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Upper Quaternary marine Skagerrak (NE North Sea) deposits: Stratigraphy and depositional environment

A contribution to OSKAP (Oslofjord-Skagerrak Project) of the Department of Geology, University of Oslo

Special editors for this issue:

Bjørg Stabell, Oslo
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EVALUATION OF DEPOSITIONAL ENVIRONMENT

Editors' note

Over the years *Norsk Geologisk Tidsskrift* has covered many diverse aspects of geosciences mostly related to problems from the area of Scandinavia. From time to time individual issues of our journal have been devoted to one topic focusing on some special problem or area.

This issue of *NGT* contains a series of papers describing a 10 m piston core of Upper Quaternary sediments from the Skagerrak. The authors were brought together by the *Oslofjord-Skagerrak Project* in a joint effort to study this unique record of the younger geological history of a key area of southern Scandinavia.

It is not easy to make a synthesis of all the different studies undertaken, so that the total picture becomes more than the simple sum of the individual contributions. It is, however, hoped that the results obtained may serve as a standard compilation for later studies of similar sediments.

A number of different methods and parameters have been used to date this core and to characterize some major changes of its depositional environment; all measurements and descriptions have been based on the same core, often also the same samples. It should henceforth be possible to correlate the individual measurements and to gain a comprehensive and well-documented understanding of the history of the core's depositional environment.

The Scandinavian land regions surrounding the Skagerrak have for many decades been areas of intensive studies relating to problems of their Quaternary history. The evolution of the depositional environments of the marine areas adjacent to these land regions has, however, for a long time been poorly understood, mainly because it has been very difficult to date this history properly and in sufficient detail. The core presented in this issue is one of very few cores from the marine area adjacent to Denmark and Norway which have been dated successfully and which have been correlated in great detail to the late Quaternary chronostratigraphy.

The Skagerrak is part of a seaway connecting the Baltic and North Seas. By unravelling and dating the late Quaternary history of its depositional environment it has also been possible to resolve certain aspects of the evolution of the adjacent epicontinental and deep-sea areas. Therefore, this core not only documents a history of local importance, but it also opens a perspective for reading and understanding signs which document the geological history of distant areas. Our work on the core also made it possible to date the two youngest acoustostratigraphic sediment units of the Skagerrak.

The sediments encountered in the core contained large amounts of components derived from the Scandinavian land areas. Certain properties of the core and their stratigraphic changes could therefore be used to make statements about the late Quaternary history of the land surfaces in southern Norway and northern Denmark.

The papers about this sediment core should therefore open up a number of perspectives for further detailed studies of the history of the Skagerrak, which some 10,000 years ago was only a fjord opening into a polar ocean, but which since then has developed into part of a wide sea region with very typical geological and oceanographic characteristics and which today is of great influence on the North and Baltic Sea depositional environments. These sea regions are today heavily used by man, and investigations into their young geological history are also important for an evaluation of the stability of this environment. The studies of this sediment core offer some insight into such questions and it therefore seemed appropriate to have them published jointly in one issue of *Norsk Geologisk Tidsskrift.*

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Preface

The submarine geology of the Oslofjord and Skagerrak in southern Scandinavia has been a subject of studies under the Oslofjord-Skagerrak-Project (OSKAP) which is being carried out mainly at the Department of Geology at the University of Oslo/Norway. The scope of the project has resulted in close cooperation with a number of other institutions in Norway, Denmark, Sweden, F. R. Germany and the Netherlands, permitting us to draw on the expertise of many colleagues and to carry out investigations which would have been impossible otherwise. In this study we present data which have been supplied by colleagues from eight institutions, namely the Geology Departments of the University of Oslo, Bergen, Copenhagen and Kiel, the Geological Survey of Denmark (Copenhagen), the Department of Applied Physics of Kiel University, the Institute of Marine Biology of Bergen University and the Department of Chemistry of Oslo University.

The sediment core, whose detailed description occupies a large part of this paper was retrieved during a 1980 cruise (Chief Scientist F. Werner, Kiel) of RV POSEIDON of the 'Institut für Meereskunde' in Kiel. The aim of this study is a very detailed and precise description of the Upper Quaternary depositional environment of the Skagerrak. We have tried to achieve this goal through the application of a diverse set of methods to one carefully selected core. The original 17 contributions have later been supplemented by a few additional studies. It has also been an aim to use jointly the same grid of 18 samples, although some investigators have later chosen to select additional samples.

Due to changes in the sedimentation rates of the cored deposits the artificial selection of the 18 sampling points resulted in a somewhat better documentation of the early record of this core than the later one; however, this turned out to be an advantage since the most important changes of the depositional environment in the study area happened during sedimentation of the lower part of the core. To preserve the individual responsibility of data generated from the core we have chosen to compile a series of papers under individual authorship rather than one lengthy manuscript. However, this approach has resulted at times in a repetition of some general aspects. We have tried to minimize repetitive sections. In a final paper under joint authorship we have tried to synthesize the main results of these studies.

The studies under OSKAP have been supported over the years by a number of funding agencies, in particular in Norway by NAVF (Norwegian Research Council for Science and the Humanities), NTNF (Royal Norwegian Council for Scientific and Industrial Research), Nansenfondet, in Germany by DFG (German Research Foundation). The papers of Erlenkeuser and Werner are part of SFB (Joint Research Programme) 95 'Interaction Sea-Sea Bottom' at the University of Kiel. Part of the stable isotope work was supported within the National Climate Program by the Minister of Research and Technology (DMFT), Germany. DEMINEX has supported the exchange of scientists between the universities of Oslo and Kiel. The help and support of all the above-mentioned institutions are gratefully acknowledged by the authors.

We thank the crew of RV POSEIDON, without whose skilful work this paper would not have been possible. Our special thanks go to Gerd Torjussen and Rønnaug Harnes who laboriously typed all the manuscripts. We are much indebted to Jorun Pedersen and Gisle Nordahl Due for carrying out most of the technical work.

H. Erlenkeuser gratefully acknowledges the expert assistance by H. Liebrenz and H. Reichenbach in sample treatment and ^{210}Pb analysis, by M. Rösler, who operated the mass spectrometer, and by H. H. Cordt, who aided in supervising the intrumental performance.

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I. Th. Rosenqvist gratefully acknowledges the help of Mr. Torgrim Jacobsen, who examined bulk samples, and of Mr. Swinder Singh, who did parallel runs from fractions finer than 2 microns. G. Qvale expresses her sincere gratitude to Dr. John Athersuch (BP Research Centre, Sunbury-on-Thames, England) for his great help with identification of the ostracods and for valuable comments and discussions.

P. Wassmann gratefully appreciates the critique and comments to the manuscript from P. Müller.

The authors also acknowledge gratefully the constructive comments of B. G. Andersen, who has reviewed most of the manuscripts.

The Authors

Introduction

Late Quaternary and modern sediments of the Skagerrak and their depositional environment: An introduction

BJØRG STABELL, FRIEDRICH WERNER & JÖRN THIEDE

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Seismic data have shown that layered Quaternary sediments of up to 200–300 meters thickness cover wide areas of the Skagerrak. Several distinct seismostratigraphic units have been discovered; their acoustic properties are similar within the individual units which can be traced at times across the entire deeper part of the Skagerrak, but which have yet to be studied and dated in detail. A 10 m long sediment core, which penetrated the first clear reflector under a 5–6 m thick apparently transparent sediment unit, is the subject of our very detailed study of the stratigraphy and depositional environment of these deposits.

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Framework of investigations

The Skagerrak is an over 600 m deep marine depression separating the southern boundary of the Precambrian Fennoscandian shield area from the Mesozoic-Cenozoic sedimentary basin further south (Holtedahl & Sellevoll 1971). This basin belongs to the seaways connecting the Baltic Sea through the North Sea with the Norwegian-Greenland Seas. Southern Scandinavia has undergone relative vertical movements of a few hundred meters in total (isostatic as well as eustatic) since the end of the last Glacial which resulted in important changes of the extent and geographic position of these seaways (Mörner 1969, Jelgersma 1979, Stabell & Thiede, this volume) and of the nature of the Baltic environments. However, the Skagerrak doubtlessly has remained a marine basin throughout the entire time span since withdrawal of the ice sheets from Jutland, and until the ice margin reached southern Norway (as documented by end moraines on the coast around southern Norway), see Fig. 1. The history of this very early development remains relatively unknown because of the lack of good sample material.

The sediments which have been deposited within the Skagerrak since the last Glacial have hitherto only been probed to a very modest degree, and the available stratigraphic data are very scarce, especially in terms of their correlatibility to the late Quaternary chronostratigraphy (van Weering 1982). In general, we only know that the uppermost few meters of the Skagerrak sediment cover consist of Holocene fine-grained marine muds which are underlain by Upper Weichselian deposits (Fält 1982, Kihle 1971, Lange 1956). It is very unclear how much sediment the cores which have been described until now represent in terms of time and what record is contained in the underlying layered (Fig. 2) very probably Quaternary sediment section, which according to seismic data in certain areas can be up to 150–300 m in thickness (van Weering 1982).

The successful retrieval of a 10 m long piston core in the central part of the Skagerrak in an area of a clearly visible seismic reflector (Fig. 3) and the apparent chance to date this reflector brought together a group of colleagues of diverse interests to study this core in some detail and to describe the sedimentary record of the central Skagerrak since the last deglaciation.

It seemed important in this attempt to use the same sample material of a core which appeared macroscopically to consist of homogenous fine-grained sediments and which after preliminary tests seemed to comprise a complete stratigraphic record from Recent back to approx. 11,000 years B.P. It seemed important to try to identify

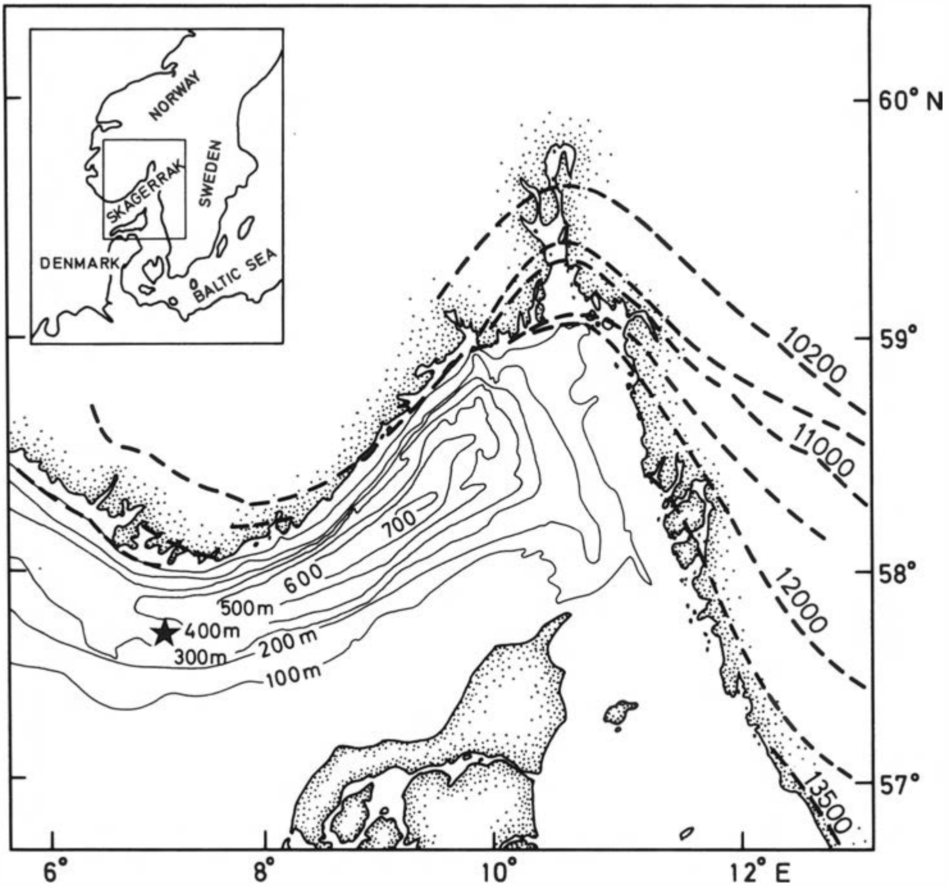


Fig. 1. Bathymetry of the Skagerrak and location of the investigated core (marked by an asterisk). Position and ages of moraines of the last deglaciation after various sources.

as many as possible of the diverse component assemblages observed in the sediment and in the same samples and to try to date the core as precisely as possible by means of relative and absolute stratigraphic methods. We emphasize in particular that we were able to establish a pollen stratigraphy and to identify some of the important 'pollen' events which are well known and well dated in southern Norway, so that we can now correlate the depositional history at the coring site with the Upper Quaternary land records in a quantitative manner which has not been available until now (Henningsmoen & Høeg, this volume).

The modern hydrography of the Skagerrak

The aim of this study is to reach some understanding of the evolution of the depositional environment of the Skagerrak during the time span covered by the stratigraphic record of the core. The youngest part of this record was expected to correspond to modern conditions which will be briefly outlined here.

The main surface water masses of the Skagerrak (Larsson & Rodhe 1979) belong to the Jutland Current which transports North Sea water along the Danish coast into the Skagerrak. It leaves the Skagerrak as the Norwegian Coastal Current with reduced salinities due to the advection of Baltic Current water, and flows along the Norwegian Coast out of the Skagerrak (Fig. 4).

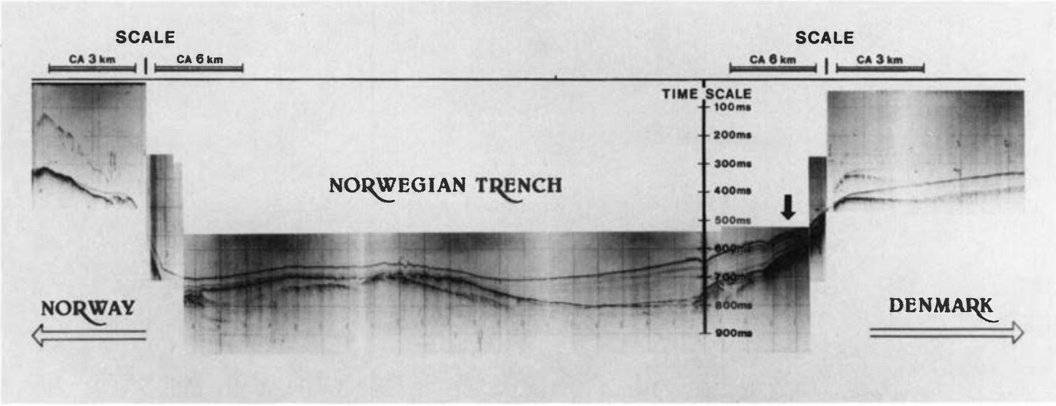


Fig. 2. Seismic reflection line across the Skagerrak/Norwegian Channel between Hirtshals and Kristiansand. Sparker data kindly provided by NOTEBY A/S, Oslo through K. Raaen. Arrow marks approximate position of coring location.

Velocities may rapidly change due to atmospheric influences (Dietrich 1951) and may be as much as 80–120 cm s^{-1} along the Danish coast. On the other hand, bottom water masses move only relatively slowly, at 10–45 cm s^{-1} (Larsson & Rodhe 1979).

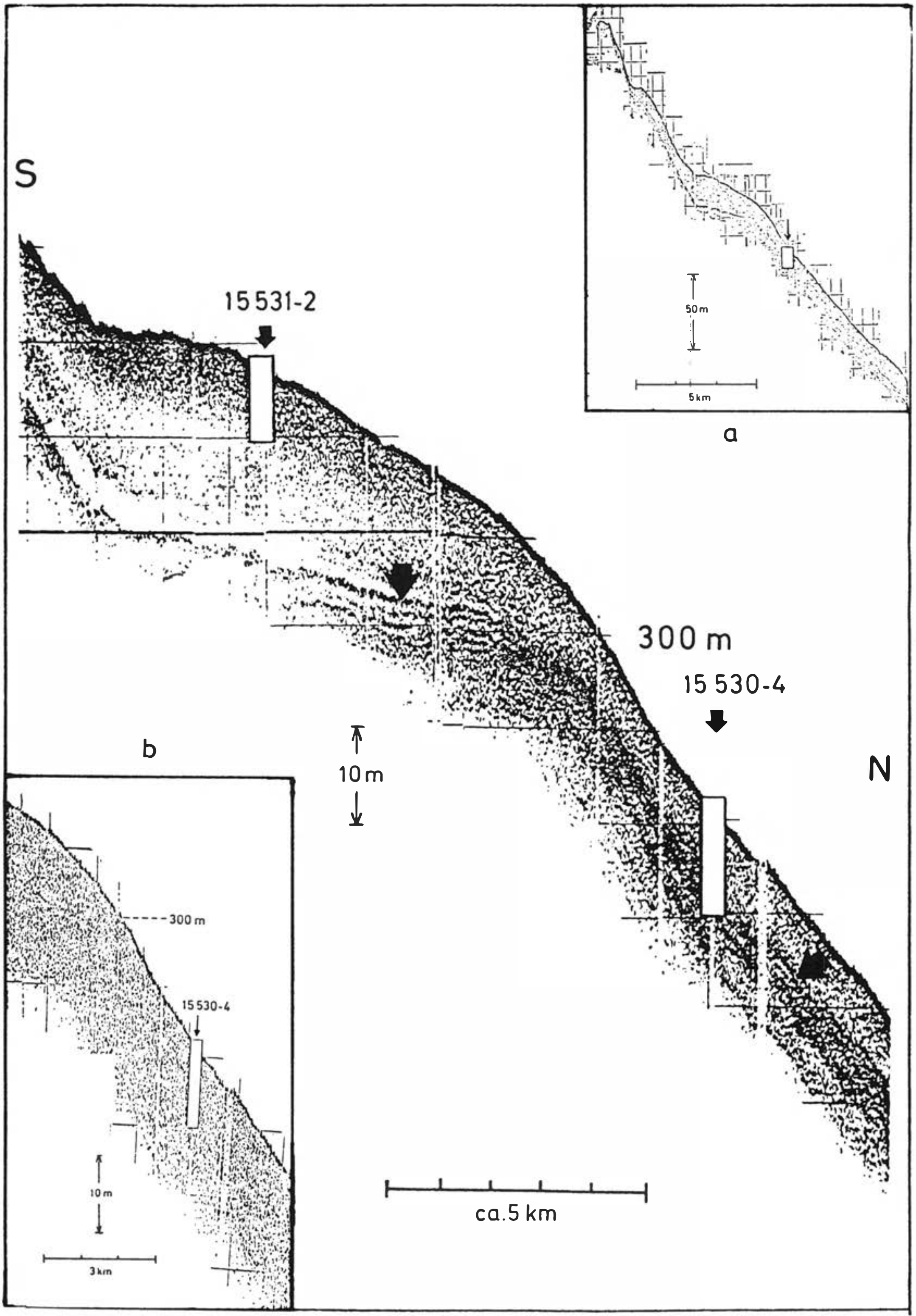
The water column of the Skagerrak is highly stratified almost throughout the entire year (Fig. 5). A stable water mass of around 6°C and 35‰ salinity fills the part of the Skagerrak which is deeper than 100 m. The shallowest part of the water column (in general < 100 m) on the other hand, is seasonally highly variable, both with respect to temperature as well as to salinity. Temperatures range in winter from < 2°C on the northern side to 5–7°C on the southern side of the Skagerrak, whereas during summer they may rise to 17°C and more. Temperature stratification is most strongly expressed during summer time. Salinities are high in the deep and in the shallow southern part of the Skagerrak where the Jutland Current flows (34–35‰). However, along the Swedish and Norwegian coasts salinities may drop well below 33‰ because of the advection of brackish Baltic Current and fresh water. The water masses with lowered salinities leave the Skagerrak as the Norwegian Coastal Current which as a wedge with its thickest part trails the Norwegian coast.

