981). The formation can be divided into a lower unit of silty, glauconitic and tuffaceous mudstone, with at least one well-defined graded volcanic ash layer, and an upper unit of relatively pure mudstone in which agglutinated arenaceous foraminifera are particularly abundant. This two-fold division is also recognisable offshore (figure 2).

2. A middle unit (early Eocene) of grey sandy carbonaceous mudstone ; this is equivalent to the Sele Formation of the central North Sea area (cf. Unit 1 of LOTT *et al.*, 1983). The grey mudstones of the Sele Formation rest with a sharp erosional contact on green mudstones of the underlying Lista Formation. They include sporadic thin volcanic ash layers. The microfauna is both restricted and sparse.

3. A top unit (early Eocene) of grey sandy carbonaceous mudstone with abundant volcanic ash layers ; this is equivalent to the lower part of the Balder Formation of the Central North Sea area (cf. Unit 2 of LOTT *et al.*, 1983) and to the Harwich Member of the London Clay. The unit is sandy at the base, and the contact with the underlying Sele Formation is probably erosional. Eighty-three ash layers have been identified in about 25 metres of sediment.

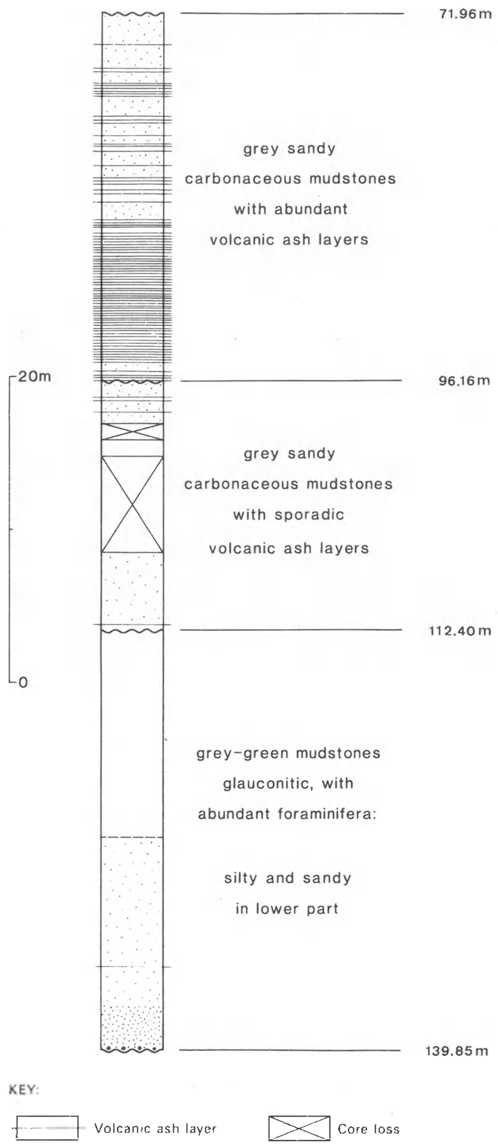

Figure 3 'Palaeocene / Eocene' sequence in the Ormesby Borehole, Norfolk (National Grid Reference TG 5148 1424) (modified from COX *et al.*, 1985).

When traced to the east (figure 4), the following changes are observed :

1. The Lista Formation shows an overall increase in thickness, with little change in facies.

2. The Sele Formation develops a local sand unit in its lower part, but otherwise shows little change in facies. Compared with the Lista Formation, the Sele Formation displays relatively consistent thickness, indicating deposition under a different subsidence regime, or perhaps gentle folding and erosion of originally more uniform Lista sediments.

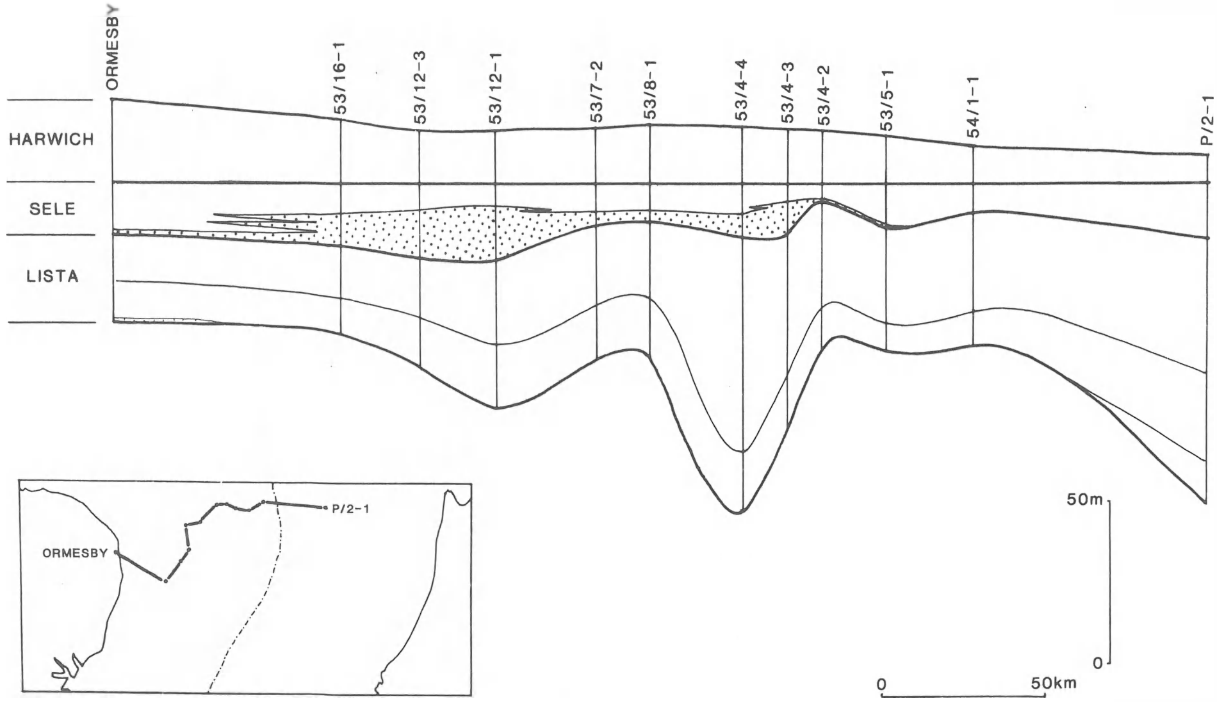

Figure 4 'Palaeocene / Eocene' sequence across the northern margin of the area.

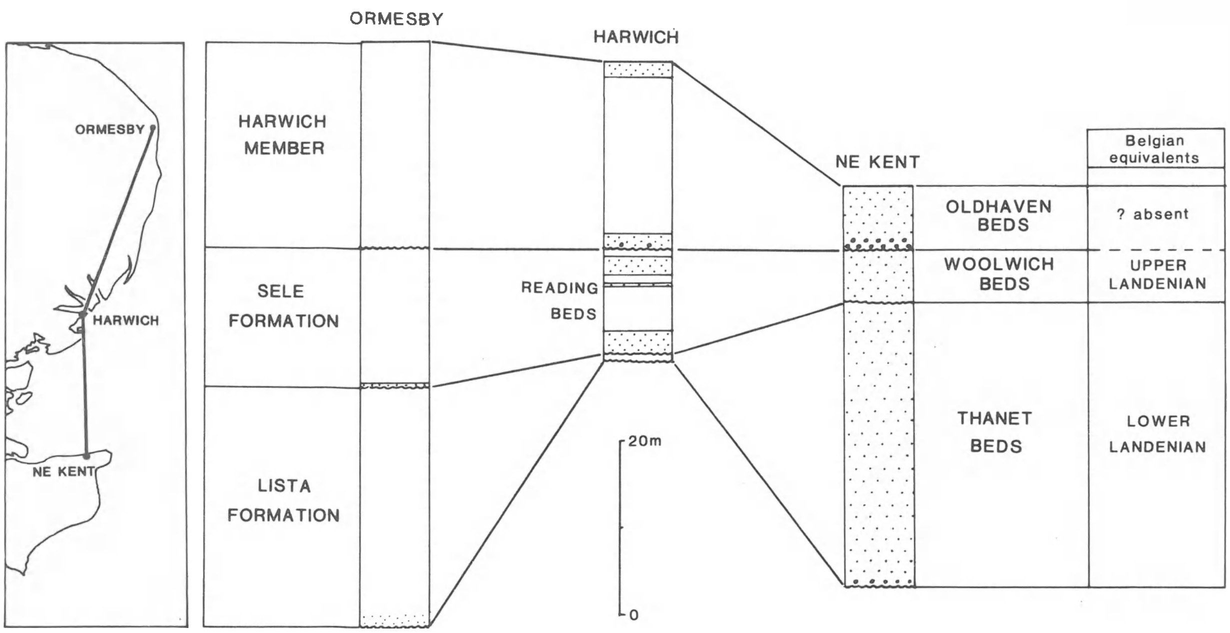

Figure 5 Palaeocene to early Eocene sequence of the onshore western margin. The Harwich sequence is based on sections from Wrabness (N.G.R. TM 172 323), Walton-on-the-Naze (N.G.R. TM 267 243), and the Shotley Gate Borehole (N.G.R. TM 2439 3460) (see figure 1).

3. The Harwich Member becomes a little thinner towards the east, due in part at least to a decrease in sand content.

When the sequence is traced southwards form Ormesby, changes in thickness and facies are much more pronounced (figure 5), and may be summarized as follows :

1. The Lista Formation thins towards the Harwich area, which seems to have been the site of a local structural high. Further to the south, the beds thicken again, but there is a rapid passage into the glauconitic fine sand facies of the Thanet Beds.

2. The grey mudstones of the Sele Formation pass rapidly southwards into yellow-brown and red-brown lacustrine and pedogenic mudstones of the Reading Beds. Further south the facies changes again, and the unit is represented by the marine glauconitic sands of the Woolwich Beds (ELLISON, 1983).

3. The mudstone and volcanic ash layers of the Harwich Member persist into the Harwich area, but the sequence is sandier and appears to be stratigraphically less complete - only 40 or so ash layers have been identified (KNOX and ELLISON, 1979) compared with 83 in the Ormesby Borehole. To the south of the Harwich area the mudstones pass laterally into muddy glauconitic sandstones and finally into the clean sublittoral sandstones of the Oldhaven Beds (KNOX, 1983).

Subsurface information on the Palaeocene/Eocene of the central offshore area is limited to two site-investigation boreholes (figure 6). These show sequences very similar to that of the Harwich area. An interesting feature of the Harwich Member in the Shipwash Borehole is the presence of an upward-coarsening pebbly unit at the base. This appears to represent a drowned beach deposit. Local conglomerates are known at a similar stratigraphical level in the Harwich area (Suffolk Pebble Red).

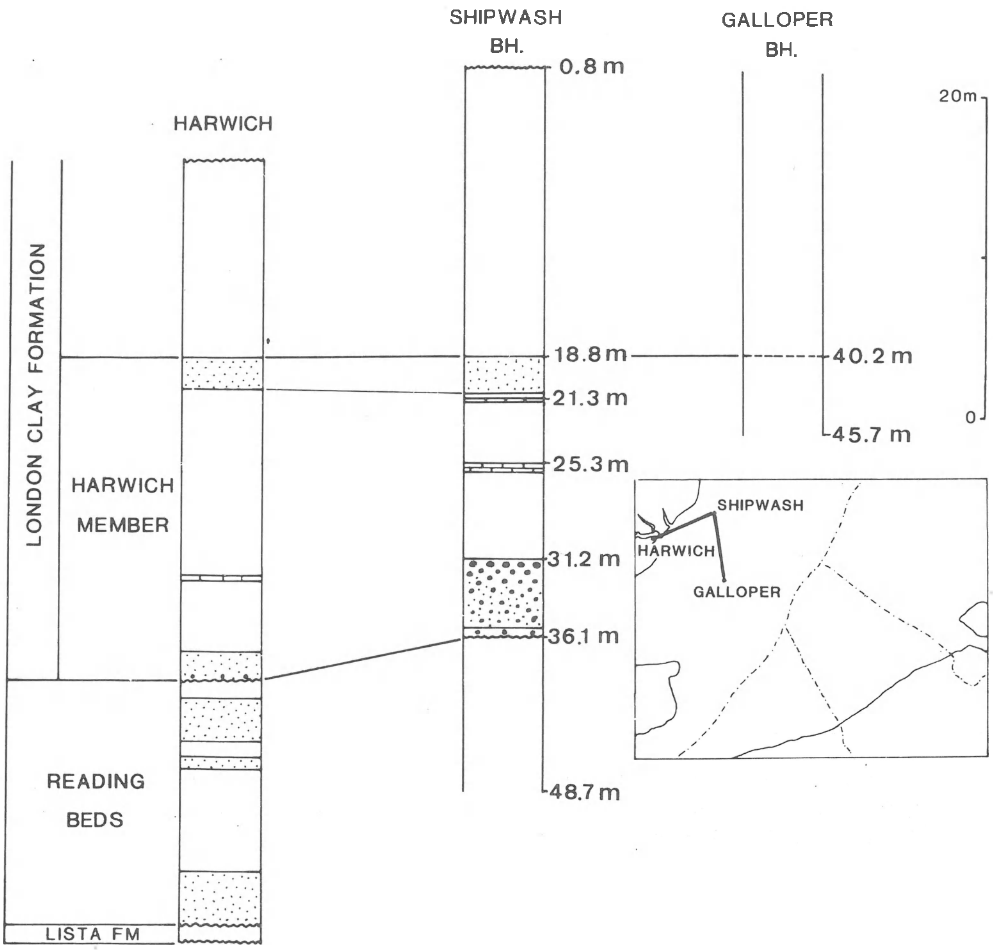

Figure 6 Palaeocene to early Eocene sequence of the central area. The Harwich sequence is composite (see capture for figure 5). Borehole depths measured from sea bed. Borehole locations : Shipwash = 52°02'20"N, 01°41'00"E. Galloper = 51°44'25"N, 01°57'35"E.

EOCENE AND OLIGOCENE

Along the northern margin of the area (figure 7), the Eocene is divisible into three units :

1. A lower unit (early Eocene) of grey carbonaceous mudstones (the upper part of the Balder Formation), with abundant *Coscinodiscus* and a sparse and restricted microfauna.

2. A middle (early Eocene) unit of well-bedded mudstones, with a relatively rich microfauna.

3. An upper unit (middle to late Eocene) of poorly-bedded glauconitic mudstones and thin sandstones, again with a relatively rich microfauna.

The Oligocene consists of brown-grey glauconitic mudstones, sandy at the base.

The nearest UK onshore sequences are in South Essex and Northeast Kent. Here the upper part of the sequence has been eroded, so that the youngest beds remaining are of early Eocene age (figure 7). Within the London Clay, the top of the Balder Formation is no longer definable on the basis of mudstone character, but the upward change from a restricted agglutinated microfauna to a richer microfauna (including planktonics) is still recognizable (at the so-called 'planktonic datum', or close to the base of Division B of KING, 1981).

The upward change from well bedded to blocky mudstones as seen in the northern boreholes appears to be represented in the south by the change from mudstone (London Clay) to sandstone (Bagshot Beds).

In the central offshore area, the London Clay has been encountered in three boreholes. Two of these (Outer Gabbard and Galloper) together provide an almost complete sequence through the formation (figure 8). An interesting feature of this sequence is that the beds are considerably thinner than their onshore equivalents,

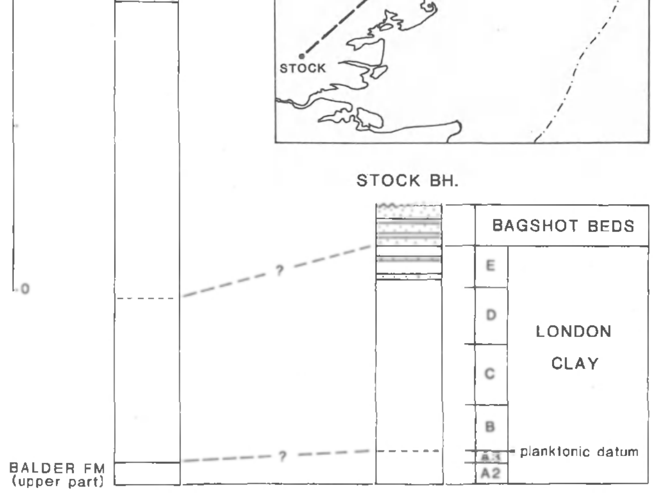

Figure 7 'Eocene' and Oligocene sequences of the northern and western margins (for 'Palaeocene / Eocene' units see figures 3-5). Borehole depths measured from sea bed or ground level. Borehole locations : 53/4-2 = 52°52'50"N, 02°47'16"E. Stock = N.G.R. TL 705 005.

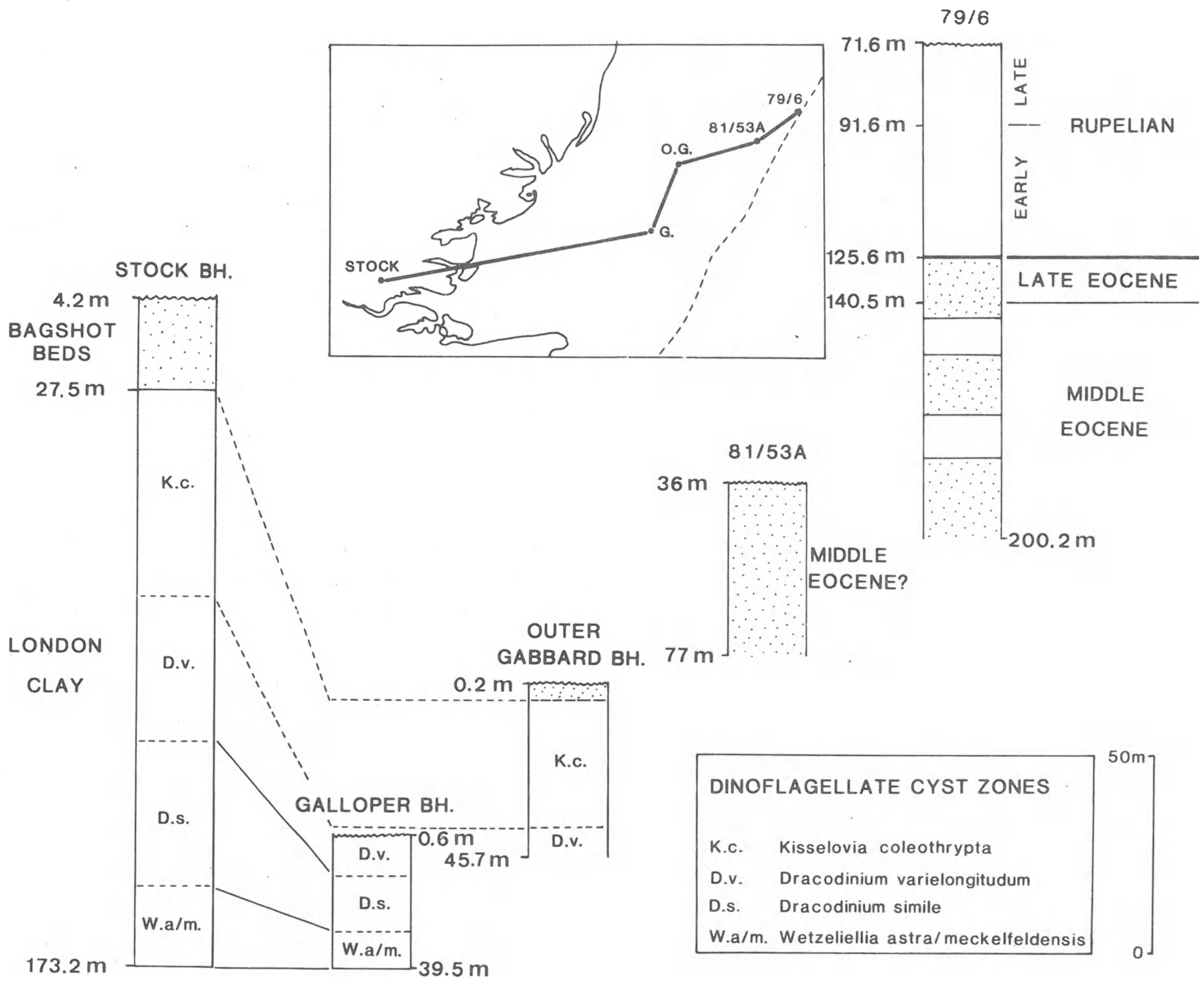

Figure 8 'Eocene' and Oligocene sequence of the central area (for 'Palaeocene / Eocene' units see figure 6). Dinoflagellate cyst zones are shown for the London Clay sequence only (the *K. coleothrypta* Zone is known to extend upwards into the sands). Borehole depths measured from sea bed or ground level. Borehole locations : Stock = N.G.R. TL 705 005. Galloper = 51°44'25"N, 01°57'35"E. Outer Gabbard = 51°59'40"N, 02°04'30"E. 81/53A = 52°05'11"N, 02°32'56"E.

as is shown not only by thinning of the entire mudstone unit, but also by thinning of individual dinoflagellate cyst zones.

Glauconitic sands of supposed middle Eocene age have been encountered in BGS Borehole 81/53A. Borehole 79/6, further offshore, provides a section through beds of latest middle Eocene and late Eocene (Bartonian) age (HUGHES, 1981); these consist of alternations of glauconitic sands and silty mudstones. Most of the Eocene beds yield rich assemblages of radiolaria and diatoms, and are thought to be of middle Eocene age, but a calcareous fauna in the uppermost sand unit indicates the presence of late Eocene (Bartonian) beds. A Bartonian age is also indicated by dinoflagellate cyst assemblages.

The Bartonian sands are overlain by 60 metres of grey-brown glauconitic mudstones. A Rupelian age for these mudstones is indicated by both foraminiferal and dinoflagellate cyst assemblages. A further refinement is made possible by the occurrence of *Cibicides ungarianus* (indicating a late Rupelian age) in the upper 20 metres, and by the occurrence of *Ceratobulimina contraria* (indicating an early Rupelian age) in the lower 35 metres.

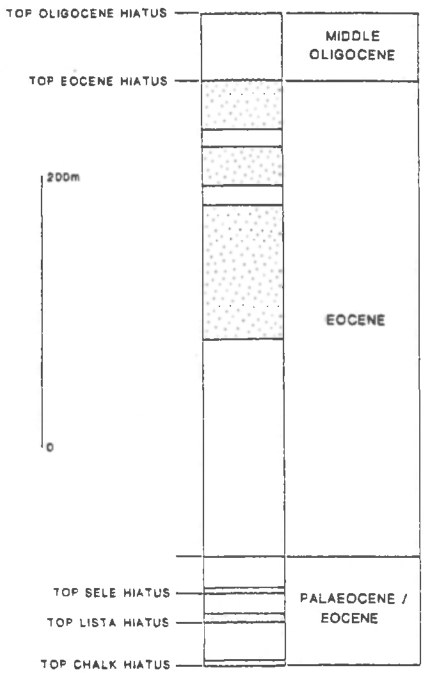

Figure 9 Composite Palaeogene sequence for the central area. The 'Eocene' sequence ranges from early to late, but the boundaries cannot be precisely located.

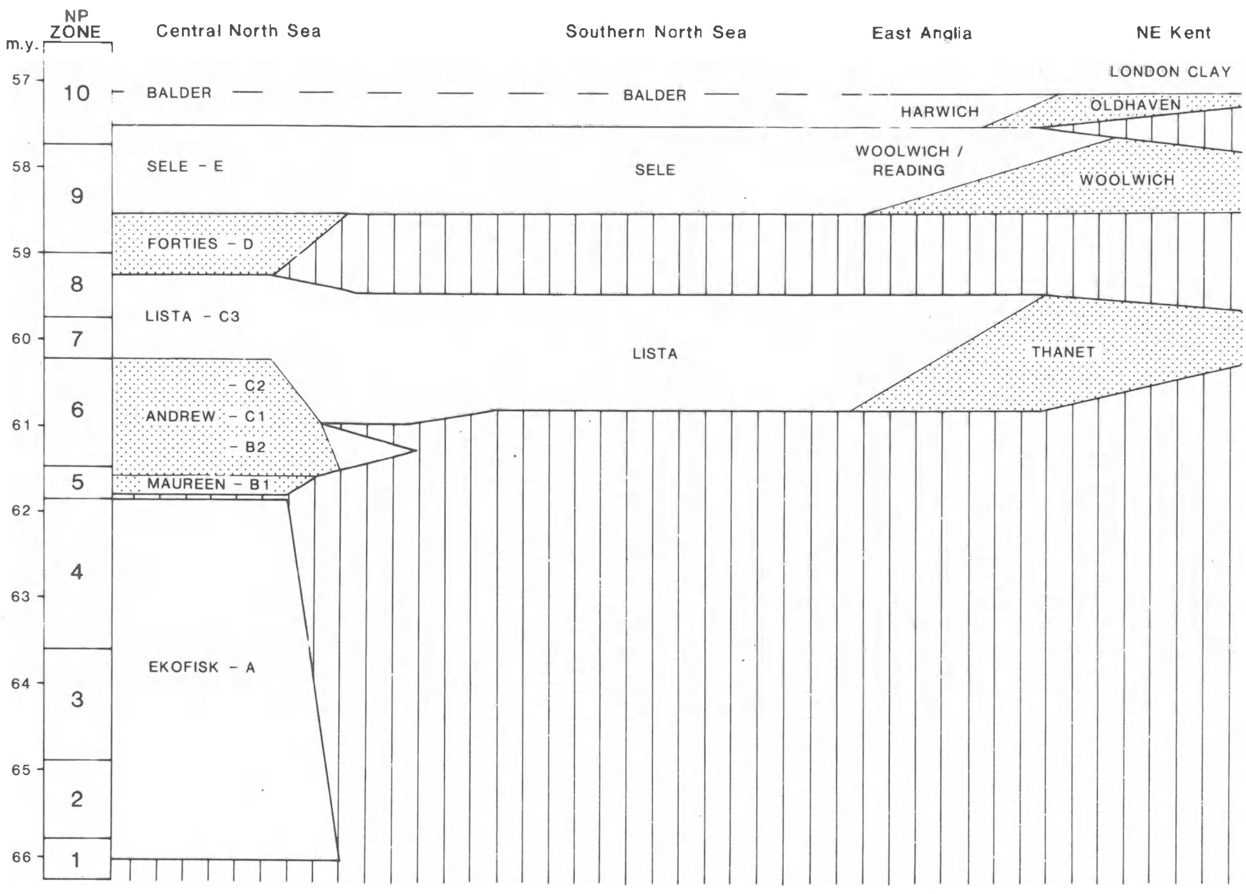

Figure 10 Chronostratigraphy of the Palaeocene and earliest Eocene beds of the UK central and southern North Sea areas, demonstrating the incomplete nature of the southern sequence. Letters and numbers in brackets refer to the Palaeocene 'units' of KNOX *et al.* (1981). Stippling indicates units dominated by sand. Chronostratigraphical and biostratigraphical scales after BERGGREN *et al.* (1985) ; correlation of lithostratigraphy with standard nannoplankton (NP) zones after KNOX (in press).

CONCLUSIONS

On the basis of all the onshore and offshore information, it is possible to draw up a probable composite Palaeogene sequence for the central area of the UK Flemish Bight (figure 9), with the following upward sequence :

1. Lista Formation : marine clays, sandy in the lower part, and often with a thin flint conglomerate at the base.

2. Woolwich and Reading beds : a basal marine transgressive sand overlain by red and brown lacustrine and pedogenic clays (Reading facies), perhaps showing transition to grey lagoonal clays (Woolwich facies).

3. Harwich Member : grey carbonaceous and sandy mudstones, with abundant volcanic ashes.

4. Marine Eocene mudstones passing upwards into alternating mudstones and sandstones.

5. Marine Oligocene mudstones.

As shown in figure 9, the sequence includes a number of hiatuses, especially in the Palaeocene / early Eocene section. The interruptions to Palaeocene/Eocene sedimentation can be related to the various tectonic events that took place preceding the opening of the Atlantic between Greenland and Rockall (KNOX *et al.*, 1981). A comparison with the central North Sea sequence (figure 10) reveals the incompleteness of the Palaeocene / Eocene sequence of the southern North Sea area. The following basin-wide events may be recognized in the form of regressive-transgressive cycles :

1. The 'end-Cretaceous' event : of uncertain duration, but involving minor erosion of the latest Maastrichtian beds in earliest Palaeocene (Danian) times. It was followed by a widespread transgression in which Danian limestones probably extended throughout the southern North Sea and into present-day onshore areas (possibly represented by the marine Montian).

2. End-Danian event : a major lowering of sea-level, with sedimentation becoming confined to the Central Graben area (Maureen Formation). It was followed by a gradual transgression (Andrew Formation), which probably extended into the southern North Sea and into present-day onshore areas (possibly represented by the Heersian of Belgium).

3. Mid-Andrew event : a lowering of sea-level, causing widespread removal of the thin lower Andrew Formation equivalents in the southern North Sea area. It was followed by a widespread transgression (Lista Formation, Thanet Beds, Lower Landenian). The transgression was accompanied by volcanism in the Hebridean region of western Scotland (KNOX and MORTON, 1983 ; 1988).

4. Top-Lista event : a major lowering of sea-level, with sedimentation becoming restricted to the Central Graben (Forties Sands). It was followed by a widespread transgression (Sele Formation, Woolwich and Reading Beds, Upper Landenian). The transgression was accompanied by renewed volcanism, representing crustal rifting to the north of Scotland (KNOX and MORTON, 1983 ; 1988).

5. Top-Sele event : a minor lowering of sea-level, accompanied by local folding and erosion of Sele Formation equivalents in marginal areas. It was followed by a limited transgression (Harwich Member, Oldhaven Beds, basal Dongen tuffite and sand equivalents). The transgression was accompanied by intense basaltic volcanism, associated with the onset of sea-floor spreading to the north of Scotland.

6. Top-Harwich event : a minor lowering of sea-level, marked by a sharp decrease in ash content and minor reworking. It was followed by progressive and widespread transgression (London Clay of Kent and Hampshire, Ieper Clay).

Following the top-Harwich event, sedimentation was more or less continuous until the late Eocene. Then a major change in subsidence and sedimentation patterns took place (end-Eocene event), with local non-deposition or erosion of late Eocene sediments in both the central basin and marginal areas. Sedimentation was renewed in the Oligocene, but with relatively thin and condensed sequences in the southern North Sea area.

It is not possible to trace the depositional history from the Oligocene into the Miocene, because the interval is represented by thin, impersistent sediments that have probably undergone repeated reworking. Furthermore, none of the borehole sequences has been cored. Information from the central North Sea area suggests that sedimentation continued with only minor interruptions into the middle Miocene, when a major change in the pattern of basin subsidence and source uplift took place, corresponding to the boundary between the Hordaland and Nordland Groups (DEEGAN and SCULL, 1977) of the central and northern North Sea area.

ACKNOWLEDGEMENTS

I am grateful to Dr. T.D.J. CAMERON (BGS) for providing information on the surface and subsurface distribution of Palaeogene units in the Flemish Bight area, and for the assistance of other members of the BGS Marine Earth Sciences Research Group. Permission to use information from the Trinity House Boreholes is gratefully acknowledged. Publication is with the approval of the Director, British Geological Survey (N.E.R.C.).

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4 Preliminary seismic-stratigraphic maps and type sections of the Palaeogene deposits in the Southern Bight of the North Sea.

J.P. HENRIET (1), M. DE BATIST (1)(2), H. DE BRUYNE (1), P. HELDENS (1), J.P. HUYLEBROECK (1), F. MOSTAERT (1), E. SEVENS (1), J.P. AUFFRET (3) and B. D'OLIER (4)

1. Renard Centre of Marine Geology, Geological Institute, State University of Gent, Krijgslaan 281/S8, B-9000 Gent, Belgium.
2. Research Assistant, National Fund for Scientific Research, Belgium.
3. Laboratoire de Géologie Marine, Université de Caen, Département de Géologie, Esplanade de la Paix 14, F-14032 Caen CEDEX, France.
4. Department of Geology, City of London Polytechnic, Walburgh House, Bigland Street, London E1 2NG, United Kingdom.

ABSTRACT

High-resolution reflection seismic investigations carried out from 1978 up to 1984 in the Southern Bight of the North Sea yield a detailed picture of the stratigraphic build-up of the Palaeogene sequences and their spatial distribution in this region. The application of classical concepts of seismic-stratigraphic analysis leads to the identification of 14 depositional sequences.

Several units display characteristic seismic facies features, which can be interpreted in terms of depositional environment and compaction history.

The intention of this paper is to set the general seismic-stratigraphic stage for the Palaeogene sequences of this region, thus yielding a marine data base which could prove useful in correlation studies with lithostratigraphic and chronostratigraphic land observations.

INTRODUCTION

Some years ago, at the 1979 Texel Meeting on Holocene Sedimentation in the North Sea Basin, three investigation teams from City of London Polytechnic, Université de Caen and Rijksuniversiteit Gent decided to join their resources for studying the superficial geology of the southern North Sea, between the coasts of northern France, Belgium and the Thames estuary. The project "Seismic Stratigraphy, Southern Bight, North Sea" (SSSBNS) was born.

At that time, the British sector was fairly well covered by sparker and boomer lines. In the Belgian and French sectors of the continental shelf however, seismic data were scarce. Only a few lines published by

HOUBOLT (1968) and BASTIN (1974), as well as harbour development surveys for Dunkerque and Zeebrugge (HENRIET *et al.*, 1978) were available.

In subsequent years, several cruises have been organized by the aforecited universities, aiming to fill the gap in the seismic coverage on the French/Belgian continental shelf. Vessels were provided by NERC in the United Kingdom and by CNRS and Wimereux Station of Université de Lille in France. In Belgium, support was granted by the Science Police Office, the Management Unit of the Mathematical Model of the North Sea and the Scheldt Estuary and the Belgian Navy. In 1982, a detailed seismic mapping of Quaternary deposits off the western Belgian coast was commissioned by the Administration of Mines, within the framework of studies of the environmental impact of dredging activities on sand banks.

This paper reviews our state of knowledge of the seismic stratigraphy of the Palaeogene substratum of the Belgian sector and adjoining areas of the Southern Bight of the North Sea, from surveys carried out up to 1984, the year of the Ghent Colloquy. As such, it presents a working base for subsequent detail studies, such as the Marine Geology programme of the Belgian Science Policy Office and the systematic, detailed mapping programme of the superficial deposits of the Belgian continental shelf, commissioned by the Belgian Geological Survey. These on-going new surveys, implemented with drilling programmes, will no doubt in the future lead to a refinement of the present stratigraphical model.

The seismic grid available by 1984 (figure 1) totalized about 7000 km. It stretched from coast to coast, bridging

Figure 1 Map showing the 1984 seismic coverage on the French, Belgian, Dutch and UK sectors of the continental shelf, between 51°-52°N and 2°-3°30'E.
Full lines indicate seismic profiles shot within the "SSSBNS"-project. Dashed lines indicate previous seismic profiles.

the former gap in the central and northern Belgian sector and linking with Borehole 79/6 of the British Geological Survey in the north, used as reference well. The apparent density of this grid should not be misleading, the represented lines being of varying quality and penetration. However, it forms a working base, which will be completed and enhanced, but which already yields a coherent picture of the stratigraphy of the Palaeogene deposits of this region.

GEOLOGICAL SETTING

The Southern Bight of the North Sea and especially its southwestern part, off the North French and Belgian coasts (area between 2-3°30'E and 51-52°N), is situated on top of the Palaeozoic London-Brabant Massif, which has formed a stable structural high during a long geological period. In post-Palaeozoic times, it was probably submerged for the first time during the Upper Cretaceous sea level high, although possibly structurally controlled local earlier depositions cannot be completely ruled out. This area has always been situated at the southern rim of the subsiding North Sea Basin, which implies that many sea level changes and/or tectonic events are reflected in the lithology and the configuration of its sedimentary cover.

The Upper Cretaceous chalk on top of the London-Brabant Massif is covered by a comprehensive sequence of Palaeogene beds. This sequence has an average dip of about 0.5° in northeastern direction, probably bound to both the subsidence movements of the North Sea Basin and to some Early- to Mid-Tertiary tectonics in the southwest (Weald-Artois anticline). Being directly exposed on the sea bed or only slightly concealed by a thin Quaternary cover, it offers in this region a unique possibility of investigation of its internal structure and facies, with a maximum of resolution.

SEISMIC-STRATIGRAPHIC APPROACH

The seismic-stratigraphic analysis of the Palaeogene was carried out in accordance with the basic principles laid by VAIL *et al.* (1977). It resulted in the identification of depositional sequences, separated by low-angle unconformities or their correlative conformities. These sequences have then been analysed in function of their seismic facies, which often resulted in a better definition of poorly-developed sequence boundaries. A further differentiation of larger sequences into sub-units could thus be achieved on base of striking seismic facies features.

The identified depositional sequences have been labeled with a character-digit symbol, suggesting their most probable chronostratigraphic identity. This preliminary chronostratigraphic interpretation is partly based on a tentative correlation with stratigraphic units in the Belgian coastal plain and the rest of the Flanders area and will presumably have to be revised as soon as micropalaeontological and sedimentological interpretations of judiciously planned boreholes on the Belgian continental shelf will be available.

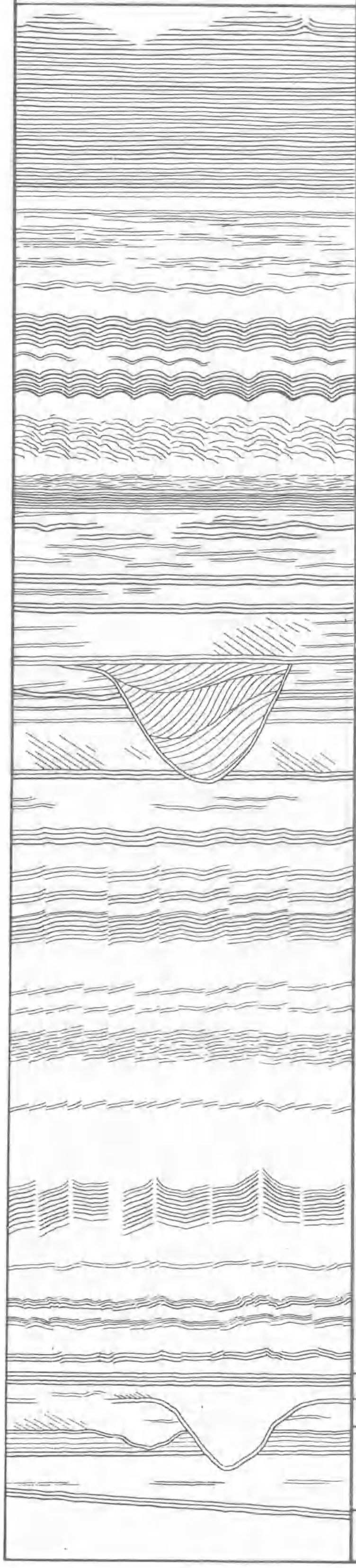
Until now up to 14 Palaeogene depositional sequences have been identified :

- R1 and R2 (Rupelian),
- P1 (Priabonian),
- B1 and B2 (Bartonian),
- L1 and L2 (Lutetian),
- Y1, Y2, Y3 and YX (Ypresian),
- T1, T2 and T3 (Thanetian).

Main reflectors are labeled with the character-digit symbol of the overlying depositional sequence, accompanied by a rank number, both separated by a dot. First order ranks are assigned to sequence boundaries, while higher rank numbers are given to significant internal reflectors.

The results of this detailed seismic-stratigraphic analysis have been compiled into a schematic

SCHEMATIC TYPE SECTION

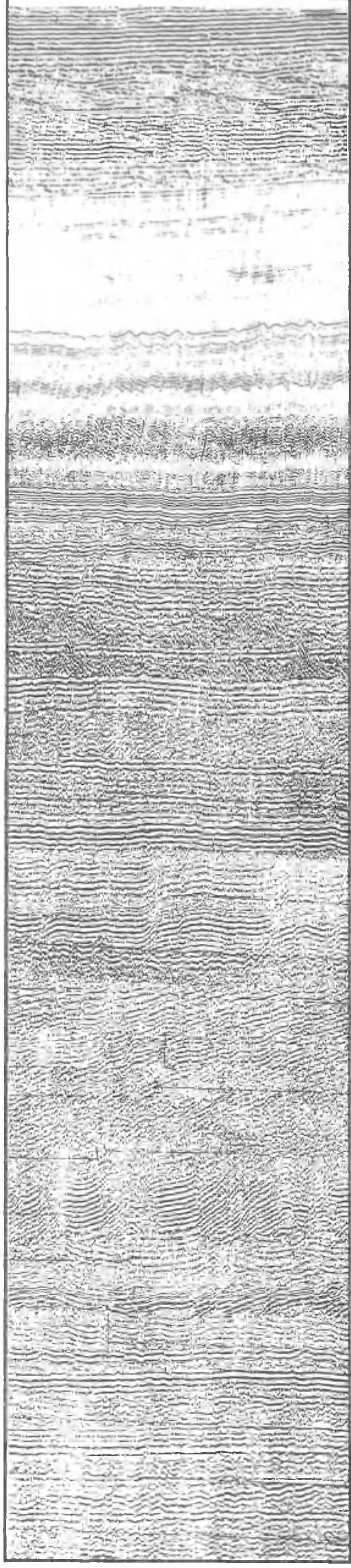


DEPOSITIONAL SEQUENCES NAME	BOUNDARY NAME	BOUNDARY TYPE	REFLECTOR DESCRIPTION AMPL.	REFLECTOR DESCRIPTION CONT.	REFLECTOR DESCRIPTION FREQ.	THICKNESS (ms) two-way time	THICKNESS Vi (m/s)	THICKNESS (m)
	N1.1	onlap/truncation	low	cont.	low			
R2						max. 120	uphole 1650 ±80	max. 99
	R2.1	concordance	medium	cont.	low			
R1						±10	1700 ?	±8.5
	R1.1	onlap/truncation	low	cont.	low			
P1						40 / 9-10		34 / 7.5-8.5
	P1.1	concordance	medium	cont.	low			
B2						8 / 8		-13.5 -
	B2.1	baselap	high	cont.	low			
B1						5 / 20 / 5 / 5 / 8		34
	B1.1	truncation	medium	cont.	high			
L2						25-35 / ±13	refraction 1580 ±40	21-30 / ±11
	L2.1	toplap	high	cont.	low			
L1						15-19	1700?	12.5-16
	L1.1	downlap/truncation	medium	cont.	low			
YX* Y3						10-20	1600	8-16
	Y3.1	baselap/truncation	variable	cont.	low			
Y2						5-15 / 20-25	downhole 1750 ±50	4.5-13 / 17.5-22
	Y2.1	downlap	high	cont.	low			
Y1						200-220	uphole 1620 ±50	162-178
	Y1.1	concordance	high	cont.	low			
T3						11-13	well correlation ±1800	10-11.5
	T3.1	downlap/truncation	medium	discont.	low			
T2						9-11		8-10
	T2.1	downlap/truncation	medium	cont.	medium			
T1						18.5-30		16.5-27
	T1.1	onlap	high	cont.	low			

Figure 2 Schematic seismic-stratigraphic sequence chart and correlation table of the Southern Bight Palaeogene deposits.
* for YX see figure 6.

DEPOSITIONAL SEQUENCES

SEISMIC FACIES



SEISMIC FACIES: TYPE SECTION / DESCRIPTION	NAME	LITHOSTRATIGRAPHIC CORRELATION
regular pattern of continuous, parallel, high- to medium-amplitude reflectors	R2	Boom Member (B)
continuous, parallel, low-amplitude reflectors	R1	Berg Member (B)
homogeneous pattern of continuous, parallel, low- to medium-amplitude reflectors continuous, parallel, draping reflectors of variable amplitude	P1	Zelzate Formation (B)
reflection-free + low-amplitude, discontinuous, draping reflector medium-amplitude, draping reflectors	B2	Meetjesland Formation (B)
reflection-free interval convex mounds of medium-amplitude, prograding, hummocky reflectors reflection-free interval hummocky reflectors with a shingled reflector on top regular set of continuous, parallel, high-frequent reflectors	B1	
2 to 3 discontinuous, subparallel, very high-amplitude reflectors discontinuous, parallel to subparallel reflectors of variable amplitude 2 continuous, high-amplitude, parallel reflectors in the south a third, discontinuous, low-amplitude reflector	L2	Oedelem Member (B)
low-amplitude, parallel reflectors or parallel-oblique clinoforms or reflection-free	L1	
low-amplitude, discontinuous, parallel reflectors or parallel-oblique clinoforms	Y3	Merelbeke Member (B)
continuous, low- to medium-amplitude, parallel reflectors reflection-free or very low-amplitude, parallel-oblique, prograding clinoforms	Y2	Egem Member (B)
undisplaced, faulted blocks with alternately tilting and downwarping bedding terminations major tilted blocks, convolute structures with broad synclines and cuspate anticlines, diapyrs block faulting, tilted and bended blocks, randomly dipping fault planes	Y1	London Clay (U.K.) or Flanders Clay (B) (Ieper Clay)
oblique or shingled clinoforms and low-amplitude, discontinuous, subparallel or hummocky reflectors	T3 T2	lagoonal to continental Landen Formation (B)
few parallel reflectors of variable amplitude, separated by reflection-free intervals	T1	marine Landen Formation (B)

seismic-stratigraphic sequence chart of the Palaeogene deposits. Descriptions of seismic facies and reflector characteristics are added and can be visualized on an accompanying synoptic seismic record, a composite of several seismogram sections acquired with comparable source signatures (figure 2). A detailed seismic-stratigraphic solid map of the Palaeogene on the continental shelf is shown on figure 3. In the following chapters the most striking features of each depositional sequence will be briefly described.

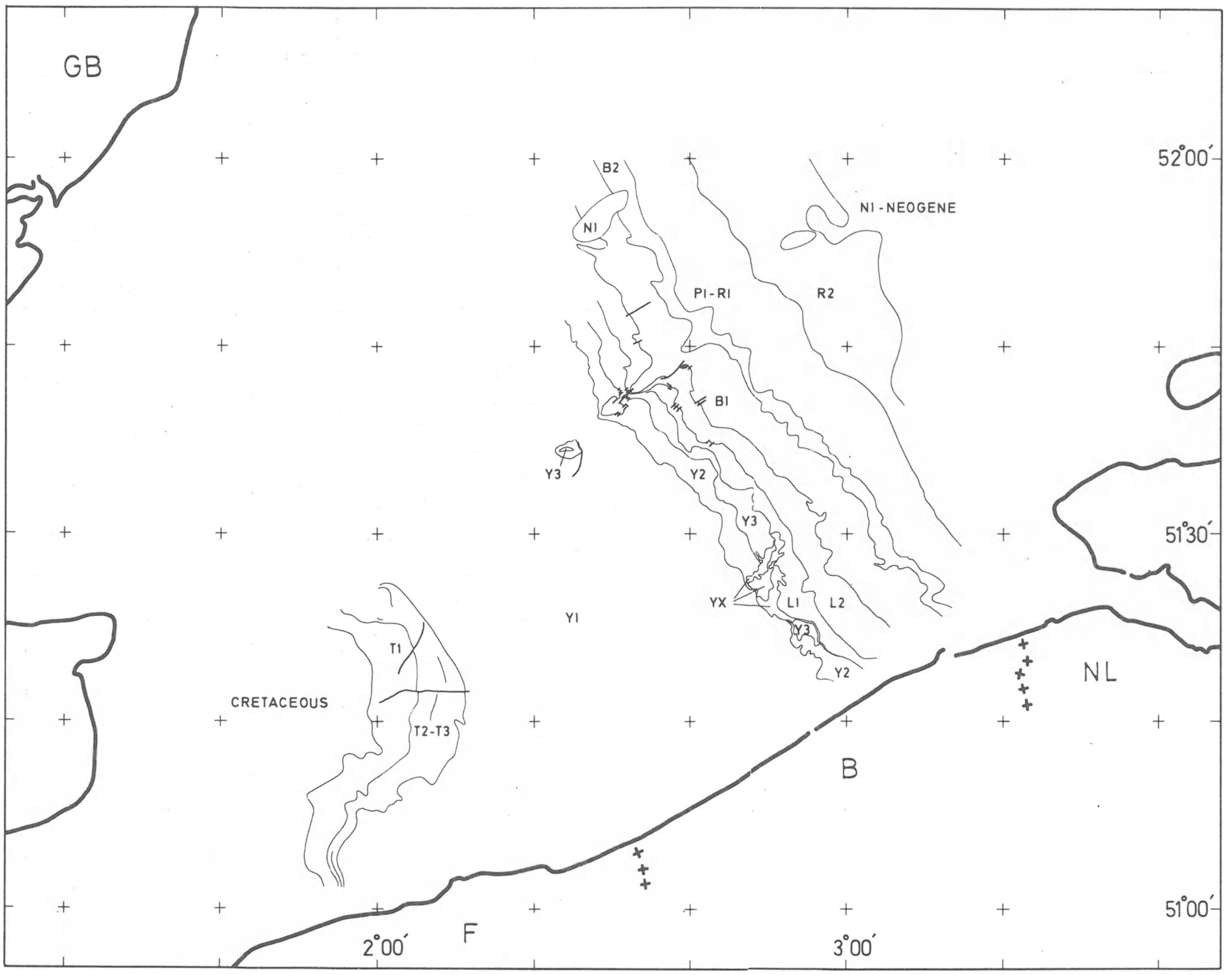


Figure 3 Seismic-stratigraphic solid map of the Southern Bight Palaeogene deposits, acquired by interpretation of seismic profiles shot within the "SSSBNS"-project. Interpretations of previous seismic lines (e.g. UK sector) have not been implemented yet.

THANETIAN

The first three seismic-stratigraphic units (T1, T2 and T3), deposited on top of the relatively thin Upper Cretaceous cover of the London-Brabant Massif, belong to the Late Palaeocene (Late Thanetian) and can be correlated with the Landen Formation in Belgium and with the Thanet and Reading/Woolwich Beds in southeastern England.

T1

The lowermost boundary op the Palaeocene depositional sequences is formed by the unconformity at the top of the Cretaceous chalk (Turonian-Senonian) : reflector T1.1. The overall width of the overlying T1-unit decreases by onlap from about 30 to 18.5 ms (two-way time) towards the southwest. This can very clearly be observed in the northwestern part of the Belgian continental shelf.

T2 and T3

The boundary with the second unit, reflector T2.1, is an erosion surface, locally with well developed gully erosion features, as observed both near the English and French-Belgian coasts. T2.1 in its turn is on some places (figure 4) scoured by the basal reflector T3.1 of the overlying sequence. The seismic facies of both sequences is often characterized by abundant prograding stratifications, hummocky reflection patterns, discontinuous, subparallel reflections and channel-fill reflection configurations (HUYLEBROECK, 1985). The total width of the T2-T3 interval is about 20-24 ms and remains fairly constant, although it may increase substantially in ravinations.

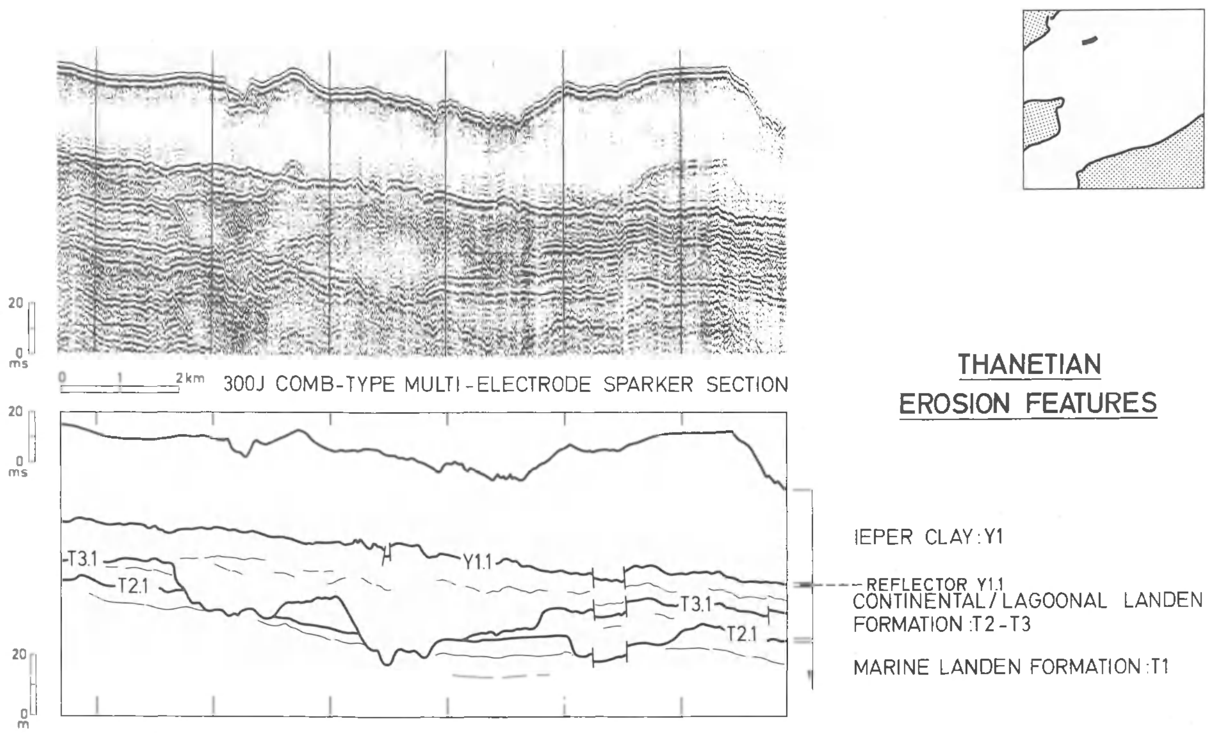


Figure 4 Analog record of a sparker section and interpreted line-drawing showing Thanetian gully erosion features on the UK continental shelf.
Vertical scales in ms are two-way time. Vertical scales in m are calculated with an average velocity of 1700 m/s.

An overlying reflector, Y1.1, independently interpreted by British as well as by Belgian investigators as the base of the London/Flanders Clay, is considered to form the upper boundary of the T3-unit, although on many seismic sections the transition between T3 and Y1 does not appear as an unambiguous unconformity.

Through correlation of a seismic calibration line off the Belgian coast with the Ostend boring (BGD 21E-41), where the boundary between the Senonian and the Landen Formation was found at -211 m and the boundary between the Landen Formation and the Flanders Clay at -174 m, an average velocity of about 1800 m/s can be proposed for the three sequences (total width between T1.1 and Y1.1 : 42 ms).

Application of this velocity model results in almost perfect fit of T1 with the marine Landen Formation, as interpreted in the Ostend boring, and of T2-T3 with the overlying lagoonal to continental Landen Formation, a correlation which was already suggested by the seismic facies.

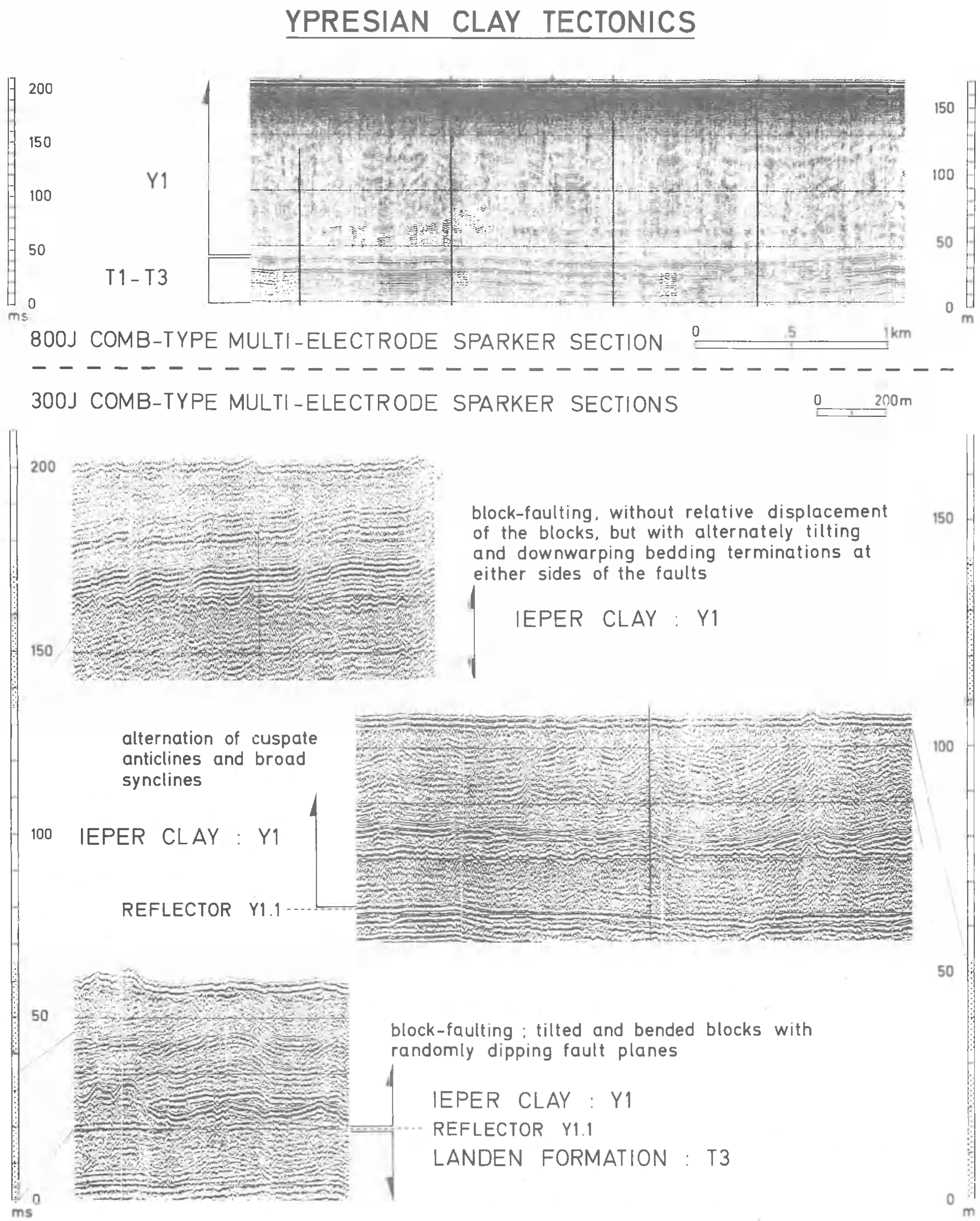


Figure 5 Analog record of a sparker sections showing Ypresian clay-tectonic features.
Vertical scales in ms are two-way time. Vertical scales in m are calculated with an interval velocity of 1620 m/s.

YPRESIAN

Y1

The next depositional sequence of major importance, Y1, is most probably entirely built up by the Flanders Clay. The Ypresian lithostratigraphic nomenclature used here is in accordance with STEURBAUT and NOLF (1986). This unit is more or less equivalent to the British London Clay. In field-geological and geotechnical use, it is also commonly referred to as Ieper Clay. Its upper boundary is formed by the distinct reflector Y2.1, one of the most significant seismic-stratigraphic markers of the entire Palaeogene in the Southern Bight of the North Sea.

The more or less homogeneous series of weak parallel reflections, typical for the seismic facies of this unit, might be partly caused by interference phenomena, from reflections at numerous laminae and thin beds of variable grain size and compaction characteristics.

The most striking features of this depositional sequence are beyond doubt the clay-tectonic deformations. These often show a remarkable vertical zonation (DE BATIST *et al.*, this volume). Some typical intervals are shown on figure 5. In some areas however the whole internal reflection pattern may become completely chaotic.

These deformations seem to be confined to the Y1-unit and gradually fade out into the overlying sediments. Their origin could be related to the compaction history of the Ieper Clay (HENRIET *et al.*, 1988).

There is no clear seismic evidence (e.g. diffractions) for the occurrence of septaria or concretions, such as those which have been described in the London Clay (HEWITT, 1982).

A typical reflector at about 8 ms above Y1.1 has been correlated with the Harwich stone band (Personal communication, B. D'OLIER), a hard horizon containing volcanic ash, which can be traced over much of the North Sea Basin and is considered by some authors to form the boundary between the Palaeocene and the Eocene (JACQUE and THOUVENIN, 1975).

Uphole shooting in boreholes in Belgium has yielded an average velocity of 1620 m/s for the Flanders Clay (HELDENS, 1983). The total thickness of the clay varies between 120-140 m in Northwest Belgium (in the Knokke boring BGD 11E-138 even 145 m) to 150 m in Southeast England. Y1's thickness on the Belgian continental shelf reaches about 170 m, when the considered velocity model is applied on a two-way reflection time of about 210 ms.

TRANSITION YPRESIAN-LUTETIAN

Off the Belgian coast, a complex group of depositional sequences heralds a distinct change in depositional circumstances, contrasting with the low-energy facies of the underlying Y1-unit. A peculiar geometry with ravinations and obliquely prograding bedding sets suggests some analogy with Late Ypresian and Lutetian units in Flanders (DE BATIST *et al.*, this volume).

Y2

The boundary Y2.1 between the Ieper Clay and Y2, the first of these depositional units, is locally marked by a weak downlap pattern. Its upper boundary is characterized by strong truncations. The total thickness of Y2 therefore is very variable. Some kilometres off the Belgian coast, it apparently has been entirely removed by erosion, leaving a basinlike depression.

Especially in the north two seismic facies units can be identified in Y2 : the aforecited lower downlapping unit (± 20 ms), apparently no longer affected by the underlying clay-tectonic deformations, and an upper one with a set of parallel reflections. This unit thins towards the south, due to erosional truncation.

Y3

The base of Y3 is formed by the northerly fading reflector Y3.1, defined by discrete baselap of internal reflectors and locally by truncation of reflectors of the underlying unit.

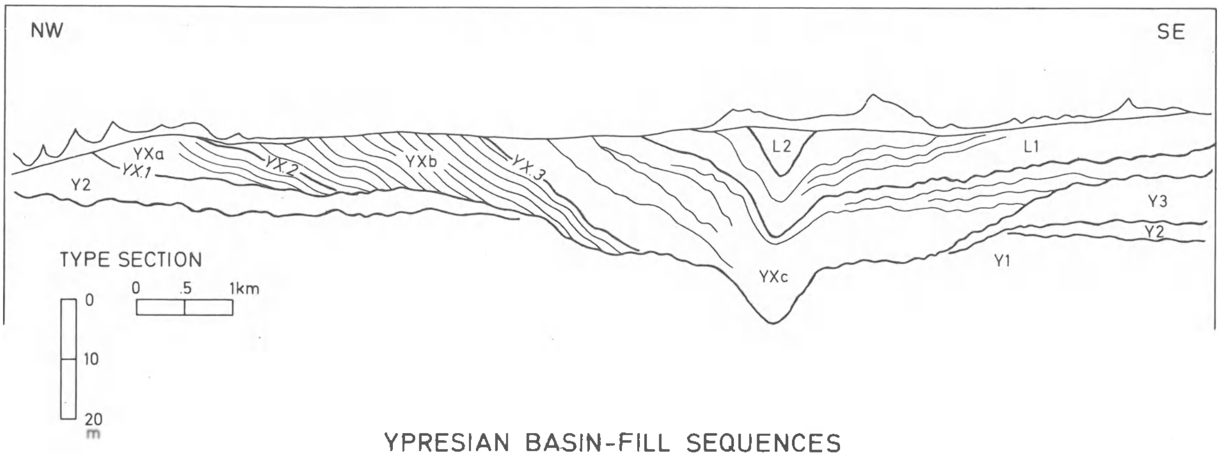
In the central part of the area, no distinction in seismic facies can be made between Y2 and Y3. Further south however, Y3 is characterized by a pattern of tangential oblique clinoforms, gently prograding towards the south.

In some areas off the Belgian coast the Y3-unit has been removed by erosion, but further to the north its thickness becomes constant.

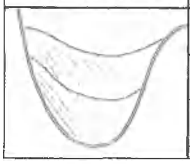
YX

The overlying deposits of the YX-sequence seem to be confined to a basinlike depression, located about 20 km north of Ostend. The basal surface is penetrating into Y3, Y2 and even into the top of the Flanders Clay. Three subsequences can be identified in a prograding basin-fill configuration, respectively named YXa, YXb, and YXc (figure 6). Their detailed configuration is described by DE BATIST *et al.* (this volume).

Neither the exact shape of the erosive basin, obscured by a narrow synclinal fold (figure 6), nor its areal extent towards the east, where it plunges under younger Lutetian deposits, could be revealed yet. Some observations suggest a circular, rather than an elongated shape.



YPRESIAN BASIN-FILL SEQUENCES



SCHEMATIC TYPE SECTION	DEPOSITIONAL SEQUENCES										
	NAME	BOUNDARY					THICKNESS			SEISMIC FACIES	LITHOSTRATIGRAPHIC CORRELATION
		NAME	TYPE	REFLECTOR DESCRIPTION			(ms) two-way time	Vi (m/s)	(m)	DESCRIPTION	
				AMPL.	CONT.	FREQ.					
	YXc						0-25	1700 ?	0-13	sigmoid or parallel oblique, prograding clinoforms of variable amplitude	Vlierzele Member (B)
		YX.3	baselap/toplap	variable	variable	medium					
	YXb						0-20		0-17	sigmoid or parallel oblique, prograding clinoforms of variable amplitude	
		YX.2	baselap/toplap	variable	variable	medium					
	YXa						0-15		0-21	sigmoid or parallel oblique, prograding clinoforms of variable amplitude	
		YX.1	baselap/truncation	variable	variable	medium					

Figure 6 Type section and seismic-stratigraphic sequence chart and correlation table of Ypresian basin-fill sequences off the Belgian coast.
Vertical scales in m are calculated with an average velocity of 1700 m/s.

LUTETIAN

L1

The base of the following depositional sequence, reflector L1.1, truncates all underlying units within 35 kilometres from the Belgian coast. Further to the north the gradual transition of Y2 into Y3 continues into L1.

Above the northern part of the erosive basin, the seismic facies of L1 is characterized by oblique, northeasterly prograding clinoforms, which laterally grade into parallel, even reflections.

L1's upper boundary, reflector L2.1, is marked by a toplap of the underlying oblique internal stratifications.

L2

The depositional sequence L2, overlying L2.1, is relatively homogeneous, although two main seismic facies units can be observed (figure 2).

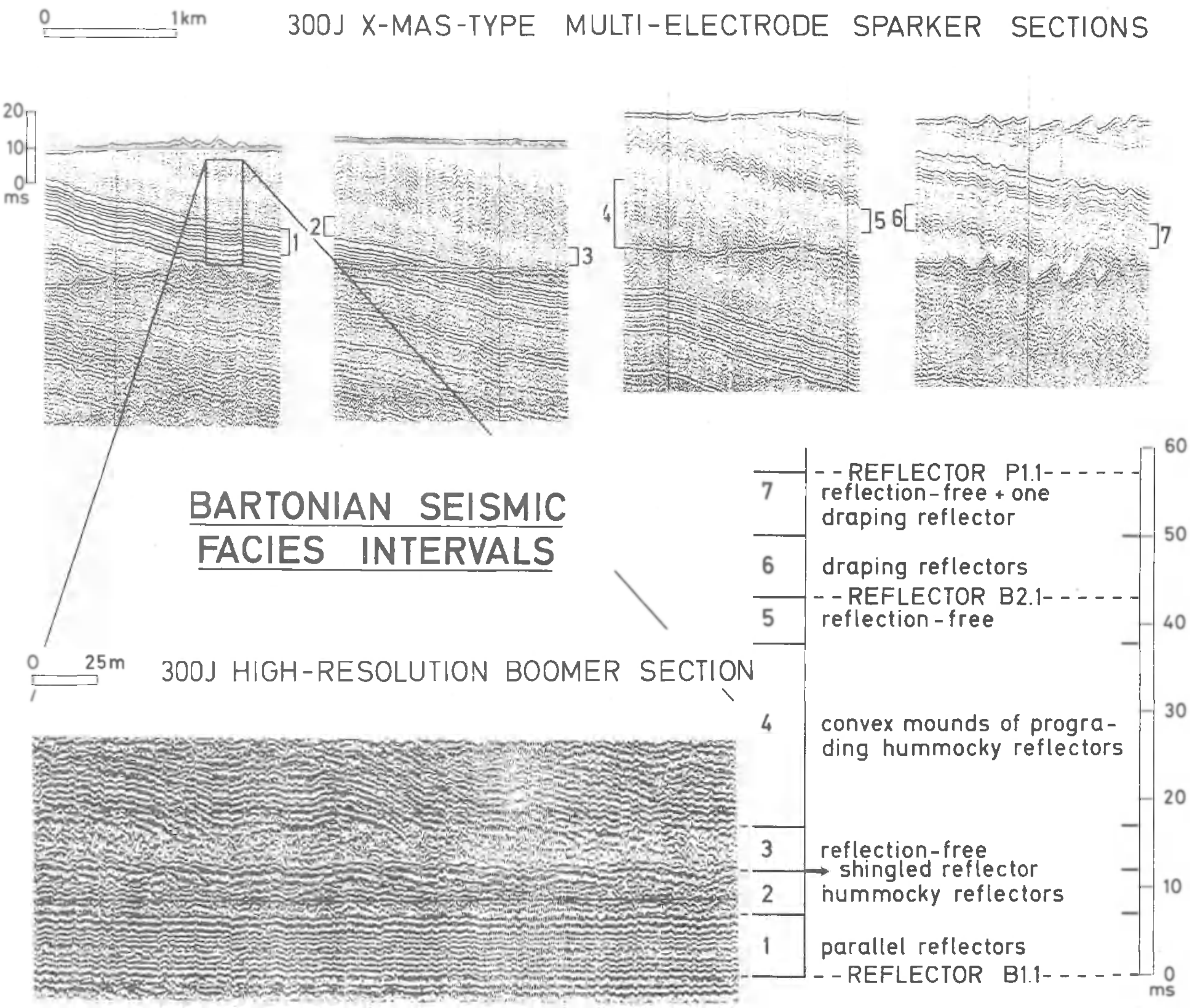


Figure 7 Analog record of a sparker and boomer sections showing 7 Bartonian seismic facies intervals. Vertical scales in ms are two-way time.

Two or three discontinuous but very strong reflectors, observed at the top of the upper seismic facies unit have been identified in a geological study for the Zeebrugge harbour extension (HENRIET *et al.*, 1981) as the glauconitic calcareous sandstone banks, which occur in the upper part of the Oedelem Member.

BARTONIAN

B1

A few milliseconds under the base of the following major sequence B1, a discrete unconformity can locally be observed in the best developed synclinal fold axes. This weakly truncating reflector is however of little interpretative use, especially by its very local character. For this reason, the first overlying continuous reflector has been taken as the lower boundary of B1, labeled B1.1. It has been correlated with the base of the Bartonian clays, the Asse Member, in the Zeebrugge harbour extension study (HENRIET *et al.*, 1978).

In B1 several seismic facies units can be identified (figure 7), probably reflecting the typical sand-clay alternation of the Bartonian Meetjesland Formation in Flanders (JACOBS, 1978).

A lower interval shows a very regular pattern of parallel reflections. By virtue of its lateral continuity and striking appearance, it forms an important seismic-stratigraphic guide. It is followed by an interval with a rather hummocky reflection pattern. On high-resolution profiles the top boundary of this interval displays a characteristic shingled configuration, built up of gently dipping, sigmoid segments prograding in eastnortheastern direction. Another almost reflection-free interval is overlain by a major sequence, showing pronounced prograding hummocky reflection configurations, terminated by convex downlaps on the reflection-free horizon. These moundlike piles of prograding reflections are followed by another reflection-free unit and are finally draped by a set of reflectors of the overlying sequence.

B2

The typically draped base reflector B2.1 of the second major Bartonian sequence B2 is locally marked by a low-angle downlap pattern.

Although lateral facies changes occur towards the northern part of the area, B2 can commonly be divided into two seismic facies units (figures 2 and 7).

The amplitude of the characteristic draping within B2 seems to be progressively decreasing towards the top of the sequence, fading out in the overlying unit. Such a phenomenon could find its origin in differential compaction processes.

PRIABONIAN

P1

The following depositional sequence, P1, has probably to be correlated with Priabonian (Tongrian) deposits (sands of the Zelzate Formation). It also consists of two seismic facies units.

The first unit is a series of parallel reflections, with a typical lateral variation in amplitude.

The second unit, also affected by lateral amplitude variations, has a very regular, homogeneous seismic facies.

RUPELIAN

R1

The R1-sequence is most likely to be correlated with the Rupelian sands (Berg Member).

Its lower boundary, reflector R1.1, is a weak erosion surface, best marked close to the Dutch coast, where it also forms the base of a discrete onlap pattern. Usually, however, little distinction can be made between the seismic facies of R1 and P1.

R2

The uppermost unit of the Palaeogene seismic-stratigraphic sequence can entirely be correlated with the Boom Member, the Rupelian clay deposit of major stratigraphic importance.

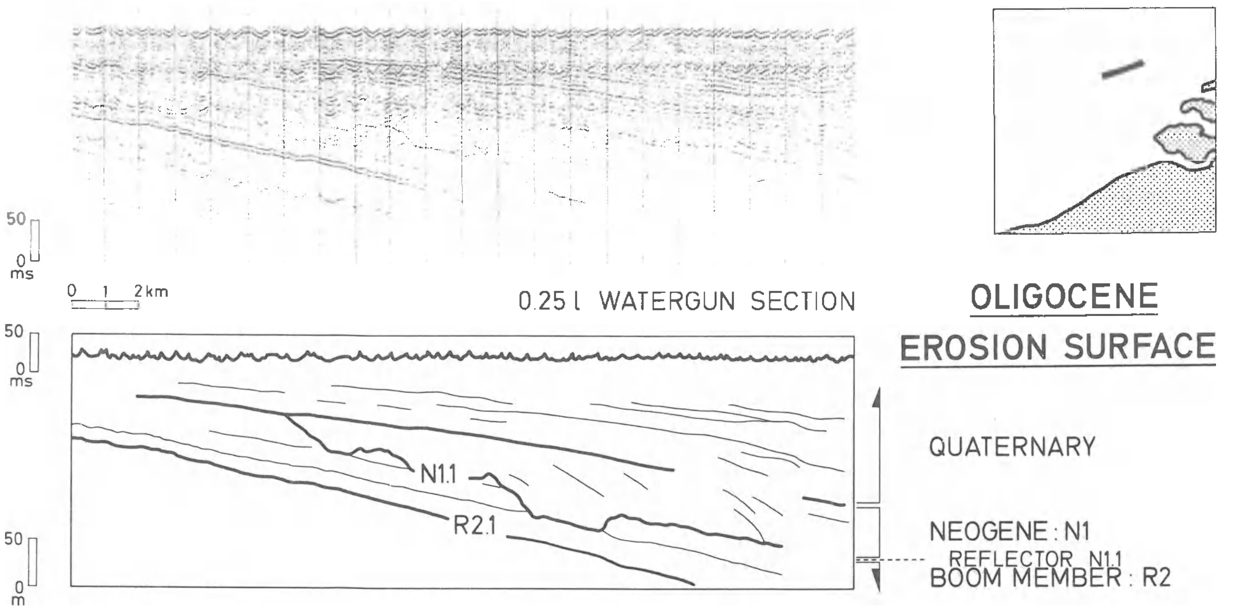


Figure 8 Analog record of a watergun section and interpreted line-drawing showing a Late Oligocene erosion surface on the Dutch continental shelf.
Vertical scales in ms are two-way time. Vertical scales in m are calculated with an average velocity of 1600 m/s.

The upper boundary of this unit is clearly defined by a remarkably well preserved erosion surface, with well-developed valley profiles (figure 8). It bears witness of the important Late Oligocene global sea level lowstand (VANNESTE, 1987), an estimated 250 m below present level (HAQ *et al.*, 1986). The freshness of this erosion profile suggests a rapid burial by the subsequent Neogene transgressions. The maximal observed width of R2 amounts to 120 ms.

The seismic facies of R2 is characterized by a very regular pattern of parallel reflectors. On very high-resolution sections, several of these horizons turn out to be built up of an alignment of diffraction hyperbolae (figure 9), each of them corresponding with the reflection at concretions or septaria, which are indeed frequently observed in the Boom Clay (VANDENBERGHE, 1974 ; VANDENBERGHE, 1978).

Up to 25 septaria levels could be identified by seismic reflection investigations on the Scheldt river, carried out for the pre-metro works and storm surge barrier project (HELDENS, 1983). The lowermost of these diffraction horizons has been taken as boundary reflector R2.1, since the base of the Rupelian clay, identified in boreholes a few metres underneath, does not show up as a reliable reflector.

Another striking feature of the Boom Clay is the occurrence of diapiric upwellings, as observed in the Scheldt estuary (figure 10). Such clay diapyrs have also been described on land by LAGA (1966). They are possibly caused by relaxation processes of underconsolidated sediments.

Detailed velocity analyses, carried out in the framework of the same site investigations in the Antwerp region, have provided an interval velocity for the Boom Clay of about 1620 m/s (SCHITTEKAT *et al.*, 1983). This low velocity could account for the distinct polarity change at the top of this sequence in the southern North Sea, where the clay is probably overlain by Pliocene and Pleistocene sands, having higher interval velocities. The important velocity pull-up of the internal B2 reflectors underneath the most important valley incisions (figure 9) also argues for such a velocity inversion.

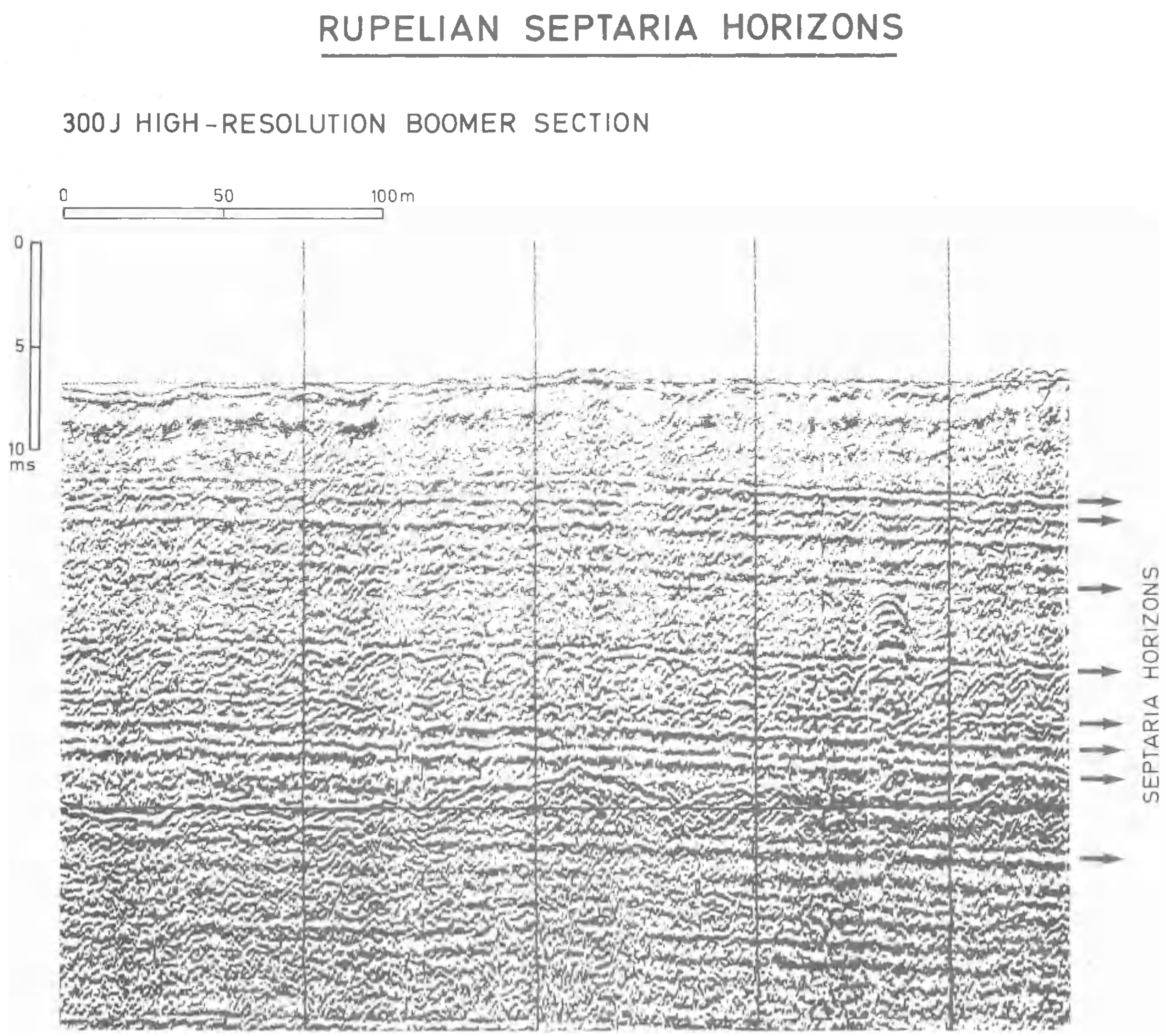


Figure 9 Analog record of a boomer section showing Rupelian septaria horizons on the Scheldt river. Vertical scales in ms are two-way time.

CONCLUSION

The detailed high-resolution reflection seismic mapping of the Southern Bight of the North Sea yields a new insight in the stratigraphic build-up of Palaeogene depositional sequences and in their spatial distribution patterns. Classic principles of seismic-stratigraphic analysis can be applied with success on these sequences, which often do not exceed a few tens of metres in thickness.

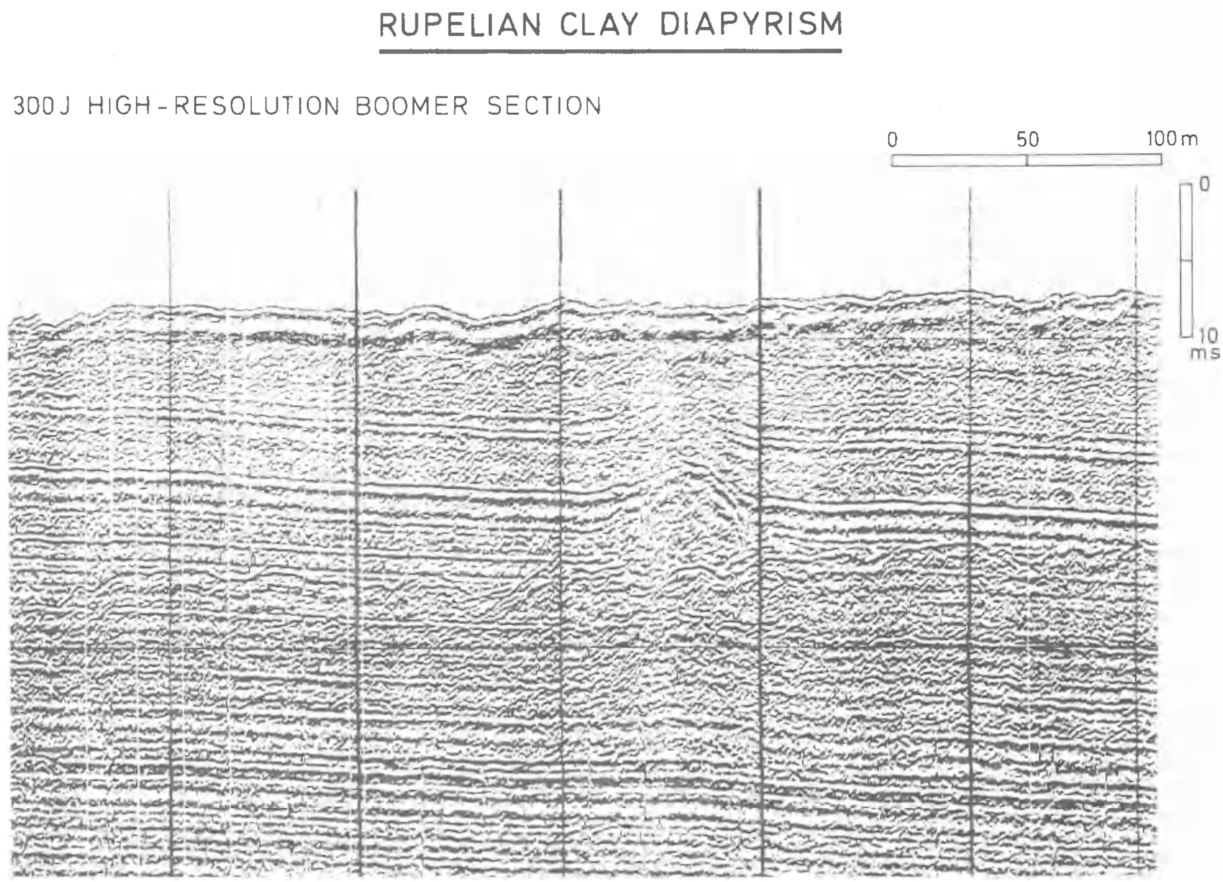


Figure 10 Analog record of a boomer section showing Rupelian clay diapirism on the Scheldt river. Vertical scales in ms are two-way time.

Many units display characteristic seismic facies features, which can be interpreted in terms of sedimentary environment and compaction history.

Correlation of this marine geophysical data base with more classic litho- and chronostratigraphic land observations should enhance our understanding of the Cenozoic geological setting and evolution of the southern North Sea.

ACKNOWLEDGEMENTS

The offshore reflection profiling has been carried out in the framework of studies supported by the Belgian Science Policy Office, the Ministry of Economic Affairs, Administration of Mines and Belgian Geological Survey and the Management Unit of the Mathematical Model of the North Sea and Scheldt Estuary.

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5 Foraminiferal Biostratigraphy of the Palaeogene in the Southern Bight of the North Sea

W. WILLEMS †

Laboratory of Palaeontology, Geological Institute, State University of Gent, Krijgslaan 281/S8, B-9000 Gent, Belgium.

ABSTRACT

Biostratigraphical correlations, mainly based on foraminiferal faunas, are proposed for Palaeogene sediments in the Southern Bight of the North Sea. The data available for correlation were : the foraminiferal zonation scheme for the Dutch onshore Palaeogene deposits established by DOPPERT and NEELE (1983), the microfaunal zonation of the Eocene-Oligocene strata in the offshore IGS Borehole 79/06 proposed by HUGHES (1981) and the succession of foraminiferal assemblages in the Palaeogene in Northwest Belgium (Flanders area) compiled by WILLEMS (in press). The correlation between the Middle Oligocene series in the IGS borehole section and those in the Dutch and Flanders areas respectively is only general as the biostratigraphy of these series is worked out in more detail in the offshore than in the onshore region. The reverse is true for the Late and Middle Eocene sequences. The correlation between the Dutch and the Flanders onshore sections is fairly well possible except for the Late and latest Eocene and earliest Oligocene.

INTRODUCTION

Foraminiferids are unicellular micro-organisms (protozoans), living in an aquatic, generally marine environment, ranging from intertidal to abyssal milieus with the central to outer shelf zone as the most suitable places for species diversity. Most of the species known, fossil and living, are benthonic (mobile or sessile). The globigerinids have a planktonic way of life, occur in shelf seas, but have their greatest development in the surface layers (0-100 m) of the oceans.

Biostratigraphy stands for the study of the fossil content from successive geologic strata. Theoretically every geologic unit could be characterized by its fossil record ; even azoic strata may be identified by the underlying or overlying fossiliferous ones. When one or more fossils turn out to be characteristic for a certain geologic sequence these index-fossils may be used to identify these strata on a regional or even interregional scale, and for time-correlation as the evolution of species and the changes in the fossil record are time-bound and not reversible.

Published data on the biostratigraphy of the Tertiary deposits in the area considered by the Colloquy (51°00'N - 52°30'N) are rather scanty. In 1972 BIGNOT discusses the microfaunal content of Early Eocene (Ypresian-Sparnacian) layers recorded from cored rock at three localities NW of Dunkerque, in the most southern part of the area considered, around 51°11'N and 51°12'N. In 1981 HUGHES publishes the results of his investigation of the Oligocene-Eocene succession encountered in the IGS Borehole 79/06 and in 1984, CAMERON *et al.* present the data on the microfaunas from the Pleistocene-Pliocene sequence occurring in three IGS boreholes (81/51, 81/50 and 81/50A) ; all four IGS boreholes are located in the northern part of the area considered, between 52°08'N and 52°30'N (figure 1).

The purpose of this paper is to compare and to correlate the microfaunas recorded in the IGS Borehole 79/06 with those known from the Flemish onshore sections in order to contribute to the interpretation of the geologic history of the area considered and to promote interest for biostratigraphic analysis. The

Palaeogene biozonation for the Dutch onshore section proposed by DOPPERT and NEELE (1983) has also been examined. This zonation, just like the succession of foraminiferal assemblages in Flanders, is projected against the calcareous nannofossil zonation accepted by the IGCP project 124 for biostratigraphical charts of the Tertiary of Northwest Europe.