To uncover the signal produced by these water masses in the sediments, we have studied distributional patterns of both planktonic and benthic organisms in the core described.

Cruise details, coring site location, general introduction to the core investigation

Core 15530-4 was sampled 8 November 1980 during a cruise with F/S Poseidon. The core was retrieved by a 9 cm diameter piston corer at 57° 40,0'N, 7° 05,5'E (Fig. 1), after a sediment echogram profile (Fig. 3) had first been taken to decide upon a suitable coring position. The coring position is located at the southern flank of the Norwegian Channel at a water depth of 325 m. Attempts were made to find a locality with both a relatively high rate of sedimentation and giving the possibility to penetrate the top layer and to recover the sediment from underneath the marked reflector which appears on the sediment echograms (Fig. 2). At the fourth attempt a 10.74 m long core was recovered and, as is evident from the detailed echogram (Fig. 3), both requirements seemed to be fulfilled.

The seismograms (Fig. 3) reveal that two, maybe three sedimentary units are present in this area which all drape a deeper lying basement of different, in part unknown nature (Holtedahl & Sellevoll 1971). The upper transparent unit is approximately 5–6 m thick on slopes but reaches more than 20 m thickness over a 5–10 km wide terrace-like flat area at 280 m water depth (Fig. 2). It is underlain by a stratified reflective unit of similar thickness. The stratified unit contains 5–6 reflective horizons of 1–2 m thickness which are separated from each other by transparent sediments of similar thicknesses. Also this unit is thickest over the flat terrace-like area at 280 m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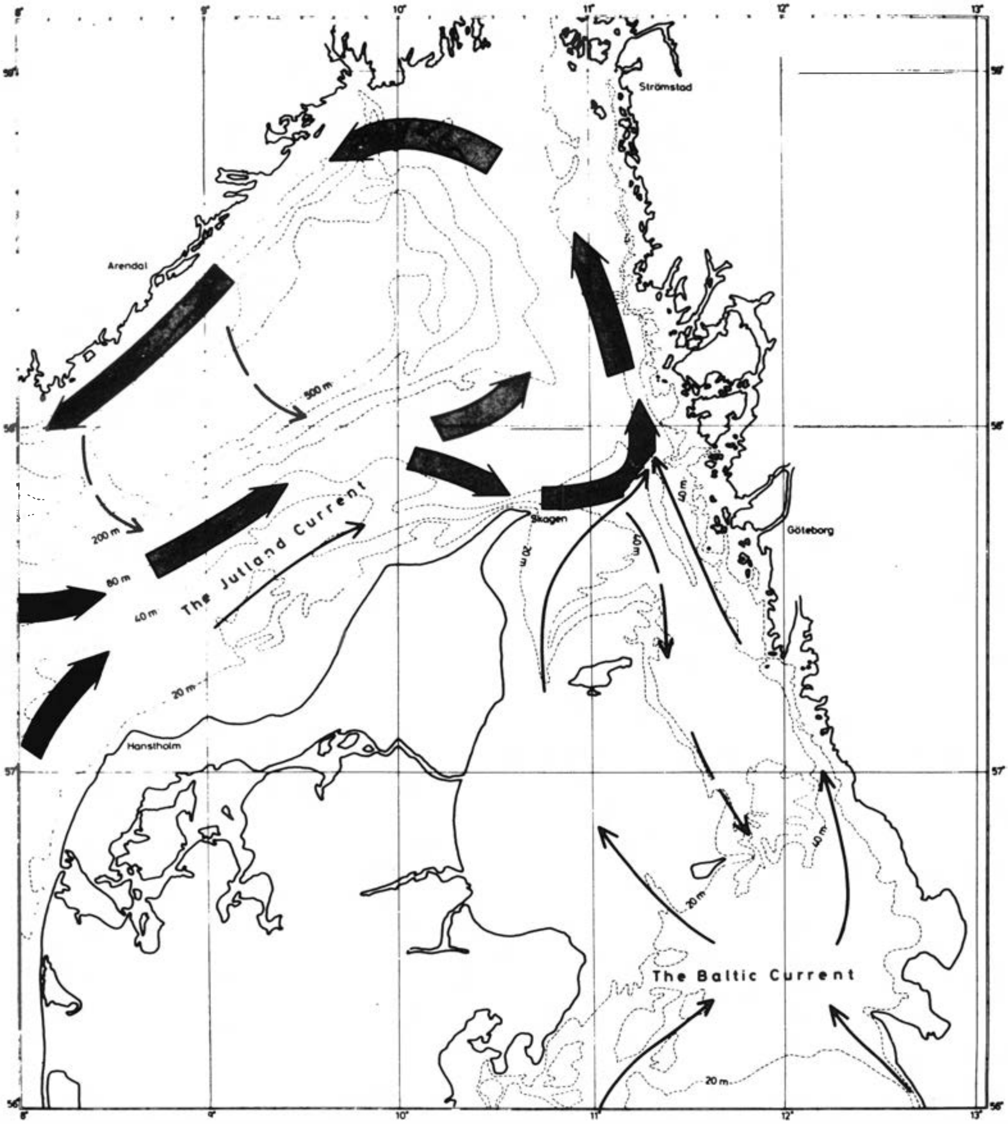

Fig. 4. Surface water circulation in the Skagerrak (from Svansson 1975).

water depth (Fig. 3c). The individual horizons of this unit trail each other parallel. The stratified unit seems to overlie another transparent sedimentary unit whose lower boundary in most areas cannot be seen on the seismograms. It is interesting to note that the seismic unit can be traced across the entire profile, but that the upper transparent unit is lacking above a narrow flat area at 220 m water depth. The upper limit of this unit seems to have been generated by erosion because a faint internal stratification is outcropping in this flat area. The sediment surface in the shallowest part of the profile seems to trail the upper boundary of the stratified seismic unit. The core described in this report penetrated the upper seismic transparent and part of the stratified units.

Fig. 3. Echosounder records (18 KHz) across the Danish flank of the central Skagerrak and across coring location (cf. Fig. 1) close to 07°E. 3A. Entire profile, 3B. Detailed record of 3A, 3C. Marked up portion of Fig. 3B with coring location. Arrow: Main reflector separating transparent and layered sections.

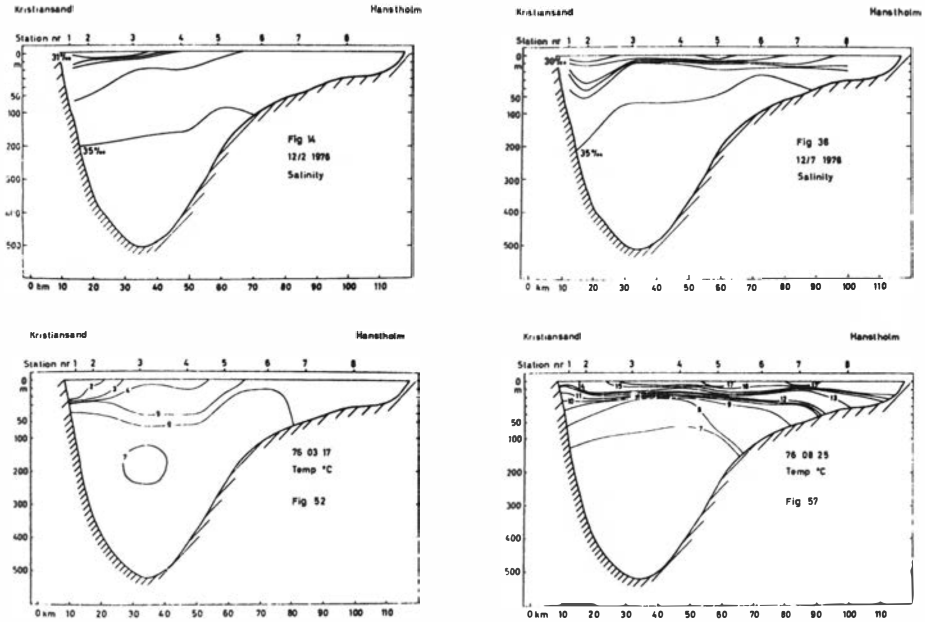

Fig. 5. Hydrographic section (after Larsson & Rodhe 1979) across the Skagerrak along a line from Hanstholm (Denmark) to Kristiansand (Norway).

The core GIK 15530-4 was opened in Kiel in December 1980. Samples were taken every 5 cm. Most papers in this report, however, present data from a set of 18 samples only. These have been selected from the upper very homogenous part of the core in 1 m intervals, from the lower part at 0.5 m intervals. Due to the sampling procedure each analyzed level represents a subsample covering 5 cm; thus, for instance, sample 100 cm represents the interval 100–105 cm. It is obvious that the large intervals between samples only allow a preliminary description of the sedimentary properties and of the stratigraphic boundaries of this core. However, the investigators deemed it important to test at first the stratigraphic qualities of this core, to compare the stratigraphic resolution which can be obtained by studying different fossil groups, and to compare this response to changes of the depositional environment before engaging in very detailed studies of selected intervals of this core.

Description and composition of the bulk sediment of core GIK 15530-4

The core contains homogenous, dark grayish green fine-grained clayey sediments down to 783 cm. Below 783 cm the sediment is pale olive gray with scattered bands of black sulfides down to 890 cm. Scattered mollusc fragments have been found in a well defined interval between 850 cm and 890 cm. At 890 cm there is a sharp boundary to a sediment characterized by zones of black sulfide more uniformly distributed than above.

The smear slide analysis of the 18 samples (Table 1) revealed that the sediments of this core consist largely of terrigenous clays with minor quantities of coarse clastic grains (mostly quartz, feldspar, mica and rock fragments). Most other components (except diatoms, see below) contribute to these sediments in only minor quantities. Of non-biogenic components beside the ones mentioned above, pyrite, micronodules and dolomite rhombs have been observed to occur in small amounts.

The biogenic particles are composed of calcareous, opaline and phosphatic remains. Only diatoms make up an important (up to 10 %) portion of the bulk sediments (they occur frequently only below the 6.6 m-level). Remains of echinoderms, gastropods, benthic and planktonic foraminifers and calcareous nannofossils contribute to the calcareous grain assemblages, whereas the opaline components have been produced by diatoms, radiolarians and sponges. Dinoflagellates, pollen

Table 1. Smear slides (visual estimate in %), x = trace

	Quartz	Feldspar	Clay	Volc. glass	Pyrite	Micro nodules	Dolomite rhombs	Echinoderms	Gastropods	Foraminifers	Nannofossils	Diatoms	Radiolarians	Sponge spicules	Dinoflagellate cysts	Pollen	Plant debris	Fish remains
0- 5	8		90						x				x					
50- 55	5	3	90		x			x	x			x		x			x	
100-105	15	3	80	x	x					x	x	x				x		
200-205	5		90		x	x				x	x		x	x				
300-305	3		95		x	x	x			x	x		x					
400-405	5		90		x		x			x	x			x				
500-505	10		85				x				x			x				
550-555	15		80		x		x			x	x	x		x				
600-605	5		90		x		x				x	x		x				
650-655	5		85		x						x	5		x	x			x
700-705	5		90			x					x	2						
750-755	10		80		x		x					8		1				
850-855	5		90				x					2		x				
900-905	10		85	x	x		x				x	3		x				
950-955	5		90									1						
1000-1005	5		90								x	x		x				
1050-1055	5		90	x			x					x						

Table 2

Grain-size distributions (cumulative weight percent of bulk sediment) in core GIK 15530-4.

>63µ	<63µ	<31µ	<16µ	< 8µ	< 4µ	< 2µ	Sample no.
	100	96	88	74	59	40	0-5
	100	95	89	82	64	51	50-55
	100	98	93	84	70	54	100-105
100	99	89	88	83	70	55	200-205
	100	99	99	88	73	58	300-305
100	98	96	89	81	69	56	400-405
100	99	94	87	76	64	52	500-505
100	99	93	88	76	65	54	550-555
100	98	91	83	70	60	48	600-605
100	99	92	83	76	65		650-655
100	99	95	88	76	65	54	700-705
100	98	90	84	72	60	51	750-755
100	98	92	84	74	63	53	800-805
100	92	81	73	63	52	42	850-855
100	96	92	85	73	60	51	900-905
100	99	99	58	51	43	40	950-955
100	98	90	82	62	59	50	1000-1005
100	99	99	89	76	66	52	1050-1055

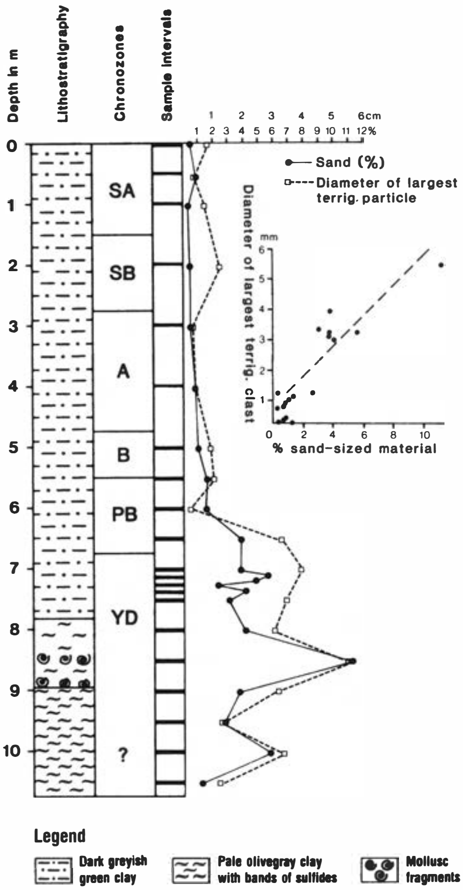

Fig. 6. Contents of sand-sized (>0.063 mm) material and diameter of largest terrigenous clastic (mostly rock fragments) particles in core GIK 15530-4.

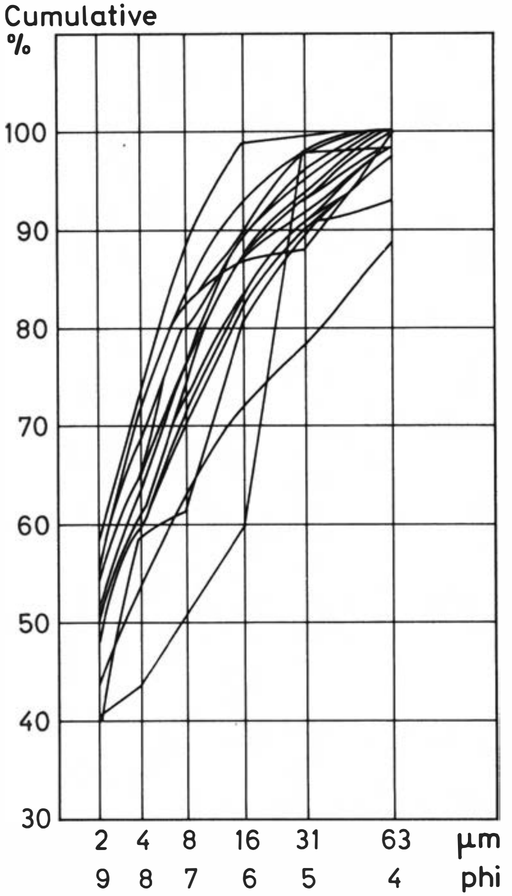

Fig. 7. Grain-size distribution of 18 samples from core GIK 15530-4.

and other plant debris (mostly fibers) contribute to the organic-walled fossil material. Fish bones have also been observed.

The distribution of sand-sized material (Fig. 6) allows us to subdivide the core into separate units, an upper one with sand contents of <1–2% of the bulk sediment, and a lower one with sand contents of 2–>10%. The boundary between the units is situated at 600–650 cm below the sediment surface and correlates to the upper boundary of the subsurface seismic reflectors visible on the seismograms across the coring location (Fig. 3). The sand contents in the lower unit are obviously not evenly distributed, but there is a sequence of horizons with variable sand and pebble contents which probably is not properly represented by the set of samples described in this report.

A conspicuous component of the sand fractions are pebble-sized, terrigenous clastic grains whose maximum size shows a close correlation to the proportion of sand-sized material (Fig. 6). These large clasts bear all characteristics of ice-rafted and ice-dropped material. They may have round or sharp edges and may be composed of quite different materials. They also float in a fine-grained matrix of sediment, although they occur more frequently in certain horizons than in others, creating a distinct stratification (Werner, this volume).

Grain-size distribution in core GIK 15530-4

Grain sizes of the sediments found in core GIK 15530-4 have been studied by means of the pipette method. The main results are given in Table 2 and Fig. 7 for the set of 18 samples which have been selected for this study. They reveal that the sediments throughout the core consist of dominantly fine-grained silty and clayey materials with no important changes in grain size to be observed throughout the core.

Conclusions

1. Seismic data reveal that the young sediments covering the deeper part of the Skagerrak can be subdivided into several acoustostratigraphic units which drape older sediments and rocks of partly unknown origin.

2. A sediment core which penetrated the upper transparent layer and part of a stratified sequence has revealed that the sediments are composed of marine clayey-silty deposits throughout. The fact that they drape a rough subsurface topography indicates that these deposits are composed of sediment particles which have settled through the water column until they reached the seafloor.

3. The lower part of the core comes from a stratified acoustostratigraphic unit which is characterized by variable quantities of ice-rafted material which again appears enriched in certain horizons. The lowermost part of this unit has not been penetrated.

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The physiographic evolution of the Skagerrak during the past 15,000 years: Paleobathymetry and paleogeography

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Stabell, B. & Thiede, J.: The physiographic evolution of the Skagerrak during the past 15,000 years: Paleobathymetry and paleogeography. *Norsk Geologisk Tidsskrift,* Vol. 65, pp. 19–22. Oslo 1985. ISSN 0029-196X.

The evolution of the paleogeography and -bathymetry of the Skagerrak has been reconstructed in a succession of synoptic maps covering the time the area was ice-covered to the present situation. The ice margin withdrew from Jutland and was situated close to the Norwegian coast sometime between 14,000 years B.P. and 13,000 years B.P. The Skagerrak was then filled with marine water but retained a fjord-like shape until about 10,200 years B.P. when the connection to the Baltic Ice Lake across Sweden opened. This seaway closed around 9,000 years B.P., but later a new connection to the Baltic basin opened through the Danish straits. After about 10,000 years B.P. the Skagerrak 'fjord' changed its shape considerably due to the transgression of the large land area which is today located under the North Sea. Its slope along the Norwegian coast, however, has showed only relatively modest changes since that time.

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The Skagerrak is a >600 m deep marine basin between the North Sea and the Baltic Sea, although it is more closely linked to the North Sea. It is located in an area which during the Quaternary was strongly affected by isostatic and eustatic changes, and which has been covered by ice for long periods. This complicated relationship has had a great impact on the geographic and bathymetric evolution of this marine basin. Although the changes can presently hardly be quantified in a proper way, we have made an attempt to develop schematic reconstructions of the paleogeography and -bathymetry of this area for the entire time span since the last Glacial, because we felt that studies of the depositional environment required a certain knowledge of the geographic framework of the basin at different times. A detailed account of the reconstruction will be published elsewhere.The very short description of our results presented here has been prepared to define some of the boundary conditions of the depositional environment documented in a long core from the outer Skagerrak which has been studied in great detail and whose data are presented in a series of papers in this issue.

Methods

The maps describing in a schematical way the extent of the Skagerrak during the last deglaciation (Fig. 1) have been constructed on the basis of ice-margin data from Lundqvist (1961), Mörner (1969, 1979), Andersen (1979) and Sörensen (1979), and sea level data from Lundqvist (1961), Jörgensen & Sörensen (1979), Jelgersma (1979), Mörner (1980), Björck & Digerfeldt (1982) and Fredén (1982).

The paleobathymetry of the Late Quaternary Skagerrak has been reconstructed by using our knowledge of its present morphology as well as of the adjacent land areas, and by applying curves of Late Quaternary relative sea-level change from the area (Henningsmoen 1979, Stabell 1980).

Evolution of paleobathymetry and paleogeography

Although the detailed history is unknown, there seems to be little doubt that the Skagerrak contained a marine depositional environment continuously after the area was deglaciated (Fig. 1). The coring site was ice-covered at 15,000 years B.P. (Fig. 1a). The ice margin withdrew from

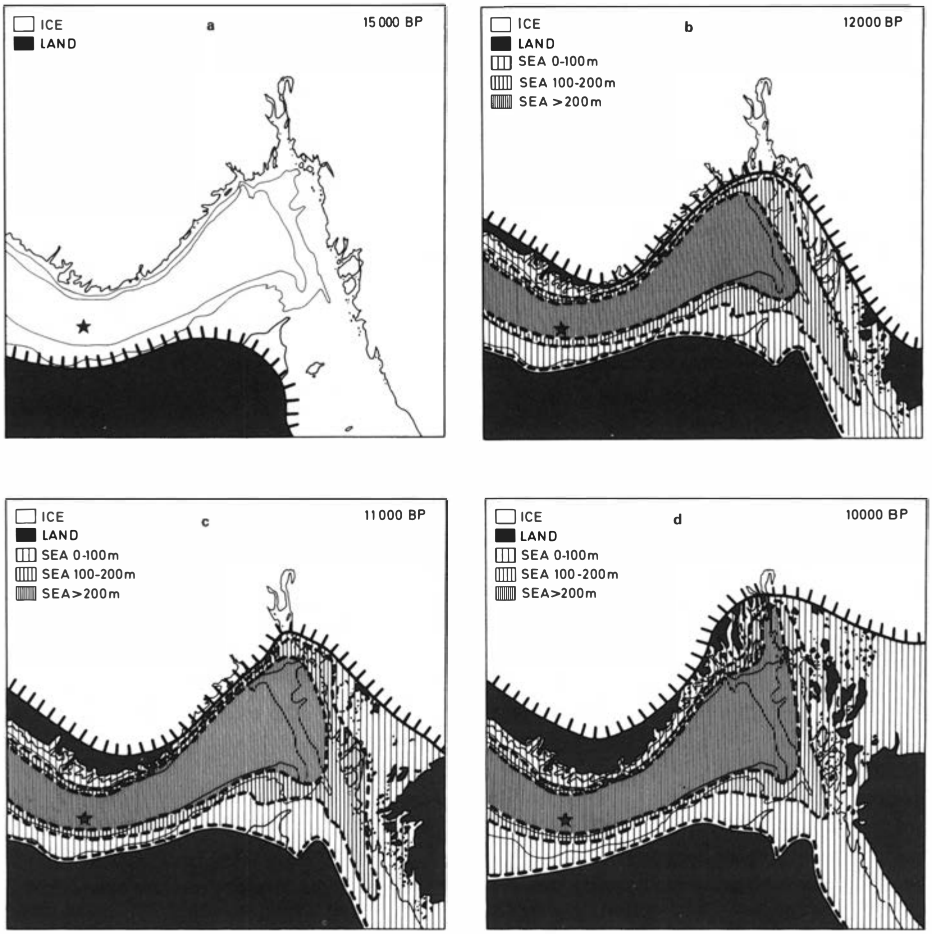

Fig. 1. Evolution of paleogeography and -bathymetry of the Skagerrak 15,000 years B.P., 12,000 years B.P., 11,000 years B.P. and 10,000 years B.P. The location of core GIK 15530–4 is marked by an asterisk.

Jutland and was situated close to the Norwegian coast sometime between 14,000 years B.P. and 13,000 years B.P. The water depth at the coring site was about 260 m at 12,000 years B.P. (Fig. 2), reaching about 285 m at 10,000 years B.P. and the present depth of 325 m at about 5,000 years B.P.

During the deglaciation and up to about 10,200 years B.P. (Fig. 1b, 1c) the Skagerrak was a deep fjord bordered with land areas to the south and a calving ice front along much of the northern and eastern flanks. A bay was situated to the southeast, in an area presently covered by the Kattegat. The Baltic Ice Lake had its outlets to this bay through the Danish straits and across the southernmost part of Sweden. The 100 m depth contour followed more or less the present coastline at 12,000 years B.P., moving inland at 11,000 years B.P. The ice front was fairly stationary along the Norwegian coast during the period 11,000 years B.P. to 10,200 years B.P., but retreated inland in western Sweden.

At about 10,200 years B.P. the ice front had withdrawn from the Billingen Hill, opening a connection between the North Sea and the Baltic Ice Lake. This resulted in a great influx of fresh water from the Baltic Ice Lake to the Skagerrak. Immediately following the drainage of the Baltic Ice Lake, marine water transgraded across southern Sweden, creating the Yoldia Sea (Fig. 1d, 3). The Scandinavian ice front retreated very rapidly thereafter and at about 9,000 years B.P. only remains of the ice sheet were located in some mountain areas. Due to isostatic uplift the con-

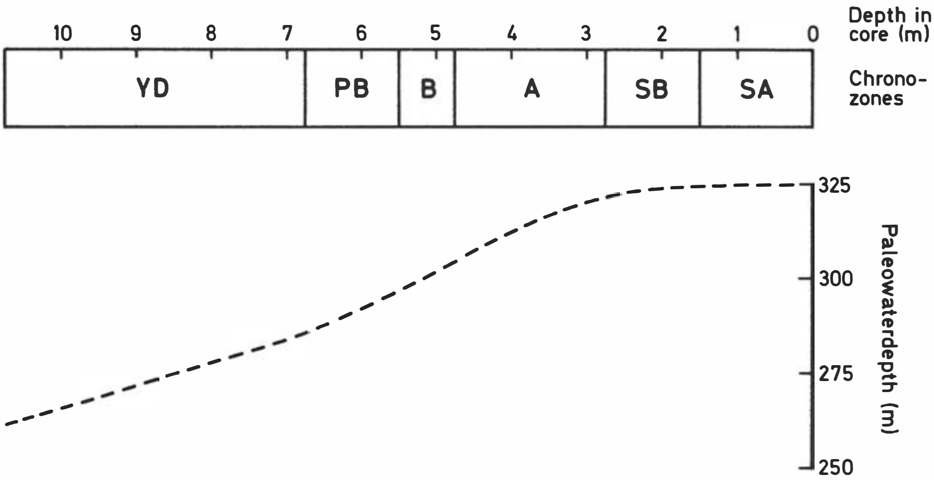

Fig. 2. Paleowater depth curve for coring location. Sea-level data from various sources.

nection across Sweden between the Baltic and the North Sea was closed at about 9,000 years B.P., and the Ancylus Lake was formed. The Ancylus Lake drained through the Danish Straits which were opened due to the eustatic transgression overtaking the isostatic rebound. At about 8,500 years B.P. marine water again entered the Baltic, forming the Littorina Sea.

Since the modern eastern and southern North Sea is generally slightly shallower than 50 m, the large land area to the south of the Skagerrak fjord was rapidly transgressed when the sea level rose above the 50 m isobase (level 50 m below present sea level, eustatic rise). This occurred in about Younger Dryas time. The area therefore started changing drastically at about that time, from coastal area of a fjord to a shallow sea with the deeper Norwegian Channel situated to the north. At about 7,800 years B.P. the English Channel opened, probably initiating a circulation pattern similar to the one at present. The eustatic rise ceased at about 5,000 years B.P.; the Littorina Sea thereafter gradually turned brackish and developed into the present Baltic Sea, see Fig. 2.

Conclusions

1. It is clear from the paleogeographic maps (Fig. 3) that the Skagerrak is the key area for understanding much of the marine evolution of the Baltic area and of the paleoclimate over Jutland and southern Norway since the last Glacial.

2. The Skagerrak was a fjord-like basin directly after deglaciation of the area, until approximately 10,200 years B.P., when the Baltic Ice Lake started to empty into it across central Sweden, and a seaway developed.

3. This seaway closed approximately 1,000 years later, but it was replaced by a seaway through the Danish straits.

4. A major change of the geographic position of the southern coastline of the Skagerrak happened when the former land region west of Jutland was inundated by the transgressing North Sea around 10,000 years B.P.

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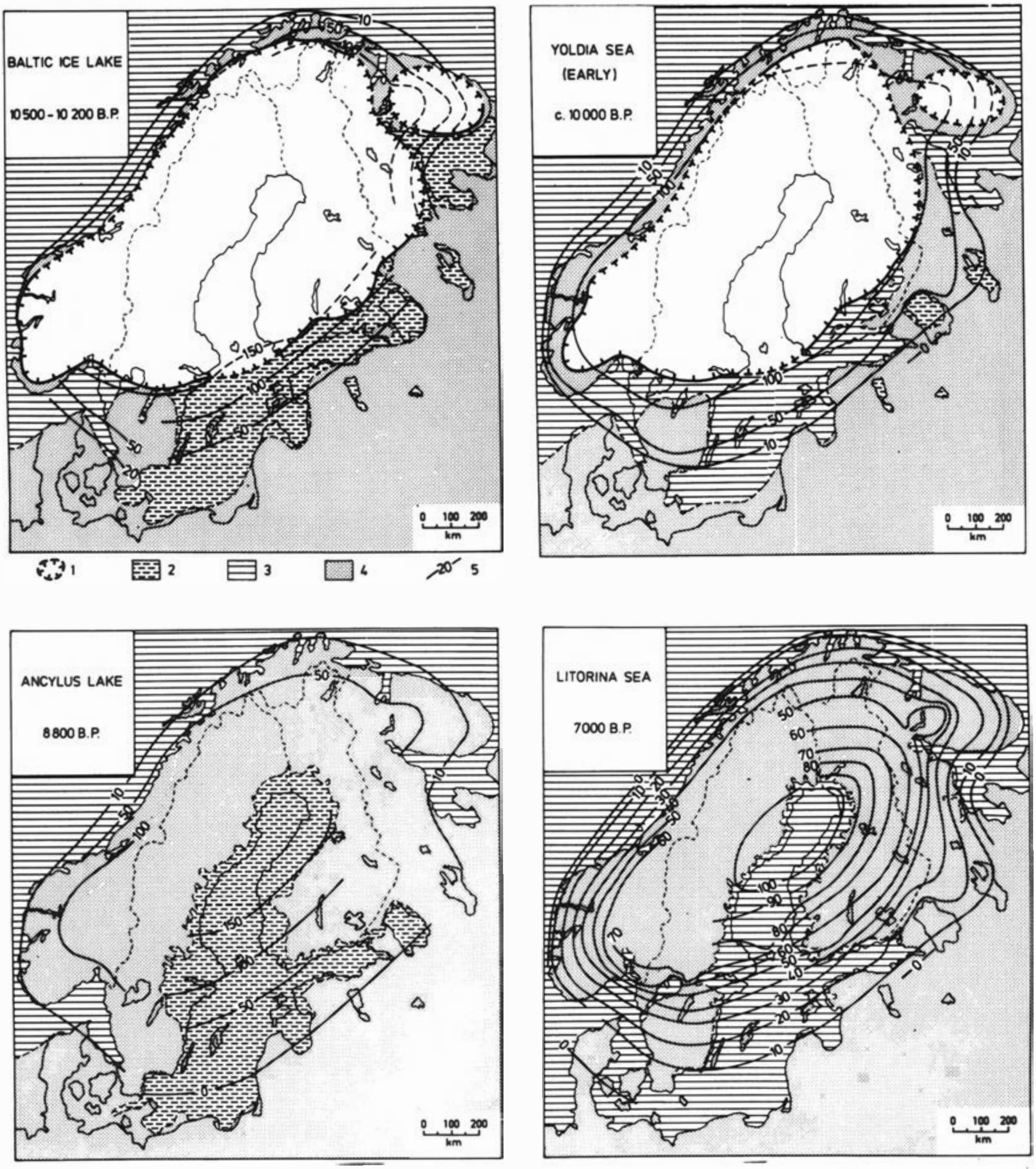

Fig. 3. Paleogeographic evolution of the Fennoscandian region (from Ignatius et al. 1981). 1 = ice margin, 2 = fresh water lake, 3 = marine, 4 = dry land, 5 = isobase with height in meters. Isobases (in meter) show present position of related strandlines with reference to present-day sea level.

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Absolute chronology

Absolute chronology: Summary core GIK 15530-4

The chronostratigraphic division of core GIK 15530-4 is based on data using four different dating techniques. The results are compared in Fig. 1. The ^{210}Pb date of -160 years at 16 cm depth indicates that the sediment surface has been cored without major loss. The division follows in general the system of Mangerud et al. (1974) and the Holocene stratigraphy is based on magnetostratigraphic and pollen-analytical datings.

The boundaries have been fixed based on linear sedimentation rates between the dated levels. The boundaries based on magnetostratigraphy deviate with maximum 50 cm from the pollen boundaries, with the exception of the boundary between Boreal and Preboreal. Here the deviation is 75 cm. For the boundaries between Subboreal and Atlantic the deviation is only 25 cm. With the exception of the Subatlantic/Subboreal boundary (SA/SB), the magnetostratigraphic ages are always younger than the pollen-analytically derived ages. It is possible that this is the case for the SA/SB boundary also, since the pollen-analytically derived boundary might have been placed slightly too low (Henningsmoen, pers. comm.).

The boundary between Preboreal and Younger Dryas (PB/YD) is also defined as the Holocene/Pleistocene boundary. It is placed at 675 cm, even though this level is dated at 10,200 years B.P. according to the pollen analysis. This boundary coincides with the biostratigraphical boundary between a cold water (polar) flora and fauna of low diversity below, and a highly diverse microfossil assemblage which indicates temperate water conditions above.

The Pleistocene part of the core could not be pollen-analytically dated properly, due to a large influx of reworked material. One radiocarbon date at 10,260 ± 280 years B.P. (T-4126) has been obtained on carbonate shells. At about the same level (895–898 cm) a peak in volcanic glass has been found. A similar ash layer from the west coast of Norway has been dated at about 10,600 years B.P., which is in good accordance with the radiocarbon date. The magnetostratigraphic ages from the Pleistocene part seem to be too old. δ^{18}O data indicate that values typical for Younger Dryas are found below 700 cm.

We have encountered considerable uncertainty in determining the age of the lowermost core section. The distribution of ice-rafted material suggested that the maximum of the Younger Dryas had been penetrated and that an older, climatically warmer interval had been reached. Extrapolating sedimentation rates from above suggests that the lowermost sediments are close to 11,000 years old. We therefore believe that the lowermost core section might contain Alleröd deposits; however, we wish to state explicitly that this interpretation is based on stratigraphically very weak data, and that further studies might result in a change of opinion.

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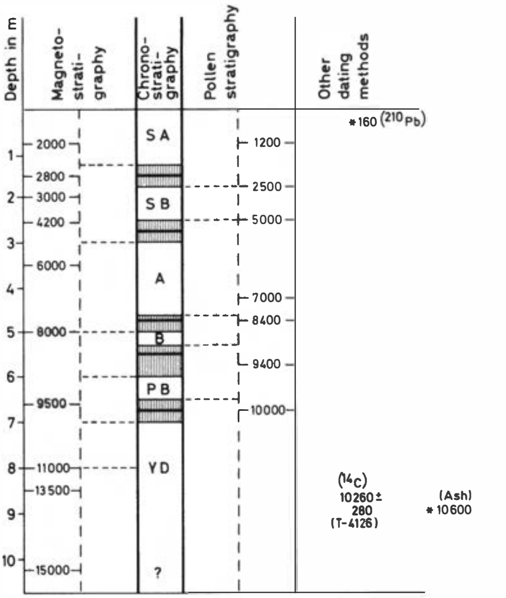

Fig. 1. Distribution of stratigraphic fix points which have been used to determine the chronostratigraphy of core GIK 15530-4.

Distribution of ^{210}Pb with depth in core GIK 15530-4 from the Skagerrak

HELMUT ERLENKEUSER

Erlenkeuser, H.: Distribution of ^{210}Pb with depth in core GIK 15530-4 from the Skagerrak. *Norsk Geologisk Tidsskrift,* Vol. 65, pp. 27–34. Oslo 1985. ISSN 0029-196X.

The distribution of ^{210}Pb with sample depth has been analysed in core GIK 15530-4 from the outer Skagerrak. Recent sedimentation rate was determined from the excess ^{210}Pb profile to about 1 mm/y during the past 160 years. A pronounced long-term variation of the ^{210}Pb background in the Upper Quaternary sediments is likely to reflect the recovery of radioactive equilibrium between ^{226}Ra and ^{230}Th and was used to estimate a mean sedimentation rate of 0.52 ± 0.1 mm/y for the last 3 or 4,000 years.

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The distribution of ^{210}Pb with sample depth in the Skagerrak core GIK 15530-4 has been analysed in order to estimate the sedimentation rates under the Recent environmental conditions and to provide data on the long-term variation of the ^{210}Pb background in the sediments during the Holocene history of the North Sea – Skagerrak depositional environment.

Methods

Due to some post-coring sediment flow, the actual sediment surface remains uncertain within ± 1 cm (range B in Fig. 1). Reference point A indicates the upper core liner rim. For ^{210}Pb analyses 1 cm thick sediment slices were taken every 1 cm between 1 and 27 cm (top sample: –1 to 1 cm; estimated weighted mean: 0.5 cm). 5 cm thick slices were sampled every 10 cm between 28 and 118 cm, and every 20 cm between 138 and 1058 cm. The samples were stored either deepfrozen or dried at 70°C. For ^{210}Pb analysis, 5 g of dry sediment were digested in aqua regia and were leached with 6 N HCl. The ^{210}Pb isotope was measured via its grand-daughter ^{210}Po, which was deposited on silver disks from a 1 N HCl solution adjusted to pH = 1.6. The alpha-disintegrations of ^{210}Po were counted by means of a surface barrier detector. Total counting yield was about 20 %. For details, see Erlenkeuser & Pederstad (1984).

Results and discussion

The results are given in Table 1 and shown in Fig. 1. The relative counting errors are less than 7 % and in most cases better than 4 %. As ^{210}Po and ^{210}Pb can be assumed to be in radioactive equilibrium, the term ^{210}Pb is used throughout the following discussion.

The ^{210}Pb profile reveals an upper section (above 13 cm) with the ^{210}Pb activity exceeding the 'background' found in the strata below. The excess ^{210}Pb is mainly supplied from the atmosphere. A (probably) minor contribution is derived from the decay of ^{226}Ra in the water column. The ^{210}Pb background in the sediment results from in-situ production due to the presence of ^{238}U and its radioactive daughters.

The radioactive decay of the excess ^{210}Pb (halflife: 22.3 y) is used for dating, assuming the initial specific excess ^{210}Pb activity of the sediment as well as the ^{210}Pb background to have been constant throughout the depositional history of the core section of interest (Nittrouer et al. 1979, Erlenkeuser & Pederstad 1984). Typically, the range of the ^{210}Pb dating method is about 100 to 150 y. A surface excess ^{210}Pb activity of 10.1 dpm/g (disintegrations per min per g of dry sediment) and a background of 0.95 dpm/g (broken line in Fig. 1) were chosen for calculating the approximate age scale shown in Fig. 1, upper x-axis.

The presence of excess radiolead indicates that the sediment surface layer has been cored without major loss. The surface excess ^{210}Pb activity, however, appears slightly too low compared to

Table 1: ^{210}Pb data vs. depth. ^{210}Po counts/min/5g dry matter are numerically equal to ^{210}Pb disintegrations/min/g (dpm/g).