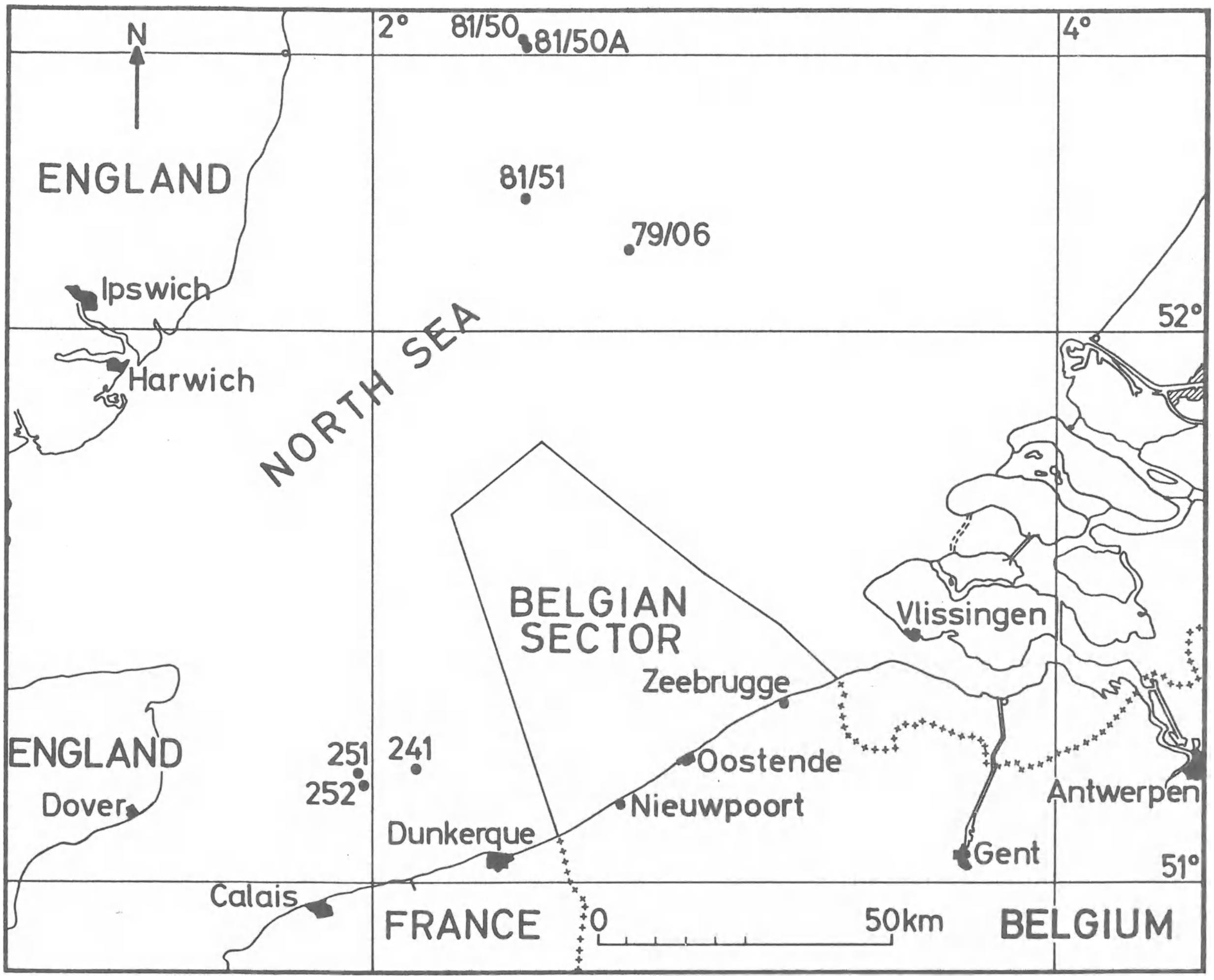

Figure 1 Localization map.

DISCUSSION

The data available for a correlation between the offshore IGS borehole section and the Flanders onshore area were respectively : the microfaunal analysis of the Middle Oligocene-Middle Eocene strata proposed by HUGHES (1981) and the succession of foraminiferal assemblages from the Flemish Palaeogene sediments, compiled by WILLEMS (in press) and which is mainly based on data published by BATJES (1958) (Oligocene) ; DROOGER (1969) (transition Oligocene-Eocene) ; KAASSCHIETER (1961) (Eocene) ; WILLEMS (1980) (Early Eocene) and MOORKENS (1982) (Palaeocene).

In both sequences Pleistocene surface layers occur and Neogene sediments are missing. The first correlation line, in down hole direction, runs between the upper part of the F-zone in Borehole 79/06 and the youngest part of the Boom Member Assemblage. It marks the top of the Late Rupelian and is based on the youngest occurrence of *Rotaliatina bulimoides, Svratkina perlata* and *Angulogerina gracilis*. In Flanders

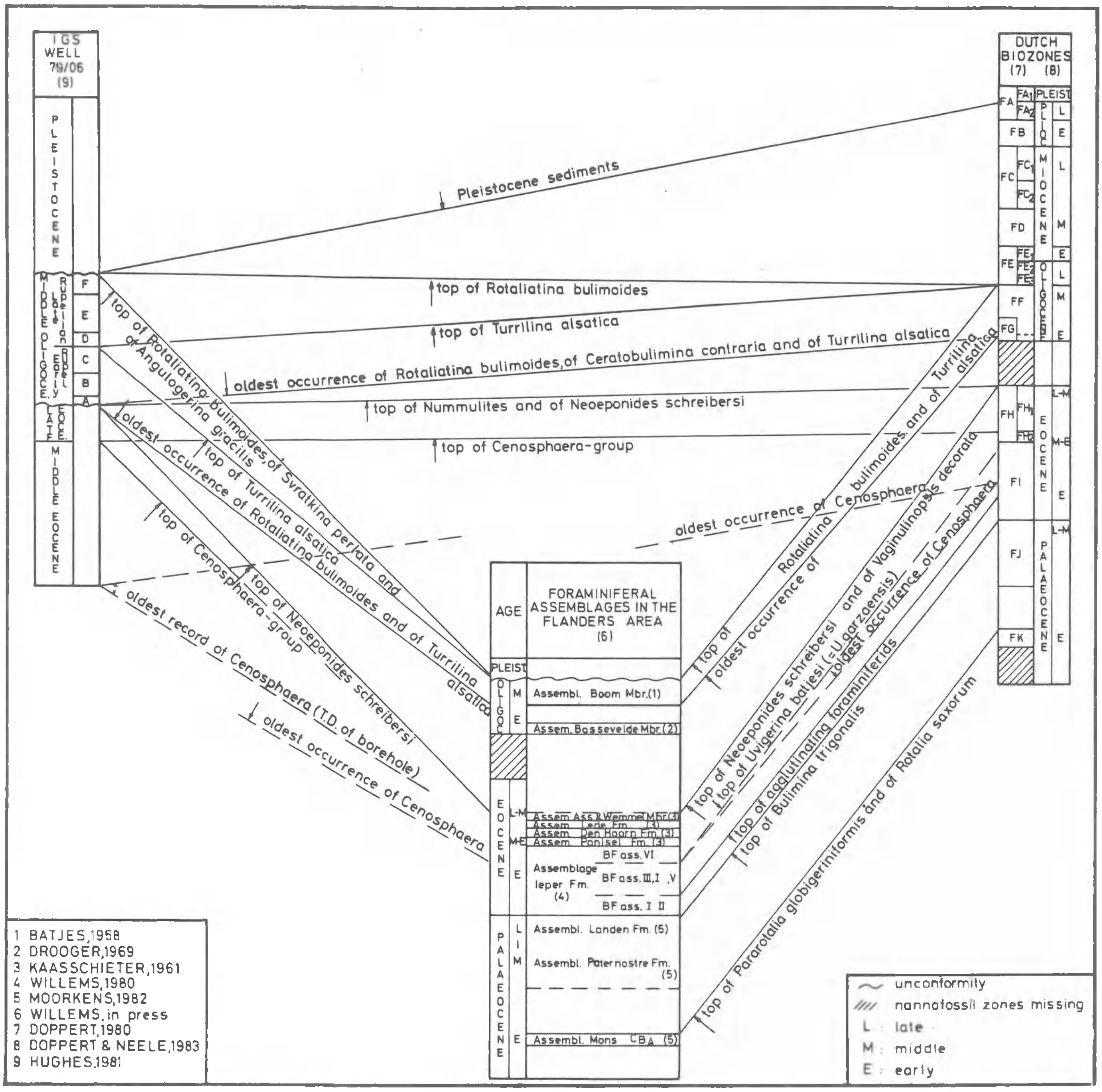

Figure 2 Foraminiferal biostratigraphic correlation table.

these marker species are accompanied by *Turrilina alsatica* (second correlation line) which ranges in the IGS borehole only from the top of zone C to the base of zone A (Early Rupelian). The range of *A. gracilis* is limited to zone F in IGS Borehole 79/06 and comprises the whole Boom Member in the Flanders area. According to HUGHES (1981) the foraminiferal assemblage of the western part of the Boom Member outcrop area (Flanders) is comparable to the one in zones A-C, while those from the eastern part (Antwerp-Kempen) reflect more affinities with the ones from zones E-F. Only the occurrence of *A. gracilis*, restricted to zone F and ranging throughout the Boom Member, disturbs the image. The third correlation line is based on the oldest occurrence of *R. bulimoides* and of *T. alsatica* marking the oldest Middle Oligocene (Early Rupelian) sediments.

The next correlation line is based on the top of *Neoeponides schreibersi* which marks the Late Eocene in the IGS borehole and is known in the Flanders area from the Asse-Wemmel Members Assemblages which are supposed to be of Middle Eocene age. This difference in age may be due to be absence of *N. schreibersi* in the Late Eocene in Flanders or by the presence in the IGS borehole of a hiatus covering more than the Early Oligocene. It is also possible that *N. schreibersi* or other typical Late Eocene microfossils occur above the Asse-Wemmel Members, in the so-called "Sand-Clay Complex of Kallo" as erected by GULINCK (1969) or the Meetjesland Formation as named by JACOBS (1978) but micropalaeontological research in Late Eocene-Early Oligocene is lacking.

The following marker forms in the IGS borehole are radiolarians of the *Cenosphaera* group, their youngest record coinciding approximately with the top of the Middle Eocene, and occurring in that well till terminal depth. Radiolarians are fully marine, planktonic protozoans with a silicious skeleton. The total range of that group is not yet known in Flanders, but its oldest occurrence so far recorded is the upper part of the Ieper Formation (Egem Member) (WILLEMS, 1981).

The data available for the correlation between the offshore IGS borehole 79/06 and the Dutch onshore Palaeogene sequence were respectively : the information provided by HUGHES (1981) on the Middle Oligocene-Middle Eocene strata and the foraminiferal letter zonation presented by DOPPERT and NEELE (1983) for the Dutch southern, central and northern Palaeogene onshore sediments.

Here also Pleistocene strata are overlying Middle Oligocene sediments in both sequences (first correlation line in down hole direction). In the Netherlands *R. bulimoides* and *T. alsatica* occur together while in the IGS borehole the range of the latter species is restricted to the older part of the Middle Oligocene sequence (second and third correlation lines respectively). The fourth correlation line marks the oldest occurrence of Middle Oligocene species *R. bulimoides, T. alsatica* and *Ceratobulimina contraria*).

The next correlation line links sediments of Late Eocene age, and is based on the youngest record of *N. schreibersi* and on the occurrence of fragments of *Nummulites*. As in both sequences the existence of a biostratigraphical gap between the Oligocene and the Eocene is accepted, it is not certain whether the total range of *N. schreibersi* is present.

The last correlation is based on the top of the *Cenosphaera* radiolarians range, indicating the top of Middle Eocene sediments. The total range of these radiolarians runs from the Middle Eocene to the late Early Eocene. In Borehole 79/06 these forms occur till terminal depth and the presence of late Early Eocene sediments in the sequence cut by the drilling can not yet be excluded definitely.

The foraminiferal zonation proposed by DOPPERT and NEELE (1983) is based on the examination by these authors of Palaeogene sediments encountered in several boreholes in the Netherlands. The succession of the foraminiferal assemblages in the Palaeogene sediments in Flanders, compiled by WILLEMS (in press) is based on literature data published by several authors from exposures and some boreholes.

Correlation is established :

1. between the Middle Oligocene FF-zone and the assemblage from the Boom Member, based on the range of *T. alsatica* and of *R. bulimoides* ;
2. between the top of the Late-Middle Eocene FH-zone and the assemblages from the Asse-Wemmel Members, based on the youngest occurrence of *N. schreibersi* and of *Vaginulinopsis decorata* ;
3. within the Early Eocene FI-zone and the assemblage from the Ieper Formation, based on the top of *Uvigerina batjesi* (= *U. garzaensis*), the oldest occurrence of *Cenosphaera* radiolarians and the presence of an agglutinating foraminiferid fauna ;
4. between the Palaeocene FJ-zone and the assemblage from the Landen Formation based on the presence of *Bulimina trigonalis* ;
5. between the Early Palaeocene FK-zone and the assemblages from the Mons Formation based on the occurrence of *Pararotalia globigeriniformis* and of *Rotalia saxorum*.

So far, no distinct microfaunas have been observed in the Dutch foraminiferal succession, similar to the Early Oligocene-Late Eocene (?) Bassevelde Member Assemblage and the Middle-Early Eocene Lede, Den Hoorn and Brussel Formations Assemblages known from the Flanders area.

CONCLUSIONS

Correlation of the Palaeogene sediments from the Dutch and Flanders onshore sections with those from the offshore sequence recorded in the IGS Borehole 79/06 is only general as the biostratigraphical results from the Middle Oligocene sediments in the offshore section are more detailed than those from the onshore sequences. The Middle Eocene sediments observed in the offshore possess a more uniform microfaunal assemblage than those from the onshore ; that also limits the possibilities for correlation.

Correlation between both onshore sequences is better except for the earliest Oligocene and the latest Eocene. Furthermore Middle Eocene microfaunas seem more diversified in the Flanders area than they are in the Dutch onshore.

Biostratigraphy can be used as a quick method in age-dating of the sediments encountered in future boreholes drilled in the Belgian sector of the southern North Sea. Meanwhile it would be of great interest, however, to have more detailed biostratigraphical work done on the Palaeogene sediments from the Flanders onshore area, using all suitable micropalaeontological disciplines.

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6 Tertiary phosphorites in the southern North Sea Basin : Origin, evolution and stratigraphic correlation

P.S. BALSON

British Geological Survey, Nicker Hill, Keyworth, Nottingham NG12 5GG, United Kingdom.

ABSTRACT

Sedimentary phosphate or phosphorite deposits are found at numerous horizons in the Tertiary sequences of the countries around the southern North Sea. Relatively little is known of the extent of these deposits in the offshore area.

These phosphorites are of two main types :
1. *Authigenic carbonate fluorapatite (francolite) concretions found dispersed at certain horizons within marine muds or muddy sands.*
2. *Conglomeratic lag or remanié deposits consisting of reworked and abraded phosphorite concretions and other phosphatic debris such as vertebrate teeth and bones, developed on prominent unconformity surfaces.*

The phosphogenic episodes represented by the occurrences of authigenic concretions may be correlated within the limited area of the southern North Sea, although the numerous stratigraphic gaps in the Tertiary sequences of the area hamper such correlation. The authigenic phosphorite occurrences correspond with periods of marine transgression.

Remanié phosphorite deposits are developed by the winnowing of such authigenic phosphorite-bearing formations, although some reworking of earlier remanié deposits also occurs. These deposits thus correspond to periods of non-deposition and stratigraphic hiatus. The remanié deposits are most conspicuously developed in the immediate vicinity of the source formation, the transportation of phosphorite material probably being restricted by the size of the concretions and the relatively high specific gravity of francolite. The concentration of phosphatic material in some of these remanié deposits is sufficient to have attracted commercial exploitation in the past.

Authigenic phosphorite is best developed in the Early-Middle Eocene and Miocene-Early Pliocene in the southern North Sea; periods of eustatic sea level rise, warm humid climates and global phosphogenesis.

INTRODUCTION

Sedimentary phosphate deposits or phosphorites are present in many sedimentary sequences throughout the world. Their economic importance as a source of phosphate for production of fertilisers and chemicals has meant that they have been the subject of many studies. The term 'phosphorite' encompasses a variety of lithological types of sediment related only by their high phosphate content contained within the mineral carbonate fluorapatite (francolite). Phosphorite deposits may consist of phosphatic pellets, phosphatised mud or limestone, phosphatic ooliths, phosphatic nodules, phosphatised teeth and bones, phosphatised excrement (coprolites), or a combination of these or other grain types, and usually, but not always, occur in marine formations. There is also a distinction between 'authigenic' phosphorite which results from the precipitation of apatite within primary pore space within the sediment and 'diagenetic' phosphorite which results from the secondary replacement of other material such as limestones. Not all phosphorite deposits form as a response to the same criteria in what is now known to be a complex set of interrelated factors (RIGGS, 1979; ARTHUR and JENKYNS, 1981; COOK, 1984). This paper concentrates on two types of sedimentary phosphorite : authigenic phosphorite concretions occurring mostly in mudstones or muddy sandstones and remanié phosphorite deposits formed as lag pebble deposits usually on unconformity surfaces by the winnowing and reworking of formations containing authigenic phosphorite concretions and other phosphatic components. Both types frequently occur within the Tertiary sequence of the southern North Sea basin.

Evidence for this paper has largely been obtained from studies of onshore outcrops around the periphery of the southern North Sea although some evidence from recent BGS offshore surveys is included.

EASTERN ENGLAND

a) London Clay Formation

The London Clay Formation (Late Palaeocene - Early Eocene) of eastern England comprises a sequence of marine silty clays, clayey and sandy silts, with subordinate sands which attains a maximum thickness of

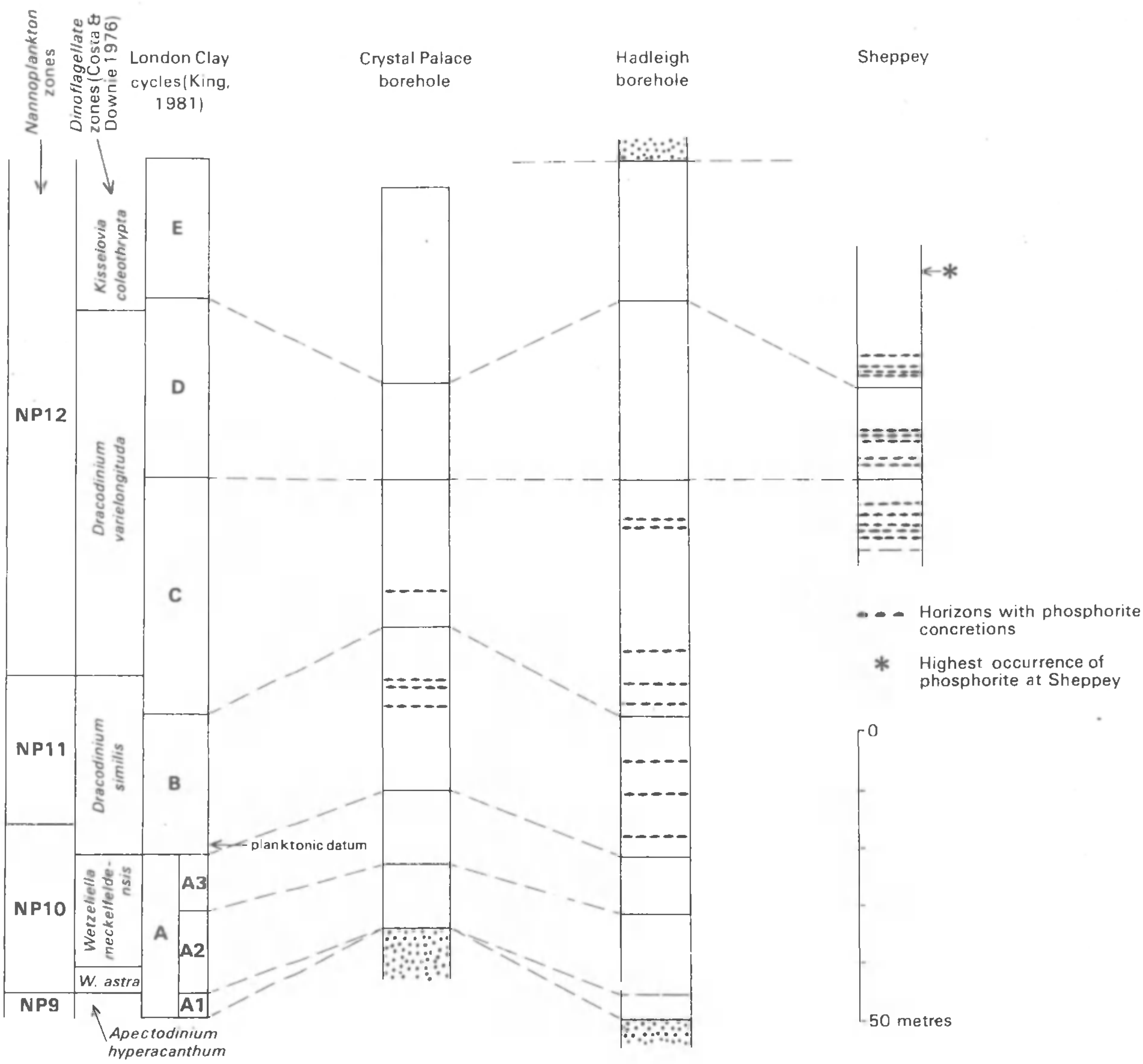

Figure 1 Correlation of BGS boreholes at Crystal Palace and Hadleigh, and cliff exposures at Sheppey (see figure 2 for locations) with palaeontological zones and with the transgressive cycles of KING (1981).

over 150 m in South Essex. The London Clay was deposited during an extensive marine transgression which also deposited the Ieper (Ypres) Clay in Belgium. KING (1981) divided the London Clay into a series of five sedimentary units designated A to E in ascending order representing individual minor transgressive cycles within the overall transgression (figure 1). Authigenic phosphorite mudstone concretions are found at several horizons within the London Clay.

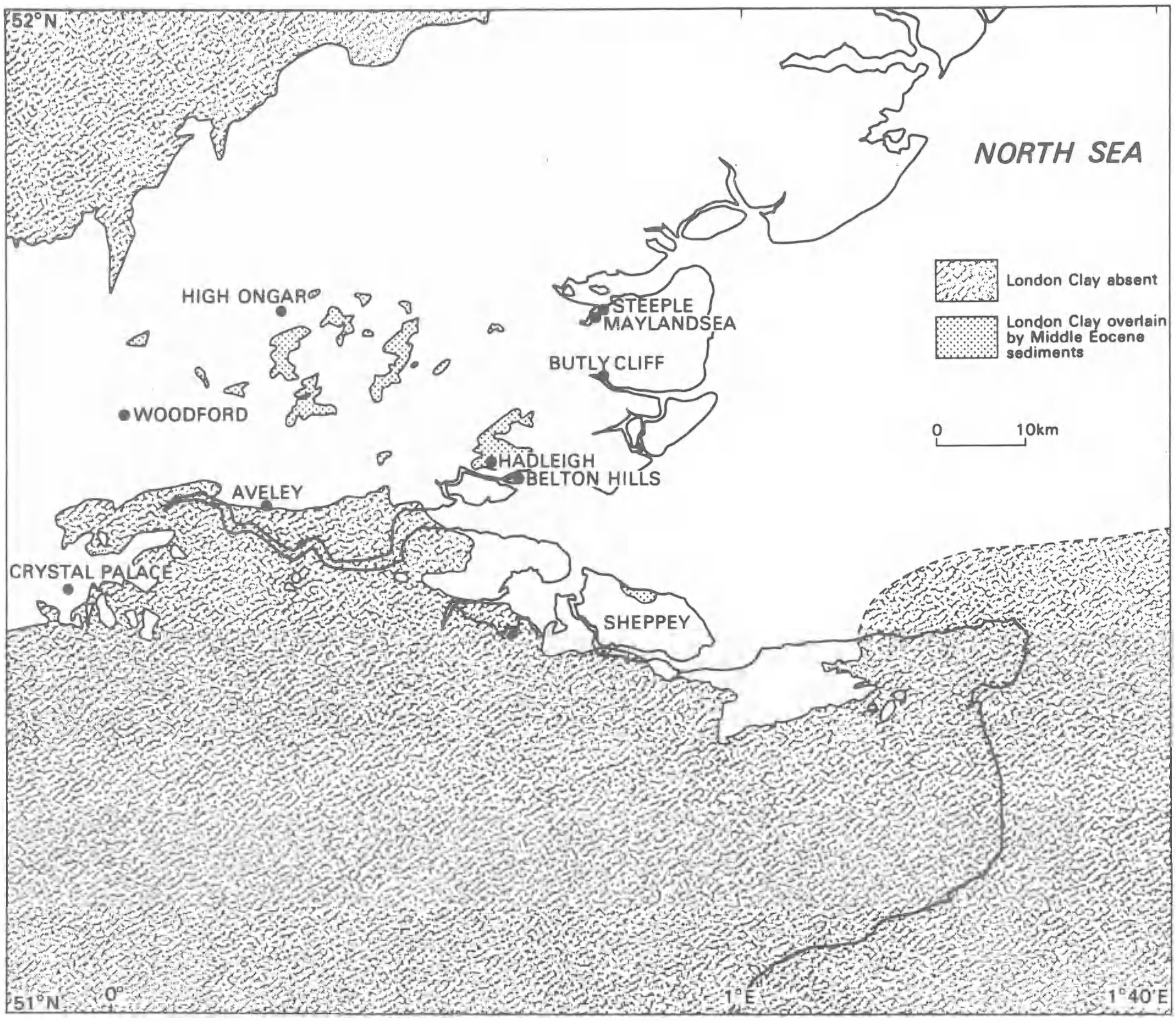

Figure 2 London Clay phosphorite localities in eastern England (see also table 1).

The occurrence of these concretions can be related to KING's units for the localities where details are available. The classic exposure at Sheppey has recently been re-examined and a lithological sequence published showing phosphorite concretion occurrences (KING, 1984). Boreholes drilled by the British Geological Survey which cored complete London Clay sequences have been re-examined and phosphorite concretion occurrences placed according to KING's units. These two sources provide the best evidence of the vertical disposition of the phosphorite concretions within the London Clay although the narrow diameter of the borehole cores (c. 90-115 mm) will inevitably mean some phosphorite levels being missed. A correlation of the sequence on the Isle of Sheppey with BGS boreholes at Crystal Palace and Hadleigh is shown in figure 1. Other occurrences at field exposures of London Clay in Essex are recorded in the literature and are summarised here in table 1. The London Clay localities at which phosphorite concretions are known to occur are shown in figure 2.

Locality	Author(s)	London Clay Division (KING, 1981)
Steeple	George & Vincent, 1977	B
Maylandsea	George & Vincent, 1982	B
Aveley	Kirby, 1974	B*
High Ongar	King, 1981	C
Woodford	King & King, 1976	C
Belton Hills	Hewitt, 1982	D
Butly Cliff, Burnham	Hewitt, 1982	D
(* C. King pers. comm. 1984)		

Table 1 Recorded occurrences of London Clay phosphorite concretions from Essex field localities.

The phosphorite concretions are concentrated within certain parts of the sequence. As far as is known phosphorite concretions are absent from division A, are commonest in B and C but are found at higher levels at localities believed to have been nearer the centre of the London Clay depositional basin such as at Sheppey where they may occur well into division E (Personal communication,C. KING, 1984). This distribution pattern is especially interesting. At the base of Division B is a horizon known as the "planktonic datum" which represents the first appearance of common planktonic foraminiferids in the English Early Eocene. Phosphorite concretions appear to be absent below but are found almost immediately above this datum suggesting this transgressive event was significant in the development of phosphorite concretions in the London Clay. The datum may represent the onset of sedimentation related to a more open connection with oceanic waters with increased of dissolved phosphorus compared with the relatively restricted marine basin represented by Division A. The London Clay transgression attains its maximum extent within Divisions B and C before a gradual regression.

The concretions have been described by BALSON (1980) and HEWITT (1982) and are generally buff-coloured, spherical to ovoid or flattened and usually between 3 and 8 cm across. The buff-coloured cortex usually surrounds a darker and harder core in which septarian fissures filled with pyrite are common (figure 3A). The darker colour of the core is due to a greater concentration of organic material within the centre of concretion. Usually the concretions have no obvious nucleus although many may have formed around burrow fills. Occasionally the concretions formed around and within the carapaces of decapod crustaceans for which the London Clay is well known (figure 3B). Indeed the preservation of a diverse and well preserved crustacean fauna is largely due to this phenomenon which has prevented the carapaces from being crushed. Other nuclei include vertebrate material such as fish vertebrae, or occasionally the roots of shark teeth (figure 3C). Phosphatised wood is occasionally found although most plant material in the London Clay is preserved by pyrite or calcite. The concretions only rarely enclose molluscan material.

b) Suffolk 'Crags'

At the base of the Crags (Pliocene and ?Pleistocene marine sands) of Suffolk is a conspicuous conglomeratic bed resting on an undulose unconformity with the underlying London Clay Formation. The pebbly material is intermixed with the lowermost Crag sediments although occasional scattered pebbles are found at other levels throughout these formations. A prominent component of these deposits are pebbles of phosphatic mudstone which are believed to have been derived from the London Clay. These pebbles are usually dark brown with a polished surface occasionally showing pitting due to marine boring organisms. The surfaces of the pebbles occasionally show thin cracks which in section are seen to be the surface manifestation of internal angular septarian fissures (figure 3D). The fissures are often empty but are more usually filled with limonite. Rarely, some pyrite may be preserved. These cracks are comparable with the pyrite-filled septarian fissures seen in the core of London Clay phosphorite concretions. After derivation from the London Clay the softer unfissured cortex of the concretion would have been removed by abrasion. The fissures would then be exposed on the surface of the abraded concretion allowing oxidation of the

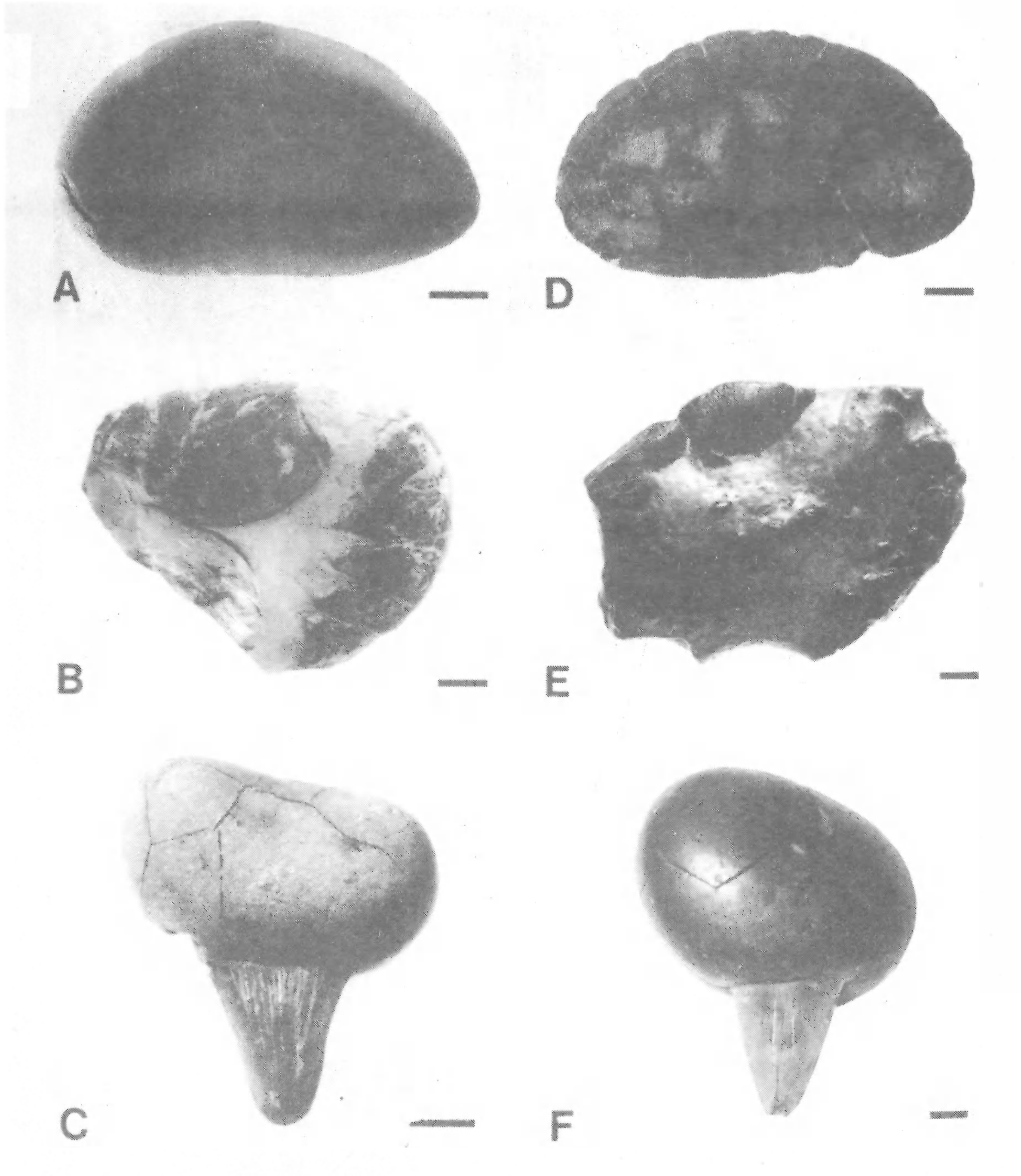

Figure 3 Phosphorite concretions from the London Clay.

A = Section through phosphorite concretion showing pyrite filled septarian cracks. Darker core due to organic pigment. Sheppey, Kent. Author's collection. Scale bar = 1 cm.

B = Phosphorite concretion enclosing lobster carapace. *Homarus gammaroides*. Sheppey, Kent. GSM 101233. Scale bar = 1 cm.

C = Phosphorite concretion around root of shark tooth. *Lamna obliqua*. Sheppey, Kent. GSM 101298. Scale bar = 1 cm.

Reworked phosphorite concretions from the Crag phosphorite deposit.

D = Section through phosphorite concretion showing limonite filled septarian cracks. Bawdsey, Suffolk. Author's collection. Scale bar = 1 cm.

E = Phosphorite concretion enclosing lobster carapace. nr Woodbridge, Suffolk. GSM 100975. Scale bar = 1 cm.

F = Phosphorite concretion around root of shark tooth. *Lamna obliqua*. Felixstowe, Suffolk. SM C45380. Scale bar = 1 cm.

GSM = Geological Survey Museum (BGS)
SM = Sedgwick Museum, Cambridge

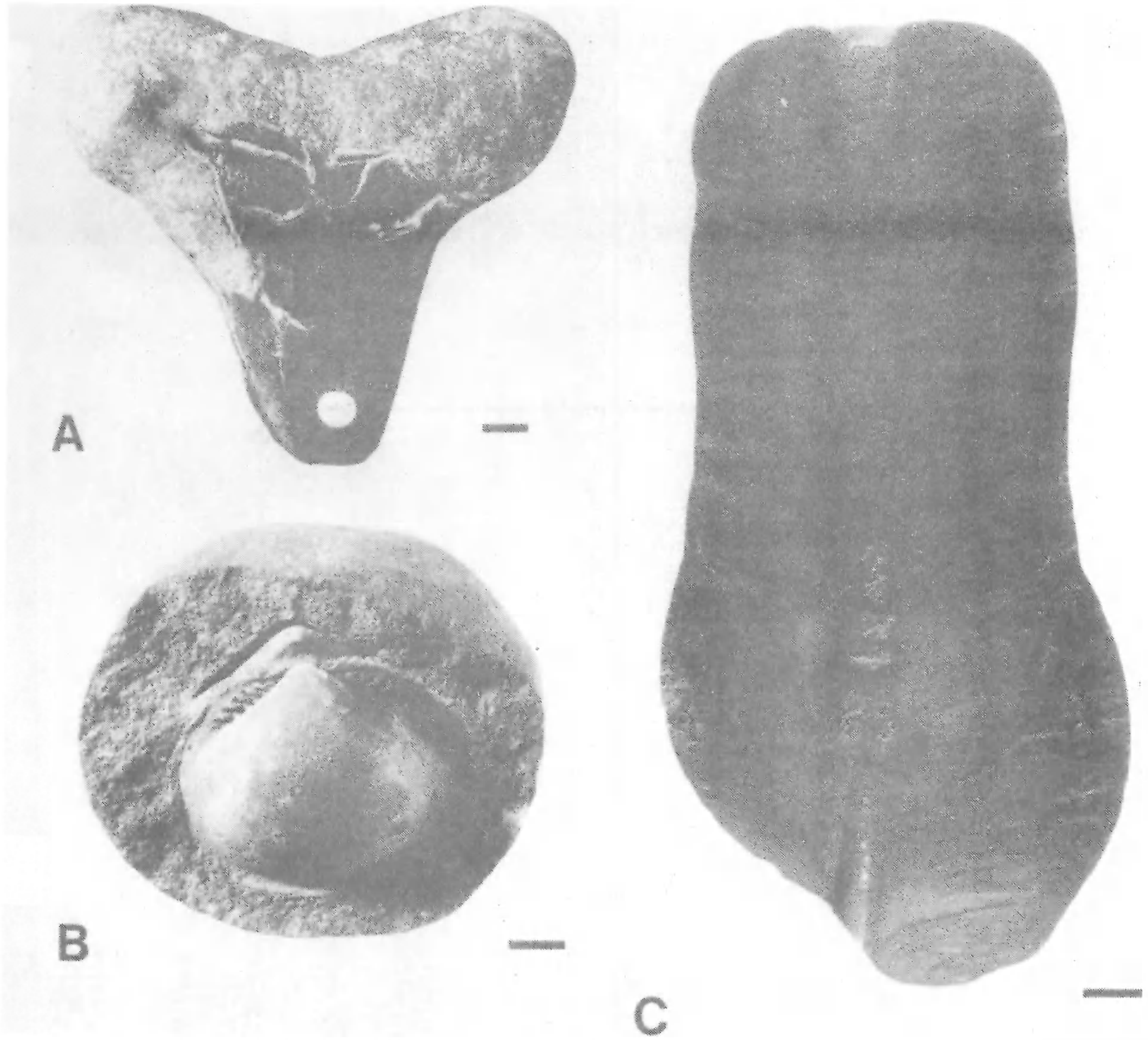

Figure 4 Ex-Miocene fossils from the Crag phosphorite deposit.

A = Phosphorite cemented sandstone around root of shark tooth. *Carcharodon megalodon.* BMNH P47380. Scale bar = 1 cm.

B = Phosphorite concretion around molluscan shell mould. *Glycymeris.* Suffolk. SM C48553. Scale bar = 1 cm.

C = Vertical section through concretion around burrow. *Tasselia.* GSM 118122. Scale bar = 1 cm.

BMNH = British Museum (National History)
SM = Sedgwick Museum, Cambridge
GSM = Geological Survey Museum (BGS)

pyrite fills. In some cases the concretion would remain held together as in figure 3D, but often the concretion would simply fall apart as angular phosphatic fragments which became rounded by subsequent abrasion. Small, smooth, sub-angular phosphatic pebbles are common in the Crag phosphorite deposits.

Some of the phosphatic nodules contain fossil material which can be identified with that found in situ in the London Clay. The derived fauna includes species of decapod crustacea (figures 3B and 3E), and shark teeth with nodules around the roots (figures 3C and 3F). The derived Eocene fauna also includes phosphatised shark and ray teeth which are not associated with phosphatic nodules.

A second phosphatic component of the Crag phosphorite deposits consists of pebbles or even large cobbles of a francolite-cemented sandstone. Cobbles of this sandstone may on breaking reveal a molluscan shell mould or other organic nuclei which led to them being termed 'box-stones'(figure 4B). Occasionally the sandstone is seen adhering to vertebrate material such as cetacean teeth or the teeth of sharks including *Carcharodon megalodon* (figure 4A). The box-stones are clearly phosphatic concretions which formed within the sediment and often centred on organic nuclei (figure 4B). The sediment preserved within these concretions consists of muddy sands or sands, in which the muddy matrix may comprise 20 to 80 % of the concretion. Some phosphatic concretions developed around burrows similar to *Tasselia ordam* described from the Merksem Sands (? Lower Pleistocene) of Antwerp (VAN TASSEL, 1964) and show an internal structure of alternating sand and mud laminae around a central vertical tube (figure 4C). This second phosphatic assemblage also includes abundant phosphatised cetacean bone and some phosphatised shark teeth without adherent matrix.

This second assemblage of phosphatic components is the only known remnant of a previously existing ?Middle Miocene deposit termed the 'Trimley Sands' by BALSON (in preparation). This formation probably consisted largely of muddy quartz sands with some areas of interlaminated sands and muds (?near-shore) indicated by the *Tasselia* concretions.

Some rare phosphatic material, such as internal moulds of ammonite chambers, is clearly Mesozoic in age.

1. Distribution of Crag phosphorite

The pebbles of phosphatic mudstone believed to be derived from a London Clay source are widely distributed within the basal Crag sediments. As already mentioned the phosphatic concretions are commonest within the central portion of the London Clay sequence so it is presumably winnowing of this part of the London Clay which has provided the derived nodules. When the distribution of the richest Crag phosphorite deposits is compared with the area where London Clay erosion is seen to have been greatest, i.e. where the least of the sequence remains, an interesting pattern can be seen. Figure 5 shows the highest preserved division of the London Clay at localities in eastern England. It will be noticed that Crag phosphorite deposits predominantly occupy an area where they immediately overlie division A and are not found where the London Clay sequence is largely complete. This supports the theory that in situ winnowing of London Clay sediments could result in the formation of lag deposits of denser components such as phosphorite concretions. Sedimentary phosphorite has a relatively high specific gravity of approximately 3.0 so would be readily concentrated in this way.

The distribution of 'box-stones' within the Crag phosphorite deposit is also interesting. They appear to be a common component only in a relatively small area centred on the coastal town of Felixstowe (figure 5). The area of greatest London Clay erosion and the distribution of 'box-stones' closely follow the line of an axis of Tertiary uplift proposed by BOSWELL (1915) which runs NW-SE approximately through Felixstowe.

Although it is not possible to say that all of the phosphatic mudstone nodules were derived from the Early Eocene, indeed there is evidence to suggest at least some being reworked from Mesozoic sources, it is likely that the vast majority were derived from the London Clay which underlies the Crag phosphorite deposits almost everywhere. Although the 'Trimley Sands' seem to have been largely sandy with subordinate mud bands, it is possible that some phosphatic mudstone nodules were derived from that formation rather than the London Clay. A source from other deposits now totally destroyed cannot be overlooked but as yet there is no evidence of such a source.

The Tertiary sequence of eastern England is summarised in figure 6. At least two phosphogenic episodes occurred during the Tertiary within this area. The first occurred in the central part of the London Clay (Early Eocene). This was followed after the early Middle Eocene by a long period of presumed non-deposition and erosion until a renewed transgression (?Middle Miocene) deposited muddy sands in an area centred on Felixstowe. The second phosphogenic episode during this transgression resulted in phosphatic concretions within the muddy sands. Another period of erosion and winnowing preceded Coralline Crag deposition during the early Pliocene which deposited carbonate-rich sands over this lag gravel to form a remanié phosphorite deposit. The subsequent deposition of the Red Crag followed much the same pattern of reworking of lag gravels into the lowermost sediments of that formation.

The eastern England phosphorites clearly show two phases of authigenic phosphorite development during transgressions in the early Eocene and ?Middle Miocene. During regressions winnowing of the enclosing sediment occurred and the dense phosphatic components were reworked more or less in situ to form

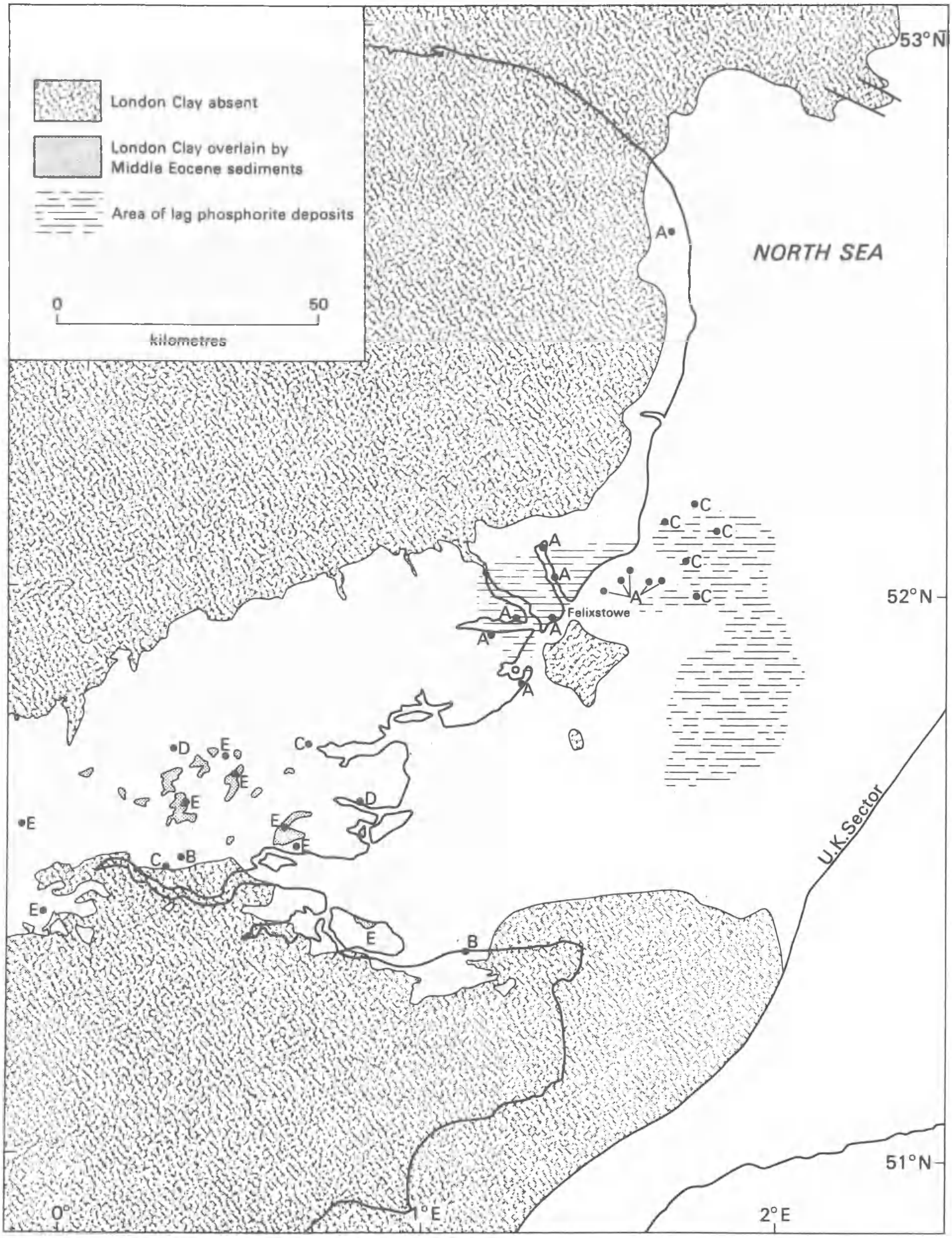

Figure 5 Geographical distribution of lag phosphorite deposits in eastern England. Onshore area is that of known rich deposits at the base of the Coralline and Red Crags. Offshore area is where there is a significant contribution by phosphorite pebbles to the modern sea bed sediments. Letters indicate uppermost of KING's (1981) cycles preserved at sites in the London Clay.

conglomeratic phosphorite deposits which were economically exploited for fertiliser manufacture during the 19th century (REID, 1890).

The sequence in eastern England can be compared with other Tertiary sequences in and around the southern North Sea.

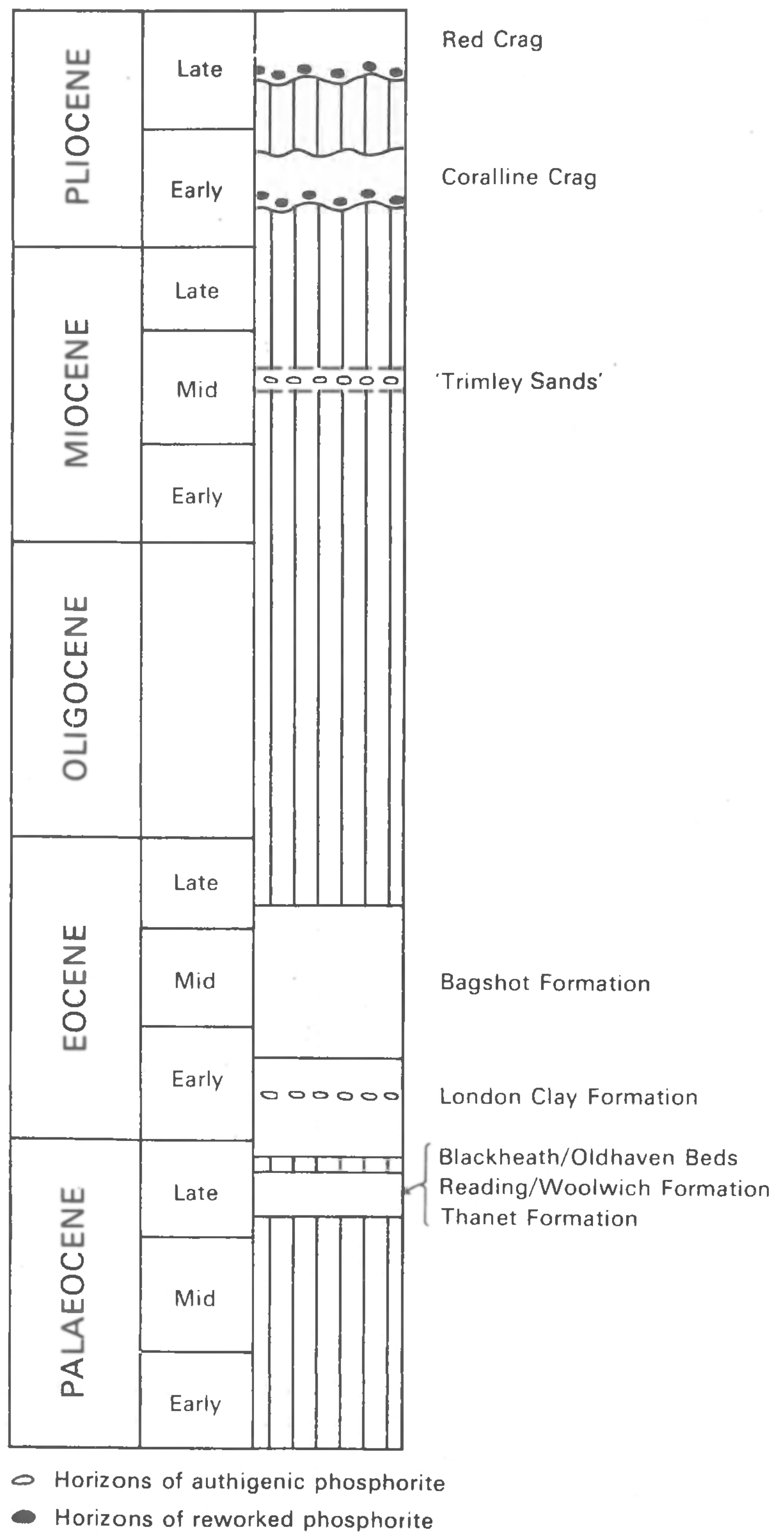

Figure 6 Phosphorite occurrences in the Tertiary of eastern England.

EASTERN NETHERLANDS

The stratigraphically lowest known phosphorite occurrence in the Tertiary of the eastern Netherlands are the in situ olive-grey phosphorite concretions in the middle part of the Brussel Sand or Marl Member (Middle Eocene) of Drente (figure 7) noted by VAN DEN BOSCH (1980) and BOR (1985). The concretions occur within a unit of calcareous, glauconitic sandstone with clayey and sandy intercalations. Reworked phosphorite concretions and shark teeth are found at the base of an overlying unit within this member. These concretions and teeth are believed to have been reworked from the in situ phosphorite below (BOR, 1985, p. 114). BOR (1985) also recorded 'small, black non-reworked phosphorite concretions' from a second borehole 56 km to the north at a similar level.

VAN DEN BOSCH *et al.* (1975) described the Tertiary sequence in neighbouring Twente (SE Overijssel) in the eastern Netherlands which includes a number of further phosphate-rich horizons. The early Rupelian (Early Oligocene) Ratum Member overlies Eocene sands and contains derived phosphorite nodules at the base. This phosphorite horizon is absent when overlying Mesozoic formations. These reworked phosphorite concretions contain a fauna which indicates the Brussel Sand/Marl Member as their source (VAN DEN BOSCH, 1980). POSTHUMUS (1923), in a description of the phosphatic nodules at the base of the Oligocene in Twente, mentioned nodules which formed around the roots of shark teeth. He also noted the similarity of the reworked fauna with Eocene faunas. PANNEKOEK (1956) noted the presence of Eocene nummulites and fish within the basal Oligocene phosphorites in East Gelderland and Overijssel. These basal Oligocene phosphorite deposits were once exploited economically (PANNEKOEK, 1956 ; DIETZ, 1960).

There appears to be no direct evidence of authigenic phosphorite development within the Oligocene of the eastern Netherlands but indirect evidence is found in the overlying Miocene sediments.

The Miste Bed (Aalten Member) of Middle Miocene age is a mainly sandy unit with a layer of abraded, hard black phosphorite nodules and phosphatised shark teeth at the base. Some of these nodules enclose moulds of molluscan shells which indicate Chattian (late Oligocene) age. A few metres above the base of the Miste Bed is a horizon of glauconitic phosphorite concretions believed to be in situ.

At the base of the sandy Delden Member (late Miocene to early Pliocene) is a deposit of reworked whale bones, shark teeth and phosphorite nodules. The vertebrate material is believed to be derived from underlying late Miocene sediments and the nodules from reworking of older parts of the Delden Member.

Within the Delden Member is a horizon of numerous phosphorite concretions. Some of these concretions formed around the roots of shark teeth. The phosphorite may also be in the form of continuous beds or slabs containing many moulds of molluscan shells. The phosphorite contains abundant goethite and glauconite. The phosphorite-bearing sediments attain a total thickness of 3.50 m. JANSSEN (1966) correlated the fauna of these phosphorite concretions with the Belgian sands of Deurne (late Miocene), but later (VAN DEN BOSCH *et al.*, 1975), he believed a correlation with the Kattendijk Sands (early Pliocene) to be more correct.

SOUTHWESTERN NETHERLANDS

HARSVELDT (1973) described the Tertiary sequence from a series of boreholes in the Zeeland area, southwestern Netherlands (figure 7). The sequence is very similar to that of nearby northern Belgium and contains a number of discontinuous phosphorite horizons. Further details of these phosphorite occurrences are given by HARSVELDT (1970) and VAN TOOR (1972). At Walcheren the lowest layer (Layer I in HARSVELDT, 1973) is about 5 m below the top of the Antwerp Sands (Middle Miocene) which here consists of glauconitic fine to medium-grained sands, with glauconite particularly abundant in the basal parts of the formation. The second layer (Layer II in HARSVELDT, 1973) is found at the base of the overlying Deurne Sands which here consist of fine to medium-grained glauconitic sands with intercalations of shell. Layer III occurs at the base of the Kattendijk Sands (Pliocene) which consists of medium to coarse-grained sands with abundant shell material.

The phosphorite nodules from these three layers are not distinguished in the descriptions and are described as consisting of glauconite and rounded quartz grains (with some subordinate feldspar) embedded in a brown cryptocrystalline apatite matrix in which apatite forms about 30-40 % of the nodule.

It seems probable that Layer I is in situ phosphorite within the Antwerp Sands, whilst II and III are lag deposits of similar material at the base of the later formations derived by winnowing of the Antwerp Sands.

Figure 7 Phosphorite localities in Holland, Belgium and northern France mentioned in the text.

ANTWERP - SINT-NIKLAAS AREA, BELGIUM

This area of northern Belgium is one of the classic areas of Neogene stratigraphy in the southern North Sea basin. A sequence from Middle Oligocene to Pliocene is preserved within the environs of Antwerp.

The oldest phosphorite deposit which is recognised in this area consists of phosphorite concretions within the uppermost part of the Berg Sands which immediately underlie the Boom Clay (Early Oligocene) at Sint-Niklaas (VANDENBERGHE, 1978). The concretions are found within a layer approximately 10 cm thick and consist of greenish-grey or brown, francolite cemented, muddy sand with a well sorted, fine sand fraction. The concretions are irregularly shaped and mostly between 3 and 8 cm diameter although some may be larger. Some of the concretions clearly formed around or within burrows or around bone fragments, whilst others contain molluscan shell moulds. VANDENBERGHE (1978) believes that the concretions may have formed within the top few centimetres of the sands during a period of non-deposition. The concretions may then have been slightly reworked causing a partial rounding before deposition of the Boom Clay. The molluscan fauna of the concretions has been equated with the Berg Sands (JANSSEN, 1981) although VAN DEN BOSCH (1981) concluded that the shark tooth assemblage of the phosphorite deposit was older than that of the Ratum Member in the Netherlands and therefore that these sands at Sint-Niklaas are not equivalent to the Berg Sands (s.s.).

At the base of the Edegem Sands (Early Miocene) is a conglomeratic horizon (Burcht Gravel) which according to DE MEUTER *et al.* (1976) contains dark small rounded flint pebbles, but also contains small, sandy phosphorite pebbles and reworked septaria fragments.

Within the Antwerp Sands (Anversian, Middle Miocene) there are levels with phosphatic concretions (DE MEUTER *et al.*, 1976 ; DE MEUTER and LAGA, 1976) within very glauconitic, moderately well, to well sorted fine to medium sands, associated with vertebrate bones and shark's teeth. These concretions are believed to be authigenic.

At Borgerhout, Antwerp and at Ramsel 32 km SE of Antwerp a thin basal gravel at the base of the Deurne (Diest) Sands (Late Miocene) contains reworked phosphorite concretions. These concretions and their enclosed fauna were described by JANSSEN and MÜLLER (1984). The concretions are very irregular in shape and show only slight evidence of transport. They are generally decalcified but may contain traces of aragonitic fossils. The fauna contained within the phosphorite concretions is of Middle Miocene age.

At the base of the Kattendijk Formation (Early Pliocene) is a conglomeratic remanié bed containing abundant phosphorite nodules many with enclosed molluscan shell moulds. They consist of francolite-cemented muddy sand with a moderately well sorted fine to medium sand fraction and are believed to have been derived from the Antwerp Sands. The nodules are associated with large fragments of phosphatised bone, some of which have adherent phosphorite-cemented sediment of the same type as the nodules.

The Kattendijk Formation itself contains abundant non-phosphatised cetacean bone which is contemporary with deposition.

WESTERN BELGIUM

The Ieper (Ypres) Clay Formation is stratigraphically equivalent to the London Clay of eastern England and is of Early Eocene (Ypresian) age. Phosphorite nodules from the Ieper Clay have been known for over a hundred years (e.g. DELVAUX, 1883). More recently they have been recorded in boreholes at Ooigem and Tielt towards the base of the Ypresian (GULINCK, 1967). It should be noted that the base of the Ypresian in Belgium is within division A2 of KING, the lowermost ash series (Harwich Member, A1) being absent (KING, 1981).

A summary of the bio- and lithostratigraphy of the Ieper Clay Formation can be found in WILLEMS (1982, table 1) (figure 8). In the Ieper Clay of western Belgium phosphorite concretions occur mostly in the Flanders Member but are occasionally found within the overlying Egem Member (S. GEETS, Personal communication, 1984) although it is not clear whether the latter occurrences are in situ. Phosphorite concretions are recorded by GULINCK between 60.5 and 141.5 m depth in the Tielt borehole and at -35.3 m and between 81.1 and 85.0 m depth at Ooigem. Unfortunately the lowermost recorded phosphorite concretion from -141.5 m in the Tielt borehole material held by the Belgian Geological Survey in Brussels has been lost. Sediment believed to have formed part of the cortex of this concretion has been proved by X-ray diffraction analysis to be siderite. The lowest confirmed phosphorite concretion is at -132.5 m. DE CONINCK (1975) examined the dinoflagellate flora from a sample from -125.0 m and correlates this (Personal communication, J. DE CONINCK, 1984) with the base of the *similis* zone of COSTA and DOWNIE

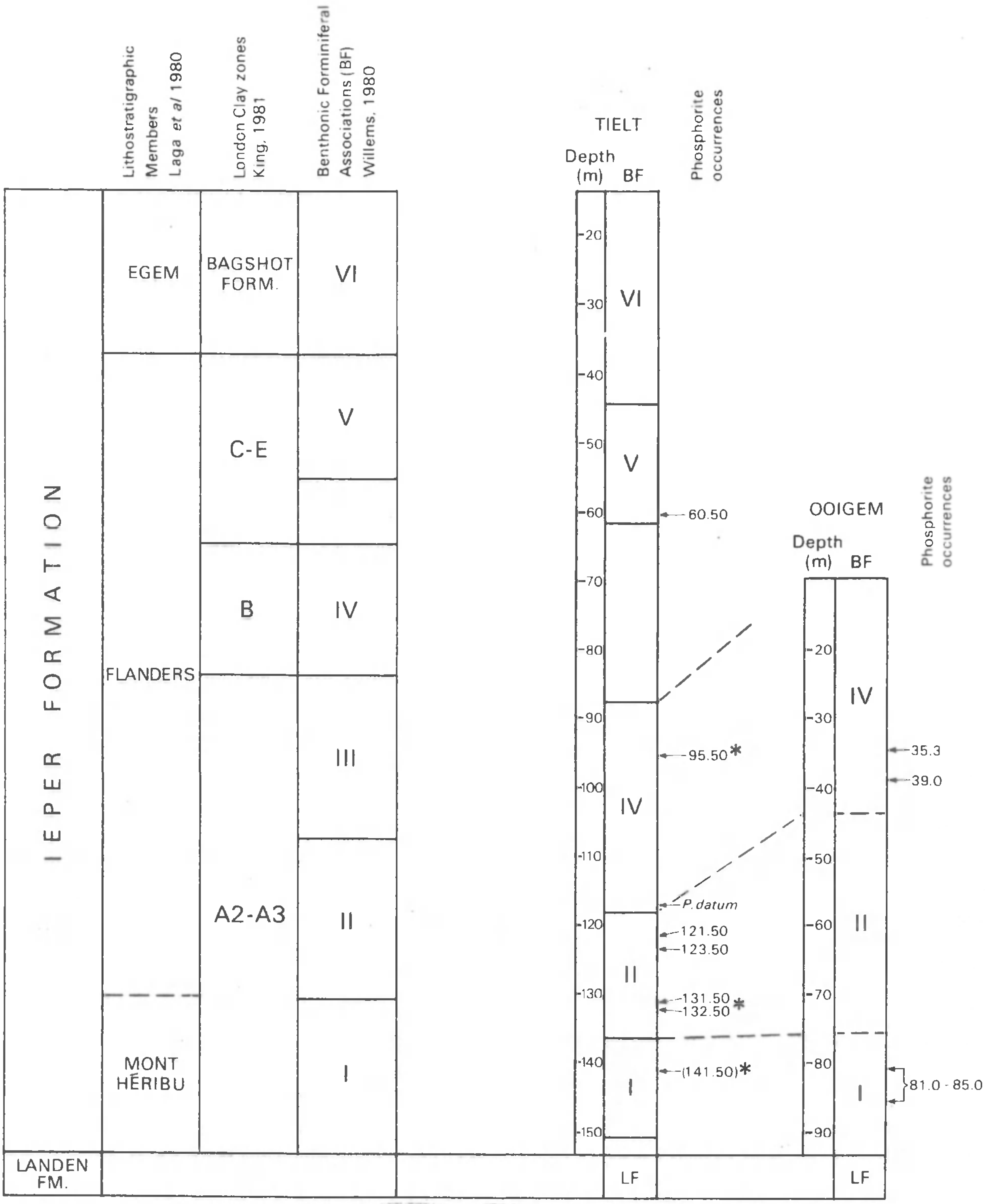

Figure 8 Correlation of phosphorite occurrences in the Ieper Formation of western Belgium. Stratigraphic correlation after WILLEMS (1980 ; 1982). Phosphorite occurrences from borehole records of the Belgian Geological Survey. Asterisks indicate samples analysed for this study.

(1976). From this evidence it would appear that the earliest confirmed phosphorite occurrence in the Tielt borehole probably occurs within the *meckelfeldensis* dinoflagellate zone which correlates approximately with the upper part of KING's division A in the London Clay. A sample studied by DE CONINCK (1975) from -151.5 m is placed within the *astra* zone (Personal communication, J. DE CONINCK, 1984). On the basis of foraminiferal evidence WILLEMS (1980) placed the planktonic datum at Tielt at a depth of 117.5 m. The evidence therefore suggests that the phosphorite occurrence at -132.5 m at Tielt occurs beneath the planktonic datum and within sediments which correlate with London Clay division A3 of KING (1981) and is therefore earlier than any recorded occurrence in eastern England. In situ authigenic phosphorite concretions also occur below the planktonic datum at Wardrecques in northern France (Personal communication, C. KING, 1984) and in northern West Germany although generally they seem to occur in greatest abundance only above this horizon.

WEST GERMANY

Phosphorite concretions and deposits occur at several horizons within the Tertiary sequence of West Germany (PAPROTH and ZIMMERLE, 1980). It is generally not clear from published descriptions which of the phosphorite occurrences are authigenic and which reworked but probable in situ authigenic phosphorite concretions occur in the Early Eocene 'Tarras' Clay of northern West Germany which is stratigraphically equivalent to part of the London and Ieper Clay Formations. The 'Tarras' (s.s.) is a non-calcareous clay unit forming part of the 'Eocene 3' Clays and corresponds to part of KING's division E although the term has probably been used more widely and arbitrarily (Personal communication, C. KING, 1985). 'Solitary phosphorite nodules' are recorded from the 'Eocene 1, tuffitserie' (which correlates with KING's Division A1) even within tuffaceous material according to WIRTZ (1939, p. 265) but the majority are concentrated in the 'Tarras' Clay. This observation would imply that some phosphorite formation occurred here at a level which predates the planktonic datum. The 'Tarras' concretions have a variable composition ranging from phosphorite to siderite with intermediate compositions between. It is interesting to note that in the 'Tarras' Clay phosphorite nodules occur in association with siderite nodules in the lowermost parts of the sequence (WIRTZ, 1939, figure 15), with phosphorite nodules occurring alone higher up. As noted in the previous section the Ieper Clay, Tielt borehole sample from - 141.5 m proved to be siderite although recorded in the borehole log as phosphorite. It may prove that a comparable situation to the 'Tarras' Clay may exist in the Ieper Clay with concretions of mixed composition or with pure siderite concretions in the lowest parts of the sequence.

PAPROTH and ZIMMERLE (1980) described the phosphorite nodules as being a few centimetres in size (up to 'fist size') and to be of varying shape but mostly oval. They generally have a light coloured exterior with a darker core caused by organic pigment. The nodules contain an irregular polygonal system of septarian cracks which are filled mostly with pyrite, barite or sparry calcite. This description is thus very similar to that of the London Clay phosphorite nodules although the septarian cracks are generally filled only with pyrite in the latter example.

Other local occurrences recorded include reworked phosphorite material at the base of the Rupelian (Early Oligocene) (DIETZ, 1960) and Burdigalian (Early Miocene) (NATHAN, 1949). GRAMANN and MUTTERLOSE (1975) and GRAMANN *et al.* (1975) described phosphatic concretions from the top of the Late Eocene which may contain carapaces of crustaceans (cf. London Clay). Further work is necessary to identify which of the West German Tertiary phosphorites are authigenic before any detailed comments on correlation can be made.

OFFSHORE SOUTHERN NORTH SEA

Very little information on phosphorite distribution is available for the offshore area of the southern North Sea. Borehole 81/50A drilled by the British Geological Survey recovered some small abraded phosphatic mudstone pebbles at the base of the Red Crag Formation (Pliocene) where it rests unconformably on Middle Eocene clays. It is interesting to note that despite the small core diameter (50 mm) several dozen small pebbles were recovered showing that the concentration of phosphatic material is high. In an area offshore of Southeast Suffolk and eastern Essex phosphatic mudstone pebbles are a common component of the gravel fraction of the modern sea bed sediments, but appear to be restricted to areas where Early Eocene sediments are exposed on the sea bed (figure 5). No confirmed phosphatic material derived from the 'Trimley Sands' has yet been found. In areas where Crag sediments underlie the modern sea bed sediments, phosphatic pebbles are apparently absent. The distribution and abundance of this offshore phosphorite deposit will be described in greater detail in a future publication.

Phosphorite correlation in the area around the southern North Sea

Phosphorite deposits are often associated with unconformities or indicators of reduced deposition. It is however important to distinguish between remanié (reworked) phosphorite deposits and in situ (authigenic) phosphorite occurrences. Phosphatic material readily forms concentrates or lag deposits due to the relative hardness (5) and the high specific gravity of francolite (2.9 - 3.1). The deposits in eastern England clearly show the correlation between previously economic phosphorite deposits and areas where the main source rock (in this case the London Clay) has been extensively eroded. Thus in this case phosphogenesis is not related to the period of unconformity, but reworking during these periods has concentrated older phosphatic material into a phosphorite deposit. Phosphatisation of limestone hardgrounds is known to occur during periods of reduced deposition but in the southern North Sea basin authigenic phosphorite, and thus phosphogenesis, occurs within transgressive depositional sequences. It is not yet known whether there is any connection between phosphorite concretion development and brief

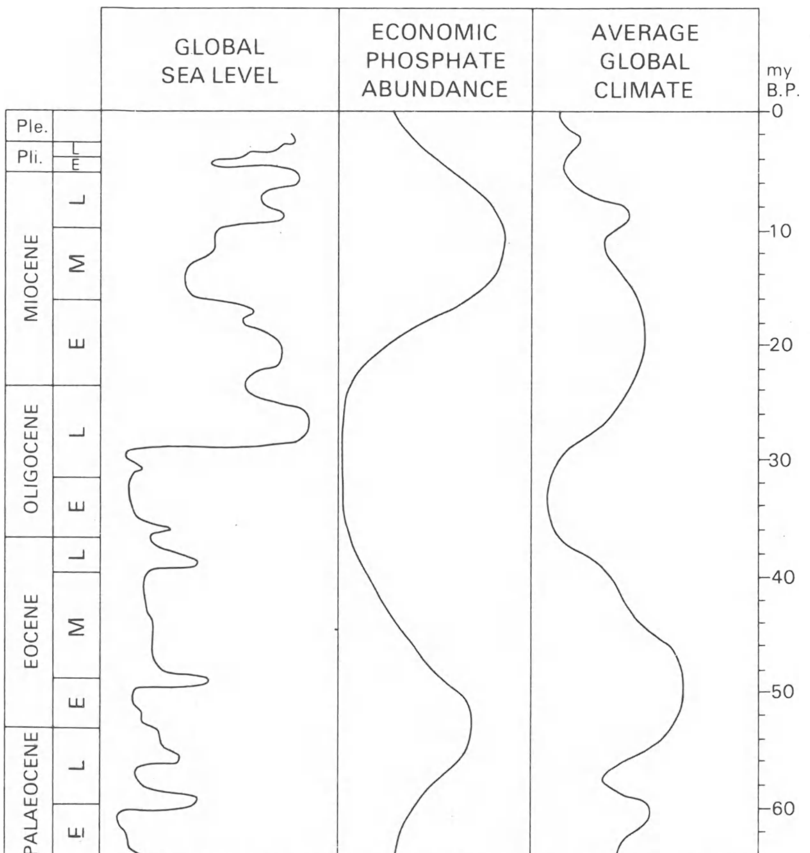

Figure 9 Correlation of the occurrence of world economic phosphate abundance with global sea level and climate (after COOK and McELHINNY, 1979). Sea level curve modified after VAIL and HARDENBOL (1979).

non-depositional episodes within sequences such as the London Clay, as was postulated by VANDENBERGHE (1978) for the 'Berg Sands' concretions, but this may be answered by future research.

It has been known for many years that phosphorite deposits are more numerous, or of greater volume, at certain periods through geologic time and in certain geographical areas (e.g. BUSHINSKI, 1966 ; COOK and McELHINNY, 1979 ; SHELDON, 1980 ; ARTHUR and JENKYNS, 1981). This had led to numerous models suggesting global constraints on the temporal and spatial distribution of phosphorite deposits such as plate tectonics, orogenic or volcanic episodes, climate, oceanic circulation, sea level changes or short term ocean current perturbations (summary in COOK, 1984). Of these factors changes of climate, oceanic circulation and sea level occur over a time scale which is comparable to the periods of Tertiary phosphogenesis and therefore may be of greatest significance. A comparison of global sea level, phosphorite abundance and climate is shown in figure 9. It is now clear that no single factor is responsible for the production of a major phosphorite deposit but that a complex interplay of factors may be involved. Although the phosphorites described in this paper are not major deposits they may in some way reflect the same combination of factors that were responsible for the formation of major phosphorite deposits. Within a relatively small basin like the southern North Sea factors such as climate, transgressive events and nutrient supply should be reasonably similar throughout the basin allowing certain generalised correlations to be made. It should be emphasized that COOK and McELHINNY's curve of phosphate abundance shown in figure 9 is based largely on economic phosphorite occurrences and would ignore such limited occurrences as those in the area discussed here. Furthermore it is based on deposits of many different types including reworked deposits rather than only authigenic phosphorite. Details of phosphorite occurrences in the area around the southern North Sea have been summarised in figure 10. From this it can be seen that on the available evidence the phosphogenic episodes in the Early Eocene and Middle Miocene may be

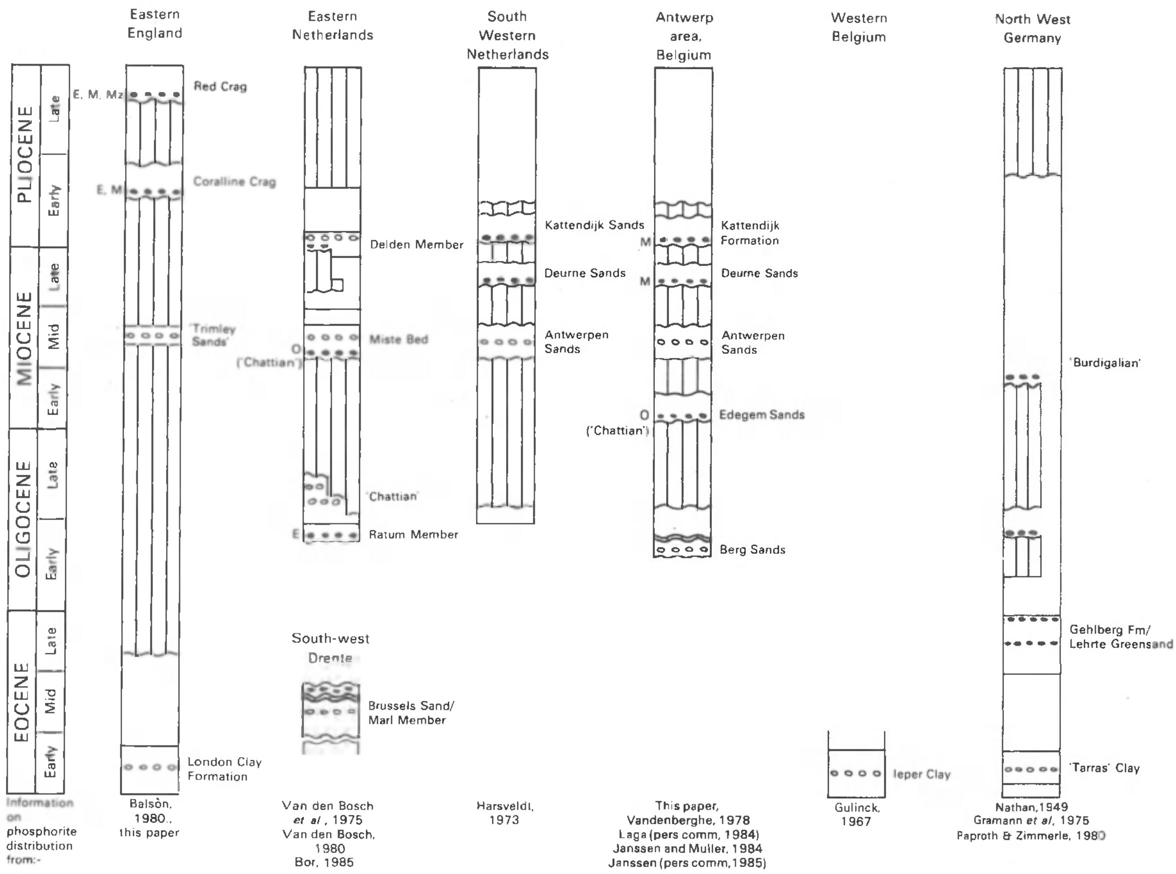

Figure 10 Correlation of principal phosphorite occurrences in the Tertiary of Northwest Europe. Known sources for derived material in reworked phosphorite deposits indicated on left of column.
M = Miocene. O = Oligocene. E = Eocene. Mz = Mesozoic.
Phosphorite symbols as in figure 6.

correlated on both the east and west sides of the southern North Sea, a distance of at least 200 km for the Middle Miocene episode and over 700 km for the Early Eocene episode. These two periods correspond closely with periods of abundant economic phosphate deposits (figure 9) ; the 'phosphorite giants' of ARTHUR and JENKYNS (1981). These two periods were also times of worldwide transgression and warm climate. The evidence suggests that the Oligocene in particular was a time of very little phosphogenesis worldwide.

In the area around the southern North Sea, there is evidence of localised occurrences of authigenic phosphorite in the Early Oligocene 'Berg Sands' at Sint-Niklaas (VANDENBERGHE, 1978 ; Personal communication, P. LAGA, 1984) and in the Late Oligocene (Chattian) of the eastern Netherlands (VAN DEN BOSCH *et al.*, 1975). Unfortunately these two horizons have no lateral equivalents in neighbouring areas and so far it has not proved possible to correlate these phosphorite occurrences beyond the limited areas from which these authors described them. In contrast, although lateral equivalents of the Delden Member (Early Pliocene) and the Brussels Sand/Marl Member (Middle Eocene) of the Netherlands do occur elsewhere in the southern North Sea basin authigenic phosphorite material has yet to be identified in these formations.

These problems of correlation are due to the incompleteness of the Tertiary sequences of this area and the rapid lateral facies changes within this relatively small sedimentary basin. Thus a phosphorite-bearing formation in one place may be equivalent to a stratigraphic hiatus only a short distance away. Further, more detailed work on the nature and age provenance of much of the reworked phosphorite material may help to alleviate this problem.

It is unclear at present therefore, whether or not extremely localised phosphorite occurrences might be due to some unusual condition within the sediment of these areas. What does seem clear is that the Early Eocene and possibly the Middle Miocene phosphorite occurrences are traceable across the southern North Sea and are therefore reflecting basin-wide events.

Phosphorites most commonly develop in marine shelf or upper slope sequences. In the southern North Sea area authigenic phosphorite material tends to be most frequent in muds or muddy sands. The authigenic phosphorites are therefore to some degree facies controlled and the correlation with times of marine transgression might be expected. It may be more constructive to examine those times of marine transgression which deposited apparently suitable lithofacies in which authigenic phosphorite deposition does not occur. One example of such a deposit is the Early Oligocene Boom Clay of Belgium and equivalent beds elsewhere. As already mentioned there may be evidence of some localised phosphorite deposition in the underlying 'Berg Sands' of Sint-Niklaas and the overlying Chattian of the Netherlands but nevertheless the Oligocene of the southern North Sea area appears to have less phosphorite development than would be expected on the basis of availability of suitable facies. It is therefore tempting to correlate this apparent paucity of phosphorite deposition with the known worldwide deficiency of economic phosphorite abundance in the Oligocene (COOK and McELHINNY, 1979). However it should also be noted that there will be a natural bias in that during a major regression such as that during the Late Oligocene, continental rather than marine facies will be better represented in those areas most accessible to land based studies.

In conclusion it should be re-emphasised that the phosphorites discussed in this paper are relatively very small in volume and geographical extent when compared to the economic deposits from which COOK and McELHINNY drew their conclusions. It is possible that very localised changes in pore water chemistry may result in the precipitation of phosphorite concretions in any marine sediment without invoking changes due to major worldwide events.

SUMMARY

A number of minor phosphorite deposits (some previously economic) and authigenic phosphorite concretion horizons occur within the Tertiary of the southern North Sea basin. The phosphorite deposits were formed as lag gravels and remanié deposits on unconformity planes by the reworking and winnowing of sediments bearing authigenic phosphorite concretions and other phosphatic material.

Where the distribution of facies and preserved sedimentary record allow, authigenic phosphorite-bearing horizons appear to be correlatable over the limited area of the southern North Sea basin and may reflect known global variations of phosphogenesis.

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Note added in proof

Further work since this paper was written suggests that the fauna of the 'Trimley Sands' of eastern England is not of Middle Miocene age but may be of similar age to the Delden Member of the eastern Netherlands i.e. latest Miocene or earliest Pliocene (BALSON, 1987).

7 Stratigraphic Occurrence of Coherent Rocks in the Eocene of Belgium

B. FOBE

Research Fellow IWONL, Laboratory of Mineralogy, Petrology and Micropedology, Geological Institute, State University of Gent, Krijgslaan 281/S8, B-9000 Gent, Belgium.

INTRODUCTION

Coherent rocks are frequently encountered in several formations of the Eocene in Belgium. They are especially abundant in the, mainly sandy, middle part, i.e. between the Flanders clay (Ypresian, Yc) and the Asse clay (Bartonian).

As they are potential reflectors for seismic waves, information about their occurrence will be given in this paper, whereas less attention will be paid to petrographic data.

GENERAL CHARACTERISTICS

Generally, the lithified sediments display some common features. Their shape is mostly platy, sometimes blocky. The thickness of the plates, which are generally lined up in distinct levels, ranges between 5 cm and 60 cm. Their width amounts to a few decimetres. In some outcrops, however, consolidated layers can be followed continuously over tens of metres.

The cement consists either of calcite, or of silica minerals (opal or chalcedony). Three main types are distinguished : calcareous sandstones, sandy limestones and siliceous sandstones. The purest limestones are the nummulite accumulations in the Ieper Formation, and the micritic Gobertange Stone in the Brussel Formation, which may contain more than 90 % calcium carbonate.

In the following chapters, the different formations will be considered systematically.

IEPER FORMATION

The Egem sands (formerly named Mons-en-Pévèle sands or Vorst sands (Yd)) contain limestone layers, consisting of cemented shells of *Nummulites planulatus*, and siliceous sandstones. HERENT (1895) mentioned four layers in Mons-en-Pévèle (France). Several levels are encountered in the subsoil of Ghent (DE MOOR and GEETS, 1973). At least one limestone layer is found in Brussels (HALET, 1930). In East Flanders, this level is sometimes encountered about 10 m below the top of the Egem sands (DELVAUX, 1884).

MONT PANISEL FORMATION

Coherent layers are found in the Pittem Member (formerly Anderlecht clay, P1 c), and in the Vlierzele sands (P1 d).

The base of the Pittem sandy clay is locally formed by a 60 cm thick, opal cemented sandstone bed. In the Egem claypit, near Tielt, this level can be followed along the whole (nearly 100 m) profile.

Above this layer, four to five thinner (10 cm) beds of platy sandstone may occur. They are separated by an interval of about 1 metre of loose sediments.

The Pittem clay is overlain by the Vlierzele sands, consisting of cross-stratified sand, with small clay seams. In some outcrops, the upper part of the Vlierzele sands shows more individualised clay layers (a few cm), alternating with sand (1 to 4 dm). These sand intervals contain platy, quartzite-like sandstone beds (10-15 cm), cemented by opal and chalcedony. This was clearly visible in the Beerlegem sandpit (now abandoned), near Ghent. In a sequence of 4 metres, no less than 11 sandstone layers were exposed.

The Vlierzele sand, with the consolidated rock levels, is also found offshore. DE BREUCK and DE MOOR (1967) mentioned its presence in a borehole on the Wenduine beach, at a depth of 27 m. In this area, eroded sandstone slabs are washed ashore.

KNESSELARE FORMATION

This formation includes the former Upper Paniselian (Aalter sands), and two underlying members (Beernem and Oedelem sands) which were recently discovered in borings (GEETS and JACOBS, 1977). These authors mentioned the presence of three lithified layers in the Beernem Member. In a borehole in Ursel, two coherent levels were found in the Oedelem sands (DE BREUCK *et al.*, 1984).

BRUSSEL FORMATION

The Brussel Formation is only found east of the Zenne river, in the Brabant province. It does not occur in the coastal area.

Some parts are rich in coherent layers. They consist of subcontinuous plates or concretions (up to 20 cm), separated by a few decimetres of loose sand. The cement may be siliceous, as well as calcareous.

LEDE FORMATION

The Lede sands contain a restricted number (three or four) of sandy limestone layers. Their thickness amounts to 50 cm.

One consolidated bed is found immediately above the base of the formation. The basal grit itself may be very rich in reworked slabs of sandy limestone, derived from the erosion of older formations.

Anyhow, the presence of the Lede sands in the coastal area is doubtful, and has not been proven yet. Probably, the formation thins out in the vicinity of Eeklo.

CONCLUSIONS

In the Eocene in Belgium, consolidated layers are sometimes encountered at the base of sedimentological units (e.g. the Pittem Member and the Lede formation). The upper part of the Vlierzele sands is sometimes very rich in sandstone beds.

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8 Stratigraphic analysis of the Ypresian off the Belgian coast

M. DE BATIST, H. DE BRUYNE, J.P. HENRIET and F. MOSTAERT

Renard Centre of Marine Geology, Geological Institute, State University of Gent, Krijgslaan 281/S8, B-9000 Gent, Belgium.

ABSTRACT

The Ypresian depositional sequences defined by high-resolution reflection seismic investigations off the Belgian coast are correlated with the main Ypresian lithostratigraphic units known along the Belgian shore line and further inland. Particular attention is paid to the detailed seismic-stratigraphic features of the Ypresian, including the seismic facies related to the deposition and compaction history of these sediments.

INTRODUCTION

The interpretation of the numerous high-resolution reflection seismic profiles shot in recent years over the Belgian continental shelf has resulted in a detailed seismic-stratigraphic description and subdivision of the Palaeogene sequences of the offshore extension of the Tertiary Belgian Basin (HENRIET *et al.*, this volume). A mere glance on the seismic-stratigraphic subcrop map of the top of the Tertiary (figure 3 in HENRIET *et al.*, this volume) reveals that the direct substratum of the larger part of the Belgian continental shelf consists of Ypresian deposits. These deposits are also known on land, where they have been the subject of detailed lithostratigraphic and chronostratigraphic studies : e.g. the revision by STEURBAUT and NOLF (1986).

Both areas of observation are only separated by a narrow gap corresponding with the nearshore zone, where seismic observations are impeded by shallow waters and gassy surficial sediments absorbing all acoustic energy. There are consequently obvious reasons for attempting to bridge this gap, keeping however in mind the fundamental differences in stratigraphic approach applied in both domains.

GEOLOGICAL SETTING

The Belgian Ypresian sequence was deposited in the so-called Belgian Basin, a bightlike extension of the Early Eocene North Sea Basin. These deposits cover most of North Belgium, including the continental shelf. In the southern and western parts of this area, Ypresian deposits are directly outcropping or but slightly concealed by a discontinuous Quaternary cover. Further north and west, they are covered by younger Tertiary sediments (figure 1).

The total thickness of this sequence increases from a few metres in the far south of the land area up to some 180 m more north, in the Knokke boring (BGD 11E-138).

In the coastal area and offshore, the strata are gently dipping towards the northeast. This regional dip is probably the combined effect of the subsidence of the North Sea Basin, the tectonic activity of the Weald-Artois Anticline and the relative uplift of the London-Brabant Massif. As a consequence, progressively younger Ypresian deposits are met at the surface from southwest to northeast.