Depth (cm)	^{210}Po-content (counts/min/5g)	Depth (cm)	^{210}Po-content (counts/min/5g)
0 - 1	10.352 ± 0.206	24 - 25	1.052 ± 0.037
1 - 2	8.978 ± 0.194	26 - 27	0.956 ± 0.053
2 - 3	6.926 ± 0.146	28 - 33	1.134 ± 0.043
4 - 5	4.326 ± 0.133	38 - 43	1.044 ± 0.027
6 - 7	2.602 ± 0.101	58 - 63	1.164 ± 0.038
8 - 9	1.699 ± 0.055	78 - 83	1.347 ± 0.041
9 - 10	1.478 ± 0.071	98 - 103	1.419 ± 0.043
10 - 11	1.629 ± 0.082	118 - 122	1.280 ± 0.039
11 - 12	1.624 ± 0.065	158 - 163	1.609 ± 0.073
12 - 13	1.255 ± 0.041	198 - 203	1.575 ± 0.043
13 - 14	0.983 ± 0.033	218 - 223	1.468 ± 0.065
14 - 15	0.844 ± 0.050	258 - 263	1.785 ± 0.047
15 - 16	0.907 ± 0.058	298 - 303	1.710 ± 0.041
16 - 17	1.053 ± 0.052	398 - 403	1.663 ± 0.066
17 - 18	1.077 ± 0.042	498 - 503	1.672 ± 0.083
18 - 19	0.937 ± 0.062	598 - 603	1.759 ± 0.026
19 - 20	1.047 ± 0.050	698 - 703	1.837 ± 0.048
20 - 21	0.821 ± 0.044	798 - 803	1.711 ± 0.046
21 - 22	1.016 ± 0.028	898 - 903	1.654 ± 0.026
23 - 24	0.946 ± 0.026	998 - 1003	1.587 ± 0.044

other cores of comparable water depth from the Skagerrak (Erlenkeuser & Pederstad 1984). Distortion or loss of the upper 2 cm layer may account for this finding. The sub-recent background (0.95 dpm/g in the present core) appears to be rather uniform in the argillaceous fine-grained sediments in the deeper part of the Skagerrak (0.8 to 0.9 dpm/g in various other cores, Erlenkeuser & Pederstad 1984).

The sedimentation rate estimated from the ^{210}Pb dates is about 1 mm/y for the excess ^{210}Pb section. Assuming the sedimentation rate to be constant, a model fitted to the data above 24 cm yields a rate of 1.15 ± 0.05 mm/y (Fig. 2). These rates are higher than the long-term average of about 0.6 mm/y derived from palynological and magnetic datings for the middle and late Holocene section of the core (Henningsmoen & Høeg, Schoenharting, both this volume). This faster sediment growth may be related to the lower state of sediment consolidation observed in the near-surface layers (Rosenqvist & Pederstad, this volume).

It is possible, however, that the (formal) sedimentation rates calculated from the excess ^{210}Pb data represent an upper limit, as bioturbation could have mixed excess ^{210}Pb into deeper strata and thus could have affected the slope of the ^{210}Pb profile (Benninger et al. 1979, Olsen et al. 1981, Christensen 1982, Officer 1982, Nittrouer et al. 1984). This effect may be particularly important when the excess ^{210}Pb is confined to the typical depth range of bioturbating organisms, i.e., roughly to the upper 10 cm of the sediment. Indeed, bioturbational effects are possibly indicated by the comparatively high ^{210}Pb values at about 11 cm of depth. So some doubt remains as to the relevance of the age scale and sedimentation rates calculated, and a more detailed evaluation including porosity changes (Rosenqvist &

Pederstad, this volume) and ^{226}Ra ingrowth (see below) have not been performed for that reason.

As shown in Fig. 1, the ^{210}Pb background has not been constant throughout the core. Different reasons may account for this phenomenon.

1. As the concentration of the uranium-supported ^{210}Pb is higher in (or possibly more efficiently extracted from) the clay and silt fraction as compared to the sand fraction, variations of the grain-size distribution will produce variations of the ^{210}Pb content measured (Erlenkeuser & Pederstad 1984). In particular, quartz particles were not digested by the chemical technique we have used.
2. The concentration of uranium and its daughter nuclides preceding the ^{210}Pb isotope varies with provenance and type of the minerals found in the sediment. A change of the source areas from which the sedimentary matter at the coring location was supplied might have occurred during the history of the Holocene sea-level rise (Bjørnstad et al., Rosenqvist, both this volume).
3. Various geochemical processes, dependent on the type of the nuclide considered and in part related to sedimentation rates, affect the behavior and the concentration of the predecessors of ^{210}Pb in the sedimentary matter and the interestitial water. Scavenging of dissolved radionuclides from the water column (Carpenter et al. 1981), redox-dependent transport and sorption processes in the sediment column (Yamada & Tsunogai 1984), diffusional exchange between the pore water and the bottom water (Imboden & Stiller 1982), or leaching from sedimentary source particulates when transported to the site of final deposition (Elsinger & Moore 1980) may have led to concentration gradients in the sediment column or to radioactive disequilibrium between the members of the radioactive family. (Numerous reports on detailed studies of many aspects of these problems have been published since about 1980.)

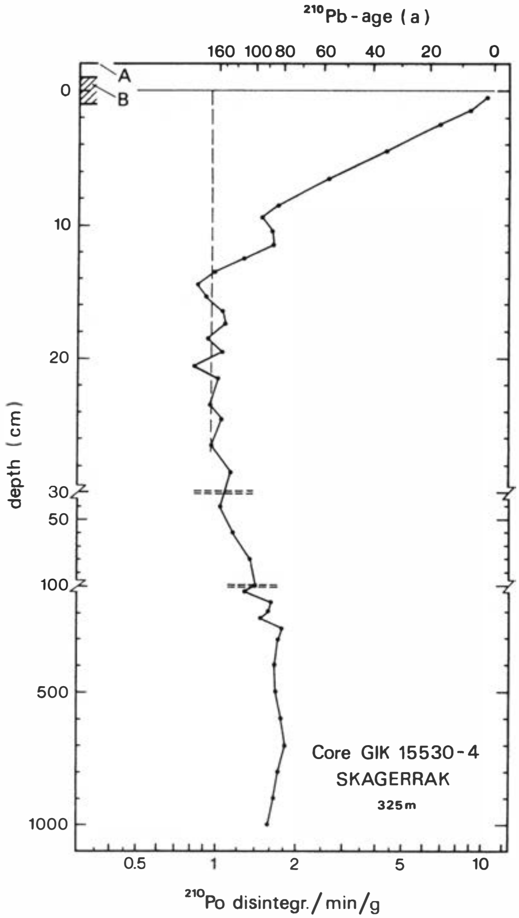

Fig. 1. ^{210}Pb distribution vs. depth in the Upper Quaternary marine sediment core GIK 15530-4 from the outer Skagerrak (57°40,0′N, 7°05,5′E, 325 m water depth). The ^{210}Pb age scale refers to the excess ^{210}Pb section above 13 cm sample depth.

Dating by supported ^{210}Pb

A deeper understanding of the processes most likely to account for the observed systematic variation of the supported ^{210}Pb along the core may be gained by comparing the ^{210}Pb data with the uranium contents analysed by Bjørnstad et al. (this volume). As a result, the ^{210}Pb profile appears to reflect ingrowth of ^{226}Ra (half-life: 1600 y) and provides a means of dating the upper few meters of the core. For easy discussion, the decay scheme of the ^{238}U natural family is shown in Fig. 3. ^{238}U makes up 99.3 % of total uranium in nature.

The correlation between the down-core distributions of total uranium and acid-extractable supported ^{210}Pb is strikingly poor. The most prominent feature of the one curve, i.e., the systematic increase with depth of acid-extractable ^{210}Pb in the 0.3 to 3 m depth interval, or the high level of total U below 8 m sample depth, is not reflected in the other (Fig. 4). This apparent disequilibrium between total uranium and acid-

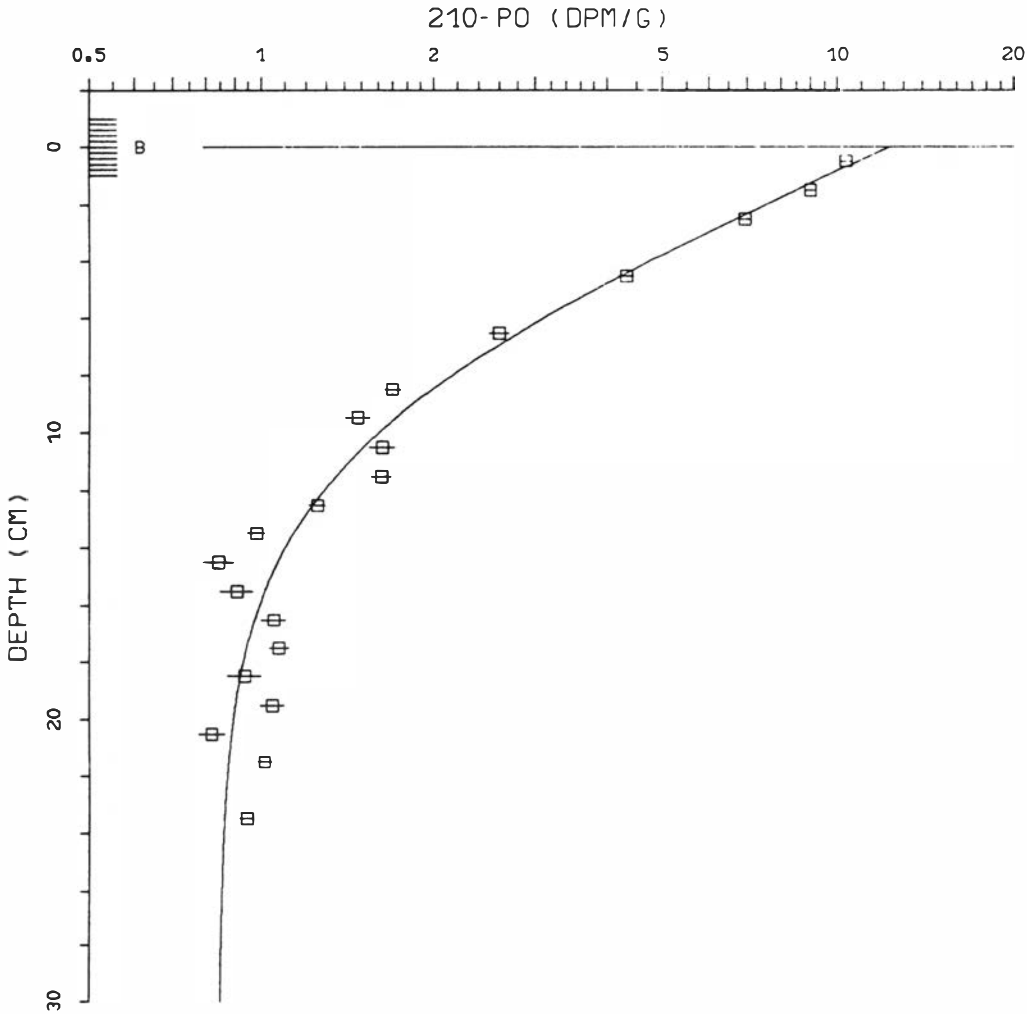

Fig. 2. ^{210}Pb profile and fitted sedimentation model (solid line) for the uppermost 24 cm of the core, assuming constant sedimentation rate (1.15 ± 0.05 mm/y).

extractable ^{210}Pb may arise from a combination of different effects.

1. Only a fraction of the total U – and of the daughter nuclides associated with it – will be accessible to acid leaching (Tilton & Nicolaysen 1957, Pliler & Adams 1962 a, b). A uranium content of 3 ppm (equivalent to a disintegration rate of 2.28 min $^{-1}$ g^{-1}), as was found in the upper part of the core, and an acid extraction yield of 77 % would match the acid-extractable ^{210}Pb activity of 1.75 dpm/g observed below 3 m, if radioactive equilibrium in the ^{238}U decay series is assumed. However, the yield of U (and associated ^{210}Pb) upon leaching depends on the type of the mineral (Pliler & Adams 1962 a, b), and may be quite low for the high-uranium bearing resistant minerals which are thought to provide the U-surplus of the strata below 4 or 5 m in the core (Bjørnstad et al., this volume). Moreover, any post depositional build-up of acid-extractable ^{210}Pb from recoil-^{230}Th (Volckok & Kulp 1957, Kigoshi 1971) or, more critical, from recoil-^{226}Ra will be undetectably small, if these high-uranium bearing particles are in the coarse-silt or sand-sized grain-size classes.
2. Uranium is well known to be dissolved from

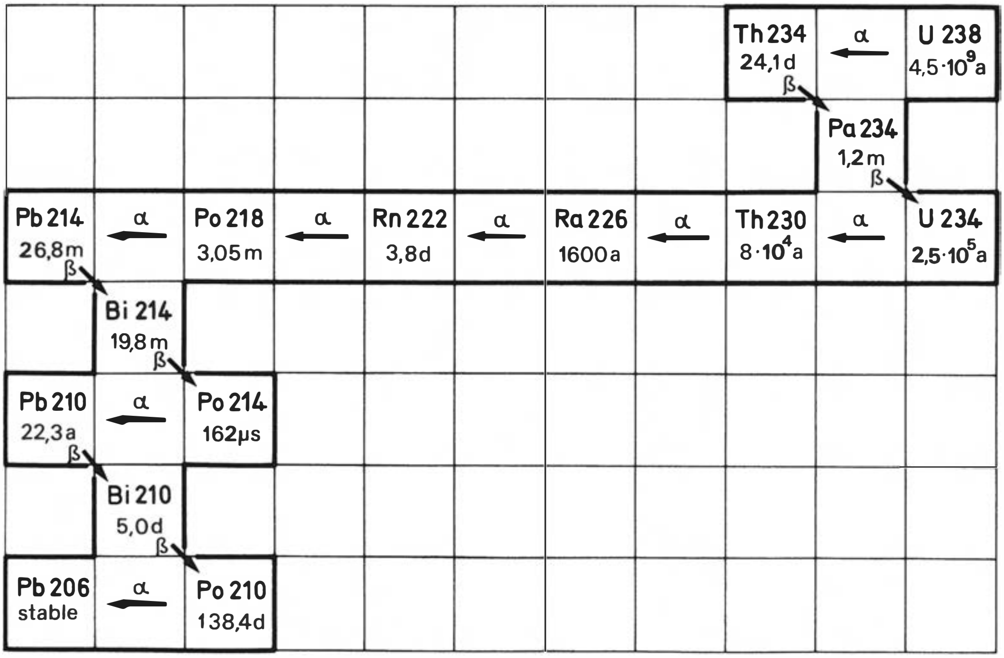

Fig. 3. U-238 natural family (after Seelmann-Eggebert et al. 1974, simplified). Key: nuclide, halflife (a:years, d:days, m:minutes, s:seconds), α, β:decay mode.

solids and to remain in solution for very long times under oxidizing conditions. Chemical weathering in the terrestrial environment thus is likely to have removed much of the uranium from the outer sites of the grains before they enter the marine environment (Martin et al. 1978, Borole et al. 1982). Although the much more reactive ^{230}Th is not likely to be lost by this process, the U-supported ^{230}Th fraction will have decayed off if U-weathering was effective long before the particles were conveyed to the sea.

The correlation between acid-leachable ^{230}Th and total U will further have become offset in the course of time, as ^{230}Th continuously delivered by decay of dissolved ^{234}U is readily adsorbed to particulate matter. For instance, Martin et al. (1978) report activity ratios significantly greater than 1 even for the total-^{230}Th/total-^{238}U pair in suspended matter and sediments of some French rivers. In the ocean, the removal of Th-isotopes from seawater due to scavenging by suspended matter is rapid and effective (Broecker et al. 1973, Aller et al. 1980) and accounts for the high excess ^{230}Th content in pelagic deep-sea sediments. This process, however, probably does not contribute very much to the specific ^{230}Th activity of the Skagerrak sediments, because in the shallow sea environment of the North Sea where much of the Skagerrak sedimentary fines come from, strong bioturbation and frequent resuspension processes greatly enlarge the size of the sedimentary reservoir dynamically participating in the interaction between the sea and the bottom.

3. Uranium, although stably kept in solution under oxidizing conditions, may be precipitated and become enriched in reducing sediments of marine or land-locked basins and estuaries (Sackett et al. 1973, Yamada & Tsunogai 1983 and references therein). High U enrichment has also been found in reducing fjord sediments, and this result may directly bear on the high uranium contents below 8 m in the present core in the sediments of Younger Dryas age, when the Skagerrak was a fjord-like feature (Stabell & Thiede, this volume). Weber & Sackett (1981) considered marine erosion of older U-rich fresh-water deposits and conveyance of this matter to the deep-water sites to account for the increased U-

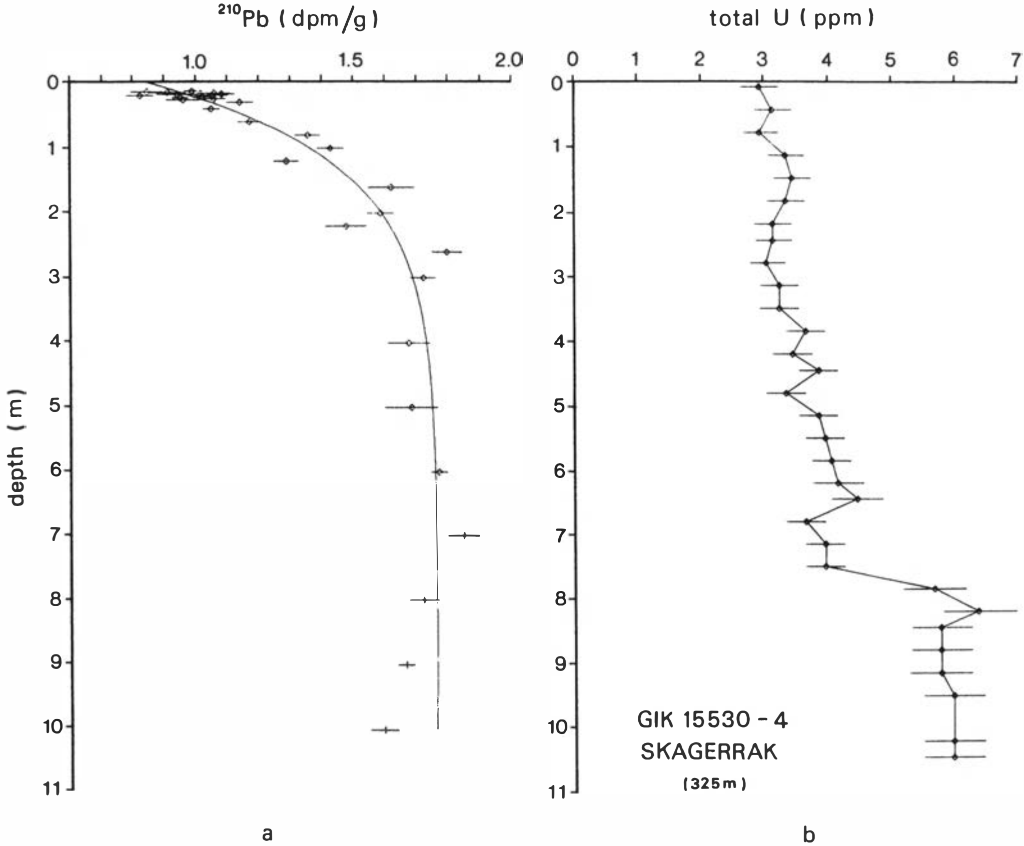

Fig. 4a. ^{210}Pb profile below 13 cm sample depth and fitted sedimentation model (solid line). The model considers the recovery of ^{226}Ra from a (constant) ^{230}Th specific activity at constant sedimentation rate (0.52 ± 0.1 mm/y). + labelled data were not included in the fit (see text).

Fig. 4b. Down-core distribution of total uranium (Bjørnstad et al., this vol.).

contents in the postglacial and early Holocene deep-sea sediments of the Orca Basin, Gulf of Mexico. The simultaneous supply of non-marine organic matter to the deep-sea was diagnosed by the stable carbon isotope ratio, which is significantly lighter in organic matter from terrestrial sites than of marine origin. Such a supply of isotopically light organic carbon is also found in the deeper layers of the Skagerrak core (Erlenkeuser, unpubl.) and parallels to some degree the variation of total uranium.

If the U enrichment observed in core 15530-4 did occur by the 'reducing' pathway, it should have taken place in postglacial times and hence is too young to have led to an appreciable build-up of the long-living ^{230}Th.

It should be mentioned, however, that the radiocarbon dates of about 20,000 y B.P. for the total organic fraction of samples below 7 m (Erlenkeuser, unpubl.) are much too old for this matter to be of postglacial origin, even if a possible hard-water effect of as much as several thousand years in the ^{14}C-age of freshwater deposits is allowed for (Willkomm & Erlenkeuser 1972, Erlenkeuser & Willkomm 1979). It thus appears that a considerable amount of non-marine organic carbon of glacial age, at least, must have been supplied to the early sediments of the present core.

4. As compared to Th, Ra is much less reactive to particle surfaces and, for instance, becomes largely desorbed when solids of terrestrial origin first contact waters of higher ionic

strength (Elsinger & Moore 1980). Moreover, ^{226}Ra has been shown to diffuse from the sediment back into the overlying water (Koscy et al. 1957, Chung & Craig 1973, Moore 1969, Li et al. 1981).

Considering all the arguments given, the leachable ^{226}Ra activity of the sedimentary particulates in the Skagerrak should be deficient as compared to the (leachable) ^{230}Th.

Although ^{226}Ra is comparatively mobile in the marine environment, it may be concentrated by planktonic organisms (Szabo 1967, Shannon & Cherry 1971) and subsequently deposited in the sediments. Koide et al. (1976) referred to this process to account for a twofold excess of unsupported ^{226}Ra over the ^{230}Th-supported background they observed in sediments underlying the high productivity surface waters off California. A similar finding is not evident in the Skagerrak sediments hitherto studied (this paper, Erlenkeuser & Pederstad 1984). Much of the sedimentary fines supplied to the Skagerrak basin originate from the shallower North Sea environment in the south. The lack of organically fixed excess ^{226}Ra in the Skagerrak sediments may therefore relate to the comparatively long time span the fine-grained particulates are available to the repeated processes of resuspension and hence chemical and biochemical attack, before ending up in the deeper lying depositories.

These numerous processes of large-scale mixing by the action of waves and bottom currents may also have had the effect that the concentration of leachable ^{230}Th in the particulate matter exported to the Skagerrak basin has remained rather uniform through the past 6 or 8,000 years at least, when the climatic regime and oceanographic conditions in the North Sea/Skagerrak area were comparatively stable.

According to the foregoing discussion the observed profile of acid-extractable supported ^{210}Pb is considered to reflect the radioactive equilibrium between ^{226}Ra and the parent ^{230}Th becoming re-established with the time since deposition, starting with a nearly 50 % deficiency of ^{226}Ra in the freshly arriving sedimentary matter.

Assuming the initial radionuclide specific activity, a_0, to have remained the same through the length of time considered, the ingrowth of ^{226}Ra may be described as

$$a = a_1 - (a_1 - a_0) \exp\left(-\frac{X}{s\,T_L}\right)$$

with a = specific activity of ^{226}Ra-supported ^{210}Pb at depth x
a_0 = ditto, at depth x = 0
a_1 = ditto, at infinite (or sufficiently great) depth
s = sedimentation rate.
T_L = lifetime of ^{226}Ra, T_L = 2308 a.

In view of the limited number of samples analyzed, the sedimentation rate was assumed to be constant. Second, grain-size effects on the specific activities (Erlenkeuser & Pederstad 1984) were considered of minor importance, as the grain-size composition, particularly the sand content, does not change very much through the middle and late Holocene sections of the core (Stabell, Werner & Thiede, this volume). Third, the radioactive loss by decay of acid-accessible ^{230}Th has been neglected, partly because this effect is not very important for the time range considered, and partly because this loss may be balanced to some extent by recoil-^{230}Th supplied from inner lattice positions of the sedimentary grains (Kigoshi 1971).

The parameters a_0, a_1, and s were obtained from a least-square fit as

$a_0 = 0.84 \pm 0.04$ dpm/g
$a_1 = 1.75 \pm 0.06$ dpm/g
$s = 0.52 \pm 0.1$ mm/y

The fit (Fig. 4a) is based on samples below the excess ^{210}Pb layer and above 6.1. m depth. The deeper lying samples have not been considered because of the different character of the sedimentary environment in the early Holocene and post-glacial time. Also, the possible increase of sedimentation rate in the upper 10 or 30 cm (see above) has not been taken into account, since the value of s does not seem to be seriously affected (but a_0 may be slightly higher than given).

The sedimentation rate obtained is sufficiently high so that effects of diffusional migration of ^{226}Ra within the sediment (Ku 1965) need not be corrected for, if a reasonable diffusion coefficient of 0.3 cm^2/y or less is used for the fine-grained argillaceous sediments of this core.

The model may be used for dating the core over a few half-lifes of ^{226}Ra, i.e., the last 3 or 4,000 years. The sedimentation rate obtained is nicely in agreement with the magnetic and palynological datings given by Schoenharting and Henningsmoen & Høeg, respectively (this volume).

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Shell material in core GIK 15530-4: Its radiocarbon age

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Bivalve shell fragments from 850–885 cm in the core have been radiocarbon dated to 10,260 ± 280 years B. P.

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A shell sample has been dated by the Laboratory of Radiological Dating, Trondheim, Norway.

The sample was collected from the two levels, 850–855 cm and 880–885 cm, with scattered shell fragments which were large enough to be visually observed. Therefore the sample dates the level 850–885 cm. The main part of the sample was an articulated specimen of *Hiatella arctica.* The fact that the valves were still attached to each other indicates an in situ deposition. *H.arctica* is an arctic type, like *Macoma calcarea,* fragments of which were also included in the dated material.

The age of the sample (T-4126) of 10,260 ± 280 years B.P. has been corrected for isotopic fractionation (to − 25‰ relative PDB) and for the reservoir effect of marine water. These two corrections just about neutralize each other. The reservoir age on Recent material from Norway is on the average 450 years (Mangerud & Gulliksen 1975). Olsson (1982) uses an estimated reservoir age of 330 ± 20 for the material from the west coast of Sweden, while the radiocarbon ages presented in the summary of that investigation (Cato et al. 1982) are uncorrected for reservoir age.

It should be noted that these estimates are based on the carbon content in the present-day sea water. The dominant factor in the variation of the apparent age within the oceans is believed to be the circulation of water masses. It is therefore difficult to reconstruct the reservoir age back in time. Mangerud & Gulliksen (1975) assumed that the changes have been small since the Atlantic Current entered the Norwegian Sea, prior to 12,000 years B.P. They do point out, however, that there is a good agreement between dates on marine shells and terrestrial plants from the Late Weichselian and that a systematic deviation can hardly exceed 200–300 years.

The radiocarbon age from core GIK 15530-4 seems to be slightly too young compared with the pollen stratigraphy and the assumed age of the peak in volcanic glass, which is considered to be about 10,500 years B.P. However, this date lies within the limit of one standard deviation for the presented radiocarbon age. Therefore the true radiocarbon age could be close to 10,500 years B.P.

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Magnetostratigraphy and rockmagnetic properties of the sediment core GIK 15530-4 from the Skagerrak

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The paleomagnetic record of the core GIK 15530-4 has been used to establish a magnetostratigraphy which can be related to European paleomagnetic standard sections covering the past 10,000, perhaps even 15,000 years. Magnetic dating of the core is less certain in the Late Weichselian, but reasonably safe in the Holocene parts of the section. A time lag of several hundred years between deposition of sediment and build-up of stable NRM is indicated. Variation of rockmagnetic properties throughout the core is mainly governed by the grain-size variation of the magnetic oxides, with smaller grain size in the Holocene part of the section.

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Paleomagnetic and rockmagnetic studies were conducted on the 10.75 m long continuous core GIK 15530-4 from the Skagerrak. The following aims were pursued in this investigation:

Firstly, to establish the age of the core from the record of stable remanent directions by comparison with the paleomagnetic field of the last 15,000 years recorded elsewhere. Sedimentation rates and possible hiatuses represented in the column might thus be resolved and results compared with other dating methods.

Secondly, to check magnetic overprinting effects by viscous magnetization and/or chemical remanent magnetization after deposition. For this last aim, identification of the magnetic minerals and the variation of these through the column were considered important.

As magnetic properties depend not only on mineralogy and geomagnetic field during and after deposition, but also on grain size, biological and lithological disturbances and diagenesis, measurements were designed to provide, at least to some degree, means for evaluation of those effects. Rock-magnetic results will be reported in detail elsewhere. Summary results and conclusions which seem important with regard to the magnetostratigraphy of the core will be presented in this paper.

Sampling and measurement techniques

The core was cut on the ship into 1 m long pieces. No azimuthal orientation marks common for all core pieces exist. The declination record therefore is continuous only within core sections. However, an attempt can be made to join declination records from adjoining sections using criteria of continuity of magnetic declination.

Samples were taken at 8 to 10 cm intervals from an 8 mm thick slab cut parallel to the core axis for x-ray radiograph analysis (Werner, this volume). Cylindrical 1″ x 1″ polystyrene beakers were pressed into the sediment orthogonally to the core axis, with common but arbitrary azimuthal orientation for all samples within each of the 1 m long core pieces. Each paleomagnetic sample consisted of 3 subsamples, pressed from the same depth interval one after the other into the beaker, which thereafter was sealed at the top and the bottom. Optical inspection as well as magnetic results demonstrated that no disturbance of any importance was introduced by this sampling technique.

Intensities and directions of natural remanent magnetization (NRM) were measured with a Digico Spinner magnetometer. AC-demagnetization was performed at peak-fields of up to 600 Oe to obtain stable remanence directions and coercivity spectra of the samples. Only in a few cases were significant changes of remanence directions found during demagnetization.

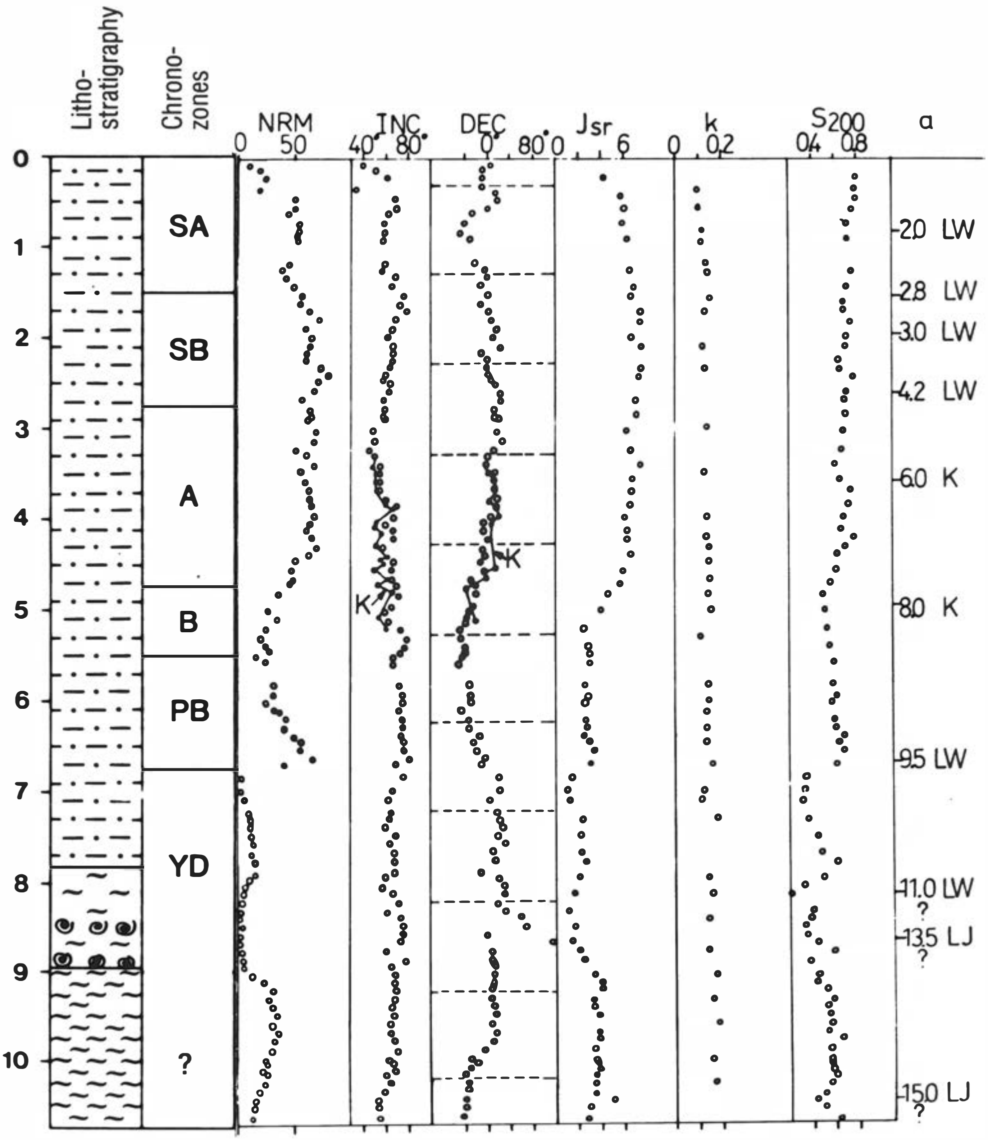

Fig. 1. Magnetic parameters of the GIK 15530-4 core from left to right: natural remanent magnetization (NRM) in 10^{-6} emu/cc, stable inclination INC (K = Kovachenko magnetic record), DEC: interpreted stable declination (horizontal dashed line shows boundaries between core pieces), saturation remanence J_{sr} (arbitrary units), low field susceptibility K in 10^{-4} G/Oe, ratio S_{200} of remanence after 200 Oe ac-demagnetization divided by NRM, 'magnetic age' in 1000 of years B.P. interpreted from correlation with Lake Windermere (LW), Lac de Joux (LJ) or 'Kovachenko' (K) magnetic records. Chronostratigraphic column from Fig. 1 in Absolute chronology, summary (this volume).

Thermal demagnetization was also performed for a number of 12 pilot samples. Measurements of low field susceptibility, saturation remanence and saturation magnetization versus temperature (J_s/T) were conducted to assist in identifying magnetic minerals and estimating their grain sizes and volume percentages.

Results of magnetic measurements

Intensity of the natural remanent magnetization (NRM) varies strongly as shown in Fig. 1 and can be used to divide the core into several magnetic subsections as described below. Comparison of saturation remanence with NRM demonstrates good correlation. This is a strong indication that mechanical disturbance of the core after the build-up of remanent magnetization as well as during the sampling process is negligible. This agrees well with the findings from x-ray radiography results (Werner, this volume).

The stability value S_{200}, the ratio of remanence after 200 Oe demagnetization and NRM, also follows closely the observed NRM pattern, with the exception of depth interval 800 to 900 cm. Susceptibility variation, however, is remarkably small throughout the column. Grain-size variation of the magnetic minerals is interpreted as the major effect controlling NRM and S_{200} relationship.

Both S_{200} values and the ratio of saturation remanence to saturation magnetization indicate that the top 450 cm contain the magnetic minerals which are mostly in single domain magnetic state, whereas below, multi-domain states become more important. The boundary between these two states depends on the grain size of the magnetic minerals and is 0.05 μm for pure magnetite and approximately 1 μm for titanomagnetite (with 35 % magnetite). Pseudosingle domain behaviour might be present for grain sizes about 10 times larger than mentioned above for the same range of composition. Thermomagnetic curves (J_s/T) indicate downhole variation in composition of the magnetic minerals, with Ti poor magnetites/maghemites above 650 cm and titanomagnetites below. The average amount of the magnetic oxides is below 0.03 vol% deduced from magnetic measurements. Considerable contribution to susceptibility stems from paramagnetic minerals. Thus the concept of susceptibility values as a measure of magnetite content has to be discarded for this core.

The following intervals can be differentiated magnetically, based mainly on the NRM record:

1) 0–50 cm: boundary zone of NRM build up;
2) 50–150 cm: undisturbed character of all magnetic properties, NRM established;
3) 150–430 cm: zone of high NRM, S_{200} and J_{sr} values, random noise slightly higher than above;
4) 430–530 cm: boundary zone of decreasing NRM, S_{200} and J_{sr} values (increasing the 'effective' grain-size of magnetites and maghemites);
5) 530–675 cm: increasing NRM, S_{200}, J_{sr};
6) 680–875 cm: small NRM, S_{200} and J_{sr} values with a relative maximum between 730 and 800 cm;
7) 875–920 cm: transitional zone in NRM and J_{sr}, maximum zone for ratio J_{sr}/NRM perhaps related to compositional variation of magnetic minerals;
8) 920–1075 cm: moderately high NRM, S_{200} and J_{sr} values.

Magnetostratigraphy of the core

The variation of stable inclination and declination values with depth can be compared with paleomagnetic records elsewhere for the time interval from 0 to about 15,000 years B.P. to arrive at conclusions about age and sedimentation rates. One of the prerequisites of the method is a reasonable noise-free paleomagnetic record. This is fufilled to a high degree with regard to inclination, except for the uppermost 50 cm. The situation for the declination is worse. The 'best fit' technique to join the adjacent core pieces together is assumed to result in a maximum error of about 20 degrees for each single fit, and the possibility of accumulative errors for the core as a whole makes the fitting attempt appear very speculative.

There are, however, two reasons which give support to the interpretive declination record in Fig. 1: The noise level of declination for each core piece is unusually low, compared to results from both marine and lake sediments elsewhere (for example Creer et al. 1979, Abrahamsen 1982). Secondly, declinations averaged over a time interval of a few thousand years tend to be close to zero. This fact provides strict limits to the fitting method.

There is strong indication that stable NRM in the core GIK 15530-4 is established only below 50 cm depth, suggesting that a time lag exists between stable NRM and the deposition of sediment. This lag may be in part due to settling effects, perhaps combined with minor lithological and bio-disturbances, providing post-depositional realignment (Tucker 1980). It is also probable that maghemitization of magnetite and titanomagnetite takes place within the upper 50 cm or more. This effect is well known for submarine basalts at ambient water temperatures and can effectively overprint a primary depositional remanence. The vexing question as to whether further chemical alteration of the NRM carrying magnetic minerals happens in zones of pyrite formation, is not yet answered. However, judging from correlation results, this effect appears to be of minor importance. The above-mentioned time lag is controlled by chemical conditions during and after deposition and is possibly different between lake and marine environments. Discrepancies in correlation of the order of several hundreds of years have to be expected because of this uncertainty.

Correlation has been attempted with the classical Lake Windermere record (Creer et al. 1979, Mackereth 1971), Lac de Joux near Lake Geneva (Creer et al. 1980) and data from south-east Europe (Kovacheva 1980). The latter was particularly useful for the period between 5000 and 8000 years and good agreement exists there between magnetically determined ages and results of pollen analysis (Henningsmoen & Høeg, this volume).

For the Late Weichselian part of the core, correlation with the record from Lac de Joux is still possible; particularly, a pronounced easterly swing of declination together with high inclination values for Lac de Joux appears to be well correlated with similar values at a depth of 875 cm in core GIK 15530-4. Biochronological dating at Lac de Joux locality provides an age of about 13,500 years for this feature. This poses a problem for our core, as the lowermost part of the core has been given an age of about 11,000 years at maximum (Absolute chronology, summary, this volume), which is 3000 years younger than extrapolated results of paleomagnetic correlation.

If the biochronological dating of the Late Weichselian section at Lac de Joux is correct, then we have either to accept the magnetic correlation and dating, or we have to assume that the secular variation record of the Skagerrak is disturbed by local effects. It is interesting to note that correlation with the Lac de Joux records would result in a hiatus defined in the Skagerrak core of about 1000 to 1500 years, approximately to be placed at 700 cm depth. An alternative, but from a paleomagnetic view-point less likely correlation with the Lake Windermere record, would give an age of 11,000 years to a depth of 825 cm in the core, more in agreement with the general datings of the lowermost part of the core.