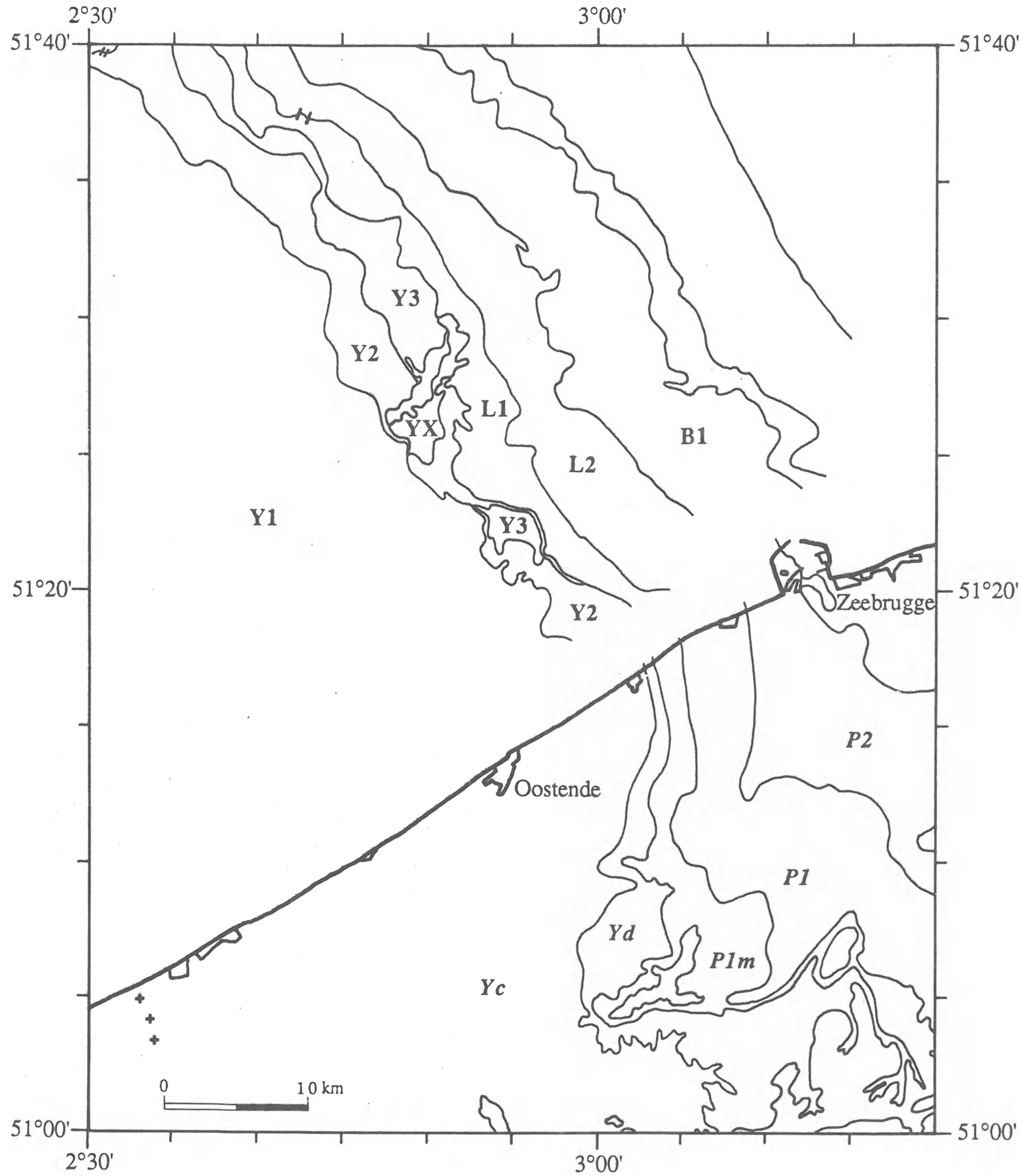


Figure 1 Combined solid map of the Eocene deposits in the Belgian Basin, illustrating the extension of offshore seismic-stratigraphic units (after HENRIET *et al.*, this volume) and onshore lithostratigraphic units. *Yc* = Ieper Clay. *Yd* = Kortemark Silt - Egem Sand. *P1m* = Merelbeke Clay. *P1* = Vlierzele Formation. *P2* = Oedelem Formation.

YPRESIAN LITHOSTRATIGRAPHY

According to STEURBAUT and NOLF (1986), two formations can be defined in the Ypresian deposits in Belgium : a lower Ieper Formation and an upper Vlierzele Formation. Both can be divided, vertically as well as laterally, into several lithostratigraphic units (figure 2). Lateral variations in lithological composition are common along the basin

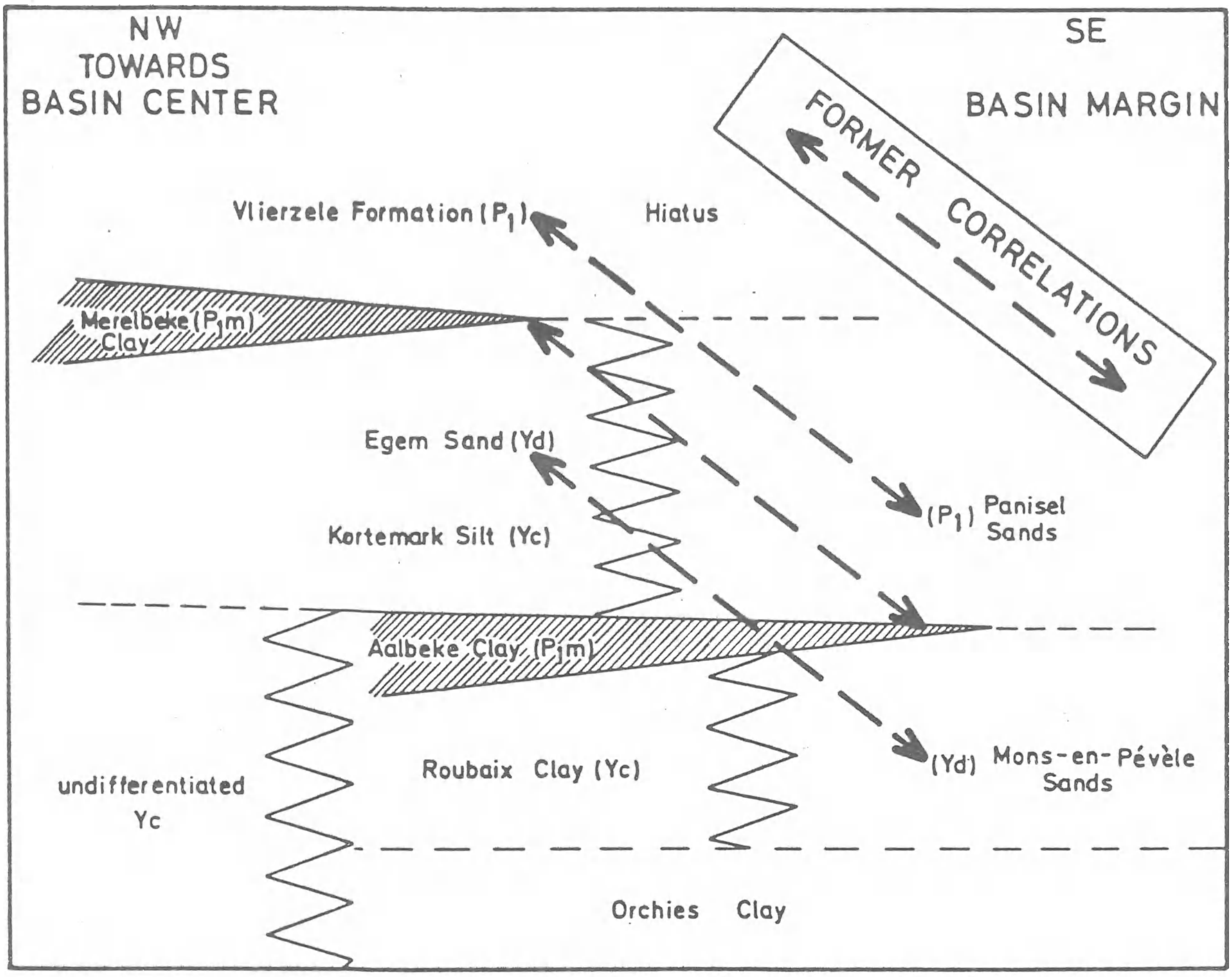


Figure 2 Correlation chart of lithostratigraphic units of the Ypresian in the Belgian Basin (after STEURBAUT and NOLF, 1986).

margins in central and southern Flanders but tend to become less explicit towards the northeast, further offshore in the palaeobasin.

In the coastal area, the Ypresian starts with a major undifferentiated clayey sequence : the Ieper Clay, grouping the lateral equivalents of the Orchies, Roubaix and Aalbeke Clay Members. Further south, the homogeneous heavy clay may become progressively more silty, locally containing silt or sand intercalations. The Ieper Clay is subcropping at the Quaternary coastal plain deposits along the west coast from De Panne to De Haan. Below the east coast, north of De Haan, its total thickness amounts to some 150 m (borings BGD 10E-30 and BGD 11E-138).

It is covered by the sandy, lateral equivalent of the Kortemark Silt Member and the Egem Sand Member, consisting of very fine sand with clay intercalations. Their maximum recorded thickness is 23 m (only 9 m in the Knokke boring BGD 11E-138).

The Ieper Formation terminates with a characteristic heavy clay layer of variable thickness (5-10 m) : the Merelbeke Clay Member. It is separated from the overlying Vlierzele Formation by a significant unconformity.

The Vlierzele Formation is quite heterogeneous and consists mainly of medium-grained cross-bedded sands. The Pittem Clay Member is a lower interval of local extension, built up of an alternation of clayey sands and sandy clays, with locally sandstone beds (boring BGD 11W-88).

CHARACTERISTICS USED IN CORRELATION

In developing a tentative correlation between offshore seismic stratigraphy and onshore lithostratigraphy all available evidence should be taken into consideration : the number of identified depositional sequences, the characteristics of their boundaries, their thickness and geometry, the seismic facies and the relative position in relation to possible onshore equivalents.

The subdivision of the offshore sequences into depositional sequences - the basic seismostratigraphic building stones - has been carried out in accordance with the principles of seismic stratigraphy (VAIL *et al.*, 1977) and the specific interpretative approach, as described elsewhere (HENRIET *et al.*, this volume). Four depositional sequences of supposed Ypresian age have thus been identified : Y1, Y2, Y3 and YX, separated by unconformities or hiatuses.

Reliable sequence thicknesses have been calculated with a generalized velocity model, compiled from a number of scattered velocity measurements carried out with various techniques (CHERLET, 1978 ; HELDENS, 1983 ; VERCOUTERE, 1987).

The seismic facies of the intervals between prominent reflectors could in some instances yield a clue for the interpretation of the depositional environment and of the early compaction history. It is also a potential lithological indicator, which however requires due borehole control. It should be noted here that the highest resolution can only be obtained in the upper 50 to 70 m below seabed, due to fundamental properties of seismic wave propagation and attenuation. As the strata dip towards the northeast, successive depth intervals can be investigated with optimum resolution for facies studies, but these are systematically shifted towards the northeast as their age decreases. This implies that a composite vertical section of seismic facies descriptions for successively younger depositional sequences (such as in figure 2, HENRIET *et al.*, this volume) always contains a certain lateral, basin-inward component.

A major criterion for offshore-onshore correlation has also been the purely geometrical fit of units on either side of the nearshore gap, using isobath maps of the base of the Quaternary (MOSTAERT *et al.*, this volume) and calculations of the local strike as guide-lines.

In correlating seismostratigraphic and lithostratigraphic units, a few considerations ought to be made about the fundamental validity of this approach. In accordance with the general principles of seismic stratigraphy as defined by VAIL *et al.* (1977), seismic reflectors do have a chronostratigraphic significance, as they generally represent bedding planes considered as isochronous surfaces. Although still point of debate, this statement seems to have proved its validity at the level of the analysis of major sedimentary basins. It has also been retained in the present approach (HENRIET *et al.*, this volume) of the Belgian continental shelf sequences. Correlating such "chronostratigraphic" boundaries with lithostratigraphic breaks might thus sound like a curse in stratigraphic practise. Two statements should be made in this respect.

First, the problem stated here is one of short-range correlation between two areas which have intensively been investigated with different techniques. Bridging a relatively narrow gap irremediably involves the best possible projection of one set of boundaries into the other one. In this way it might be better to speak of a "best fit" between two sequences rather then of a correlation, in the full significance of this word. Hence this bridging operation carried out on a very local scale does not automatically imply a recognition of a long-range identity of lithological boundaries and seismic reflectors. The problem of evaluating the true nature of a seismic reflection in this regional stratigraphic domain can only be solved with adequate borehole control in the offshore environment. The alternative of applying high-resolution reflection seismics of similar standards on land does not seem to be a short-range perspective, both for fundamental reasons of wave propagation in unsaturated media (a particular aspect of the more general problem of so-called "static" corrections on land) and for economical reasons.

A second remark is that the seismic-stratigraphic analysis of deep hydrocarbon-bearing basins (VAIL *et al.*'s original approach) automatically involves the use of waves with a dominant wavelength (and hence resolution) of a couple of tens of metres. In high-resolution reflection seismics, the dominant wavelength is in the metre range, thus providing a much finer tool possibly offering a higher potential for imaging lithological contrasts. This emerging idea deserves further attention.

THE Y1 DEPOSITIONAL SEQUENCE

The boundary between the Ypresian sequences and the underlying Thanetian (Palaeocene) deposits is locally marked by a weak but clearly defined unconformity.

Above this unconformity, the whole of the Y1 sequence is characterized by a lack of true reflecting horizons. Only at the base, about 6 to 7 m above the unconformity, a relatively consistent reflector can be identified, which can be traced all over the Southern Bight and which correlates in the Thames estuary with a hard horizon of volcanic ash, the so-called Harwich stone band. This ash marker is known as a prominent reflector also further north in the North Sea Basin.

The relative absence of true reflecting horizons, which can be identified with any seismic source and configuration, does not mean the absence of any possibility of imaging the structure of the Y1 sequence, which can be identified with the Ieper Clay. Indeed when high-resolution seismic sources are properly tuned, in other words when the right energy is delivered in the right frequency band and due attention is paid to pulse shaping, a remarkable interference composite emerges from this homogeneous clay unit . Such a composite represents the sum of all weak reflection responses from the many, subtle interfaces and laminations within the clay sequence (HENRIET *et al.*, 1982 ; HENRIET *et al.*, 1988). Such interference patterns closely mould

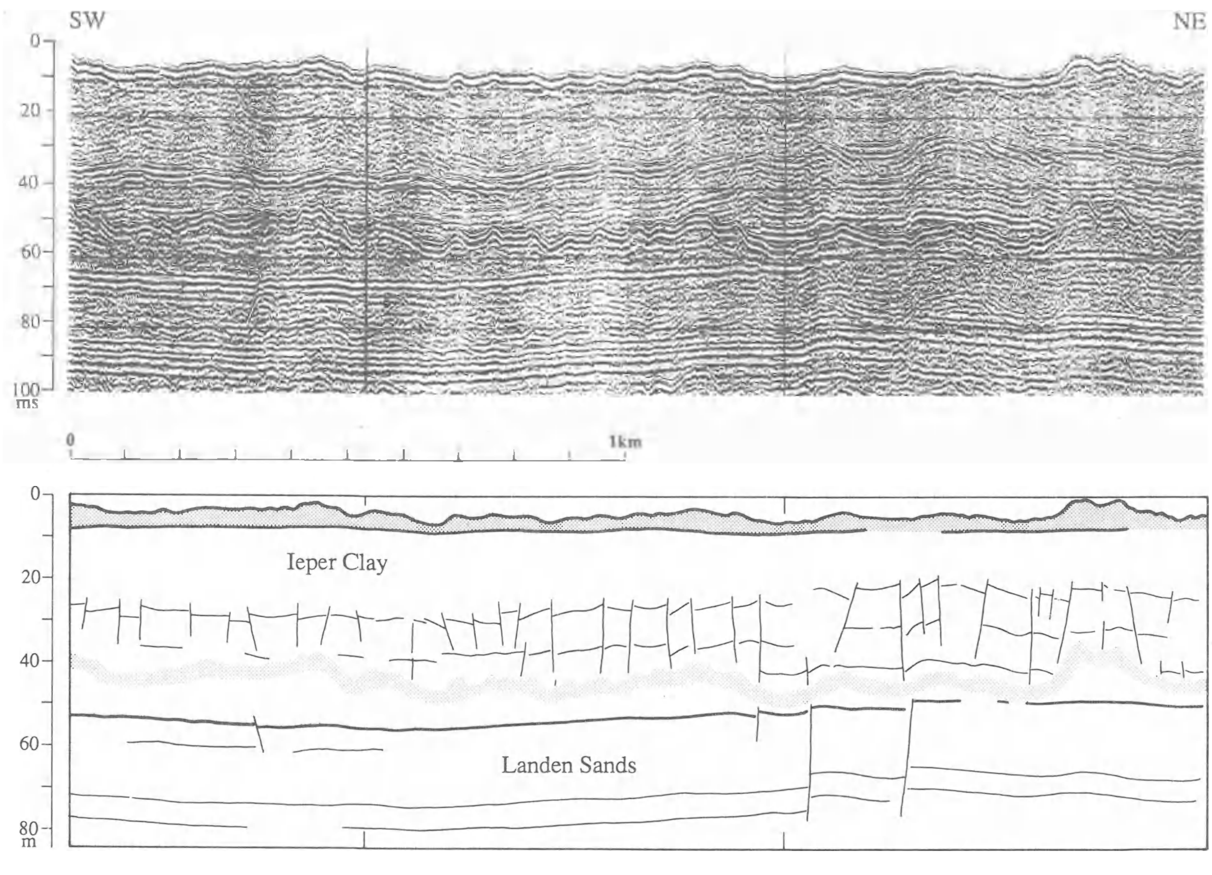


Figure 3 Analog-recorded sparker section and interpreted line-drawing showing the lower interval of Ypresian clay-tectonic deformations. Approximate localization : N 51°09.50', E 02°15.00'.
Vertical scales in ms are two-way time. Vertical scales in m are calculated with an interval velocity of 1620 m/s.

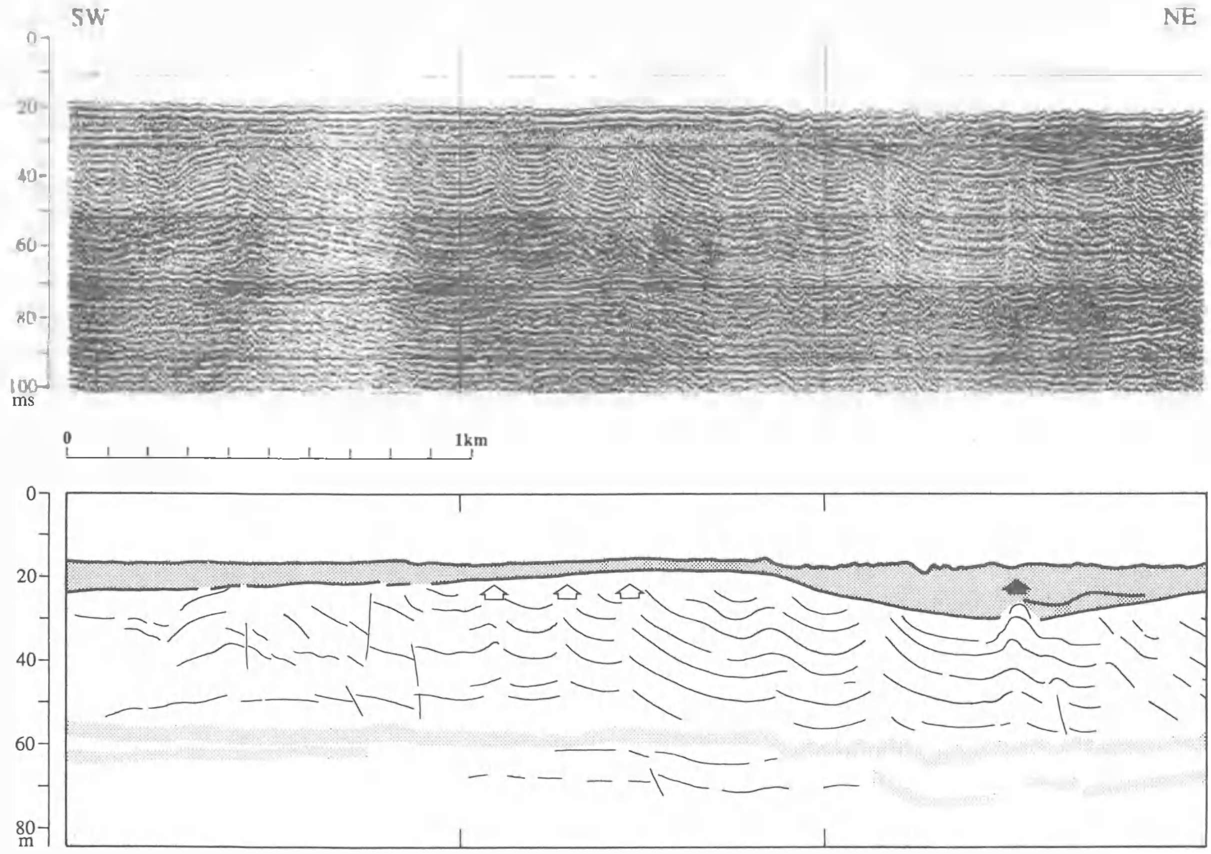


Figure 4 Analog-recorded sparker section and interpreted line-drawing showing the second interval of Ypresian clay-tectonic deformations. Approximate localization : N 51°16.00', E 02°24.00'.
Vertical scales in ms are two-way time. Vertical scales in m are calculated with an interval velocity of 1620 m/s.

the structure of the clay beds and allowed the discovery of a most intriguing and extensive set of internal deformations of these clays.

The style of these deformations may vary from place to place, being more chaotic in some areas and regular in others, but a remarkable observation is a distinct kind of vertical zonation, with a superposition of intervals with different but related deformations.

In a lower interval, up to some 25 m above the undisturbed basal reflector, intense block-faulting may be observed, with tilted and bended blocks and apparently randomly dipping fault planes (figure 3). The average throw amounts to a couple of metres.

In a second interval, stretching from about 25 metres up to at least 70 metres above the clay base, the movement initiated in the lower interval amplifies and develops into a convoluted pattern, consisting of a festoonlike alternation of broad, rounded synclines and narrow, cuspate anticlines (figure 4). These anticlines often develop into diapyrlike features. The apparent wave-length of the convoluted structure varies between 200 and 300 metres and the amplitude of the clay waves ranges from 2 to 10 metres.

Further upward in the stratigraphic sequence, the convoluted structure becomes increasingly disturbed, giving way to a pattern of faulted blocks, sometimes with a dominant tilt direction (figure 5). Such patterns of faulted blocks with dipping fault planes have also been observed on land, e.g. in the "Koekelberg" quarry in Marke (figure 6).

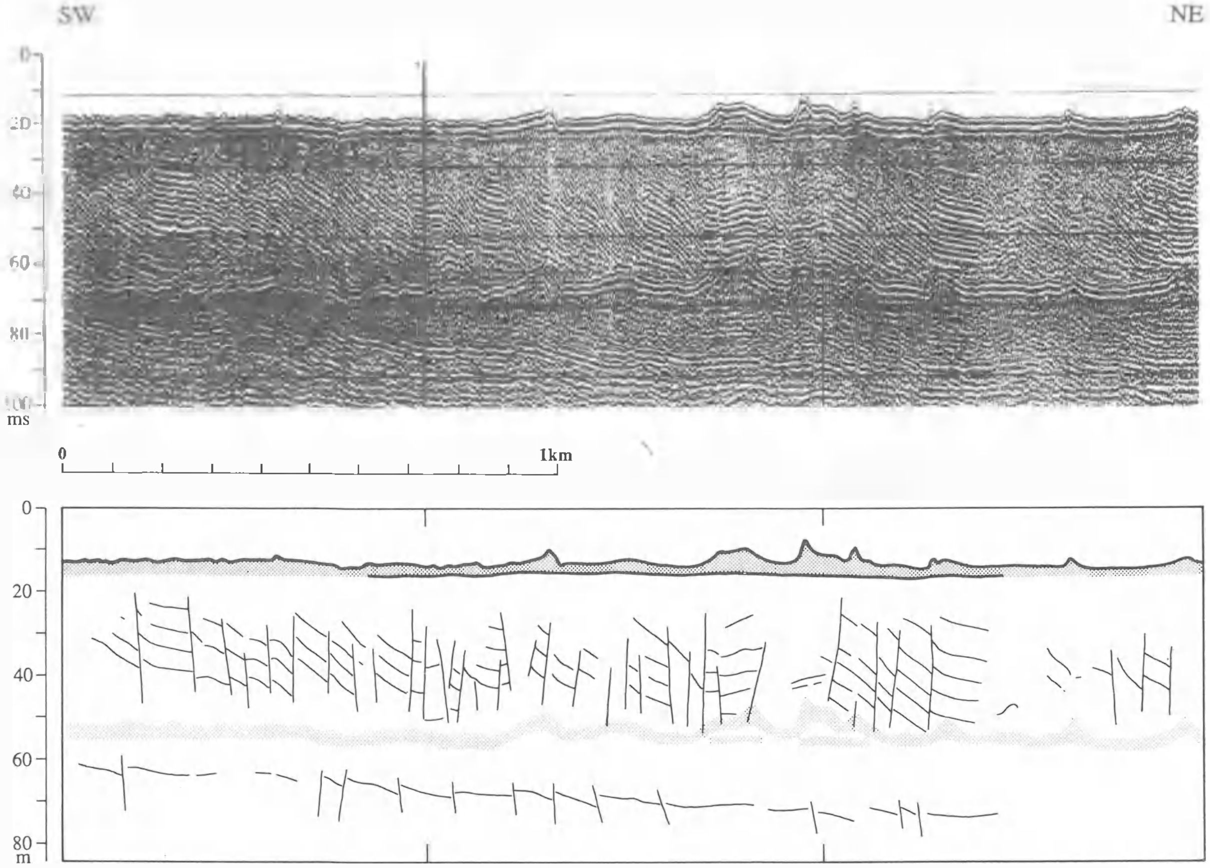


Figure 5 Analog-recorded sparker section and interpreted line-drawing showing the third interval of Ypresian clay-tectonic deformations. Approximate localization : N 51°18.00', E 02°26.00'.
Vertical scales in ms are two-way time. Vertical scales in m are calculated with an interval velocity of 1620 m/s.

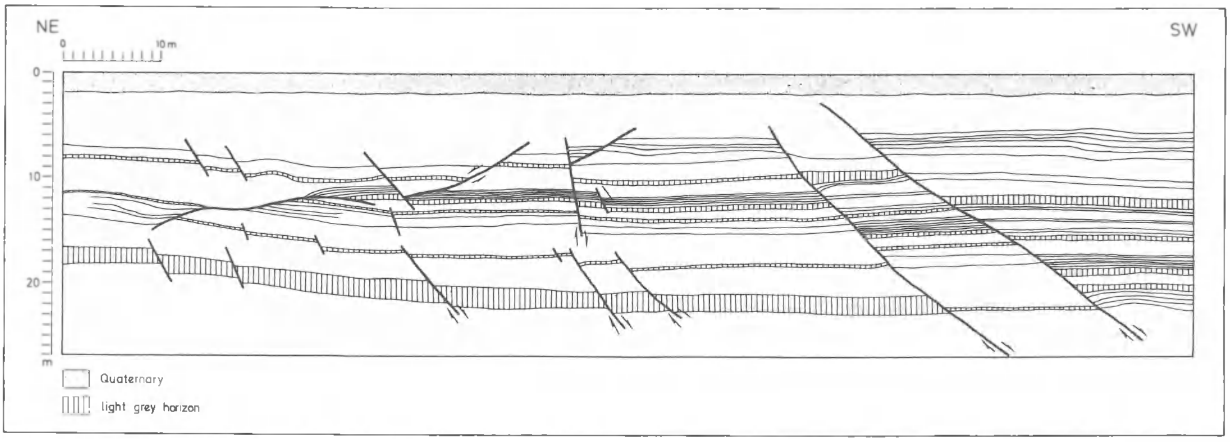


Figure 6 Faults in the "Koekelberg" quarry, Marke (Belgium) (VAN VAERENBERGH, 1987).

Approximatively at the level where the Ypresian clay should grade into the Ypresian sands, another peculiar deformation pattern is observed (figure 7). Here some well defined reflectors are again affected by faults, but the deformation consists of alternately tilting and down-warping bedding terminations, without significant tilting or displacement of the blocks themselves, between the faults. Tilting and down-warping segments associated with each fault point away from each other, in opposition with bedding deformations caused by normal drag associated with block-faulting.

A genetic model for these features has been proposed (HENRIET *et al.*, 1988), implying a build-up of undercompaction and density inversion due to a self-sealing mechanism of the compacting clay body, the subsequent development of a central clay wave fitting a Rayleigh-Taylor-type instability as well as of brittle deformations in the more compacted base and top intervals and finally a relaxation of the pore water overpressure, freezing the deformations in the shape nowadays observed. This whole process was probably completed short after the deposition of the Ieper Clay unit and of the covering Egem Member, as shown by the amplitude of the deformations, fading progressively away in the overlying sand layers.

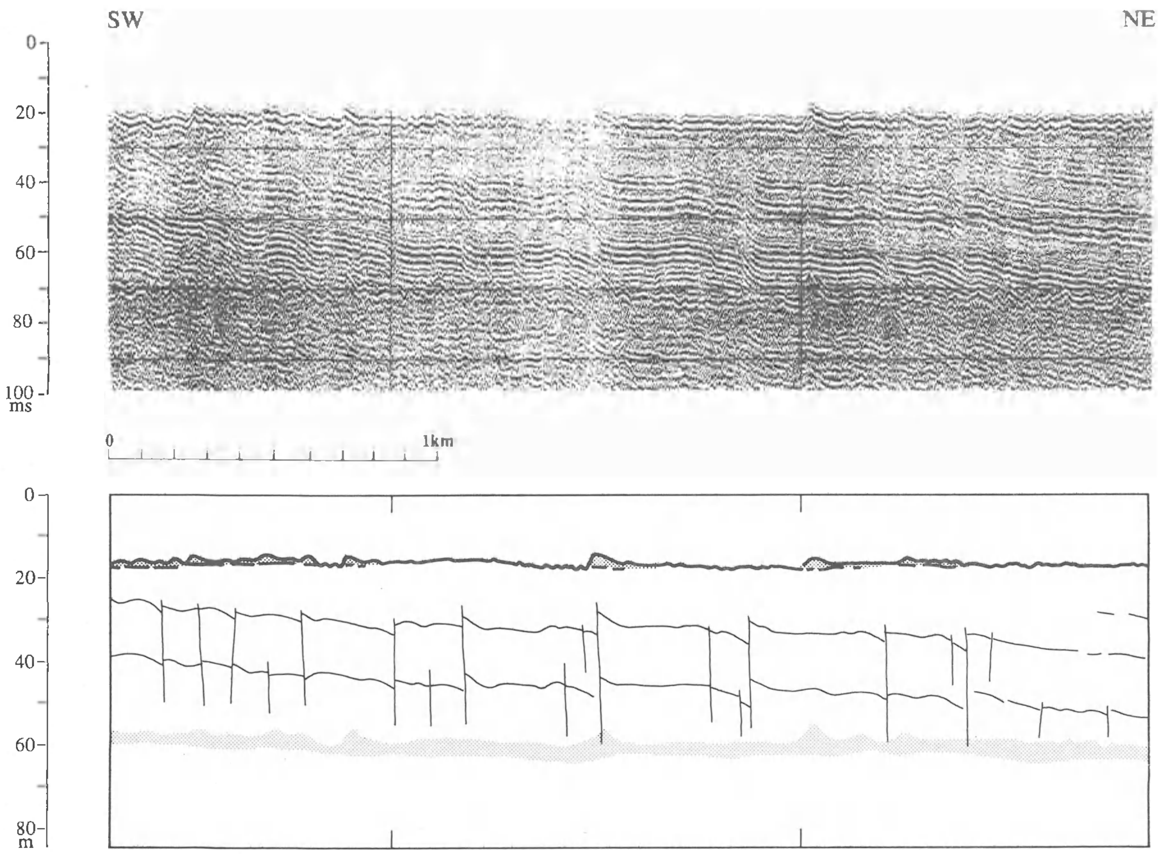


Figure 7 Analog-recorded sparker section and interpreted line-drawing showing the upper interval of Ypresian clay-tectonic deformations. Approximate localization : N 51°31.50', E 02°39.50'.
Vertical scales in ms are two-way time. Vertical scales in m are calculated with an interval velocity of 1620 m/s.

THE Y2 DEPOSITIONAL SEQUENCE

The boundary between the Ieper Clay and the following depositional sequence, Y2, can only be identified close to the coast, where a very discrete downlap can be observed on the distinct base reflector Y2.1. This weak unconformity disappears towards the basin centre, where the transition becomes gradual, but the reflector itself remains one of the strongest, most characteristic seismic-stratigraphic markers of the entire Palaeogene off the Belgian coast.

The total observed stratigraphical thickness of Y2 ranges from some 30 m in the north to an average of 15 m closer to the coast, where the sequence clearly has been truncated. Deep scouring can locally be observed (figures 8 and 9).

In general, two seismic facies units can be identified within Y2. A lower one, some 18 m thick, may consist of very weak, sometimes prograding reflector elements, but is usually completely reflection-free. Further to the north it is overlain by a second seismic facies unit, composed of a set of parallel reflectors of changing amplitude.

In accordance with its seismic facies characteristics the entire sequence can be correlated with the Egem Member : a lower interval of homogeneous, fine sands and an upper one, containing *Nummulites*-enriched horizons and calcareous sandstone banks.

Both units can also geometrically be connected, despite the strong northeasterly deflection in the outcrop pattern observed when the offshore seismic-stratigraphic map is compared with the onshore lithostratigraphic map (figure 1). This deflection is caused by a local depression in the erosion surface at the top of the Tertiary (MOSTAERT *et al.*, this volume ; DE VOS, 1984).

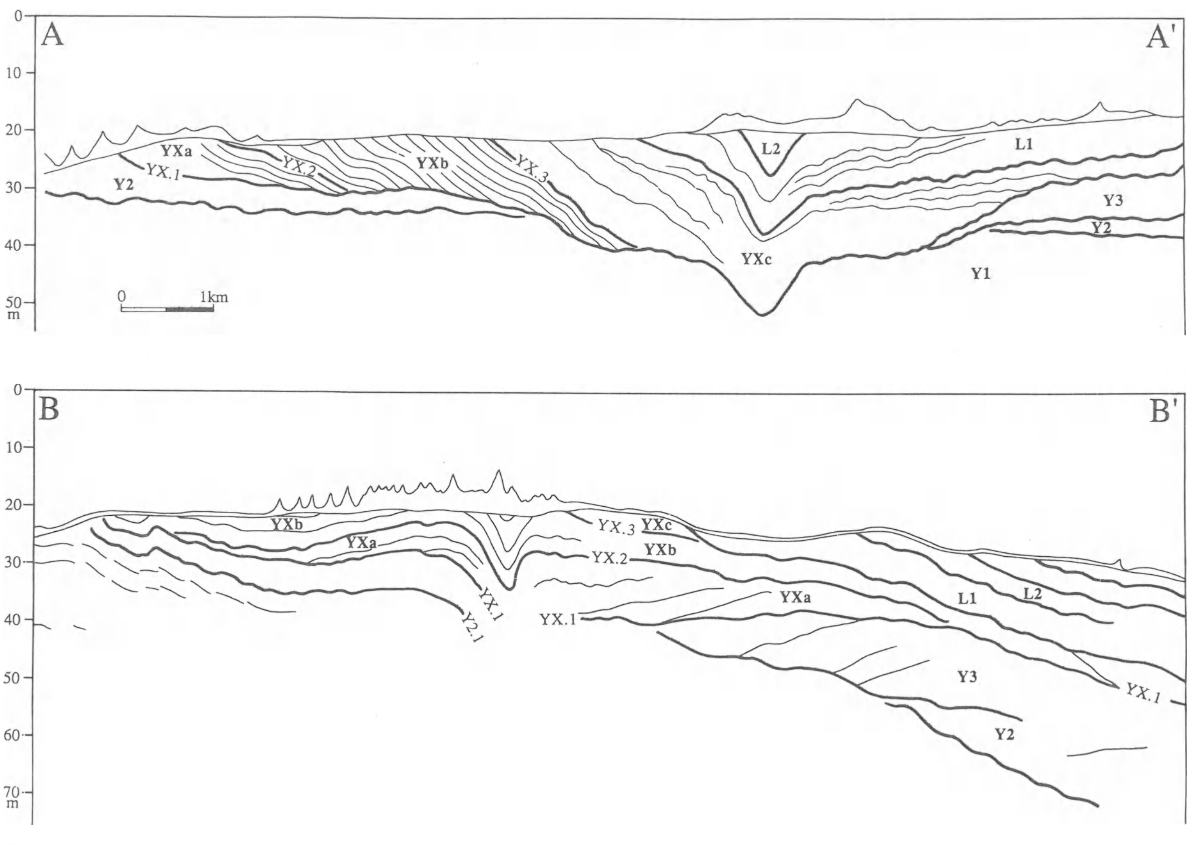
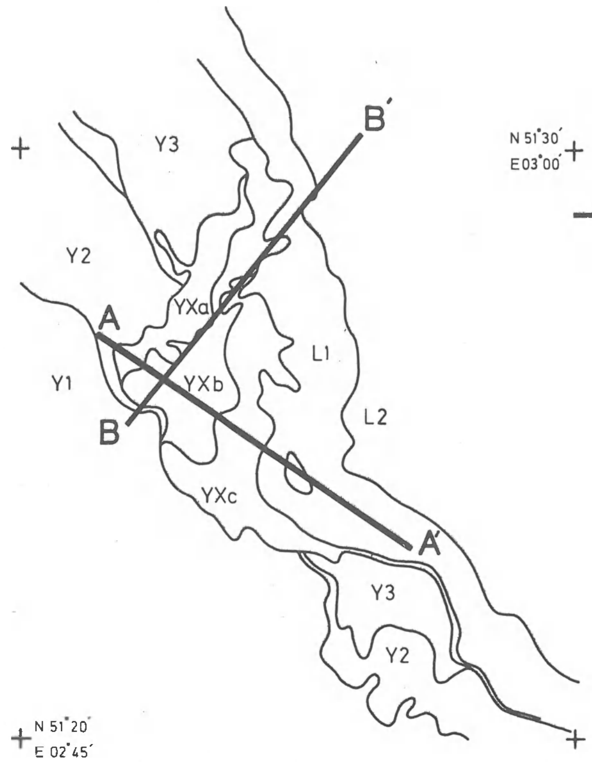


Figure 8 Type sections AA' and BB', illustrating successive erosive phases and the configuration of the basin-fill subsequences north of Ostend. For approximate localization see figure 9.
Vertical scales in m are calculated with an interval velocity of 1700 m/s.

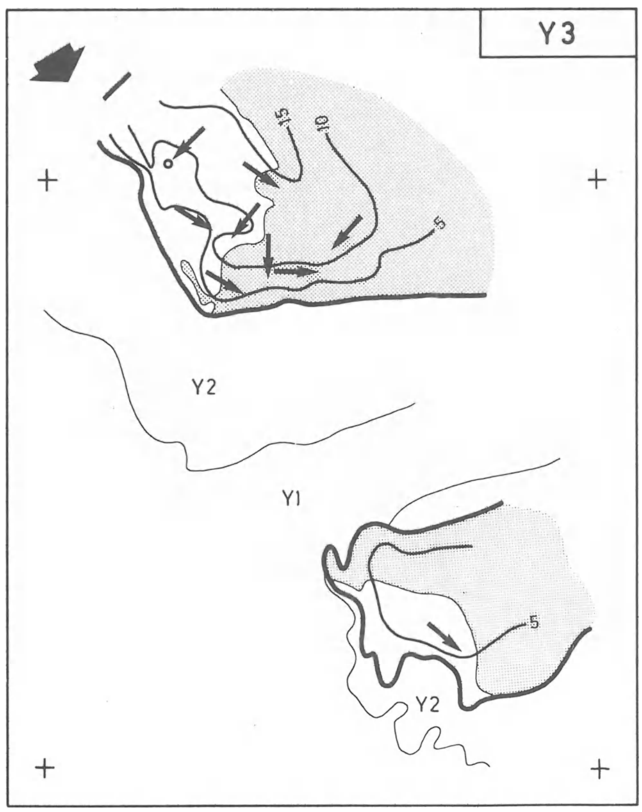


YPRESIAN BASIN-FILL SEQUENCES

SUBCROP MAPS OF
Y3, YXa, YXb, YXc TO L1;
LAPOUT PATTERNS AND ISOPACHS

SYMBOLS :

- isopach contour line (5)
- subcrop extension
- covered by overlying Tertiary units
- apparent downlap
- apparent onlap
- apparent concordance
- direction of progradation



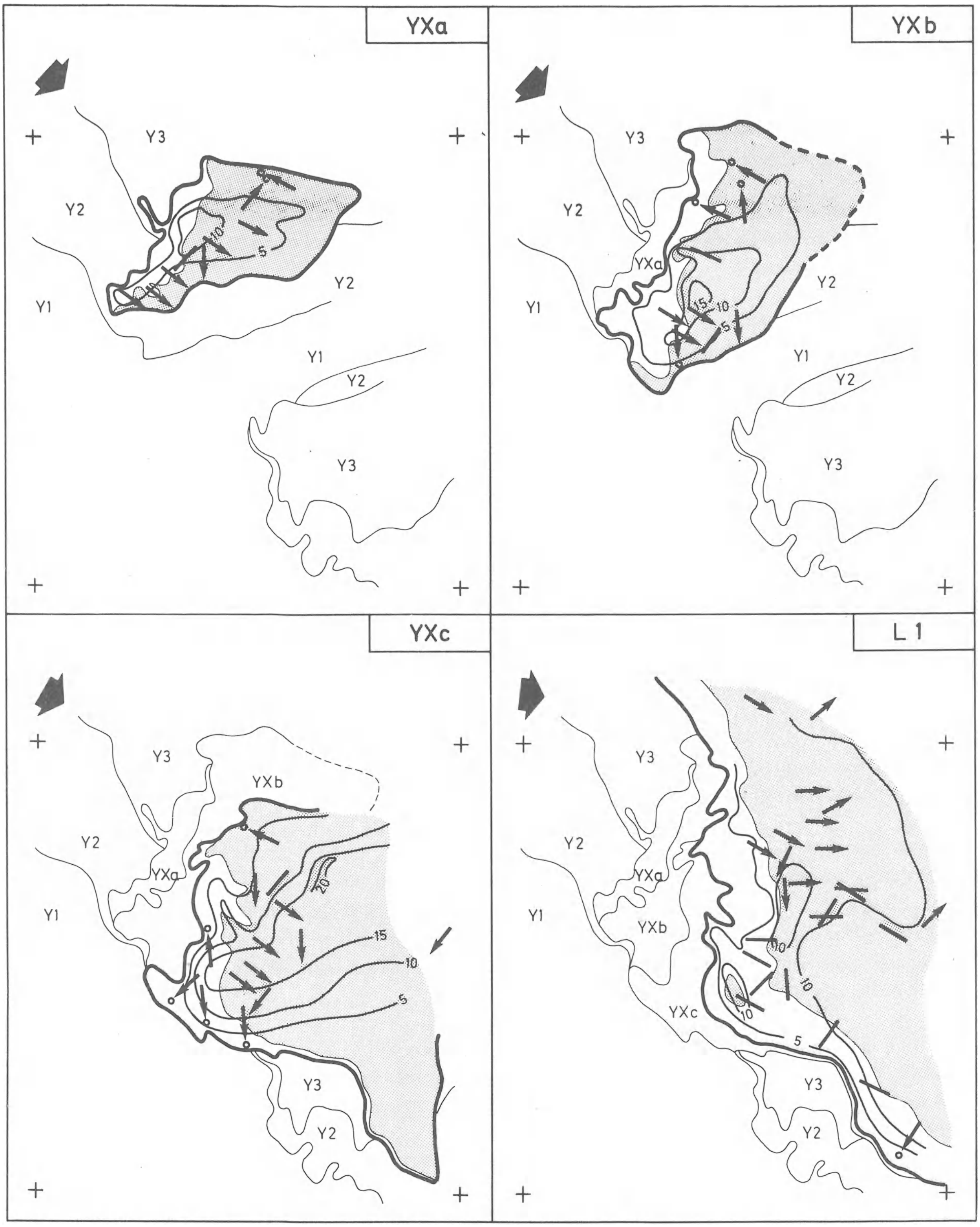


Figure 9 Successive subcrop and isopach (on base of an interval velocity of 1700 m/s) maps of Y3, the YX basin-fill subsequences and L1, illustrating seismic facies and lapout patterns.

THE Y3 DEPOSITIONAL SEQUENCE

The following depositional sequence, Y3, is separated from the underlying Y2-unit by a distinct erosion surface, which is best developed in the southern part of the area, where locally also some subtle baselaps can be observed in Y3.

The observed thickness reaches about 15 m in the north, but is usually less than 10 m further to the coast, where even in some areas this sequence has been entirely removed by erosion (figures 8 and 9).

In the southern part of the area (figure 9), the seismic facies is characterized by weak, roughly southwards prograding reflectors. Further to the north however, it shows some analogy with the underlying unit with parallel reflectors of variable amplitude. In this area the transition of Y1 into Y2 and Y3, and even into the overlying L1-unit (cf. HENRIET *et al.*, this volume), is very gradual, without noticeable unconformity or abrupt change in seismic facies.

On land, the previously mentioned Egem Member is overlain by the massive, homogeneous clay of the Merelbeke Member. Such a lithology could account for the fact that on the offshore isobath map of the base of the Quaternary (MOSTAERT *et al.*, this volume) the Y3-outcrop often appears as a significant positive feature. Furthermore, the boundary between the Ieper Formation and the Vlierzele Formation is known as an important unconformity (STEURBAUT and NOLF, 1986) ; a similar consideration can be made for the top of the Y3-sequence. Therefore a correlation of Y3 with the Merelbeke Member is suggested.

THE YX DEPOSITIONAL SEQUENCE

The last depositional sequence of supposed Ypresian age, YX, occurs only locally, as a complex fill of an erosive depression (DE BRUYNE, 1984).

This remarkable erosion feature is located about 20 km north of Ostend. It has a predominantly westsouthwest-eastnortheast orientation and it was scoured into the underlying units Y3, Y2 and even into the top of the Ieper Clay, over a width of more then 10 km. The exact shape of this erosive structure and its extent towards the east remains uncertain, due to its burial under younger Tertiary deposits, which leave it beyond the range of high-resolution seismic probing. Isopach evaluations (figure 9) suggest a circular, rather than an elongated shape.

YX can be divided into three subsequences in a southeasterly prograding basin-fill configuration, respectively named YXa, YXb and YXc (figure 8). Their seismic facies (see also figure 6 in HENRIET *et al.*, this volume) consists of abundant, tangential or parallel obliquely prograding reflectors. A detailed analysis of the internal structure and lapout patterns of Y3, the three YX-subsequences and the overlying L1-sequence (figure 9), shows a progradation, swinging from a southsoutheasterly over a southeasterly to an almost easterly direction. This rotation is associated with a southeasterly migration of the depocentres.

Due to be isolated character of this depositional sequence, it is not easy to link it with any onshore lithostratigraphic unit. However, there are some striking analogies between these deposits of limited areal extent and typical seismic facies, with the abundantly cross-bedded, ravinating Vlierzele Formation in the coastal plain (MOSTAERT, 1985). The typical stratigraphical configuration also recalls some parallelism with the Brusselian event in central Belgium, although on another scale and in an apparently different stratigraphical position.

CONCLUSION

On base of a detailed analysis of the offshore seismic data off the Belgian coast and a comparison with the onshore geology, it has been possible to establish a tentative correlation of the identified seismic-stratigraphic depositional sequences and the recently revised Ypresian lithostratigraphical units.

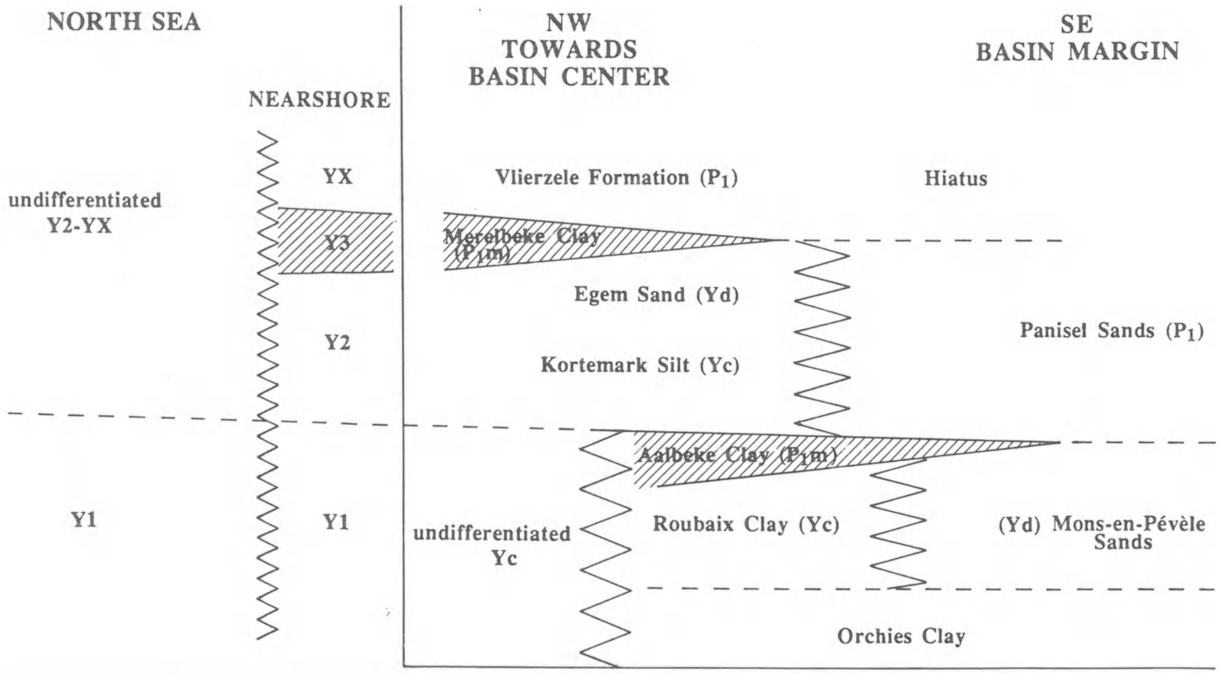


Figure 10 Stratigraphic correlation chart of the Ypresian in the Belgian Basin, ofshore and onshore.

This correlation is presented on a litho-, chrono- and seismic-stratigraphic correlation chart (figure 10). Although all observations indicate that for instance the short-range lateral and vertical changes in lithofacies frequently observed in the Flanders area and offshore close to the coast tend to become less explicit towards the palaeobasin's centre, future borehole control will still be necessary for ascertaining the exact sedimentological composition as well as geologic age of all identified seismic-stratigraphic units.

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9 Neogene deposits of the UK sector of the southern North Sea (51°-53°N)

P.S. BALSON

British Geological Survey, Nicker Hill, Keyworth, Nottingham NG12 5GG, United Kingdom.

ABSTRACT

Neogene deposits have a limited distribution in the UK sector of the Southern Bight of the North Sea. Miocene deposits onshore are restricted to ferruginous sands within solution pipes in North Kent (Lenham Beds) and nodules of apatite-cemented sandstone ('box-stones') found within the conglomeratic phosphorite deposits at the base of the East Anglian Crags. Pliocene deposits are more widespread and have been sampled offshore in boreholes and vibrocores. Equivalents of the onshore Lower Pliocene Coralline Crag and Upper Pliocene/Lower Pleistocene Red Crag have been identified. Preservation of pre-Red Crag Neogene deposits in this area appears to be largely due to resistance to erosion caused by early lithification of the sediment or protection within solutional or erosional hollows.

INTRODUCTION

Neogene deposits are known to have only a limited extent in eastern England but a much more complete sequence is known to be present in Belgium and the Netherlands on the eastern margin of the southern North Sea basin. Until recently little was known of the extent or lithologies of Neogene formations from the offshore southern North Sea area. Between 1978 and 1984 the British Geological Survey completed a survey of the UK sector of the southern North Sea which has enabled the distribution of Neogene formations to be outlined and the lithologies identified. The survey consisted of a grid of seismic traverses with a line spacing typically between 8 and 10 km. Formations identified on the seismic records were sampled using a variety of coring techniques ranging from short cores around a metre long to shallow boreholes to a depth of approximately 110 m. Further details of the BGS survey in this area can be found in BALSON and CAMERON (1985).

MIOCENE DEPOSITS

To date no Miocene deposits have been found in the UK offshore sector of the Southern Bight of the North Sea. Pliocene deposits offshore rest with a conspicuous unconformity on Palaeogene (Eocene) formations. On the adjacent land area of eastern England only scattered remnants of Miocene formations are found. The best known of these are the Lenham Beds of North Kent (figure 1, A). The Lenham Beds consist of ferruginous sands and ironstones found within solution pipes into the underlying Upper Cretaceous Chalk at localities around Lenham to the east of Maidstone. The sands have been leached and contain no calcareous fauna but a large fauna of marine molluscs is known from shell moulds. The relatively poorly preserved fauna and the elevation of the Lenham Beds over 180 m above present day sea level has led to speculation as to their exact age. RASMUSSEN (1966) interpreted the molluscan fauna as Pliocene while SHEPHARD-THORN (1976) has suggested that the Lenham Beds sediments were glacially transported during the Pleistocene to their present position. CURRY *et al.* (1978) however, believed that the Lenham assemblages are most likely correlatable with the Belgian Deurnian (Diestian) and are thus of Late Miocene age.

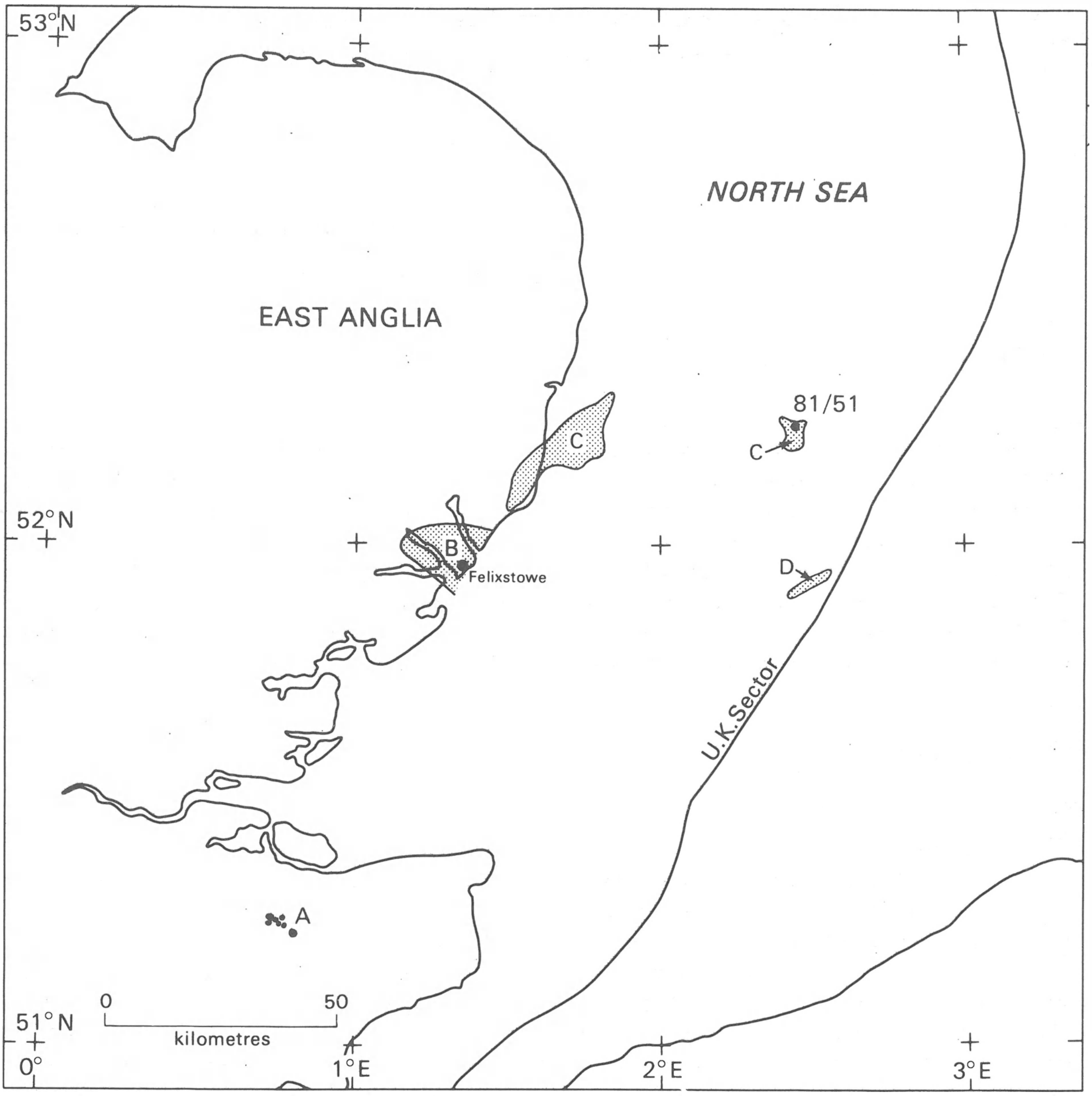

Figure 1 Map showing occurrence of Miocene and early Pliocene deposits in eastern England and the UK sector of the southern North Sea.
A = outcrop of Miocene Lenham Beds. B = area of occurrence of Miocene phosphatic sandstone concretions ("box-stones") in the Crag phosphorite deposits and inferred area of outcrops of 'Trimley Sands'. C-D = outcrop of Early Pliocene.

The second Miocene deposit of which there is evidence in eastern England is found only as remnants within the conglomeratic, phosphatic remanié deposits ('coprolite bed' or 'Suffolk Bone-bed' in LANKESTER, 1868), at the base of the East Anglian Coralline and Red Crags (Pliocene). These remnants occur as rounded nodules, 5-20 cm in diameter, of an apatite-cemented sandstone . The nodules contain a marine fauna including many species of mollusc although, as in the case of the Lenham Beds, subsequent dissolution of calcium carbonate has left only moulds of the molluscan shells. The nodules are believed to have formed as early diagenetic concretions around sites of organic decay at shallow depths in a deposit of muddy sands (BALSON, 1980). These nodules were termed 'box-stones' by LANKESTER (1870) after the use of the word 'boxes' applied by quarry workers during the mid-19th century phosphate exploitation of these deposits. The term referred to those nodules which broke open like a 'box' to reveal a shell mould. The name 'box-stone' is an inappropriate term for these nodules for reasons summarised by BALSON (in preparation) who proposes that the term 'Trimley Sands' be applied to the deposit of muddy sands of which these concretions are the sole known surviving remnants. These nodules have a relatively limited geographical extent within the more widely distributed Crag phosphorite deposit. The distribution of these nodules (figure 1, B) is believed to be closely related to the original depositional extent of the 'Trimley Sands'. There has been little modern work on the correlation of the 'Trimley Sands'. LANKESTER (1868 ; 1870) correlated the 'box-stone' fauna with the Black Crag of Antwerp known more recently as the Antwerp (Antwerpen) Sands of Middle Miocene (Anversian) age. Some confusion exists arising from the fact that the Black Crag of Antwerp was included within the 'Diestien' during LANKESTER's time. The Diestian stage was later redefined (TAVERNIER and DE HEINZELIN, 1961) as late Miocene and thus no longer includes the Antwerp Sands.

PLIOCENE DEPOSITS

Pliocene deposits are more extensive than Miocene sediments in eastern England and have offshore outcrops in the UK sector of the southern North Sea.

The oldest of these deposits on land is the Coralline Crag of Suffolk believed to be of Early Pliocene age (KING, 1983). The Coralline Crag is a formation of marine, skeletal carbonate sands and muddy sands found in an elongate outcrop near the coast of eastern Suffolk. The outcrop continues offshore as a poorly defined area extending some 14 km from the coast (figure 1, C). Vibrocores taken a short distance offshore where the Coralline Crag is exposed at, or very close to, seabed have proved the same facies as that seen onshore; a limonite-stained molluscan or bryozoan-rich coarse skeletal sand. In onshore sections the Coralline Crag consists of a lower portion of unaltered and largely unlithified calcareous muddy sands overlain by a lithified, calcite-cemented, porous limestone from which aragonitic debris has been selectively removed by dissolution and which typically shows large scale cross-stratification. The aragonite dissolution and consequent lithification is believed to have been largely facies controlled, affecting mainly the coarser, porous, cross-stratified facies. This selective lithification may have served to preserve the morphology of a Pliocene sand bank or banks (BALSON, 1983). Stable isotope evidence (^{13}C and ^{18}O) together with field and evidence suggests that the diagenetic alteration and lithification of the Coralline Crag occurred soon after deposition and may have taken place under marine conditions possibly even before the regression which ended Coralline Crag deposition. This early lithification may have been very important in preserving the Coralline Crag sediments as the unlithified lower facies are almost always overlain by the more resistant lithified facies. Without this resistant 'cap' the entire formation might have been eroded during the regression.

The lithified upper portion now forms a low topographic ridge onshore which dips to the northeast and continues offshore where it is overlapped by more recent Crag sediments. The Coralline Crag offshore shows the same limonitic, yellowish-orange staining and selective aragonite dissolution indicating that these are not modern subaerial weathering phenomena. Shallow seismic records across the offshore Coralline Crag outcrop show gently dipping internal reflectors which may support a sand bank model for the unit's deposition. The seismic profiles indicate a total thickness of about 25 m for the formation.

Further offshore, BGS Borehole 81/51 (figure 1) recovered 16.3 m of Lower Pliocene sediments (CAMERON *et al.*, 1984) which may be equivalent in age to the Coralline Crag. The lower limit of this small lens of Lower Pliocene was not penetrated but from seismic evidence it is believed that the 16.3 m recovered may be close to the unit's total thickness. The extent of the lens is not accurately known; a rough approximation is shown in figure 2. It does not appear to be continuous with the Coralline Crag outcrop to the west. The Lower Pliocene sediment in Borehole 81/51 is an olive-green, glauconitic muddy sand in which the sand fraction (55-65 % of the total sediment) is well sorted, fine to very fine ($Mz = 2.75 - 3.57\ \phi$). The sediment has a variable $CaCO_3$ content of between 18 and 44 % and is thus much less calcareous than the Coralline Crag sediment onshore where the $CaCO_3$ content may be over 90 % and is generally $>$ 50 %. Macrofauna is scarce but includes the gastropod *Nassarius (Amyclina) labiosus*. The dinoflagellate assemblage was described by CAMERON *et al.* (1984) although the flora is not significantly different from the overlying Red Crag formation. The assemblage indicates an outer neritic environment of deposition.

About 30 km to the south of this outcrop is a further small isolated body of Neogene sediment infilling a NE-SW aligned valley eroded into Eocene formations (figure 1, D). The valley is approximately 2-5 km across, 12 km long and is infilled with over 70 m of presumed Neogene sediments. Two vibrocores into the topmost sediments of the infill have recovered glauconitic muddy sands with glauconite grains comprising between 10 and 50 % of the sand fraction. The sand fraction (73-90 % of the total sediment) is moderately well sorted and fine grained ($Mz = 2.48\text{-}2.57\ \phi$). $CaCO_3$ content is only 7-15 %. In one sample an abundant macrofauna included molluscs, and fragments of bryozoans, echinoids and *Terebratula*. The fauna appears superficially similar to that of the lowermost Coralline Crag facies onshore. The mollusc and bryozoan material appears relatively unabraded although corroded to a greater or lesser degree. The microfauna includes abundant Foraminifera which indicate an Early Pliocene age for the sediment and are comparable with assemblages of the Luchtbal Sands in Belgium and lower facies of the Coralline Crag of the UK mainland (Personal communication, M. HUGHES, 1986).

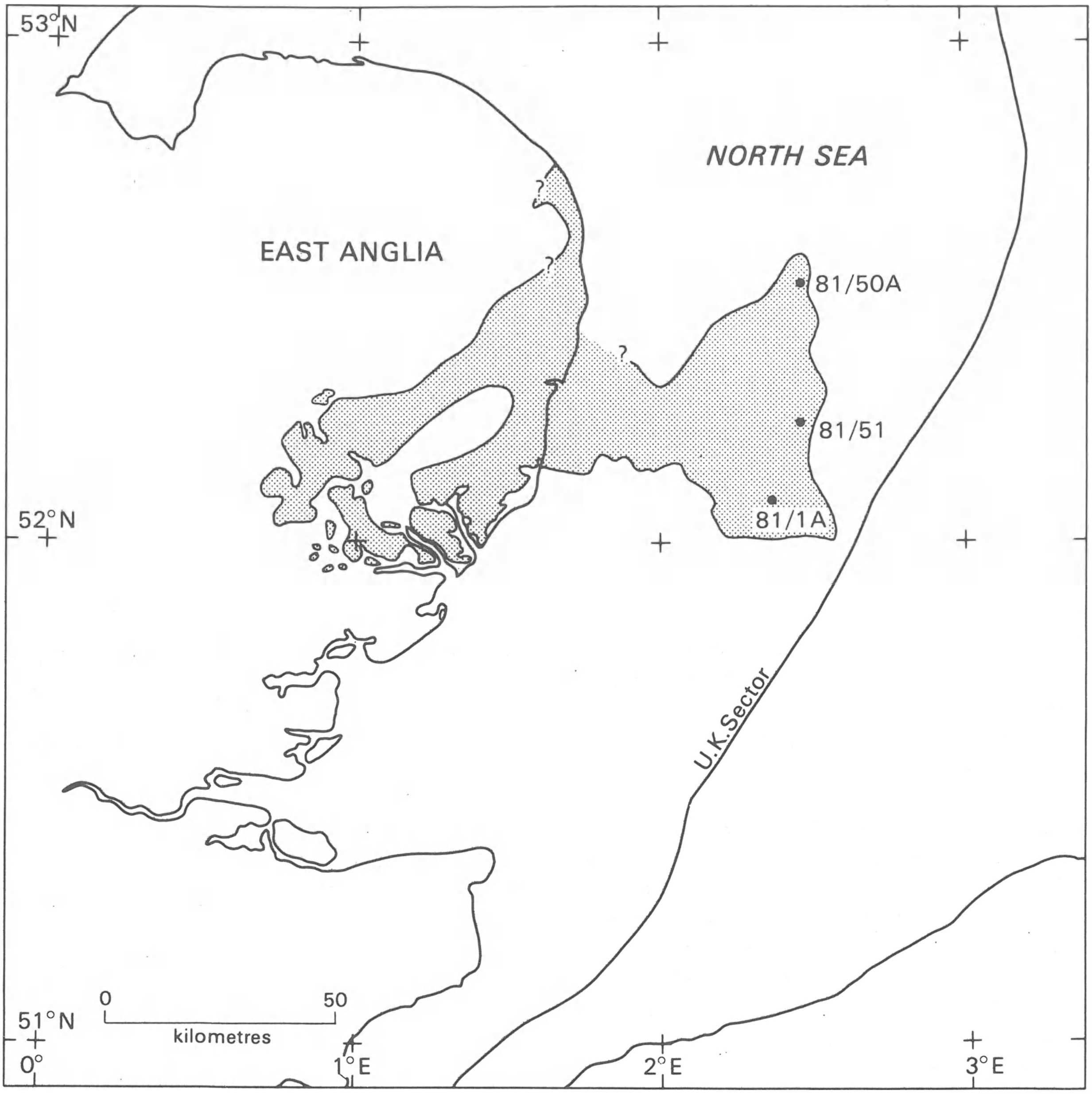

Figure 2 Map showing outcrops of Late Pliocene deposits (Red Crag Formation) in eastern England and the UK sector of the southern North Sea.

The age of the Red Crag of eastern England has long been a source of controversy. The base of the Pleistocene has variously been placed at the base of this formation (e.g. BADEN-POWELL, 1950 ; BOSWELL, 1952), within it (e.g. VAN DER VLERK, 1950) or above it (e.g. CURRY *et al.*, 1978 ; CAMERON *et al.*, 1984). For the purposes of this paper the Red Crag is considered as Pliocene in accordance with the most recent studies which correlate the southern North Sea Plio-Pleistocene stratigraphy with the palaeomagnetic time scale.

Onshore the Red Crag covers a large area of East Anglia (figure 2) and consists of reddish-brown (hence the name) limonite-stained shelly sands and sandy gravels. A number of sedimentary facies are present ranging from marine sand wave facies with large-scale cross-stratification to nearshore tidal flat and even beach deposits.

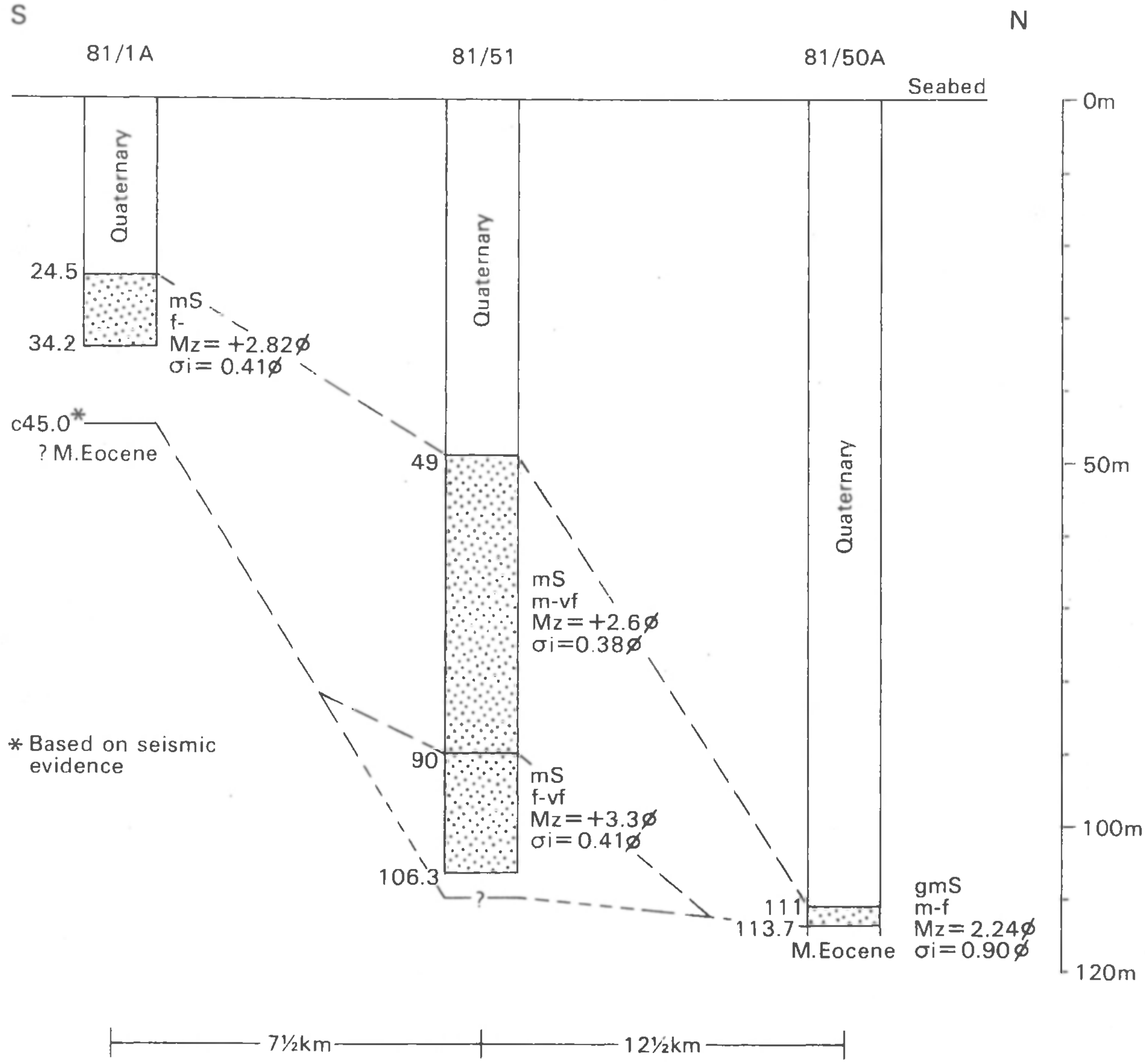

Figure 3 Summary logs showing lithologies and correlation of Pliocene sediments in BGS Boreholes 81/1A, 81/51 and 81/50A. Grain-size parameters are those of FOLK and WARD (1957).
FOLK & WARD, 1957. Brazos River Bar, a study in the significance of grain-size parameters. *J. Sediment. Pet.* 27, 3-26.

Offshore vibrocores off the coast of Suffolk have recovered similar shelly sand sediments. These are usually olive-grey in colour but may be locally oxidised near the seabed to the same red-brown colour seen onshore. Three BGS boreholes further to the east have recovered sediments believed to be equivalent in age to the Red Crag (CAMERON *et al.*, 1984), underlying a Quaternary cover which thickens to over 110 m at the northern limit of the outcrop (figure 3). The boreholes proved a maximum thickness of 51 m of sediment consisting mostly of olive-green, glauconitic muddy sand where sand may constitute between 72 and 91 % of the sediment. The sand fraction is fine (2.33-2.95 ϕ) and well sorted. In Borehole 81/51 there was a slight tendency to coarsening upwards e.g. from Mz = 2.77 ϕ at 84.3 m to 2.33 ϕ at 54.85 m. The sediments have a $CaCO_3$ content of between 34 and 39 %. In Borehole 81/50A at the extreme northern limit of the outcrop the Red Crag sediments consisted of 1.7 m of gravelly muddy sand with abundant small pebbles of phosphatic mudstone at the base overlying Middle Eocene muds. This basal facies is similar to that seen at the base of the Crags onshore (BALSON, this volume). The Red Crag Formation shows dipping internal reflectors on seismic profiles with some minor erosion surfaces and a possible bank structure on one profile (CAMERON *et al.*, this volume).

SUMMARY

The Neogene succession of the UK sector of the southern North Sea is characterised by deposits of shallow marine sediments punctuated by stratigraphic gaps. These gaps make accurate correlations difficult. In the Netherlands on the eastern margin of the late Tertiary basin a thick and much more complete sequence of marine sediments is known. The Breda Formation (mainly Miocene) reaches a thickness of 60 m and the overlying Oosterhout Formation (mainly Pliocene) is up to 176 m thick (NEDERLANDSE AARDOLIE MAATSCHAPPIJ B.V. and RIJKS GEOLOGISCHE DIENST, 1980). There is evidence that offshore in the Netherlands sector of the southern North Sea the Miocene may be over 100 m thick (BRITISH GEOLOGICAL SURVEY and RIJKS GEOLOGISCHE DIENST, 1984).

Syn-depositional tectonic subsidence which continued and increased into the Pleistocene (ZAGWIJN and DOPPERT, 1978) allowed a thick Neogene sequence of over 200 m of shallow marine sediments to accumulate in the centre of the basin (east of median line) while uplift of the basin margins resulted in periods of erosion and non-deposition in the marginal areas (e.g. eastern England and UK sector of southern North Sea). Neogene marine sediments were deposited in the UK sector during transgressive episodes but were subsequently probably largely removed by erosion. The preservation of the pre-Red Crag Neogene sediments in this area appears to be largely fortuitious. The Lenham Beds have been preserved by being let down into solutional hollows which formed in the underlying Chalk. The 'Trimley Sands' have been almost totally destroyed leaving behind only the previously enclosed large, dense phosphorite concretions as a component of a winnowed lag gravel deposit (the Crag phosphorite deposits). Preservation of much of the Coralline Crag onshore appears to be due to early carbonate lithification, while offshore early Pliocene sediments have been preserved only within large erosional hollows.

ACKNOWLEDGEMENT

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10 Upper Pliocene and Lower Pleistocene stratigraphy in the Southern Bight of the North Sea

T.D.J. CAMERON (1), C. LABAN (2) and R.T.E. SCHÜTTENHELM (2)

1. British Geological Survey, 19 Grange Terrace, Edinburgh EH9 2LF, United Kingdom.
2. Rijks Geologische Dienst, Spaarne 17, Haarlem, The Netherlands.

ABSTRACT

The British Geological Survey and Rijks Geologische Dienst have jointly compiled a 1:250,000 map of the Quaternary sediments of the Southern Bight of the North Sea between 52° and 53°N and 2° and 4°E. The Upper Pliocene and Lower Pleistocene succession in this area contains an upward transition from marine to deltaic and fluviatile sediments and is more than 350 m thick at 53°N, 3°30'E. The succession is clearly punctuated by unconformities on seismic profiles, defining the boundaries between six separate seismo-stratigraphic units. Each unit has characteristic lithology and seismic facies and there are significant differences in one or more components of the analysed fossil assemblages of dinoflagellate cysts, foraminifera, or pollen and spores between succeeding seismic formations. The unconformities each indicate an interval of stratigraphic hiatus within the Lower Pleistocene offshore. The total succession describes a history of repeated and widespread marine transgression and regression in the Southern Bight area during the early Pleistocene.

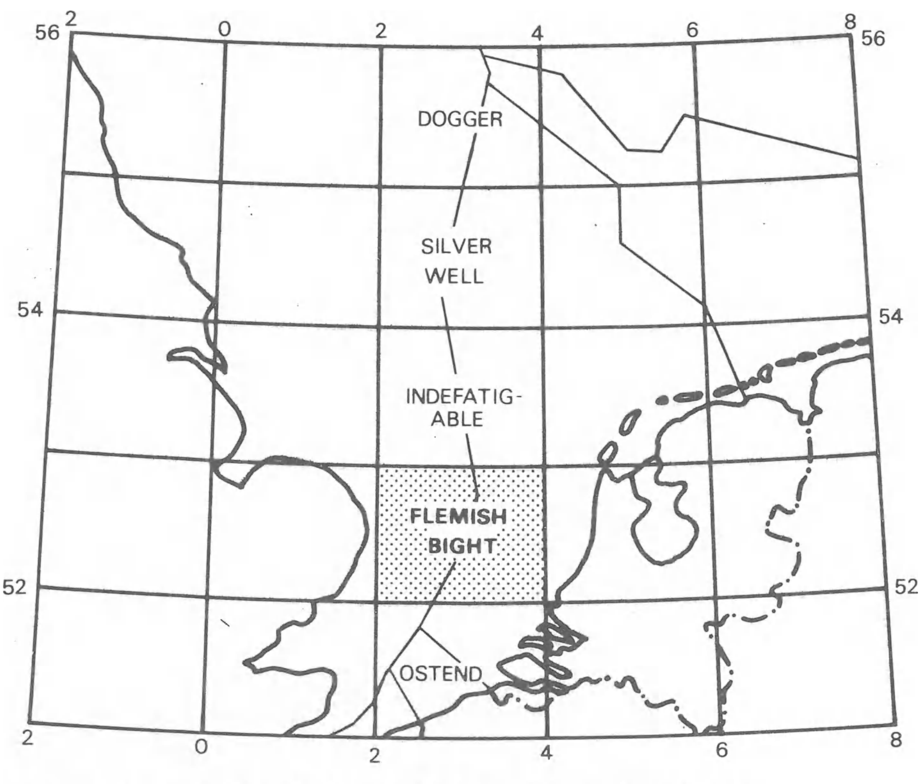


Figure 1 Map showing the location of the Flemish Bight sheet in the southern North Sea.

INTRODUCTION

Since 1980, the British Geological Survey and Rijks Geologische Dienst have been collaborating in the compilation of geological maps at a 1:250,000 scale of the UK and Netherlands sectors of the southern North Sea. Maps illustrating the Solid Geology, Quaternary Geology and Sea Bed Sediments of the Flemish Bight sheet (52°-53°N, 2°-4°E) are now published and it is planned that sets of the remaining sheets which straddle the international median line (figure 1) will be published progressively between 1985 and 1990.

The Quaternary map of Flemish Bight shows the subcrop of Pleistocene and older sediments beneath a cover of Holocene sediments. Eocene, Upper Pliocene and Lower Pleistocene sediments are at or close to the sea bed in the southwestern quadrant (figure 3) while Middle and Upper Pleistocene sediments cover the remainder of the sheet and are described separately (CAMERON *et al.*, this volume).

The Flemish Bight map is based on reconnaissance seismic and sampling surveys, carried out in the UK sector mainly between 1978 and 1981 and in the Netherlands sector between 1968 and 1982. The Pleistocene has been subdivided into a succession of seismo-stratigraphic units. Each unit displays unconformable relationships with the preceeding unit across the grid of seismic profiles and each has characteristic lithology and seismic facies. The seismic units have been given formation status on the map and new formation names, except where there is good correlation with established onshore stratigraphy.

The stratigraphic relationships and seismic facies of the formations have been determined from high resolution seismic profiles in the UK sector of Flemish Bight and examples of these are illustrated in figure 5. A 4 kJ sparker or 5 cubic inch air gun and 500 J or 1 kJ sparker were operated along a network of north-south and east-west seismic lines between 6 and 10 km apart (figure 2). The higher energy source yielded geological information from the sea bed to the base of the Tertiary in most lines, with a maximum penetration of 900 ms (two-way time), through between 700 and 800 m of sediment. The lower energy sparker provided better resolution of seismic reflectors in the uppermost 100 ms (80 - 90 m) of the profiles. A comprehensive sparker survey of the Netherlands sector is planned for 1986, but interpretation of the two

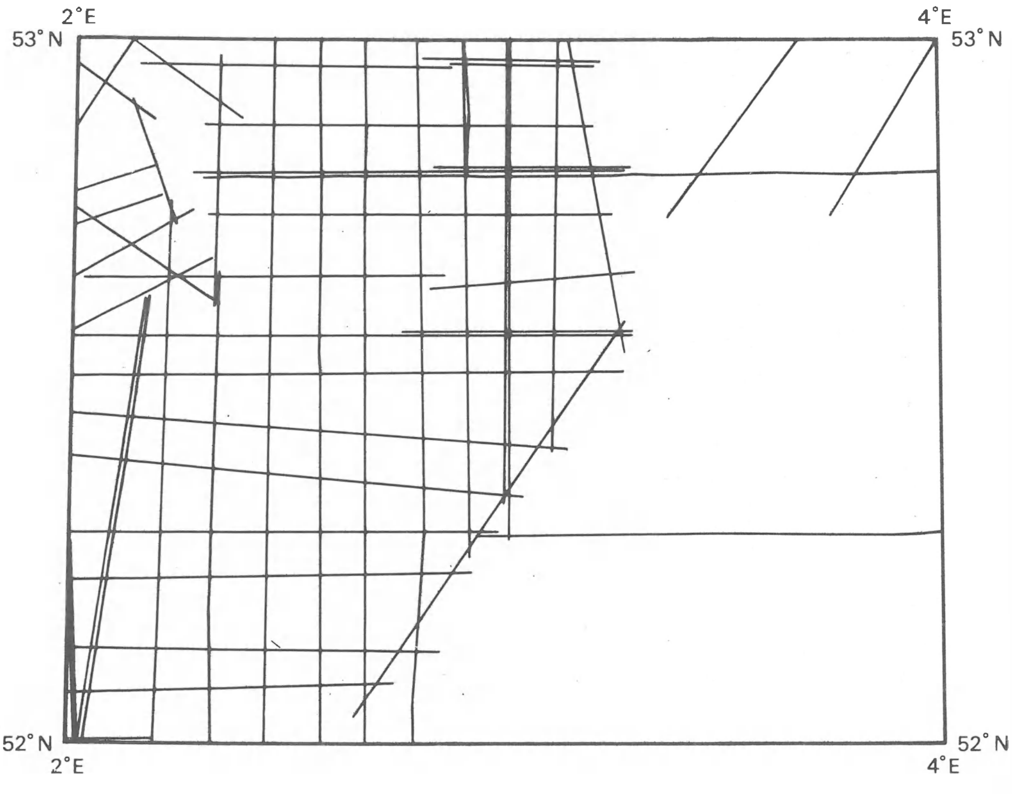


Figure 2 Map showing sparker and air gun survey lines in Flemish Bight.

east-west and two northeast-southwest profiles presently available has already confirmed the stratigraphic relationships observed in the western half of the sheet.

The unprocessed seismic returns comprise a composite of primary and multiple reflectors, but the multiples are easily identified ; they bear an imprint of the uneven sea bed topography characterised by sand waves between 3 and 12 m in amplitude in Flemish Bight. Depths and thicknesses on the geological sections and isopachyte maps have been calculated from the seismic profiles using an average acoustic velocity of 1.82 km/s for the Quaternary, derived by comparing the recorded thickness of the sediments in adjacent commercial boreholes with their apparent thickness in the profiles. Water depths in Flemish Bight are mostly between 20 and 60 m.

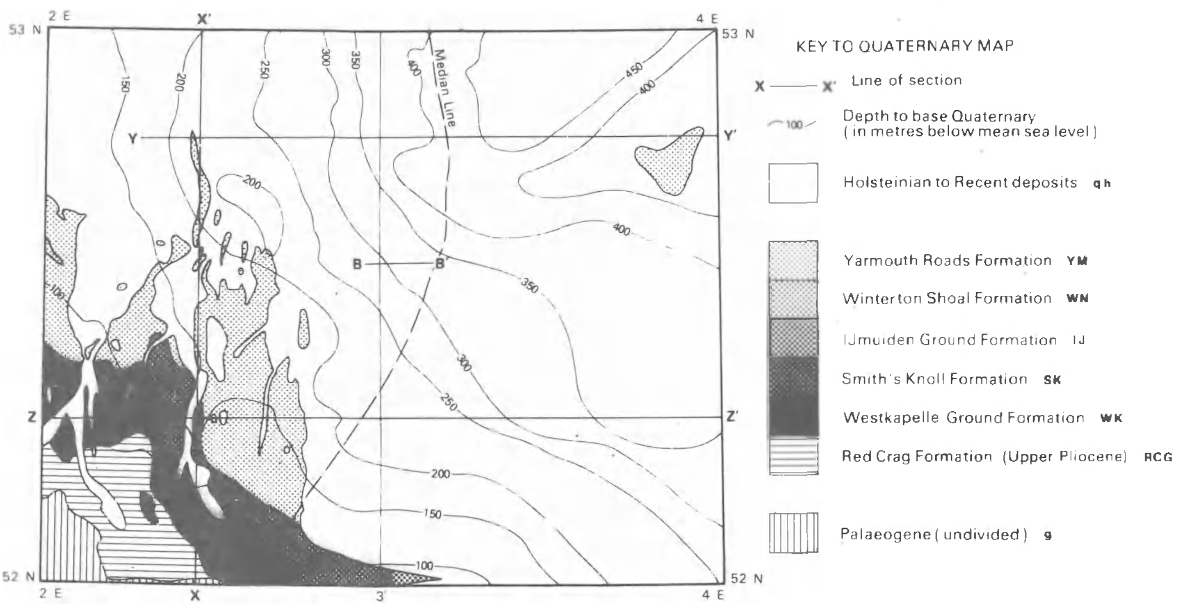


Figure 3 Quaternary Geology of Flemish Bight sheet simplified from published 1:250,000 map.

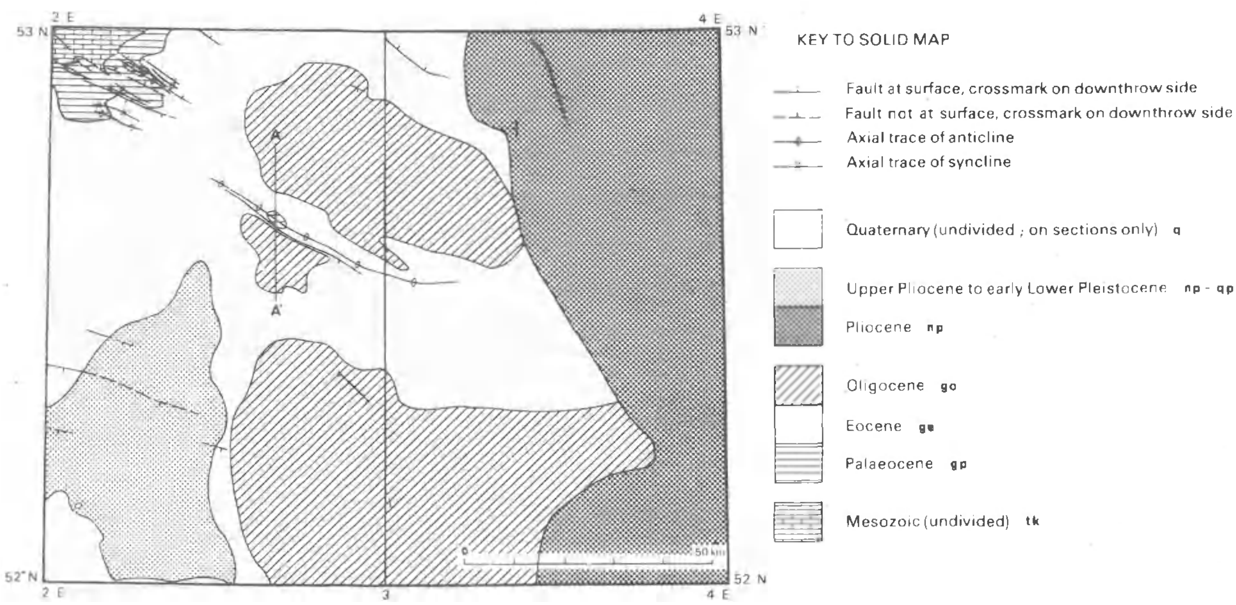
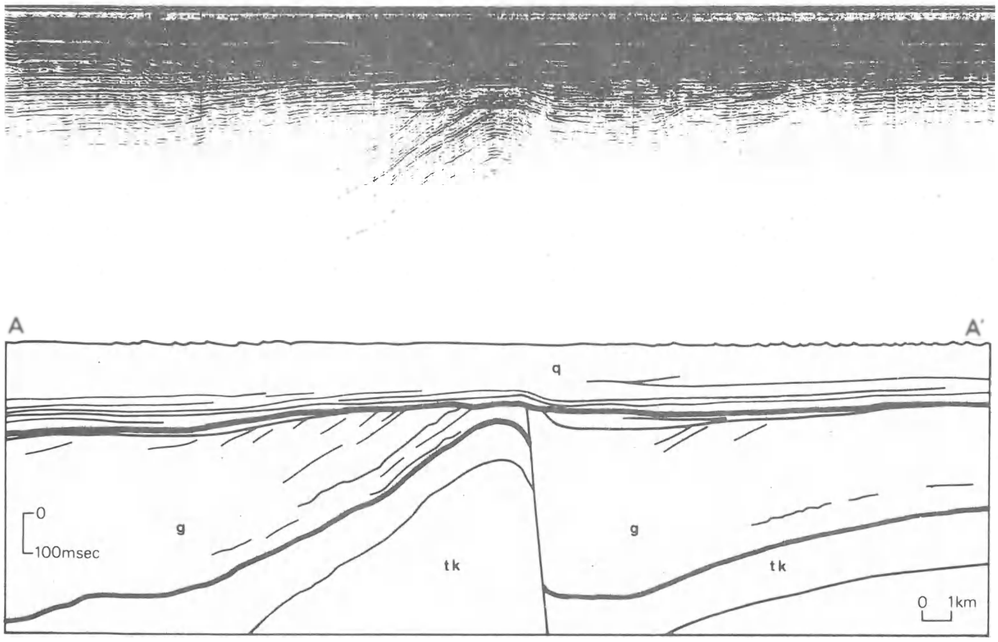


Figure 4 Solid Geology of Flemish Bight sheet simplified from published 1:250,000 map.

a)



b)

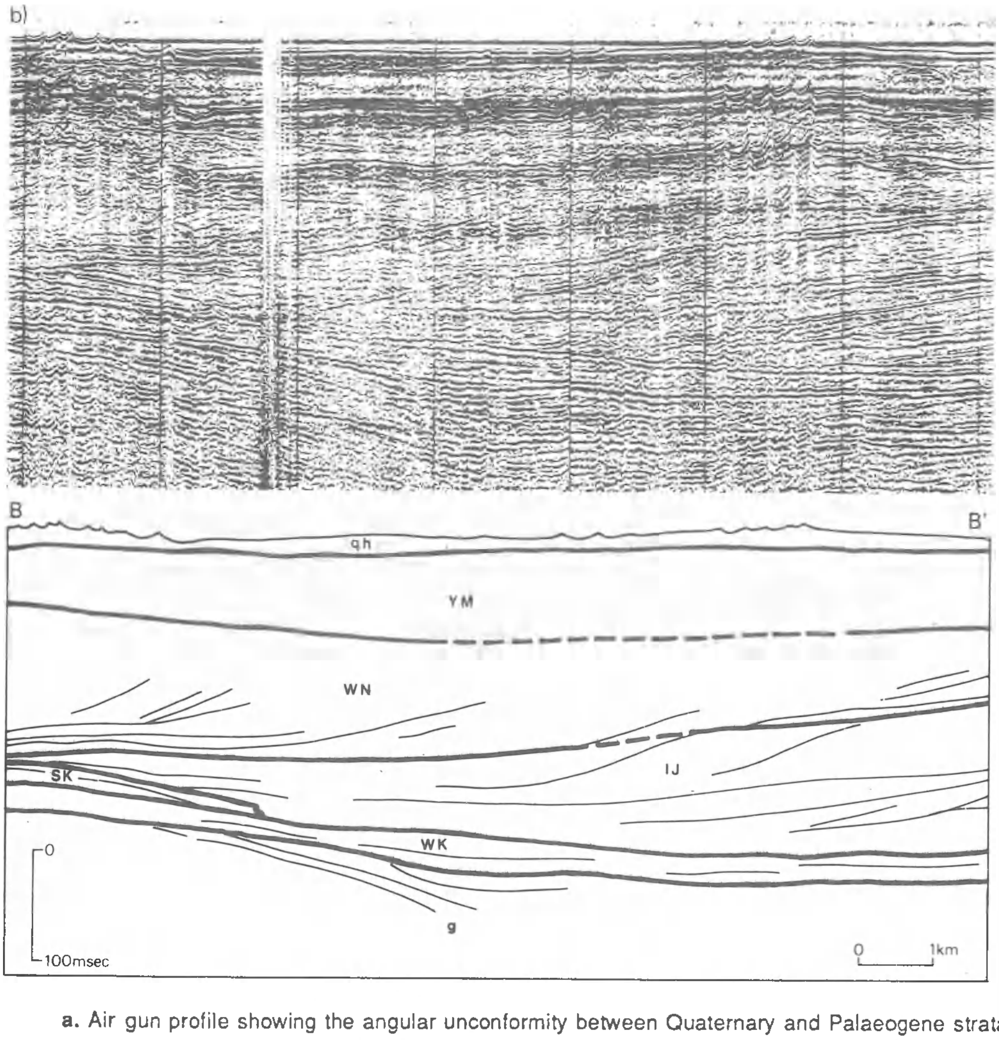


Figure 5 **a.** Air gun profile showing the angular unconformity between Quaternary and Palaeogene strata in Flemish Bight.
b. Air gun profile showing the relations between some of the Lower Pleistocene formations in the centre of the Flemish Bight.

The Lower Pleistocene sediments have been sampled in four boreholes drilled by the British Geological Survey in 1981, and detailed analyses of the dinoflagellate cyst, foraminiferal and pollen assemblages of each formation have been published by CAMERON *et al.* (1984). Three of the formations have been additionally sampled in the southwest in vibrocores, with a maximum penetration of 6 m, and these have yielded valuable information on the lithology of the youngest unit, the Yarmouth Roads Formation, which is poorly represented in the boreholes. Samples of the uppermost layers of the early Pleistocene were obtained from six non-confidential boreholes in the Netherlands sector.

The Quaternary rests on Oligocene or gently folded and faulted Palaeocene and Eocene sediments in most of the UK sector of Flemish Bight, but overlies Mesozoic sediments in the northwest and Upper Pliocene sediments in the southwest, and rests on an eastward thickening accumulation of Pliocene and Miocene sediments in the Netherlands sector (figure 4). Where the base of the Quaternary rests on Eocene or older sediments, there is clearly an angular unconformity on seismic profiles (figure 5a) ; where it rests on Oligocene sediments, the unconformity is only slightly angular. On the down-hole logs of commercial boreholes in the UK sector, the unconformity is commonly emphasised by an anomalously high gamma peak suggesting a concentration of derived phosphate pebbles or glauconite in a lag deposit at the base of the Pleistocene.

Upper Pliocene seismic reflectors are truncated by the base Quaternary surface in the southwest, confirming that the Pliocene/Pleistocene boundary is a slightly angular unconformity in the UK sector. The unconformity is less clear in the Netherlands sector, and the transition between late Pliocene and early Pleistocene sediments on the eastern margin of the sheet may be conformable, as in the Netherlands (VAN STAALDUINEN *et al.*, 1979).

The Quaternary thickens northeastwards to more than 400 m within Flemish Bight (figure 3), and its basal unconformity has an average dip of 1:270 towards the northeast. There is no clear evidence to indicate that faults displacing Tertiary and older sediments (figure 4) were reactivated during the Quaternary period in either the UK or Netherlands sectors of Flemish Bight, unlike sheets to the north, where the base of the Pleistocene has been displaced locally by as much as 100 m by reactivation of deep-seated faults on the margins of Permian salt domes.

The seismic facies, stratigraphic relationships and geometry of each formation are considered in ascending stratigraphic order. Although the Red Crag Formation may be entirely Upper Pliocene in age offshore, it is included with the Lower Pleistocene in this discussion in that the equivalent sediments in East Anglia are variously regarded as Upper Pliocene (CURRY *et al.*, 1978), Lower Pleistocene (MITCHELL *et al.*, 1973), or may straddle the Pliocene/Pleistocene boundary (OAKLEY, 1949).

STRATIGRAPHY

a) Red Crag Formation

The Red Crag Formation is an outlier of the extensive Upper Pliocene deposits of the Netherlands sector of Flemish Bight. The formation has an uneven and erosional base, cutting into eastward-dipping Eocene sediments in the southwestern quadrant of the sheet, and into Lower Pliocene calcareous, muddy sands in the vicinity of Borehole 81/51. Its upper boundary is an unconformity, and there is clear seismic evidence of a stratigraphic hiatus between deposition of Upper Pliocene and succeeding Pleistocene sediments in the UK sector (figure 6a). Southwest of the Lower Pleistocene outcrop, the Red Crag Formation subcrops beneath, and is truncated southwestwards against the basal Holocene surface ; the formation may have extended into the Outer Thames area before removal by erosion during the late Pleistocene or recent times. If there was a connection eastwards with the equivalent sediments in the Netherlands sector, the evidence suggests this connection was removed by erosion during early Lower Pleistocene times.

The Red Crag Formation has lenticular geometry, is bounded by convex basal and relatively planar upper surfaces, and has a maximum thickness of 70 m in Flemish Bight (figure 7a). The formation has weak to moderate amplitude reflectors inclined gently towards the northwest on seismic profiles ; locally it is

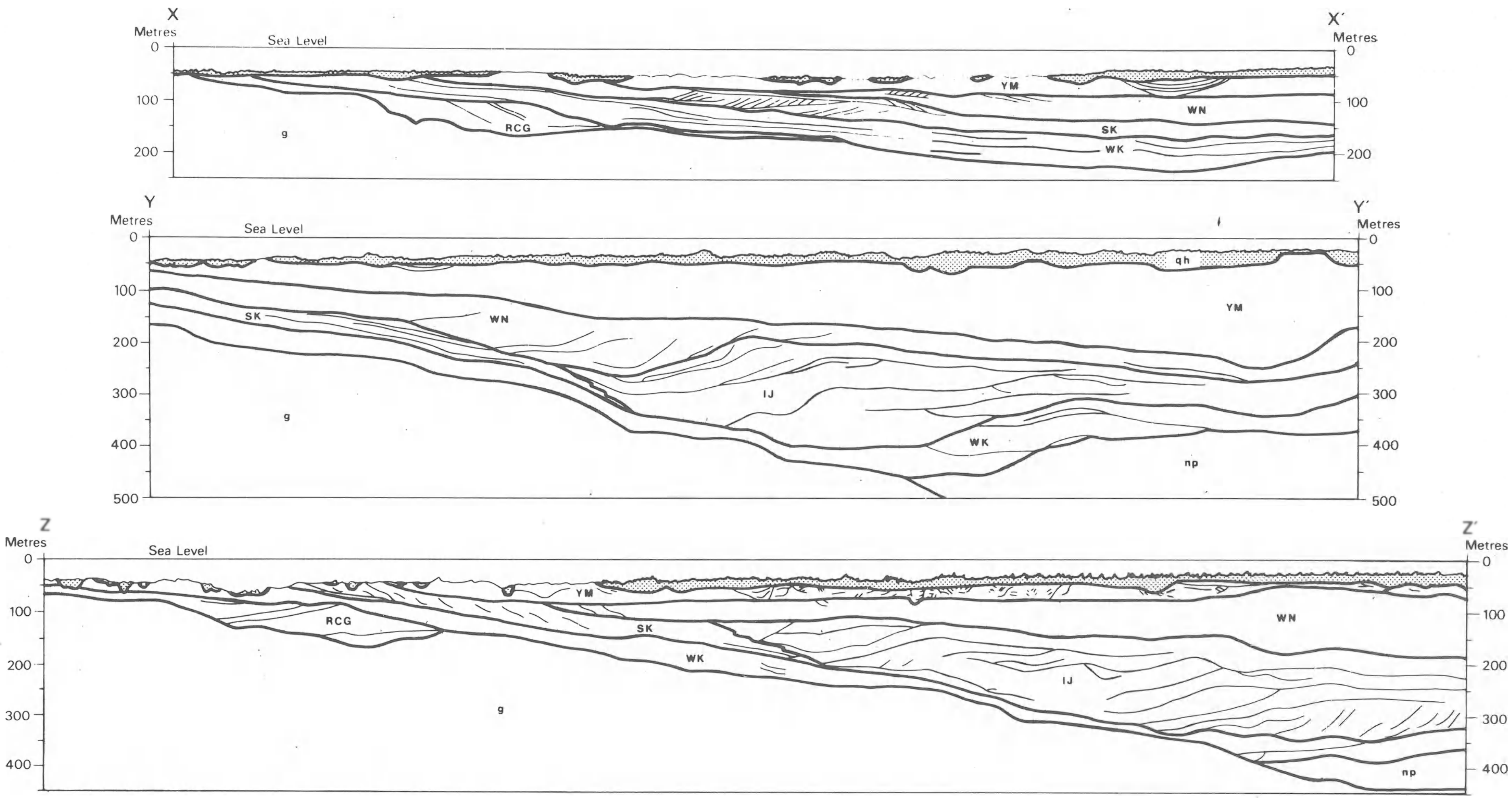


Figure 6 Sections showing the general relations of the Upper Pliocene and Lower Pleistocene formations along the lines indicated in figure 3.

structureless. The reflectors variously describe sigmoid, tangential oblique or parallel oblique progradational patterns (cf. MITCHUM *et al.*, 1977), terminating down-dip mainly towards the northwest ; in the east the reflectors terminate locally by transgressive onlap towards the east or southeast. There are minor erosion surfaces within the formation on some profiles, and a sand bank has been observed on one profile ; this sand bank has an amplitude of 10 m and its internal reflectors have an apparent dip towards the south. The total seismic facies indicates deposition in a high energy, shallow marine sedimentary environment.

In shallow cores in the west, as in East Anglia, the Red Crag Formation comprises fine- to medium-grained, glauconitic sand and contains abundant shell gravel. Unlike East Anglia, the sand typically has an olive-green colour in Flemish Bight, but it is locally oxidised at surface outcrop. Further east, in Boreholes 81/1A and 81/51, the sediments are fine-grained, relatively muddy, and include up to 30 % finely comminuted shell sand. The foraminiferal assemblages in Borehole 81/51 confirm correlation with the Red Crag deposits of East Anglia (FUNNELL and WEST, 1977) and are closely comparable to those of the Lillo Formation in Belgium (DE MEUTER and LAGA, 1976) and to the Upper Pliocene FA_2 subzone of the *Elphidiella hannai-Cribrononion excavatum* foraminiferal biozone in the Netherlands (DOPPERT, 1980).

b) Westkapelle Ground Formation

The sediments of the Westkapelle Ground Formation were deposited during the most extensive transgression of the North Sea into the Southern Bight in early Pleistocene times. The formation overlies Upper Pliocene, Palaeogene and Mesozoic sediments in Flemish Bight ; it is itself overlain in the east by the IJmuiden Ground Formation and in the west by the Smith's Knoll and Yarmouth Roads Formations, and by Holocene deposits. Where it subcrops beneath the Holocene, the Westkapelle Ground Formation is truncated southwestwards against the basal Holocene surface. There are no indications of littoral deposits in seismic profiles or borehole returns in Flemish Bight, and the formation probably extended much further southwest of its present outcrop during Lower Pleistocene times.

The Westkapelle Ground Formation forms a sheet-like deposit between gently undulating basal and relatively planar upper surfaces, is mostly between 25 m and 50 m thick in the UK sector (figure 7b), and has a maximum thickness of 75 m in Flemish Bight. In seismic profiles, the formation is characterised by reflectors of weak to moderate amplitude, parallel or slightly divergent, and laterally continuous for up to 40 km. Some of the reflectors terminate by transgressive onlap onto Tertiary strata, especially south of 52°15'N and in the north-central part of the sheet. There are no reflector truncations against the formation's upper surface. The seismic facies indicates deposition in an open shelf marine environment, and in a significantly lower-energy tidal regime than the Red Crag Formation. There are local sub-units of parallel oblique prograding reflectors on the western periphery of the sheet, inclined towards the eastnortheast, which suggests slightly higher energy sedimentation in the west.

The Westkapelle Ground Formation contains an upward transition from predominantly argillaceous to arenaceous sediment. In boreholes in the southwest, the basal olive-grey or grey-brown silty clay has intercalations of very fine- or fine-grained, glauconitic and extensively bioturbated sand, in beds up to 20 mm thick. The sands thicken and become predominant upwards, become relatively muddy and have a sparse molluscan fauna of thin-walled bivalves ; there are only minor clay intercalations in the topmost 15 m in Borehole 81/50A. Foraminiferal analyses support the lithological evidence of a general shallowing of sea level towards the top of the formation.

Pollen spectra obtained from Borehole 81/50A suggest correlation of the Westkapelle Ground Formation with the early Pleistocene Thurnian Stage deposits in East Anglia (CAMERON *et al.*, 1984). The clays in Boreholes 81/50A, 81/51 and 81/53A have reversed magnetic polarity which thus dates the sediments within the Matuyama epoch, but there are indications of two normal polarity events towards the base of the formation, possibly the "X" and Réunion events. The foraminiferal assemblages in the Westkapelle Ground and succeeding formations in Flemish Bight are markedly different from those in Upper Pliocene sediments in the Netherlands, and are referable to the Lower Pleistocene FA_1 subzone of the *Elphidiella hannai-Cribrononion excavatum* biozone (DOPPERT, 1980).

c) Smith's Knoll Formation

The Smith's Knoll Formation rests unconformably on the Westkapelle Ground Formation in much of the UK sector of Flemish Bight, and is overlain by the Winterton Shoal Formation, or by the Yarmouth Roads Formation in the west above a slight angular unconformity (figure 6c). The formation subcrops beneath and is truncated against the basal Holocene surface in the southwest, and may have extended some distance southwest of its present outcrop during Lower Pleistocene times. The Smith's Knoll Formation passes laterally eastwards into, and was deposited penecontemporaneously with the delta-related sediments of the IJmuiden Ground Formation.

The Smith's Knoll Formation has planar upper and lower surfaces and describes a sheet-like geometry, is mostly between 20 and 50 m thick, and has a maximum thickness of 55 m (figure 6c). The formation has a wedge-like eastern termination under the IJmuiden Ground Formation (figures 6b and 6c) over a zone 5 km wide in the south, but up to 15 km wide in the north. In the south, the upper surface of the wedge is defined by a high amplitude reflector dipping at 1:125 towards the east ; the reflector may indicate a marked lithological contrast between the two formations along their common boundary. In the north, the reflector is inclined at 1:400 or less towards the east and has lower amplitude, suggesting that the formations here are of relatively similar lithology.

Along seismic profiles in the west, the Smith's Knoll Formation is characterised by oblique prograding reflectors, inclined towards the eastnortheast in the south and mainly towards the southeast further north (figure 7c). These inclined reflectors are typically parallel, but locally have tangential down-dip terminations, and on some profiles there are two or even three overlapping sets of inclined reflectors, prograding towards different directions. Further offshore, the formation passes through a reflector-free facies into high amplitude, continuous, parallel or slightly divergent horizontal reflectors, suggesting a transition into the basin from a delta-front to a lower-energy open shelf marine sedimentary facies.

In boreholes and cores in the west, the Smith's Knoll Formation comprises olive-green, muddy, fine-grained, glauconitic sand, with minor intercalations of silty clay and a few beds of gravelly or shelly sand. Their seismic texture suggests that the sediments may become relatively argillaceous towards the northeast. Dinoflagellate cyst and pollen assemblages indicate correlation with the Antian stage deposits of East Anglia, although pollen evidence suggests that the sediments in Borehole 81/50A may not be precise time equivalents of those in the type Antian section at Ludham (WEST, 1961).

d) IJmuiden Ground Formation

The IJmuiden Ground Formation rests unconformably on the Westkapelle Ground Formation in much of the Netherlands sector of Flemish Bight, traces laterally westwards into the Smith's Knoll Formation, and is itself unconformably overlain by the Winterton Shoal Formation. Northwards of 52°50'N, the boundary between the Smith's Knoll and IJmuiden Ground Formations becomes indistinct on seismic profiles, they are both represented by open shelf marine facies, and the sediments have all been referred to the IJmuiden Ground Formation on the Indefatigable sheet to the north.

The IJmuiden Ground Formation is characterised by continuous, high amplitude, sigmoid progradational reflectors on seismic profiles, inclined towards the westnorthwest. Each reflector terminates tangentially down-dip onto the formation's basal unconformity, or terminates by transgressive onlap where the formation oversteps slightly across the basal beds of the Smith's Knoll Formation (figure 5b) in the west. The formation's upper surface truncates relatively few reflectors (figures 6b and 6c), and most reflectors become diffuse eastwards by decrease in amplitude. Each successive pair of reflectors bounds a lenticular sedimentary unit, which is itself either structureless or has low amplitude prograding reflectors dipping more steeply towards the westnorthwest. The total seismic facies describes a transition into the basin from delta-front to prodelta to open shelf marine sedimentary environments westnorthwestwards across the eastern half of Flemish Bight.

The IJmuiden Ground Formation forms a lenticular sedimentary unit ; it is bounded by uneven, but approximately sigmoidal upper and lower surfaces and is made up of a composite overlapping sequence of lenticular units. The maximum sediment accumulation within each successive unit is translated towards the northwest, slightly oblique to the direction of progradation. The formation's maximum total thickness of

200 m is in an axial basin of increased sediment accumulation trending diagonally across the eastern half of Flemish Bight (figure 7d).

The IJmuiden Ground Formation comprises fine- or very fine-grained sands, with intercalations of silty clay probably becoming increasingly abundant towards the west and northwest. In Borehole S1-63 in the Netherlands sector, close to the south of Flemish Bight, the sediments are interbedded sands and clays and have been dated as Upper Tiglian in age.

ZAGWIJN (1974) demonstrated that the widespread early Pleistocene marine regression, which occurred towards the end of the Tiglian interglacial stage in the Netherlands, was caused by a massive seaward expansion of the deltas of the Rhine, Meuse and westward flowing North German river systems into the North Sea Basin ; there are no marine Upper Tiglian deposits in the Netherlands. The authors interpret the major influx of delta-related sediments in the IJmuiden Ground Formation as representing the direct offshore expression of this initial phase of delta growth in the Southern Bight. Northwestward prograding delta-front and prodelta seismic facies are indeed characteristic of Lower Pleistocene sedimentation in the central area of the North Sea, at least as far north as 55°30'N, and delta-related sedimentation continued offshore until the later stages of the "Cromerian Complex" interglacial (ZAGWIJN, 1974). The IJmuiden Ground and Smith's Knoll Formations represent separate supply of sediments from the European and British land areas respectively.

e) Winterton Shoal Formation

The Winterton Shoal Formation rests unconformably on the IJmuiden Ground Formation in most of the Netherlands sector, oversteps westwards across the Smith's Knoll Formation in the UK sector of Flemish Bight, and is overlain by the Yarmouth Roads Formation, with evidence of a slight angular unconformity in the southwest. The formation does not extend to the western margin of the map, but there may be deltaic or estuarine sediments of equivalent age in East Anglia.

Towards its base, the Winterton Shoal Formation is characterised by moderate to high amplitude reflectors on seismic profiles, concave upwards towards the westnorthwest. Each reflector terminates tangentially against the basal unconformity towards the westnorthwest, becomes diffuse eastwards by decrease in amplitude, while the upper part of the formation is structureless. The formation includes deltaic, delta-front and prodelta seismic facies, representing continued expansion of the delta complex of the Rhine, Meuse and North German river systems into Flemish Bight. The formation is predominantly deltaic in the east, and becomes fully marine towards the northwest, in the Indefatigable area. Oblique prograding seismic reflectors inclined towards the northeast are present on some profiles in the west, representing a subsidiary component of sediment supply from the British landmass.

The Winterton Shoal Formation comprises a lenticular sedimentary unit, bounded by an approximately sigmoidal lower surface, and an undulating upper surface inclined gently towards the northeast (figures 6b and 6c). The formation has its maximum thickness of 130 m in an open, north-south trending basin (figure 7e), 25 km west of the thickest accumulation of sediments in the IJmuiden Ground Formation.

The Winterton Shoal Formation comprises fine- and medium-grained sands, silts and clays, and may become relatively argillaceous towards the northwest. One sample obtained from Borehole 81/50, where the formation is only 2 m thick, yielded a pollen spectrum similar to Baventian Stage assemblages in East Anglia (WEST *et al.*, 1980) and the formation may be post-Upper Tiglian in age.

f) Yarmouth Roads Formation

The Yarmouth Roads Formation overlies the Winterton Shoal Formation in most of Flemish Bight, but oversteps across the Smith's Knoll Formation and the uppermost beds of the Westkapelle Ground Formation in the west and southwest respectively (figure 3). The base of the formation is a discontinuous, but locally high amplitude reflector on seismic profiles, becoming continuous north of 53°N, in the Indefatigable area. There is an angular unconformity between the Yarmouth Roads Formation and overlying Middle and Upper Pleistocene and Holocene deposits (figures 6b and 6c). The uppermost beds have been dated as Waalian in age in the west, but are mostly "Cromerian Complex" in age in the Netherlands sector.

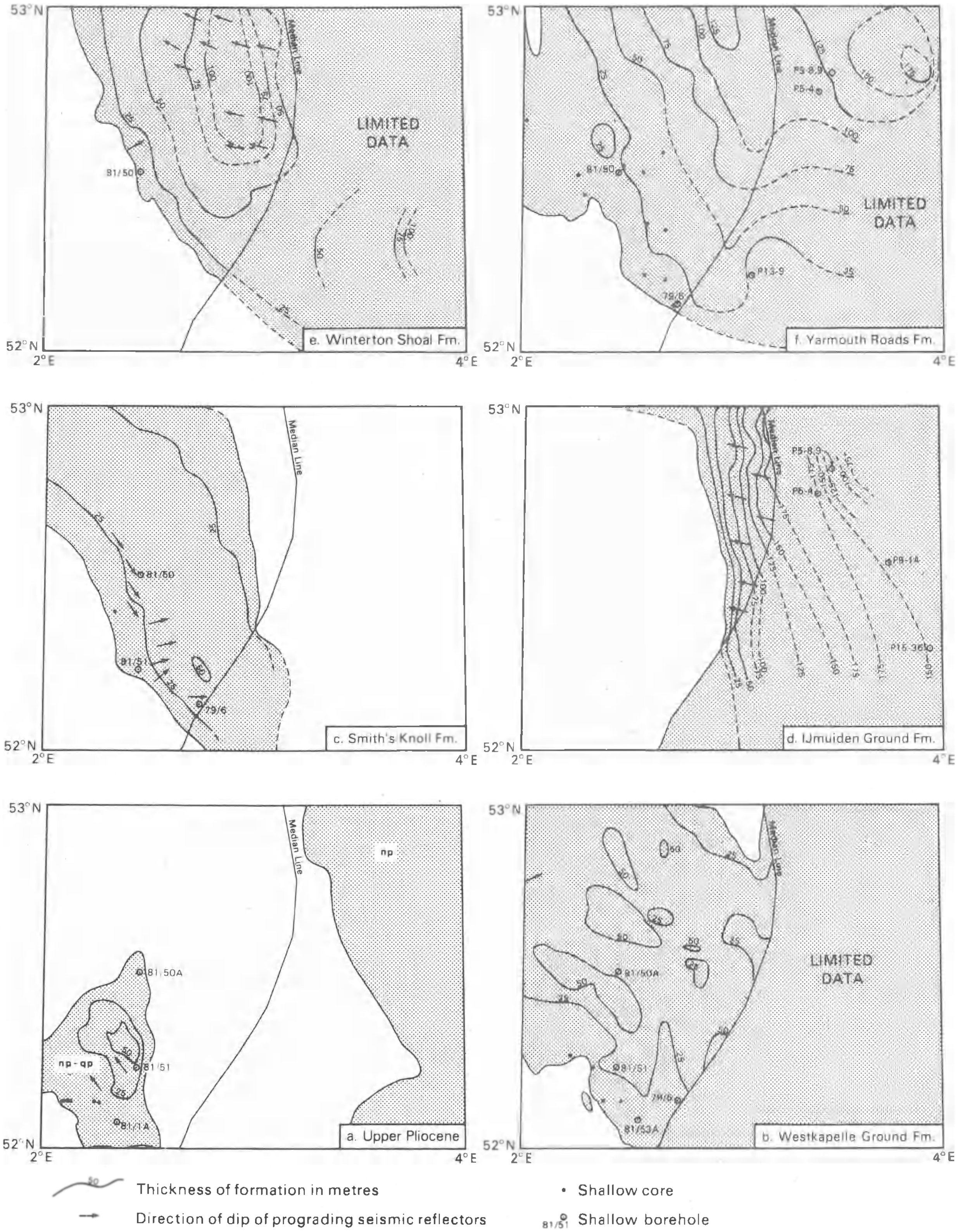


Figure 7 Maps showing the extent and thickness of each of the Upper Pliocene and Lower Pleistocene formations in Flemish Bight.

The Yarmouth Roads Formation is structureless or has chaotic reflector configuration on seismic profiles. Small-scale, poorly-defined channels have been observed on some sparker records. Its seismic texture suggests that the formation comprises a deltaic facies, and though the sediments may be predominantly fluviatile, there may be intercalations of estuarine, intertidal and coastal deposits. The formation was deposited as the Northwest European delta complex expanded across the entire Southern Bight area and possibly amalgamated with river systems flowing eastwards and northwards from the British landmass, during the Waalian Stage. The uppermost Lower Pleistocene sediments comprise a deltaic seismic facies throughout the southern North Sea, at least as far north as 55°N ; the sediments represent the final phase of deltaic infill of the basin, which was initiated during Upper Tiglian times in the south.

The Yarmouth Roads Formation is a wedge-shaped unit, bounded by an undulating lower surface inclined gently towards the northeast, and an upper erosion surface between 40 and 80 m below mean sea level in Flemish Bight. The formation has its maximum thickness of 180 m in the northeastern quadrant of Flemish Bight (figure 7f).

In shallow sea-bed cores in the west, the Yarmouth Roads Formation comprises fine- and very fine-grained, grey-green sand, with laminations of soft, olive-grey silty clay and occasional partings rich in reworked organic detritus. Some of the cores contain layers of mud pebbles and, more rarely, well-rounded cobbles of flint. The carbonate content of the sediments is very low, typically less than 2 %, and although there are local intercalations of marine shelly sand, the formation has yielded no calcareous microfauna.

FORMATION	DINOFLAGELLATE CYSTS	FORAMINIFERA	POLLEN & SPORES
Yarmouth Road Formation			Mixed coniferous-deciduous forests Abundant herbaceous pollen Sparse ericaceous pollen
Winterton Shoal Formation	*Tectatodinium pellitum* *Protoperidinium* spp.	*Elphidium excavatum* *Elphidiella hannai* *Cibicides lobatulus* (Group)	Coniferous forests Abundant herbaceous pollen Abundant ericaceous pollen
Smith's Knoll Formation	*Spiniferites* spp. *Operculodinium israelianum* *Operculodinium centrocarpum* *Tectatodinium pellitum* *Protoperidinium* spp.	*Elphidiella hannai* *Elphidium excavatum* *Cibicides lobatulus* (Group) *Ammonia batavus*	Mixed coniferous-deciduous forests Reduced herbaceous pollen Abundant ericaceous pollen
Westkapelle Ground Formation	*Tectatodinium pellitum* *Operculodinium israelianum* *Spiniferites* spp. *Operculodinium centrocarpum*	*Elphidium excavatum* *Elphidiella hannai* *Cassidulina reniforme* *Cassidulina teretis* *Cibicides lobatulus* (Group)	Coniferous forests Abundant herbaceous pollen Abundant ericaceous pollen
Red Crag Formation	*Tectatodinium pellitum* *Spiniferites* spp. *Operculodinium centrocarpum* *Protoperidinium* spp.	*Elphidiella hannai* *Cibicides lobatulus* (Group) *Elphidium macellum* *Pararotalia serrata* *Elphidium excavatum* *Textularia decrescens*	

Figure 8 Table showing the major components of the fossil assemblages analysed from boreholes in the UK sector of Flemish Bight. The species of dinoflagellate cysts and foraminifera in each formation are listed in approximate order of numerical abundance, though their proportions do fluctuate within all of the formations, and particularly so in the Westkapelle Ground Formation.

DISCUSSION

Upper Pliocene and Lower Pleistocene sedimentation in the Southern Bight was punctuated by four major unconformities, each of which has been identified across a grid of seismic profiles in the Flemish Bight sheet, and each of which separates formations of contrasting seismic and sedimentary facies in the offshore succession. In boreholes in the UK sector, in which five of the six formations have been sampled, the unconformities coincide with significant changes in one or more components of the analysed fossil assemblages of dinoflagellate cysts, foraminifera, or pollen and spores. The boundaries between the formations represent intervals of significant hiatus within the offshore succession, suggesting episodes of widespread regression of the early Pleistocene North Sea from the Southern Bight area.

Detailed analyses of the microfauna and microflora in the boreholes have been published by CAMERON *et al.* (1984) the results are summarised in figure 8.

The assemblages of dinoflagellate cysts are made up of a restricted number of species. Each of the species has a stratigraphic range from Pliocene to Recent in the North Atlantic, but some have not been recorded from deposits younger than Lower Pleistocene in age in the North Sea Basin. There are significant fluctuations in the relative proportions of individual species of dinoflagellate cyst in Flemish Bight, but some of these may relate to changes in oceanic water mass circulation during the Lower Pleistocene period, while some reflect local changes in environment. *Spiniferites*-dominated assemblages do appear to be characteristic of Antian (Upper Tiglian) deposits in East Anglia and in the offshore boreholes, but in general, the dinoflagellate cysts are not sufficiently diagnostic to assist in dating sediments within the Pliocene to Lower Pleistocene period in the Southern Bight.

There are significant differences between the assemblages of benthonic foraminifera of the Red Crag and succeeding formations in Flemish Bight, between Red Crag and Norwich Crag assemblages in East Anglia (FUNNELL and WEST, 1977) and between FA_2, Upper Pliocene, and FA_1, Lower Pleistocene, assemblages in the Netherlands (DOPPERT, 1980). The differences reflect a general decrease in species diversity across the Pliocene/Pleistocene boundary in the southern North Sea Basin. There are minor differences in the relative proportions of individual species between the Lower Pleistocene assemblages in the boreholes and equivalent sediments in East Anglia and the Netherlands, but these most likely reflect local differences in the sedimentary environment, particularly water depth, turbidity or salinity, at the respective sites.

Pollen and spore assemblages have proved particularly useful in correlating marine Lower Pleistocene deposits across the North Sea Basin, but there are difficulties in reconciling the terrestrial, pollen, and marine, dinoflagellate cyst and foraminiferal, evidence on early Pleistocene palaeoclimate (CAMERON *et al.*, 1984). Oceanic heath and herbaceous pollen taxa are well-represented throughout the succession in Flemish Bight and in equivalent sediments in East Anglia (WEST, 1961). There are significant differences in the arboreal components of the pollen spectra above each of the unconformities in Flemish Bight. The mechanisms by which the pollen is translated to an offshore site are not fully understood, but in successive formations, the forests in the peripheral catchment areas appear to have been dominated alternately by coniferous species - *Alnus, Pinus, Picea* and *Betula* - or by mixed coniferous-deciduous species, including additional elements of *Quercus, Carpinus, Ulmus* and *Juniperus*.

The geometrical relationships and seismic and sedimentary facies of the sequence of formations in Flemish Bight reflect the combined influences of differential isostatic movements between the southern North Sea Basin and adjacent land areas, eustatic fluctuations in sea level, and changing rates of sediment supply during the Lower Pleistocene period. There is evidence of differential tectonic movements within Flemish Bight in the increase in thickness of Lower Pleistocene sediments to more than 350 m in the northeastern quadrant of the map.

The earliest isostatic movements are recorded in the slight angular unconformity between the Red Crag and Westkapelle Ground Formations in Flemish Bight, and there may also be a stratigraphic hiatus between the Red Crag and succeeding formations in Britain (VAN VOORTHUYSEN *et al.*, 1972). Pliocene deposits are widespread in the Netherlands sector and are conformably overlain by basal Lower Pleistocene sediments in the east, and the evidence suggests a slight relative uplift of the western margin of the Southern Bight during late Pliocene or early Lower Pleistocene times.

Widespread marine transgression at the base of the Westkapelle Ground Formation was followed by a phase of uniform subsidence throughout the Southern Bight area, and deposition of a sheet-like unit of open-shelf marine sediments in Flemish Bight. The formation describes a shallowing- and coarsening-upwards sequence, but this may simply reflect a general increase in the influx of clastic sediment into the southern North Sea during this period, rather than eustatic or isostatic influences. The western half of Flemish Bight continued to subside uniformly during deposition of the Smith's Knoll Formation, but tectonic subsidence and rates of sedimentation accelerated significantly in the east as the Northwest European delta complex expanded into the Southern Bight during Upper Tiglian times, allowing accumulation of up to 200 m of delta-related sediment offshore in the IJmuiden Ground Formation. Perhaps the massive influx of clastic sediment into the Netherlands and the southern North Sea may register a significant isostatic uplift of the principal sediment source areas during the Lower Pleistocene period.

During the deposition of the Winterton Shoal and Yarmouth Roads Formations, the delta complex continued to expand into and eventually spread across the entire Southern Bight area before "Cromerian Complex" times. Basinal subsidence was enhanced at first in the north-central area of Flemish Bight, but then more generally in the northeast as the deltafront moved northwards out of the Southern Bight, probably during Waalian times. The angular unconformity between the Yarmouth Roads Formation and overlying deposits suggests differential uplift and erosion in the southwest since the Lower Pleistocene period. The Southern Bight and its peripheral land areas appear to have been a relatively stable tectonic environment during Upper Pleistocene and Holocene times.

The four regional unconformities that have been identified in Flemish Bight indicate four episodes of widespread marine regression and transgression within the Southern Bight during the Upper Pliocene and Lower Pleistocene period. The lowest, Pliocene/Pleistocene unconformity, most likely represents a combination of isostatic and eustatic influences on regional sea level. Each of the Lower Pleistocene unconformities could have resulted from fluctuations in regional isostatic movements, but they display mainly non-angular relationships, and more probably record the effects of eustatic change in sea level on sedimentation on the southern periphery of the North Sea Basin. Periods of maximum climatic deterioration may be expected to correspond with regional lowering of sea level and marine regression from the Southern Bight area. Periods of climatic amelioration would then be represented by regional rise in sea level and marine transgression into the Southern Bight, possibly with some slight erosion of previous deposits. As the Northwest European delta complex expanded into the North Sea, the maximum transgression of each climatic amelioration was displaced progressively further northwards across the Southern Bight during the Lower Pleistocene period.

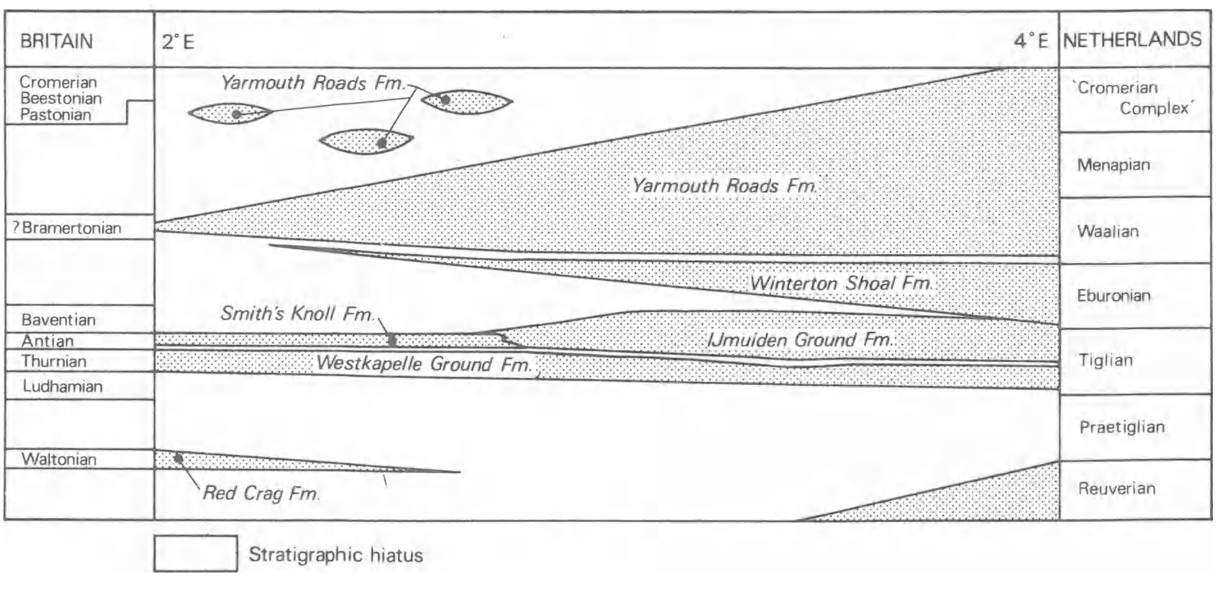


Figure 9 Suggested chronostratigraphic correlation of Flemish Bight formations with the Pliocene and Lower Pleistocene stages of Britain and the Netherlands.

The Upper Pliocene and Lower Pleistocene succession in Flemish Bight shows an upwards transition from shallow marine to open-shelf marine to delta-related marine and non-marine sedimentary facies. The history of sedimentation is punctuated by four regional unconformities which represent four significant intervals of stratigraphic hiatus offshore, and the marine succession offers at best a fragmentary record of climatic fluctuation in the Southern Bight during Lower Pleistocene times.

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11 Quaternary shelf deposits and drainage patterns off the French and Belgian coasts

F. MOSTAERT (1), J.P. AUFFRET (2), M. DE BATIST (1)(3), J.P. HENRIET (1), A. MOONS (1), E. SEVENS (1), I. VAN DEN BROEKE (1) and M. VERSCHUREN (1)(3)

1. Renard Centre of Marine Geology, Geological Institute, State University of Gent, Krijgslaan 281/S8, B-9000 Gent, Belgium.
2. Laboratoire de Géologie Marine, Université de Caen, Département de Géologie, Esplanade de la Paix 14, F-14032 Caen CEDEX, France.
3. Research Assistant, National Fund for Scientific Research, Belgium.

ABSTRACT

High-resolution reflection seismic data recorded in the Southern Bight of the North Sea have allowed to draw a first general reconstruction of the topography of the base of the Quaternary deposits. The morphological aspects of this erosion surface have been confronted with the knowledge of the Quaternary history of the Belgian coastal plain, yielding a hypothesis about the origin of the palaeolandforms and of the present-day offshore morphology. The evolution of this erosion surface is also related to the present-day knowledge of the river terraces in the coastal backland.

INTRODUCTION

The erosional surface at the base of the Neogene and Quaternary deposits has been analysed in a limited area of the Southern Bight of the North Sea, encompassing the Belgian coastal plain, the Belgian continental shelf, the French continental shelf in the Strait of Dover and the southernmost offshore area of the Netherlands. Most data have been derived from reflection seismic profiles, calibrated in terms of seismic velocities with a few borehole data. The results are presented as an isohypse map of the complex post-Palaeogene erosion surface (figure 1) and an isopach map of the post-Palaeogene cover sediments (figure 2).They are based on the seismic data-base, available in 1985. Meanwhile, the Renard Centre of Marine Geology (Gent University) has continued its exploration efforts in the Southern Bight, which will in the future lead to a more detailed model of the Quaternary evolution of the coastal plain and its offshore extension.

In the major part of this region, the surface which truncates the regular sequence of Palaeogene strata coincides with the base of the Quaternary deposits. In the Strait of Dover, this erosional surface also truncates Mesozoic deposits. In the extreme northeastern part of the considered area, the surface truncating the Palaeogene sequences however does not correspond with the base of the Quaternary. In this region, the Palaeogene strata have a larger dip of 2-3°NE and are covered by a thick sequence of Neogene and Quaternary deposits. As in this area the base of the Quaternary cannot be recognized from the seismic data alone, without borehole control, the northeastern part of the map (figure 1) displays the erosional surface at the top of the Palaeogene sequences.

The maps shown have also been implemented with morphological data of the base of the Quaternary deposits and of their thickness in the Belgian coastal plain, as compiled from boring data and literature (BAETEMAN, 1981 ; DE BREUCK and DE MOOR, 1974 ; DE PRET, 1983 ; DEVOS, 1984 ; HENRIET *et al.*, 1981 ; MOSTAERT, 1985 ; SOMME, 1979).

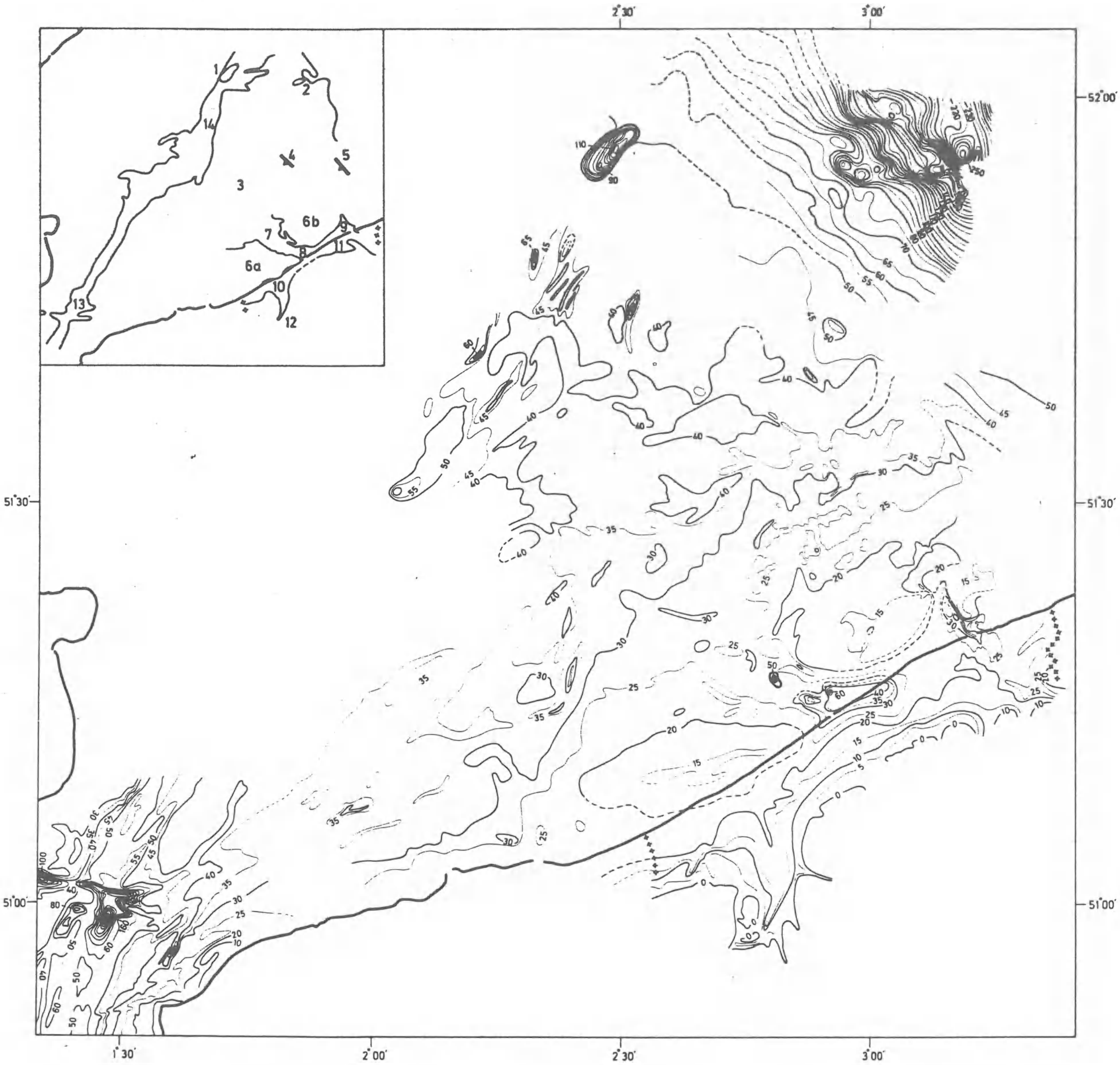

Figure 1 Topography of the erosional surface at the base of Neogene or Quaternary sediments in the Southern Bight of the North sea.
Morphological units : 1. Murray pit with Pliocene fill. 2. Buried valleys (Late Oligocene ?). 3. Abrasion slope. 4. Bartonian cuesta. 5. Rupelian cuesta. 6. Marginal platform. 7. Ostend valley. 8. Sepia pits in Ostend valley. 9. Zeebrugge pit and Bartonian cliff. 10. Western coastal palaeovalley. 11. Eastern coastal palaeovalley. 12. Yzer valley. 13. Dover Strait scour hollows. 14. Axial channel (see also BRIDGLAND and D'OLIER, this volume).

GENERAL FEATURES OF THE EROSION SURFACE

The description of the erosional surface truncating the Mesozoic and Palaeogene sediments will in this paper be limited to that area, where it effectively forms the base of the Quaternary and recentmost deposits. For a description of the morphology of this erosion surface below the Neogene cover on the Dutch continental shelf, we refer to VANNESTE' s thesis (1987).

The origin of the erosion surface at the base of the Quaternary deposits is polygenetic. The erosion took place under variable climatic conditions and in marine as well as fluvial circumstances.

In the present-day offshore area, the morphology of the erosion surface clearly shows a central, regular abrasion slope, bound to the northwest by a northeast-trending axial channel. This axial channel can be morphographically linked to the complex Strait of Dover channel system. The Belgian nearshore zone and coastal plain show a seaward dipping marginal slope, interrupted by the Yzer valley system and by a coastal valley floor, further referred to respectively as the eastern and western coastal palaeovalleys. A marginal platform limits the palaeovalleys to the north except for the central part of the Belgian coastal area, where it is transected by the so-called Ostend valley, with an east-west orientation.

THE CENTRAL ABRASION SURFACE

The slightly seaward dipping open marine abrasion surface is found at a depth of -15 m close to the present coastline and at depths of -35 to -40 m further seaward. It is characterized by a specific microrelief, sometimes reflecting structural aspects of the Tertiary substratum. The general trend of the surface seems to be independent of the strike and dip of the Tertiary layers and of the existing tectonic deformations.

The largest part of this surface, situated to the west of the Ostend valley, is modeled in Ieper Clay (London Clay). There the platform is quite regular, the main anomalies being rock-head depressions of low amplitude, some of which are filled up by Quaternary deposits (KIRBY and OELE, 1975). A detailed study of the seaward dipping slope allowed to detect continuous changes of dip probably indicating erosion in separate phases.

To the east of the Ostend valley the Tertiary substratum consists of alternating sand, clay and sandstone layers (HENRIET *et al.*, this volume). This lithological diversity is reflected in some morphological features on the erosion surface, such as remnants of cuestalike ridges. Scattered depressions occur as well. The asymmetric cuesta ridges have probably not been pre-modeled by subsequent rivers but should be regarded as the product of preferential scouring of the erodable sands in tidal channels, a phenomenon frequently accompanied by local overdeepening features. Locally the Asse Clay (Eocene - Bartonian) and the Boom Clay (Oligocene - Rupelian) are clearly exposed. In most cases, these ridges are covered by Quaternary deposits (sand waves and sand banks) and are consequently not reflected in the present-day sea floor morphology.

Systematic seismic profiling on the Zeeland Ridges (Thornton, Akkaert and Goote Bank) has revealed bedrock elevations generally located at their southwestern end. The Akkaert Bank and especially the Goote Bank are largely located on Tertiary bedrock swells and are characterized by relatively minor Quaternary accumulations.

These phenomena, together with the location of some sand banks along slope breaks on the abrasion surface suggests a wide-spread bedrock morphological control on sand bank setting and geometry in the Southern Bight of the North Sea. However, the active present-day erosion processes observed in channels lying between the sand ridges may also suggest that the bedrock swells are relict highs, partly modeled after the emplacement of the sand bank and protected from erosion by the thick sand cover. The overall orientation pattern of the sand ridges seems anyhow to be essentially controlled by tidal processes, and that since the opening of Dover Strait about 8,000 years BP.

Another characteristic feature of the erosional surface is the presence of remnants of palaeovalleys. The gully systems and their fills, which are of fluvial origin, existed before the final (Holocene) open marine scouring took place. Due to this erosion, only scarce remnants are left which makes it difficult to reconstruct the areal distribution pattern of these palaeoriver systems. The most typical palaeovalley is found under the Thornton Bank and has a northwestward trend. Based on the fact that fluvial incision in the Belgian coastal area did not reach relatable depths until the Saalian times, an older age of these palaeovalley remnants and their fills may be excluded. Except for the Ostend palaeovalley, correlation of these remnants with equivalents on land has not yet been possible.

THE NORTHEASTERN SLOPE

The regular norhteastward dip of the base of the Quaternary in the northeastern part of the study area (not represented on figure 1) corresponds with an Early Quaternary erosion surface. Marine Icenian deposits and Tegelen Clays covering this surface may be expected, as is the case in the laterally equivalent land position in Walcheren (VAN RUMMELEN, 1965). One of the internal Quaternary reflectors of the seismic profiles follows the dip of the central abrasion slope, which is of more recent age (Late Quaternary).

The present-day sea floor morphology does not reflect the northeastern slope at all. In this area the Quaternary sequences reach considerable thicknesses. It must be kept in mind that the erosion surface of the northeastern part of the study area, drawn on figure 1, corresponds with a post-Oligocene denudation related to an important sea level lowering (VANNESTE, 1987). It shows valleylike incisions which underline a continental fluvial impact.

THE AXIAL CHANNEL

The mid-Southern Bight drainage channel is characterized by a weakly pronounced southern margin. The channel has an average depth of 50 m. The transition to the abrasion platform is gradual. Several drift-filled hollows seem to be confined to this major gully system. The most spectacular rock-head depression is the Murray pit. This broad elliptical depression, scoured in Bartonian clays and Eocene sands, has been filled by eastward dipping drift deposits, of which some are of Pliocene age (BALSON, this volume). This implies that the first modeling of a central Southern Bight palaeovalley is at least of Pliocene age. The existence of a connection of the Southern North Sea with the ocean through the Channel in Late Tertiary time seems to be confirmed by marine faunas (FUNNEL, 1972). However it is generally assumed that the connection of the North Sea with the Channel through the Strait of Dover did not come into existence in recent geological times before Late Pleistocene (DESTOMBES *et al.*, 1975 ; KELLAWAY *et al.*, 1975 ; ZAGWIJN, 1979).

Very little information is available about the age and the origin of the thin and scarce infillings of the axial channel. The presence of overdeepened hollows indicates tidal scouring which does however not preclude a fluvial origin. Marine and fluvial erosion and infilling phases have alternated. The channel could have been an important southward orientated river system in glacial periods, when northward drainage was blocked by the glaciers.

THE STRAIT OF DOVER

The Strait of Dover, connecting the Channel with the North Sea, is situated on a structural westnorthwest-eastsoutheast-trending high, the Weald-Artois Horst, which has formed a barrier between the northern Anglo-Belgian basin and the southern Anglo-Paris basin from Middle Eocene up to recent geological times. Deposits which could be correlated with the erosion phases which formed the Strait of Dover are scarce :

some channel fills or remnants of terraces have been described (AUFFRET and ALDUC, 1977). Three drowned terrace levels, which could be correlated with secondary alluvial infillings, have been identified.

The pre-existence of a structural inversion landscape (Boulonnais) and the transversal fracturation of the Boulonnais are considered as important factors in the development of an initial drainage pattern.

The presence of so-called "fosses" (Fosse Dangeard and Fosse de la Bassurelle) indicates that tidal action was a major factor of remodeling of the initial drainage pattern and of the creation of Dover Strait itself. The Fosse Dangeard is located on erodable sands and clays of Aptian-Albian and Wealdian age. A combination of lithological and structural factors has probably controlled the overdeepening processes. According to HAMILTON and SMITH (1972), the "fosses" must be related with tidal scouring during periods of withdrawal or penetration of marine action in pre-existing channels, due to sea level changes. Moreover, it may be expected that tidal currents are amplified in the narrow strait when communication is achieved between the Channel and the Southern Bight, because of the decalation of the tidal waves in the separate basins.

THE COASTAL PALAEOVALLEYS AND THE OSTEND VALLEY

The seaward dipping topography of the Tertiary substratum on the southeastern margin of the coastal plain is steeper than the general slope of the offshore abrasion platform. This marginal slope shows little structural influences (MOSTAERT and DE MOOR, this volume) and runs 10 km landward of the present-day coastline except along the Yzer valley, where this slope retreats further landward. The Yzer valley joins the palaeovalley which runs parallel with and underneath the present-day coastline. Smaller tributaries as the Waardamme river join the eastern branch of the coastal valley.

These coastal palaeovalleys show a striking asymmetric cross profile. The northern flank is little exposed and not higher than -10 m. The southern flank corresponds with the steeper and higher coastal margin reaching at some places levels of +20 m. The asymmetry is a result of marine erosion, as will be developed further. The average depth of the valley floor is about 25 to 30 m.

The northwest trending Ostend valley was probably related to the so-called Flemish valley via the eastern branch of the coastal palaeovalley. In offshore direction the valley morphology is obliterated by the general northward dipping abrasion surface. The longitudinal profile of the coastal valleys and of the Ostend valley is fairly irregular. The Sepia pits off Ostend, scoured to -62 m in the Ieper Clay, and the depression under the outer harbour of Zeebrugge, flanked by a steep Bartonian cliff, probably are tidal scouring structures created in ancient estuarine environments.

The Quaternary history of the eastern part of the Belgian coastal plain is quite different from that of the western part. In the western coastal plain, marine deposits of Middle Pleistocene age have been found (the Herzeele Formation). These deposits have been laid down in a sea which may have penetrated from the southwest through an open Strait of Dover (SOMME, 1979 ; PAEPE and BAETEMAN, 1979) into a precursor of the Yzer valley. Due to later sea level changes, renewed fluvial erosion took place in the Yzer valley. These erosive processes have apparently not been able to remove all older marine deposits, hence leaving a fringe of tidal flat sediments.

Such Middle Pleistocene marine deposits are totally lacking in the eastern coastal plain. The eastern coastal valley and the Ostend branch got not shaped before the Saale glacial period (MOSTAERT and DE MOOR, this volume).The maximal deepening took place during the Eemian period, when the sea breached into the valley. Tidal scour hollows have been formed, comparable with those found in the present-day Western Scheldt estuary. Both in the eastern and in the western coastal plain, the Eemian open marine environment penetrated further landward than during the Holocene. As a consequence the northern flank of the coastal valleys got fairly but not entirely flattened by open marine erosion.

It is not fully clear yet whether the Yzer valley ever joined the Ostend valley along the western coastal valley, or kept a totally independent evolution with drainage to the northwest via the western outlet of the valley, as was the case during earlier marine inundations.

On base of these palaeogeographic considerations, the age of the modeling of the northward dipping abrasion surface can be brought back to Eemian times. This may as well be the age of the deepest scouring of the Ostend valley and of the Sepia pits. The study of the Quaternary infillings of the Sepia pits, which is now in progress, may shed some new light on this problem. Anyhow, thick Eemian deposits are preserved in the eastern coastal plain. The present-day nearshore ridges also contain Eemian sediments.

Figure 2 Thickness of Quaternary deposits or Quaternary plus Neogene deposits (NE edge of the study area)

The Weichselian drainage pattern, which seems to be directed to the central Southern Bight deeps, has locally eroded the Eemian abrasion platform. Also Holocene erosion influenced by sea level rise partly uncovered the Eemian abrasion platform. The re-opening of the Strait of Dover caused a spectacular and complex redistribution of the sediments available in the basin and created new sand supplies. These changes in the sedimentary environment, accompanied by differential erosion and deposition form the context which lead to the development of a rather flat basal plain covered with sand banks and sand wave fields. The modeling of the Tertiary substratum is still taking place for instance in the erosive gullies between the sand banks. There the Tertiary strata are exposed or lay beneath a thin gravel veneer. The map showing the thickness of the Quaternary deposits (figure 2) clearly shows that a substantial part of the offshore area is characterized by little or no Quaternary cover. This map also clearly reflects the distribution pattern of the Holocene sand banks. It is only in the northeastern part of the study area, in the present-day coastal plain and in the nearshore zone that the pre-Holocene Quaternary deposits get a sizable importance.

CONCLUSION

Seismic high-resolution surveys in the Southern Bight of the North Sea have allowed to study the present-day morphology of specific stratigraphical boundaries and may lead to palaeoenvironmental and palaeomorphological reconstructions. The palaeomorphological surface which has been investigated here is a result of fluvial and dominantly open marine scouring under varying sea level conditions since the end of the Tertiary, a process which has been achieved in a rather short period of time. The morphological features which have been described and if possible explained may contribute to the interpretation of older, more important erosion plains, reflecting major long-term sea level changes.

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12 Middle and Upper Pleistocene and Holocene stratigraphy in the southern North Sea between 52° and 54°N, 2° to 4°E

T.D.J. CAMERON (1), R.T.E. SCHÜTTEN-HELM (2) and C. LABAN (2)

1. British Geological Survey, 19 Grange Terrace, Edinburgh EH9 2LF, United Kingdom.
2. Rijks Geologische Dienst, Spaarne 17, Haarlem, The Netherlands.

ABSTRACT

A late Quaternary stratigraphy has been established for the southern North Sea by the British Geological Survey and Rijks Geologische Dienst during joint compilation of 1:250,000 maps of the Quaternary Geology and Sea Bed Sediments of the area between 52° and 54°N, and 2° and 4° E. The deposits of the three regional glaciations of Northwest Europe are represented offshore. The Elsterian subglacial valley system of northern Germany continues across the southern North Sea towards eastern England, and the valleys, which are commonly more than 100 m deep offshore, have been filled mainly by Elsterian glaciolacustrine and fluvioglacial deposits. There was no connection across the southern North Sea between the Saalian and Weichselian ice sheets of northern Europe and eastern England. Periglacial conditions prevailed offshore during both of these glacial periods, except in the northwest, where up to 15 m of till was deposited during late Weichselian times. Marine sands and clays were deposited in the southern North Sea during the Holsteinian and Eemian interglacial periods. The sediments of the Holocene transgression have been subdivided into six formations on the BGS/RGD maps, based on their lithology and shell content.

INTRODUCTION

Fully glacial conditions extended to Northwest Europe for the first time during the Middle Pleistocene and initiated a major change in the style of sedimentation in the southern North Sea. The marine and deltaic sedimentary facies which characterise the pre-glacial Lower and early Middle Pleistocene succession of the Southern Bight of the North Sea are overlain by a relatively complex association of glacial, proglacial, periglacial and interglacial marine and intertidal deposits. The stratigraphic relationships of these late Quaternary deposits have been established during joint compilation by the British Geological Survey and Rijks Geologische Dienst of 1:250,000 maps of the Quaternary Geology and Sea Bed Sediments (Holocene Geology) of the UK and Netherlands sectors of the North Sea. As the deposits of all three of the regional glaciations are poorly represented within the Flemish Bight sheet (52°-53°N, 2°-4°E) the area described in this paper has been extended northwards to include data from the Indefatigable sheet (53°-54°N, 2°-4°E, figure 1).

The methods of geological survey in the UK sector of the southern North Sea have been summarised elsewhere (CAMERON *et al.*, this volume). A range of low-frequency sparker or air gun sources was operated along each line in a seismic grid of mainly north-south and east-west lines in Flemish Bight and Indefatigable (figure 2), and has provided geological information from the sea bed to below the base of the late Quaternary succession, here mostly less than 50 m below sea bed. Higher frequency pinger or boomer sources were operated simultaneously along all lines to provide better definition of near-surface seismic reflectors and have proved particularly effective in resolving bedding surfaces within the relatively argillaceous sedimentary units.

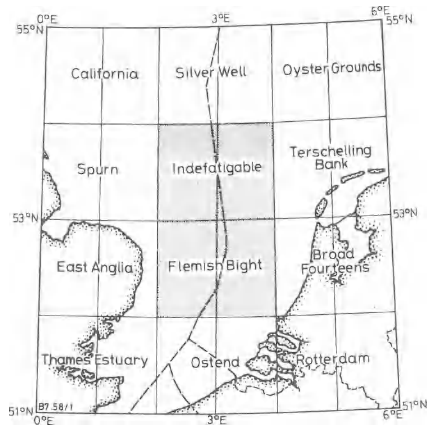

Figure 1 Location of Flemish Bight and Indefatigable sheets described in the text, and of their adjacent sheets in the southern North Sea.

In the Netherlands sector, a high-resolution seismic profiler, the Sonia, was operated along a network of northwest-southeast and northeast-southwest traverses mostly about 14 km apart (figure 2), and has provided geological information to 40 m below the sea bed. Additional grids of relatively closely spaced traverses have been surveyed across some of the principal areas of geological interest. A 4 kJ sparker survey of the Indefatigable sheet was completed during 1983, and it is planned to extend this coverage on the Flemish Bight sheet during 1986.

Interpretation of the seismic profiles has provided the framework for establishing a late Quaternary succession offshore. Extensive sampling programmes have been carried out in both the UK and Netherlands sectors of the area to collect samples and cores of all of the seismic units identified in the profiles. The uppermost, Holocene sediment layers have been sampled at 273 stations in the UK sector using the BGS vibrocorer, which is capable of obtaining up to 6 m of relatively undisturbed core in argillaceous lithologies; over 500 grab samples have also been taken in the UK sector. Cores and samples have been collected at about 660 stations in the Netherlands sector using a vibrocorer or the Geodoff air-lift/counterflush system capable of obtaining disturbed samples from up to 10 m below the sea bed. Middle or Upper Pleistocene sediments have been sampled from beneath the Holocene cover in 57 % and about 50 % of the vibrocores and air-lift samples in the UK and Netherlands sectors respectively. About 520 grab samples have been used to study the nature of the surface sediments in the Netherlands sector. Some of the cores and samples have been obtained - especially in the Netherlands sector - from site investigation boreholes and these have yielded important evidence on the total number of glacial and interglacial stages represented in the offshore deposits. The returns from these boreholes and from many of the vibrocores have been analysed by RGD for their pollen content to enable correlation with the onshore succession established in the Netherlands. Analyses of the molluscan faunas have assisted in correlation and palaeo-environmental interpretation of the marine beds.

The upper Middle and the Upper Pleistocene deposits of the southern North Sea have been subdivided into ten sedimentary units in Flemish Bight and Indefatigable, and each of these has been allocated formation status on the Quaternary maps. Their correlation with the climatic stages of Britain and the Netherlands is summarised in figure 3. Six of these formations comprise the glacial, periglacial and proglacial deposits associated with the Elsterian (Anglian), Saalian (Wolstonian) and Weichselian (Devensian) glacial stages. Two of the formations represent the marine deposits of the Holsteinian (Hoxnian) and Eemian (Ipswichian) interglacial stages. The remaining formations comprise an extensive unit of late Eemian to early Weichselian brackish-marine and lacustrine clay, and a unit of fluviatile, gravelly sand deposited in southeast Flemish Bight during low sea-level stands. Some of the formations are separated by erosion surfaces indicating intervals of stratigraphic hiatus within the late Quaternary succession offshore. All of the formations have a characteristic lithology, easily distinguished from the deposits of the preceding and successive units in boreholes and cores.

The Holocene sediments of the area have been subdivided into six sedimentary units in the joint BGS/RGD 1:250,000 series of Sea Bed Sediment maps. These have been allocated formation status on the maps and comprise six distinctive sedimentary facies relating to the early Holocene transgression of the North Sea across the Southern Bight. Visual inspection of the vibrocores has enabled one of the formations on the

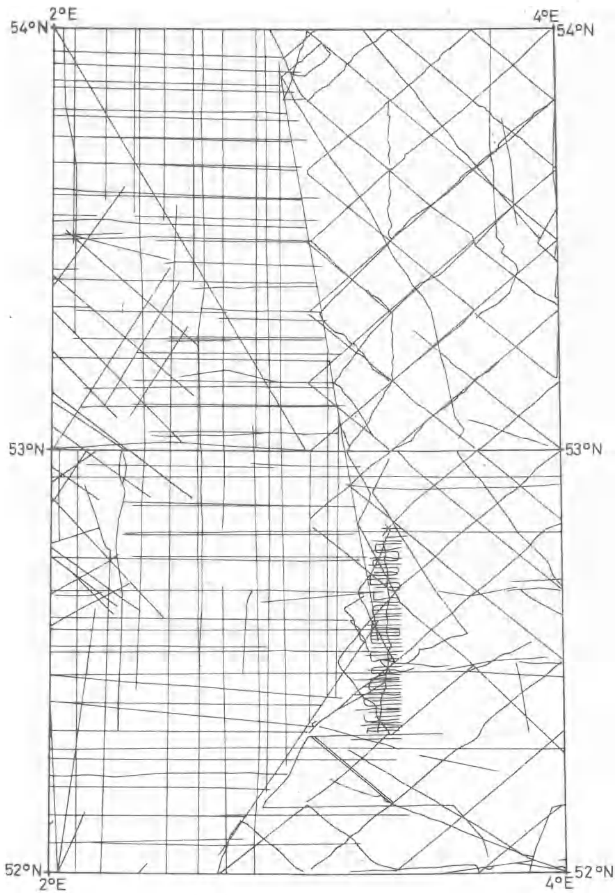

Figure 2 BGS and RGD seismic survey traverses in the Flemish Bight and Indefatigable sheets.

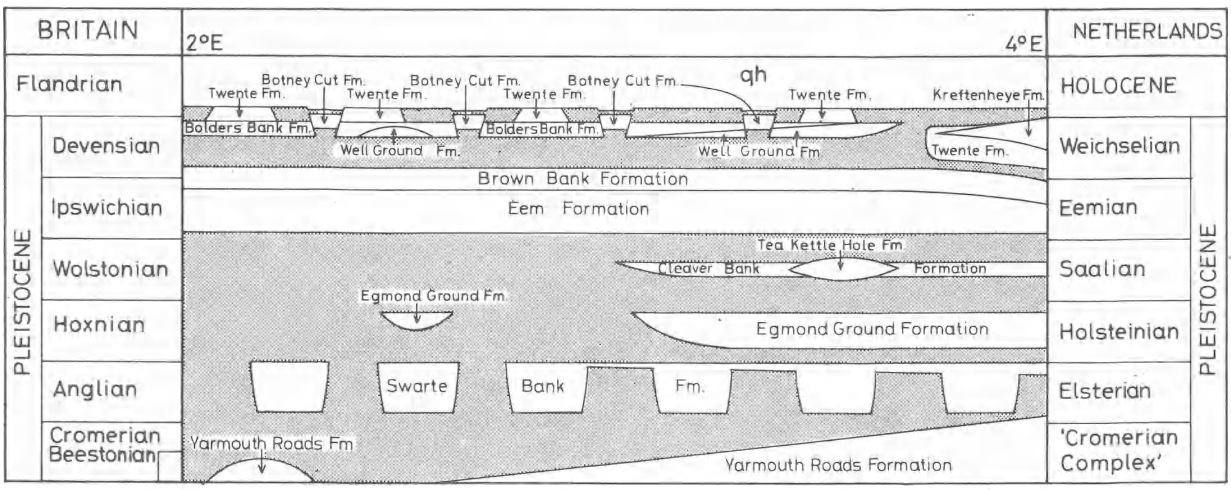

Figure 3 Chronostratigraphic diagram showing the correlation between the late Middle and Upper Pleistocene formations of the southern North Sea with the late Quaternary climatic stages of Britain and the Netherlands.

Indefatigable sheet, comprising marine, reworked glacial sands, to be further subdivided into two members, based on their grain size and clay content.

MIDDLE AND UPPER PLEISTOCENE

a. Swarte Bank Formation (Elsterian)

The base of the late Quaternary succession in the southern North Sea is a gently undulating erosional unconformity mostly less than 40 m below the sea bed (less than 80 m below sea level). On the Indefatigable sheet, this surface is incised by a complex system of anastomosing valleys, many of which have been eroded to depths of greater than 100 m into Lower Pleistocene and Tertiary sediments. The valleys (figures 4 and 5) trend mainly NNW-SSE, are up to 23 km wide and have a maximum depth of 450 m (510 m below sea level) at 53°16'N, 3°E, and all have a conspicuously uneven longitudinal profile. The Swarte Bank Formation comprises the non-marine sediments which completely filled most of these valleys in the UK sector, which partially or completely filled all of the valleys in the Netherlands sector, and which were deposited between some of the valleys in eastern Indefatigable. The formation is principally of Elsterian age, is overlain by Holsteinian marine sediments in and around many of the valleys in the Netherlands sector, but is mainly overlain by younger Pleistocene sediments in the UK sector.

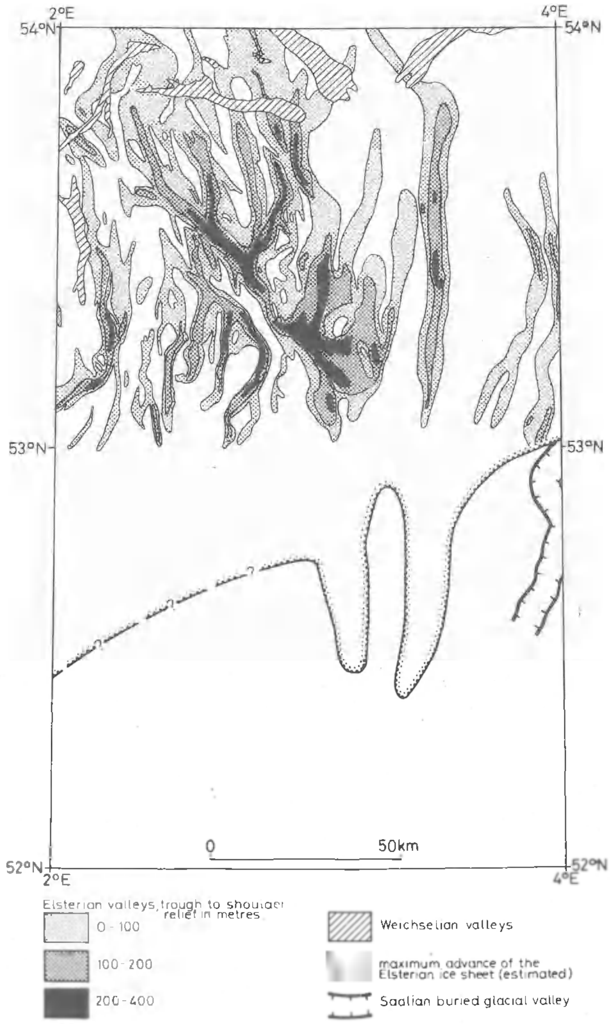

Figure 4 Map of the Elsterian, Saalian and Weichselian glacial valleys in the Flemish Bight and Indefatigable sheets.

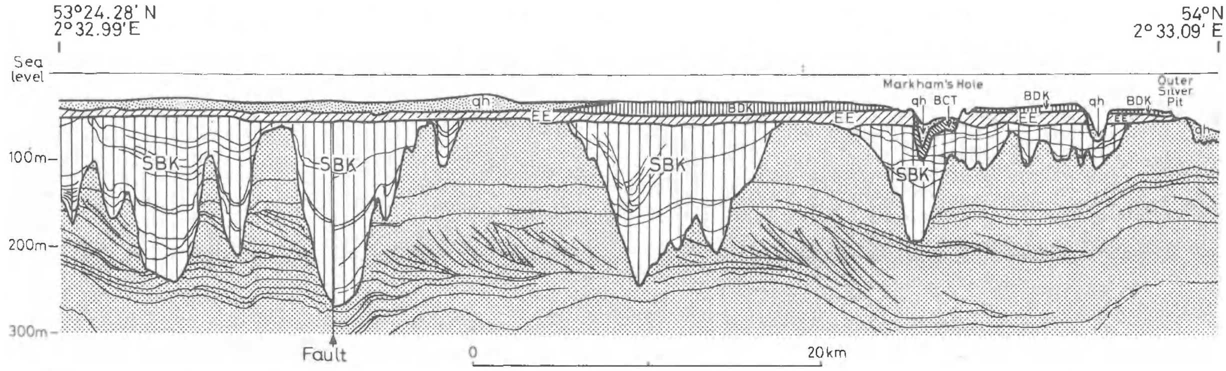

Figure 5 Geological cross-section along the line indicated on figure 8, based on interpretation of an airgun profile. The formations are : SBK = Swarte Bank Formation. EE = Eem Formation. BDK = Bolders Bank Formation. BCT = Botney Cut Formation. qh = Holocene sediments (undifferentiated).
Along this line of section, the Elsterian valleys have been cut into Lower Pleistocene sediments.

There is a broad similarity in the style of infill of the valleys on the seismic profiles. All of the valleys contain a basal unit which is structureless or has a chaotic reflector configuration, has zones of reduced acoustic penetration, and may comprise gravelly coarse sand, slump beds or, locally, till deposited penecontemporaneously with valley incision. This facies has completely filled many of the shallower valleys in the south, but represents less than 10 % of the fill of some of the deeper valleys. The upper surface of this unit is an uneven, moderate or high amplitude seismic reflector.

The second phase of infill is characterised by sub-parallel layering on the seismic profiles, draped over topographic irregularities on the surface of the underlying unit. The layering is defined by high amplitude reflectors which are up to 60 ms apart (two-way time), but which occur at similar intervals and depths in adjacent valleys. Many of the valleys evidently comprised an interconnecting network of open depressions during this phase of infill. Between these reflectors, the sediments are either structureless or have discontinuous, low amplitude reflectors on the profiles. In site investigation boreholes in the Netherlands sector, these sediments have proved to be glacio-lacustrine clays with laminae of silty clay and beds of silt and very fine- or fine-grained sand.

A third component of infill occurs above an erosion surface cut into the parallel-bedded sediments in some of the valleys in the east. This unit is mostly less than 30 m thick and is characterised by gently-inclined, low or moderate amplitude reflectors on the seismic profiles. The reflectors are inclined mainly towards the west and the sediments may comprise a late Elsterian delta-related influx of fine-grained sands and clays. The intervalley deposits in the Netherlands sector are fluvioglacial sands, silts and clays.

The Indefatigable palaeovalleys have similar dimensions, morphology and infill to, and are the direct offshore continuation of the Elsterian subglacial valley system of Northwest Europe, described by EHLERS *et al.* (1984). They are considerably larger than the valleys of equivalent age cut into Upper Cretaceous rocks in East Anglia (cf. WOODLAND, 1970). EHLERS *et al. (op. cit.)* observed that their base levels are much too deep for fluviatile erosion alone to have been effective, and deduced that they are subglacial features, eroded under very high hydrostatic pressure by sudden outbursts of meltwater through the base of an ice cap. As such, they record the earliest extension of land ice from Northwest Europe into the southern North Sea.

The southern limit of the Elsterian ice sheet (figure 4) is not easily defined from the offshore evidence. No valleys of comparable dimensions to the Indefatigable system have been observed south of 53°N, though shallow valleys in the Middle Pleistocene regional unconformity may locally contain sediments of Elsterian age. OELE (1971) recorded the local occurrence of widely-disturbed Lower Pleistocene sediments in the central part of the Netherlands sector of Flemish Bight and attributed these to ice-pushing during the Saalian glaciation. Recent mapping has indicated that the Saalian ice sheet had a relatively limited extent offshore, and the disturbed sediments more likely record ice-push deformation during Elsterian times. Ice-push ridges have not been observed in the UK sector of Flemish Bight, and the Anglian (Elsterian) tills of East Anglia do not extend eastwards of the present British coastline.

b. Egmond Ground Formation (Holsteinian)

By the onset of the Holsteinian interglacial stage, most of the Elsterian valleys in the UK sector, and almost all of the valleys in the Netherlands sector of the southern North Sea had been totally filled by the non-marine sediments of the Swarte Bank Formation. The Egmond Ground Formation comprises the marine sediments, locally several tens of metres thick, which completed the infill of the remaining valleys during the Holsteinian transgression, and which were deposited more widely between the valleys in the Netherlands sector, particularly on the Indefatigable sheet. Outside the valleys, the formation is less than 25 m thick. The Egmond Ground Formation subcrops mainly beneath Eemian marine sediments, but has been sampled beneath Saalian proglacial sediments in northeastern Indefatigable, and comprises marine, sparsely shelly, very fine-, fine- and medium-grained sands with thin beds of silt and clay.

Holsteinian sediments have not been sampled in any of the vibrocores or boreholes in the UK sector of the area. Seismic evidence indicates that if present, they occur as the uppermost fill of some of the Elsterian valleys adjacent to the median line, but that even here they may be only a few metres thick. Outside the area, marine clays dated as Hoxnian (Holsteinian) in age by FISHER *et al.* (1969) have filled an Elsterian valley 50 km north of East Anglia (BALSON and CAMERON, 1985), suggesting that the Holsteinian transgression may have extended more widely across the UK sector, but that the sediments have been largely removed by later Pleistocene erosion.

c. Tea Kettle Hole Formation (Saalian)

During the Saalian glacial stage, an extensive ice sheet covered most of Northwest Europe (EHLERS *et al.*, 1984). Opposing views have been expressed by STRAW (1983) and SUMBLER (1983) as to whether a contemporary ice sheet covered eastern England, but the offshore evidence indicates that periglacial conditions prevailed over most of the southern North Sea during this period. Sediments of Saalian age have not been sampled in the UK sector of the area. The Tea Kettle Hole Formation comprises the well-sorted, very fine- or fine-grained, and only sparsely micaceous, wind-blown sands which have been sampled beneath Eemian marine beds in vibrocores and site investigation boreholes in southeastern Indefatigable. The formation has a maximum thickness of 6 m and may have formed a more extensive deposit prior to the Eemian transgression.

d. Cleaver Bank Formation (Saalian)

At its maximum offshore extent, the Saalian ice sheet encroached from the Netherlands to deposit till to within 14 km of the eastern margin of the Indefatigable sheet and formed ice-pushed ridges of Middle Pleistocene sediments on the eastern margin of Flemish Bight. The Cleaver Bank Formation comprises proglacial silty clays or fluvioglacial very fine- to fine-grained, micaceous outwash sands with beds of silt and clay which were deposited beyond the ice front in northeastern Indefatigable. The formation subcrops beneath Eemian marine sediments and is locally up to 20 m thick.

e. Eem Formation (Eemian)

Marine sands and clays of the Eem Formation are widespread in the Indefatigable sheet and in the eastern half of Flemish Bight and are mostly between 5 and 20 m thick (figure 6). These deposits are structureless or have low amplitude, sub-horizontal reflectors on the seismic profiles, and rest above a gently-undulating erosion surface on formations ranging between Lower Pleistocene and Saalian in age. The Eem Formation passes upwards into late Eemian to early Weichselian brackish-marine deposits or is overlain by Holocene marine deposits in most of Flemish Bight and in southern and eastern Indefatigable, but is overlain by late Weichselian till in northwestern Indefatigable.

No marine beds of the E1 and E2 pollen zones as defined by ZAGWIJN (1961) have so far been recorded from the southern North Sea (JELGERSMA, 1979). The Eem Formation comprises fine- or medium-grained, locally gravelly sands of the E3 to E6 pollen zones in the Netherlands sector (OELE and SCHÜTTENHELM, 1979) and these sediments have a markedly higher shell content than the deposits of the Holsteinian and

Holocene transgressions. Though relatively few samples of the formation have been obtained in the UK sector, in northwestern Flemish Bight 8 vibrocores have proved that its uppermost beds are very fine- and fine-grained muddy sands and clays of intertidal aspect, sparsely shelly, and contain a late Eemian, E6 pollen assemblage (Personal communication, J. DE JONG).

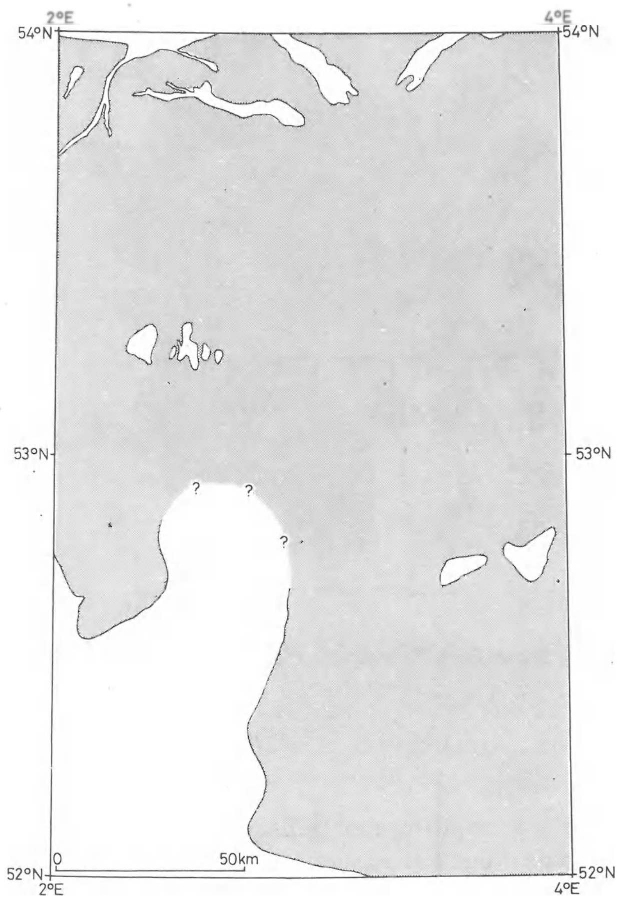

Figure 6 Mapped extent of the Eem Formation in the Flemish Bight and Indefatigable sheets.

f. Brown Bank Formation (late Eemian to early Weichselian)

Towards the end of the Eemian interglacial stage, regional sea level in the southern North Sea fell by 30 m from its Eemian maximum to between 20 and 30 m below its present level (JELGERSMA, 1979) during the initial deterioration in the climate of Northwest Europe towards the fully glacial conditions of the Weichselian Stage. The Brown Bank Formation records a further fall in sea level to about 40 m below present during early Weichselian times.

The Brown Bank Formation subcrops extensively beneath Holocene marine sediments in the central area of Flemish Bight and in southern-central Indefatigable (figure 7), is generally between 5 and 15 m thick, and has a maximum thickness of 20 m. The base of the formation is poorly defined acoustically or is represented by a low amplitude, sub-horizontal reflector on the seismic profiles where it overlies Eemian sediments. Where the formation overlies Lower and Middle Pleistocene sediments in western Flemish Bight, its base is defined by a gently undulating, high amplitude seismic reflector, describing an erosion surface between 40 and 60 m below sea level. The boundary between the dominantly sandy sediments of the Eem Formation and the clays of the Brown Bank Formation lies within the E6 pollen zone in northwestern Flemish Bight (Personal communication, J. DE JONG), but may be slightly diachronous on a regional scale.

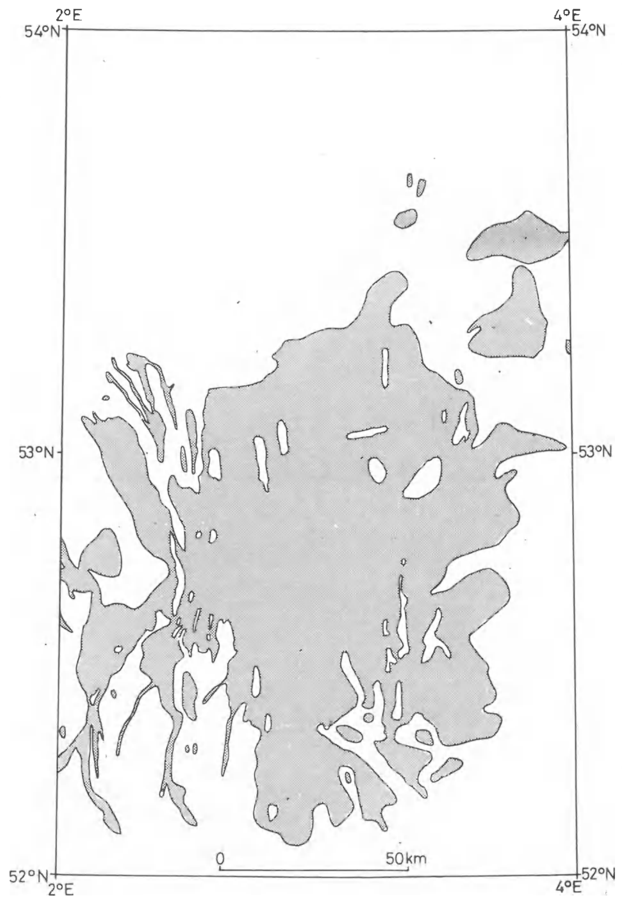

Figure 7 Mapped extent of the Brown Bank Formation in the Flemish Bight and Indefatigable sheets.

The Brown Bank Formation has a strongly-layered appearance on high resolution acoustic profiles. Its layering is draped over topographic irregularities suggesting low-energy deposition. A unit comprising up to 3 m of relatively structureless sediments is present at the base of the formation in some of the depressions in the west. On many profiles across the centre of the Flemish Bight sheet, the formation's internal structure is completely masked within discrete zones by acoustic blanking, and there is a reduction in the amplitude of deeper reflectors beneath these zones on some of the sparker records. Similar acoustic blanking is common within soft, argillaceous sediments in the central and northern North Sea and was attributed by FANNIN (1980) to scattering of the acoustic energy from the seismic source by interstitial gas. It seems likely that the sediments of the Brown Bank Formation are similarly gas-charged within these zones, which have a total areal extent of approximately 500 km^2 on the Flemish Bight sheet (BGS/RGD 1:250,000 Quaternary map).

The Brown Bank Formation comprises a distinctive grey-brown silty clay containing lenses and pods of very fine-grained muddy sand and variable amounts of finely-disseminated iron sulphide. Specimens of the gastropod *Hydrobia ulvae* have been recovered from only a few out of more than 100 vibrocores taken, and most of the cores contain no macrofauna. Angular pebbles up to 0.5 cm in diameter have been observed in some of the cores and are interpreted as dropstones. Two of the vibrocores penetrated beds of gravelly, shelly, coarse-grained sand close to the base of the formation. The layering observed in the seismic profiles does not appear to correspond to any significant lithological variation in the cores.

Primary lamination is present in about 10 % of the cores recovered and is defined by slight variations in colour, grain size, carbonate and sulphide content. Although mainly sub-horizontal, there are zones of intense deformation in some of the cores in which the lamination describes upright or recumbent folds and occasionally has a crenulated appearance. The fold hinges are commonly truncated by the core margins, indicating that the deformation is not an artefact of the coring process. Microfaults with displacements of a few millimetres have been observed in some cores. The zones of deformation are underlain and overlain by undisturbed bedding, suggesting that they represent expulsion of pore water or gas rather than cryoturbation.