In the Late Weichselian record of GIK 15530-4, no short-lived geomagnetic reversal has been found, such as reported by Mörner et al. (1971) from Sweden which has been dated at 12,350 years B.P. This 'Gothenburg excursion' and similar recordings from Lake Erie, N. America (Creer et al. 1976) have later, however, been attributed to sediment slumping effects (Thompson & Berglund 1976) and cannot any more be used unambiguously to define the maximum age of the lowermost part of the core. Further correlation with other cores from the Skagerrak is needed to clarify the problem of Late Weichselian dating and application of the Lake Windermere – Lac de Joux reference section.

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Pollen analyses from the Skagerrak core GIK 15530-4

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Pollen analyses from the core indicate a Holocene age for the sediments down to ca. 675 cm, and a Late Weichselian age for the sediments below this level. The Holocene section demonstrates a vegetational development in accordance with the general development known from the surrounding land areas, and pollen-analytical datings are based on ^{14}C datings from these areas. The *Betula* rise occurs at 675–650 cm, representing ca. 10,200 years B.P. *Pinus* and *Corylus* rise shortly below and above 575 cm, respectively, this level representing ca. 9 400 years B.P.. *Alnus* rises at 500–450 cm, ca. 8,400, and *Tilia* starts at 450–400 cm, probably about 7,000 years B.P.. *Tilia* and *Ulmus* decline shortly below and above 250 cm, respectively, indicating an age of approximately 5,000 at about 250 cm. A *Corylus* decline between 200 and 150 cm may represent ca. 2 500, and the *Picea* rise between 75 and 50 cm occurred at maximum 1 200 years B.P..

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Pollen distribution in marine sediments is considered to reflect the main features of the vegetational history of the inland sources of the pollen (e.g. Robertsson 1982 and literature cited therein). In this case, where land is reasonably near on 3 sides, such an assumption seems justified. The locality of our core is situated about 100 km from the nearest point of Denmark, 250 km from Sweden, and only about 40 km from Norway, cf. Fig. 1.

As pointed out by Groot & Groot (1966) there are special problems implied in pollen analysis on marine, minerogenic sediments, such as low pollen frequency, state of preservation, irregular dispersion etc. The presence of reworked pollen may represent a limitation to the interpretation of pollen in marine sediments. In this case, pre-Quaternary pollen and spores are present throughout the core, but generally only as a few per cent of the palynological material. They are not indicated in the diagram (Fig. 2), due to inaccurate registration of this category. Younger rebedded pollen represents a more serious problem in the core. There are no colour differences characterizing rebedded Quaternary pollen as is generally the case in older material (Stanley 1966). The majority of the Scandinavian pollen types are also the same through at least Eem, Weichsel and Holocene, so plant extinctions are of little help in this question. One has to judge from the context for a tentative separation of rebedded and primary pollen of these types (cf. below).

Bioturbation effects may also represent a problem. In this case, bioturbation is present (Werner, this volume), but it seems to cause only minor vertical disturbances, not least considering the vertical distances between the pollen samples.

Methods

The 28 pollen analysed samples from the core were acetolyzed and HF-treated in the traditional way according to Fægri & Iversen (1975).

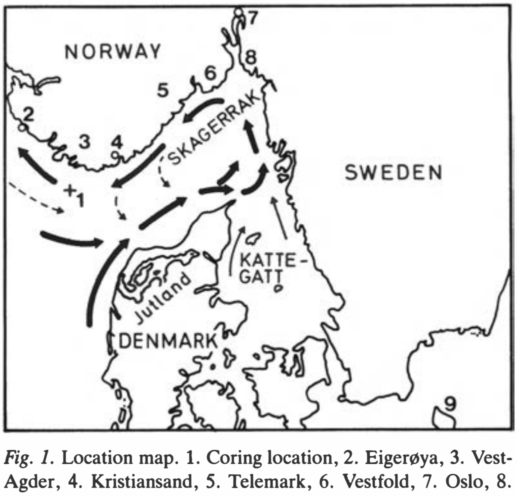


Fig. 1. Location map. 1. Coring location, 2. Eigerøya, 3. Vest-Agder, 4. Kristiansand, 5. Telemark, 6. Vestfold, 7. Oslo, 8. Østfold, 9. Bornholm. Current pattern after Svansson 1975.

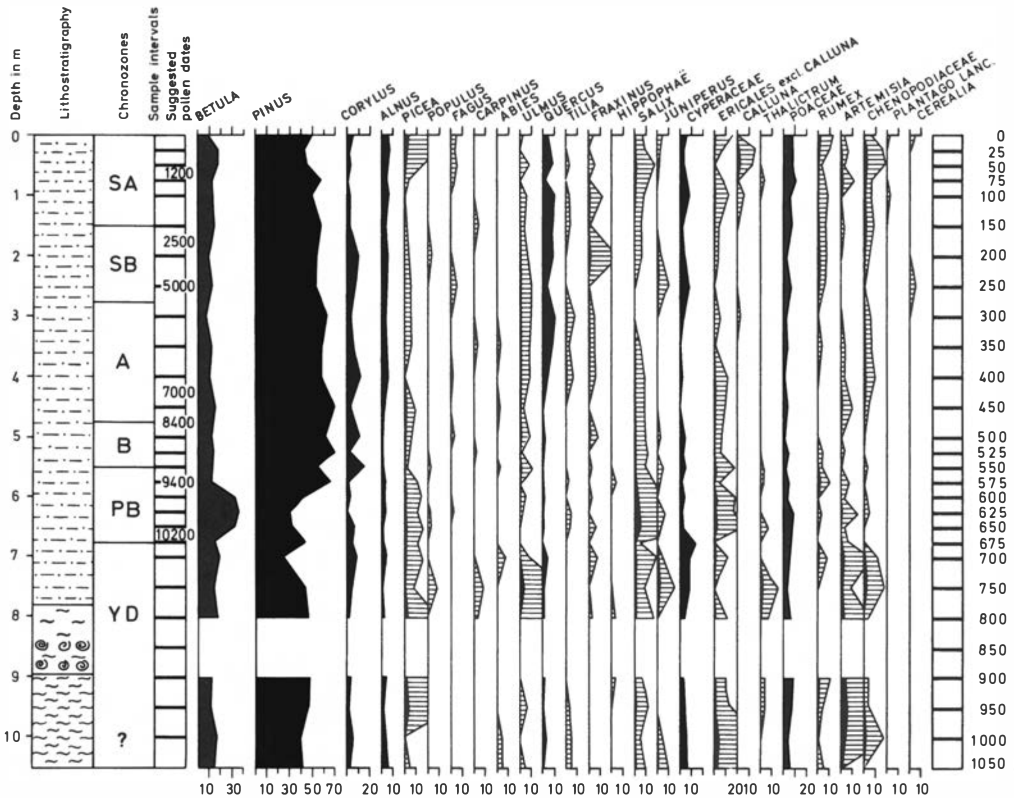


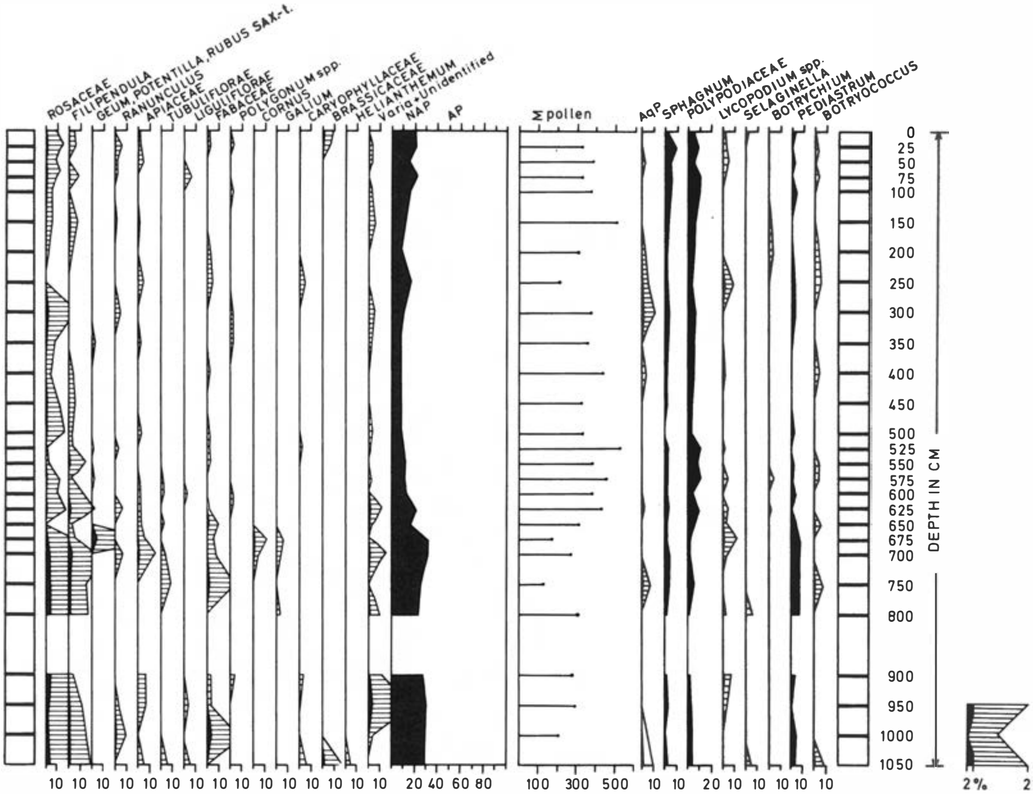
Fig. 2. Pollen diagram, core GIK 15530-4 from the outer Skagerrak. Lithostratigraphy, see Stabell et al., this volume Fig. 6.

(Some previous preparations of similar Skagerrak/Oslofjord material with heavy liquids gave similar or inferior results as compared to the traditional way of preparation.) All samples except the one from 850 cm contained enough pollen for analysis, and a pollen sum of 300 or more was reached in most samples. Pollen preservation was generally good, and no indication of corrosion was necessary in the diagram.

The pollen diagram (Fig. 2) is constructed in a traditional way as a resolved relative diagram (Fægri & Iversen 1975). Two different scales are used, the normal scale being shown in black, the other – 10 × enlarged – is horizontally hatched. All pollen types are included in the pollen sum, except pollen from aquatic plants (AqP). This, as well as spores and algae, is calculated separately with Σ pollen + the taxon in question as the calculation basis. Arboreal pollen (AP) is shown to the left in the diagram, then follows the non-arboreal pollen (NAP), including pollen from dwarf shrubs, wind pollinated and insect pollinated herbaceous plants. The ratio between NAP and AP is also expressed, before Σ pollen and the taxa not included therein.

The pollen diagram below 575 cm

A natural base for interpreting the pollen diagram is provided by the *Betula* maximum between 575 and 675 cm. This maximum is a conspicuous feature in the diagram, and is assumed to represent the Preboreal *Betula* maximum, the beginning of which is generally considered to have occurred 10,200–10,300 radiocarbon years B.P. in the area in question. This is more or less contemporaneous with the transition between Late Weichselian and Holocene (cf. Mangerud et al. 1974), and is an important feature for dating the core pollen-analytically. There is good accordance between this dating and the results from investigations of other fossil groups in the same core material. They, too, demonstrate a conspicuous change in the fossil assemblages be-



(Fig. 2 cont.)

tween 650 and 700 cm (e.g. Dale, Bjørklund, Nagy & Qvale, Thiede, all this volume). The change demonstrates a transition from cold to warmer conditions, and is interpreted as the transition from Late Weichselian to Holocene. A sedimentological indication of cold conditions below about 680 cm is shown by Stabell et al. (this volume); from the sediments below this level they describe pebble-sized terrigenous clastic grains which they classify as ice dropped material.

The *Betula* maximum is accompanied by a *Salix* maximum, a slight rise in Poaceae, and a *Pinus* minimum – in good accordance with a Preboreal date. In addition there are, however, small amounts of pollen from warmth-demanding trees, *Quercus, Ulmus, Tilia, Fraxinus* and *Corylus,* together with *Alnus* and *Picea*. These are considered to be rebedded pollen, originating from morainic material – and also indicating that a certain amount of the other pollen types may be rebedded as well.

The sediments below 675 cm contain even more of the warmth-demanding species than the Preboreal ones do, together with a noticeable amount – 30 to 40% – of non-arboreal pollen, and with a lower pollen frequency than is the case in the material above 675 cm (only estimated, not quantitatively measured). Given that these sediments are of Late Weichselian age, the contemporaneous land areas near the Skagerrak would have been partly ice covered (cf. maps Fig. 1 in Stabell & Thiede, this volume). The vegetation on the ice-free areas would have been rather open, consisting mainly of non-arboreal plants and shrubs, with at most scattered groups of pioneer trees. The pollen from demanding climax forest trees obviously does not represent this vegetation, but is rebedded pollen from an earlier interglacial. Similar occurrences of rebedded material are clearly demonstrated, first by Iversen (1936), and later by, for instance, Donner & Gardemeister (1971) and Hafsten (1963). The relatively high content of non-arboreal pollen in these old sediments may, on the other hand, very well represent mainly the open, contemporaneous vegetation – also regarding the floristic aspect.

The considerable amount of *Pinus* pollen present in the Late Weichselian part of the core is probably not representative of the contemporaneous vegetation. *Pinus* is considered to have grown in south-eastern Sweden (Berglund 1966) and on Bornholm (Iversen 1967) during parts of the Late Weichselian, i.e. during the favourable Allerød time. Parts of the above-mentioned pine pollen may come from here, but most of it probably is rebedded. *Pinus* is often very well represented in morainic material, as shown already by Iversen (1936), and is thus likely to be found where such material is deposited. Corresponding results from Late Weichselian material are presented by de Jong (1981). In his cores from the Skagerrak and the North Sea, he finds high NAP values (ca. 30%), dominating *Pinus* and high values (ca. 10%) of warmth-demanding trees in Late Weichselian sediments.

There seems to be no systematic variation in the pollen flora within this lower, ca. 4 m thick sediment section of the core, and no conclusions can be drawn from this material concerning the Late Weichselian vegetational or climatic changes. During Late Weichselian time, when eroded material was amply supplied by the melting ice not far away, rebedded pollen material probably played a greater role in the sediments than it did later on. Moreover, the influence of rebedded pollen will be very important in a pollen diagram where the contemporaneous pollen production is low, as was presumably the case when the lower section of the core was deposited. The Preboreal vegetation produced considerably more pollen than the Late Weichselian one because it was denser and more arboreal, and also because the ice retreated far inland in Norway and Sweden during this time, and left large areas to be very rapidly covered by vegetation. This means that the influence of rebedded and long-distance transported pollen would be relatively less in Preboreal sediments than in Late Weichselian ones, although it is still noticeable, cf. the elements of warmth-demanding trees still present together with the *Betula* maximum.

The diagram above 575 cm

Rebedded pollen is certainly also present throughout Boreal and younger sediments, but the contemporaneous pollen production was then probably large enough to dominate over the rebedded contribution. This seems corroborated by the fact that the pollen curves from the Holocene section follow the general scheme which has been demonstrated for the area around the Skagerrak-Kattegat by several pollen-analysts (Iversen 1967, Berglund 1966, Nilsson 1964, Fries 1951, Hafsten 1956, Danielsen 1970, Høeg 1982a, and others). Consequently, the pollen curves from the Holocene part of the core are assumed to give a far better reflection of the vegetational development on the surrounding land than is the case from the Late Weichselian part.

The *Betula* maximum is followed by an increase in *Pinus* and *Corylus,* interpreted as representing the transition to the Boreal time. The high *Pinus* percentages in the Holocene section of the core are certainly partly representative of pollen production on adjacent land, but they also probably partly demonstrate an overrepresentation, as is often found in marine sediments (e.g. Fægri & Iversen 1975, Traverse & Ginsburg 1966). This buoyant pollen type is apt to be transported over long distances through the air and in water.

The *Corylus* rise is dated to ca. 9,000 radiocarbon years B.P. in western Jutland (Iversen 1967), ca. 9,900 in southern Sweden (Berglund 1966, Nilsson 1964), ca. 9,200 in the Oslo area (Nydal et al. 1970, Hafsten 1972), ca. 9,600 in western Sweden (Påsse 1983), ca. 9,300 in Vestfold (Henningsmoen 1980), ca. 9,400 in the Kristiansand area (Høeg 1982a) and ca. 9,900 on Eigerøya (Simonsen, referred in Thomsen 1982). It is reasonable to assume an age of the *Corylus* rise (575–550 cm) in the core of about 9,400 years B.P., in accordance with the corresponding date from the nearest land area.

The *Alnus* rise occurs between 500 and 450 cm. According to Wenner's summarizing (1969) of this event, the tree established itself astonishingly contemporaneously over large parts of Sweden, viz. some hundred years before 8,000 years B.P. This is in agreement with Danish (Iversen 1967) and Norwegian reports, e.g. for the Oslo area ca. 8,200 (Hafsten 1972), Vestfold between 8,000 and 8,700 (Nydal 1959, 1962, Henningsmoen 1979a), the Kristiansand area ca. 8,400 (Høeg 1982a). An age of ca. 8,400 years B.P. is assumed for the *Alnus* rise in the Skagerrak core – again in accordance with the date from the nearest land area.

Tilia is present in very small percentages, and its occurrence is statistically uncertain. It rises between 450 and 400 cm. This indicates an age of about 7,000 years B.P. in view of the reports listed in Høeg (1982a), spanning from ca. 6,400 in Østfold to ca. 7,700 in the Kristiansand area. From southern Sweden there are corresponding dates of about 7,200 (summarized in Eriksson 1979), and nearly 8,000 years B.P. from Denmark (Iversen 1967). An age younger than the ones from Kristiansand and Jutland is tentatively used here, not least because there is about 1 m sediment between the rise in the *Alnus* and *Tilia* curves.

Tilia declines between 300 and 250 cm; a similar *Tilia* decline has been radiocarbon dated from Vestfold to about 5,500 years B.P. (T-2433 in Henningsmoen 1979a). *Ulmus* declines slightly later, between 250 and 200 cm; a corresponding *Ulmus* decline is found in the Oslo and Østfold areas (Hafsten 1956, Danielsen 1970) and dated in Østfold to about 5,000 years B.P. (Griffin et al. 1980). It is interpreted as representing the transition between Atlantic and Subboreal times, more or less corresponding to the chronozone transition (Mangerud et al. 1974). At about the same time – or slightly later – the first occurrences of 'cultivation' indicators are found in the above-mentioned areas, and also in Telemark (Høeg 1982b). Similar development is shown in diagrams from Sweden and Denmark (Berglund 1969, Iversen 1967, Troels-Smith 1982) for the same time, and the level about 250 cm is assumed to represent a time around 5,000 years B.P. A single cereal pollen was found at level 250 cm in our core; this is in good accordance with the aforesaid.

The transition Subboreal-Subatlantic represents a climatic deterioration, generally considered to have occurred about 2,500 years ago. This transition is very vaguely represented in our material. One would expect the Subatlantic climate to manifest itself by a decrease in the warmth-demanding trees. Such a decrease is very vague, but it is possible that the *Corylus* decline between 200 and 150 cm reflects this deterioration.

Some per cent of *Picea* pollen are found throughout the Late Weichselian and Preboreal parts of the core, and – in smaller amounts – through the rest of the Holocene part; not until near the level 75 cm does the *Picea* curve rise again. The Late Weichselian and Preboreal *Picea* is considered to be redeposited, and the same may be the case with the small amounts of the Holocene *Picea*. The *Picea* from the upper 75 cm, on the other hand, probably reflects the establishment of the species in southern Norway and western Sweden. Since it did not establish itself in southernmost Sweden and Denmark in the Holocene, these areas may be excluded.