There is abundant evidence of bioturbation in the cores. Burrows are particularly apparent within the more sulphide-rich layers, are emphasised by their sulphide rims giving the sediment a mottled appearance, and have diameters of up to 3 mm. Approximately 50 % of the sand in the cores occurs as discrete pods of similar dimensions in bands up to 2 cm thick within the clay, and these are interpreted as representing bioturbation of continuous sand layers within a primary laminated sequence.

A marginal facies of finely-laminated greyish-brown clay with beds of silt and very fine-grained sand has completely filled a system of north-south trending channels between 2 and 15 km wide (figure 7) and up to 20 m deep in western Flemish Bight. These sediments are relatively water-saturated and were extensively deformed by the coring process ; they are locally current-laminated, and contain no evidence of bioturbation.

The sedimentary facies represented within the main outcrop of the Brown Bank Formation indicate that most of the area was covered by a shallow lagoon with only limited access to the open sea during late Eemian and early Weichselian times. The laminated sediments in the west were deposited in estuarine or fluviatile channels which flowed into the lagoon from the southwest. The lagoonal sediments pass upwards into finely-laminated lacustrine sediments in the Netherlands sector indicating that the lagoon was cut off completely from the open sea when regional sea level had fallen to about 40 m below present.

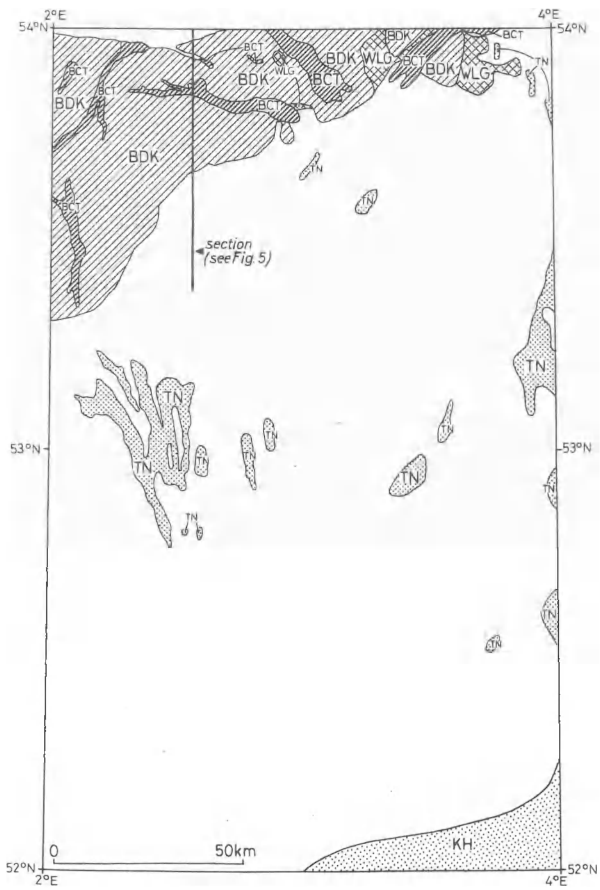

Figure 8 Map of Weichselian deposits in the Flemish Bight and Indefatigable sheets.
Weichselian formations are : BCT = Botney Cut Formation. BDK = Bolders Bank Formation. WLG = Well Ground Formation. KH = Kreftenheye Formation. TN = Twente Formation.

g. Kreftenheye Formation (late Weichselian)

During the low sea-level stand of the Weichselian glacial stage, fluviatile sedimentation associated with the Rhine and Meuse rivers extended offshore into parts of the Southern Bight (figure 8). The Kreftenheye Formation comprises the fluviatile fine- and medium-grained, locally gravelly sands which were deposited in southeastern Flemish Bight during late Weichselian times. This dating is based on layers of pumice related to well-dated volcanic eruptions in the Eifel area, West Germany. The Kreftenheye Formation sands are up to 8 m thick and commonly contain an Eemian molluscan fauna reworked from the underlying marine beds of the Eem Formation.

h. Bolders Bank Formation (late Weichselian)

At its maximum extent, the Weichselian ice sheet covered most of eastern England north of 53°N, and late Weichselian till is at or close to the sea bed over extensive areas to the north of East Anglia (BALSON and CAMERON, 1985). This till, the Bolders Bank Formation, extends into northwestern Indefatigable (figure 8) where it overlies the marine interglacial sediments of the Eem Formation, has a maximum thickness of 15 m, but is less than 5 m thick to the east of 3°E. The till is mainly structureless on the seismic profiles and has a flat or gently undulating base between 39 and 45 m below mean sea level. In shallow cores, the Bolders Bank Formation is a uniform, stiff, greyish-brown sandy clay with pebbles derived mainly from the Upper Palaeozoic and Mesozoic rocks of eastern England, but there are minor igneous and metamorphic components of Scottish or Scandinavian origin.

i. Well Ground Formation (late Weichselian)

The late Weichselian fluvioglacial deposits of the Well Ground Formation (figure 8) are laterally equivalent to, but locally underlie the till of the Bolders Bank Formation in northeastern Indefatigable, and they may also be present locally beneath the till in the west. The sediments are predominantly micaceous, very fine- or fine-grained sands with intercalations of silt and clay. The formation has a maximum thickness of 5 m.

j. Botney Cut Formation (late Weichselian to early Holocene)

A system of late Weichselian valleys (figures 4 and 8) occurs mostly within the outcrop limits of the Bolders Bank Formation. The valleys are up to 80 m deep, have a maximum width of 8 km, and have been eroded through the Weichselian till into older Pleistocene sediments. Although their dimensions are much smaller than the Elsterian system, these valleys are similarly thought to have formed by the outbursts of large volumes of meltwater through the base of the decaying Weichselian ice sheet. Many of the valleys have

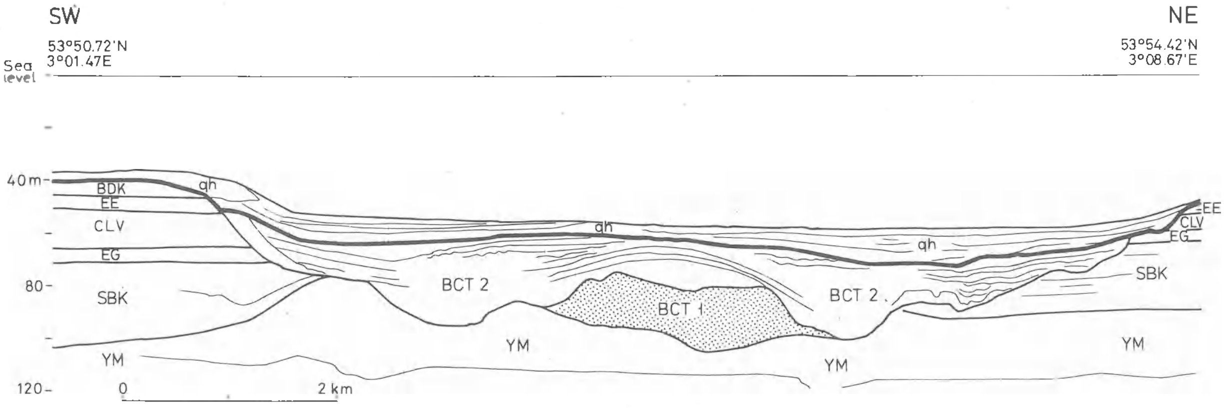

Figure 9 Geological cross-section across the southern part of Botney Cut. The cross-section is based on interpretation of the seismic profiles in the area.
The formations are : YM = Yarmouth Roads Formation (Lower and early Middle Pleistocene). SBK = Swarte Bank Formation. EG = Egmond Ground Formation. CLV = Cleaver Bank Formation. EE = Eem Formation. BDK = Bolders Bank Formation. BCT1 = lower, seismically structureless unit of Botney Cut Formation. BCT2 = upper, parallel-bedded unit of Botney Cut Formation. qh = Holocene deposits (undifferentiated).

been completely filled by the sediments of the Botney Cut Formation and overlying Holocene deposits, but some, such as the Botney Cut and Markham's Hole, have been only partially filled and remain open as linear depressions up to 40 m deeper than the surrounding sea floor.

The Botney Cut Formation can be separated into two discrete seismic and sedimentary facies in all of the valleys. The lower unit (figure 9) is structureless on the seismic profiles, is up to 15 m thick, and may comprise poorly-sorted, gravelly coarse sands. The upper parallel-bedded unit is up to 35 m thick, consists of very soft, slightly sandy mud with partings of fine sand, and was probably deposited in a glacio-lacustrine environment. There are zones of acoustic blanking on seismic profiles across Markham's Hole and Botney Cut suggesting that the sediments are locally gas-charged.

k. Twente Formation (Weichselian to early Holocene)

South of the Weichselian glacial maximum, periglacial conditions prevailed from Weichselian into early Holocene times. The periglacial sediments of the Twente Formation may have been deposited throughout most of the Southern Bight during this period, but they were largely reworked during the Holocene transgression, and are now mainly preserved as scattered outliers (figure 8), for instance beneath Holocene topographic ridges. In northwestern Flemish Bight, the formation was penetrated beneath early Holocene peat in three vibrocores and comprises moderately-sorted, fine-grained sand with minor intercalations of clay. Close to the north, in southwestern Indefatigable, two vibrocores have proved up to 30 cm of fine-grained, wind-blown sand overlying the Bolders Bank Formation, confirming that the sediments here are of late Weichselian or early Holocene age. The Twente Formation is up to 5 m thick in the Netherlands sector, comprises well-sorted wind-blown sand, and includes in southeastern Indefatigable, some slump deposits, sands deposited by streams, and stringers of gravel derived from adjacent Saalian deposits.

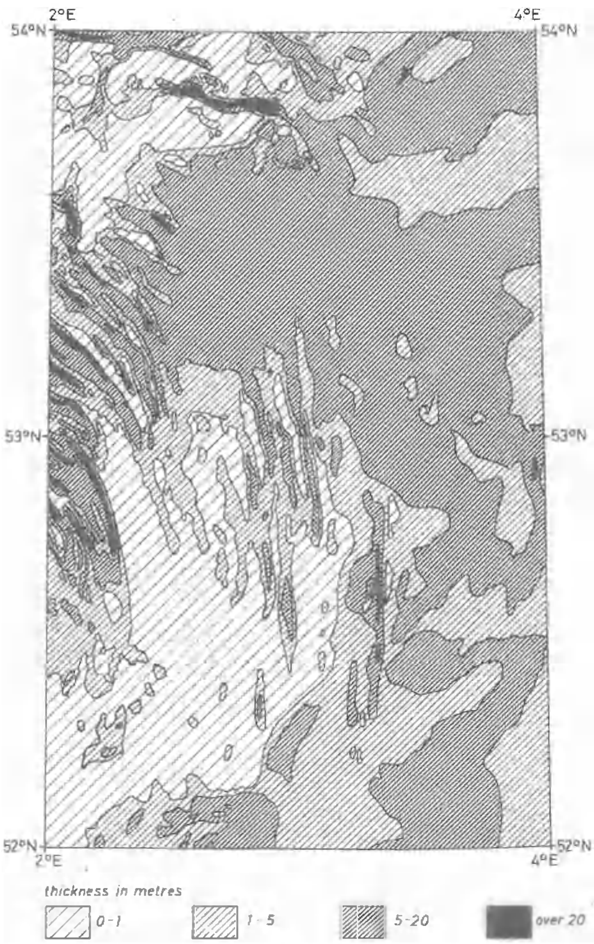

Figure 10 Thickness of the Holocene sediments in the Flemish Bight and Indefatigable sheets.

HOLOCENE

Regional sea level began to rise once more at the close of the Weichselian ice age, and the earliest brackish-marine incursion into the Southern Bight of the North Sea may have occurred as early as 10,000 years BP (EISMA *et al.*, 1981). With continuing sea level rise, tidal flat sedimentation became more widespread between 8000 and 7000 years BP and the environment over most of the southern North Sea became fully marine after c. 7000 years BP (EISMA *et al., op cit.*). The Holocene sediments in the area are generally between 5 and 15 m thick (figure 10). They are up to 30 m thick in the sand banks in the west and in sea bed depressions, but are less than 5 m thick where they overlie the till of the Bolders Bank Formation in the northwest.

The Holocene sediments have been subdivided into six formations on the BGS/RGD Sea Bed Sediment maps of the Flemish Bight and Indefatigable sheets.

a. Elbow Formation

Intertidal and shallow subtidal sediments are widespread at the base of the Holocene succession in the Netherlands sector of the Southern Bight and are mostly between 1 and 6 m thick. The Elbow Formation (OELE, 1969) comprises fine- or very fine-grained, grey, muddy sands and clays and contains a characteristic *Spisula subtruncata* bivalve assemblage (SPAINK, 1973). A brackish-marine clay is present beneath the sands in much of eastern Indefatigable and locally overlies early Holocene peat. The contrast in lithology and macrofauna across its upper boundary suggests a stratigraphic hiatus between the Elbow Formation and the overlying sediments.

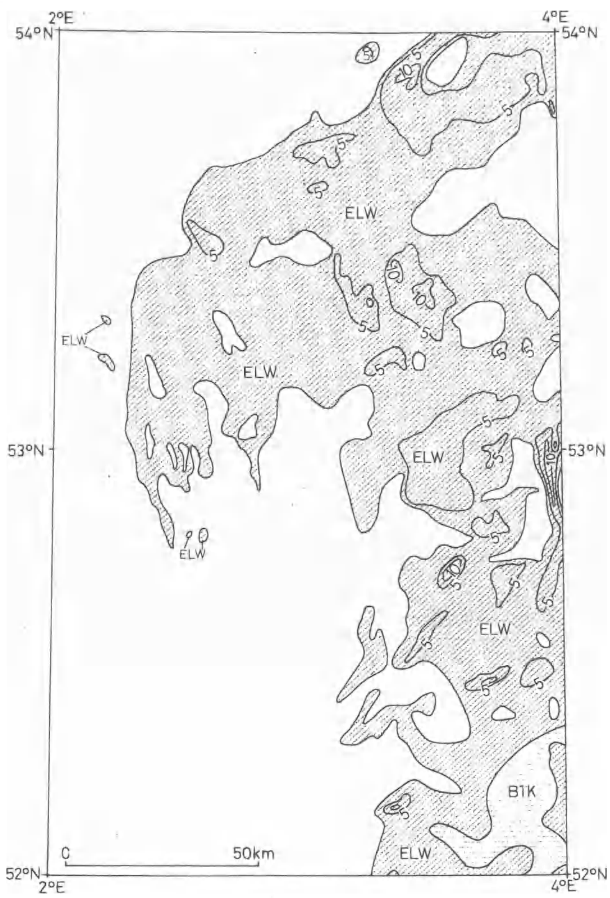

Figure 11 Map of the basal Holocene deposits in the Flemish Bight and Indefatigable sheets, with contours of their thickness in metres.
The formations are : ELW = Elbow Formation. BTK = Buitenbanken Formation.

In the UK sector, the Elbow Formation is restricted to southern Indefatigable and northwestern Flemish Bight (figure 11), comprises muddy sands and clays, and is mostly less than 5 m thick. A basal discontinuous peat bed is present in the west at c. 37 m below mean sea level and is interbedded with fluviatile and wind-blown sands. A sample of this peat from the Leman Bank area in southern Indefatigable has yielded a radiocarbondate of 8425 ± 170 years BP (JARDINE, 1979).

b. Buitenbanken Formation

The Buitenbanken Formation is restricted to southeastern Flemish Bight (figure 11), where it traces laterally into, but may be locally younger than the Elbow Formation. Its sediments comprise medium-grained sands, occasionally fine- or coarse-grained, and locally gravelly. The sands contain a nearshore *Spisula subtruncata* bivalve assemblage (SPAINK, 1973), but commonly have a significant component of reworked Eemian and Lower Pleistocene shells. The formation is mostly less than 2 m thick and has a maximum thickness of 5 m.

c. Bligh Bank Formation

A blanket deposit of medium-grained or fine- to medium-grained, clean, yellow-brown sands with local mud laminae is present at sea bed throughout Flemish Bight, extends into southern Indefatigable, and comprises the Bligh Bank Formation on the BGS/RGD maps (figure 12). These marine sands are commonly gravelly at their base and contain a sparse, but characteristic *Angulus pygmaeus* bivalve assemblage (SPAINK, 1973). There are abundant sand waves up to 12 m in amplitude throughout Flemish Bight, and the Bligh Bank Formation includes the sediments both within these and within a system of linear sand

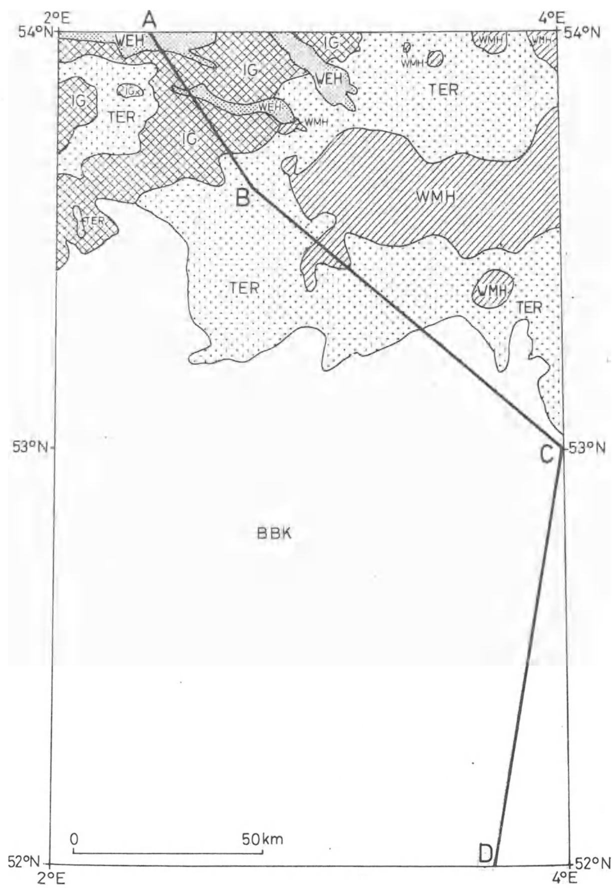

Figure 12 Map of the uppermost Holocene deposits in the Flemish Bight and Indefatigable sheets.
The formations are : IG = Indefatigable Grounds Formation. TER = Terschellingerbank Member of Nieuw Zeeland Gronden Formation. WMH = Western Mud Hole Member of Nieuw Zeeland Gronden Formation. WEH = Well Hole Formation. BBK = Bligh Bank Formation.

banks up to 30 m in height to the northeast of East Anglia. Between the sand waves and sand banks, the formation is mostly less than 5 m thick in Flemish Bight, and is between 1 and 10 m thick in Indefatigable.

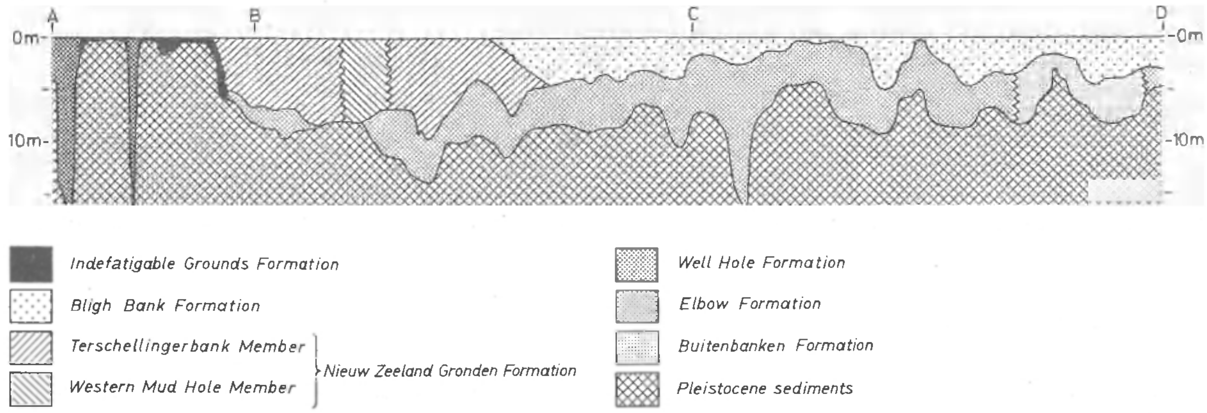

Figure 13 Schematic cross-section through the Flemish Bight and Indefatigable sheets, showing the typical thicknesses and stratigraphic relations of the Holocene units ; the line of section is indicated in figure 12.

The 1:250,000 BGS/RGD maps of sea bed sediment distribution are based on laboratory analyses of the uppermost 10 cm of the Holocene sediment layer. These analyses have indicated that the surface sands of the Bligh Bank Formation become slightly finer grained towards the northeast, which is the direction of net sediment transport in Flemish Bight (JOHNSON *et al.*, 1982), but there is no significant vertical grain size variation within the cores. Detailed sampling across individual sand waves in the Netherlands sector has revealed that the surface sediments become coarser grained from the troughs between adjacent sand waves towards their crests (unpublished RGD data). Most of the sediments of the Bligh Bank Formation were probably derived by reworking of Middle and Upper Pleistocene fluviatile and marine sands from between the Straits of Dover and southern Flemish Bight.

d. Nieuw Zeeland Gronden Formation

The Nieuw Zeeland Gronden Formation occurs mainly in the centre and east of the Indefatigable sheet. The formation consists of sandy and silty marine sediments which were probably derived by reworking from Pleistocene glacial and periglacial deposits, and has been subdivided into two members on the BGS/RGD maps.

The Terschellingerbank Member consists of grey or olive-grey, slightly muddy sand and contains a sparse, open marine molluscan fauna. The grain size of the sand decreases from 130-220 μm in the north to 100-180 μm in the south. Adjacent to the subcrop of the Bolders Bank Formation, the member locally contains intercalations of gravel. The Terschellingerbank Member varies between 1 to 10 m in thickness, and passes gradationally southwards into the Bligh Bank Formation in southern Indefatigable.

The Western Mud Hole Member in east-central Indefatigable consists of olive-grey, muddy, fine- or very fine-grained sands or sandy muds, and contains an open marine molluscan fauna including abundant *Turritella communis*. The member is typically between 2 and 5 m thick, but increases to about 7 m in thickness towards the west.

e. Indefatigable Grounds Formation

The Indefatigable Grounds Formation (figure 12) comprises a thin veneer, mostly 0.2-2 m but locally up to 5 m thick, of yellow-brown, gravelly, fine- to coarse-grained sand and sandy gravel overlying the Weichselian till in northwestern Indefatigable. The sediments contain a mixed assemblage of intertidal and subtidal molluscs (SLIGGERS, 1985). Side-scan sonar images of the sea bed have indicated that the gravelly sediments are locally overlain by ribbons of less gravelly sand.

f. Well Hole Formation

The Well Hole Formation occurs as the uppermost unit of fill within the partially filled late Weichselian valleys of Botney Cut, Markham's Hole and the Outer Silver Pit in northwestern Indefatigable. The sediments rest above an erosion surface on the Botney Cut Formation in the centre of the valleys (figures 9 and 12) and on older Pleistocene sediments along the valley margins. The Well Hole Formation consists of very soft, very fine- to fine-grained muddy marine sand or sandy mud, which is locally laminated. The formation is mostly between 5 and 20 m thick, but is more than 20 m thick in areas within Markham's Hole and locally along the southern margin of the Outer Silver Pit.

DISCUSSION

Deposits related to all three of the principal glaciations of northwestern Europe are represented in the area, and their distribution and sedimentary facies can be used to reconstruct the palaeoenvironment of the southern North Sea during these separate ice ages. Eustatic fall in sea level caused the regional shoreline to retreat to well outside the Flemish Bight and Indefatigable sheets during all three glaciations ; sea level at the maximum of the last ice age seems to have been approximately 110 m below its present level (JANSEN *et al.*, 1979). There was a continuous cover of land ice from the Netherlands across the northern part of the area to eastern England during Elsterian times. The deeply incised Elsterian valleys in Indefatigable are thought to have formed by the outbursts of large volumes of meltwater through the base of this ice sheet, most likely during the early stages of glacial decay.

There was no connection across the southern North Sea between either the Saalian or Weichselian ice sheets of Northwestern Europe with those of the British Isles during the two subsequent glaciations; the Saalian till of western Europe extends to only 40 km offshore of the present western coastline of Holland. Periglacial conditions prevailed in most of the southern North Sea through Saalian and Weichselian times.

Eustatic rise in sea level caused the North Sea to transgress across the area during the climatic ameliorations at the onset of the Holsteinian, Eemian and Holocene stages. EISMA *et al.* (1981) have shown that the Holocene transgression did not become extensive until about 9,000 years BP. The Eemian transgression in the Southern Bight may have been delayed until the E3 pollen zone (JELGERSMA, 1979). The precise timing of the transgression of the Holsteinian sea is not yet known.

The late Quaternary succession of the southern North Sea contains a much wider range of sedimentary facies than the Lower and early Middle Pleistocene deposits described by CAMERON *et al.*(this volume). This reflects a marked increase in the fluctuation of regional palaeoclimate and sea level since pre-Elsterian times. There was a significant reduction in the supply of clastic sediment from the peripheral land areas to the southern North Sea basin, initiated during Middle Pleistocene times ; the late Quaternary succession is mostly less than 50 m thick offshore, whereas the Lower and early Middle Pleistocene sediments are commonly more than 200 m thick. It is more difficult to assess whether the southern part of the North Sea continued to subside relative to the adjacent land areas throughout this period, although the Southern North Sea is at present some 40 m or more deeper than during early Middle Pleistocene times, when sedimentation was still mainly fluvial (CAMERON *et al.*, this volume).

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13 Eemian and Holocene sedimentary sequences on the Belgian coast and their meaning for sea level reconstruction

F. MOSTAERT and G. DE MOOR

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

INTRODUCTION

The Quaternary deposits of the eastern part of the present-day Belgian coastal plain comprise two major marine sequences : Holocene and Eemian sedimentary layers. These sequences are composed of coastal, nearshore, perimarine and tidal flat deposits.

In this paper emphasis is put on the comparison of these marine sequences.The lateral and vertical distribution pattern of Eemian and Holocene lithofacies together with the sedimentary characteristics of the units allowed to compare local palaeogeographical conditions during both interglacial periods. Furthermore this paper deals with the interpretation of important palaeomorphological surfaces, which might explain the pattern of the different palaeoenvironments of sedimentation conditions. The substantial amount of data makes it possible to draw conclusions with regard to the deduction from the sedimentary record of coastline migration and relative sea level changes.

Conclusions are based on numerous test borings, on the study of temporary excavations and on the reinterpretation of other available geological data. The techniques applied during the field and laboratory investigation included a detailed analysis of the sedimentary characteristics in the various deposits. Mollusc shell assemblages, micropalaeontological aspects, sedimentary structures, granulometrical and petrological characteristics made it possible to get a fair impression of large- and small-scale facies variations and of environmental changes both in time and regionally.

EEMIAN AND HOLOCENE DEPOSITS OF THE EASTERN PART OF THE BELGIAN COASTAL PLAIN

The present-day Belgian coastal plain is bordered to the south by the uprising Pleistocene and Tertiary substratum. The outcropping deposits consist of Holocene intracoastal tidal deposits and of present-day beach and dune sediments (figure 1). In this text attention is focused on the southern margin of the coastal plain near to Brugge.

Apart from the Holocene sequences, important Eemian marine deposits were identified most of the time underneath fluvioperiglacial and aeolian sediments of Weichselian age. Often Eemian deposits are immediately covered by Holocene marine layers. No older Quaternary marine deposits were found in the eastern coastal plain.

Both the Eemian and Holocene sequences reflect very much the same depositional conditions. The area was characterized by the existence of a barrier coast with intracoastal tidal sedimentation under rising sea level conditions. The impact of sea level lowering during the Weichselian with the associated fluvioperiglacial erosion, did not succeed in scouring completely the Eemian transgressive sequence. It is obvious that the Holocene sequence corresponds to a transgressive prism as well. Till now, regressive interruptions, if any at all, were rather unimportant.

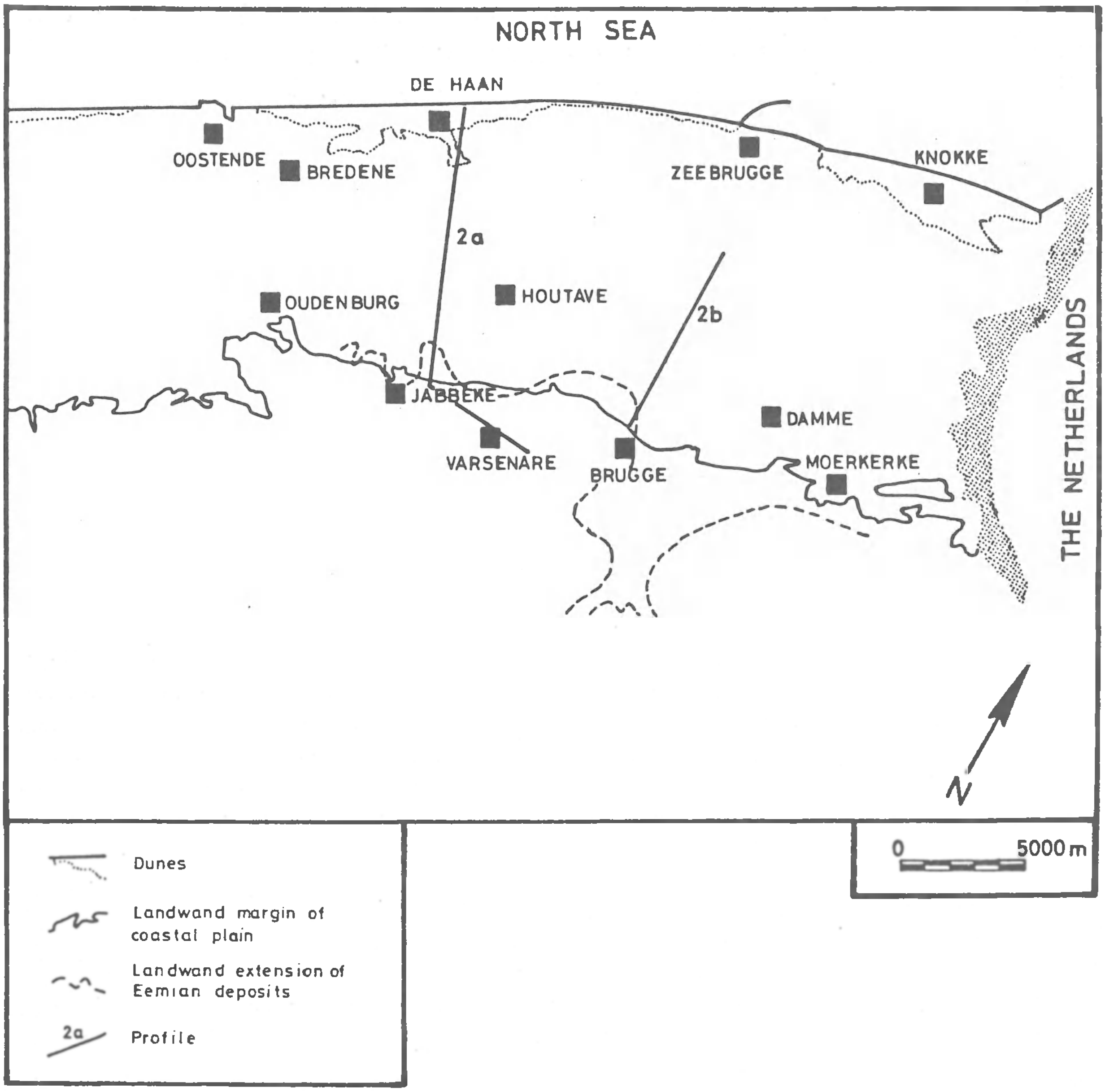


Figure 1 Situation of the study area.

The major difference between Eemian and Holocene sequences concerns the areal and vertical distribution pattern of the sedimentary units and the corresponding depositional environments. This can be proved by means of the synthetical geological profiles (figures 2a and 2b) and by means of maps with specific facies distribution (figures 3a, 3b and 3c). An attempt is made to visualize the palaeogeographical changes during Eemian phases periods with specific sea level stands.

The maximum landward extension of intertidal sediments, reflecting the highest interglacial sea level stands, was very much the same for both the Eemian and Holocene (figure 1). However, Eemian tidal flats locally penetrated further to the south than during the Holocene period in a valley system in the neighbourhood of Brugge. A Late Weichselian aeolian sand ridge dammed this valley system, preventing southward penetration of Holocene tidal impact. Although highest sea level stands reached comparable heights during both interglacials, the most southern margin of open marine Eemian conditions was once situated about 7 km south of the most landward Holocene coastal barrier system (figure 3c). This explains the rather restricted area with preserved Eemian intracoastal marine deposits contrary to the Holocene intracoastal extension corresponding to 95 % of the present-day coastal plain.

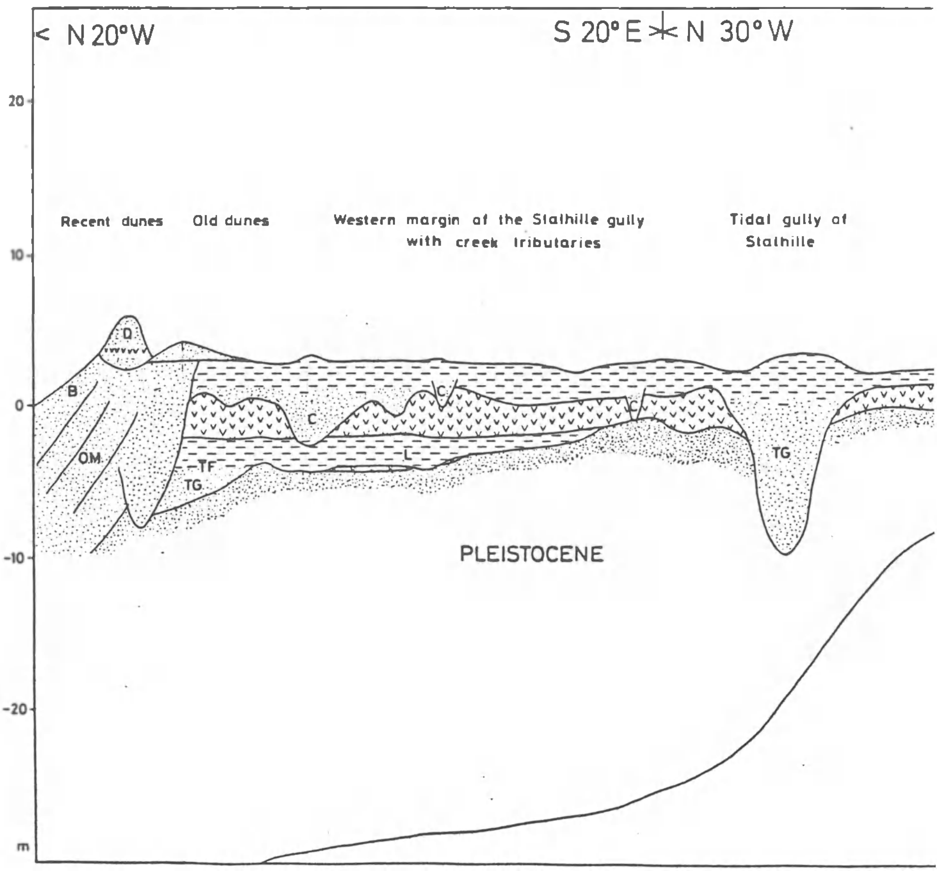
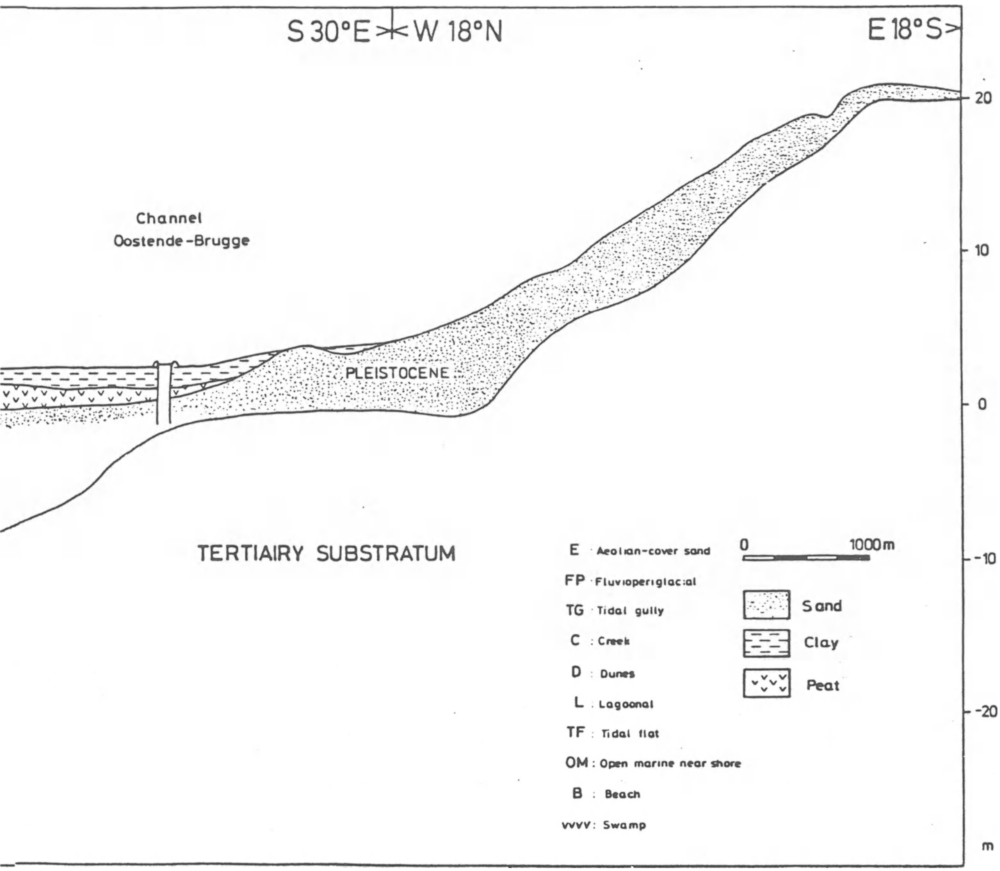


Figure 2a Holocene sequences.


S 30°E
W 18°N
E 18°S
Channel
Oostende-Brugge
PLEISTOCENE
TERTIAIRY SUBSTRATUM
20
10
0
-10
-20
m
0
1000m
E : Aeolian-cover sand
FP : Fluvioperiglacial
TG : Tidal gully
C : Creek
D : Dunes
L : Lagoonal
TF : Tidal flat
OM : Open marine near shore
B : Beach
vvvv : Swamp
Sand
Clay
Peat

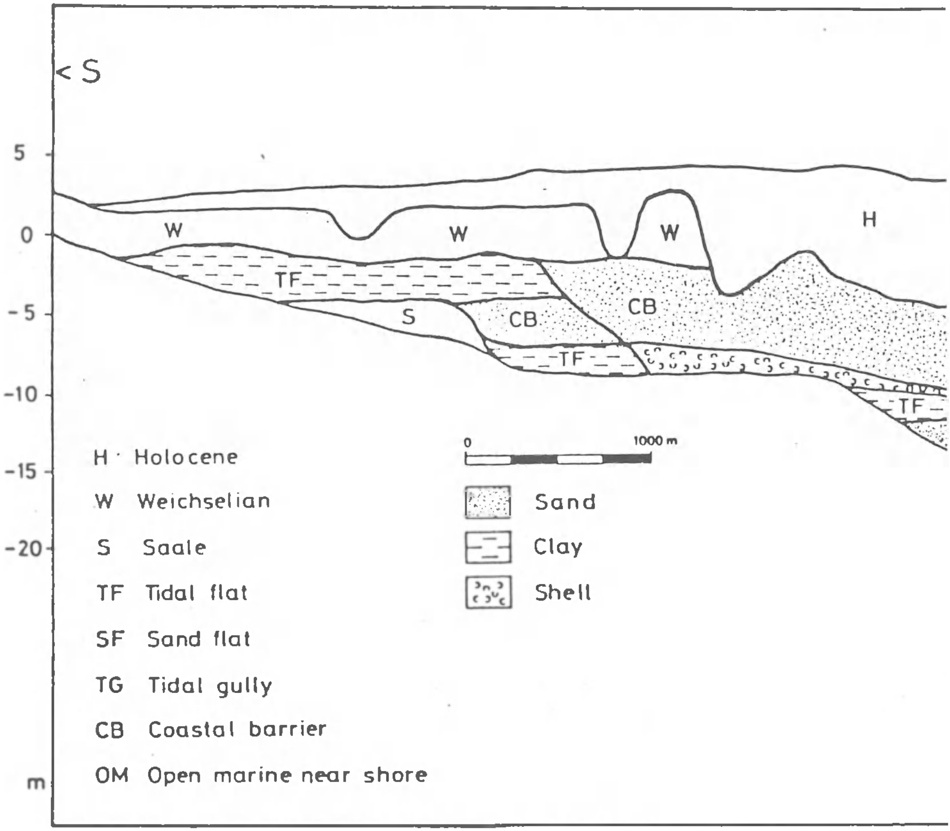
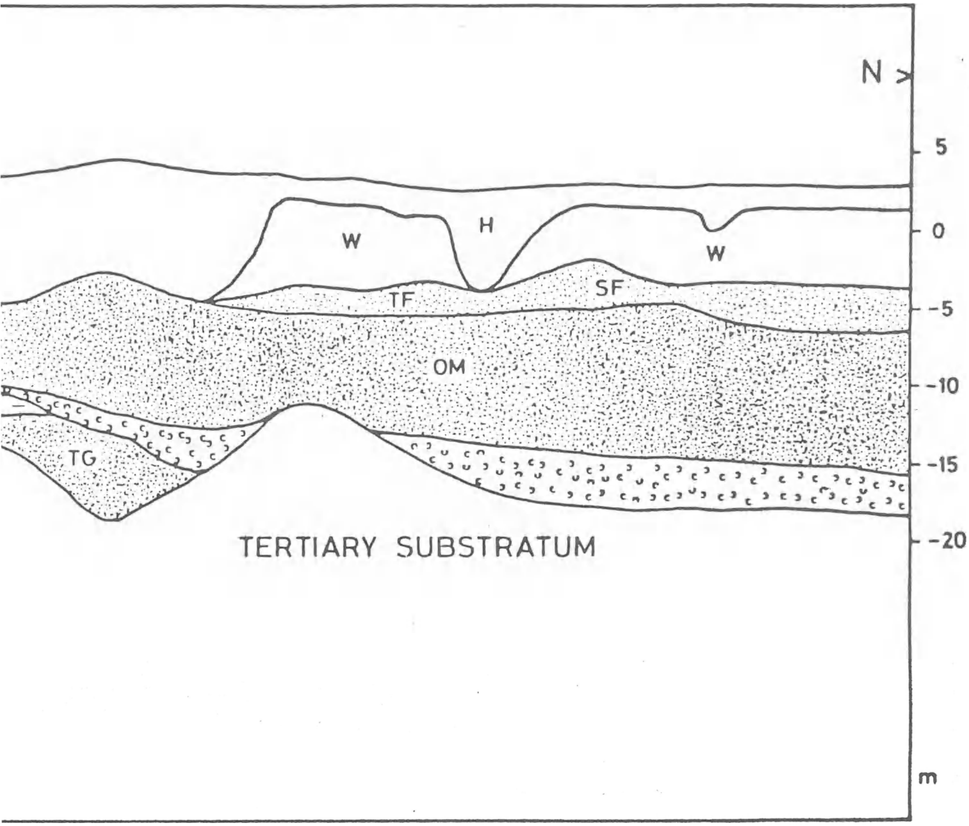


Figure 2b Eemian sequences.



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TERTIARY SUBSTRATUM

The synthetical sections (figures 2a and 2b) illustrate the stratigraphical position of the lithological sequences with the interpretation of the corresponding environmental conditions. In this part of the Belgian coastal plain, the thickness of the Eemian remnants is far more important than that of the Holocene strata. The Eemian layers form a wedgelike sedimentary body which thickens to the north. The northward thickening Holocene layers show important irregularities corresponding to tidal gully infillings and to the Upper Holocene beach and nearshore deposits.

Contrary to the Holocene sequences little or no important tidal gully deposits could be traced in the Eemian intracoastal area. The most important Eemian tidal gully corresponds to the Eemian estuary of the Reie-Waardamme river (figures 1 and 3a). Most of those sandy gully infillings were more susceptible to erosion during the Weichselian period than the surrounding clayey mudflat deposits. Therefore the deepest Weichselian river incision corresponds more or less to the course of an Eemian tidal channel in so far that nowadays little marine gully sediment is left in situ. The Weichselian fluvioperiglacial sandy sediments contain reworked marine mollusc shells which originate from the Eemian tidal channel and tidal flats.

The intracoastal Holocene sequence is characterized by an alternation of peat and tidal flat deposits, while extended Eemian peat layers are almost lacking. Rather thick Eemian mudflat sequences are found, showing continuous tidal flat conditions during sea level rise. Remnants of Eemian coastal barriers, important enough to prevent the area from tidal action have not been detected, while Holocene coastal barriers protected the hinterland temporarily to such an extent that swamps could develop. The absence of dune remnants in the presumed coastal area of the Eemian period might give an extra indication of the poorly developed coastal protection in comparison with the existing Holocene one, even during active transgression phases.

During the Eemian, the region of Brugge was characterized by an exposed tidal flat merely protected by low beach ridges comparable with the present-day German Bight conditions (REINECK *et al.*, 1968), while the conditions during the Late Holocene transgressive phases resembled present-day Wadden Sea conditions with larger tidal range however. During peat development and during the so-called Subatlantic (Late Holocene) regressions the coastal barrier system was nearly closed.

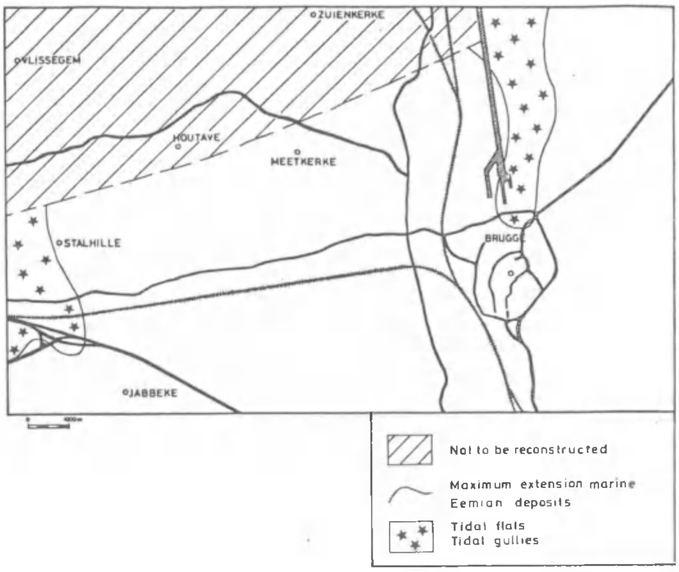
RECONSTRUCTION OF THE LANDSCAPE EXISTING BEFORE THE LATE QUATERNARY MARINE INGRESSIONS

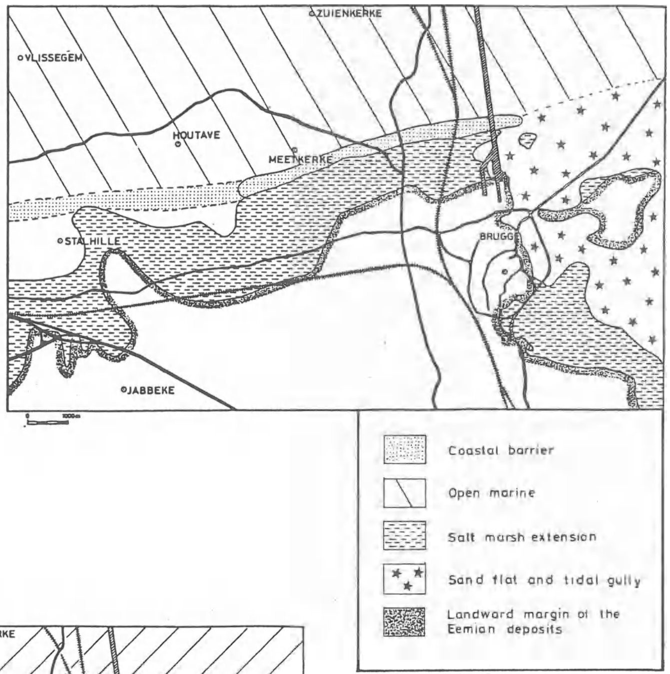
The palaeorelief which got influenced by the Eemian transgression was totally different from the Early Holocene landscape. This explains the major differences of the areal and vertical distribution pattern of the Holocene and Eemian sedimentary environments. The Eemian marine ingression attacked regressively on old, partly infilled valley system. As soon as sea level reached -13 m OL (OL = Ostend Level), tidal flats already developed in the immediate surroundings of Brugge (figure 3a).

Holocene tidal deposits came into existence on a palaeosurface of Weichselian fluvioperiglacial deposits covered with a thin coversand sheet and with east-west orientated coversand ridges. This seaward dipping topography was lying high enough to prevent the largest part of the eastern coastal plain from marine influence until the Dunkerquian I flooding (2500 BP).

The pre-Eemian landscape

A geomorphological map of the basal surface of the Quaternary deposits is presented (figure 4) indicating the morphography, the genesis and the age of the dominant erosion phases. The pre-Eemian morphology is partly reflected in the palaeosurface of the base of the Quaternary. Before the Eemian transgression reached the area of Brugge, a valley and interfluvium morphology existed. This morphology is more or less preserved in the southern part of the studied area. In this region only minor modifications of the erosion surface at the top of the Eocene sediments have taken place due to Eemian and Holocene tidal gully erosion or due to Weichselian fluvioperiglacial phenomena and mass movements.





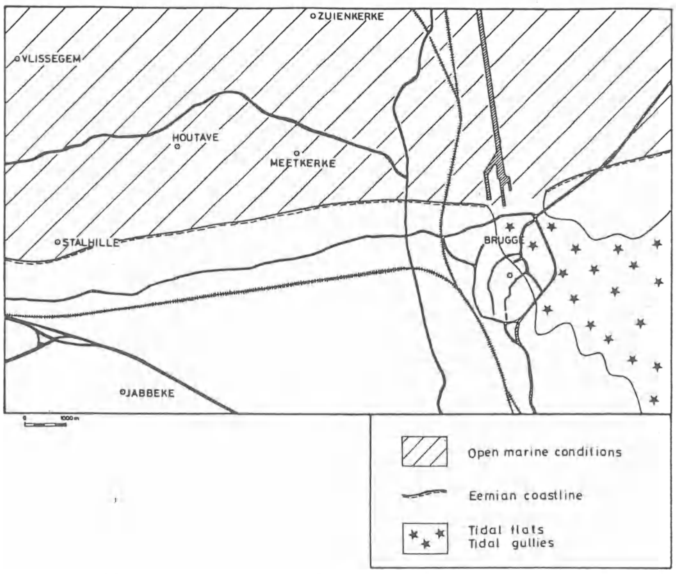


Figure 3 **a.** Sea level -10 OL.
b. Sea level -5 OL.
c. Eemian highest sea level stands.

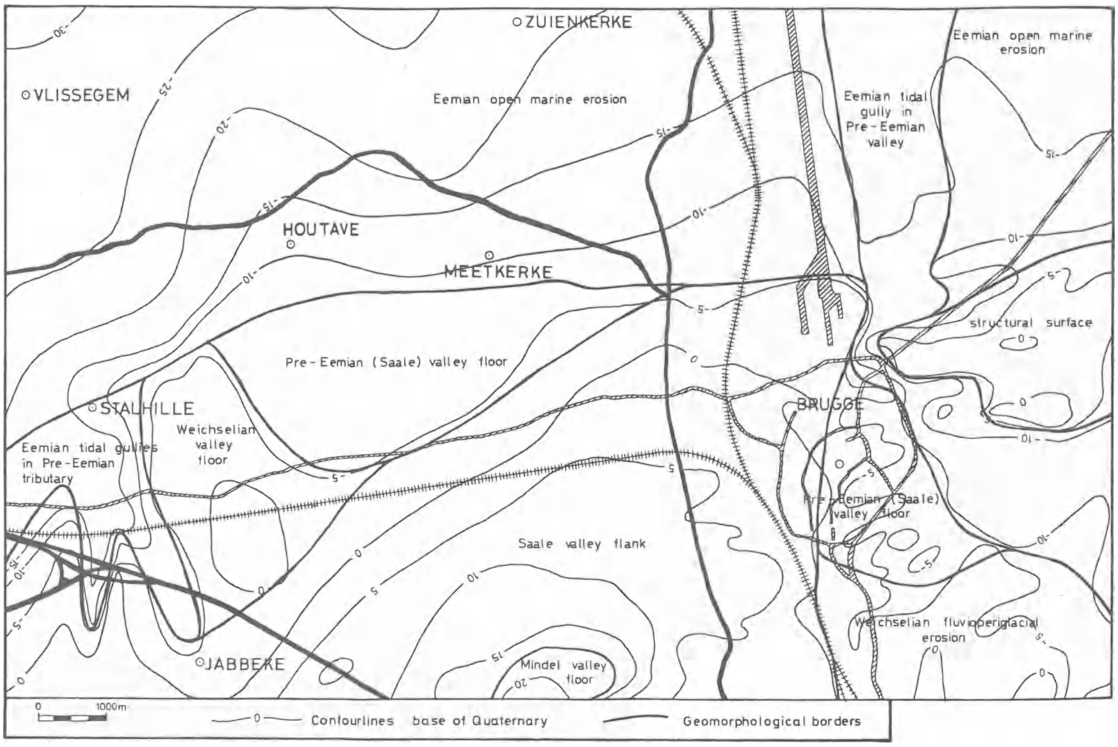


Figure 4 Geomorphology of the surface at the base of the Quaternary.

The Quaternary erosion phases have been conditioned by the lithological characteristics of the Tertiary substratum (figure 5). This substratum consists of slightly northeasterly dipping (1 %) alternating clay and sand units of some 5 to 20 m thickness each. The hills to the southwest of Brugge (St.- Andries), reaching an elevation up to 20 m OL were protected from fluviatile erosion by a semi-continuous sandstone layer which covers rather erosion-sensitive sands. Southeastward of Brugge (Oedelem), resisting Bartonian clay layers occur giving rise to a scarp-like landscape. To the northeast of Brugge (St. Kruis) Upper Paniselian shell layers with intercalated sandstone levels could not be removed by fluviatile nor by marine erosion.

Apart from the Waardamme Valley, a smaller pre-Eemian river system existed to the west of Jabbeke and Stalhille. This valley incision lays in front of a now buried cuestalike ridge formed by a 8 m thick clay layer (Merelbeke clay). This river system came into existence before Eemian marine ingressions reached the area after the general northward dipping slope of the Tertiary substratum already existed.

Both the Waardamme and the Jabbeke Valleys are tributaries of a larger valley system running parallel to and underneath the present-day coastline (the Ostend Valley). This palaeovalley runs about perpendicular to the strike of the slightly northeast dipping Tertiary layers. To the west, the valley joins an analogue valley underneath the western coastal plain. This buried valley system found its way to the North Sea area northwesterly of Ostend (MOSTAERT *et al.*, this volume). The relationship of this pre-Eemian valley system with the Flemish Valley, situated further to the east is not clear yet.

The map (figure 5) combining the Tertiary lithology and stratigraphy with the topography of the base of the Quaternary shows that the Waardamme river found its way to the north eroding sands (Vlierzele sand), sand-clay layers (Oedelem-Beernem-Aalter sands) and clays (Bartonian, Asse clay). This river axis followed a pre-Holsteinian consequent river direction. Further south of Oedelem, the Waardamme river system has a subsequent orientation. The localization of a consequent river system in the surroundings of Brugge breaking through the Bartonian cuesta is made possible by lithological facies changes and geometrical changes within the Bartonian layers.

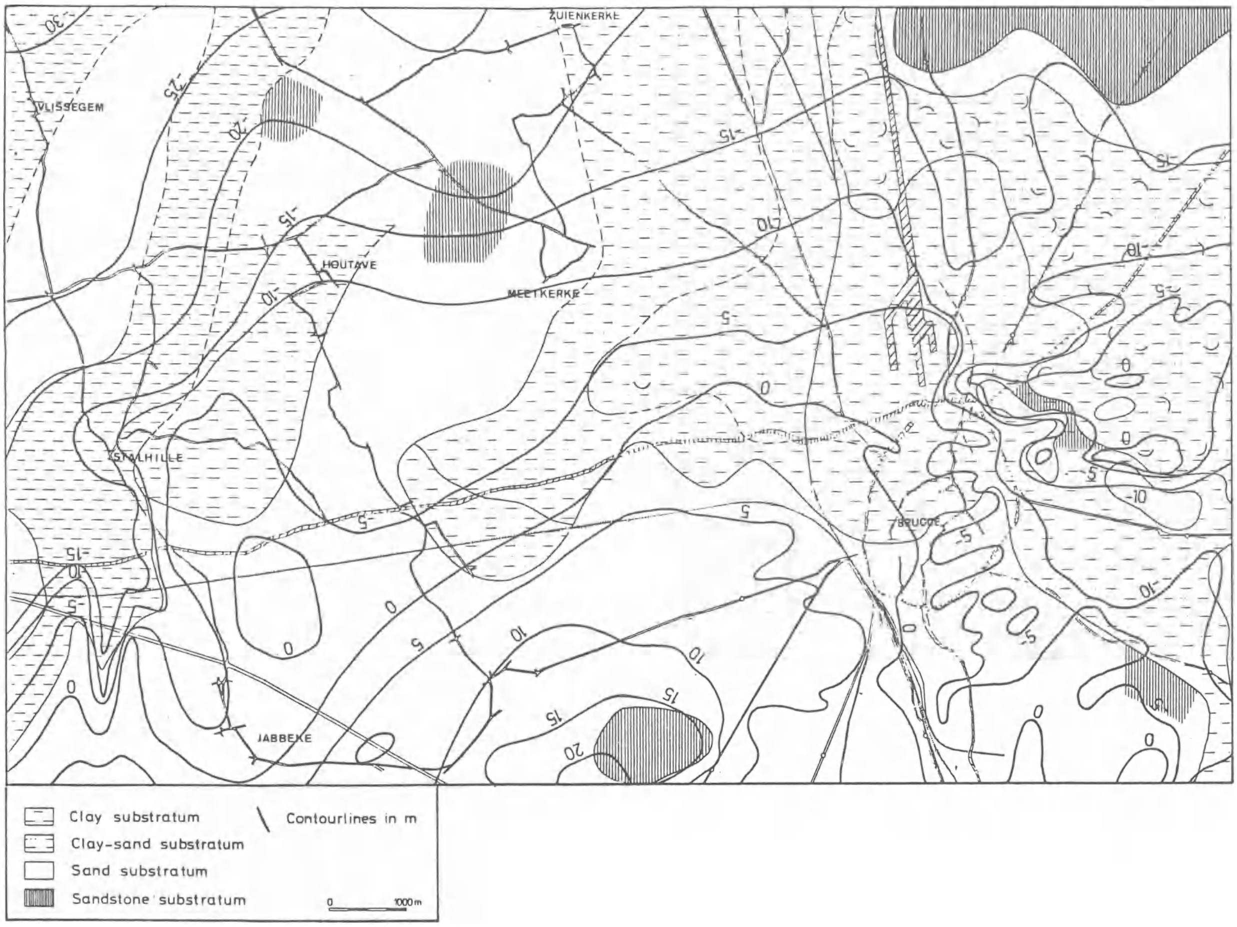


Figure 5 Lithology of the Tertiary substratum and base topography of the Quaternary.

The part situated higher than -7 m OL existed before the Eemian sea reached the area. Originally it corresponded with what was the southern flank of a Saalian river system. During the Eemian and especially during the Weichselian period only slight modifications of this slope occurred due to mass movements and cryoturbations. Below -7 m OL Eemian erosion deepened and steepened the original Ostend Valley. As no Saalian sediments have been found below -7 m OL in the Waardamme tributary and in this part of the Ostend Valley we suspect marine Eemian erosion to have deepened this area from -7 m to more than -20 m OL. The area where Eemian salt marsh deposits immediately covered the pre-Eemian sediments without previous marine erosion prove that minimum incision depth of the pre-marine fluviatile landscape was at least -7 m OL.

The pre-Eemian landscape of the southern part corresponds approximately to the erosion surface of the base of the Quaternary. The existing older Quaternary deposits consisted of thin fluviatile valley infillings. In the area considered, the net effect of the Saalian period was lowering by erosion. Early Eemian fluviatile action might as well be responsible for the modelling of the erosion relief on which marine sediments came into existence.

Eemian tidal impact first occurred in the Waardamme Valley by regressive tidal gully erosion. Tidal gully erosion caused overdeeping of the original valley which allowed later tidal flat deposition at levels beneath the pre-marine valley floor level.

The palaeomorphology at the beginning of the Holocene

A detailed geomorphological map of the base of the Holocene deposits has not yet been achieved : a conceptual framework is presented on figure 6. This map illustrates the basal surface of the main Holocene peat layer which corresponds to the base of the Holocene except for the dotted area (figure 6) where older marine Holocene tidal flat and lagoonal deposits exist. Three main zones can be distinguished :

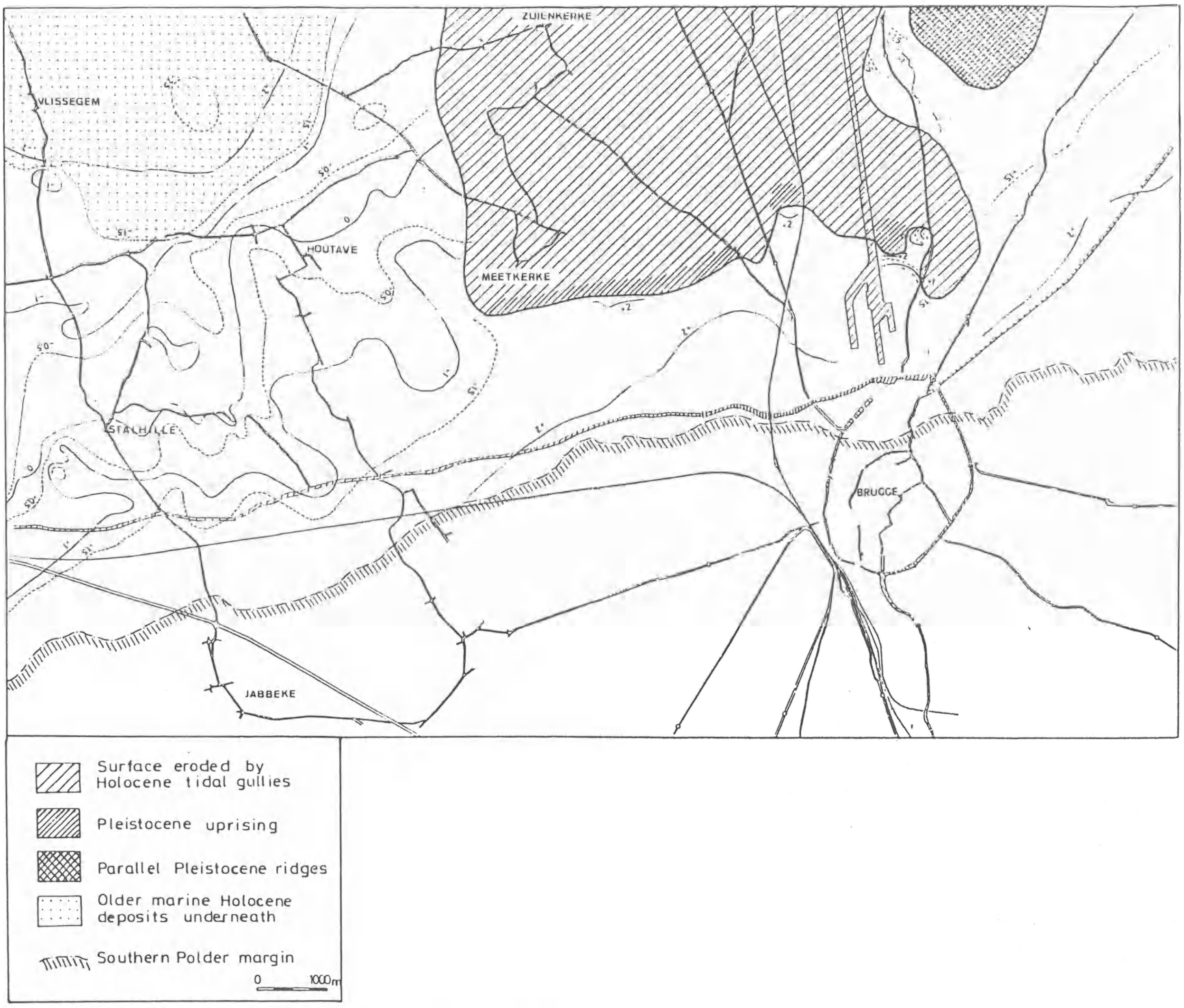


Figure 6 Base of the main peat layer. Reconstruction of the Pleistocene substratum.

- The area where Holocene phenomena were restricted to soil formations.

  This zone is bordered to the north by the coastal plain. It consists of an outcropping coversand ridge running east-west and it is merely interrupted by the small river system of Brugge, the relict of the large Weichselian Waardamme river.

- The region where the Late Weichselian landscape is preserved underneath a Holocene cover.

  Most of the times this Pleistocene substratum is covered by peat and/or salt marsh deposits. Often the pre-existing Early Holocene soil profile is preserved underneath the cover. Where older profile Holocene marine deposits occur underneath the main peat layer (dotted area on figure 6) the original Late Weichselian landscape is still intact as well.

- The area where former Holocene, laterally migrating tidal gullies or coastal processes shaped the Late Pleistocene palaeosurface.

The first two zones allow to reconstruct the Early Holocene landscape which did not change until marine influences reached the area or until peat growth started. It was a rather flat, slightly seaward dipping fluvioperiglacial surface. Southward, flat and low aeolian ridges with east-west orientation existed (e.g. north of the Moeren of Meetkerke), culminating in the most important ridge Gistel-Brugge-Stekene. This large ridge is a combination of several subridges with interjacent depressions. Microridges and depressions with wavelengths of 100-200 m and height differences of less than 2 m are found north of Dudzele and in Zeebrugge. In this landscape the unimportant Waardamme River system found its way to the north. The incision depth since 5600 BP was never below +3 m in Brugge and -2 m in Zeebrugge.

The rather flat and rather high lying surface of the Pleistocene substratum was extremely favourite for peat development due to rising water table conditions associated with sea level rise. Before 5600 BP, tidal impact only reached the extreme western part of the eastern coastal plain. There, tidal flat and lagoonal conditions existed. Extensive peat growth started from 5600 BP, which is obviously earlier than in the western part of the coastal plain. While sea level rose, peat development penetrated further to the south reaching about the present-day coastal plain border. Generally speaking, the eastern coastal plain is not flooded before 2500 BP, when mean sea level reached at least 0.5 to 1 m OL. This Subatlantic transgression occurred on a very flat peat landscape reaching top levels of +2 m, exceptionally up to 3 m OL. It can be proved that tidal action did not disappear from the coastal plain since 2500 BP until medieval reclamation. The Roman regression corresponds with local enlargement of the salt marsh area. Salt water influence still reached the southern border of the coastal plain. A post-Roman enlargement of some gullies could be derived from the sedimentary record, while other tidal gully systems did not change at all.

SEA LEVEL CHANGES

Both the Holocene and the Eemian sedimentary layers came into existence under varying sea level conditions. The relative elevation and stratigraphical position of mudflat and salt marsh deposits, of coastal plain peat layers, of subtidal and intertidal open marine and intracoastal facies, were the major criteria for Eemian and Holocene sea level reconstruction. Far more detailed sea-level studies can be produced for the Holocene sequences while absolute ^{14}C-datings of sea level indicators are potentially available because of the appearance of peat and shell layers corresponding to critical levels. Here, abstraction is made of the absolute datings and the same interpretation method was applied for both the Holocene and Eemian sequences of the study area. Figure 7 confronts Holocene and Eemian sea level interpretation of the area. The interpretation of the Eemian sea level indicators is based on assumptions of tidal range possibly varying between 2.5 and 5 m. The assumed Holocene tidal range is an extrapolation of the present-day range (± 4 m). Eemian sequences indicate at least one sea level lowering during the period of Eemian highest sea level stands. This lowering is deduced from the immediate superposition of subtidal open marine sediments by supratidal salt marsh deposits indicating important sea level lowering and seaward migration of the coastal barrier system (MOSTAERT and DE MOOR, 1985). Only minor sedimentary indications are left of the definitive sea level retreat at the end of the Eemian interglacial period.

Sedimentological, stratigraphical arguments and absolute datings cannot prove the existence of general sea level lowering during the Holocene. Palaeogeographical changes as the general appearance or disappearance of swamp conditions giving rise to peat layers, so-called regression and transgression phases can all be explained without the necessity of accepting real sea level changes. Local fluctuations of the tidal levels within the tidal flats itself might have occurred due to a combination and interaction of factors which are all related with the changing flood basin morphology. The flood basin effect certainly makes sea level interpretation of intracoastal zones very difficult as it can be very important (up to 2 m in the estuary of the river Scheldt, cf. KIDEN, this volume). The flood basin effect changes in time and is influenced by a number of factors, e.g. by tidal gully infilling or enlargement, differential sedimentation and erosion. The flood basin effect influences these sedimentation levels which are important high water indicators (e.g. the top of salt marsh deposits). The cyclicity of the so-called Dunkerque transgressive phases and their appearance over larger areas seem to indicate the control of a general process like sea level changes. However, the sedimentary sequences of the eastern coastal plain cannot entirely prove a general flooding at once in distinct transgression phases and even if so, other long-run mechanisms than sea level changes could have controlled the mechanism : open marine sediment dynamics, coastal dynamics, tidal range changes. Earlier investigations might have confused local gully development or successive gully shiftings with transgressions.

No such small-scale palaeogeographical changes due to minor high water level changes and corresponding with varying flood basin effects could be traced in the Eemian sequences. This is probably due to the exposed character of the area.

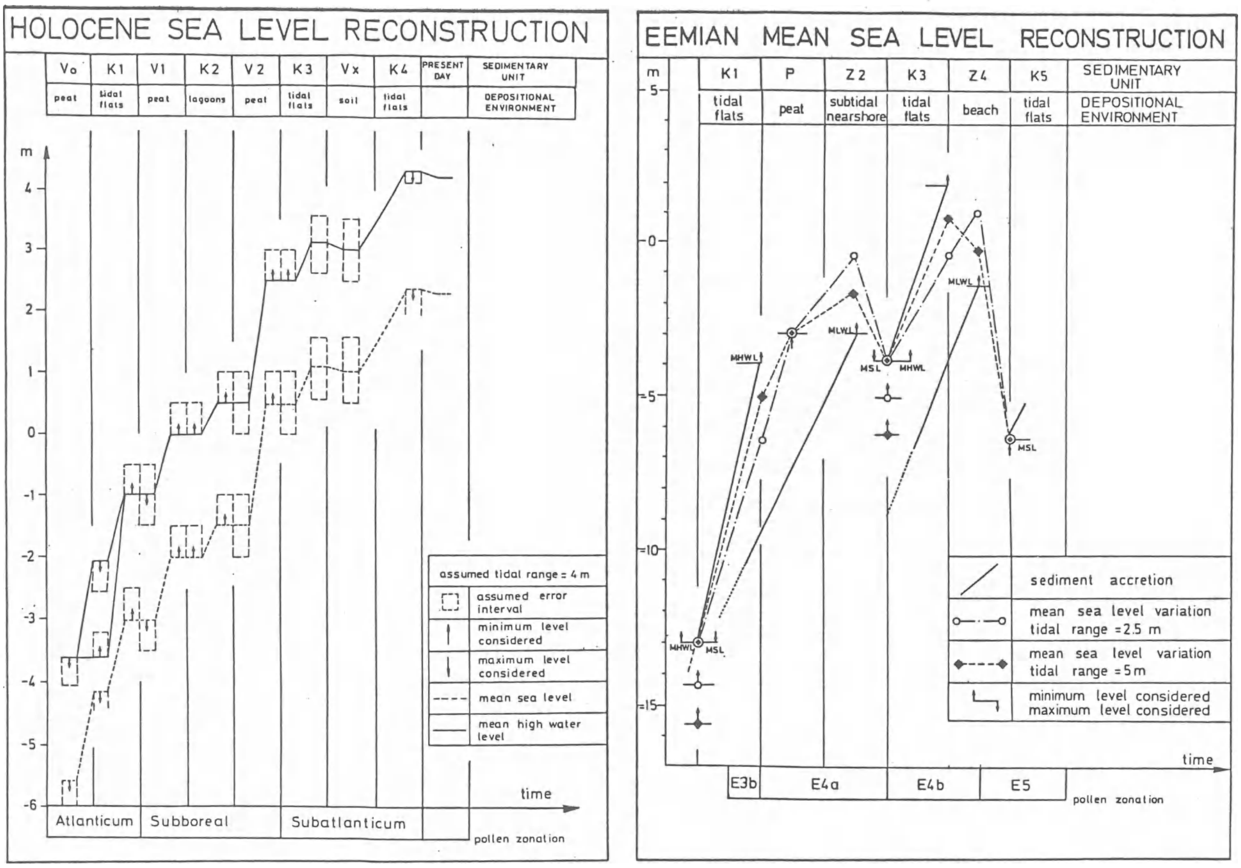


Figure 7 Sea level reconstruction.

CONCLUSIONS

The Eemian and Holocene sequences show important differences indicating different palaeogeographical conditions. These differences are highly conditioned by the existing landscape before the marine ingressions.

Another important factor with consequences for the variable palaeogeographical conditions is the appearance of the coastal barrier. During the Holocene a protecting coastal barrier system existed allowing peat development in the hinterland while the poorly developed Eemian barriers implied continuous tidal action.

The palaeogeographical consequences of a net sea level lowering during the Eemian could be detected and compared with Holocene transgressive and regressive effects which did not necessarily correspond with real sea level fluctuations.

The stratigraphical and sedimentological approach, the inventory of the facies, allowed to reconstruct sea evel and coastal migration for regions and for sequences of which no absolute datings are available.

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14 Modern deposits, quasi-deposits and some Holocene sequences in the Southern Bight, North Sea

A.H. STRIDE

23 Sussex Road, Petersfield, Hampshire GU31 4LB,

United Kingdom

ABSTRACT

The nature of the modern offshore marine deposits of the Southern Bight is clarified by removing from consideration sediment that has still not reached its final site of deposition. Many sand banks seem to fit into this category, particularly many of those that are located within the invaginations of the Southern Bight Bed-Load Parting. Late-stage evolution of the sand sheet facies is also discussed. The varied types of modern deposit of the present-day Southern Bight are used as a partial guide for a review of the diachronous deposits of the Holocene marine transgression. The flooding of the barrier separating the Southern Bight from the North Sea seems to be associated with a period of erosion that is recognisable at least in the vicinity of the Zeeland Ridges.

INTRODUCTION

The Southern Bight of the North Sea is a particularly fruitful region for which to make a study of modern marine sedimentation, as much is already known about sediments, bedforms, faunas, and of both long-term observations and numerical simulations of water movements. As a result it has already been possible to formulate some initial sedimentation process-response models. Moreover, the large size of the sand banks and the relatively good sorting of some of the sands means that their ancient analogues are of considerable interest as potential reservoirs for oil (HOUBOLT,1968).

Short-zone sediments are relatively well known, because of their widespread accessibility. The near-coast deposits of the Southern Bight have been singularly well revealed in recent years in some deep sections on land and within coffer dams in estuaries. These sections show the degree of detail that should be sought in studies of the related deposits of the offshore zone.

The present paper airs some of the problems concerned with Holocene offshore marine sedimentation in the Southern Bight and suggests some tentative solutions. It seeks to distinguish between the real deposits of the present sea and the quasi-deposits, which (although recognisable as mappable units at the present time) probably have a low preservation potential. The improved understanding of the real deposits of the present sea aids understanding of the transgressive deposits and sequences of the early Holocene. It also facilitates recognition and understanding of analogous deposits in the stratigraphic record.

MODERN OFFSHORE MARINE DEPOSITS

a. Recognition of offshore sand facies

The known net sand and gravel transport directions for the Southern Bight greatly facilitate the finding of modern offshore tidal current deposits for that region. The nature of most of the abundant data for the net transport directions has been summarised elsewhere (JOHNSON *et al.*, 1982) and need not be repeated in detail here. These directions are supported by more recent data (SMITH, 1984a ; 1984b).

The essential element of the net sand transport pattern is the presence of the Southern Bight Bed-Load Parting, with southwestward net transport to south of it and northerly net transport to north of it (figure 1). The general outline of this bed-load parting has been confirmed by independent numerical simulations of maximum bed shear stress directions as well as of net sand transport directions (JOHNSON *et al.*, 1982). The apparent simplicity of the bed-load parting shown on the shear stress map near to the sand bank and

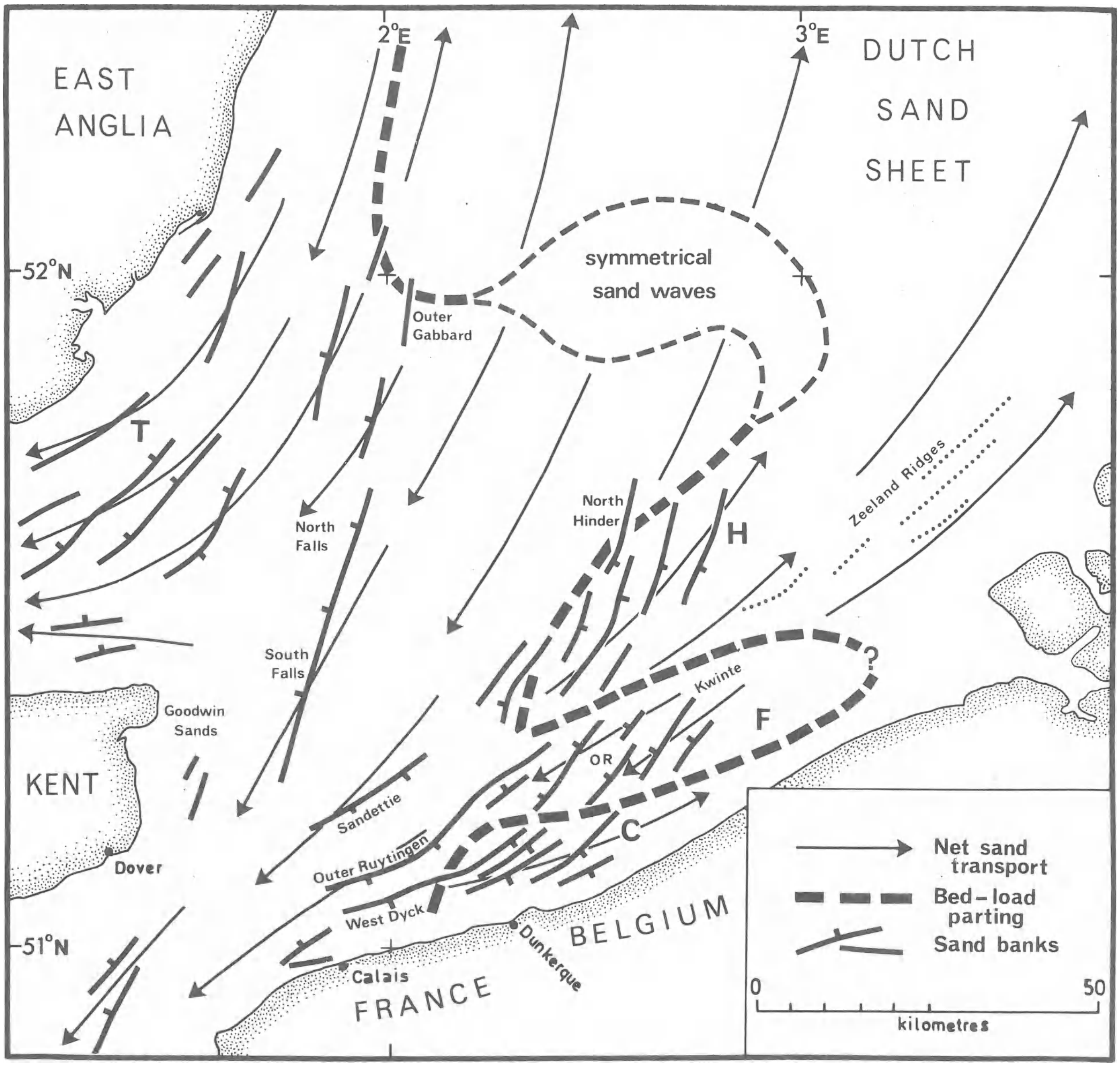

Figure 1 Net sand transport directions from the Southern Bight Bed-Load Parting. A tick indicates the steeper side of a sand bank, where known.
C = coastal sand banks near to Belgium (the nearshore banks are not shown). F = Flemish Banks. H = Hinder Banks. T = Thames Approaches Banks. OR = Outer Ratel Bank. After JOHNSON *et al.* (1982).

channel complex off Belgium is thought to be due to the inadequate representation of the relief by mean depth over each grid rectangle and by use of a constant friction factor.

Comparison of the northerly and southwesterly net sand transport directions with respect to the regional variation in tidal current strength (HOWARTH, 1982) reveals two strongly contrasting situations. To the north of the Southern Bight Bed-Load Parting the tidal currents weaken progressively in the direction of net northerly sand transport. In contrast, on the south side of that bed-load parting the currents strengthen progressively along the line of southwesterly net sediment transport. These differences are reflected in the grain sizes of the sediments on the two paths. On the northerly route the grain sizes of sand that can be transported decrease progressively northwards (except off Texel). In contrast, on the southerly route the grain sizes of the sediment that can be transported increase progressively southwestwards, and can include gravel as well as sand. The deposits of these two paths are, thus, very different (STRIDE *et al.*, 1982). The grain sizes of the sand accumulating along the northerly route fine to the north, so that the grain size at any point is essentially in equilibrium with the speed of the associated tidal currents. In contrast, to the south of the bed-load parting there is a lag gravel that should coarsen progressively southwestwards, in keeping with progressively increasing peak tidal current strength in that direction. In addition, any sand on that path will move along the whole route to its southern ends in the Thames Approaches or Dover-Calais Strait, unless it can accumulate as banner banks, say, with respect to fixed obstacles on the intervening floor. The grain sizes of the sand at these locations are not in equilibrium with tidal current strength, but the sand banks at the end of the route are presumable in some sort of dynamic equilibrium with the currents. The only comprehensive theoretical treatment and computer simulation (HUTHNANCE, 1982) shows that the long axes of sand banks will generally be aligned at small angles to the peak tidal flow directions, just as is found at sea for these existing sand banks. Thus, it can be assumed on reasonable grounds that these sand banks are in keeping with modern tidal currents and can have been formed by them. Indeed, there is no need to invoke stronger currents at low sea levels to account for them.

The markedly different conditions required for making the contemporaneous sand deposits to north and south of the bed-load parting show that the sand sheet and sand bank facies are alternative facies (STRIDE, 1984), in the sense used by MIDDLETON (1973) for other sediments.

The presence of shells of *Angulus pygmaeus* in the late Holocene sands of the Southern Bight (where living examples are rare today) has been cited as evidence that these shells were probably transported from the English Channel (where they are relatively common) along with sands and pebbles of chalk and chert (LABAN and SCHÜTTENHELM, 1981). This presumed northward transport is discounted for three reasons. First, it is thought to be more likely that the molluscs reached the Southern Bight as spat, carried there by the known slow northerly drift of water from the English Channel. The molluscs would have grown and thrived while conditions in the Southern Bight remained suitable. Secondly, the shells are believed to be too fragile to withstand such long transport (Personal communication, J.B. WILSON, 1984), especially as they would have been moved, presumably, by currents strong enough to transport the associated pebbles as far as 52°N (LABAN and SCHÜTTENHELM, 1981). Thirdly, the supposed strong net transport of sediments up to pebbles in size from Dover-Calais Strait northwards is contrary to the overwhelming mass of evidence of net southerly transport of sand and gravel, referred to above. The same southwesterly transport is thought to be applicable for the past 5,000 years, at least. Accordingly, it is suggested that a northern source should be considered for the chalk and chert pebbles.

b. The Dutch sand sheet facies and its approach to equilibrium

Only a few general comments need to be made here about the nature of the sand sheet facies, as its recognition has been discussed above (this section, a) it has been described in some detail elsewhere (STRIDE *et al.*, 1982). However, some suggestions are made about how the bed morphology may change as the grain sizes approach equilibrium.

First, the Dutch Sand Sheet shows a longitudinal grain size gradient. Secondly, the sand is essentially well sorted to very well sorted. Thirdly, sand waves are still present on much of the Dutch Sand Sheet, so that cross-bedding is present within it. Fourthly, from a knowledge of the conditions required to make sand waves in flumes it can be accepted now that small sand waves up to one metre high could be built at times by existing currents on all parts of the Dutch Sand Sheet where sand waves are known to be present today. Migration of small sand waves leads to the evolution of the large sand waves on which they stand, so the large ones can also be considered as evolving under modern conditions. The average height of these large sand waves decreases progressively northwards to a minimum of about 3 m (STRIDE, 1970). It is considered that the longitudinal scatter in these heights may partly indicate variations in the relative abundance of particular grain sizes that must be accommodated at any particular sites.

The overall northward decrease in the height of the large sand waves is thought to provide a measure of the approach of the grain sizes to equilibrium with the current strength. Thus, the best sorted sands are likely to be nearest to their equilibrium position with respect to current strength, so that they have the smallest of the large sand waves on them. When a particular grain size has almost reached its equilibrium position the large sand waves on it are expected to be smoothed out by the destructional forces of storm waves and non-tidal currents. The semi-random transport directions for sand provided by these water movements will then exceed the ability of the tidal currents to provide a dominant net sand transport direction any further north. Thus, the flattening of the large sand waves may begin at the fine sand end and will presumably migrate towards the coarser grain sizes when these approach their equilibrium positions, also.

The internal structure of a large sand wave that has been progressively reduced in height until it has been smoothed out, has not yet been determined in modern seas. However, this structure may be analogous to the observed progressive northwards longitudinal change in profile of the large sand waves in the direction of net sand transport. Thus, in vertical sequence there could be a progressive decrease in the angle of the lee slope (master-bedding) from about 15° to about 3° say (on the Dutch Sand Sheet) as the height of a crest becomes progressively reduced. The master bedding should also become planar. There would also be concurrent changes in the nature of the crossbedding due to the presence of smaller sand waves. The grain size of the sand at any site should perhaps only show a small upwards increase. Ultimately the sand wave facies should be overlain by a sand ripple facies. Exploration of the large sand waves is clearly required in order to test out these suggestions about internal structure and to obtain details. For the moment, the upper parts of the large sand waves are considered to be quasi-deposits, for although many sand deposits in the stratigraphic record show cross-bedding, attributed to sand waves, the morphology of sand waves themselves has not been recognised anywhere in that record. There is some resemblance between the supposed internal structure of these sand waves and a model proposed for the Tertiary Roda Sandstone (NIO, 1976). However, the analogy may not be valid as the height of the cross-bedding and its steepness in the Roda Sandstone are far in excess of these known in the Dutch Sand Sheet, and the depositional environments may well be different in the two cases.

The shelly fauna of the Dutch Sand Sheet shows marked longitudinal variations along its net sand transport-deposition path, in conjunction with the known longitudinal variation in bed morphology. Thus, the number of species and the number of individuals increase markedly on a longitudinal traverse from the zone of large sand waves, through the zone of small sand waves, to the zone of rippled sand (WILSON, 1982). This change is presumed to result from the increasing stability of the sands in that direction. When the large sand waves decay there should be an upwards increase in the richness of the fauna.

c. Sand bank facies

Little information need be given about modern sand banks or the development of sand bank facies as these have been described in some detail elsewhere (STRIDE *et al.*, 1982). The sand in a sand bank represents the full range of grain sizes that can be swept up into it from the transport path that leads to it. Thus, the sorting need not be good (HOUBOLT, 1968, figure 22). The sand may be more or less homogenous throughout a sand bank because of its process of growth. Sand waves are only associated with evolving sand banks and the expected cross-bedding representing them has been observed in cores.

As these sand waves are located in a region of relatively strong currents the associated cross-bedding of the sand bank facies should be generally steeper than in the case of the cross-bedding of the sand sheet facies (this section, b).

HOUBOLT (1968) provided the classic block diagram of an offshore tidal current sand bank. This shows master-bedding with a dip of only a few degrees, together with the high angle cross-bedding now recognised as due to sand waves. The model has been subsequently updated (STRIDE *et al.*, 1982) to allow for the differences between the head and tail of a sand bank and for the presence of aprons of large sand waves that extend beyond both ends of a bank (CASTON, 1981). Modern high-resolution continuous - reflection profiles give a clearer idea, than hitherto, of the internal structure of such a sand bank and thus of its evolution (Personal communication, G.F. CASTON, 1984). Such sand banks are considered to be quasi-deposits as they are still evolving.

The facies model proposed for an offshore tidal sand bank that has become moribund (STRIDE *et al.*, 1982) shows that the bank has progressively lost its relatively steep lee slope of 5° or so and has lost its sand waves. Its crest has been progressively eroded and the material spread out on its slopes. It is supposed that the large symmetrical sand waves that were present in its aprons during its active phase (CASTON, 1981) will have left traces of a distinctive style of cross-bedding at these locations in the deposit. The angle of the master-bedding of these sand banks should presumably decrease upwards during preservation, as in the case of the sand waves of the sand sheet facies.

The cross-sectional asymmetry of sand banks in an estuary, such as the Thames Estuary, implies that they may migrate across it. Nevertheless, they will still remain in a region of strong tidal currents. Preservation of these sand banks should not be associated, therefore, with lessening of the lee slope angles or with loss of sand wave structures (as in the case of the offshore sand banks, mentioned above). Where there is a continuing supply of sand to an estuary the sand banks must first build up to sea level, after which they can only grow laterally and develop a relatively flat upper surface. Ultimately, it is expected that neighbouring sand banks will unite. Their surfaces are likely to pass through a tidal flat phase before becoming dry land. Thus, the estuarine style of sand bank facies should be distinguishable from the offshore style of sand bank facies, mentioned earlier.

It is presumed that the North and South Falls and Sandettie Bank are examples of sand banks that are migrating towards Dover-Calais Strait. If this is the case then they should be considered as quasi-deposits. Alternatively, it remains to be proven that they are banner banks, tied to fixed obstacles on the sea floor, as has been suggested. The Outer Gabbard Bank (figure 1) may also be migrating southwards, like the other sand banks, just mentioned. It is also, seemingly, losing sand from its southern end, if the sand waves thereabouts are a fair guide (CASTON, 1981). It is not known whether there are any other sand banks that are also losing sand from their tails.

There are numerous sand banks in the Approaches to the Thames Estuary but remarkably few in Dover-Calais Strait, although both regions are located at the end of the same southward net sand transport path. This contrast may perhaps indicate that some sand on reaching the Dover-Calais Bed-Load Convergence can escape from it, to feed the sand banks off the coast of France or the Goodwin Sands off the Kent coast (figure 1). Tracer studies might resolve these questions.

The Goodwin Sands are presumed to be a banner bank. The anti-clockwise "rotation" of this sand bank during a period of about 100 years (CLOET, 1954) implies that there should be easterly dipping foresets within its southeastern end and westerly dipping foresets within its northwestern side.

There are few data available about the shelly faunas of the majority of the sand banks. The shelly fauna of offshore sand banks, with sand waves on them, appears to be as impoverished as the shelly fauna in the zone of large sand waves of the sand sheet facies (WILSON, 1982). Presumably the shelly fauna of offshore sand banks should become more varied and more abundant as such a sand bank becomes moribund, because of the decreasing amount of bed disturbance. However, this may not be the case for the estuarine sand banks, because of the continuing strength of the tidal currents affecting them. So, the fauna of these sand banks should remain impoverished.

THE SIGNIFICANCE OF THE HINDER, FLEMISH AND COASTAL SAND BANKS OFF BELGIUM

a. Introduction

Asymmetrical sand waves are present on these sand banks (HOUBOLT, 1968), so these banks must still be evolving at the present time. A remarkable feature of the sand banks is that they lie within, and are separated by, three deep invaginations of the Southern Bight Bed-Load Parting (figure 1). Moreover, there are no obvious supply routes along which sands could have passed towards these sand banks (but see also this section, c), as there is in the case of the sand banks that have been discussed already.

b. Future evolution implied by sand bank asymmetry

The asymmetrical cross-sections of these sand banks imply that they should be migrating laterally, as a result of removal of sand from their gentle slope and deposition of that sand on their relatively steeper lee slope. This type of migration can be shown by the internal structure of sand banks. In addition, the sand banks are assumed to migrate lengthwise at the same time, but little is known of the process. The effective time scale of both types of change is presumably measured in hundreds or even thousands of years, as no clear trends have been revealed by repeated surveys of these banks (VAN CAUWENBERGHE, 1971).

The asymmetrical cross-sections of the Hinder Banks imply that, in general, they should be moving east and north, away from the Southern Bight Bed-Load Parting (figure 1). It is thought that once they migrate outside the bed-load parting zone of finely-balanced tidal currents they will degenerate progressively. If the outermost two Zeeland Ridges can be used as a guide then most of the Hinder Banks may become progressively aligned with the peak tidal flow, such that sand waves on the two sides cease to migrate towards the crest but, instead, migrate along the length of the bank and then move away from it, travelling in a northeasterly direction. In contrast to the other members of the group the North Hinder Bank is more or less symmetrical in cross-section and so can be expected to remain where it is and even to grow in size. It would appear to be located at a local bed-load convergence within the overall Southern Bight Bed-Load Parting.

The asymmetrical cross-sections of the Flemish Banks imply that they should be migrating approximately west and south (figure 1). Indeed, there is a westerly increase in the height of successive banks in this group. The ground between successive members of these sand banks appears to be progressively swept clearer of sand in the same direction (HOUBOLT, 1968, enclosure 1). The implication is that upbuilding of these sand banks increases in importance westwards, in keeping with the increasing strength of the tidal currents in that direction. These sand banks are expected to maintain their identity as they move westwards away from the Southern Bight Bed-Load Parting. The bulk of their sand will presumably reach the Dover-Calais Bed-Load Convergence. In contrast to the other members of the group, the lower part of Kwinte Bank (figure 1) is known to show horizontal reflectors and so presumably represents a mass of uneroded older material, on top of which the sand bank-proper is situated. Thus, the western lee slope of the bank-proper, which is locally as steep as 7°, extends downwards as an erosion surface that cuts across the basal part of the bank (DE MOOR, this volume).

The cross-sectional asymmetry of the outermost of the coastal group of sand banks (figure 1), seawards of Belgium, implies that they should be heading east and shorewards. Successive echo-sounder surveys have revealed some morphological changes in these sand banks (VAN CAUWENBERGHE, 1971) but the expected long term trend is not yet evident.

The sand banks nearest to the coast of Belgium (not plotted on figure 1) have been shown by successive echo-sounder surveys to be migrating in an easterly direction (VAN CAUWENBERGHE, 1971).

The overall location of the Hinder, Flemish and coastal groups of sand banks, with respect to one another, may be taken to lend some support for the movement directions that are implied by their asymmetry in cross-section. Thus, the implied east and north movements of the Hinder Banks should have separated them from the Flemish Banks, which should be moving in westerly and southerly directions. Indeed, there is a useful shipping channel between the ends of these two groups of sand banks at the present time. This channel should widen in the distant future, if the above interpretation is correct. In contrast, there is only a small gap between the ends of the Flemish Banks and the outermost of the coastal sand banks, as would be expected from their growth directions partly towards one another. Moreover, the southern ends of the Flemish Banks are wide and Outer Ratel Bank (figure 1), for example, has a flat top (HOUBOLT, 1968, enclosure 2). It is not yet clear whether the northern ends of the outermost of the coastal sand banks are also wide or flat topped, as might be expected. It is conceivable, on present evidence, that the southern tips of some of the Flemish Banks may unite with the northern tips of some of the outermost of the coastal sand banks to create flat topped banks which remain in situ, as in the case of North Hinder Bank. On present evidence the majority of the Hinder Banks and Flemish Banks are therefore considered to be quasi-deposits.

The expected shoreward motion of the outermost of the coastal sand banks may perhaps increase the volume of the nearshore sand banks, or at least make good the losses of sand at any point, due to the seeming easterly migration of these sand banks, as shown by repeated echo-sounder surveys (VAN CAUWENBERGHE, 1971). Indeed, it seems conceivable that the seaward growth of dunes during the medieval and modern times between Dunkirk and Calais (BRIQUET, 1930, in SOMME, 1969) may perhaps be a direct result of the continuing supply of sand to the shore from seaward. Possibly, even the infilling of the wide embayment under the neighbouring coastal plain may be partly due to the same sand transport system towards the land, if that commenced early enough.

c. Supposed origin of the Hinder, Flemish and coastal sand banks

There are few obvious signs of sand transport paths that could have been the supply routes for sand moving towards the Hinder, Flemish and coastal series of sand banks that lie seaward of Belgium. This situation contrasts markedly with the known net transport paths associated with the Norfolk Sand Banks (outside the region of present consideration), the Thames Approaches Sand Banks, as well as with those in Dover-Calais Strait. Nevertheless, the sand banks off Belgium also represent sizeable bodies of sand, so a source of sand needs to be identified, if they are to be considered as modern features. One clue to the origin of this sand may be given by the internal structures of Kwinte Bank (DE MOOR, this volume) and by the seemingly analogous structure of at least the adjacent Outer Ratel Bank (Personal communication, G. DE MOOR, 1984) (figure 1). The lower parts of these two sand banks are separated from the upper parts by flat-lying reflectors. It is tentatively suggested, therefore, that the lower parts of these two sand banks are the remaining portions of a once more extension body of older Holocene sediments, for example. Furthermore, it is suggested that erosion of that older deposit, in the positions of what are now channels, may have provided some of the sand that has been built up as sand banks. There must have been some size-sorting going on at the same time to account for the different grain sizes of the sand banks in the region as a whole that have been shown by HOUBOLT (1968).

An additional mechanism of providing sand for the construction of the sand banks is suggested below. It involves sand brought into the region as a result of the operation of the occasional non-tidal currents. Such currents would act regionally and so might help to bring about the observed differences in grain sizes between sand banks, referred to above.

The Hinder, Flemish and coastal sand banks are seen as the natural mode of accumulation for sand, wherever it is trapped in regions of strong tidal currents. However, the apparent migration directions of these sand banks imply that in the long term they can mostly move away from their present locations (contrary to the inability of sand banks in the Thames Estuary to move far because of the limit set by the adjacent coast). Thus, it is considered likely that in the case of most of the Hinder, Flemish and possibly the coastal sand banks the trapping of sand is only temporary. The trapping mechanism is thought to result from a combination of the sinuosity of the Southern Bight Bed-Load Parting (which almost surrounds each group of banks), in conjunction with the conflict in sand transport directions caused by tidal and non-tidal water movements. Thus, the Southern Bight Bed-Load Parting is attributed to tidal currents : the usual small excess of 5 cm/s say, of flow one way or the other being sufficient to determine a net sand transport direction. In addition to the tidal currents there is the usual slow net mean drift of water northwards up the Southern Bight. Of more significance than the latter are the short-lived currents existing during storm surges. The numerically modelled peak depth mean values of these in the vicinity of the sand banks may range between 25 and 50 cm/s (FLATHER and DAVIES, 1978) in a severe individual storm and the estimated 50 year extreme depth mean value is as high as 60 cm/s in this region (Personal communication, R.A. FLATHER, 1982). Such strong currents will temporarily override the bed-load parting due to the tidal currents. For a few hours at a time the strong surge currents, almost equalling the tidal currents in strength plus storm wave enhancement of sand transport rates, will cause much sand to be moved to northeast or southwest approximately even across the bed-load parting. These occasions of sand transport by the combined currents will undo some of the work of the tidal currents on their own at some sites and increase their effect at other sites. Such sand may contribute to the size of the banks temporarily. Only gradually will the sand be moved away from the bed-load parting by the tidal currents. It would be of interest to explore this suggested mechanism, possibly by tracers and quantitative studies.

DEPOSITS OF THE EARLY HOLOCENE MARINE TRANSGRESSION

a. Introduction

The period being considered is from about 9,500 years BP to about 5,000 years BP, representing a sea level rise of about 55 m (JELGERSMA, 1979). Sand transport paths of these times are not known and so cannot be used to guide the search for the associated deposits. Nor have numerical simulations of tidal range and tidal currents been attempted, so there is no guidance from these as to the processes operating for successive sea levels.

For the period from about 9,500 years BP to just before about 8,300 years BP it has been assumed (JELGERSMA, 1979; JANSEN *et al.*, 1979) that the sea reached the Southern Bight only from the south, via a more restricted Dover-Calais Strait. As sea level rose there could then have been a slow northerly drift of sea water into the growing Southern Bight. This inflow could have carried mud in suspension, as well as facilitating the northwards spread of the marine fauna from the English Channel. If there were any tidal currents it is reasonable to expect them to have been stronger in Dover-Calais Strait than elsewhere.

At around 8,300 years BP, when the sea stood at about 30 m below present level, it united with the North Sea (JANSEN *et al.*, 1979; JELGERSMA, 1979). After that event the tides in the main North Sea would give another cause for tidal currents to be present in the still relatively narrow Southern Bight. On general considerations these tidal currents might have been stronger at that time than they were for the same ground by 5.000 years BP, when the sea was deeper. The narrow Southern Bight would, also, have begun to be exposed to the waves and non-tidal currents of the North Sea. It is therefore suggested that there was probably a substantial increase in tidal current activity as well as in other water movements after about 8,300 years BP. This change should be recognisable in a changed style of sedimentation, additional to any catastrophic effects at that time.

By 5,000 years BP the tidal currents and sand transport pattern may well have resembled rather closely those of the present time (when the sea level is only 5 m higher), so that similar types of offshore deposits could be expected to those described in section "Modern offshore marine deposits".

The most valuable source of information about the period of rising sea level, under consideration, comes from the chance finding of deposits.

b. Early Holocene diachronous deposits

A variety of deposits of early Holocene ages have been sampled.They can best be understood by reference to the Holocene sea level recovery curve (JELGERSMA, 1979). This is based on peats of known ages that represent much of the period of the early Holocene (and extend up to present day sea level). The deeper-lying samples were taken in the Southern Bight, while the younger ones were taken from below the adjacent land. The samples show conclusively that the peat and overlying brackish water deposits are diachronous, in keeping with the rising sea level. There may well have been diachronous nearshore and offshore deposits prior to 8,000 years BP, but none of these seem to have been recognised.

For the period of the marine transgression after about 8,000 years BP, nearshore sands have been recognised for a range of depths, where they are preserved beneath the Zeeland Ridges. These sands contain the *Spisula subtruncata* mollusc association, which is analogous to the nearshore fauna occuring near Holland at the present time (LABAN and SCHÜTTENHELM, 1981). They can be considered as diachronous, as in the case of the underlying brackish water deposits and peats. They are, in turn, overlain by sands with the *Angulus pygmaeus* mollusc association of deeper water origin, that form much of the Zeeland Ridges, and are found in the large sand waves located further west (LABAN and SCHÜTTENHELM, 1981). The base of these younger sands lie at depths below the present sea level of about 45 m in the west and 30 m in the east, so they are presumably diachronous also. The same authors state that the two sands are separated by a strong "horizontal" reflector as seen on continuous reflection records. This reflector occurs within the Zeeland Ridges, and is located just below, to several metres below the troughs of the large sand waves occuring further west.

Parts of this diachronic succession are considered in more detail below. First, the basal peats are overlain locally by a thin clay layer, but more generally off Holland they are overlain by fine sand with thin clay laminae, the Elbow and Calais Deposits of OELE (1971). In addition, *Cardium edule* shells have been found widely in the Southern Bight. They cover an age range from 9,560 years BP to 5,910 years BP (EISMA *et al.*, 1981), and are clearly diachronous. The morphology of these shells, as well as the salinity implied by their composition, show that for the most part they originated under brackish water conditions. The sea level recovery curve based on these shells follows the recovery curve based on the peats, but is stated as generally being some 8 m deeper. Those authors concluded, therefore, that the shells had been washed out of the tidal flats where they had grown and had then accumulated nearby in the bottom of tidal channels some 8 m deeper. Actually, the sea level recovery curve based on these shells (EISMA *et al.*, 1981, figure 5) can be seen to begin its divergence from the accepted recovery curve based on peats at a depth of about 40 m below present sea level. It is concluded, therefore, that at this time of about 8,700 years BP any supposed tidal channels were too shallow to be detected and that deeper ones only developed later on during the marine transgression. Indeed, this rather indirect evidence might be used to suggest very tentatively that tidal range only began to reach significant values just before the Southern Bight was united with the North Sea. This event seems to have occured at about 8,300 years BP, for a depth below present sea level of about 30 m (JANSEN *et al.*, 1979). Maybe the union took place somewhat earlier than this time, if due allowance has not been made for the full thickness of later Holocene deposits in this region. The minimal apparent tidal range in the ancestral Southern Bights up to about 8,700 years BP might be taken to imply the presence of only weak tidal currents offshore except perhaps in the bottleneck of the early Dover-Calais Strait. Thus, sand banks and sand waves may not have been formed in some of the ancestral Southern Bights (contrary to what has been assumed by some workers).

The strong horizontal reflector, that is known to be widespread in the Dutch sector of the Southern Bight up to 52* N at least, is considered to indicate the changed current conditions that resulted from the union of the Southern Bight with the North Sea. On published sections (LABAN and SCHÜTTENHELM, 1981) it can be seen that this reflector reaches down even into the Pleistocene sediments, so that at some locations there has been a considerable amount of erosion. Indeed, the diachronous series of *Cardium edule* shells that are so widespread in the Southern Bight (EISMA *et al.*, 1981) may result from erosion that occured during the period represented by the reflector (rather than at the times when the old tidal flats and supposed tidal channels were evolving, as mentioned above). Thus, it is argued now that these shells occur at a deeper level than would be expected by their ages because up to about 8 m, say, of sediment below them has been removed during the period of late Holocene erosion. If this interpretation is correct it also accounts for the seeming lack of any nearshore and offshore sediments prior to 8,000 years BP, as well as the seeming lack of any later offshore sediments equivalent in age to the nearshore sands with the *Spisula subtruncata* mollusc association. Indeed, it was presumably the removal of much of the older Holocene deposits, as well as some underlying material that has supplied the sediment that makes up the later Holocene deposits. The process of removal and sorting out of this earlier material is still continuing (as discussed in section 'Modern offshore marine deposits, a').
Clearly, a choice needs to be made between the two proposed explanations for the depth of occurence of the shells.

It might be supposed that any sand banks of the early Holocene in the Southern Bight would be relatively easy to locate because of their internal structure or because they are likely to be localised bedforms with appreciable height. D'OLIER (1972) has suggested that the cross-bedded sediments, that partly fill a drowned channel beneath the Thames Approaches, may indicate the presence of a sand bank. He now calls for samples and for dates.

Certainly, low sea level, presumed Holocene, sand banks had been recognised by their relief in the Celtic Sea (STRIDE *et al.*, 1982) and these have subsequently been sampled and dated as late Devensian to early Holocene (PANTIN and EVANS, 1984). Old sand banks have been identified tentatively for the southern North Sea outside the Southern Bight. Although their size and shape, the absence of sand waves on them and their location in regions of weak tidal currents are in keeping with such an interpretation they do require sampling and dating if their nature is to be established adequately. It has also been suggested that there are some early Holocene sand banks in the Southern Bight itself. The argument used is that the depth at which the base of a sand bank is located gives an indication of its maximum possible age : some of these sand banks might, therefore, have formed as early as about 8,300 years BP (JELGERSMA, 1979). This interesting suggestion requires support from dated material if it is to be established, as these deeper-lying sand banks of the Southern Bight all carry sand waves and so can be accounted for by means of modern processes (discussed in section 'Modern offshore marine deposits'). At the present time the tidal currents of the Southern Bight, especially south of the Southern Bight Bed-Load Parting, are strong enough to move much sand each year, so that any old sand banks that had been exposed there would soon have ceased to show their original form, as they would be adapting to the present tidal current regime. At some sites in the Southern Bight there may even be scope for converting a sand sheet facies of a lower sea level tidal current origin into modern sand bank facies, or elsewhere converting old sand bank facies into modern sand sheet facies simply because the net sand transport directions and the geographical distribution of tidal current strength have changed (see also section 'Modern offshore marine deposits').

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15 A preliminary correlation of the onshore and offshore courses of the Rivers Thames and Medway during the Middle and Upper Pleistocene

D.R. BRIDGLAND (1) and B. D'OLIER (2)

1. Nature Conservancy Council, Northminster House, Peterborough, Cambridgeshire, United Kingdom.
2. Department of Geology, City of London Polytechnic, Walburgh House, Bigland Street, London E1 2NG, United Kingdom.

INTRODUCTION

Study of Pleistocene deposits in the coastal fringe of Essex has a long history, although the area has attracted little attention compared with the adjacent Thames Valley and East Anglian regions. The area comprises low-lying land, predominantly formed of London Clay overlain by a succession of gravel terraces in various stages of dissection, descending eastwards and culminating in coastal marshes and mudflats (figure 1).
Offshore, the submarine environment is dominated by large sandbars (some parts of which dry at low spring tides) separated by tidal channels, principally floored by coarse sand and gravel lags.

The earliest description of the gravels in this area was by WOOD (1866), who regarded them as marine. WOOD's term "East Essex Gravel" has been incorporated into modern lithostratigraphic terminology (below). WHITAKER (1889) interpreted these deposits as a continuation of the gravels of the Thames Valley and suggested that the Thames and Medway flowed together across the area. Later GREGORY (1922), impressed by the presence of large quantities of Lower Greensand chert in these gravels, believed they were deposited, in late Tertiary times, by rivers flowing down an extended northern slope of the Weald. More recently, GRUHN *et al.* (1974) confirmed the strong southern affinities of the gravel and suggested that the Medway was the main agent in their deposition. These authors were the first to recognise terraces in the area between the Thames and Blackwater. This area has recently been resurveyed by the Geological Survey, as a result of which four numbered terraces and several drift-filled channels have been recognised (Geological Survey New Series sheets 241 & 258/9 ; LAKE *et al.*, 1977 ; HOLLYER and SIMMONS, 1978 ; SIMMONS, 1978 ; BRISTOW, 1985 ; LAKE *et al.*, 1986).

Drowned river channels, incised mainly into the London Clay bedrock, have been traced offshore, although their recognition is complicated by the effects of marine planation (SPURRELL, 1892 ; D'OLIER, 1975). Using reflection seismic profiling techniques, such features have been recognised even when filled with fluvial or later marine sediments. In addition, up to four aggradational terraces have been recognised, running approximately parallel to the main buried channel which underlies the central part of the Thames Estuary (figure 4).

PREVIOUS RESEARCH IN THE CLACTON AREA

The Clacton area has received considerably more attention from geologists than other parts of eastern Essex, largely because of the occurrence there of important fossiliferous channel deposits, first discovered by JOHN BROWN OF STANWAY in the late 1830's. The area was first mapped by the Geological Survey during the late 19th century, when the various drift deposits were divided into gravel and loam of 'Glacial' and 'Post-Glacial' ages, the former largely covering the Tendring Plateau (the peninsula between the Colne

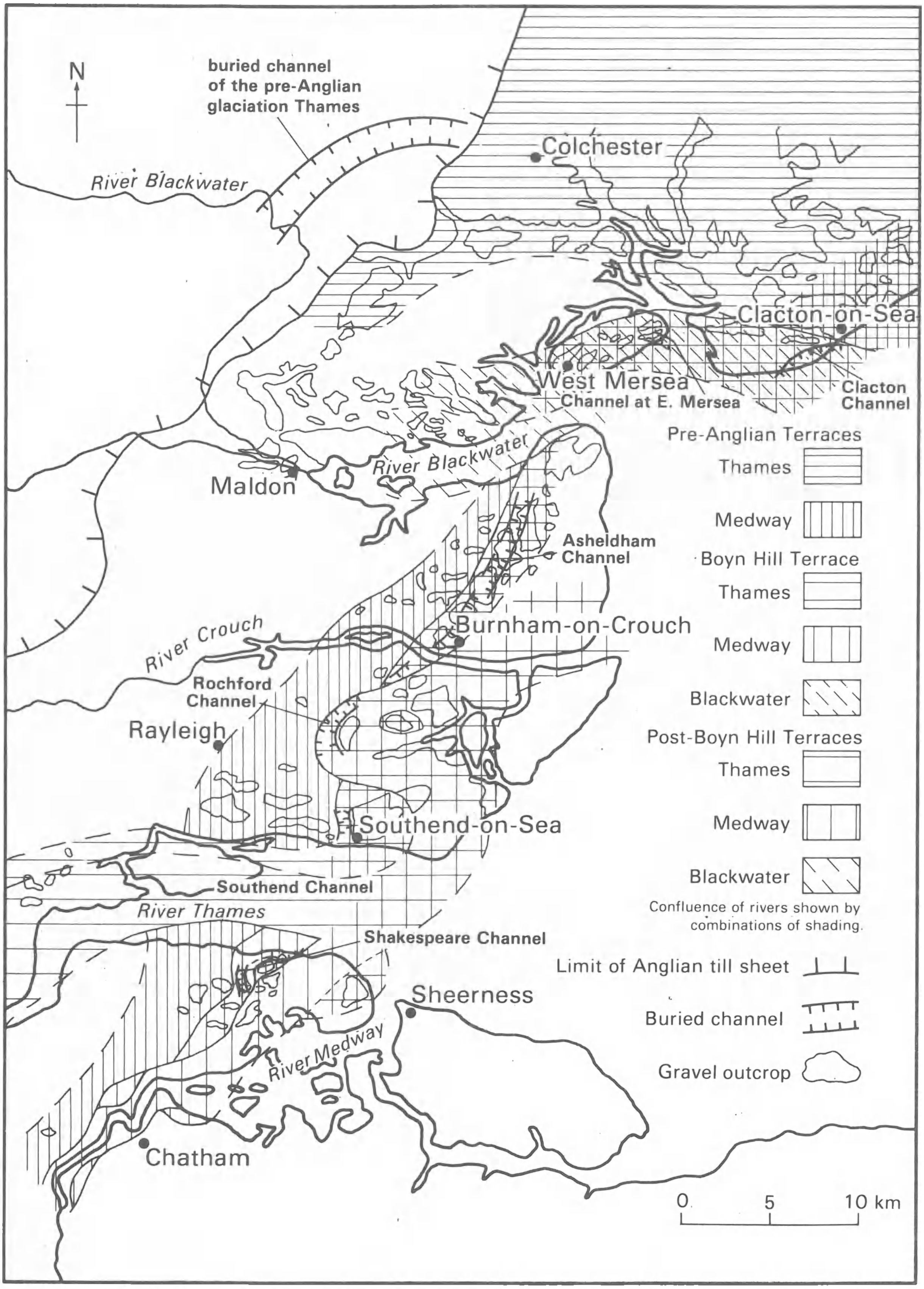


Figure 1 Map showing the distribution of gravel terraces in eastern Essex.

and Stour estuaries) while patches of 'Post Glacial' drift were recognised fringing the valleys of the Colne and Stour and in the extreme southeast of the area, between St. Osyth and Clacton (Geological Survey Old Series sheet 48 ; DALTON, 1880). DALTON later reported (1902 ; 1904 ; 1908) that deposits at Walton-on-the-Naze, Harwich, Mersea Island and Tollesbury, formerly regarded and mapped as 'Glacial Gravel', might in fact be of 'Post-Glacial' age, and suggested that they represented an extended Thames, "when its mouth was in a common estuary with all other East British and many West European rivers" (DALTON, 1904). This view was reinforced by OAKLEY and LEAKEY (1937).

The Clacton Channel deposits have been described by numerous authors since their discovery, notably WOOD (1866), FISHER (1868), DALTON (1880), WARREN (e.g. 1923 ; 1924 ; 1955), OAKLEY and LEAKEY (1937) and SINGER *et al.* (1973). WARREN also described other Pleistocene deposits in the area. Originally, he suggested (1923 ; 1924 ; 1933) that the widespread gravels of the Tendring Plateau could be attributed to the Boyn Hill Terrace of the Thames, but the discovery of Cromerian deposits channelled into the Plateau Gravels at Little Oakley led him later to conclude (WARREN, 1940) that the deposits were very much older than any of the Lower Thames terraces. This interglacial site has recently been studied in detail and its Cromerian age confirmed (BRIDGLAND, GIBBARD and PREECE, in preparation). WARREN (1955) reinterpreted the gravel outcropping in the cliffs north of Clacton, his 'Holland Gravel', as a "pre-Northern Drift" (pre-Anglian glaciation) Thames deposit, which he believed could be traced from the present Middle Thames Valley across central Essex to Holland-on-Sea. GLADFELTER (SINGER *et al.*, 1973) suggested that the Clacton Channel gravel was reworked from local sources and therefore post-dated the diversion of the Thames from its former route across central Essex. However, he later interpreted it as a Thames-Mole deposit pre-dating this diversion (GLADFELTER, 1975). ROSE *et al.* (1976) included the gravels of the Tendring Plateau in their Kesgrave Formation, which they interpreted as a Beestonian periglacial river deposit laid down by the early Thames.

During the period 1977-1980 an attempted reappraisal of the gravels throughout eastern Essex was carried out by one of the authors (D. BRIDGLAND) based on morphostratigraphic and lithostratigraphic techniques.

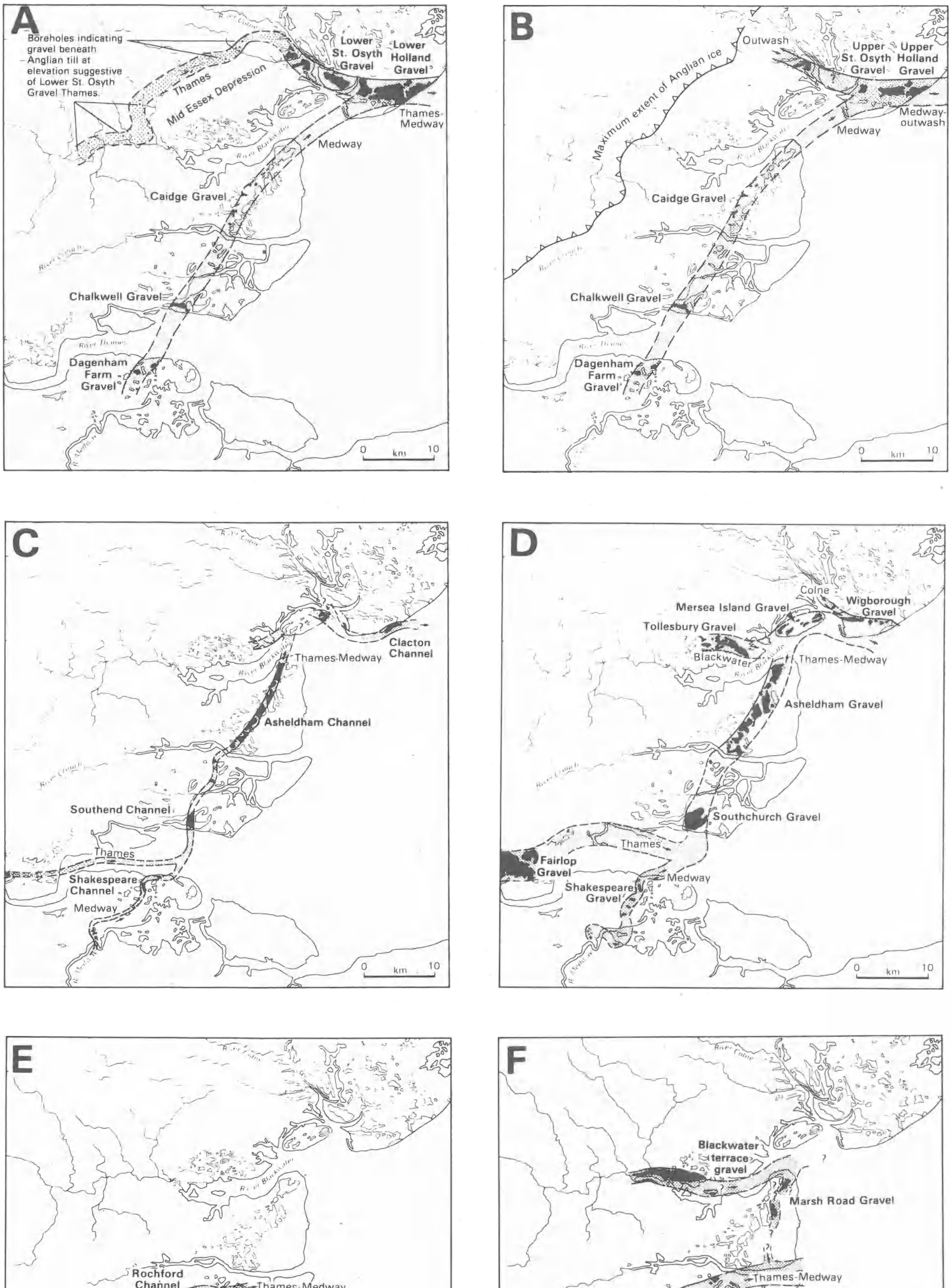
THE SUCCESSION IN EASTERN ESSEX (SOUTH OF THE BLACKWATER)

Samples of the various gravels in this area were subjected to clast-lithological analysis following methodology outlined in BRIDGLAND (1986a). This enabled two distinct lithostratigraphical divisions to be recognised, the 'High-level' and 'Low-level East Essex Gravel Formations'. In addition the various gravel bodies studied were related to ordnance datum by levelling, enabling the subdivision of the deposits into a number of distinct aggradational units. Following the example of GIBBARD (1985) with the equivalent sequence in the Middle Thames, these are given member status (table 1). It is believed that the channels recognised by the Geological Survey can be related to this sequence of gravel aggradations (BRIDGLAND, 1983a ; 1983b) (figures 2 and 3).

THE HIGH-LEVEL EAST ESSEX GRAVEL FORMATION

This formation comprises a number of individual aggradations, each characterized by a composition almost entirely limited to local (35-86 %) and southern (13-65 %) rock types, the local material comprising flint from the Chalk and flint pebbles from the Palaeogene of the London Basin, while the southern component is made up largely of chert from the Folkestone and Hythe Beds (Lower Greensand), supplemented by sandstones, siltstones and ironstones from the Hastings Beds of the central Weald and rare arenaceous lithologies from the Lower Greensand (BRIDGLAND, 1980 ; 1983a ; 1986b). Occasional 'exotic' material (derived from outside the London Basin and the Wealden area) was also encountered, mainly cherts of Jurassic or Palaeozoic origin and quartzose lithologies. These are thought to have been derived from pebble beds within the Cretaceous of the Wealden area.

These deposits are, on the basis of their gravel content, attributed to the River Medway. They have a composition virtually indistinguishable from Medway gravels studied in North Kent (BRIDGLAND, 1980 ;


A
Boreholes indicating gravel beneath Anglian till at elevation suggestive of Lower St. Osyth Gravel Thames.
Thames
Mid Essex Depression
Lower St. Osyth Gravel
Lower Holland Gravel
Thames-Medway
Medway
Caidge Gravel
Chalkwell Gravel
Dagenham Farm Gravel
0 km 10
B
Maximum extent of Anglian ice
Outwash
Upper St. Osyth Gravel
Upper Holland Gravel
Medway-outwash
Medway
Caidge Gravel
Chalkwell Gravel
Dagenham Farm Gravel
0 km 10
C
Clacton Channel
Thames-Medway
Asheldham Channel
Southend Channel
Thames
Shakespeare Channel
Medway
0 km 10
D
Colne
Mersea Island Gravel
Wigborough Gravel
Tollesbury Gravel
Blackwater
Thames-Medway
Asheldham Gravel
Southchurch Gravel
Thames
Fairlop Gravel
Medway
Shakespeare Gravel
0 km 10
E
Rochford Channel
Thames-Medway
Thames
Medway
0 km 10
F
Blackwater (terrace) gravel
Marsh Road Gravel
Thames-Medway
Rochford Gravel
Corbets Tey Gravel
Thames
River Thames
Newhall Gravel
Medway
0 km 10

1983a ; BRIDGLAND and HARDING, 1985), although they generally contain slightly less southern material, particularly the predominantly soft Hastings Beds lithologies. This probably reflects the dilution of the southern component by the addition of local material (especially flint pebbles from the Palaeogene), picked up directly or from tributaries, in the area of the present Thames Estuary.

The High-level East Essex Gravel aggradations are believed to have been laid down by the Medway at a time when the Thames flowed to the north of the present Blackwater, as occurred prior to the Anglian glaciation (ROSE, ALLEN and HEY, 1976 ; GIBBARD, 1977 ; BRIDGLAND, 1980 ; 1983a ; 1983b). Six aggradational units have been recognised within the High-level East Essex Gravel Formation, while degraded remnants of others also exist (BRIDGLAND, 1983a) (figures 2 and 3). These have been defined and named according to the recommendations of HEDBERG (1976) ; detailed descriptions and formal definitions will be published elsewhere.

THE LOW-LEVEL EAST ESSEX GRAVEL FORMATION

The composition of gravels belonging to this formation differs from the early Medway aggradations in a number of ways. They contain a rather higher proportion of local material (74-94 %) and a significantly smaller proportion of southern rock-types (5-25 %). The most marked change, however, is the presence in these lower deposits of a small but consistent exotic component (0.5-3 %), comprising vein quartz, ortho- and metaquartzite, Carboniferous chert, *Rhaxella* chert and igneous rock-types (BRIDGLAND, 1983a ; 1986b). This exotic suite is identical to that recognised in Lower Thames gravels in the Tilbury area (BRIDGLAND, 1980 ; 1983a).

The presence of this characteristic exotic component, and the continued occurrence of fairly abundant southern material (including occasional soft Hastings Beds lithologies), suggests that the Thames and Medway have both contributed to the Low-level East Essex Gravel aggradations, which are therefore interpreted as Thames-Medway deposits (i.e. laid down by the Thames downstream from its confluence with the Medway), as first envisaged by WHITAKER. Three individual aggradations have been recognised within this formation, as well as a number of associated drift-filled channels (including those described by LAKE *et al.*, 1977). As with the High-level East Essex Gravel deposits, different lithostratigraphic nomenclature is applied on either side of the Crouch estuary (table 1).

The highest of these Thames-Medway aggradations, represented by the Southchurch and Asheldham gravels, remains relatively intact north of the Crouch, where it appears to represent a complete floodplain and buried valley rather than the expected west bank terrace remnant only. This, and the poor development of later deposits in this area, suggests that the river's course north of the Crouch significantly changed following deposition of the Asheldham Gravel. The river probably then turned eastwards, along the approximate line of the modern Crouch estuary. This is of potential importance (a) for correlation with the offshore terrace sequence and (b) for determining the origins of this enigmatic estuary, which appears over-large for the insignificant River Crouch (GREGORY, 1922 ; GREENSMITH and TUCKER, 1971).

Figure 2 Series of palaeogeography maps showing drainage evolution in eastern Essex.
A = Immediately prior to the arrival of Anglian ice in central Essex.
B = Anglian glacial maximum (Lowestoft Stadial), with the Thames valley blocked by ice.
C = Immediately following the diversion of the Thames into the Medway (Late Anglian/Hoxnian); the excavation and infilling of the Clacton Channel system.
D = The early Wolstonian terrace aggradation which overlies the channel system shown in C (equates with the Boyn Hill Gravel of the Middle Thames).
E = Mid-Wolstonian course of the Thames-Medway during downcutting of the Rochford Channel; the valley north of the Crouch is now abandoned.
F = The mid-Wolstonian terrace aggradation which overlies the channel shown in E; later courses follow the line of the modern estuary more closely.

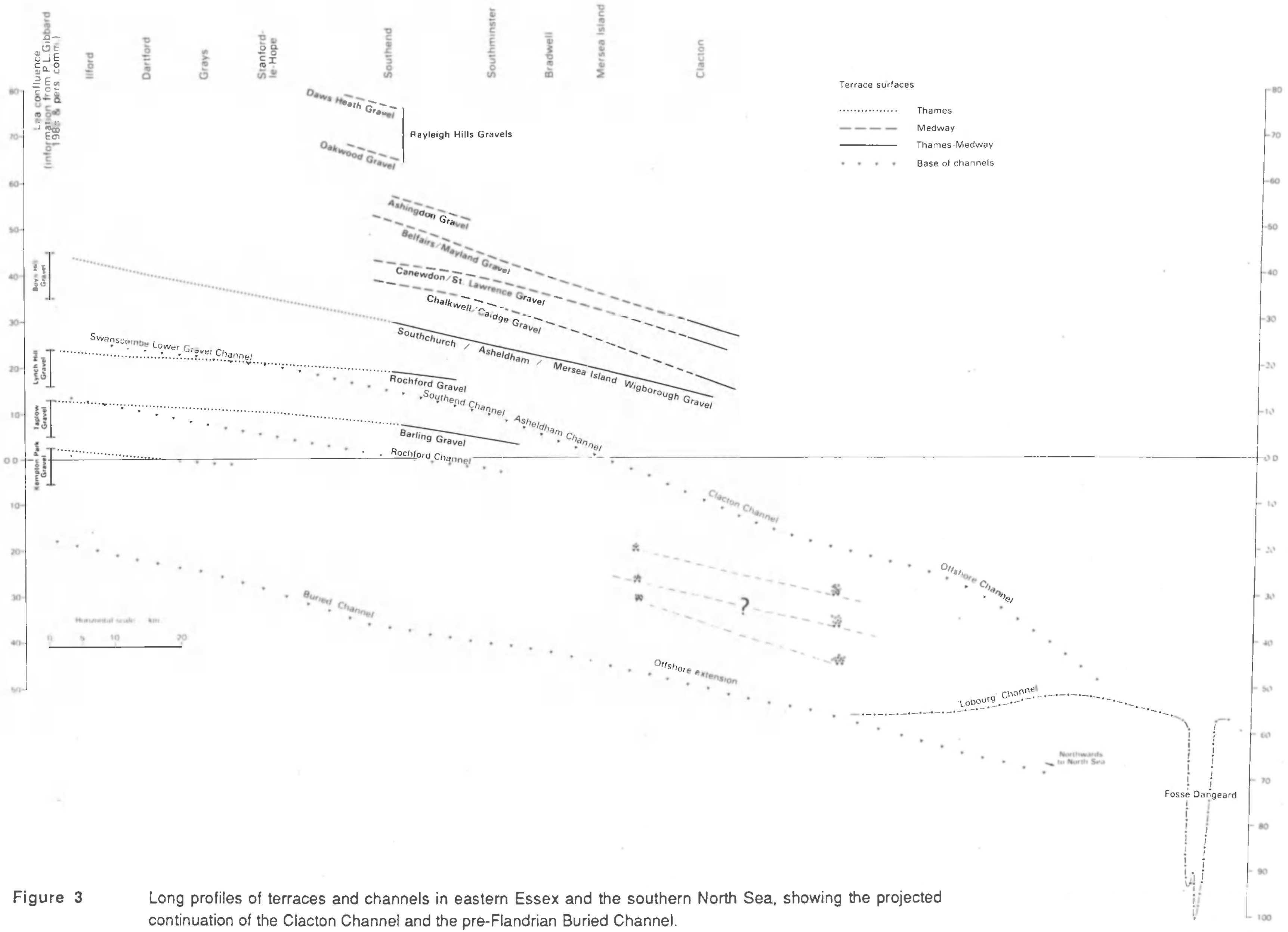


Figure 3 Long profiles of terraces and channels in eastern Essex and the southern North Sea, showing the projected continuation of the Clacton Channel and the pre-Flandrian Buried Channel.

High-level East Essex Gravel Formation			
Southend area	**Dengie Peninsula**	**Clacton area**	**Proj. elevation** *
Claydons Grav.			> 75 m
Daws Heath Gr.			c 70 m
Oakwood Grav.			c 60 m
Ashingdon Gr.			52 m
Belfairs Grav.	Mayland Gravel	K	45 m
Canewdon Gr.	St. Lawrence Gr.	K	38 m
Chalkwell Gr.	Caidge Gravel	K	33 m
Low-level East Essex Gravel Formation			
Southend area	**Dengie Peninsula**	**Clacton area**	**Proj. elevation** *
Southchurch Gr.	Asheldham Gr.	Wigborough Gr.	25 m
(Southend Ch.)	(Asheld. Ch.)	(Clacton Ch.)	(11 m)
Rochford Grav.			13 m
(Rochford Ch.)			(- 3 m)
Barling Gravel			5 m
(Shoeburyness Channel)			(- 15 m)
Submerged terraces yet to be differentiated (Thames-Medway buried channel)			(- 30 m)

* Projected elevation of gravel surface (channel floor) in position of present Crouch estuary, in metres above OD.

K Equivalent Kesgrave Formation gravels recognised (Bridgland, Gibbard & Preece, in preparation).

Table 1 Gravel lithostratigraphic units (and associated channels) in eastern Essex. Data for channels in parentheses.

The appearance of the River Thames (in Thames-Medway form) in eastern Essex, as indicated by the marked change in gravel composition between the High- and Low-level East Essex Gravel Formations, is believed to coincide with the diversion of the river by Anglian ice from its old northern course (through the Vale of St. Albans to northern Essex) into its modern valley through London (GIBBARD, 1977 ; 1979). Following this diversion the Thames appears to have adopted the former course of the River Medway across eastern Essex, implying that it was actually diverted into the Medway, possibly by way of an existing tributary of that river in the position of the modern Lower Thames (BRIDGLAND, 1980 ; 1983a).

THE OFFSHORE EVIDENCE FOR LATER FLUVIAL EVOLUTION

The submerged and buried channel of the Thames-Medway, with up to four associated submerged terraces, can be traced offshore, turning northwards as it passes out under the modern estuary (figure 4). Some 25 km east of Foulness Point it is joined by the buried channel of the Crouch. This Crouch channel is only marginally smaller, in terms of breadth and depth of incision, than the Thames-Medway channel it joins. The area of their confluence, where marked overdeepening has occurred, is of critical importance for the interpretation of these low-level terrace and channel sequences. The Thames-Medway turns eastwards here and continues to a position around 1° 40′ E and 51° 35′ N, where it meets another large channel trending NNE-SSW. This channel is largely open, perhaps because most of the sediments which once filled it have been removed by marine erosion. It has been variously interpreted by past workers as a Palaeo-Rhine channel (OELE and SCHÜTTENHELM, 1979), the result of tidal scour (ZAGWIJN, 1979), an ice-scour channel (DESTOMBES *et al.,* 1975) or a glacial lake overflow channel (SMITH, 1984). The southwestern end of this feature occupies the Straits of Dover (figure 4).

THE SUCCESSION IN THE CLACTON AREA

The sequence recognised south of the Blackwater estuary can be directly related to the later elements (i.e. at lower elevations) of a flight of gravel terraces covering the peninsula between the Colne and Stour estuaries.

A detailed study of gravel composition in the Clacton area has broadly confirmed ROSE *et al*'s. interpretation of these as Kesgrave Formation Thames deposits (characterized by a relatively high exotic content, at around 20 %). However, it has been found that these early Thames gravels pass eastwards into equivalent Thames-Medway deposits, the change determined by a significant increase in southern material at the expense of exotics (BRIDGLAND, 1983a). The site of the contemporary confluence has been located with some precision in the lowest of several 'Kesgrave' terraces recognised in this area, at St. Osyth. Here the (Thames) St. Oysth Gravel changes eastwards over a few hundred yards into the (Thames-Medway) Holland Gravel.

A further Thames aggradation lower (and therefore younger) than any of the Kesgrave Formation deposits has also been recognised on the Tendring Plateau. This aggradation, the Wigborough Gravel, is limited to the extreme southeastern fringe of the peninsula, from which it can be traced upstream to Mersea Island (figure 2) (BRIDGLAND, 1983a). Plotting of long profiles indicates that the Wigborough Gravel is a continuation of the Asheldham Gravel from south of the Blackwater (figure 3), indicating that it post-dates the diversion of the Thames into its modern valley through London. The well-known Clacton Channel deposits are thought to represent the basal part of the aggradation which culminated in the deposition of the Wigborough Gravel (BRIDGLAND, 1983a). This aggradation started in the late Anglian (basal Clacton Channel Gravel) and culminated in the deposition of the Wigborough Gravel in the early Wolstonian, the Hoxnian being represented by the extensive interglacial sequence in the Clacton Channel (1953). These deposits are considered to be downstream equivalents of those filling the Southend and Asheldham Channels, south of the Blackwater (see above).

CORRELATION OF THE VARIOUS AGGRADATIONS IN THE EASTERN PART OF THE LONDON BASIN : THE SOUTHEND/BURNHAM/CLACTON CHANNEL 'MARKER LEVEL'

Subdivision into individual gravel aggradations within each formation, in each geographical area, is based, in the almost total absence of other evidence, on elevation. No landform-morphological studies have been attempted, however ; individual gravel aggradations are recognised and correlated according to the elevation of the in situ sediment bodies. Correlation between separate areas such as the fairly small peninsulas between the various estuaries in eastern Essex would, if based on this method alone, be fraught with difficulty, since the estuaries themselves represent considerable gaps in which no evidence is preserved. Fortunately, the events of the Anglian glaciation, when the Thames was diverted, provided an invaluable 'Marker Level' within the gravel succession in eastern Essex, in the form of the change from High-level to Low-level East Essex Gravel deposition. This "Marker Level" is represented on the gravel map by the Southchurch/Asheldham/Mersea Island/Wigborough Gravel aggradation. This represents the earliest mappable appearance of gravel with the characteristic Low-level East Essex Gravel Formation (post-diversion Thames-Medway) composition in eastern Essex. Correlation with the Thames valley (BRIDGLAND, 1983a) suggests that this is a downstream continuation of the Boyn Hill Gravel. However, beneath this aggradation in eastern Essex, covering a more limited area, is a continuous buried channel system, the Southend/Asheldham Channel. Tracing of the Thames-diversion 'Marker level' north of the

Blackwater enables the correlation of this channel with the famous Clacton Channel, which contains deposits yielding interglacial fauna and flora and important Palaeolithic artefacts (PIKE and GODWIN, 1953 ; WARREN, 1955). This Southend/ Asheldham/Clacton channel has also been recognised at East Mersea, on Mersea Island and has been traced upstream in the Medway valley in the form of the Shakespeare Channel (BRIDGLAND, 1983a). In the Lower Thames Valley its upstream equivalent may be represented by the channel containing the Lower Gravel and Lower Loam at Swanscombe (BRIDGLAND, 1980 ; 1983a), which also contains Clactonian artefacts and has yielded Hoxnian fauna (WYMER, 1968 ; KERNEY, 1971). In the present paper the authors attempt to demonstrate that this important feature has now been recognised in the offshore sequence.

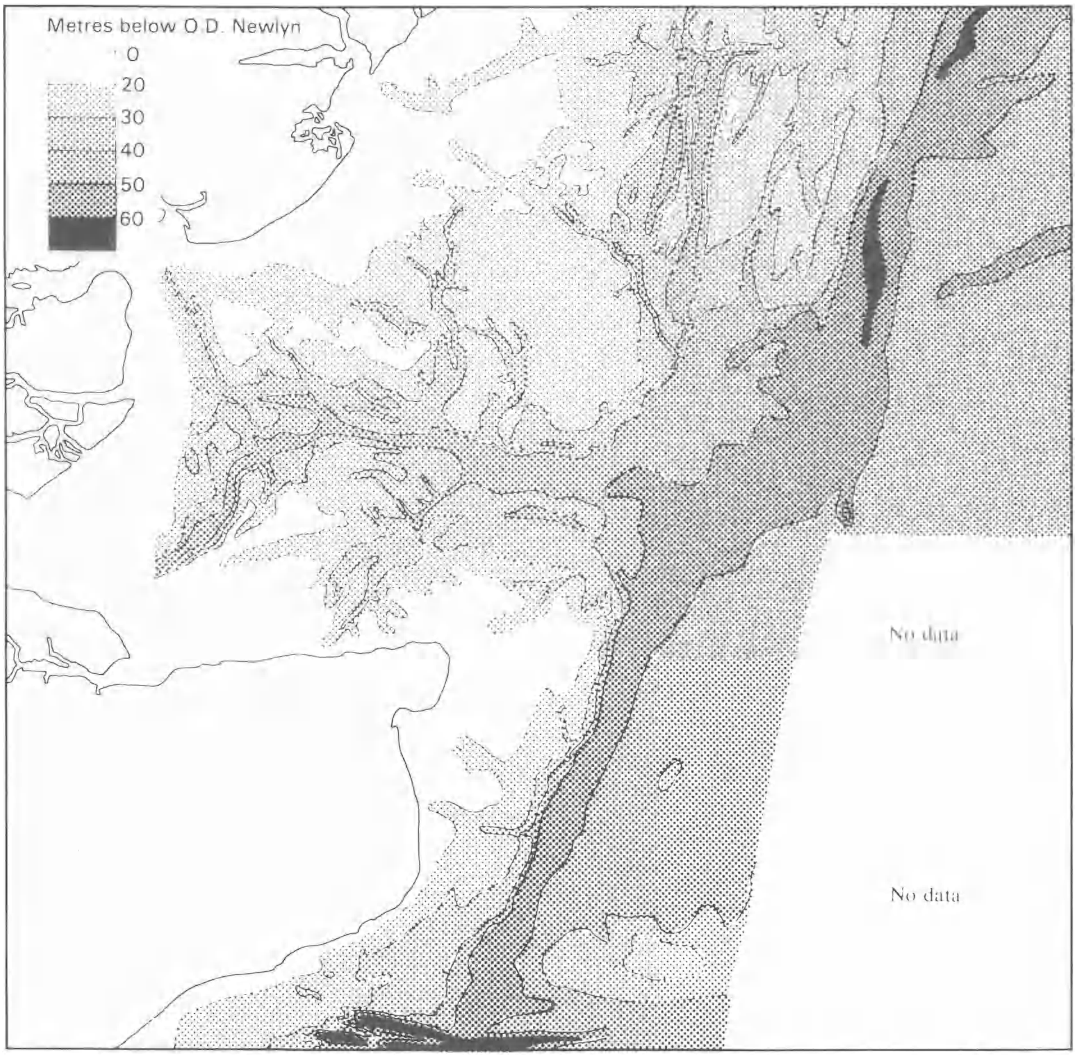


Figure 4 Bed-rock surface contour map of the area offshore from eastern Essex, showing the buried and submerged valley system revealed by reflection seismic profiling.

THE OFFSHORE EVIDENCE FOR A CONTINUATION OF THE CLACTON CHANNEL

Immediately offshore from the Clacton Channel coastal site, no buried valley can be observed, as the London Clay in this area has been subjected to recent marine planation. The sea-floor is therefore below the projected level of the Clacton Channel (figure 3) for an offshore distance of several kilometers. Some 5 km ENE of Clacton, however, a channel has been recognised, incised 2-5 m below the main planation surface and declining eastwards (figure 5). Plotting of the base of this buried channel (as an extension of the onshore terrace/channel long profiles diagram) indicates that its elevation is consistent with interpretation of the feature as a continuation of the Clacton Channel. The channel continues in an ENE direction until, at 1° 30′ E, extensive dissection of the southern bank by later small channels running southwards to the Crouch/Thames-Medway buried channels makes its course difficult to follow. However, it can be picked up again at 1° 35′ E, 51° 52′ N, from where it runs north-eastwards, before turning to the east at 1° 54′ E, 51° 57′ N. It finally enters the large NNE-SSW channel described above at 2° 7′ E, 51° 57′ N.

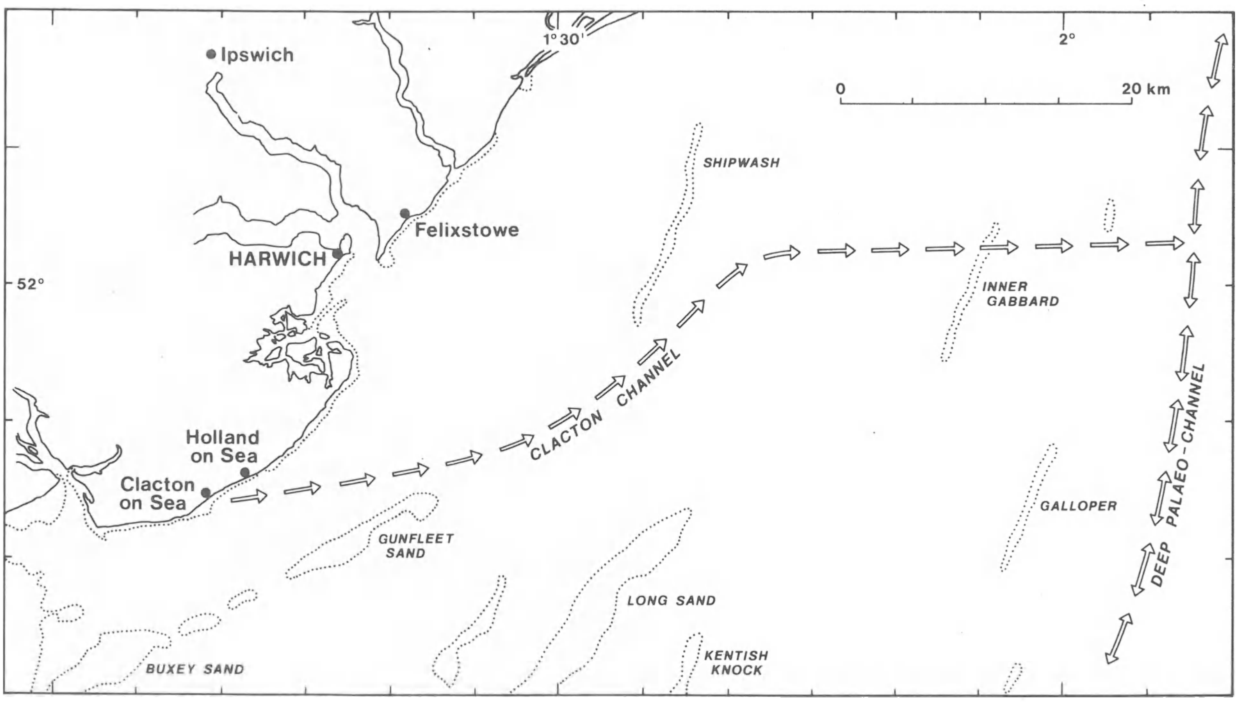


Figure 5 Map showing the trend of the offshore extension of the Clacton Channel described in the text.

DISCUSSION

The excavation of the Southend/Asheldham/Clacton Channel by the Thames-Medway is considered to have closely followed the diversion of the Thames, by Anglian ice, southwards into its modern valley. This channel can be traced from Swanscombe, where its base is at c.23 m OD, through Southend, Southminster, west of Bradwell and through East Mersea to Clacton (figures 1 and 2), where its base has already fallen to (at least) -6 m OD (figure 3). The offshore continuation of this channel, now recognised, has a base level of -15 m 5 km offshore from Clacton, falling to -40 m OD 65 km further east, where it joins the large palaeo-channel previously described (figure 5). The downstream profile of this extended Clacton Channel appears to steepen as it approaches the central North Sea (figure 3), a possible indication of post-Hoxnian subsidence of this sedimentary basin, as suggested by several previous workers (ZEUNER, 1945 ; WOOLDRIDGE and HENDERSON, 1955 ; GRUHN *et al.*, 1974 ; GREENSMITH and TUCKER, 1980). Subsidence has not, however, been sufficiently active to prevent the tracing of the various terraces and channels between the Lower Thames valley and the eastern North Sea.

After the infilling of this channel system and deposition of the overlying terrace deposits (Southchurch/Asheldham/Mersea Island/Wigborough Gravel) the river shifted southwards, initiating the course broadly followed by all later channels and terraces, culminating in the submerged Thames-Medway buried channel (figures 2 and 4). In the earliest manifestation of this new route, the Rochford Channel, the river followed a highly sinuous course (figure 1). It was possibly in the northern arm of the large Rochford Channel meander that the River Crouch came into being, perhaps as a result of this loop being cut off at a later date. Work is continuing with the aim of correlating the later terraces and channels of the onshore area with their corresponding features offshore.

ACKNOWLEDGEMENTS

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

P. KIDEN

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

ABSTRACT

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

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

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

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

INTRODUCTION

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

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

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

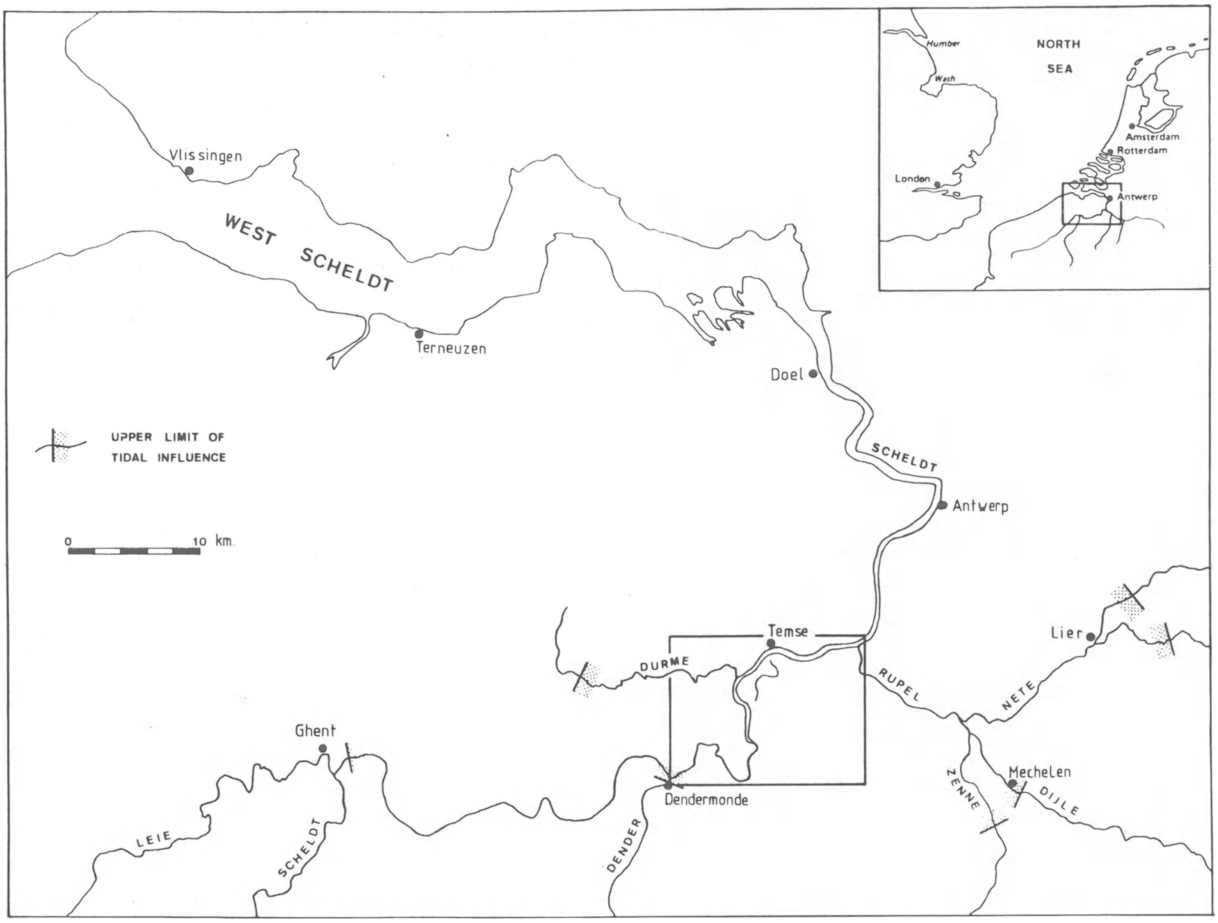

Figure 1 Location map of the study area.

OVERVIEW OF THE HOLOCENE EVOLUTION OF THE RIVER SCHELDT

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

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

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

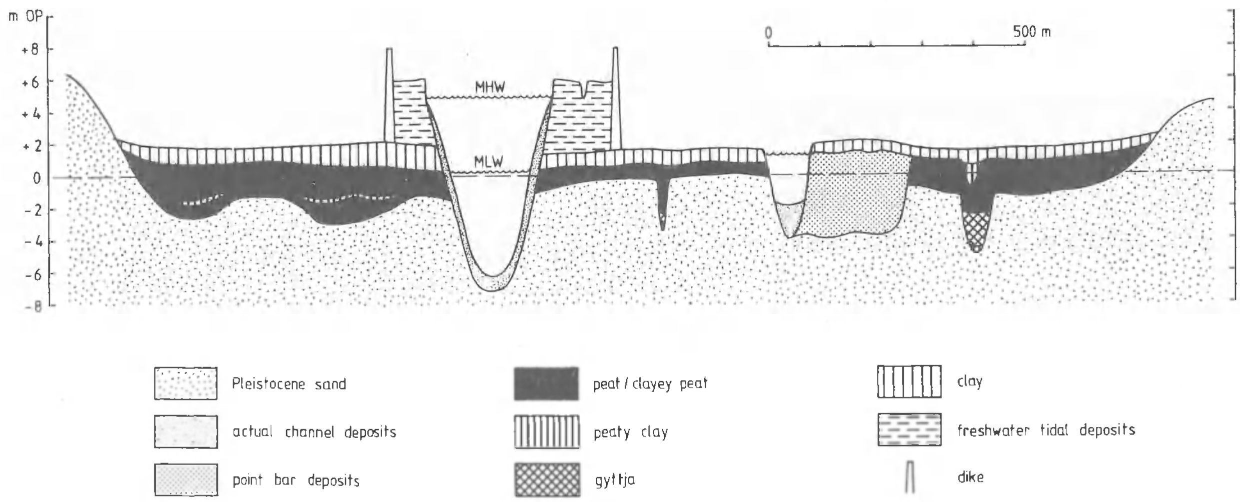

Figure 2 Synthetic geological cross-section of the Holocene deposits of the alluvial plain of the river Scheldt in the study area.

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

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

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

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

LATE HOLOCENE EVOLUTION OF THE TIDAL PART OF THE RIVER SCHELDT

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

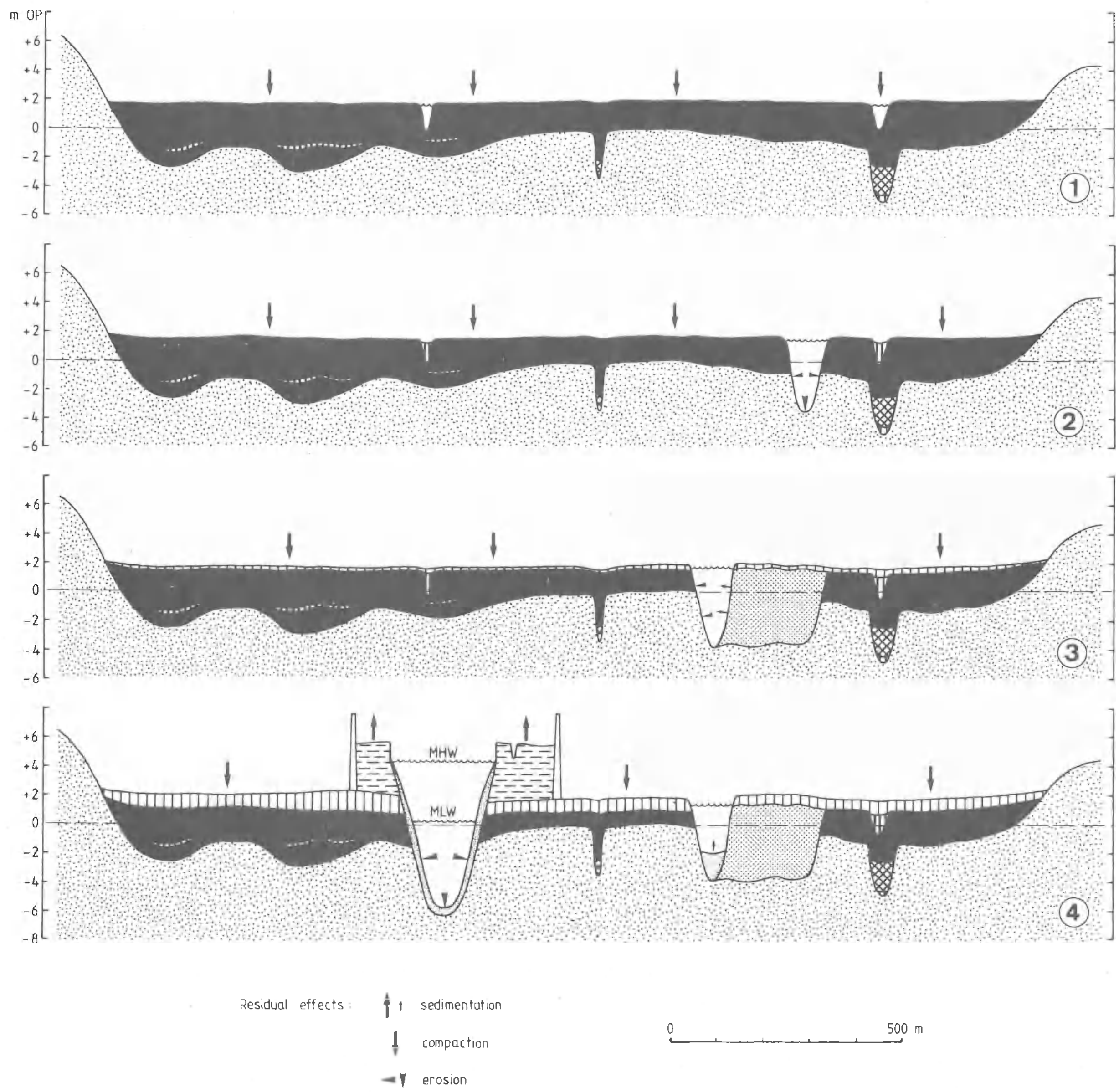

Figure 3 Schematic reconstruction of the Late Holocene evolution of the Scheldt alluvial plain in the study area.

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

1. Accumulation of clayey peat

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

2. Incision of river channel and end of peat formation

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

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

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

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

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

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

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

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

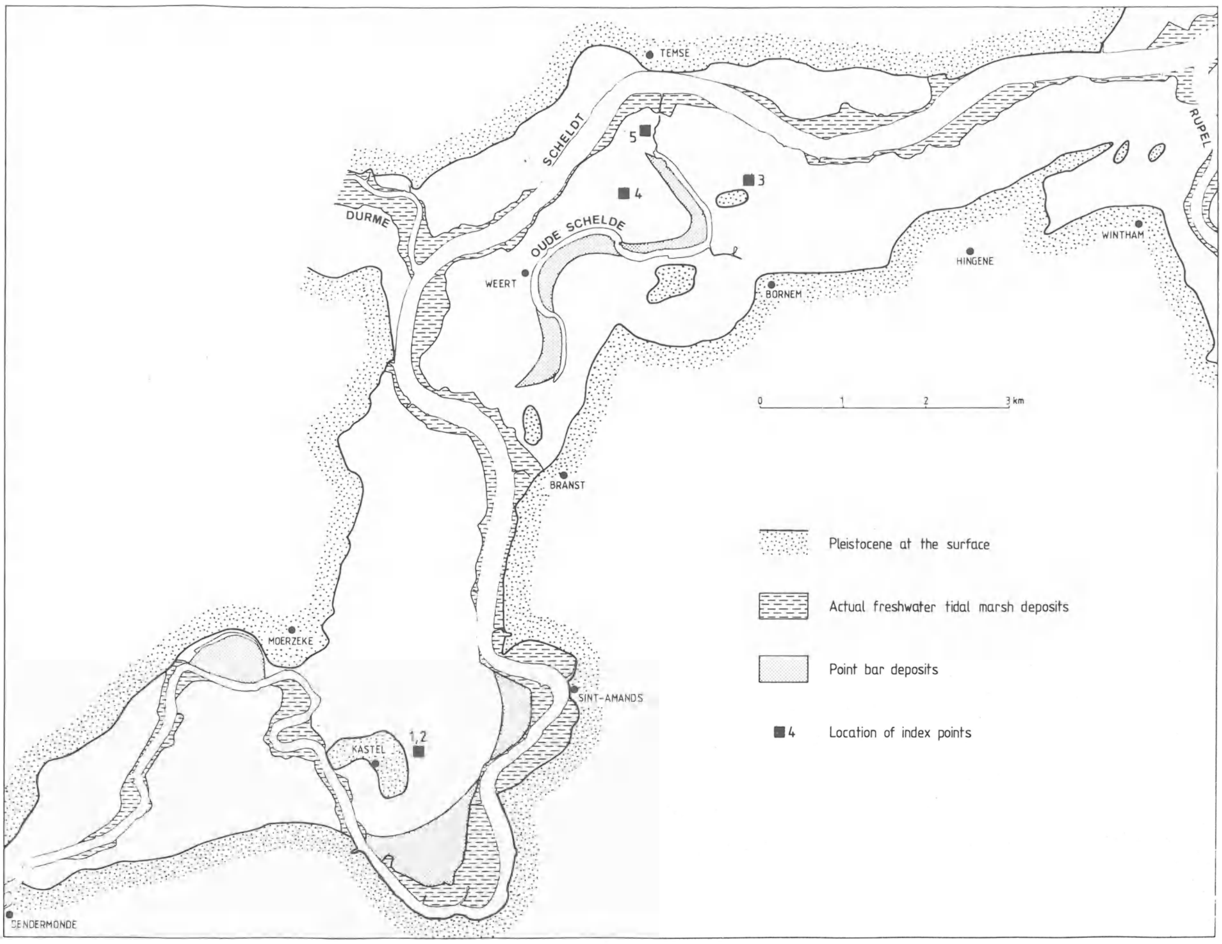

Figure 4 Map of the study area showing late Holocene point bar and freshwater tidal marsh deposits and location of MHW index points in the alluvial plain of the river Scheldt.

4. Penetration of tidal action

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

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

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

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

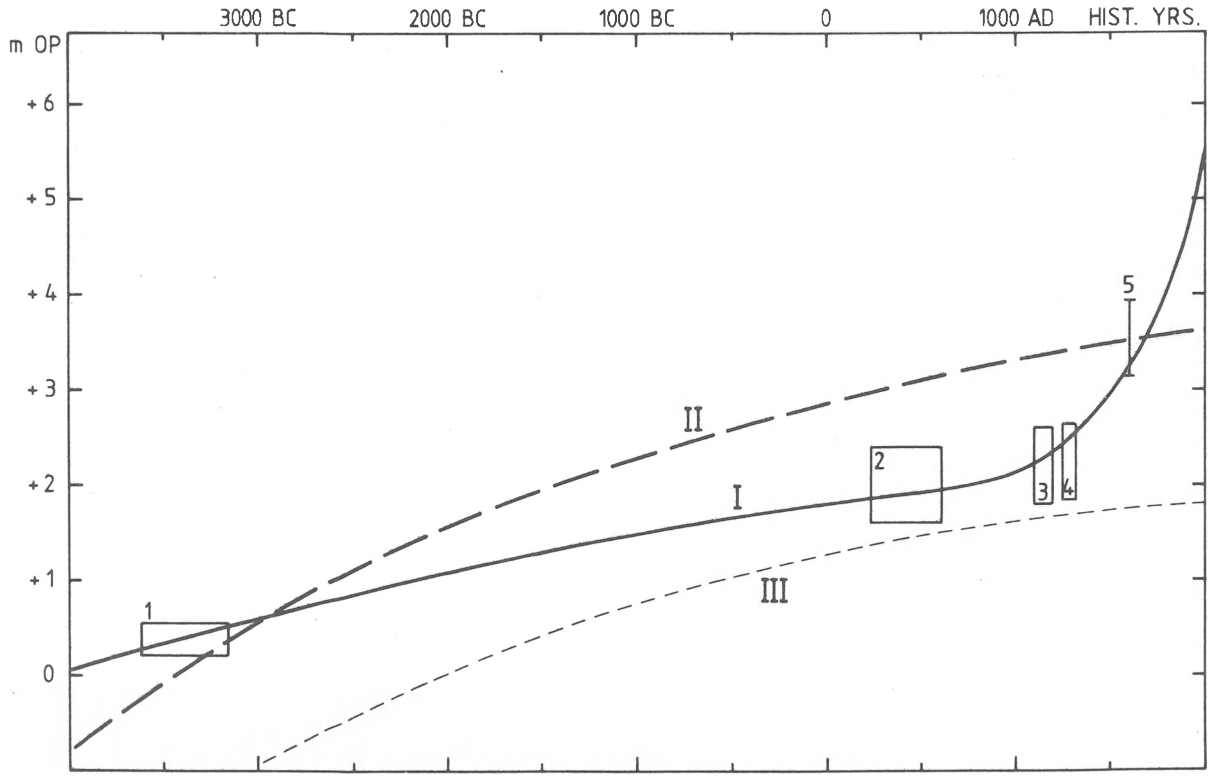

Figure 5 Time-depth diagram with error boxes of MHW index points and local and coastal water level curves. Curve I represents local MHW level, while curves II and III respectively indicate coastal MHW and MSL. A full explanation is given in the text.

THE LATE HOLOCENE FLOODBASIN EFFECT IN THE SCHELDT TIDAL RIVER SYSTEM

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

1. Available time-depth data in the study area

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

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

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

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

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

2. Method of curve construction

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

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

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

3. Interpretation

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

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

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

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

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

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

4. Discussion

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

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

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

CONCLUSIONS

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

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

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17 Maintenance on the Flemish Banks

G. DE MOOR

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

ABSTRACT

The paper analyses present day residual short term morphodynamics and sediment dynamics during the period 1982-1986 on the Kwinte Bank, one of the Flemish Banks using volumetric monitoring, visual bank profile comparison and mapping of bottom load transportation pathways, diagnostic bedform characteristics, departing from detailed sequential bathymetric profiling along fixed reference transverses, high precision navigation and positioning, echosounder record normalization, uniformization and equalization as well as from areal side scan sonar recording of bottom topography. It aims at a quantified approach of shape and location change and of sediment migration problems.

It presents numerical data and a typology about volumetric evolution trends on the bank and on different bank parts. It advances a dynamical subdivision of the bank on sedimentdynamical and geomorphological grounds. It puts forward the existence of different sand transportation mechanisms such as residual uppiling from both surrounding swales by deflection of opposite directed residual bottom currents and providing bank maintenance, vertical sediment exchanges and longitudinal wave-like sediment transport along the bank. It advances an alternation of periods of local upbuilding using sand supplies at least partly eroded from other bank parts. It advocates a residual erosive nature of the steep bankside and alternations of the steep side on both sides of kinks due to differences of evolution stage in a longer term bank position oscillation pattern, using therefore internal sedimentary structures and sequential map analyses.

THE BANKS AND THEIR ENVIRONMENT

Off the Belgian coast in the Southern Bight of the North Sea, lies a remarkable complex of large offshore sand banks, reaching lengths of tens of kilometers, widths up to a few kilometers and relative elevations up to 20 m (figure 1).

Differences in orientation and position allow to group the banks into several systems. The Coast Banks are nearshore and run parallel to it. The Flemish Banks have a more SW-NE direction, the Hinder Banks are further seawards and have a more or less S-N direction, the West-Zeeland Ridges lie more eastwards and run in a more W-E direction.

These banks are separated by channels with maximum depths of about 30 m below MLLWS. The general top level rises shorewards. In their southern parts some of the banks have their crests even at less than 3 m below the low water level.

The channels in between the banks generally present a slight offshore sloping. Channels on both sides of a single bank sometimes have their bottoms at quite different altitudes. This especially occurs on both sides of the Kwinte Bank (figure 2a). Some channels show a low parabolic threshold at their shoreward side. This occurs in the somewhat shallower Negenvaam channel. Moreover bottom sediment in the Kwinte channel consists of coarser sands than in the adjacent Negenvaam, while in both channels sands are coarsening in an offshore direction.

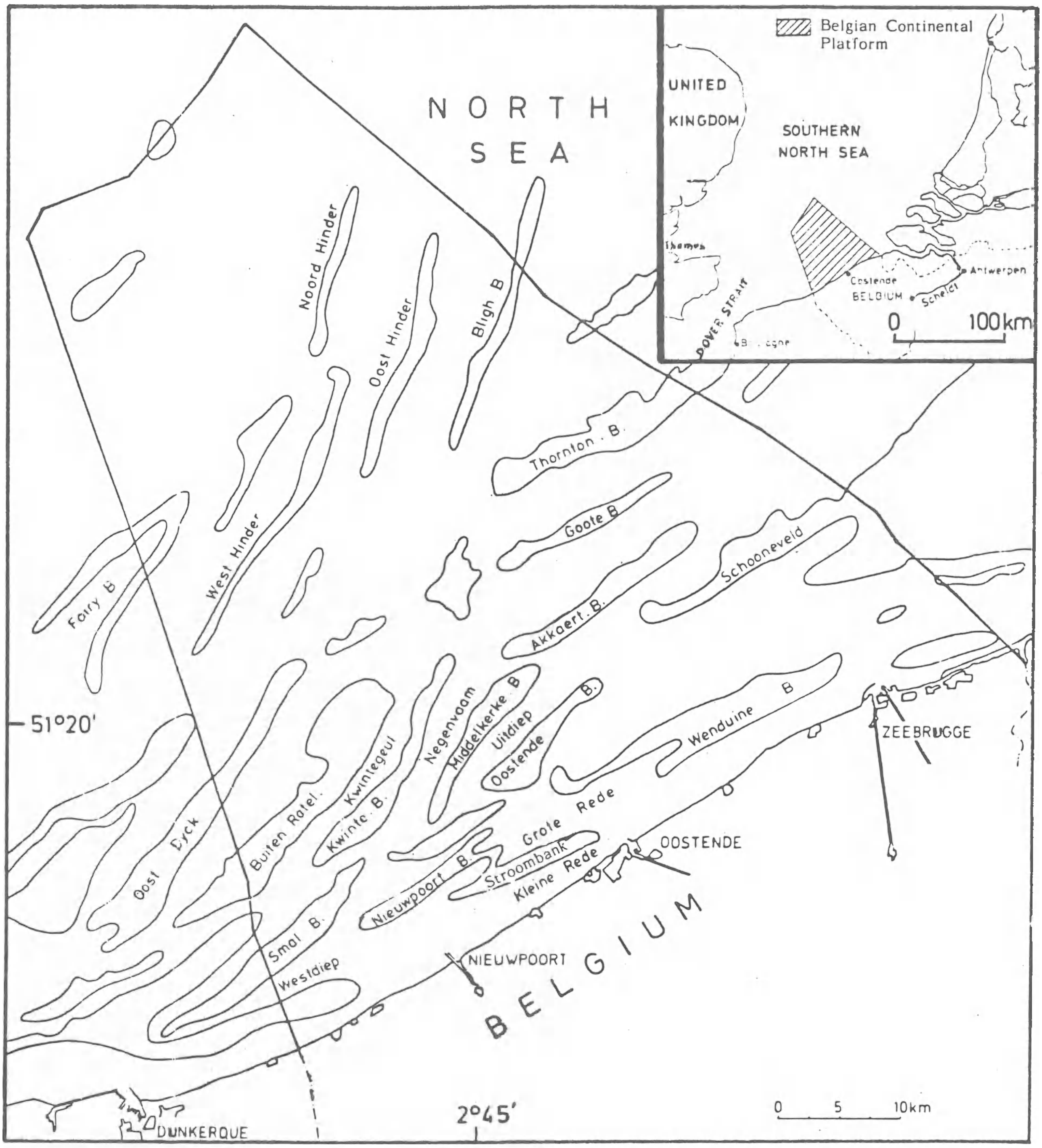


Figure 1 The Belgian Continental Platform and the different sand banks.

Most of the banks are rectilinear, a few are parabolic. Such one is the Fairy Bank. Many present a longitudinal articulation varying from a slight kink to a S-shaped planform. This is the case with the Kwinte Bank, the Westhinder Bank and others.

On the offshore side of such an articulation the Buiten Ratel Bank presents a remarkable widening into a bank head.

Most of the banks show a distinct transversal asymmetry (figures 2a and 2b). The slope intensities however are much less important than apparently recorded on echosounder profiles. This is due to the low height/length scale ratios on such records. The steep sides generally do no exceed 5 to 7 %, while the slight slopes generally are inclined at about 2 to 2.5 %. Some banks present a differently directed asymmetry on both sides of a longitudinal articulation. This is distinctly the case with the Westhinder Bank (figure 2b).

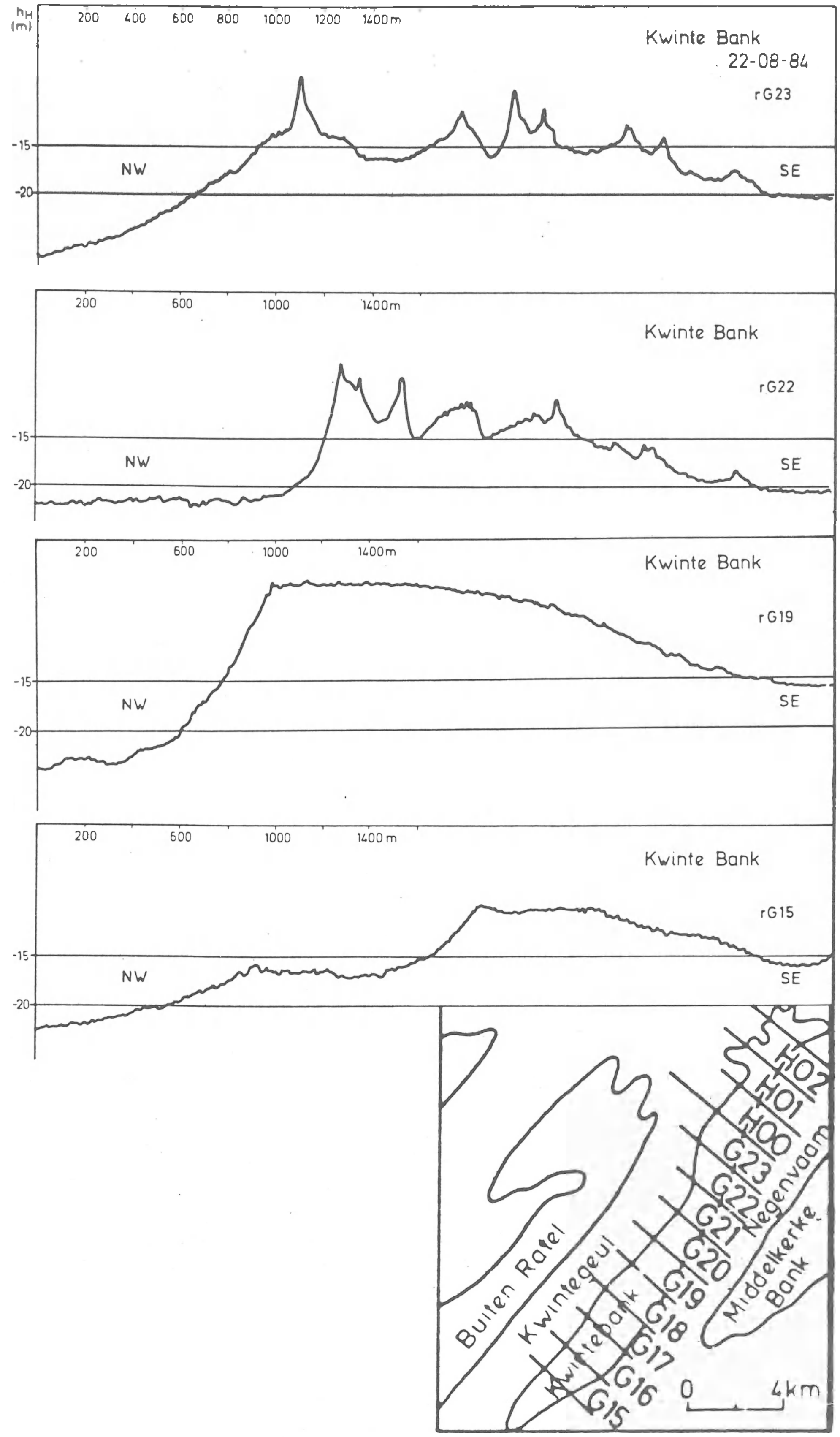


Figure 2a Summer transverses of the Kwinte Bank on 22 August 1984 from the northern part (rG23, rG22) over the central part (rG19) to the southern part (rG15). The northern part is characterized by high asymmetric sand waves climbing both bank sides ; the central part shows a flattened top zone ; the southern part presents a western bank side terrace. Remark the general asymmetry with a steeper residual erosive western side and a residual accretion on the slighter eastern side, as well as the depth difference between the western Kwinte swale and the eastern Negenvaam.