Picea invaded Scandinavia from the north-east, and it was established in Vestfold about 1,200 B.P. (Henningsmoen 1979b). It was established in Østfold (Griffin et al. 1980) and in southern Telemark at about the same time, but a considerable delay in *Picea* occurrence is demonstrated towards the north-west in Telemark (Høeg 1978). We do not know when it arrived further south along the Norwegian coast, but since Fægri (1950) found a very recent *Picea* invasion at its present occurrence limit in Vest-Agder, we may suppose an invasion delay from Vestfold/Telemark towards the Kristiansand area. This means that the *Picea* occurrence in the Skagerrak core may be at most about 1,200 years old, but perhaps some hundred years younger.

Fagus is more weakly represented in the material than *Picea,* but it has a continuous curve which corresponds to the *Picea* curve in the upper part of the core. *Fagus* occurs spontaneously in a rather restricted area in Norway, mainly in Vestfold and along parts of the southern coast. In Vestfold, *Fagus* and *Picea* were established more or less at the same time (Henningsmoen 1980), whereas its first occurrence along the southern coast is unknown. In Jutland, *Fagus* is present in small amounts from late Subboreal and it shows a clear increase about 1,400 years B.P. (Iversen 1967). Danish *Fagus,* consequently, may appear earlier in our core than *Fagus* from Norway. The rather simultaneous occurrence of *Fagus* and *Picea* in the diagram may indicate a mainly Norwegian origin of the upper *Fagus* pollen grains, whereas the few older grains may come from the south or be redeposited.

Pollen transport

An important question – but difficult to answer – is: From where did the pollen come to the core locality, and how? As early as 1919, Hesselmann demonstrated the ability of the pollen grains to get transported by wind over long distances. He studied the pollen rain on the Gulf of Bothnia, and found considerable amounts of pollen in pollen traps stationed on board ships. His results have later been corroborated and supplemented by numerous other scientists.

Wind may have been the main transporting agent for the pollen material in the core, and all the areas around the Skagerrak/Kattegat are possible sources for wind-transported pollen. The diagram (Fig. 2) seems, however, to have a mainly northern rather than a southern character. The *Pinus* dominance throughout the Holocene – although partly a result of marine overrepresentation – may indicate a northern origin (cf. for instance the Boreal diagram from Doggerbank in Behre & Menke (1969), where the relative amount of *Corylus* compared to *Pinus* is far higher than in corresponding parts of our core). The same applies to the presence of *Picea,* and to the very low percentage of *Fagus* in relation to *Picea.* As stated above, the Norwegian south coast, being the nearest land area, is assumed to be the main contributor to the airborne pollen rain.

However, water transport by currents obviously also took place, as indicated by the presence of rebedded pollen, where water transport must have been at work at least during the last phase. The same applies to the freshwater algae *Pediastrum* and *Botryococcus.* Whether the algae originally came from streams and rivers, or from morainic material, they must finally have been transported to the coring locality by water currents and not by air.

The main current pattern (Fig. 1) allows transport from both northern and southern directions. The admixtures of pre-Quaternary pollen point to some transport from the south and the west, and the dinoflagellate cysts as well as coccoliths give stronger indications of such transport (Dale, Mikkelsen, this volume). The mineralogical investigation (Rosenqvist, this volume) indicates that the whole core contains material from both south and north, but with the southern elements dominating in the Holocene section, and with a considerably larger northern contribution in the Late Weichselian than in the Holocene section. A northern origin of the greater part of the Late Weichselian sediments would agree well with our assumption that a large part of the pollen in these sediments comes from morainic material. In this case, the transporting agents for this part of the pollen must have been ice and water, whereas the primary pollen may have been mainly wind transported.

In conclusion we tentatively suggest that a large part of the pollen comes from the northern side of the Skagerrak, that wind transport was important, and that water transport probably took place both from northerly and southerly directions.

Dating methods, comparison of results

The magnetometric measurements (Schoenharting, this volume) agree fairly well with the pollen datings. Best correspondence is found about 5,000–8,000 years B.P., whereas the youngest datings represent a certain deviation (cf. Fig. 1 in Absolute Chronology, summary, this volume). As described above, the pollen-analytic datings in our core are based on radiocarbon datings of pollen zones from surrounding land areas. Other factors, like statistical errors and admixture of rebedded pollen, add to the uncertainty; consequently, discrepancies in correlation of some hundreds of years may be expected from the pollen-analytical point of view. Corresponding views concerning the magnetometric datings are expressed by Schoenharting (op.cit.).

The radiocarbon shell dating 10,500 years B.P. from level 850–885 cm (Stabell, this volume) as well as the ash dating ca. 10,600 from 895–898 cm (Bjørklund, this volume), agree well with the pollen dating ca. 10,200 years B.P. at 675 cm. This implies a high sedimentation rate during Younger Dryas, which seems probable.

Oxygen and carbon isotope data (Erlenkeuser, this volume) seem to be in good accordance with the pollen results as far back as conclusions can be drawn by pollen analysis of the core, viz. back to the beginning of the Preboreal time.

The ^{210}Pb dating (Erlenkeuser, this volume) gives the very important information that the upper part of the core is complete or very nearly so. It indicates a sedimentation rate of about 1 mm/year for the upper 16 cm. The rate derived from the *Picea* curve is not unambiguous, since the beginning of this curve is rather approximately dated. With an age of 1,200 years B.P. at 75 cm, we arrive at a sedimentation rate of 0.6 mm/year, but since this is based on a maximum age, the sedimentation rate may be somewhat higher. On the other hand, a younger *Picea* rise will add to the discrepancy between the magnetometric and the pollen-analytical datings.

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Stable isotopes in benthic foraminifers of Skagerrak core GIK 15530-4: High resolution record of the Younger Dryas and the Holocene

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Erlenkeuser, H.: Stable isotopes in benthic foraminifers of Skagerrak core GIK 15530-4: High resolution record of the Younger Dryas and the Holocene. *Norsk Geologisk Tidsskrift,* Vol. 65, pp. 49–57. Oslo 1985. ISSN 0029-196X.

Stable carbon and oxygen isotope analyses of benthic foraminifers from Skagerrak core GIK 15530-4 (325 m water depth) yield a high resolution record of the Younger Dryas and the Holocene. The Holocene $\delta^{18}O$ variations ($\pm 0.3‰$) nicely reflect the main epochs of the NW European climatic history since late Boreal time. They particularly reveal the warmest phase in the first half of the Atlanticum, a warm late Subboreal, the climatic deterioration about the Subboreal/Subatlanticum boundary, and the coolest phase of the Little Ice Age. The $\delta^{18}O$ data suggest Holocene temperature variations in the Skagerrak deep water of about $\pm 1.3°C$, which may reflect temperature changes of the source water (presumably the deeper North Atlantic Drift Water body in the Norwegian Sea) or a varying degree of more local deep-water formation by winter cooling. A direct response of the benthos $\delta^{18}O$ to the melt water discharge from Scandinavia into the Skagerrak or to the late glacial history of the Baltic Sea is not evident. The $\delta^{13}C$ profile reveals long-term variations which are thought to reflect 'aeration' of the Skagerrak deep water story at the coring location and appear to respond to climatically controlled thermo-haline stratification phenomena in the Skagerrak or the Norwegian Sea.

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The northern Atlantic Ocean plays a key part in the Quaternary history of the earth's climate. The large glacial to interglacial contrasts in the oceanography of the Norwegian/Greenland Sea are evidenced by the faunal and isotopic records of deep-sea cores (Jansen et al. 1983 and references therein). The Norwegian Sea can be expected to sensitively respond to the less dramatic variations of the Holocene climate as well.

Although the position of the present core appears somewhat distant from the occurrences in the North Atlantic ocean, the Skagerrak deep water seems closely coupled to the upper mid-water story of the Norwegian Sea and hence may relate to the North Atlantic Drift Water. This may give the isotope signals recorded by the benthic foraminifers a broader significance. Short-term oscillations of the climate may still be well apparent at the comparatively low water depth of the present core. The fast accumulation of the Skagerrak deposits promises a high resolution of the Holocene environmental variations, which are blurred in slowly growing deep-sea sediments due to bioturbation. The chance to compare high resolution marine sedimentary records from northern latitudes of the Atlantic sector with the comparatively well-known history of the NW European late glacial and Holocene land-climate makes the study of the Skagerrak sediments particularly attractive.

Materials and methods

Stable isotope analyses were performed on the benthic foraminifers *Cassidulina laevigata* and *Elphidium excavatum*. Planktonic species are rare and are mainly confined to the upper 4 m of the core (Thiede, this volume). *C. laevigata* is restricted to the upper 6.75 m, while *E. excavatum* occurs below 4 m (Nagy & Qvale, this volume). In the overlap range, both species were analysed at numerous levels. A total of 207 samples, 122 of *C. laevigata* and 85 of *E. excavatum,* were prepared from 5 cm thick slices taken nearly every 5 cm. According to the average sedimentation rate of about 0.6 mm/y in the Holocene and 4 mm/y in the late glacial section, the samples represent a time interval of ca. 80 y and 13 y, respectively. About 63 % of both sample batches have been analysed up till now.

18 to 40 well preserved specimens were select-

ed from each sample. The samples were ultrasonically cleaned in methanol, dried at 50°C, and reacted under vacuum at 50° with 100% phosphoric acid. The extracted CO_2 gas was analysed for isotopic composition on a VG Micromass 602 D mass spectrometer. The results are given in the usual δ-notation and refer to the PDB scale. For *E. excavatum* the analytical precision ($2\sigma_{10}$-values, calculated as recommended in the spectrometer manual) is generally better than 0.08‰ in $\delta^{18}O$ and 0.06‰ in $\delta^{13}C$. Of the thin-walled low-weight *C. laevigata,* sample gas amounts were rather small in most cases, and the analytical noise becomes more pronounced and variable. It ranges between 0.05 and 0.35‰ (average: 0.12‰ in $\delta^{18}O$ and between 0.02 and 0.20‰ (average : 0.08‰) in $\delta^{13}C$. The $2\sigma_{10}$-'error'-bars are included in the presentation of the results in order to indicate the statistical relevance of the isotope variations found. A strict theoretical treatment shows that the $2\sigma_{10}$ figures is – on the average – somewhat greater than the standard (rms) deviation, the precise relation depending on the number of sample to reference gas cycles run in the particular analysis.

The foraminiferal samples were too small to allow distinct shell size classes to be analysed. Shell size ranged between 200 and 500 μm and for many samples was between 250 and 350 μm. Due to size-dependent 'vital' effects which may particularly appear in the carbon isotope ratio, some of the $\delta^{13}C$ scatter found may result from a varying grain-size composition. However, a systematic variation of the grain-size spectra with sample depth was not observed in the present core.

Results

Oxygen isotopes

The results of the isotope analyses are shown in Fig. 1 and 2. The chronostratigraphy used in Fig. 2 is based on the fixpoints the members of the working group agreed upon (Absolute chronology, summary, this volume), but the uncertainty of the age-scale below 9 m is well realized.

On a first view, the $\delta^{18}O$ profile displays the last 3 of the 4 main steps seen in the benthos $\delta^{18}O$ record of the last glacial and Holocene in well-resolved and ^{14}C-dated deep-sea cores from the North Atlantic (Duplessy et al. 1981, Sarnthein et al. 1982):

1. Termination I A: the first phase of deglaciation and the corresponding decrease of $\delta^{18}O$, occurring between about 16,000 and 13 or 12,000 years B.P. (conventional ^{14}C years).
2. An intervening stage of nearly constant or slightly re-increasing $^{18}O/^{16}O$ ratios (ca. 13 or 12,000 years B.P. to 10,000 years B.P.)
3. Termination I B: the second step of deglaciation, ending at about 8 000 years B.P.
4. The Holocene stage with comparatively light $\delta^{18}O$.

The deepest sample of the Skagerrak core possibly falls in late Term. I A, but more probably marks late Allerød or early Younger Dryas (see below). Younger Dryas sediments range between 10 m (certainly 9 m) and 7 m depth in the core. The beginning decrease of $\delta^{18}O$ at about 7 m coincides with the onset of the Preboreal Chronozone, which appears well established by the bio- and chronostratigraphical framework available for this core. Further upward the isotope profile reveals highly resolved Holocene $\delta^{18}O$ variations, the most prominent features of which nicely coincide with the main phases of the Holocene history of the NW European land-climate (Lamb 1977, chap. 16), if for this first discussion the light $\delta^{18}O$ values are taken to stand for an as yet indistinctive term 'warmth'. In this sense, the $\delta^{18}O$ profile reflects the climatic amelioration beginning with the Preboreal, the main warm period of the Holocene in the first half of the Atlanticum, the climatic deterioration at the end of this period, the restored warmth in Subboreal time, the climatic drop towards the Subatlanticum, and the youngest cold phase of the Little Ice Age.

Three very light $\delta^{18}O$ values closely group at about 8 m depth. The pronounced excursion of $\delta^{18}O$ towards heavier values around 3.8 m within the warmest epoch of the Atlanticum coincides with a phase of possibly rapid sedimentation (as suggested by ^{14}C data, Erlenkeuser, unpubl.) and might reflect sediment redeposition (see also below).

The significance of the smaller $\delta^{18}O$ variations is more difficult to assess in view of the analytical noise, sample spacing, and the comparatively rough and not unambiguous datings. For instance, the ^{14}C ages of the total organic matter (Erlenkeuser, unpubl.), although offset by several thousand years from the magnetic/palynological dating, suggest some fluctuation of the sedimentation rate in the Holocene section of the

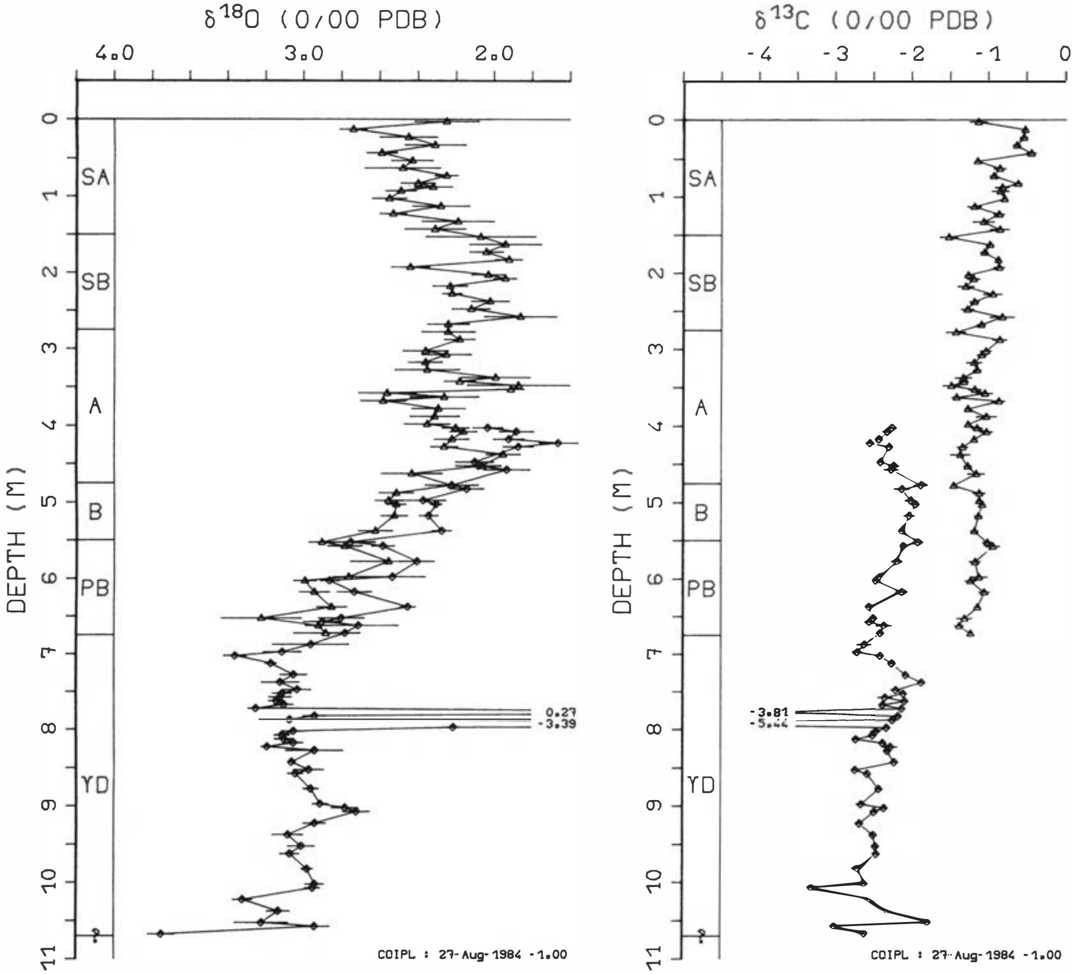

Fig. 1. Stable oxygen and carbon isotope profiles on the benthic foraminifers *Cassidulina laevigata* (△) and *Elphidium excavatum* (◇) in Skagerrak core GIK 15530-4 from 325 m water depth. Error bars indicate about 1 σ noise of the sample gas analysis. Positions of the chronozone boundaries are approximate. SA = Subatlanticum, SB = Subboreal, A = Atlanticum, B = Boreal, PB = Preboreal, YD = Younger Dryas.

core. The details still have to be worked out.

Where the $\delta^{18}O$ profiles of the two analysed foraminiferal species overlap (4.0 to 6.75 m) they appear to run in parallel. However, *E. excavatum* seems lighter by 0.21 ± 0.03 ‰ than *C. laevigata* (pairs from 20 levels, standard (rms) deviation : 0.12 ‰). Isotopic differences between different species from nominally the same habitat are well known to exist (Duplessy et al. 1970, Dunbar & Wefer 1984) and are considered to reflect species-dependent but otherwise constant 'vital' effects. In the present core section, however, with its general upward trend toward lighter oxygen isotope ratios, preferential redeposition of *C. laevigata* from older sediments with 'more glacial', i.e., heavier $^{18}O/^{16}O$ composition, would lead to phase shifts and could have produced the observed systematic isotope difference as well. The low-weight shells of *C. laevigata* might become more easily reworked than is the case for the thick-walled *E. excavatum*. Such redeposition phenomena may depend, for instance, on the stratification of the water column (activity of internal waves) or on transgression phases, and must not appear with continuous intensity throughout the whole core.