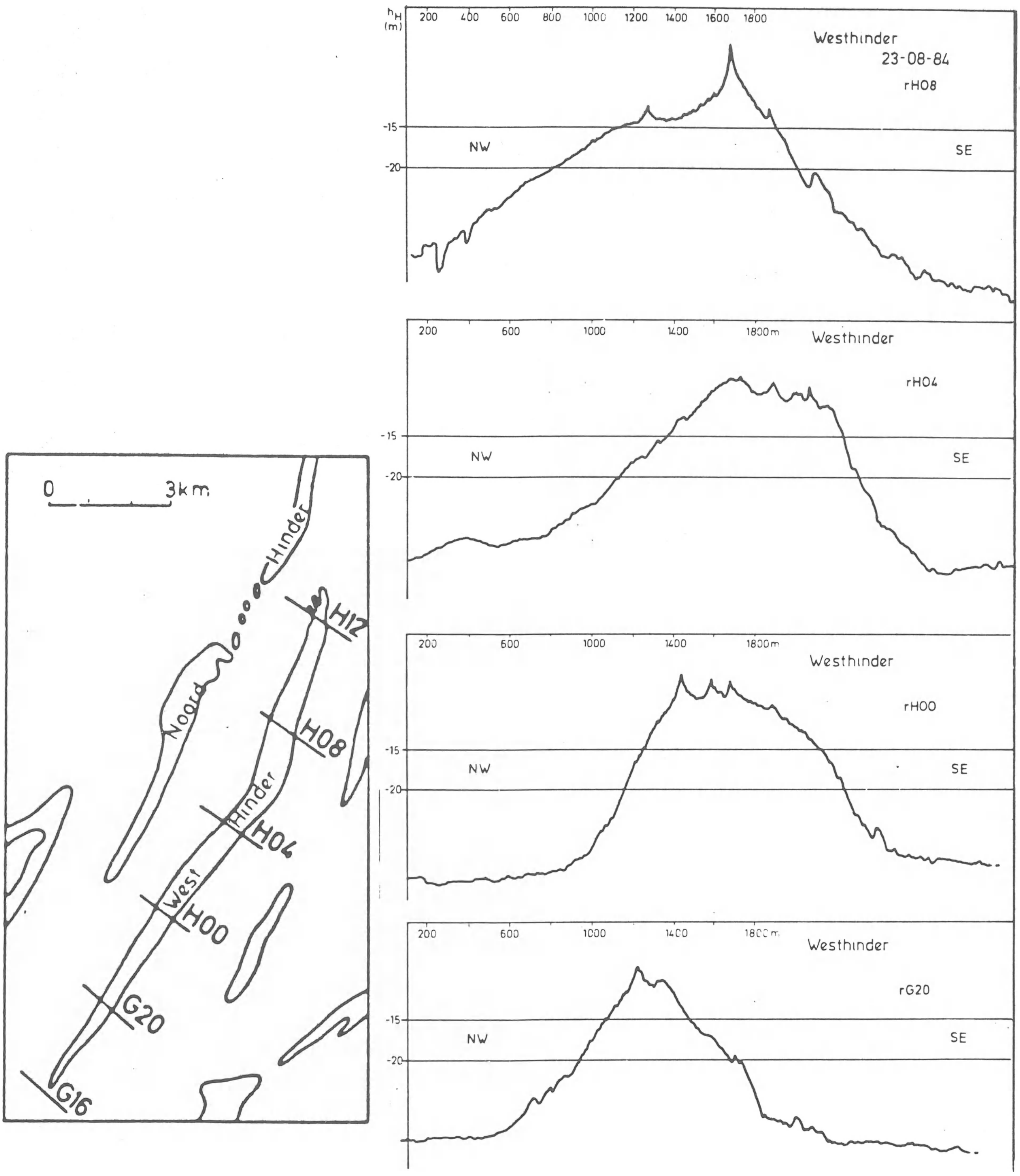


Figure 2b Summer transverses of the Westhinder on 23 August 1983. Remark the opposite asymmetry on both sides of the kink near transverse rH06, pointing towards two bank parts in a different stage of evolution.

Most banks and channels are completely or partly covered with fields of linear bedforms of various types : asymmetric and symmetric sand waves, reaching even heights of 10 m (north of the Westhinder) ; megaripples ; ripples, often covering each other (figures 2a and 2b). These bedforms move in various directions. This is indicated by the various orientations of the steep progradation side and by the differences in the strike of the linear bedforms as they are shown on side scan sonar registrations (figure 3). This strike is considered as running perpendicular to the residual bottom load transportation direction.

That sediment dynamical interpretation of bedform asymmetry however seems not applicable to the sand banks themselves as high resolution subbottom profiling on some of the Flemish Banks (DE MOOR, 1983 ;

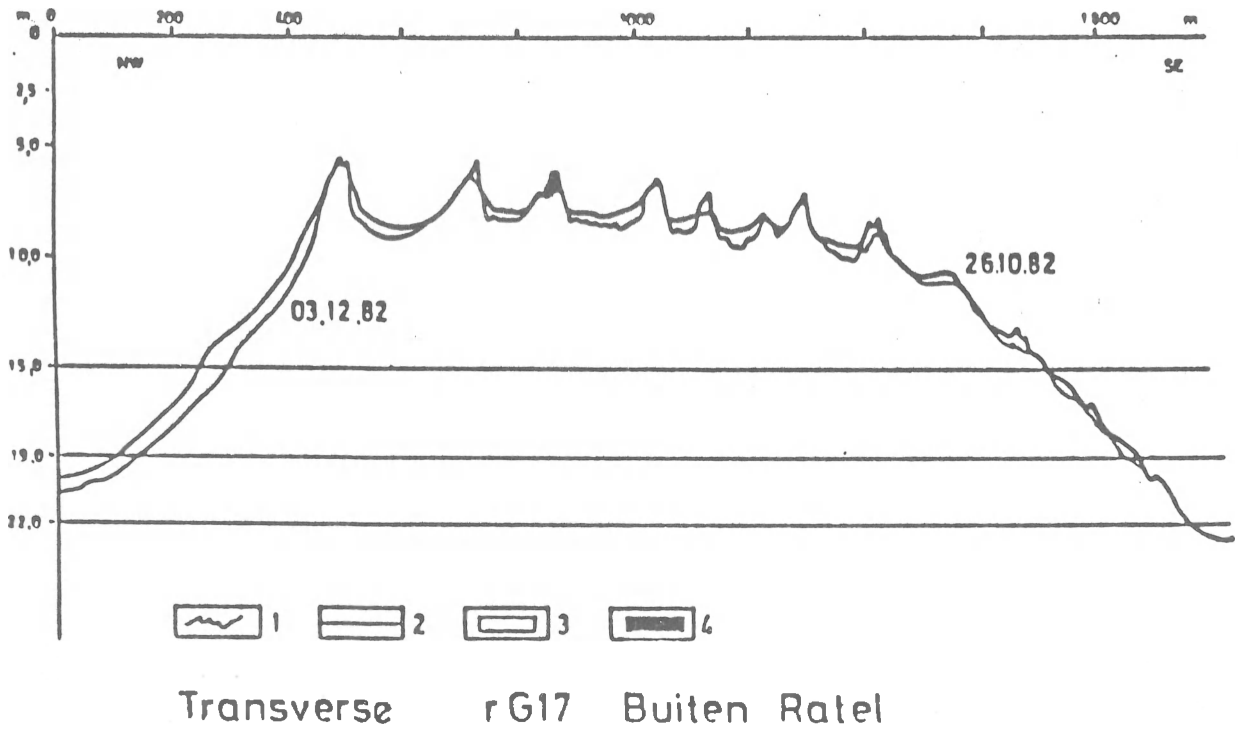


Figure 2c Superposition of two successive bathymetric profiles along transverse rG17 on the Buiten Ratel Bank. Development of longitudinal erosion furrows between sand waves on top of the Buiten Ratel Bank. 1 = sequential profiles. 2 = reference levels. 3 = losses. 4 = gains.

1985a ; 1985b ; 1986) indicates internal sedimentary structures beneath the steep side, distinctly cut by that steep slope, while beneath the slightly sloping side internal sedimentary structures present some parallelism with the slope itself (figure 4). Therefore the steep slope of the Flemish Banks sand banks is not a progradational but a residually erosional slope and the slight slope is one of recent aggradation. Moreover, the fact that there are no indications of present day residual displacements of banks of any importance, at least in the Flemish Banks zone, but that on the contrary arguments arise in favour of a restricted medium term oscillation of the banks around their longitudinal axis (VAN CAUWENBERGHE, 1971 ; DE MOOR, 1985a ; 1986), enhances the idea of a medium term oscillation mechanism (DE MOOR, 1985) and points towards a residual position stability of these banks. Moreover differences in the stage of oscillation on both sides of a longitudinal articulation, could be an explanation for the existence of a different asymmetry on both sides of such kinks and even a reason for the formation of kinks themselves.

The micromorphography of the bank summits generally varies along their length axis. On many banks, such as the Kwinte Bank, the Buiten Ratel Bank, the Westhinder Bank and others, the central part is quite flat and at a relatively high level. The landward side and especially the seaward front part of these banks are covered with remarkable sand waves, having their axis more or less parallel to the bank axis and their summits at a lower level than the central part of the bank. On both sides of the Kwinte Bank, the prograding steep sides of the sand waves are more or less directed towards the bank axis, indicating a residual progradation of these bedforms towards the bank summit and a possible uppiling of sand at the bank top (figure 2a).

Such continuing uppiling would lead to an emergence of the bank, which however is not known to have occurred, at least not in the last decades. It suggests a mechanism of reworking of the uppiled sand by wave activity, especially by storm waves. This idea is supported by the occurrence of seasonal variations of bedform characteristics, by the occurrence of short living erosional furrows (figure 2c) running longitudinally between sand waves, as indicated by sequential bathymetric profiling (DE MOOR, 1983 ; 1984), and by the fact that the top level possibly is related to a mean wave base level, changing seasonally.

In the whole area surficial sediments consist of Holocene marine sands (Bligh Bank Formation).

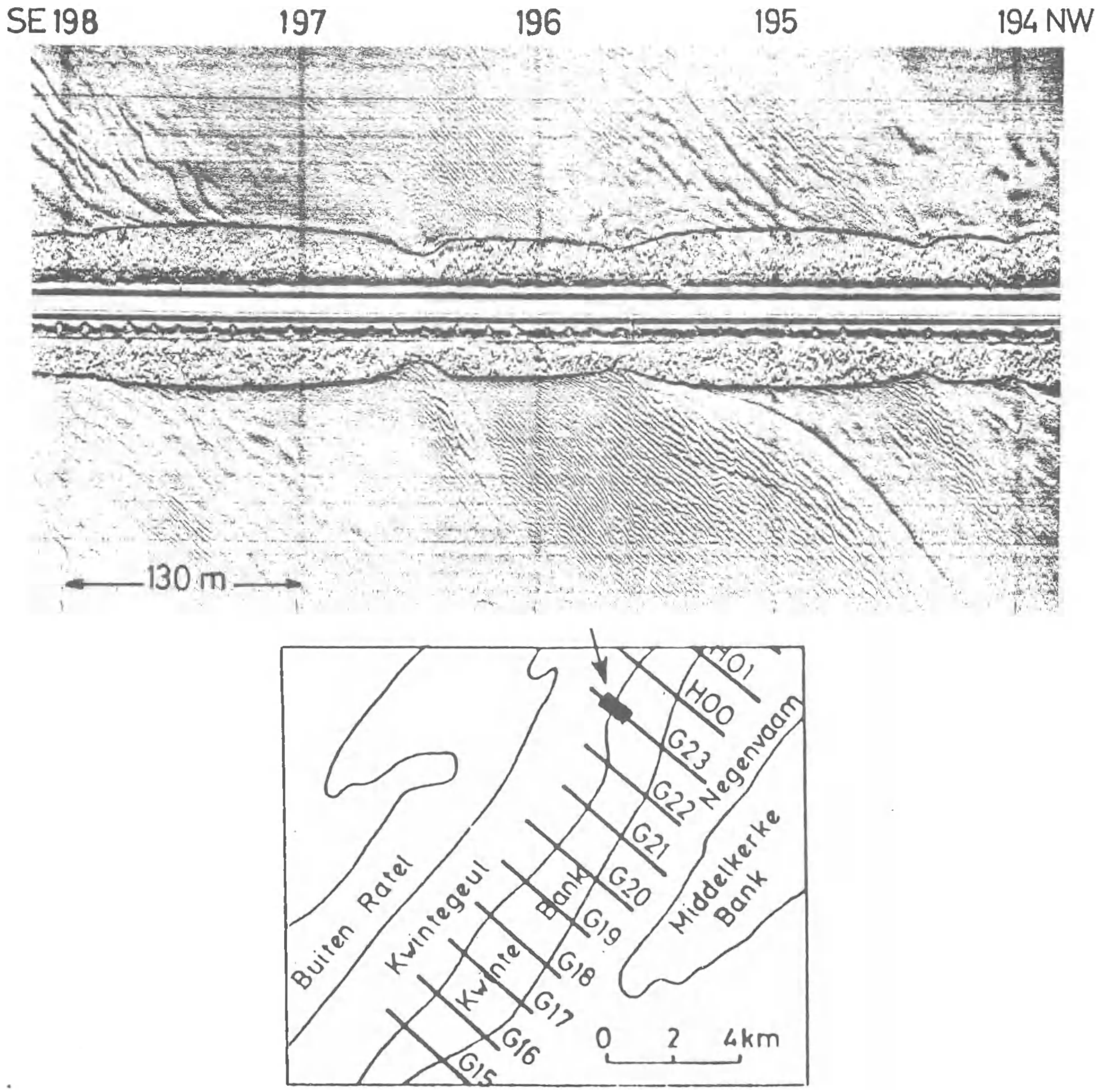


Figure 3 Northwest side of Kwinte Bank side scan sonar record along transverse rG23 on 2 December 1982. v = ship's heading. t = transducer depth. b = sea floor depth below t.

On the Flemish Banks the sediment coarseness seems not merely related to the bank or channel position as in both cases there is a seaward coarsening. Nevertheless in between the Flemish Banks themselves channel bottom sediments often are less coarse than the adjacent bank sands and strips of coarser sands seem to cross obliquely the morphography. Locally the swales present mud patches of varying importance, and gravelly spots as well.

A much denser sampling, to be worked out simultaneously over a large area, and a more accurate geographical, morphographical and bathymetrical positioning would be needed to relate granulometric characteristics to bedform types and to possible sand transportation paths.

Acoustic reflection data (DE MOOR, 1983 ; 1985), borings (CARPENTIER, 1970 ; RIJKS GEOLOGISCHE DIENST, 1968) and seismic reflection data (MARECHAL and HENRIET, 1983) indicate that the banks consist of relatively thick, mostly sandy quaternary sediments with thicknesses sometimes up to 20 m or even more, while in the deeper channels the older substratum often is at shallow depth. Locally the older

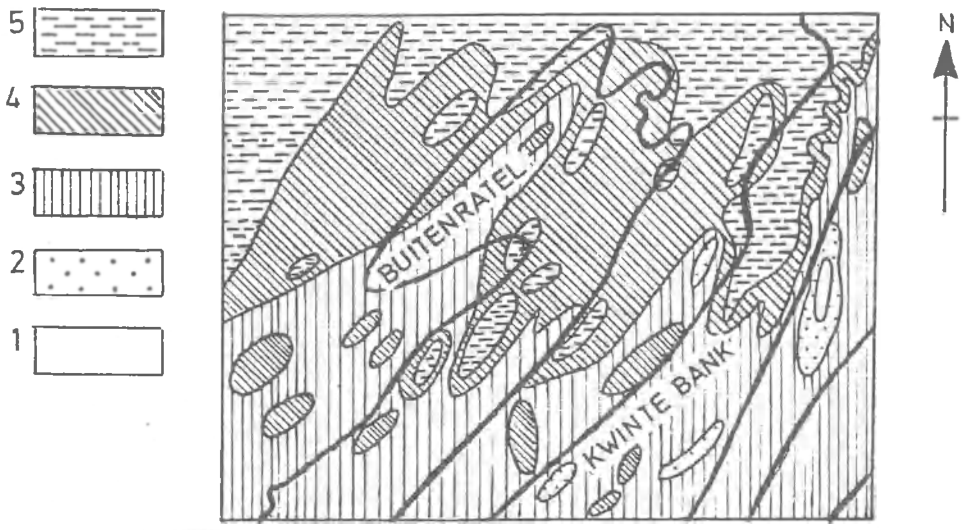


Figure 4 Granulometry sand fraction 50 µm - 4 mm. Graphic mean (FOLK and WARD).
1 = 50-100 µm. 2 = 100-200 µm. 3 = 200-250 µm. 4 = 250-300 µm. 5 = 300-500 µm.

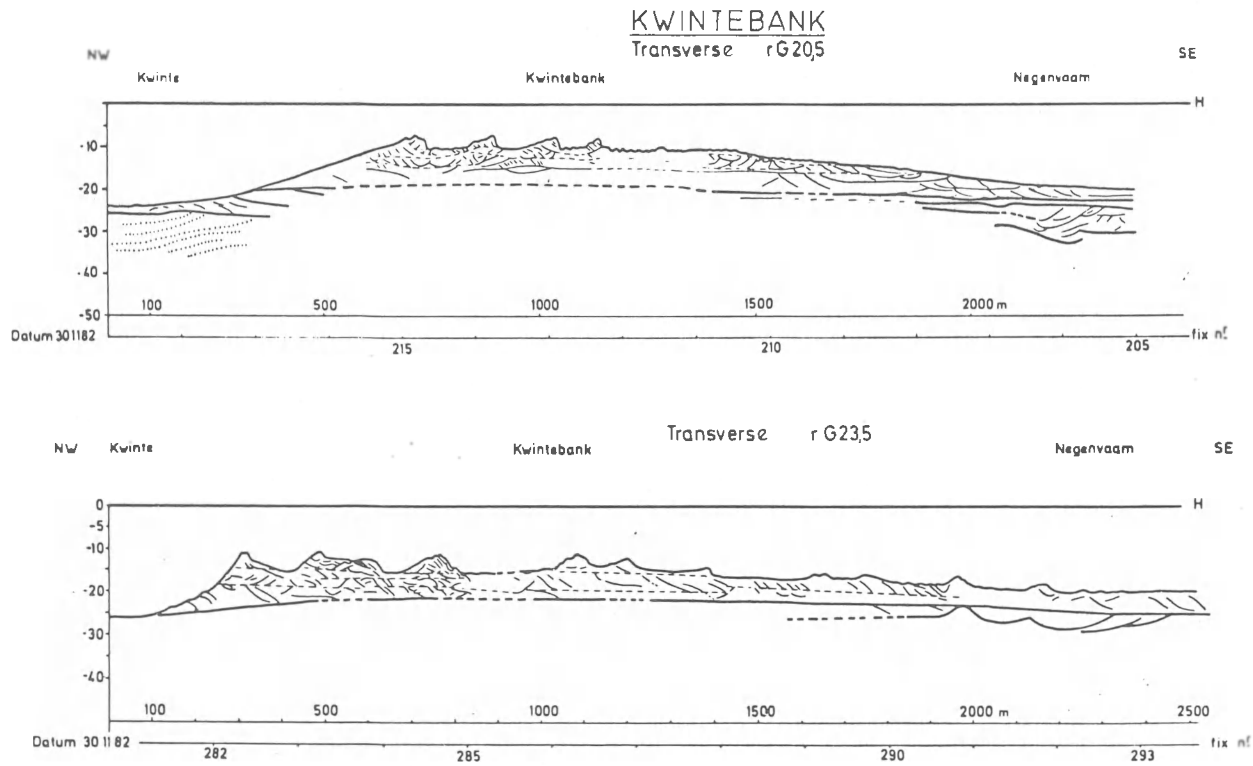


Figure 5 Subbottom profiling of the internal sedimentary structures of the Kwinte Bank along transverses rG20.5 and rG23.5 on 30 November 1982, showing the structural difference between both bank sides.

substratum is at greater depth under the channel bottom than below the bank. This depth pattern suggests the existence of low subbank cores. Figure 5 provides a model of internal structure of the Kwinte Bank.

The geological constitution of banks and channels, and their morphographical characteristics put the problem of the erosional, depositional or transportational nature of the banks.

The whole area has a diurnal macrotidal regime and is the seat of strong semidiurnal anticlockwise rotating tidal currents reaching peak velocities of 2.5 kts.

At Ostend tidal amplitude reaches about 3 m during mean neap water periods and about 4.6 m during spring tides. Due to the tidal command by the Norfolk amphidromic point, to its eastward approach of the coastline, and to the phase superposition of the Channel and the North Sea tidal waves, tidal range decreases north- and eastwards off the Belgian coast.

Eb peak current velocities are lower than flood peak current velocities and eb peak currents run in directions slightly different from opposite to the flood peak ones as shown by the elliptic tidal roses (VAN CAUWENBERGHE, 1981). As the tidal wave moves anticlockwise, the mean ebb currents run to the SW, the mean flood currents to the NE. This was the reason why VAN VEEN (1935) considered the Northern Flemish Banks as breached eb parabolas and the Southern Flemish Banks as breached flood parabolas.

Moreover detailed analysis of tidal current velocity and direction data, recorded on progressive vector diagrams proves the existence of residual water displacements. Together with the rotational character of the tidal current and the differences in duration and peak velocity of ebb and flood currents, residual water displacements are at the origin of a complex mechanism of sand sorting and transportation in different directions according to granulometry.

PROBLEMS AND AIMS

The banks off the Belgian coast present all the classic problems of the offshore bank morphology. They raise numerous questions related to the various mechanisms involved in the different aspects of their genesis, their evolution, their planform, transversal morphology, pattern and alignment, their shape and dimensions, their bedform components, their position, shape and volume stability and their survival and maintenance in a highly energetic environment, as well as about the origin, the pathways and the supply mechanisms for sands.

Despite the numerous aspects, the global problem presents 7 main topics :

1. the typological and geographical inventory of the submarine landforms and bedforms ;
2. the genesis of the banks during the pre-Atlantic Holocene transgression ;
3. the evolution of the banks during the fluctuations of the Holocene high sea level stand ;
4. the morphographic characteristics of the banks ;
5. the present day morphodynamics of the banks, considered in the framework of different factors ;
6. the present day sediment dynamics, especially the sediment transportation pathways and their relation to hydrodynamics ;
7. the stability and maintenance of banks over different terms and the possible meaning of present day maintenance mechanisms upon the genesis, evolution and survival of the banks over long terms.

Since the pionneering work of VAN VEEN (1935), introducing echosounder sensing, and that of HOUBOLT (1968), introducing side scan sonar detection and subbottom profiling, the Flemish Banks became a reference area for shelf sand banks.

Many authors studied one or several aspects of the morphogenetic complexity of these banks, often departing from the morphological analysis of hydrographical maps. Moreover the results of investigations on similar sand banks, especially the East Anglian Banks, cannot be overlooked. A review of the numerous genetic theories that have been advanced is given by DE MOOR (1986).

In order to understand landforms, processes and evolutions it is essential to distinguish short, medium and long term components as well as local and regional, instantaneous and residual aspects and to take into account the megatidal environment with its semidiurnally rotating currents, presenting velocities continuously changing between more or less opposite, unequal and varying eb and flood peaks and slack minima, as well as the impact of the wave characteristics.

It is not the purpose of this paper to deal with a more detailed bathymetric mapping, nor to focus on the geological structure of the banks and their origin and age, but to obtain *quantitative data* about the present day short term residual morphodynamics and sediment dynamics by using *geomorphological techniques* in monitoring the basic volumetric variations and the related shape and location changes, and in evaluating the sediment migrations on ground of their morphological repercussions. Hereby the attention is more directed towards the *quantified description* of the present day phenomena and especially towards present day short term evolution trends, than upon their explanation, which undoubtedly will require detailed hydrodynamical data.

Moreover attention is paid to the possible meaning of these present day dynamics and maintenance for the interpretation of the internal sedimentary structure of the banks and for the explanation of their long term evolution.

Morphodynamics is the study of the changes in dimensions - especially volume -, in shape and in position of landforms.

Sediment dynamics is the study of the intensity, the pathways and the mechanisms of the migrations of sediments.

Residual means that we will consider the changes resulting from successive positive and negative movements after numerous cycles of tidal, wave and current actions and over periods of increasing length.

The Kwinte Bank, one of the Flemish Banks, has been chosen as testing area, especially because it is, since 1978, an area of continuous sand extraction at a more or less controlled rate of about 400,000 m^3/year, mainly localised on its northern offshore part.

METHODOLOGY AND RESULTS

Within that geomorphological and quantitative approach the fundamental way to tackle the problem of present day short term residual morphodynamics and sediment dynamics consists in comparing qualitatively or quantitatively successive total or partial cross-sections along representative reference transverses of the bank, and in inferring transport pathways either from bedform characteristics indicating the residual migration direction of the bottom load transport or from the geographical distribution of the sediment losses and gains.

Qualitative evaluation of volume, shape, dimension and position changes of a bank transverse can be obtained by superposition and visual comparison of successive comparable cross-sections along a reference transverse. A posteriori the changes in profile can be expressed quantitatively.

Evaluation of changes in the total volume can be approached quantitatively by computing the differences between unit-volumes of the total cross-section along a reference transverse and above a reference level obtained at different dates (total unit-volumes).

The unit-volume corresponds to the area of the vertical cross-section times a width of 1 m. Unit-volumes are only computed above reference levels cutting the profile at least on both sides of the bank and bankwards of the bank's base-concavity.

Evaluation of shape and position changes of the bank cannot be inferred from the total unit-volumetric changes along the cross-section. It can be obtained by comparing successive unit-volumes for bank-halves delimited by a fixed position reference, or by comparing unit-volumes for total or half-bank reference slices delimited by reference levels, or by comparing unit-volumes for reference columns delimited by reference verticals (partial unit-volumes) (figure 6).

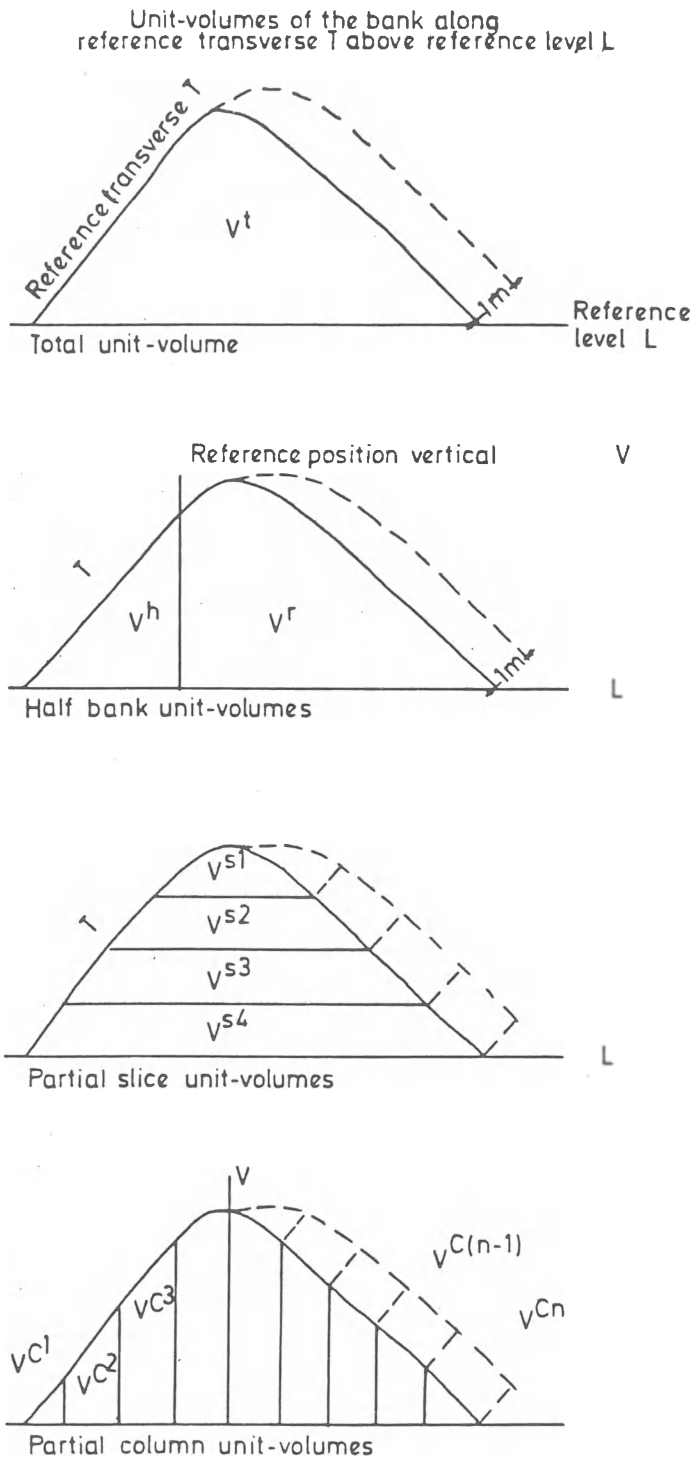


Figure 6 Kwinte Bank. Unit-volumes of the bank along reference transverse T above reference level L.

1. Basic operations

Therefore this morphometrical approach comprises two basic operations.

1.1. Monitoring of bank profile variations

This departs from sequential bathymetric profiles, sailed within a very short time in successive campaigns along fixed representative reference transverses in a sufficiently dense network. They are recorded with high depth and position accuracy and geographical reliability, and with high vertical and areal resolution, which means high registration frequency. They have to be transformed into comparable hypsometric profiles, especially because of variations of the ship's bottom velocity and because of variations in the sea level due to the continuous vertical tidal movement.

Such original echosounder transversing is used instead of successive hypsographic map analysis and comparison because of the long period required for the construction of such maps so that every set of depth-data, covering a larger area such as one bank, no longer is built up by sufficiently comparable data and possibly because of the maps resolution as well.

Moreover the use of older maps and sounding data is subject to great care because of possible inaccuracy of positioning and sounding data.

1.2. Mapping of the geometric characteristics of linear bedforms

Type and height, but especially strike and direction of the steep progradation side of linear bedforms are considered as diagnostic criteria for a reconstruction of the residual bottom load transportation pathways.

The analysis and the mapping of the geometric characteristics of bedforms can hardly be worked out on the base of bathymetric profiles, as echosounder-records generally provide deformed images of the landforms and bedforms. This is due to the scale ratio and its possible variations provoked by variations in ship's speed, but especially to the impossibility to know sufficiently the relative orientation of the bedform's strike in relation to the ship's track, unless extremely time taking operations. For similar reasons, and especially because of the need for an extremely dense depth station or depth profile network, detailed bedform mapping by echo-sounding can hardly be used.

Therefore mapping of surficial sedimentary structures has been worked out on areal registrations of bottom topography by side scan sonar techniques.

2. Main steps

The whole operation then comprises 6 main steps.

2.1. Navigation and positioning

Repeated accurate navigation exactly along the reference transverses and high frequency positionings are the fundamental prerequisites for the total operation. Sailing at constant bottom speed is very important but not indispensable, because on echosounder registrations scale corrections for velocity variations are possible.

The sequential bathymetric profiles have been sailed along *loxodromic* lines between fixed extreme W-points situated along red Decca lines of the Decca 5B network, which in the Flemish Banks area are nearly straight lines transversal to the banks at distances of about 1 NM (figure 7). In between 1982 and 1986 18 successive sailings along 8 reference lines situated between red Decca 5B lines rH02 and rG16 have been worked out on the Flemish Banks.

Navigation has been worked out using radio-electronic hyperbolic systems : before 1984 Decca Navigator, since 1984 video track plotter aided Decca Navstar, since 1987 combined with simultaneous video track plotter assisted pseudo-hyperbolic Syledis support.

Positions have been taken every 60 seconds on Decca Navigator before 1984 ; afterwards every 30 seconds simultaneously on Decca-Navstar, on Decca Shipmate and on Toran, using the Belgian Toran chain, and since 1987 on a Belgian Syledis chain. Positionings have been attented by on-line data aquisition of the different system coordinates and other nautical parameters such as ship's longitudinal bottom velocity and ship's heading, by their monitor display, by on-line video track plotting, by immediate control of geographical reliability of the hyperbolic positionings, and by data storage by ship's HP 1000 minicomputer followed by data transfer and fix positioning by compatible laboratorium HP 600A minicomputer. All registrations have been provided with 1 minute time marks, corresponding with

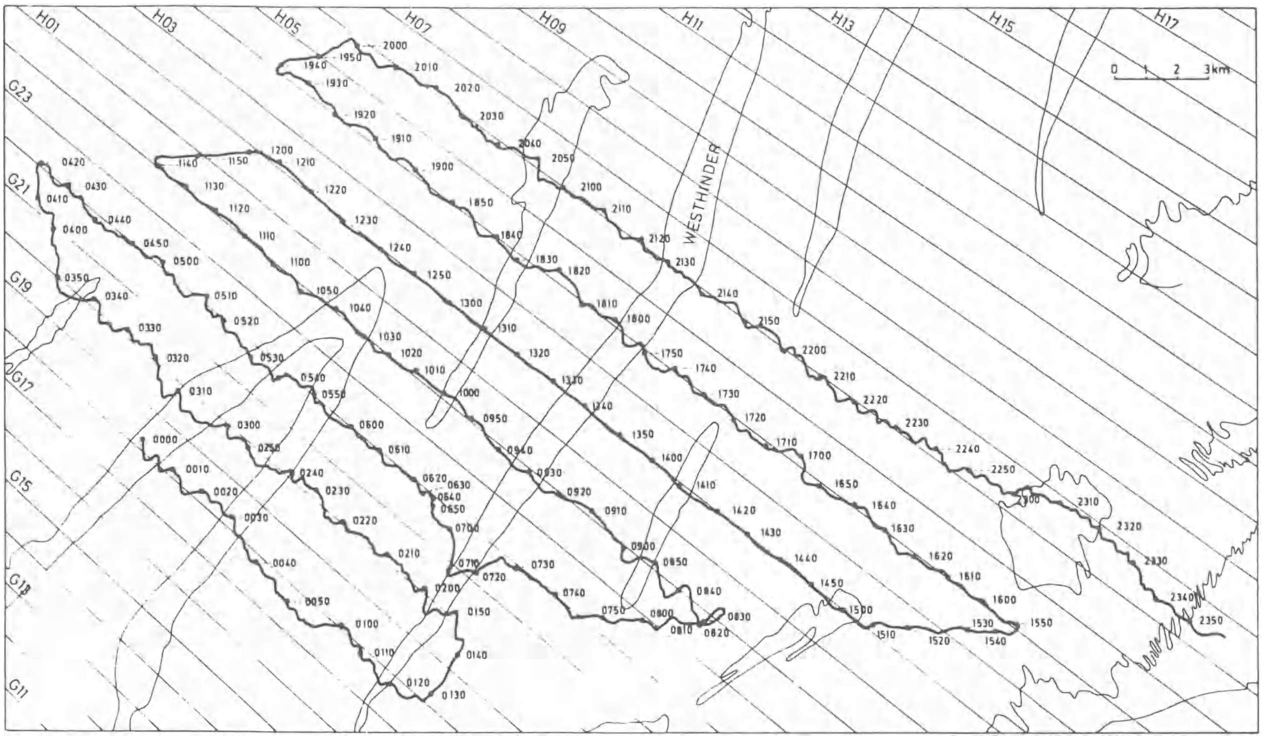


Figure 7a Track plot for "Belgica" on 26 November 1986 between 00h00 and 23h50 (GMT+1) in UTM 31 coordinates computed from NAVSTAR geographical coordinates, based on Decca 5B received coordinates. Positionings every 30 seconds, fixes shown every 20 minutes.

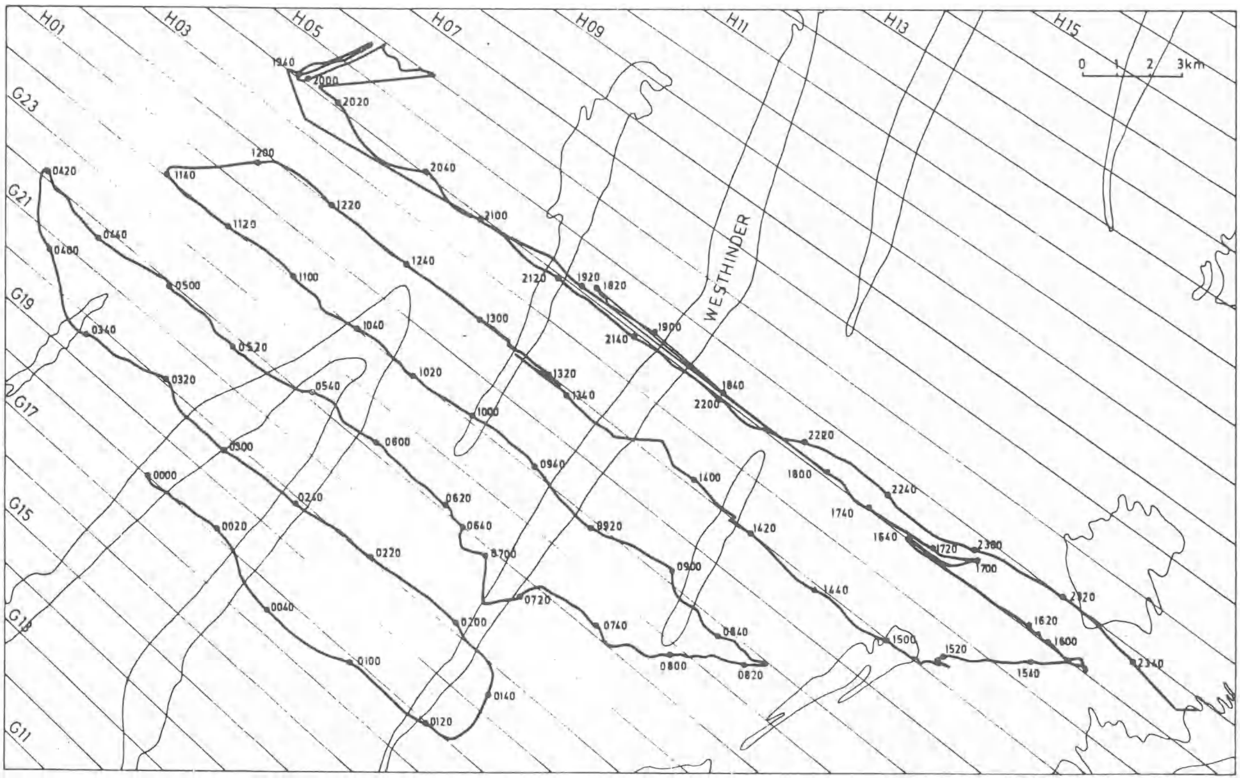


Figure 7b Track plot for "Belgica" on 26 November 1986 between 00h00 and 23h50 (GMT+1) in UTM 31 coordinates, based on TORAN-coordinates. Positionings every 30 seconds, fixes shown every 20 minutes. Between 09h00 and 15h00 the plotted track is geographically reliable and shows little deviation from the reference transverse. Between 19h40 and 23h00 deviation due to lane-slip. At 07h10 deviation because of navigation necessities.

positionings, easying cartographic elaboration and time/length scale transformation and length scale uniformisation.

Theoretical accuracy of the Decca 5B network in this area varies between 15 and 25 m, theoretical accuracy of Toran between 2 and 5 m, Syledis around 2 m.

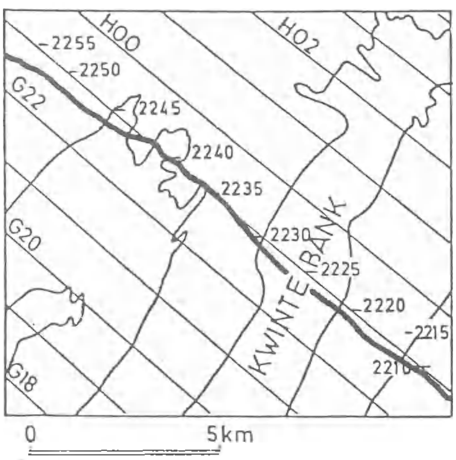
With fixed error corrections the simultaneously received different hyperbolic coordinates are transformed on-line into geographical coordinates and displayed. This allows an immediate control upon the geographical reliability of the hyperbolic positions and navigation (figure 8). Navigation and position control on board are completed by radar positionings.

Because of possible impact of variable errors upon hyperbolic positions, further laboratory evaluation of the geographical reliability has been worked out by logical analysis of the position fix plots, and by their comparison with the longitudinal bottom speed and the ship's heading registered simultaneously with the positions.

Moreover, in order to avoid geographically unreliable positions due to overnight instability of the Decca and Toran systems, bathymetric profiling on these systems has been stopped between sun set and sun rise.

Bottom velocity and ship's course have been used to correct the profile for small but real deviations from the reference track (course correction), at least when no doubt existed between real deviations and apparent deviations due to system instability.

No profile sections have been used when situated at more than 100 m away from the reference track.



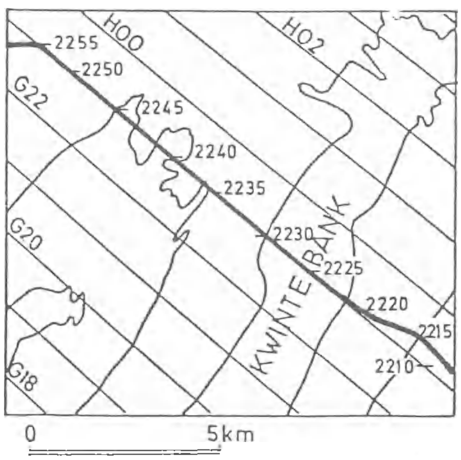


Figure 8a Kwinte Bank. Track plot for "Belgica" along reference transverse rG23 on 24 February 1987 between 22h00 and 23h35 (GMT+1) in UTM 31 coordinates computed from SYLEDIS coordinates. Positions shown every 20 minutes, registered every 30 seconds.

Figure 8b Kwinte Bank. Plot for the same track for "Belgica" along reference transverse rG23 on 24 February 1987 in UTM 31 coordinates computed from NAVSTAR geographical coordinates simultaneously registered.

2.2. Bathymetric profiling

This step comprises accurate bathymetric echosounding registration with high depth and areal resolution along the reference transverses, backed by reduction, correction, uniformization and equalization operations in order to make the profiles normalized and comparable.

Depth profiling has been worked out with Atlas Deso X echosounder before 1984 and, since 1984, with Atlas Deso XX echosounder, working simultaneously on 30 and 210 kHz wave frequency. In case of double reflections, the 210 kHz signal has been used. The vertical resolution is about 10 cm and the digital reading error 10 cm. The profiles have been sailed at a ship's bottom velocity of 10 kts.

In order to obtain comparable profiles usable for morphological analysis, for volumetric computation and for visual comparison by superposition, several corrections are introduced on the bruto echosounder record : tidal reduction every 10 minutes for vertical tidal movement during the profile recording, by a tide reduction procedure (VAN CAUWENBERGHE, 1977), using tidal curves for Ostend or Zeebrugge and taking into account a waterlevel correction for differences in tidal phase between observation station and tidal gauge location, as well as for the difference in tidal amplitude at equal tidal phase. The general zero datum level used for tidal reduction and for bathymetry is the local MLLWS, as defined by the Belgian Hydrographic Service of the Coast ;

- time/length scale transformation ;
- possibly length scale uniformization for varying ship's bottom speed by a speed correction procedure ;
- possibly course correction for real but short course deviations ;
- length scale equalization on sequential profiles ;
- drawing of reference level and reference position marks.

All registrations have been preceded by a sound velocity adaption and a transducer depth adaption.

There have been no real time wave compensations. No profiles have been used for visual comparison or for bedform analysis when wave amplitude exceeded 1 m because of interference. Use of profiles for volumetric computation has been stopped at wave amplitudes of 2 m.

2.3. Volumetric computations

Volumetric monitoring uses several parameters, three of them being important in this evolution of morphographical changes and sediment mobility.

The unit-volume (V) of the bank or bank part profile along a reference transverse above a given reference level or between reference levels at a fixed date, has already been defined. It can be worked out for the total profile (total unit-volume) or separately for morphographical components, for reference depth slices or for reference length columns of the profile (partial unit-volumes) (figure 6).

The unit-volumetric change (ΔV) is the change in unit-volume for a transversal profile over an interval between any two sailings. The interval unit-volumetric change (ΔV_i) gives the volumetric change during the interval between two successive sailings. The residual unit-volumetric change (ΔV_c) provides the volumetric change between the unit-volume at any sailing in a series and the initial unit-volume. With periods of increasing length this parameter expresses more and more the residual change.

Volumetric monitoring rests upon chronological comparison of successive unit-volumes along a reference track and upon geographical comparison of unit-volumes for different reference tracks but for equal periods over a larger area, a bank e.g.

Such chronological comparison is hindred by differences in length and date of the intervals and by occasional ups and downs of the unit-volume. Geographical comparison is hampered by differences in dimensions and in shape-factors of the transverses.

Chronological normalization is possible by relating the unit-volume or the unit-volumetric change to the length of the interval. It leads generally to asymptotization. Geographical normalization is possible by relating the unit-volumes or the unit-volumetric changes to a reference unit-volume V_R specific for the transverse and the level.

A quantitative and geographical evaluation of morphodynamics and sediment dynamics is obtained using the residual mobility index ΔF_c for periods of equal length and date.

$$\Delta F_c = \frac{\Delta V_c}{V_R} = \frac{V_i - V_I}{V_R}$$

This residual mobility index corresponds to the ratio of the residual unit-volumetric change versus a specific reference volume. It relates the unit-volume at a certain moment as well to the initial as to the reference volume. It is a relative value expressed in % of the reference volume. It can be computed for the total unit-volume, as well as for partial unit-volumes.

Such a reference volume, provisionally used here, is the arithmetic mean of the unit-volume at the initial day (V_1) and that of the last of the measurements.

Daily means for such ratios can be used for one constant interval: $\Delta D_C = \Delta F_C . N^{-1}$.

Table 1 gives a list of the different volumetric parameters obtained for the transverse rH01 on the Kwinte Bank above the level -15 during the successive campaigns of the period November 1982 - November 1986. Tables 2 and 3 give similar information respectively for the reference transverses rH00 and rG22.

2.4. Residual volumetric evolution trend

A first qualitative approximation of the residual unit-volumetric evolution along a bank transverse and above a fixed reference level can be obtained by visual analysis of a time series of corresponding unit-volumes measured at successive data along that transverse. A qualitative idea of the volumetric evolution of a bank, and especially of possible geographical differences in the evolution and therefore of possible sediment transportation, can be obtained by a comparative visual analysis of time series of unit-volumes obtained for

Bank : KWINTE BANK — Initial date : 30-11-82
Reference transverse : rH01 — Reference level : -15 m — V Reference : 610 m^3

Date order	Date	N_i	N_c	V_i	$\Delta V_i = V_{i+1} - V_i$	$\Delta V_c = V_i - V_1$	$\Delta F_c = \frac{\Delta V_c}{V_R}$	$\Delta D_c = \frac{\Delta F_c}{N_c}$
1	30-11-82			679				
2	07-05-83	158	158	888	+209	+209	+0.3462	+0.002168
3	10-06-83	34	192	748	-140	+69	+0.1131	+0.000589
4	13-06-83	3	195	-				
5	12-07-83	29	224	854	+106	+175	+0.2869	+0.001281
6	07-09-83	57	281	566	-288	-113	-0.1852	-0.000659
7	22-08-84	340	621	790	+224	+111	+0.1820	+0.000293
8	28-05-85	279	900	874	+84	+195	+0.3197	+0.000355
9	06-06-85	9	909	474	-400	-205	-0.3361	-0.000369
10	11-09-85	97	1006	465	-9	-214	-0.3508	-0.000349
11	13-09-85	2	1008	362	-103	-317	-0.5197	-0.000516
12	28-11-85	76	1084	392	+30	-287	-0.4705	-0.000434
13	11-02-86	75	1159	541	+149	-138	-0.2262	-0.000195
14	11-11-86	273	1432	464	-77	-215	-0.3525	-0.000246
15	12-11-86	1	1433	182	-282	-497	-0.8148	-0.000569

V_R = Reference Volume — V_1 = Volume at initial date
V_i = Volume at given date order i — N_i = Number of days between two successive dates
V_{i+1} = Volume at next date order — N_c = Number of days at given date since initial date

Table 1 Kwinte Bank. Observation dates and volumetric parameters for reference transverse rH01.

Bank : KWINTE BANK				Initial date : 30-11-82				
Reference transverse : rH00				Reference level : -15 m		V Reference : 1173 m^3		
Date order	Date	N_i	N_c	V_i	$\Delta V_i = V_{i+1}-V_i$	$\Delta V_c = V_i-V_1$	$\Delta F_c = \frac{\Delta V_c}{V_R}$	$\Delta D_c = \frac{\Delta F_c}{N_c}$
1	30-11-82			1081				
2	07-05-83	158	158	-				
3	10-06-83	34	192	2535	+1454	+1454	+1.2396	+0.006456
4	13-06-83	3	195	1238	-1297	+157	+0.1338	+0.000686
5	12-07-83	29	224	1287	+49	+206	+0.1756	+0.000784
6	07-09-83	57	281	1080	-207	-1	-0.00085	-0.000003
7	22-08-84	340	621	1501	+421	+420	+0.3581	+0.000577
8	28-05-85	279	900	1656	+155	+575	+0.4902	+0.000545
9	06-06-85	9	909	828	-828	-253	-0.2157	-0.000237
10	11-09-85	97	1006	1260	+432	+179	+0.1526	+0.000152
11	13-09-85	2	1008	1428	+168	+347	+0.2958	+0.000293
12	28-11-85	76	1084	1252	-176	+171	+0.1458	+0.000134
13	11-02-86	75	1159	1453	+201	+372	+0.3171	+0.000274
14	11-11-86	273	1432	1313	-140	+232	+0.1978	+0.000138
15	12-11-86	1	1433	961	-352	-120	-0.1023	-0.000071

V_R = Reference Volume — V_1 = Volume at initial date
V_i = Volume at given date order i — N_i = Number of days between two successive dates
V_{i+1} = Volume at next date order — N_c = Number of days at given date since initial date

Table 2 Kwinte Bank. Observation dates and volumetric parameters for reference transverse rH00.

different representative transverses, all above the same reference level and taken at the same dates. Such data can be worked out for total or for partial unit-volumes. Working it out for specific sections of the bank transverses it is even possible to approach shape and position changes.

Figure 9 shows such a time series of total unit-volumes above level -15 for several transverses on the Kwinte Bank during the period 1982-1985, while the bank was under sand extraction.

The mere volumetric change between the beginning and the end of a period, even after normalisation, no more provides sufficient indications for an evolution trend. Determination of evolution trends by time series of unit-volumes or other volumetric parameters as well encounters the problem that the volumetric data often show a high variability from one date to another, shorter terms giving mostly important variations. That short term variability can partly be explained by short term sediment migrations, whose residual effects are mildered by fast alternation of phases of intense deposition and erosion or it can possibly be caused by real deviations from the reference transverses. Moreover there are the differences in length and dates of the intervals which hardly can be anticipated by chronological averaging.

To get a numerical expression for the evolution trend of the present day short term residual morphological change and residual sediment mobility along different transverses, and which at the same time allows geographical comparison, ΔF_c/time graphics are drawn for each of the total or partly reference transverses above a same reference level and at identical successive date.

Figure 10 shows the ΔF_c/time graphics for the total cross-section above level -15 for 4 different transverses on the Kwinte Bank over tne period 1983-1986. These graphics illustrate distinctly the existence of different evolution types along the Kwinte Bank.

Bank : KWINTE BANK Reference transverse : rG22				Initial date : 30-11-82 Reference level : -15 m			V Reference : 1988 m^3	
Date order	Date	N_i	N_c	V_i	$\Delta V_i = V_{i+1}-V_i$	$\Delta V_c = V_i-V_1$	$\Delta F_c = \frac{\Delta V_c}{V_R}$	$\Delta D_c = \frac{\Delta F_c}{N_c}$
1	30-11-82			1320				
2	07-05-83	158	158					
3	10-06-83	34	192	1568	+248	+248	+0.1247	+0.000649
4	13-06-83	3	195					
5	12-07-83	29	224	1973	+405	+653	+0.3285	+0.001466
6	07-09-83	57	281	1709	-264	+389	+0.1957	+0.000696
7	22-08-84	340	621	2333	+624	+1013	+0.5096	+0.000821
8	28-05-85	279	900					
9	06-06-85	9	909					
10	11-09-85	97	1006					
11	13-09-85	2	1008	2274	-59	+954	+0.4799	+0.000476
12	28-11-85	76	1084	2657	+383	+1337	+0.6725	+0.000620
13	11-02-86	75	1159	2484	-173	+1164	+0.5855	+0.000505
14	11-11-86	273	1432	2183	-301	+863	+0.4341	+0.000303
V_R = Reference Volume V_i = Volume at given date order i V_{i+1} = Volume at next date order					V_1 = Volume at initial date N_i = Number of days between two successive dates N_c = Number of days at given date since initial date			

Table 3 Kwinte Bank. Observation dates and volumetric parameters for reference transverse rG22.

Using the graphic distribution a correlation line can be inferred.

Using this correlation line a trend for the residual mobility index, hence for the annual mean relative residual unit-volumetric change, can be deduced (ΔF_c = R in % of V_R). As it corresponds to a percentage of the reference volume it expresses morphological change.

Knowing the specific reference volume for each reference transverse and reference level, the trend of the relative annual mean residual unit-volumetric change for the given transverse and level can be recalculated into a trend for the absolute annual mean (A in m^3/year) over the considered period. This value expresses more directly the amount of residual sediment displacement.

Table 4 shows the total mean annual R and A values for several reference transverses on the Kwinte Bank over the period 1983-1986, and for the bank above level -19.

Nothing of course is provisionally known about the future evolution of the evolution trend itself. Therefore longer observation is needed, especially in order to detect a possible cyclic character.

Figure 11 shows distinctly the geographical differentiation of present day short term residual morphodynamics and sediment dynamics along the Kwinte Bank. The northern edge shows a trend of low sediment loss, but due to the low available sand volume its morphological repercussion is not important. The northern part of the bank is characterised by a trend of important sediment gain which increases southward and which corresponds to important morphological changes as the available sand volumes still are not very important. This sediment gain strikes especially as this part of the bank mainly supports the sand extraction and it suggests sand convergence and possibly uppiling.

The central part of the bank shows longitudinal alternations of important sand gains and losses over distances of less than 1 NM. They correspond to relatively less important morphological changes as the available sand volumes in this central part are quite important. Moreover the geographical differentiation suggests longitudinal sand exchanges between the central and the northern part and a possible longitudinal wavelike sand migration mechanism in the central part, despite its flat general morphology (figure 2). The important terrace occurring more to the south (figure 2a) suggests that the southernmost part of the western bank side is the main zone of residual erosion.

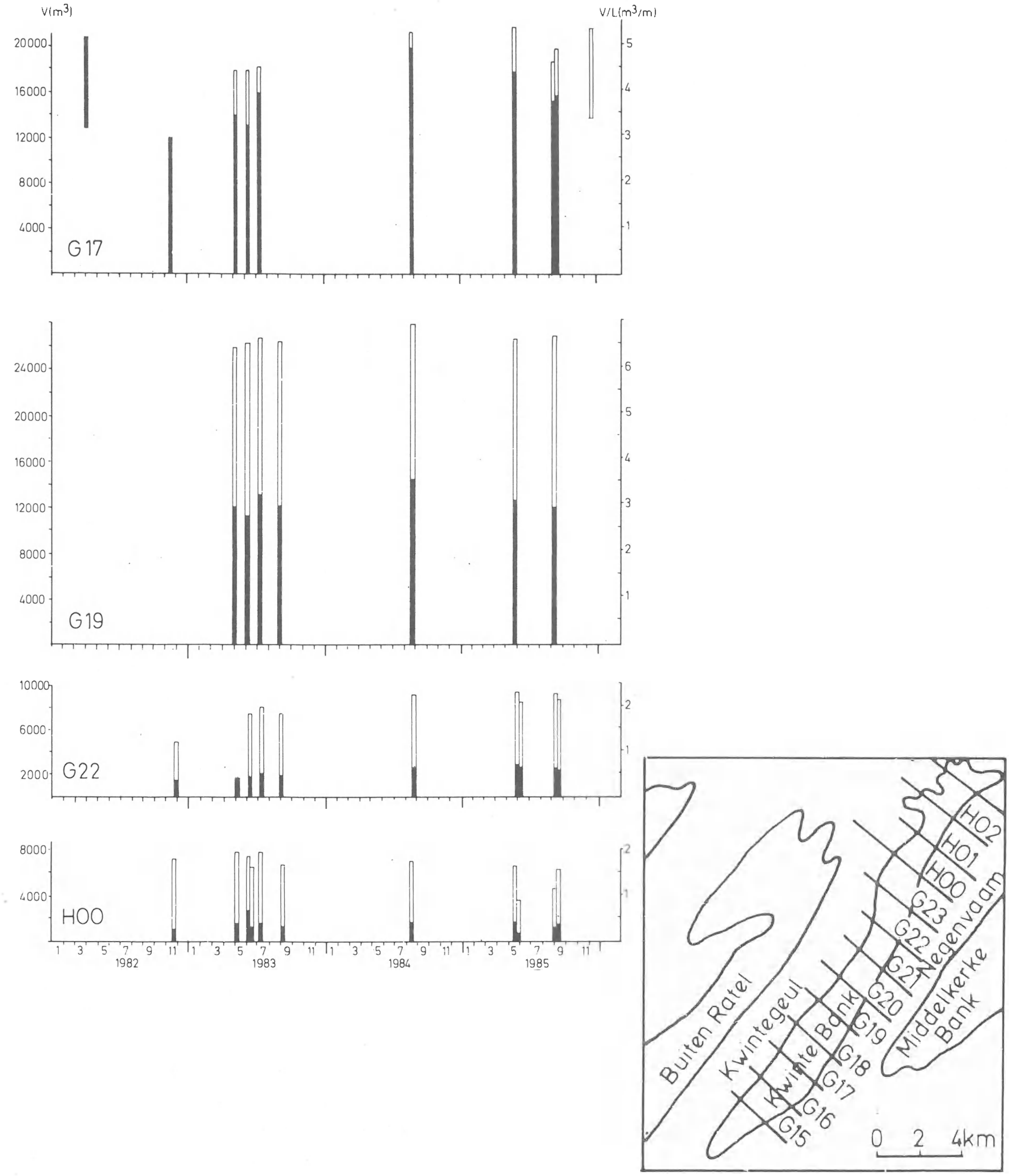


Figure 9 Time-series of unit-volumes above reference level -15 (MLLWS) along several reference transverses on the Kwinte Bank between November 1982 and August 1985. V = unit-volume along the given transverse. V/L = unit-volume per metre planimetric transversal length at the reference level -15.

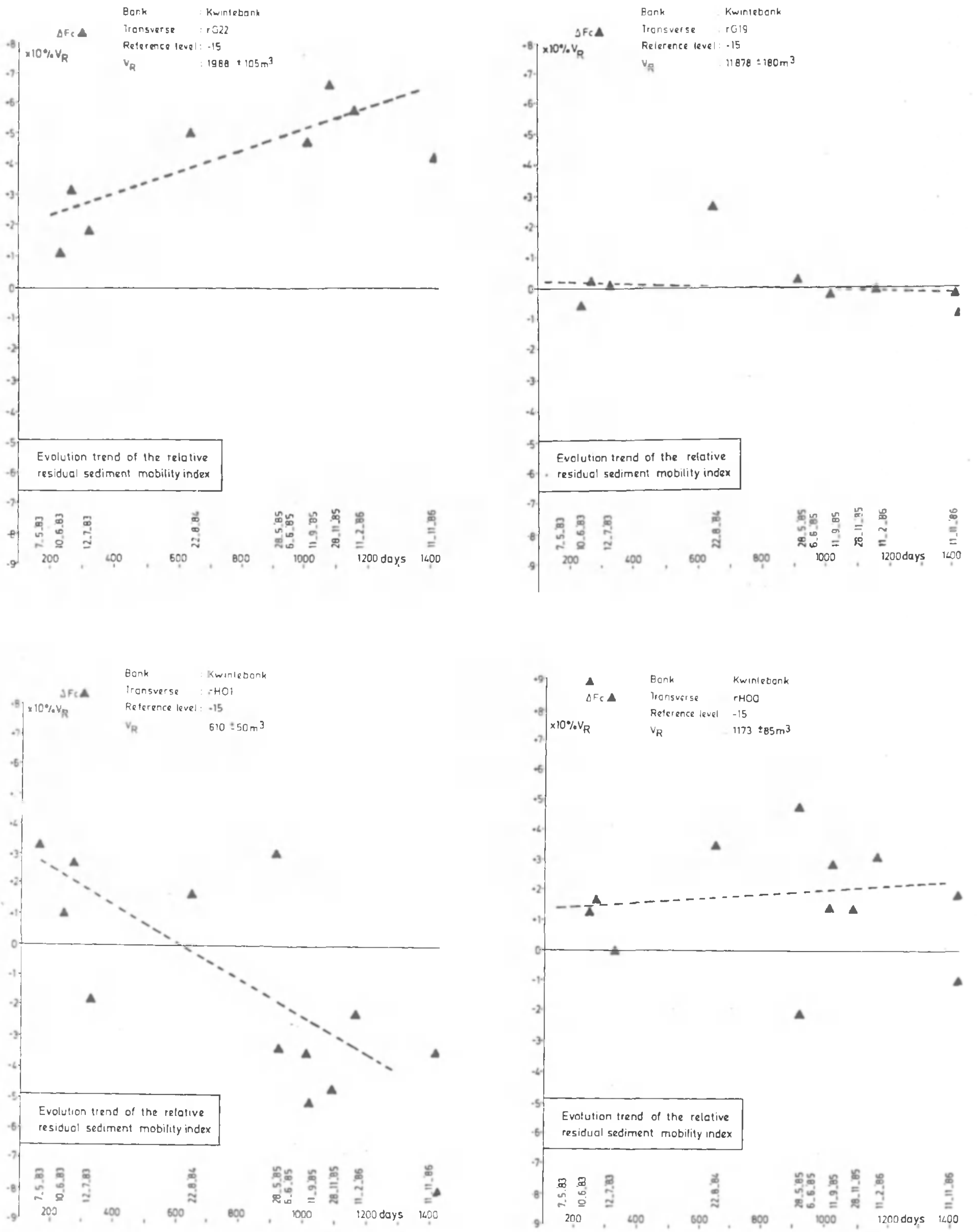


Figure 10 Kwinte Bank. Evolution trend of the relative residual sediment mobility index along different reference transverses for the period November 1982 - November 1986.

Bank : KWINTE BANK Reference level : Volume above level -19. Total bank. Period : 1983-1986				
Along Reference Transverse	Mean bank length at given level (m)	Reference Volume (m^3)	Annual mean relative unit-volumetric change (% of V_R/year)	Annual mean absolute unit-volumetric change (m^3/year)
	L_B	V_R	R	A
rH01		3569	-1.4 %	-50
rH00		5308	+2.8 %	+148
rG23		8621	+2.1 %	+181
rG22		6859	+5.1 %	+349
rG21		10871	-5.0 %	-544
rG20		13027	+1.3 %	+169
rG19		19417	-1.4 %	-272
rG18				
rG17				
rG16				

Table 4 Kwinte Bank. Residual unit-volumetric evolution trend for the total bank along different reference transverses.

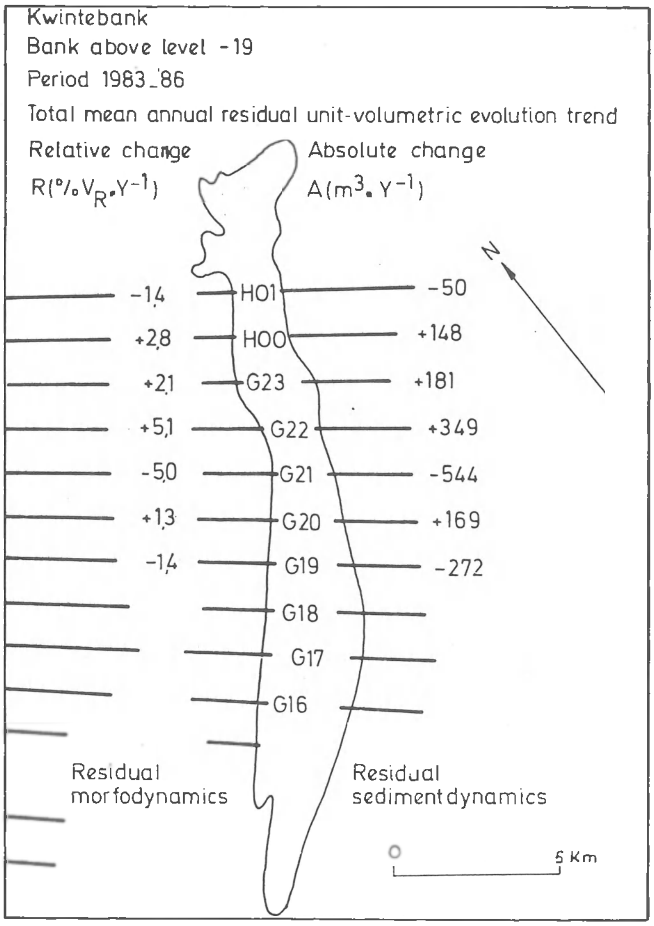


Figure 11 Kwinte Bank. Geographical differentiation of the total annual mean residual unit-volumetric evolution trend above level -19 (MLLWS) along several reference transverses on the Kwinte Bank for the period November 1982 - November 1986.

2.5. Evaluation of shape and position changes and of sediment exchanges

Changes of total unit-volume above a chosen reference level along a reference transverse yield an indication for breakdown or for building up of the bank above that level, along the chosen transverse and for the given period. Moreover comparison of the total unit-volumetric evolution in successive transverses along the bank shows geographical differences in that evolution. Moreover it allows a first assessment of sand mobility by possible longitudinal sand migrations along the bank, or possible preferential transversal sand supplies, sand losses or sand passages. The total unit-volumetric changes however do not show where and how volumetric changes affect the bank's shape, height and location. A bank can migrate while keeping a constant volume. Total volumetric changes do not allow to indicate what side knows residual deposition or residual erosion, neither how bank asymmetry develops, nor what is the role of bedform displacements. They do not indicate if growth occurs by heightening or widening, neither if decline is caused by lowering or by narrowing, nor what role plays arrival, growth, decline or disappearance of large bedforms. They give no information about the specific evolution on different hypsographical or morphographical parts of the bank, especially on the differences in evolution of the bank's top zone, its sides, its foot zone, neither what possible sediment exchanges could exist vertically along the transverses or horizontally at different levels.

Residual changes in shape and in location and related residual sediment displacements can be approached on two ways :

1. Visual comparison of sequential depth profiles by their superposition, after they were made comparable by scale equalization and drawing of reference location and reference level marks. The former correspond to cross points of the reference transverse with hyperbolic coordinate lines or with reference meridians or reference parallels. Reference levels are parallel with the zero datum level.

   Figure 12 shows an example of such superposition of two successive depth profiles along the reference transverse rG22 on the Kwinte Bank between November 1982 and July 1983. There is hardly any location change of the bank over that period in this station, but there are strong changes in shape, due to the migration of large sand waves, especially from the western top convexity toward the slight eastern slope, where westwards upslope climbing sand waves are encountered. This is a first clear indication for a bank maintenance mechanism by sand uppiling.

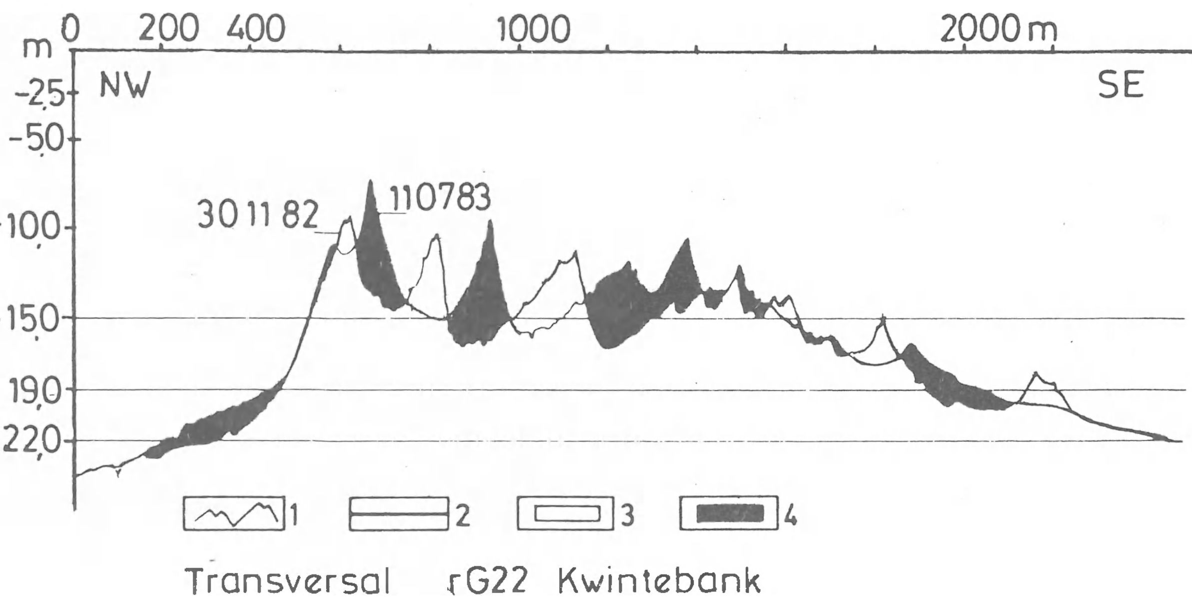


Figure 12 Kwinte Bank. Visual comparison of sequential depth profiles along reference transverse rG22 on the Kwinte Bank.
1. Profile. 2. Reference levels. 3. Loss. 4. Gain.

Bank : KWINTE BANK Reference level : Volume above level -15. TOPZONE bank sides. Period : 1983-1986				
Along Reference Transverse	Mean bank length at given level (m)	Reference Volume (m^3)	Annual mean relative unit-volumetric change (% of V_R/year)	Annual mean absolute unit-volumetric change (m^3/year)
	L_B	V_R	R	A
rH01	500	610 ±50	-22.0 %	-134
rH00	850	1173 ±85	+2.3 %	+33
rG23	950	1858 ±95	+8.5 %	+158
rG22	1050	1988 ±105	+12.0 %	+238
rG21	1250	5129 ±125	-8.5 %	-438
rG20	1390	7000 ±139	+6.0 %	+420
rG19	1760	11878 ±179	-1.4 %	-166
rG18				
rG17	2950	14518 ±295	-2.0 %	-290
rG16	2100	9256 ±210		

Table 5 Kwinte Bank. Residual unit-volumetric evolution trend for the bank's topzone along different reference transverses.

Bank : KWINTE BANK Reference level : Volume between -15 and -19. LOWER bank sides. Period : 1983-1986				
Along Reference Transverse	Mean bank length at given level (m)	Reference Volume (m^3)	Annual mean relative unit-volumetric change (% of V_R/year)	Annual mean absolute unit-volumetric change (m^3/year)
	L_B	V_R	R	A
rH01		2959	+2.8 %	+84
rH00		4235	+2.7 %	+115
rG23		6763	+0.3 %	+23
rG22		4871	+2.2 %	+111
rG21		5742	-2.0 %	-113
rG20		6027	-4.2 %	-251
rG19		7539	-1.4 %	-106
rG18				

Table 6 Kwinte Bank. Residual unit-volumetric evolution trend for the lower bank sides along different reference transverses.

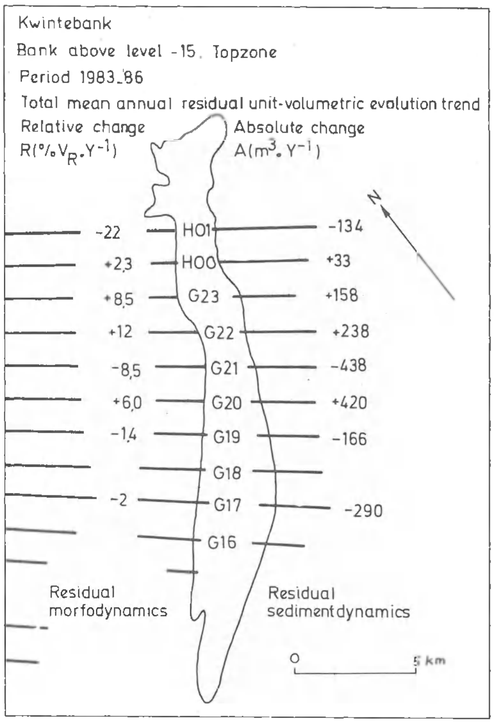


Figure 13 Kwinte Bank. Geographical differentiation of the annual mean residual unit-volumetric trend on the top zone (above level -15) along several reference transverses for the period November 1982 - November 1986.

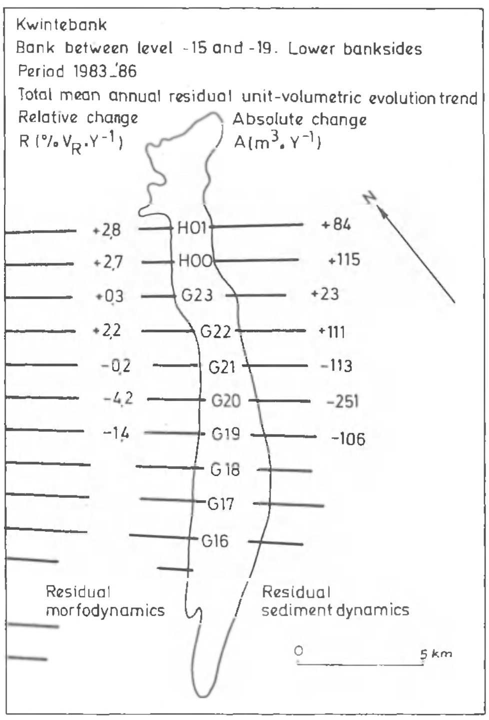


Figure 14 Kwinte Bank. Geographical differentiation of the annual mean residual unit-volumetric trend on the lower bank sides (between level -15 and -19) along several reference transverses for the period November 1982 - November 1986.

2. Monitoring of partial unit-volumetric change of morphological sections, of slices between two successive reference levels, especially if they are computed separately for both bank halves, or of columns, along a reference transverse may provide a numerical approach towards an evaluation of the shape evolution and the location changes of the bank, giving simultaneously indications about the sediment migrations involved.

   The slice method has been used for the different transverses on the Kwinte Bank merely for the total top zone above -15 and for the total lower bank part between levels -15 and -19, but without differentiation of both bank sides.

   Table 5 shows the R and A values for the top zone above level -15, along several reference transverses on the Kwinte Bank during the period 1983-1986.

   Table 6 gives similar information for the lower bank sides between levels -15 and -19.

   Figure 13 shows the geographical differentiation in the present day short term morphodynamics and sediment dynamics along the top zone (above -15) of the Kwinte Bank. Figure 14 provides similar information for the lower bank sides (between -15 and -19).

   During the period 1983-1986 there was a trend towards much more important total residual sediment dynamics in the top zone of the Kwinte Bank than on the lower bank sides, with the exception of the offshore end of the northern part (rH00).

   Residual sediment loss was relatively very important in the top zone on the northern edge, while there the lower bank sides knew accretion.

   The trend of the residual sediment dynamics in the northern part (between rH00 and rG22) points to important sand gains in the top zone, increasing southwards, and to important sand gains on the lower bank sides as well, although there the gain is most important on the offshore side. Of course there is no differentiation between both bank sides. Sand extraction in this northern part is at the present not leading to interruption of the local bank growth process, but possibly lowers the natural growth rate.

   In the central part, the residual sand mobility in the top zone is very high and alternatively positive and negative. On the lower bank sides it is still important but indicates continuous residual sand losses. The question is if this lower part contributes to sand supply on the northern part of the bank and if there is a vertical sand exchange with the top zone and the bank side foot.

   In the southern part there is a present day short term trend towards sand losses, at least in the top zone. No quantitative data are available for the lower bank sides. Bathymetric profiles (cf. figure 2a) in this southern part suggest important residual erosion on the higher parts of the west side and a rather stable situation on that lower bank side. It is however not clear whether the terrace which occurs there on the lower bank side is due to a vertical differentiation of the sediment dynamics, or to differential erosion because of a more resistant bank core.

2.6. Areal side scan sonar profiling, elaboration of bedform maps and residual bed load transportation pathways

Residual bottom current diagnostic characteristics of bedforms, especially strike and direction of steep progradation slopes of megaripples and sand waves are used to map the residual bed load transportation pathways on a geomorphological base.

Typology, strike and steep side direction of the bedforms have been mapped on the base of areal Klein side scan sonar registrations, sailed partly along the reference transverses, partly along other tracks, because bedforms are best registered on strike parallel courses as they then give maximal reflections.

All registrations have been obtained using a 100 kHz frequency signal emission with a transducer towed at 5 m depth at about 15 m on the starboard side of the ship and about 40 m behind the positioning antenne, and generally with slant ranges of 75 or 100 m and at a ship's bottom speed of 4 kts. All have been provided

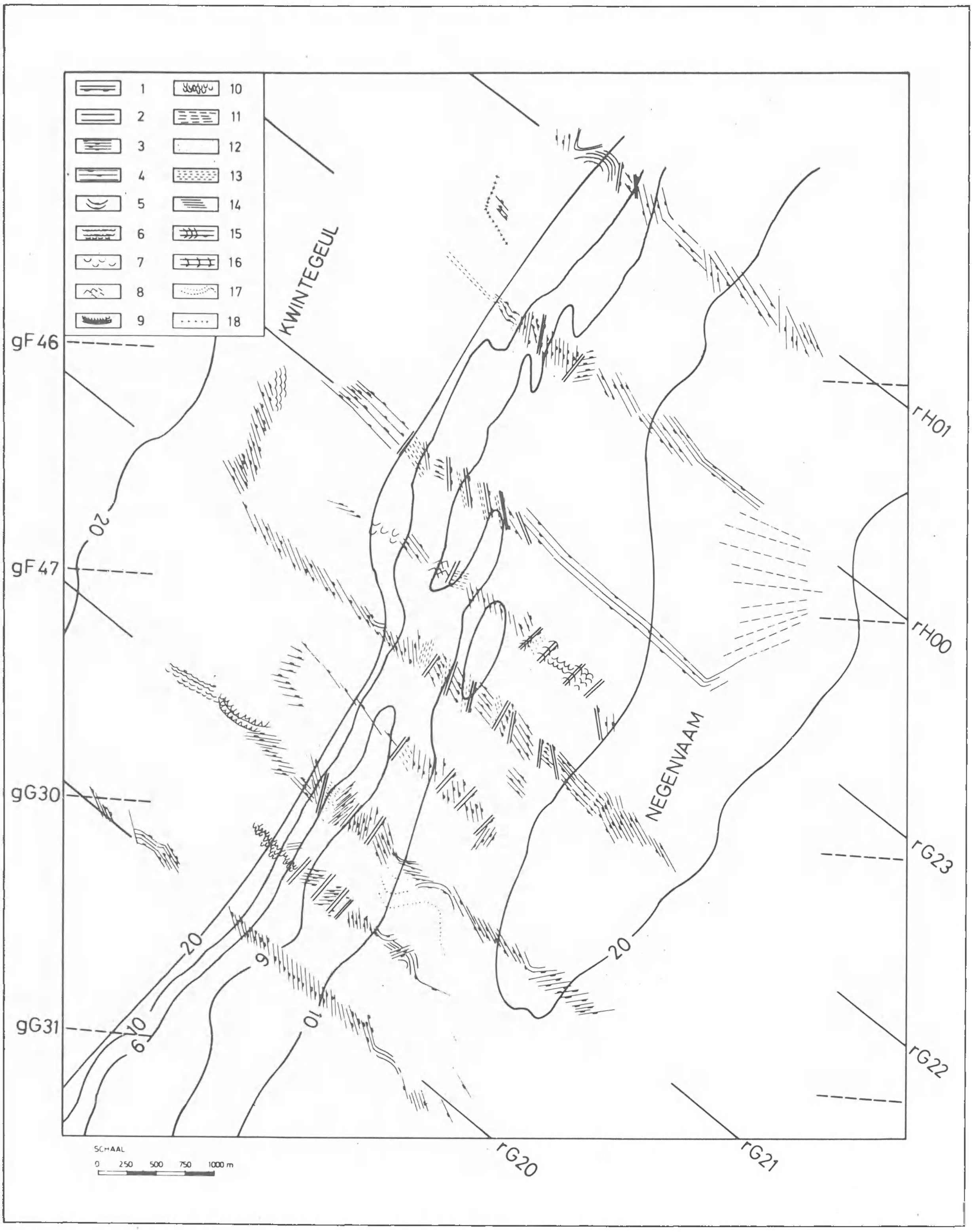


Figure 15 Kwinte Bank, 30 November 1982. Map of surficial bedform characteristics inferred from side scan sonar records along several reference transverses. 1 = asymmetric sand wave, steep side indicated. 2 = asymmetric sand wave. 3-4 = asymmetric linear megaripple. 5 = parabolic megaripple. 6 = asymmetric sinuous megaripple. 7 = barchanoid megaripple. 8 = rhomboid megaripple. 9-10 = linguoid megaripple. 11-12 = field with medium (11) or small (12) ripples. 13 = possible megaripples. 14 = symmetric small ripples. 15-16 = sand waves with megaripples or ripples. 17 = possible sinuous megaripples. 18 = sand dredging spur.

with 1 minute position marks easying cartographic elaboration, distorsion corrections and possibly length scale corrections. Strike-distorsions because of ship's speed have been corrected using the Flemming-method (FLEMMING, 1982). The resulting morphographic bedform map is transformed into a residual bed load transportation pathway map using the current diagnostic bedform characteristics.

Figure 3 provides an example of a side scan sonar registration on the northwestern side of the Kwinte Bank.

Figure 15 shows a map of bedform characteristics inferred from side scan sonar registrations sailed transversally at distances of half a Decca-lane (0.5 NM) over the central and the northern part of the Kwinte Bank on 30 November 1982.

Megaripples and sand waves run more or less transversally in the central zone of the Kwinte channel and Negenvaam and move residually in opposite directions, offshore in the eastern part of the Kwinte, landwards in the westward part of the Negenvaam. This corresponds to the residual bottom current pattern advanced for the current pattern along some of the East Anglian Banks by CASTON and STRIDE (1982).

On the lower bank sides and higher up the slopes, lagging behind, they gradually change their orientation from bank transversal to more and more bank parallel. Moving upslope towards the top zone, but in opposite direction on both sides of the bank, their crests become more and more parallel to the crest of the bank itself while climbing the bank sides.

On the top zone of the northern and northern central part, especially on the western side, they form large asymmetric sand waves. Having their steep progradation side directed towards the bank centre, these sand waves indicate a residual sand movement towards a convergence axis, situated near the eastern top convexity. There symmetric sand waves occur locally, indicating by their position a dominancy of sand supply by sand waves from the west and a certain equilibrium position in the residual sand supply from both sides.

In the central part, where the summital zone is less deep, a flat surface develops with smaller bedforms probably in relation with differences in energetic level and impact of waves and tidal currents.

The occurrence of large sand waves seems not restricted to zones with important residual sand supply but perhaps to areas with higher sand mobility. Only a more detailed mapping of bedform characteristics and one of unit-volumetric evolution trends would possibly allow to answer the question of the relationship between sand waves on one side and bruto and residual sediment dynamics on the other.

Figure 16 provides a map of the residual bottom currents, corresponding with residual sediment pathways.

The analysis of the sediment pathways lets little doubt about the existence of an uppiling mechanism that in a more or less discontinuous and varying way provides maintenance to the bank, although it seems not to be the only mechanism of sediment transportation on the bank, and certainly is counteracted as well by erosional phases.

The residual bottom current direction in the Kwinte channel and in the Negenvaam channel, both adjacent to the Kwinte Bank, suggest that the first channel is flood dominated and knows a seaward loss of fines, while the second channel is eb dominated. This explains the uppiling from both sides, the formation of the threshold at the landward side of the Negenvaam and the finer character of the sediments in this less energetic eb dominated swale.

This residual sediment transportation pathway model differs from the eb dune interpretation of the northern Flemish Banks (VAN VEEN, 1935).

As to the origin of the gradual direction change of currents and bedforms along the bank side, it can only partly be related to Coriolis effects while refraction due to friction can not be omitted and phase differentiation of the tidal wave on both sides of the bank might induce level differences.

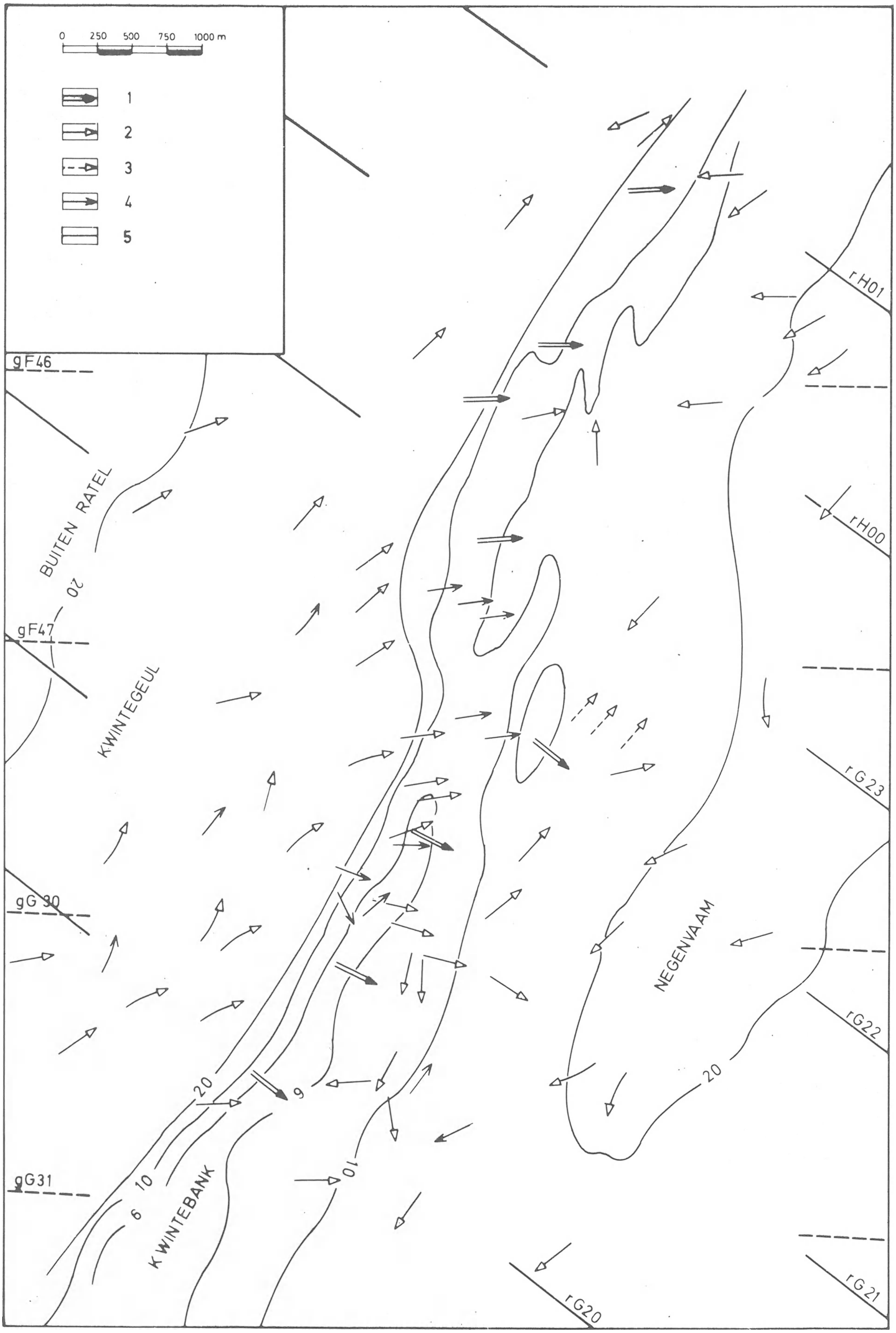


Figure 16 Kwinte Bank, 30 November 1982. Direction and intensity of residual bottom currents around the Kwinte Bank, and corresponding residual bottom load transportation pathways as indicated by asymmetry, strike and type of bedforms, inferred from side scan sonar registrations. 1 = residual sand migration by sand waves. 2 = residual sand migration by linear megaripples. 3 = residual sand migration by linguoid and parabolic megaripples. 4 = residual sand migration by megaripples upon slight sand wave slopes (cf. figure 3). 5 = contour lines.

CONCLUSIONS

1. A detailed quantitative monitoring of present day short term residual morphodynamics and sediment dynamics - especially changes in volume, shape and location of sand banks on a geomorphological basis becomes now possible. Qualitative approach by visual comparison remains necessary.

   Both approaches however put strict prerequisites on accurate navigation and reliable high frequency positioning ; on high resolution bathymetric recording and its very accurate elaboration unto comparable profiles ; on areal observation of bottom topography, especially of bedforms ; on computer facilities on board and in the laboratory.

2. The present day morphodynamics and sediment dynamics - at least on the Kwinte Bank - are quite complex.

   There are different types of evolution trends varying not only from one place to another along the bank (over distances in the 1 NM range) but as well in the different hypsographical zones of the bank. Possible asymmetry between both sides has not yet been investigated by the quantitative approach.

   There are places where the whole of the top zone knows accretion while the lower bank sides are declining, others with a contrary evolution and still others show a similar evolution. Hence that the intensity of the evolution for total bank transverses can be lower than for parts of them.

   Chronologically the period 1983-1986 shows at least 4 types of evolution trends on the Kwinte Bank :

   1. rH01 type : high medium residual mobility index ΔF_c and low reference volume. Medium sediment loss but strong residual morphological changes by erosion, as the total available volume is small.
   2. rG22 type : medium reference volume and medium ΔF_c Relatively high volumetric increase and important morphological effects.
   3. rH00 type : low ΔF_c and medium reference volume. Slow volumetric increase and low morphodynamics.
   4. rG19 type : low ΔF_c but important available sand volume and large dimensions. Slow volumetric decrease but trend towards important residual sediment loss.

   These trends can not be used for long term prediction, as the evolution of the evolution trend itself is unknown, especially because of the probable cyclic and interrelated character. These trends should not be projected to other areas.

   Geographically, during the period 1983-1986, the Kwinte Bank was characterized by different dynamical zones :

   1. **the northern edge** with intense residual breakdown in the top zone above -15 but relatively important accretion (the highest relative one of the whole bank) on the lower bank sides. This can partly be due to sand redeposition after reworking by waves from the top zone.

   2. **the northern part** (between rH00 and rG22) is morphologically relatively low but sand mobility is shown by sand wave migration. The top above -15 shows a trend to accretion, southward gradually increasing to high values especially around the transverse rG22. On the lower bank sides the evolution trend is slightly positive, but accretion rate increases to the north, to join the northern edge at that level.

The top zone accretion is remarkable as most of the sand dredging is done on that section. This certainly proves the existence of a process of maintenance due to a residual sand supply. Apparently the direct supply is mainly caused by uppiling along both sides of the bank, as indicated by the bedform mapping, and probably shows an eastward dominancy due to a residual crest overrunning from the west. The general sediment dynamics pattern on the lower bank sides and in the central bank part however suggests a further supply by a wavelike longitudinal sand transport and possibly the existence of transversal sediment transportation pathways. The general sediment dynamics pattern, the bedform pattern and the bank morphology indicate that the sands are at least partly supplied by erosion on the southwestern side of the bank, partly provided by internal reworking within the bank itself, such as due to transversal or longitudinal furrowing and to wave truncation, and that they are partly of external origin.

3. **the central part** which is a higher, flatter and more voluminous part of the bank. Between rG21 and rG19 it knows complex dynamics. The evolution trend in the top zone varies largely at short distances (1 NM) between relatively important losses (-8.5 % V_R/year) and moderate gains (+ 6 % V_R/year), pointing at a longitudinal wavelike sediment passage or possibly at some transversally preferential sand migration.

   The morphodynamics are less important than the sediment migration, because of the greater available sand volume. Remarkable is that here the lower parts show an inversed evolution trend, what could indicate unequal vertical exchanges.

4. **the southern part** (rG18 to rG15) is still massive and shows some sand waves. Its western side presents erosional terrace features on the lower bank sides, while the eastern side leans against the Negenvaam treshold. Top and bank sides present a trend of moderate unit-volumetric decrease (1 to 2.5 % V_R/year). The low short term morphological change corresponds to important sand migration. On longer term the erosion however seems to affect distinctly the morphology of the top zone of the western side by erosion (with the formation of a terracelike feature), while the eastern side seems to know accretion as indicated by the threshold crossing the Negenvaam. It is however unknown if the terracelike feature is due to vertical difference in the dynamics, or forms a structural feature due to local presence of a more resistant core in the lower part of the bank.

3. There are different **mechanisms of sand transportation** :

   1. **Residual sand uppiling** on both sides from the lower parts and from the swales by bank side deflection of dominant eb or flood bottom currents, the Kwinte side being flood dominant, the Negenvaam side being eb dominant. That residual sand migration does not always eliminate local residual erosion and it interacts with other mechanisms of sand transportation. Therefore the migrating sands are partly reworked from the bank by the uppiling itself or supplied after intrabank transportation processes, and partly supplied by the channel bottom currents themselves, being then of an external origin or supplied primarily by the bank erosion itself.

   2. **Internal longitudinal** wavelike sand transport as indicated by the sediment dynamics pattern in the central part but as well by occasional erosion furrows.

   3. **Internal vertical** sand exchanges between top zone and lower sides probably by wave truncation alternating with renewed uppiling.

4. On longer terms the present day short term sediment dynamics pattern seems to indicate a geographical alternation of periods of local upbuilding (by heightening and probably asymmetric broadening) of bank parts using sand simultaneously, and at least partly, eroded from other bank parts, as shown by the erosive terrace on the southwestern side of the bank, and transported at least partly by intrabank mechanisms. Therefore we assume that locally sand deposits within the bank can greatly differ in age laterally at short distances.

5. Moreover there are geomorphological as well as internal sediment-structural characteristics indicating that the steep western side of the Kwinte Bank is one of present day residual erosion, despite intermittent deposition and sediment migration, and that the slightly sloping eastern side of the Kwinte Bank is one of a present day residual accretion.

6. This suggests a shifting and probably oscillation of the bank or of bank parts around its axis within a multidecade periode. Such medium term processes, elements of a longer term stability of the bank have been advanced on the base of comparisons of sequential hydrographical maps (VAN CAUWENBERGHE, 1975) or of inferred bathymetric profiles (DE MOOR, 1985).

Longitudinal differences in advancement of the oscillation mechanism and in the intrabank sediment shifting explain the occurrence of kinks and the possible inversion of transversal asymmetry on both sides of the kink.

7. The present day maintenance of the bank results from a very complex dynamic equilibrium comprising variations in position, in transversal and in longitudinal shape of the bank.

Reconnaissance of the internal sedimentary structures by high resolution ORE-subbottom profiling along the reference transversals on the Kwinte Bank in 1982-1983 (DE MOOR, 1983 ; 1985 ; 1986), reveals that, with the exception of the eastern side, most of the bank shows a subhorizontal makroplanar structure with internal diagonal progradation structures (figure 4), sometimes alternatively directed to one side and to the other one, and that sometimes the interfaces are not plane but present a buried sand wave aspect (figure 17).

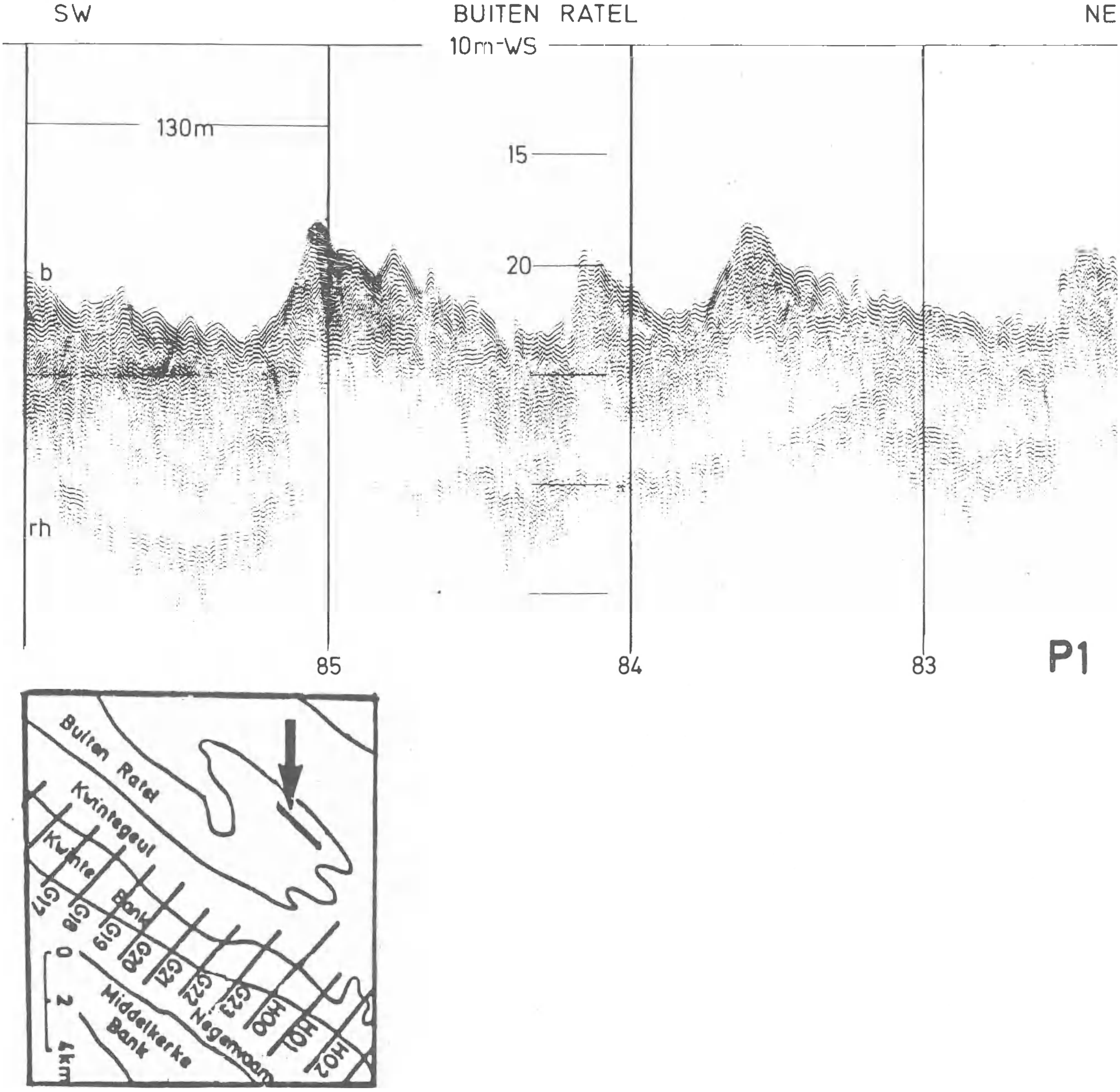


Figure 17 Buiten Ratel Bank. ORE Subbottom profile on the northwest side of the bank between reference transverse rG20 and rG22 on 2 November 1982. b = sea floor. rh = buried sand wave surface. -WS = below sea surface.

Hence it is assumed that the process of uppiling supporting the present day maintenance has also been acting during the upbuilding of the bank during the Holocene sea level rise because at each step of sea level rise, sands have been residually uppiled unto the new and higher wave base.

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18 Dynamics of gravel in the superficial sediments of the Flemish Banks, southern North Sea

J. TYTGAT

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

ABSTRACT

Although it is obvious that the gravels in the Flemish Banks consist of three main deposits, it is very hard to stake them out. Due to the Holocene marine transgressions the gravels are becoming a mixture of southern coming, Rhine and Meuse, and local clasts. In consequence of the tidal stream generated erosion, a coarse gravelly lag deposit develops in the interbank channels.

Both grain-size and petrographic data demonstrate gravel paths leaving these channels. They indicate a net NNE transport of fine gravel oblique to the crest of the ridges.

INTRODUCTION

Gravel occurrences in the Belgian part of the North Sea seem to have been first noted by VAN BENEDEN (1862). Since then, many authors have reported gravels coupled to a large diversity of interpretations. It is the purpose of this paper to describe and interpret some textural and petrographic aspects of the superficial sediments of the Flemish Banks, especially to detail the coarse fractions.

The technique employed during the investigation, consists of sampling the area using a Van Veen-type sampler. A first reconnaissance is followed by a closer spacing in areas where at least 1 % gravel occurs. These samples are analysed on their grain size distributions, gravel lithology, carbonate content, macro- and microfauna.

All sample stations are positioned by means of Decca-navigation (figure 1).

DISTRIBUTION OF COARSE SEDIMENTS IN THE FLEMISH BANKS

The isoconcentration contours of the gravel content trend approximately parallel to the bathymetric contours. The highest concentrations (up to 50 % and more) of coarse sediments (shells included) occur in the interbank channels or swales (figure 2). These percentages decrease over the slopes to the top of the ridges.

This general pattern produces some divergences. On top of the Kwinte Bank and Buiten Ratel nearly no gravel at all occurs. Both northern parts however, contain fields with coarse sediments up to 16 % and more. These fields make part of a tongue and ribbon formy pattern leaving the swales (figure 2). The highest concentrations (> 32 %) in this configuration occur in the channels down the steeper NW sides of the ridges. They decrease in a NNE direction, slightly oblique to the crest of the sand ridges.

Decalcification of the samples does not give any important relative changes of concentration.

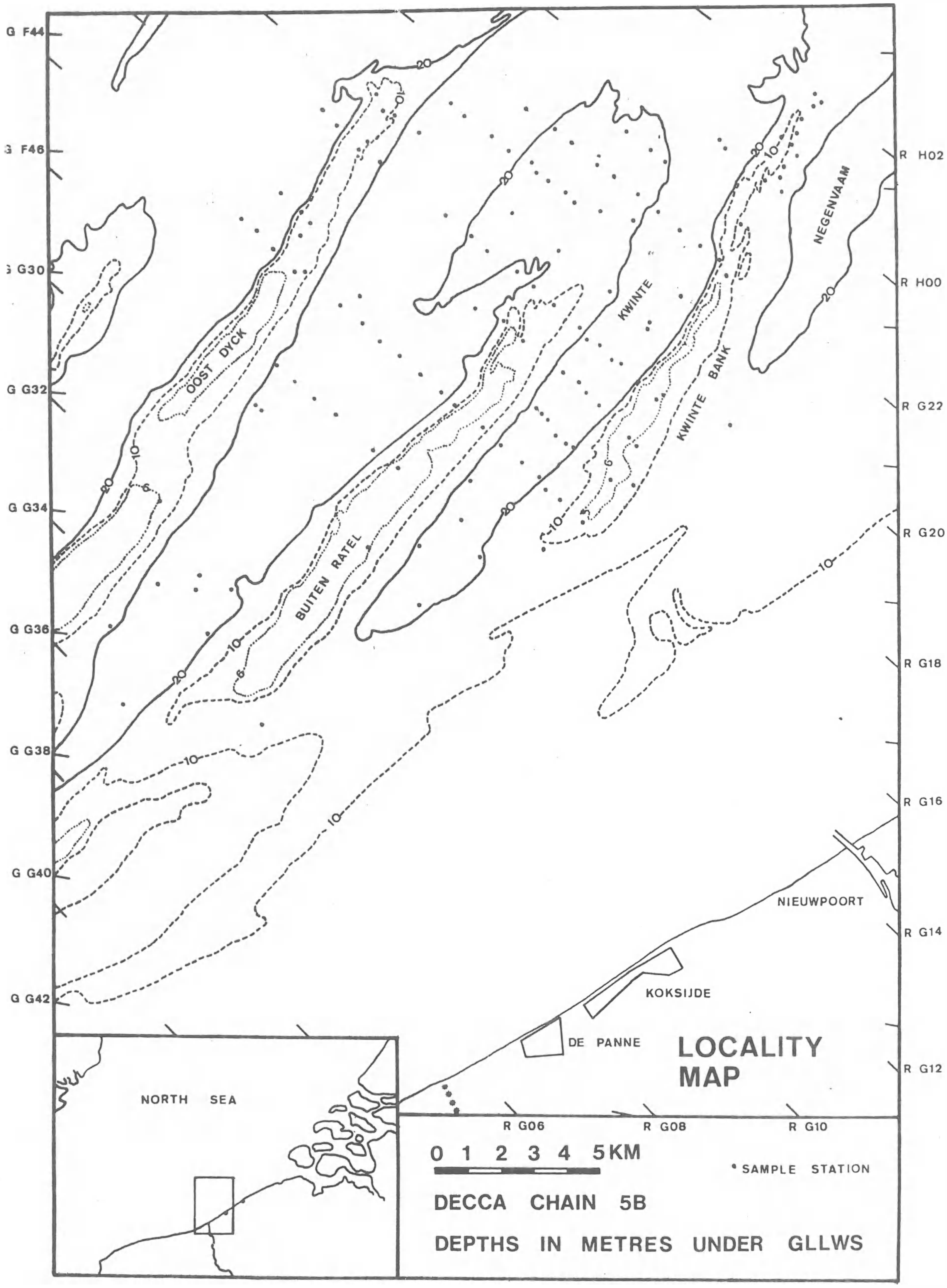

Figure 1 Linear sand banks of the Flemish Banks area. Location of sample stations.

Looking closer to the maximum grain size, the general view of the bathymetric contours is also reflected in these data (figure 3). The coarsest sediments (> 32 mm) lie in the channels down the NW slopes. It is remarkable that the sediments on the ridges, containing less than 1 % coarse sediments, can have maximum grain sizes of 16 mm and even more. This at first sight, however apparent, contradiction is explained by the presence of shells. These are, in proportion to their sizes, very light. The ribbon and tongue formy pattern still exists, although that one on top of the Kwinte Bank is not very obvious. Only a few spots indicate a coarser sediment. These patterns become accentuated after having picked out the shells. This means that the maximum sizes of the sediments on the ridges are those of the shells and shell debris. The maximum sizes in the channels and in the tongue and ribbon pattern are determined by the sizes of the lithic gravel.

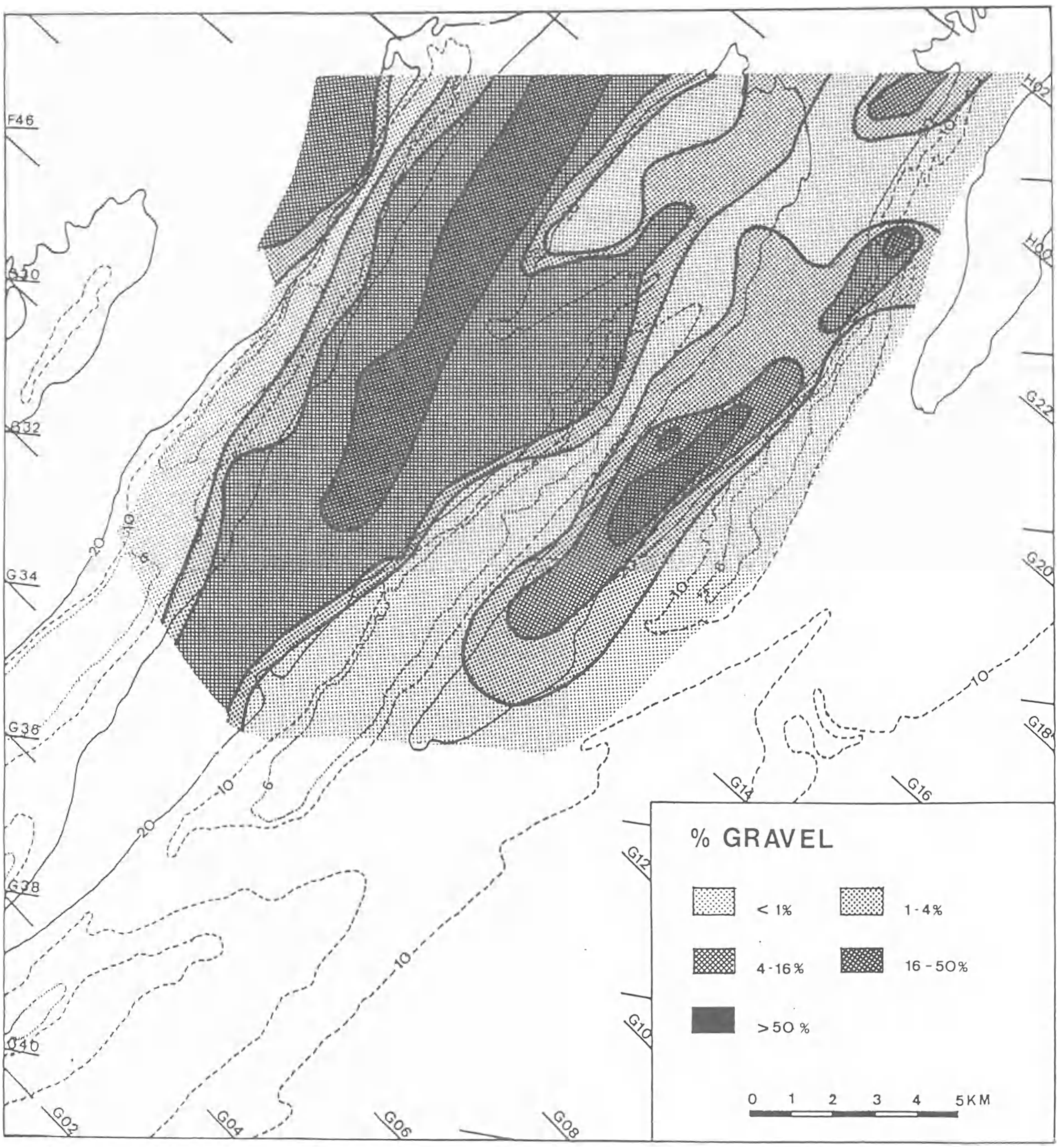

Figure 2 Distribution of the gravel in the Flemish Banks. Percentages are calculated for the total weight of the sample.

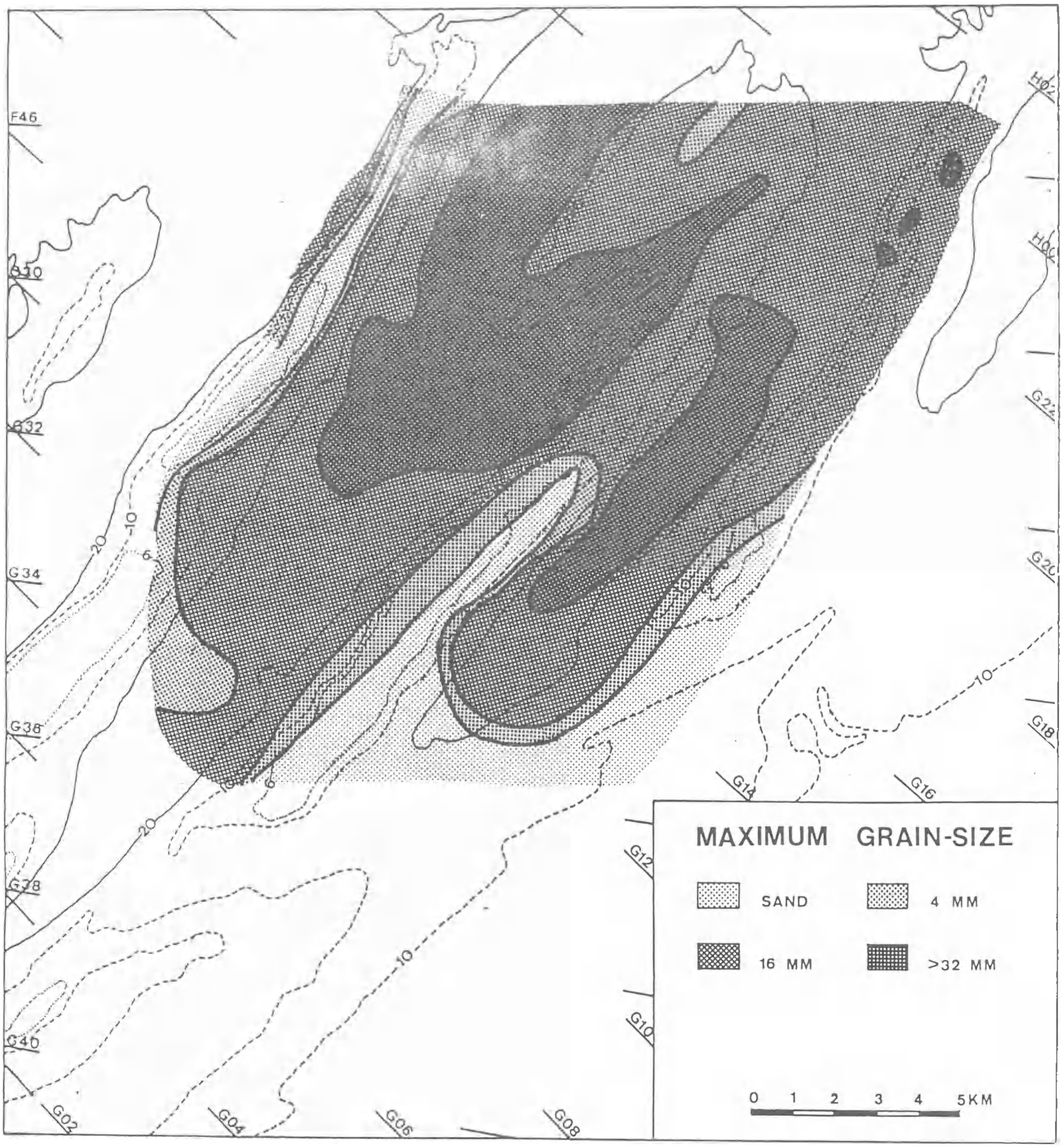

Figure 3 Distribution of the maximum grain sizes.

GRAIN SIZE DISTRIBUTION OF THE SEDIMENTS

Granulometric data are collected and analysed in function of the gravel characteristics. Since the total grain size distributions of the sediments are polymodal, the mean and standard deviation are determined according to FOLK and WARD (1957). This is done for both decalcified and non-decalcified samples, as well for the total sample as for the sand fraction. There is a general increase of grain size from the south to the north of the area. The superficial sediments on top of the ridges consist of fine to medium, moderately to well sorted sands. The graphical mean and standard deviation remain unchanged after decalcification. This can be explained by the small variance between the mean sizes and sorting of the lithic and carbonatic fractions. Due to the lack of gravel, there is neither a real difference between the mean of the total sample and that of the sand fraction.

The channels contain coarse (Buiten Ratel channel) to fine (Kwinte) sands. This difference is owing to the presence of shell debris in the sand fractions of the sediments in the Buiten Ratel channel. In consequence of the difference in the mean of the lithic and carbonatic sand fraction, the sands in the channels are poorly to moderately sorted. The variables also differ strongly after decalcification.

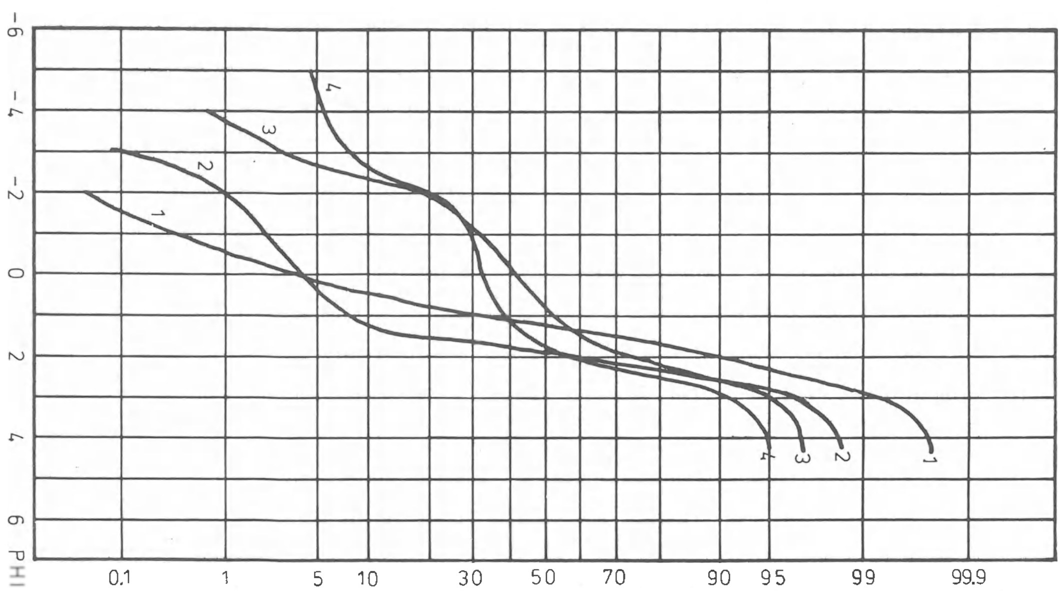

Figure 4 Grain size distribution types along the tongue formy pattern over the Buiten Ratel. For location, see figure 6.

Particular attention is drawn to the sedimentological characteristics of the samples in the tongue and ribbon formy patterns. For that purpose four mean grain size distributions are studied along the tongue over the Buiten Ratel (figure 4). Down the steep NW side of the Buiten Ratel the sediments contain about 30 % gravel, most lithic, and about 5 % mud. There is a pronounced gap in the distribution between the grain sizes -2 and 1 ϕ (distribution type 4). Along the tongue, following the NNE direction, there is a clear evolution of this distribution. The grain size gap gets filled up starting with the coarser sizes. Gravel and mud percentages decrease and the maximum grain sizes diminish. At the end of this tongue the sediment distributions are almost unimodal (type 1). The only gravel size sediments are the biogenetic carbonates.

PETROLOGY OF THE GRAVEL

The petrographic composition of the gravel fraction of all samples has been studied. For that purpose each pebble is put in one of the petrographic classes.

The flint pebbles are cloudy, dark or brown and consist of microcrystalline silica. The clasts are very well rounded. They originate from Cretaceous deposits outcropping in the Strait of Dover.

The weathered flints are white or palish brown. They usually have a nucleus of flint. The crust or 'pattina' is composed of tripalite. Weathered flints occur frequently in the Diestian Formation and in Tertiary and Quaternary base gravels.

The sandstone pebbles are mostly fine and well rounded. A distinction is made between two large groups :

1. Coarse sandy sandstone, brown, containing grit. These pebbles contain limonite-stained and limonite-coated intraclasts. This sandstone is very similar to that one of the Diestian formation.

2. Grey to green sandstone containing glauconite grains. In the past this gravel is always mentioned as a local gravel belonging to the Mont Panisel Formation (LERICHE, 1948 ; TESCH and REINHOLD, 1946 ; VERBEECK, 1954). By means of the lithological characteristics of these clasts and using the microfauna, a first subtype can be correlated to the Gardes Formation (Aptian-Albian, Lower Cretaceous) in northern France. The second subtype can be considered as a part of the Mont Panisel Formation (Eocene).

The quartz grains occur mainly as a very fine gravel, mainly yellow and white. They are mostly very well rounded and sometimes even polished. More to the south, quartz grains are not noticed in the Sandettie-Fairy Bank gap (KIRBY and OELE, 1975). Previous investigations consider quartz pebbles to be a Rhine and Meuse gravel, which is dated as Holstein and early Saale (VAN DER LIJN, 1949 ; ZANDSTRA, 1959) and is strongly reworked by the Holocene transgression.

Clay balls are nearly only found in small quantities and in the interbank channels. A first type consists of fine sandy clay, well rounded, brown or brownish red. These are very similar to the fine matrix of the Diestian Formation. The second type varies from fine to medium gravel. These clay balls are grey. They are extremely sensitive for desaggregation. Pollenanalysis points to a Pleistocene-recent age (Personal communication, C. VERBRUGGEN and J. DE CONINCK).

The class of the fossils consists of a large number of *Nummulites* which are very difficult to study because of their abraded characteristics. The shark teeth are dated as Eocene (*Isurus novus, Galeorhinus minor*). One *Rugosa* specimen is found on top of the Kwinte Bank. This fossil occurs in rocks of Ordovician to Permian age (often in Devonian in Belgium).

The crystalline pebbles occur as a fine gravel, very well rounded and nearly spherical. They mainly consist of granite and alkalifeldspar granite. Thin sections of larger pebbles indicate a strong alteration of biotite and feldspar progressively from the center to the outer edge of the clasts. Some larger pebbles are fractured without one spur of weathering.

	ODG	OD	BRG	BR	KW	KWB
flint	0.18	0.45	6.51	0.36	1.07	0.23
w. flint	0.30	0.28	3.51	0.31	0.77	0.14
sandst.	0.12	0.10	1.94	0.10	0.19	0.04
quartz	0.14	0.08	1.45	0.11	0.24	0.06
fossils	0.01	0.01	0.01	0.01	0.02	0.00
crystal.	0.04	0.02	0.09	0.03	0.08	0.02
clay (grey)	0.01	0.00	0.23	0.00	0.34	0.00
clay (brown)	0.10	0.03	0.34	0.04	0.24	0.04
others	0.02	0.05	0.39	0.06	0.51	0.04
shells	2.96	1.63	6.30	2.13	3.50	1.15

Table 1 Percentages of each class in the morphological units of the Flemish Banks. ODG = Oost Dyck channel. OD = Oost Dyck. BRG = Buiten Ratel channel. BR = Buiten Ratel. KW = Kwinte. KWB = Kwinte Bank.

The class 'others' consists of rare gravel types or macroscopical non-determinable clasts, including the class of the limestones. This rock is often used as shipload and for the construction of breakwaters. The rest of the class consists mainly of waste, such as oil nodes, pieces of iron and even munition. In some cases they amount to 10 % of the gravel sample (table 1).

STATISTICAL ANALYSIS ON THE PETROGRAPHIC COMPOSITION

Samples with similar petrographic compositions are grouped using cluster analysis. This is a straightforward, logical, pair-by-pair comparison between samples. One of the objectives of this multivariate study is to eliminate extreme petrographic variables based on the mean and the standard deviation. This elimination has rejected 22 samples out of 150 for further calculations. The results are reproduced in table 2 and figure 5.

cluster	1	2	3	4
flint	0.02	0.52	2.32	0.45
w. flint	0.02	0.37	1.75	0.39
sandst.	0.01	0.15	0.63	0.19
quartz	0.01	0.12	0.46	0.14
fossils	0.00	0.01	0.02	0.01
crystal.	0.00	0.04	0.09	0.04
clay (grey)	0.00	0.01	0.00	0.52
clay (brown)	0.00	0.05	0.34	0.14
others	0.00	0.08	0.26	0.11
shells	0.41	3.11	7.12	3.81
total	0.47	4.45	13.37	5.77

Table 2

Cluster 1

This cluster groups the samples with poor gravel content. The coarse fraction consists of fine shells and shell debris along with very small quantities of lithic gravel. This cluster occurs exclusively on top of the ridges and in the shallow southern parts of the swales.

Cluster 2

Cluster 2 contains about 4.45 % gravel which consists mainly of shells, flint and weathered flint. The Tertiary input is indicated by the presence of sandstone, brown clay balls and Tertiary fossils. These samples occur in the same configuration as the tongue and ribbon formy patterns.

Cluster 3

Cluster 3 is an assemblage of gravelly sediments which consist mostly of shells, flint and weathered flint. In comparison with the previous clusters, high percentages of quartz, crystalline rock fragments and brown clay balls occur which are typically exotic gravels. The local input comes from the relatively high amounts of sandstone (mostly Eocene) and fossils. These samples all occur in the deepest parts of the swales.

Cluster 4

This cluster contains about 5.77 % gravel and is, beside the shells, characterized by the dominance of grey clay balls. These samples are all taken down the steep NW slopes of the ridges.

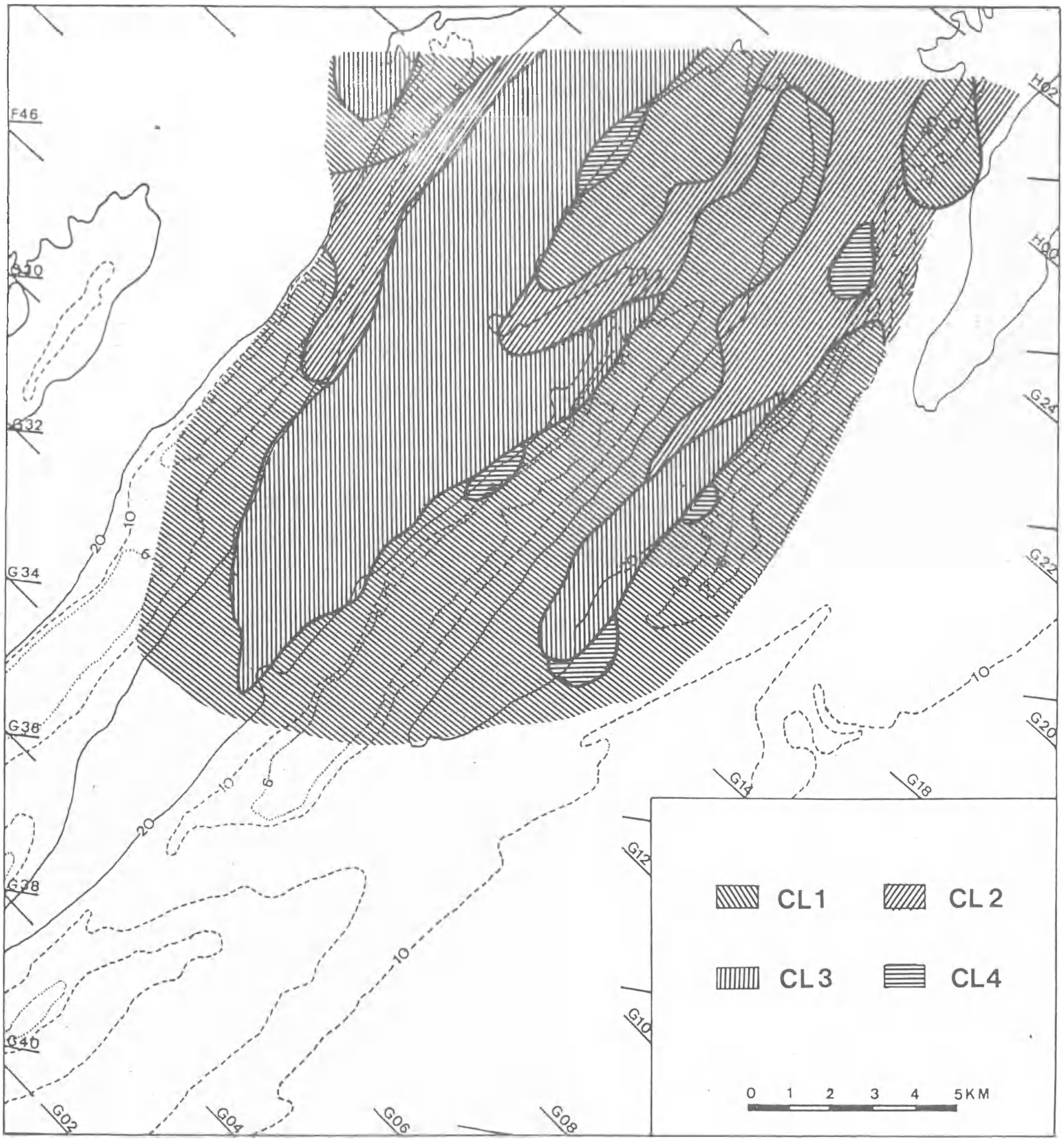

Figure 5 Distribution of the petrographic clusters.

INTERPRETATION AND DISCUSSION

The area under investigation is characterized by a strong tidal activity and tide generated sediment mobility. Maximum ebb and flood currents are nearly parallel to the ridges and the interbank channels (figure 6). Consequently erosion dominates and a gravelly lag deposit develops in the channels. Local higher concentrations of gravel in the channels can be connected with a deepening at the same place by locally more intensive erosion. As a result, this erosion causes a blow out of the sediments between -2 and 1 ϕ while the finer fractions fill the pore spaces of the gravel bed.

As there is a clear evolution of the grain size distribution (type 4 to 1) along the tongue over the Buiten Ratel, the conclusion is made that this gravel is taken up in the bed load transport. The composition and the grain

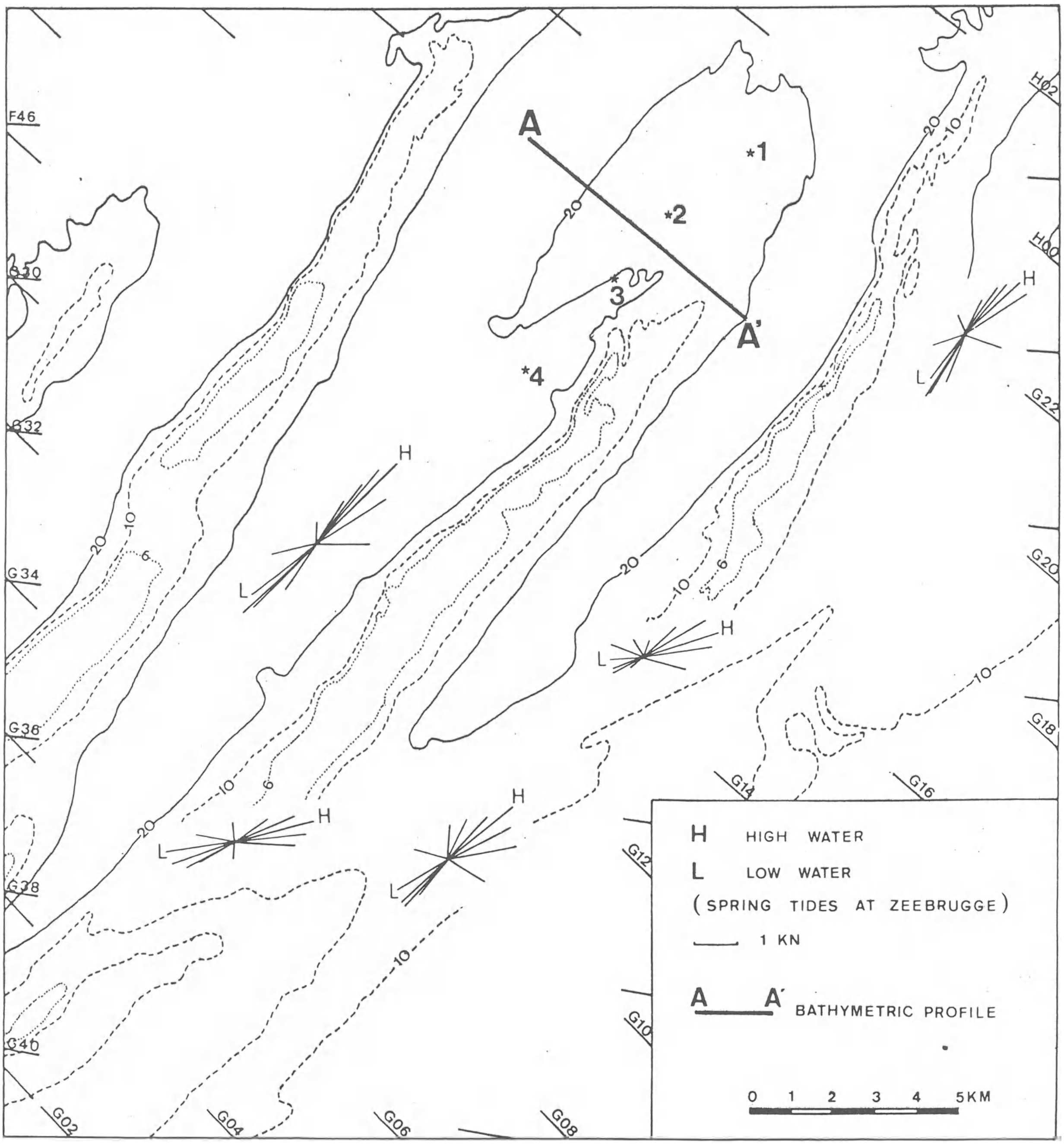

Figure 6 Tidal current estimates. Location of section A-A' (see figure 7). Location of grain size distribution types (see figure 4).

sizes suggest that the gravel in this tongue corresponds with the gravel hiatus in the distribution type 4 localised in the Buiten Ratel channel.

The presence of these gravel paths goes together with the presence of large sand wave trains indicating a same direction of transport. Only the sand waves present on the NW slopes of the Buiten Ratel and the Kwinte Bank (figure 7) seem to have embodied fine lithic and medium carbonatic gravels. Coarse layers in sand waves are also noticed in cores from the Sandette-Fairy Bank gap (KIRBY and OELE 1975). Occurrence of coarse sand and fine gravel in sand waves is one of the characteristics of the masterbedding, being the steeper sand wave face and recording the erosion during the ebb stream (ALLEN, 1982).

The local low concentrations of shells in the channels and the higher ones on the ridges are caused by the transport. The size distributions of the shells are very well sorted and the modus coincides with the

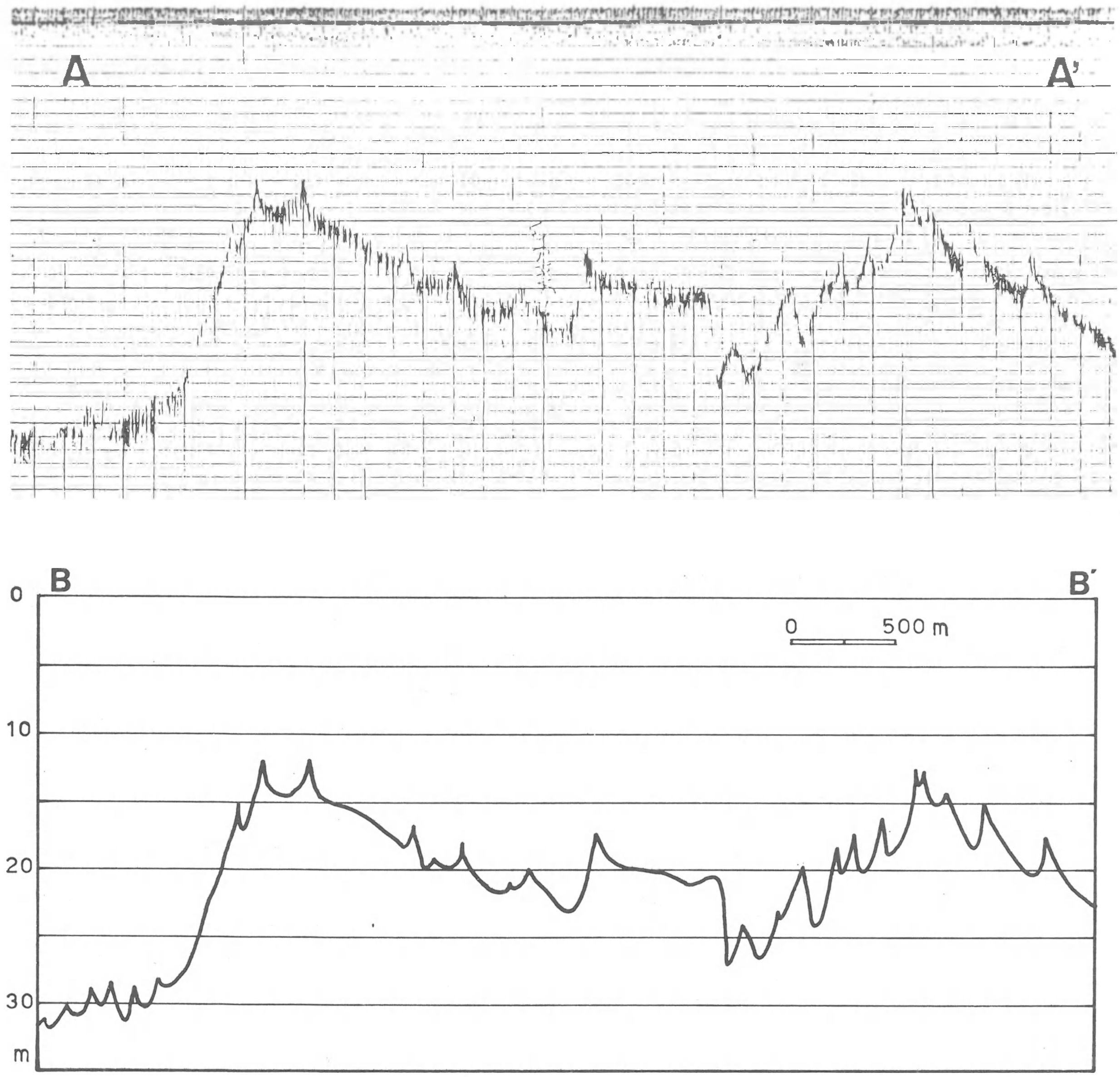

Figure 7 Echo-sounder profile across the Buiten Ratel showing the abundance of asymmetric sand waves and the local deepening over this ridge. For location see, see figure 6.

hydraulic equivalent sizes of the fine gravel. This means that if the critical velocity is reached to transport fine gravel, nearly all shells are carried off. Only a few bigger or living shells remain behind in the swales.

Along the transport paths the decreasing tidal current causes the deposition of the gravel starting with the coarsest clasts. Near the end of these paths, the sand gets dispersed by the increasing wave effectiveness to the top of the ridges. Due to deficit on velocity measurements, it is not clear whether this transport occurs during the normal flood current or only under tempestuous, flood favourable conditions.

The transport paths take care of local deepenings over the ridges which can easely be found back on the bathymetric maps. These deepenings leave the channels at kinks formed in those ridges.

The grey clay balls occur completely independently. This can be explained by the origin. The great sensitivity for desaggregation excludes transport which means that this gravel is strictly local. It is only found in the swales where it is eroded. It is a very young gravel which will be destroyed very fast due to the abrasion of moving sand and water.

SUMMARY

The transport of fine gravel is incorporated in the build up and maintenance mechanism of the Flemish Banks. This incorporation occurs during the flood stream and the transport of gravel is therefore wholly unidirectional. Both grain size and petrographic data support the idea that this is a recent transport feeded by sediments from the interbank channels. Although it is obvious that the gravels in the Flemish Banks consist of three main deposits, it is very hard to stake them out. Due to the Holocene marine transgressions the gravels are becoming a mixture of southern coming, Rhine and Meuse, and local clasts.

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19 A comparative study of sedimentological parameters of some superficial sediments on the Flemish Banks

J. LANCKNEUS

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

INTRODUCTION

The Flemish Banks have been sampled several times (VAN MIERLO, 1899 ; GILSON, 1900 ; GULLENTOPS *et al.*, 1972 ; BASTIN, 1974 ; GULLENTOPS *et al.*, 1976). The density of the sample points during most of those campaigns was severly restricted due to the great extent of the sampling area. The results of those campaigns made it possible to obtain a good picture of the distribution of the seabottom sediments of the North Sea, but nevertheless little was known about the local grain-size characteristics of one particular sand ridge or swale.

During the past 2 years (1982-1984) a new investigation program has been directed by Dr. G. DE MOOR (DE MOOR, 1985) in which sampling was restricted to the sand ridges and swales of the Flemish Banks area (Kwinte Bank, Buiten Ratel Bank, Oost Dyck, Negenvaam, Kwinte geul, Ratel geul, Oost Dyck geul) (figure 1).

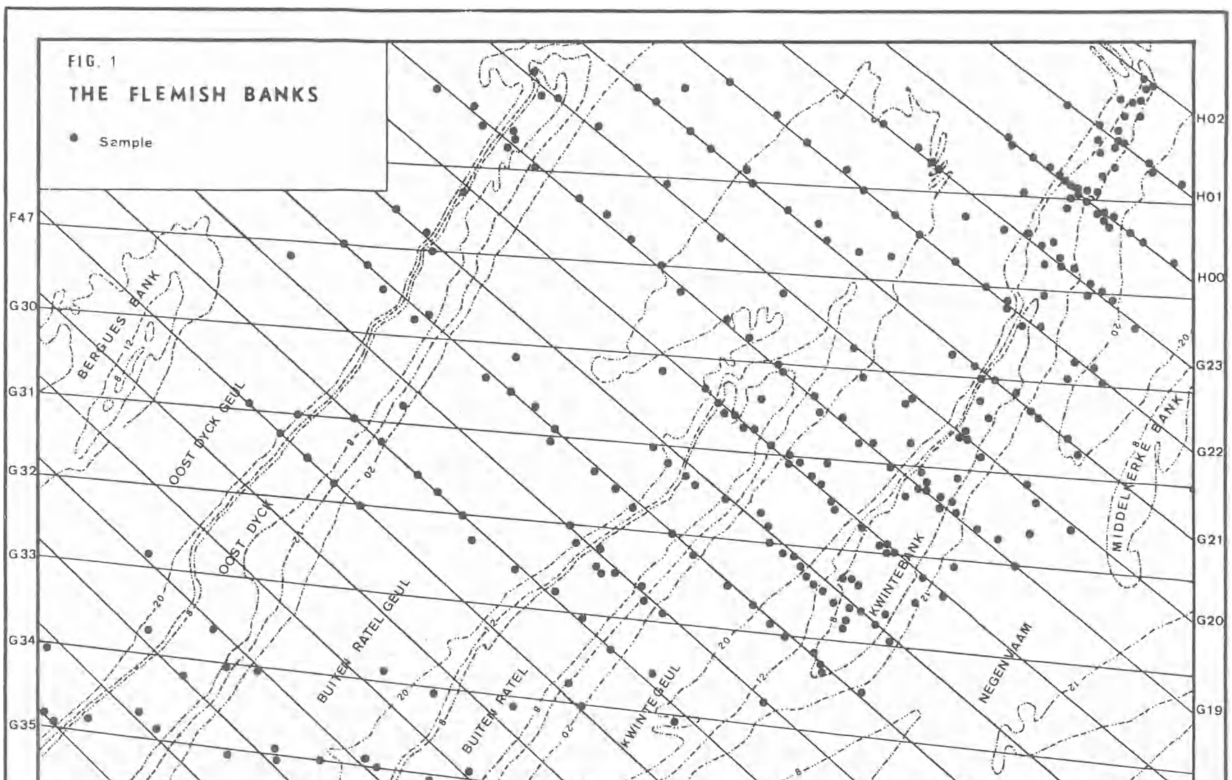

Figure 1 Sample locations on the Flemish Banks.

Some 320 samples of the superficial sediments were taken with a Van Veen bottom sampler (figure 1). This dense sample net in a relative small area has allowed us to obtain a good picture of the distribution of the sediments on a rather large scale. Each sample was analysed and different grain-size parameters were calculated. The parameters used in this study were for each sample :

- mean, calculated according to FOLK and WARD, determined on a sample of ± 2000 g (< 4 mm) without elimination of the $CaCO_3$ ("natural" mean) ;
- mean, determined after the complete elimination of the $CaCO_3$ ("decalcified" mean) ;
- $CaCO_3$ content (in %) of the sediment (4 mm - 50 μm) ; determined by disolution of the carbonates by 2N HCl ;
- silt-clay fraction (in %) ;
- gravel content of the sediments (> 4 mm), determined on a sample of ± 4 kg, without decalcification, including shell fragments and petrographic gravel.

METHODS

The obtained parameters can best be studied by a graphic representation, which can be performed in different ways.

In most cases the values of the parameters are divided up into classes. Each class is represented by a different colour. Maps can then be drawn in which an area of a particular colour groups all the values falling into the chosen class intervals.

The map representing those classes gives a good overall picture. Nevertheless, the choice of the class intervals often happens in an arbitrary way and the picture completely depends on this choice.

Furthermore those intervals are artificial and they do not correspond to possible existing changes in sediments.

For those reasons another approach was chosen. A cluster analysis was selected to separate all values of each parameter into groups (DAVIS, 1973).

Cluster analysis is a technique in which objects are placed into a hierarchy in such a way that objects with the highest mutual similarity are placed together. Then those groups or clusters are associated with other groups, which they most closely resemble, and so on, until all of the objects have been placed in a complete classification scheme.

In this study the chosen similarity coefficient was the squared euclidean distance. This coefficient is computed by

$$d_{ij} = \sqrt{\frac{\Sigma_{k-l} (X_{ik} - X_{jk})^2}{m}}$$

where X_{ik} denotes the kth variable measured on object i, X_{jk} is the kth variable measured on object j and m is the number of variables. The clustering technique between groups here used is Ward's method (WARD, 1963). In our particular case, the cluster analysis was first performed on each parameter and 4 final clusters were taken in consideration, the silt-clay fraction excepted, where only 3 clusters were used because of the very low silt-clay content of the sediments.

As we are mainly interested in the behaviour of each morphological unit separately we will consider the sand ridges and swales as separate populations. To avoid some extreme high or low values, which can heavily alter the global picture, we do not consider data larger or smaller than 1.96 times the standard deviation of each population.

The samples belonging to a same cluster are grouped together in a class represented by a particular point-density symbol (see following figures). When the distribution of values is highly irregular, like for example in the swales, some classes can not be drawn. In this case each sample just represents by symbol the cluster to which it belongs.

An average is calculated for each cluster to make a numeric comparison possible between the different populations.

The following tables represent those averages. Only in a few cases all 4 values of each population are given; in most cases only the minimum and maximum values are represented.

DISCUSSION OF EACH PARAMETER

	Subpopulations			
	Lowest value	Low value	High value	Highest value
Sand ridges				
Kwinte Bank	224	289	332	404
Buiten Ratel	237	294	348	473
Oost Dyck	238	292	363	435
Swales				
Negenvaam	206	223	243	306
Kwinte geul	210	316	457	599
Ratel geul	245	310	380	543
Oost Dyck geul	200	273	341	429

Table 1 Mean (FOLK and WARD, 1957), determined on samples without decalcification (mm).

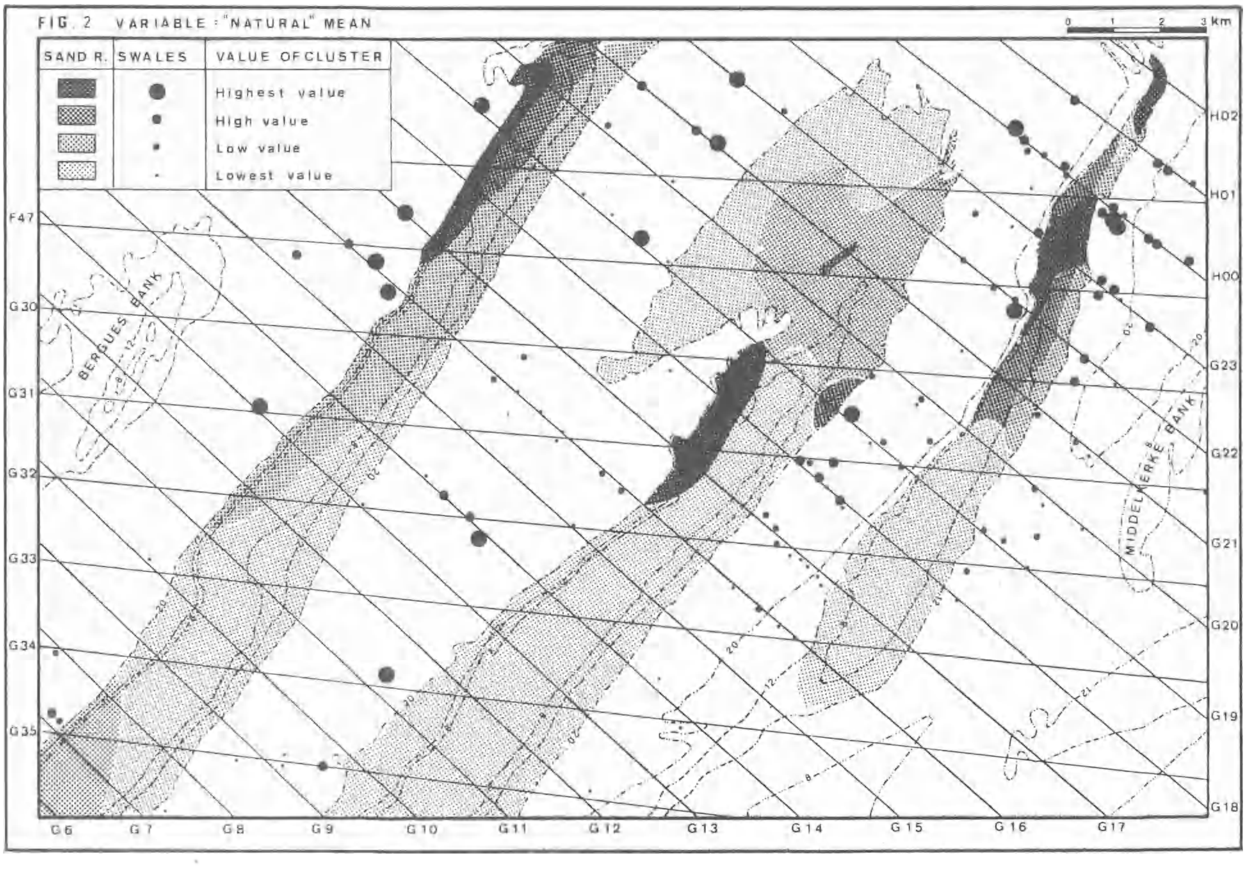

Figure 2 Variable : "natural" mean.

1. Mean, determined on sediments without decalcification ("natural" mean) (table 1, figure 2)

a. Sand ridges

The sand on the three sand ridges becomes gradually coarser towards the NE end of the ridge. The mean of the Kwinte Bank, for example, changes from 224 in the south to 404 μm in the north. The Buiten Ratel and the Oost Dyck show a similar behaviour. We also observe on the northern part of the ridges that the slopes of the ridges present coarser sand than the centrel part of the ridge ; furthermore, the western slope always shows coarser sand than the eastern one.

A comparison of the obtained values between the sand ridges indicates a strong likeness between all subpopulations of the three ridges.

b. Swales

The swales, contrary to the regular pattern found on the sand ridges, are characterized by a somewhat arbitrary distribution of values. The Ratel geul has its high values all located in the central part of the swale, while the Oost Dyck geul contains the coarsest sand near the western flank of the Oost Dyck.

The values in the Negenvaam and Kwinte geul do not seem to follow a regular pattern.

If we compare the different swales, we notice the presence of the coarsest sand in the Kwinte geul.

2. Mean, determined on decalcified samples ("decalcified" mean) (table 2, figure 3)

a. Sand ridges

The local pattern of the "decalcified" mean on the sand ridges is very similar to the one just observed with the "natural mean". The mean of the superficial sediment raises from the SW end of the ridge to the NE end. Furthermore, we notice the same grain-size changes on the slopes, as observed with the "natural" mean.

Compared to the two other sand ridges, however, the Oost Dyck now contains the coarsest sand in all four subpopulations.

	Subpopulations			
	Lowest value	Low value	High value	Highest value
Sand ridges				
Kwinte Bank	214	248	290	353
Buiten Ratel	223	243	271	303
Oost Dyck	241	282	340	373
Swales				
Negenvaam	183	219	261	302
Kwinte geul	217	254	287	354
Ratel geul	229	257	301	377
Oost Dyck geul	225	268	328	382

Table 2 Mean (FOLK and WARD, 1957), determined on decalcified samples (mm).

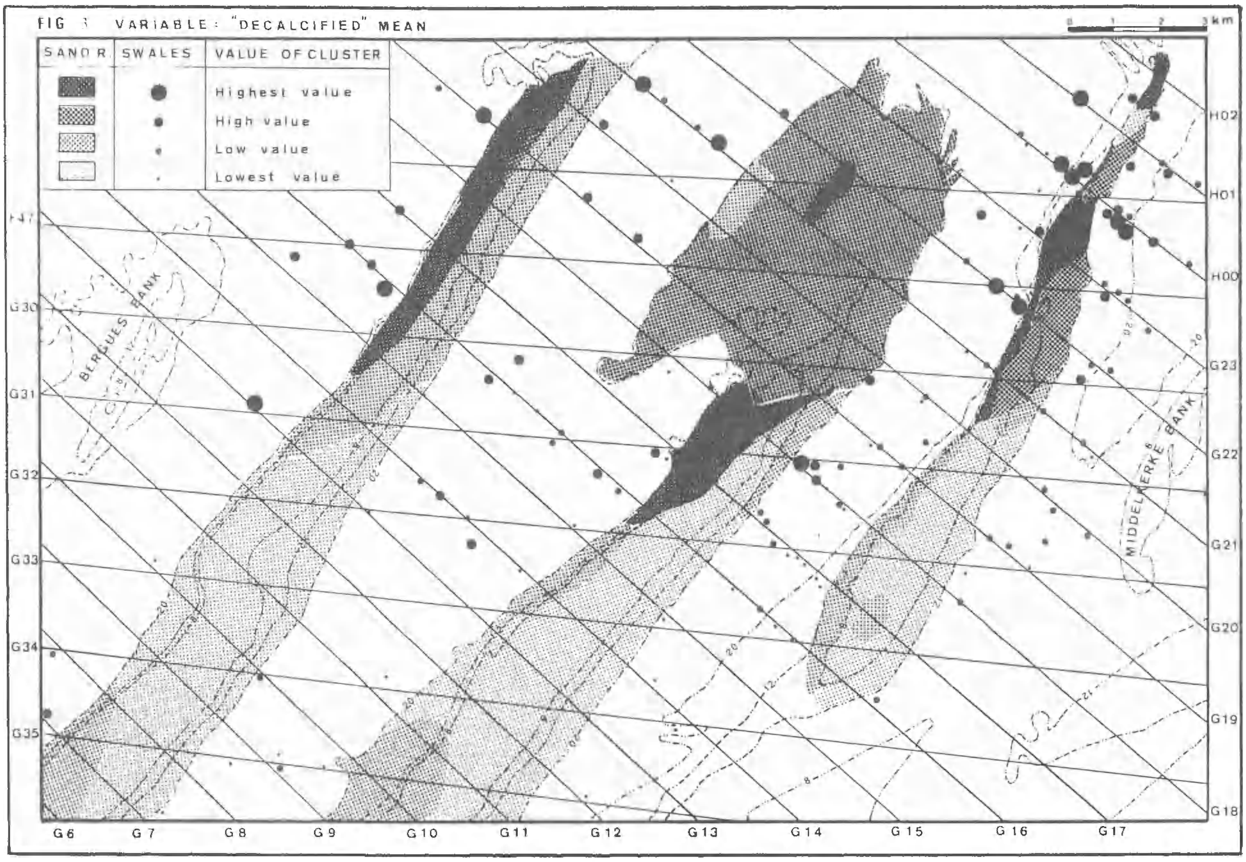

Figure 3 Variable : "decalcified" mean.

b. Swales

The distribution of the values in the swales follows an irregular pattern.

However, a comparison of all subpopulations of each swale shows a systematic increase of the mean from the SE (Negenvaam) to the NW (Oost Dyck geul). An opposite picture was obtained from the "natural" means, in which the Kwinte geul presented the coarsest sediment with an occurrence of finer sediments, to the Negenvaam as well as to the Oost Dyck.

So it seems that the pattern modifies and simplifies itself after decalcification of the sediments.

It can be summed up as follows :
- a coarsening of the sediment on the sand ridges towards the NE end of the ridges ;
- on a smaller scale : a global coarsening, as well in the swales as on the ridges, from the SE towards the NW.

3. $CaCO_3$ content (table 3, figure 4)

a. Sand ridges

The $CaCO_3$ percentage on the sand ridges increases to the NE end of each ridge.

At the same time we note a general reduction of the $CaCO_3$ content from the Kwinte Bank and the Buiten Ratel, which show similar values towards the Oost Dyck.

	Subpopulations	
	Lowest value	Highest value
Sand ridges		
Kwinte Bank	7 %	32 %
Buiten Ratel	7 %	35 %
Oost Dyck	4 %	19 %
Swales		
Negenvaam	9 %	17 %
Kwinte geul	13 %	42 %
Ratel geul	9 %	40 %
Oost Dyck geul	7 %	27 %

Table 3 CaCO3 percentage (%). (<4 mm)

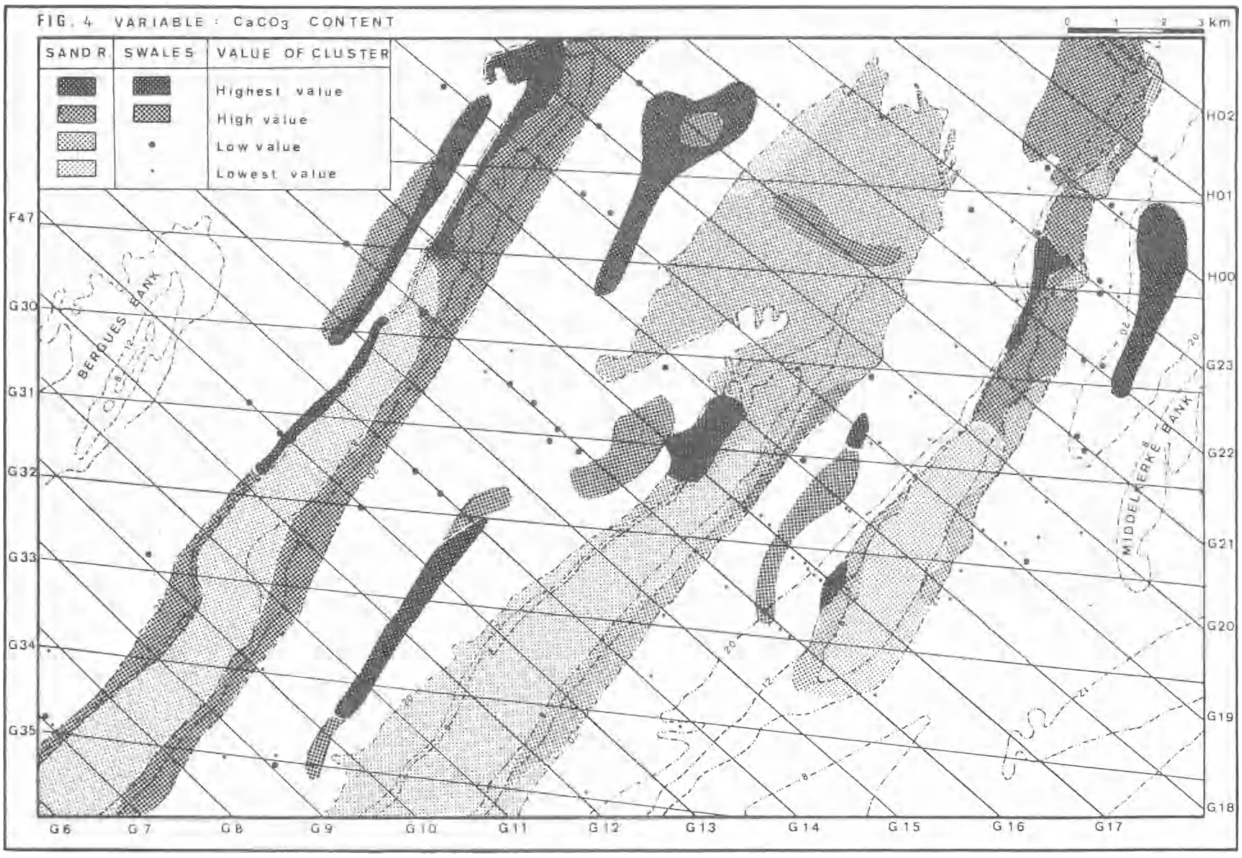

Figure 4 Variable : CaCO3 content.

	Subpopulations	
	Lowest value	Highest value
Sand ridges		
Kwinte Bank	0.1 %	7 %
Buiten Ratel	0.2 %	3 %
Oost Dyck	0.0 %	2 %
Swales		
Negenvaam	0.3 %	2 %
Kwinte geul	2 %	27 %
Ratel geul	3 %	42 %
Oost Dyck geul	1 %	19 %

Table 4 Gravel percentage (%). (>4 mm)

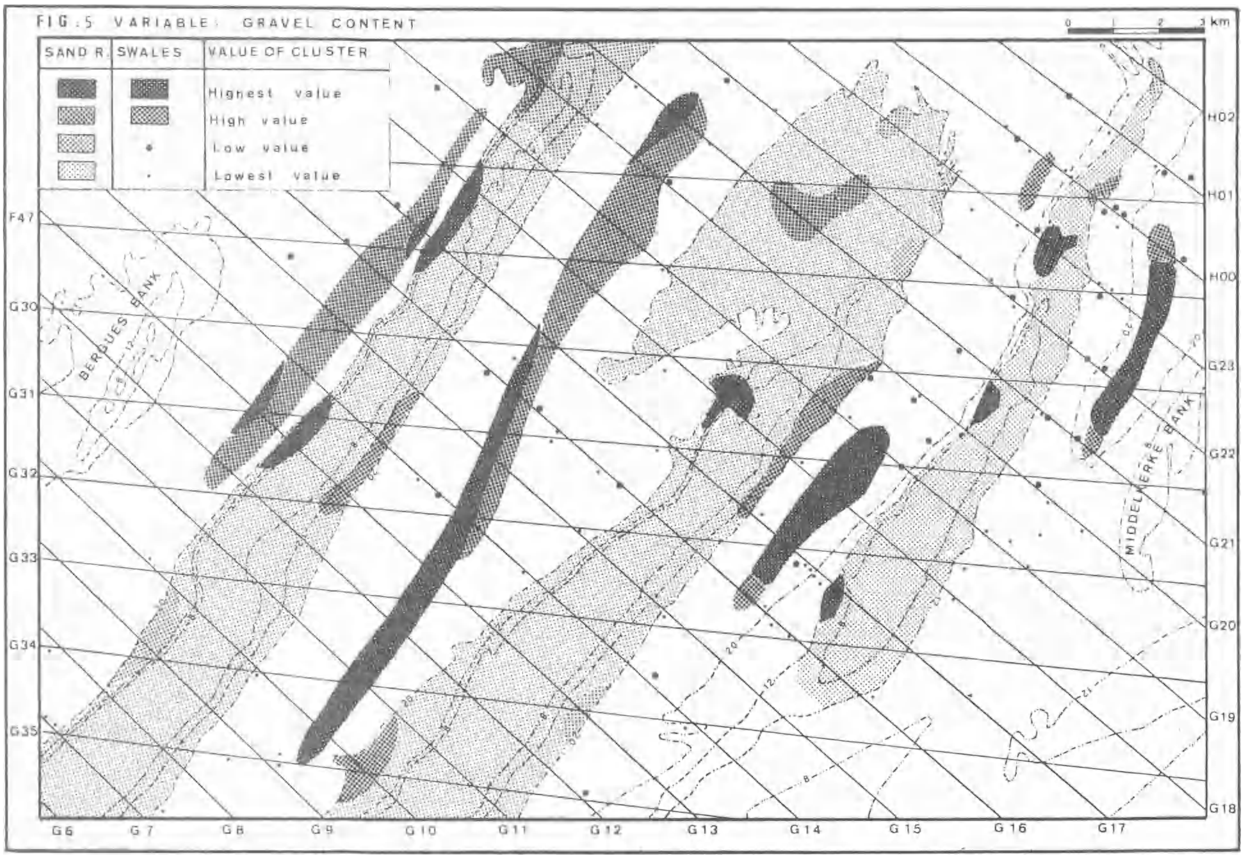

Figure 5 Variable : gravel content.

	Subpopulations	
	Lowest value	Highest value
Sand ridges		
Kwinte Bank	0.5 %	3 %
Buiten Ratel	0.5 %	2 %
Oost Dyck	0.6 %	1.1 %
Swales		
Negenvaam	2 %	8 %
Kwinte geul	1 %	4 %
Ratel geul	1 %	3 %
Oost Dyck geul	0.6 %	1.2 %

Table 5 Silt-clay fraction (%).

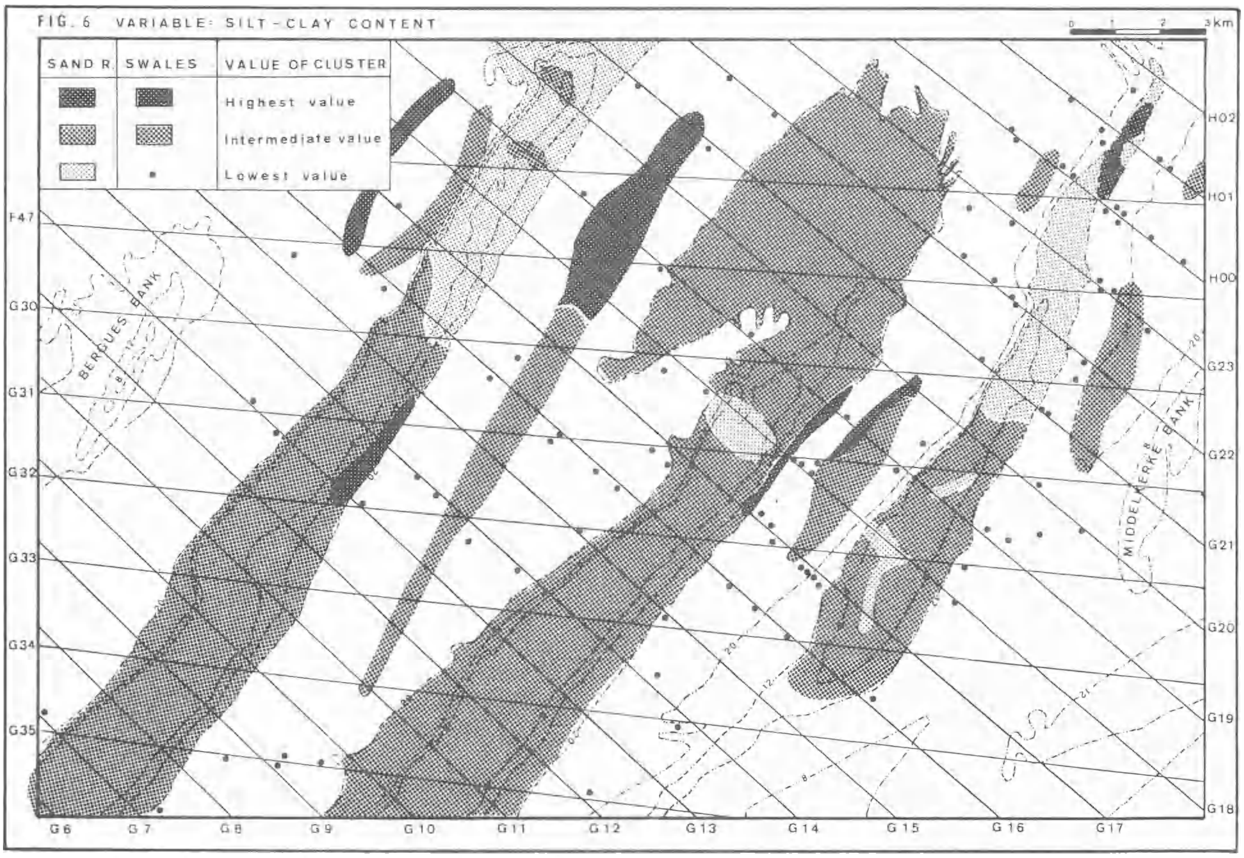

Figure 6 Variable : silt-clay content.

b. Swales

The swales contain systematically more $CaCO_3$ than the banks. Both Negenvaam and Kwinte geul show a general increase of the values towards the central parts of the swales.

The pattern in the Ratel geul is somewhat more capricious but it follows more or less the 20 m depth line of the Buiten Ratel. It is difficult to express an opinion about the Oost Dyck geul due to the lack of taken samples in the area.

The values of the subpopulations increase gradually and systematically from the Oost Dyck geul towards the Kwinte geul, where it runs up to a maximum of 42 %. In this respect the Negenvaam forms an exception because of its lowest maximum value of 17 %.

4. Gravel (table 4, figure 5)

a. Sand ridges

The gravel content on the ridges is rather uniform ; some higher values occur towards the NE end of the ridge and on the slopes of the ridges.

On a smaller scale, the values augment towards the SE, where they reach a maximum value on the Kwinte Bank with 7 %. Those high figures are mainly due to shell fragments. As a result, the patterns of occurrence of the gravel and the $CaCO_3$ are very likely.

b. Swales

The gravel content is significantly higher in the swales than on the ridges. Here too we observe the same pattern as with the $CaCO_3$. In other words, we note a general increase of the values towards the central parts of the swales.

The Ratel geul displays the highest values while the Negenvaam is characterised by extremely low values. The sampling net was not dense enough in the Oost Dyck geul to enable us to come to a conclusion about its gravel characteristics.

5. Silt and clay fraction (table 5, figure 6)

a. Sand ridges

The Kwinte Bank and the Oost Dyck display the same pattern in which the silt-clay content decreases from the SW towards the NE end of the ridge. The Buiten Ratel does not correspond to this image because of its uniformity on the whole ridge (1 %), with the only exception of a small area located at the north end of the sand ridge at -8 m. This same location is also characterised by very high values of mean, $CaCO_3$ and gravel content.

The measured values in silt and clay fraction on the three ridges differ little from each other.

b. Swales

The swales show higher maximum values than the ridges. The Kwinte geul and the Ratel geul are once again characterised by a higher content of silt-clay in the central parts of the swales. Those high values follow almost the same pattern as the maximum values of the gravel.

A comparison between different swales indicates a maximum value for the Negenvaam followed by a progressive decrease towards the Oost Dyck geul.

CONCLUSIONS

The "natural" mean values are strongly affected by two opposite factors :
1. a coarsening of the quartz sand towards the NW, where little $CaCO_3$ is found ;
2. an increase in coarse shell fragments towards the SE.

Both factors cause a similarity in "natural" mean between the sediments of the whole area. The important values of the "natural" mean of the Kwinte geul, apparently higher than those of the Oost Dyck geul, are also due to high $CaCO_3$ percentages. In this respect one can refer to the high positive correlation coefficient of 0.89 between the "natural" mean and the $CaCO_3$ content, measured on the whole area of the Flemish Banks. So we can conclude that the "natural" mean, can be hasardous to use in comparing, for example, the local energy levels of the environment. Further conclusions can be taken :

- for the sand ridges : towards the NE end of the sand ridges, the means, the $CaCO_3$ content and the gravel percentage increase, while the silt-clay fraction decreases. On a smaller scale the sand coarsens from the SE towards the NW and the values of gravel and $CaCO_3$ increase in the opposite direction. The Flemish Banks have a very low and almost similar percentage of silt-clay.

- for the swales : the grain-size changes in the swales occur in a perpendicular way to the axis of the swales. On a smaller scale the parameters behave almost in the same way as on sand ridges : the sand coarsens, from the SE towards the NW ; the silt-clay fraction and the $CaCO_3$ percentage increase in opposite direction. The Ratel geul presents the highest percentages of gravel.

CLUSTER ANALYSIS WITH 4 PARAMETERS ON THE KWINTE BANK

The preceding cluster analyses were realised with only one variable, which was one of our 4 parameters. The next analysis was achieved with 4 variables : "decalcified" mean, $CaCO_3$ content, silt-clay and gravel percentage. Numeric values are also calculated for the 4 final clusters of the Kwinte Bank. Our 4 parameters are expressed in different types of measurement units (%, μm) therefore, we must transform the original values of our parameters to Z-values (DAVIS, 1973). Those are obtained as follows :

$$Z_b = \frac{X_b - m}{s}$$

where X_b is the bth value of the population and m and s represent the mean and standard deviation of the population. In this way we obtain from each sample, 4 Z-values, one for each parameter, from which an average value M_i can be taken. A numeric value (M) for the population is obtained by taking the average of all M_i values belonging to the cluster.

Table 6 shows us the obtained numeric values M on the Kwinte Bank and the results of the cluster analysis are shown in figure 7. Each of the 4 final clusters contains samples which are more or less similar in sedimentological characteristics. The most southern cluster embraces sediments characterised by a small mean, by low percentages of $CaCO_3$ and gravel and by a high content in silt-clay. The low values of our first 3 parameters are responsible for the negative numeric value (- 0.58), just slightly increased by the higher silt-clay content. The most northern cluster comprises sediments characterised by a high mean, high percentages of $CaCO_3$ and gravel and a low content in silt-clay. Those mainly high figures are reflected in the positive numeric value M (1.30). The opposite behaviour of the silt-clay content in respect with the other 3 parameters can clearly be deduced from table 7.

CLUSTER ANALYSIS WITH 4 PARAMETERS ON THE FLEMISH BANKS

Till now we have always considered the sand ridges and the swales as separate populations. It could be interesting to compare the sediments of the sand ridges with those of the swales.

A cluster analysis with the same 4 parameters than the previous analysis has been carried out on all samples of the area.

The result with 8 clusters was retained. Four of those clusters group exclusively samples of the swales and in particularly of the center parts of the swales. The remaining 4 clusters contain all the samples of the sand ridges. Samples from the slopes of the sand ridges fall also in this second category. This regular separation of the sand ridges from the swales can be appreciated on figure 8. On this figure, the 8 clusters are fused into 2 main populations, one grouping the 4 clusters of the swales, the other containing the 4 clusters of the ridges.

This image could probably still be improved by considering other variables, like for example the sorting. We can conclude from those results that basic differences do exist in grain-size characteristics between the sand ridges and the swales.

	Subpopulations			
	Lowest value	Low value	High value	Highest value
Kwinte Bank	-0.58	-0.26	0.07	1.30

Table 6 Numeric values of the 4 clusters on the Kwinte Bank.

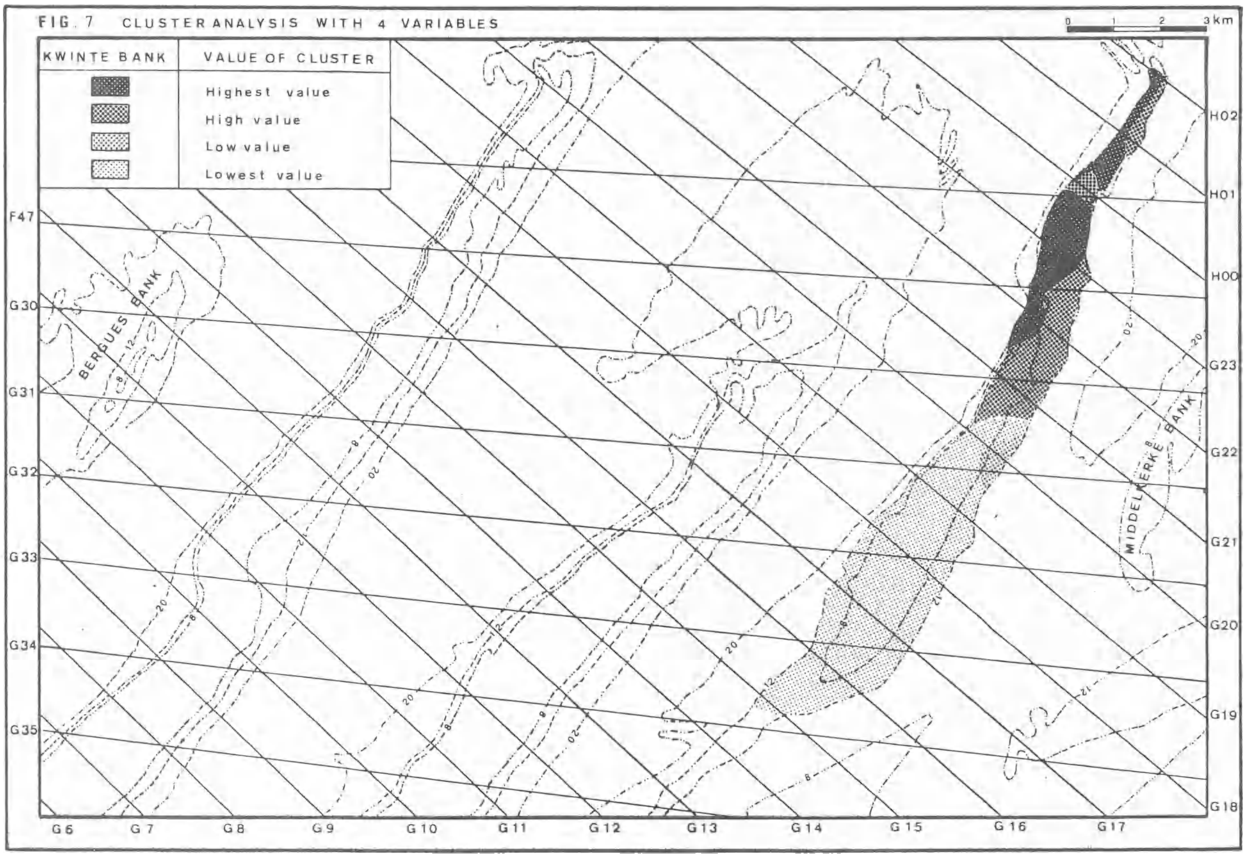

Figure 7 Cluster analysis with 4 variables.

Swales	Mean	$CaCO_3$	Silt-clay
$CaCO_3$	0.87		
Silt-clay	-0.37	-0.18	
2 mm	0.55	0.63	-0.11

Table 7 Correlation of parameters.

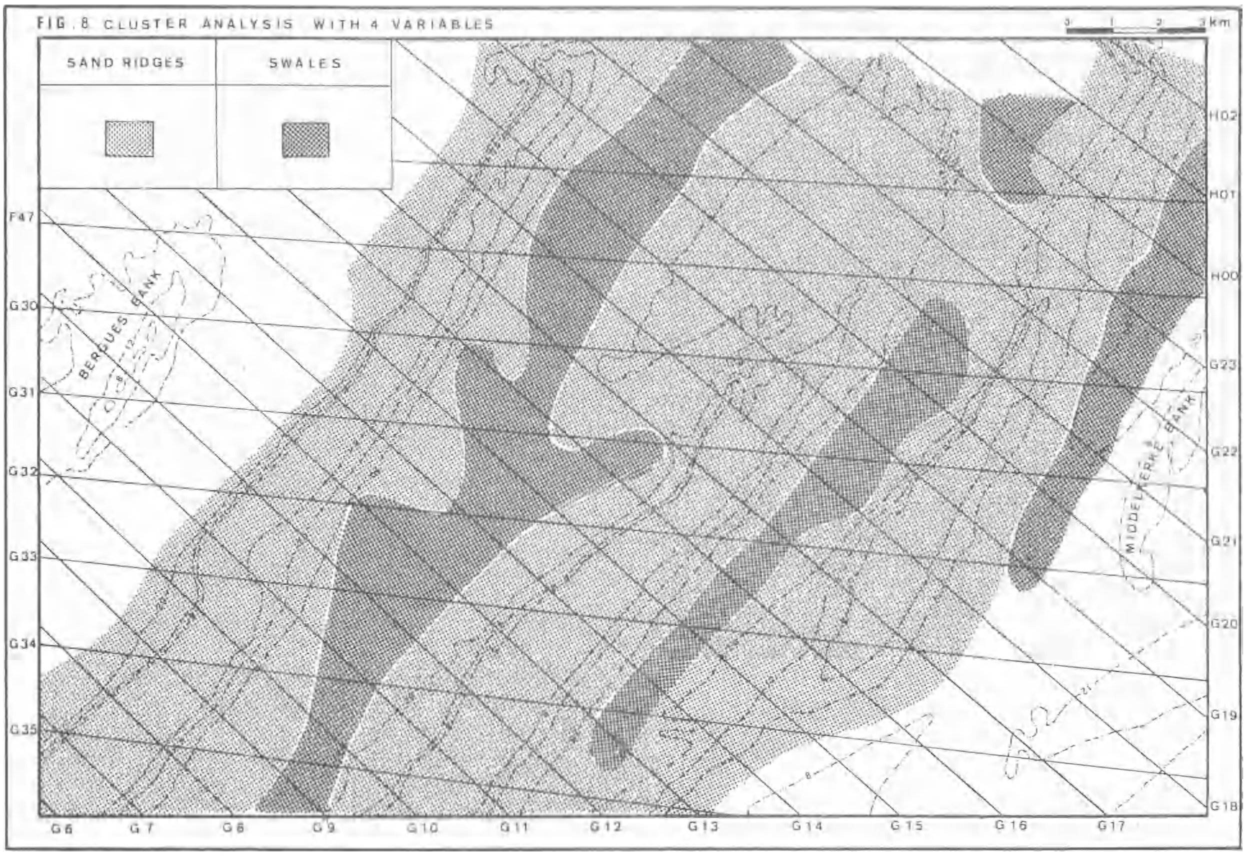

Figure 8 Cluster analysis with 4 variables.

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