With respect to $\delta^{18}O$, *C. laevigata* appears to form its shell closely in 'isotopic' equilibrium

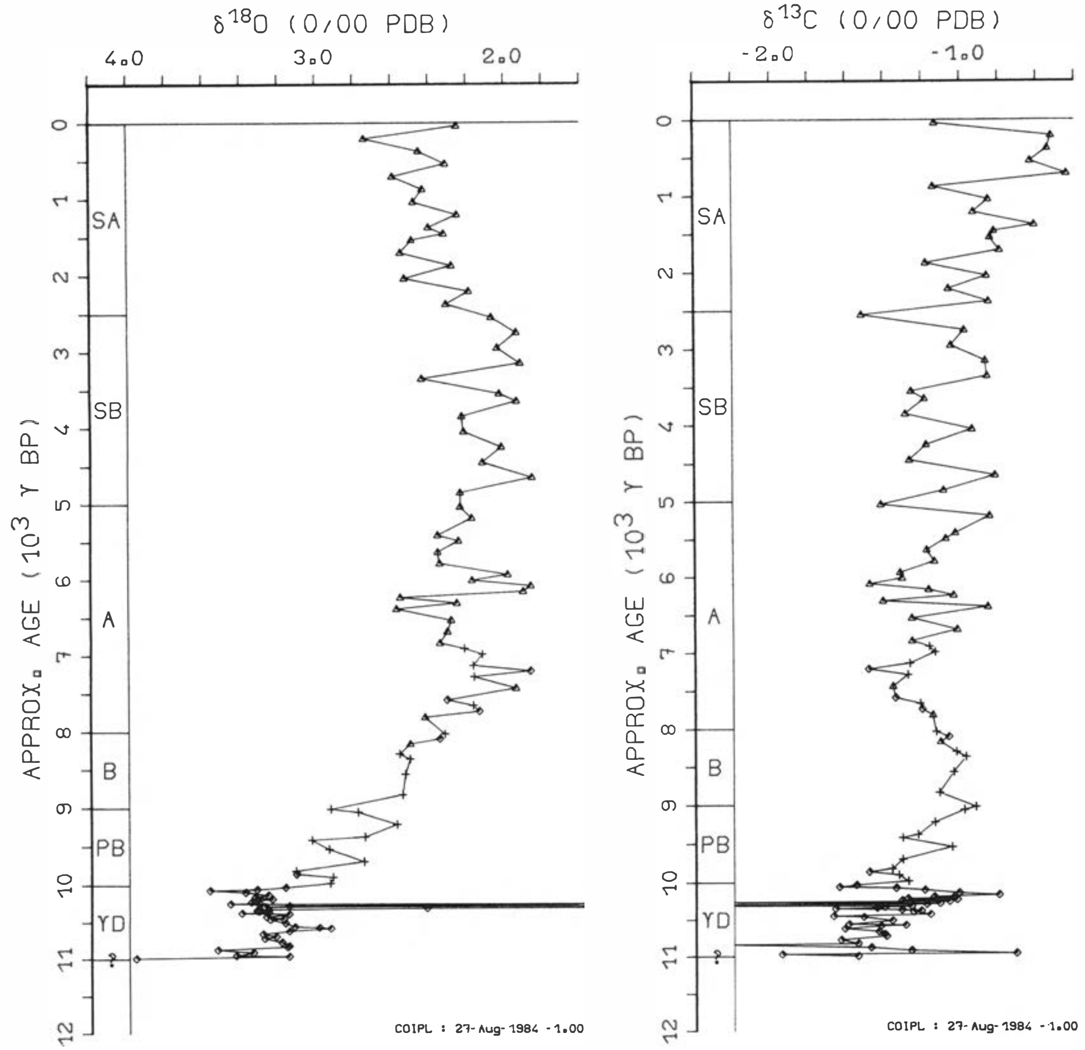

Fig. 2. $\delta^{18}O$ and $\delta^{13}C$ distribution with age in Skagerrak core GIK 15530-4 (325 m). The profiles combine the data of the two foraminifer species analysed: △*Cassidulina laevigata,* ◇*Elphidium excavatum* (normalized data), + (unweighted) average where both species were sampled from the same depth. *E. exc.* data were normalized to *C. laev.* by adding 0.21‰ and 1.06‰ to the measured values of $\delta^{18}O$ and $\delta^{13}C$, respectively. Ages were obtained by linear interpolation between the chronostratigraphical fixpoints adopted (see: Absolute chronology, summary; this volume). The age of the basal sediments is rather uncertain.

(sensu Shackleton 1974) with the ambient water. The uppermost sample (5–10 cm, $\delta^{18}O = \delta_c = 2.25 \pm 0.17$‰) yields an isotope temperature of 7.7 ±0.8°C in close agreement with the actual oceanographic data (Larsson & Rodhe 1979), which show salinities slightly above 35‰ and temperatures of about 7°C in the Skagerrak deep water with little seasonal variance. The isotope temperature was calculated using the $^{18}O/^{16}O$ ratio of Norwegian Sea water ($\delta^{18}O = 0.26$‰ vs. SMOW, i.e., $\delta_w = 0.04$‰; salinity S = 35.2‰, Craig & Gordon 1965) and the paleotemperature equation of Shackleton (1974),

$$t\,(°C) = 16.9 - 4.38\,(\delta_c - \delta_w) + 0.1\,(\delta_c - \delta_w)^2.$$

This equation is physically more relevant and more precise for the low temperature range than the often used (Epstein-relation Epstein et al. 1953). The 10–15 cm sample ($\delta^{18}O = 2.74 \pm 0.07$‰)

from late Little Ice Age sediments yields a temperature of 5.8 ± 0.3°C (assuming salinity $S = 35.2$‰).

For later discussion, it is noted here that also the planktonic foraminifer *Globoquadrina pachyderma* (sin) appears to calcify in isotopic equilibrium with respect to $\delta^{18}O$. This is indicated by the results of Kellogg et al. (1978) and the isotopic agreement with *C. laevigata* in core 25–09 from the S Norwegian Sea (Jansen & Erlenkeuser, in prep.).

Carbon isotopes

In $\delta^{13}C$, *C. laevigata* is heavier than *E excavatum* by 1.06 ± 0.05‰ (20 pairs, rms deviation: 0.22‰). This difference probably reflects vital effects, which in general seem more pronounced in the carbon isotope ratio than in $\delta^{18}O$. Part of the vital effect may depend on the rate of metabolism, and small (young) specimens often reveal lighter ^{13}C than large (adult) forms of the same species (Dunbar & Wefer 1984). The isotopic range may be of the order of 0.5‰. These effects may in part account for the comparatively large scatter of $\delta^{13}C$ in core GIK 15530-4 as the samples were not selected for shell size. The $\delta^{13}C$ profiles nevertheless reveal long-term variations which appear to be related to climatically controlled environmental conditions (see below).

Discussion

Oxygen isotopes

The oxygen isotope composition of foraminiferal carbonate is tagged to the $^{18}O/^{16}O$ ratio of the ambient water and is further controlled by the temperature during calcification. The $\delta^{18}O$ differences in the ocean water mainly result from evaporation and precipitation processes and, locally, from the admixture of (isotopically light) fresh- or meltwater (Craig & Gordon 1965). Seasonal variations of the oceanographic data are averaged in the $\delta^{18}O$ of the total shell according to the growth rhythm of the species considered.

Today the Skagerrak deep water salinity is about the same as in the Norwegian Sea. According to its density, the Skagerrak deep water remains largely unaffected by the seasonal variability of salinity and temperature in the Skagerrak surface waters. This fundamental situation is expected also to hold for the past. The variation through time of the benthos $^{18}O/^{16}O$ ratio in the deep Skagerrak thus appears to reflect oceanographic changes of the Norwegian Sea. This may account for the apparent lack of correlation between the $\delta^{18}O$ record of the present core and the surface hydrography of the Skagerrak, which must have undergone large changes in the course of the Scandinavian deglaciation and in particular with the possibly abrupt drainage of the Baltic Ice Lake by the end of the Younger Dryas (Olausson 1982 b, Fredén 1982).

The mean ocean water oxygen isotope composition which has changed by about -1.6‰ since the glacial, stabilized with the end of deglaciation at about 8,000 y BP. Little further change has occurred in the Holocene since, because the eustatic sea level changes remained small. Accordingly, the residual variations of $\delta^{18}O$ by about ± 0.3‰, as recorded in core 15530-4 since Boreal time around an average of ca. 2.3‰, may be tentatively ascribed to temperature variations (ca. ± 1.3°C) of the S Norwegian Sea upper midwater, which probably feeds the deep Skagerrak. From a less regional point of view, these variations might reflect climatically controlled changes of the oceanographic situation in the northern ocean (temperature, circulation speed and pattern, deep water formation). Salinity changes may also exert rather strong effects in $\delta^{18}O$, as the relation between water- $\delta^{18}O$ and salinity is rather steep in the Norwegian-Greenland Sea ($d\delta_w/dS = 0.61$‰/‰, Craig & Gordon 1965). Finally, deep water production on a more local scale by cooling of comparatively saline water in the shallower parts of the adjacent North Sea might provide another mechanism to account for increased benthos $^{18}O/^{16}O$ ratios for climatic periods when strong winters occurred more frequently.

The $\delta^{18}O$ change between the average post-Boreal $\delta^{18}O$ level (ca. 2.3‰, on *C. laevigata*) and the Younger Dryas period ($\delta^{18}O = 3.0$ to 3.2‰, on *E. excavatum*) amounts to about 0.7 or 0.8‰ (referring to the present). This difference represents the net shift summarizing the global $\delta^{18}O$ decrease of the ocean water through the deglaciation period of Term. I B and the environmental signal arising from a change of temperature and a variation of salinity (as far as the latter change differs from the global effect of deglaciation on salinity).

The ice-cap effect may be evaluated from the benthic isotope records of well-resolved North Atlantic deep-sea cores, which reveal a change by 0.7 or 0.8‰ (downcore) through Term. I B (Duplessy et al. 1981, Sarnthein et al. 1982). However, the isotopic deglaciation signal of Term. I B may be smaller by 0.3‰ because the temperature of the North Atlantic Deep Water (NADW) appears to have increased by (at least) 1.3°C since glacial time (Duplessy et al. 1980). This warming is related to the modern mode of NADW formation, whose onset was probably in Younger Dryas time (Jansen & Erlenkeuser, in prep.) and became fully established thereafter. The increase of NADW temperature thus probably took place during Term. I B.

Accordingly, the environmentally related change, Δ_E, which is left in the Skagerrak benthos isotope record amounts to about 0.2 to 0.5‰, or is 0.4 to 0.7‰ if allowance is made for a possible vital effect between the faunal species referred to. Combining, in a linearized differential form, Shackleton's (1974) paleotemperature equation (which yields a δ_c vs. temperature coefficient of $d\delta_c/dt = -0.267$‰/°C at $t = 4$°C, this latter choice not being very critical) and the slope of δ_w with salinity ($d\delta_w/dS = 0.61$, Craig & Gordon 1965), the allowed variations of temperature and salinity are found interrelated as

$$\Delta_E = 0.61\ \Delta S - 0.267\ \Delta t$$

Here, Δt (°C) denotes the difference of the former temperature from the present and ΔS is the departure of the former salt content from the recent one beyond the global effect of glaciation on salinity.

In the extreme case of $\Delta t = -8$°C, ΔS is between –3.2 and –2.4‰. If, on the other hand, ΔS is limited to a probably more realistic range of $\Delta S = 0$ to -1‰, the temperature change falls between –1 and –5°C. This leaves the possibility that the Skagerrak deep water during the Younger Dryas was nearly as warm as it is today.

However, these results do not consider the possibility of bottom water production by sea ice formation in the Skagerrak of the Younger Dryas (see below). Deep water supplied by this process would have the lighter isotopic composition of the melt-water affected surface layer. Temperatures estimated as above might come out too high by 2.7 to 3.3°C if this effect is not taken into account. Accordingly the bottom water temperature at the coring site might have been close to 0°C in the Younger Dryas. Adopting this value in turn suggests salinities –1 to –2‰ below 'normal' (in the sense of the definition of ΔS).

Establishing a time control for the pre-Preboreal section of the core on the basis of the oxygen isotope record appears difficult and rather speculative at present. During deglaciation, the Norwegian Sea has undergone thorough changes of circulation, temperature, and salinity probably at all water depths (Jansen et al. 1983, Sejrup et al. 1984, Jansen & Erlenkeuser in prep.). Lateral gradients in temperature or salinity were recorded for the S Norwegian Sea of that time (Sejrup et al. 1984), so that the source and hence the isotopic composition of the Skagerrak deep water cannot be assessed with confidence. Interestingly, the $\delta^{18}O$ level in the Younger Dryas sediments of the Skagerrak core is closely the same as in core 31–33 from the S Norwegian Sea (Sejrup et al. 1984).

According to paleoclimatic studies in S Norway and paleoceanographic reconstructions from faunal and isotopic evidence in sediment cores from the Norwegian Sea (for references see Jansen et al. 1983), North Atlantic water entered the Norwegian Sea as early as 13,000 years B.P., forming at least a subsurface or mid-water body which also persisted through the climatic deterioration of the Younger Dryas. The comparatively heavy $\delta^{18}O$ value at the base of our core might thus be taken to trace a pre-Bølling deposit. An age of 13,000 years B.P. could then be tentatively assigned at 10.6 m depth, and the onset of the Allerød (ca. 11,800 years B.P.) perhaps at 10.1 m. It should be noted, however, that there is no direct time control pinpointing the end of Term. I A as defined by the local isotope record in the cores from the S Norwegian Sea (Sejrup et al. 1984).

On the other hand, cores 31–33, 31–36 (Sejrup et al. 1984) and K 11 (Duplessy et al. 1975) consistently reveal – although by one sample in each core only – an isotopically light $\delta^{18}O$ peak at the end of the (locally defined) Term. I A. As the planktonic foraminifer *Globoquadrina pachyderma* (sin.), which was analysed in these cores, is a deep-dwelling species (Kellogg et al. 1978), this isotopic result might have some relevance as regards the isotopic composition of the source water feeding the Norwegian Trench and the deep Skagerrak. A light $\delta^{18}O$ signal of about 0.5‰ in magnitude which then should appear below the Younger Dryas zone is not evident in the Skagerrak core. Therefore it seems possible that the Allerød stage has not been cored. The faint trace of a boreal fauna appearing at the base of the core and tailing off upward (Nagy & Qvale, this volume) as well as the decline of some thermophile plant species according to the pollen profiles (Henningsmoen & Høeg, this volume) may further indicate that the Allerød sediments have not been retrieved. So I suggest placing the age of 11,000 years B.P. at 10.7 m depth. I realize that none of these arguments is convincing because for these deep core sections the apparent effects of reworking can hardly be asessed.

For further insight into the nature of the late glacial isotope record of the Skagerrak core, regard is given to the work of Olausson (1982 b), who reports late glacial $\delta^{18}O$ profiles (on *Elphidium excavatum*) for several cores from basins of the Swedish Skagerrak coast. The early pre-Holocene sediments in these cores were deposited below 30 m water depth (Solberga core: below 50 m) and therefore should be sheltered to some degree from variable salinity of the surface water. The $\delta^{18}O$ values in the deeper sections of the cores are about 3 to 3.5‰. They are lighter than the glacial oxygen isotope composition (4.4 to 5.0‰) of the deep-dwelling *G. pachyderma* (sin) in the Norwegian Sea, but are in the same range as the deep samples of the Skagerrak core (3.0 to 3.7‰, below 10.1 m) and also compare with the isotope ratio of *G. pachyderma* (sin.) from the intermediate section between Term. I A and I B in cores 31–36 ($\delta^{18}O$ range: 3.5 to 4‰), 31–33 (2.7 to 3.3‰), 25–29 (3.5 to 4‰), and K 11 (3.2 to 3.7‰) from the Norwegian Sea (Kellogg et al. 1978, Sejrup et al. 1984). Such a situation of a comparatively uniform $\delta^{18}O$ distribution in the late glacial sub-surface waters of the Norwegian Sea and the Skagerrak might have prevailed in times when local melt water supply was not excessively high, i.e., before the Baltic Ice Lake drained into the Skagerrak.

Further up the Swedish cores, a pronounced decrease in $\delta^{18}O$ appears to mark the increasing influence of melt water now beginning to be discharged into the Skagerrak. Olausson (1982 b) dates this isotopic change at about the Older Dryas/Allerød boundary. However, this date seems to be in conflict with the much younger ^{14}C dates of shell fragment samples from these cores (Moltemyr and Solberga cores, Olsson 1982). This discrepancy is even harder to understand because such radiocarbon dates tend to be too old rather than too young on account of unrecognized redeposition and reservoir age effects. Based on the radiocarbon analyses, in particular on those of the Solberga samples, the said $\delta^{18}O$ decrease should occur at about the Allerød/Younger Dryas boundary, and thus appears to coincide with the suspected age of the $\delta^{18}O$ decrease at the base of the Skagerrak core. (Interestingly there are further details of isotopic correspondence between the Skagerrak sediments and the Swedish cores, particularly the deeper lying Solberga core.)

Accordingly, the onset of the Younger Dryas is suspected to be accompanied by a *decrease* of the benthos oxygen isotope composition in the Skagerrak. Recalling the cold waters prevailing below the surface of the Skagerrak of Younger Dryas time according to the faunal evidence, the decrease in $\delta^{18}O$ might be accounted for by surface freezing. This process increases the density of the water left, allowing it to sink, whereas the isotopic composition is preserved or is shifted, if at all, toward lighter values (Craig & Gordon 1965). Sea ice formation hence provides a mechanism to bring isotopically light water from a meltwater affected surface layer down to greater water depth. Disregarding possible effects of temperature changes on the benthos isotopic composition, a decrease in $\delta^{18}O$ by 0.7‰ corresponds to the salinity of the surface water being 1.1‰ lower than in the deep water displaced. This estimate is based on the modern δ_w vs. salinity relation (Craig & Gordon 1965), but possibly should be lowered by 10 or 20 % for the late glacial time when melt waters probably had lighter $^{18}O/^{16}O$ ratios than today and hence would have exerted a more pronounced influence on the isotopic composition of the sea surface waters. Thus some margin is left to allow for temperature changes in connection with the bottom water replacement.

Still, deep water formation by surface freezing can hardly account for all of the $\delta^{18}O$ change recorded in Olausson's cores below the Younger Dryas/Preboreal boundary. Either there existed a marked melt water run-off to the Skagerrak already in Younger Dryas time, or increased reworking started with the climatic deterioration at the end of the Allerød, bringing isotopically light foraminifer shells from littoral and sublittoral low salinity environments down to the deeper parts of the coastal basins.

Of the 5 samples taken between 7.75 m and 8.00 m, three exhibit very light $^{18}O/^{16}O$ ratios partly accompanied by very light carbon isotopes too. If these samples are autochthonous, the oxygen isotope composition suggests salinities of ca. 23 to 27‰, which must have occurred down to about 280 m at least, i.e., the former water depth of the core (Stabell & Thiede, this volume). The low salinity events should have lasted a few decades at most, according to the average sedimentation rate of this core section.

The light isotope samples appear well below the Preboreal chronozone still within the Younger Dryas. Although they are younger than the 8.5 m level which is dated by the ash layer and the radiocarbon sample below at about 10,500 years B.P. (Stabell, this volume), they still seem too old to be associated with the rapid and probably dramatic break-through of the Baltic Ice Lake water to the Skagerrak close to 10,200 years B.P. (Olausson 1982 a). Nevertheless, this point deserves closer inspection, as short term variations of sedimentation rate may not have been recognized.

Other hypotheses have to consider the build-up of a giant ice-barrier across the Norwegian Trench by icebergs or rapidly advancing glaciers. Interestingly, the age of the 'low salinity' events closely coinicides with the maximum westward extension of the glaciers in the Hordaland area in late Younger Dryas (Mangerud 1977). It is possible that the low $\delta^{18}O$ samples represent allochthonous assemblages of specimens redeposited (ice rafted?) from low salinity shallow coastal areas, a well-known habitat of *E. excavatum* (van Weering & Qvale 1983). Isotopically light *E. excavatum* were reported from late glacial sediments of the Swedish west coast (Olausson 1982 b).

Carbon isotopes

In spite of the comparatively large scatter between adjacent samples, long-term variations are evident in the $\delta^{13}C$ profiles. The $^{13}C/^{12}C$ ratio of the water is non-conservative. In the deeper water column, the release of CO_2 by decomposition and degradation of (isotopically light) organic matter decreases the $^{13}C/^{12}C$ ratio of the dissolved inorganic carbon species and hence of the foraminifers calcifying therefrom. The carbon isotope ratio hence depends on the history of the water body considered, in particular on the productivity at the ocean surface under which the water has proceeded since the time it was cut off from gas exchange with the atmosphere. The accumulation of isotopically light carbon dioxide in the water is inversely related to the availability of dissolved oxygen (Williams et al. 1977, Kroopnick 1980).

In this sense, the variation of the $^{13}C/^{12}C$ ratio of benthic foraminifers may be taken to reflect a change of the extent to which the deep water site is 'aerated'. Clearly, the nature of the $\delta^{13}C$ variations is complex and may have other causes besides 'aeration' effects.

In general, phases of lighter $^{13}C/^{12}C$ ratios in the benthic record of the Skagerrak core appear to coincide with 'warm' climatic conditions. Comparatively low $\delta^{13}C$ values are found between ca. 10 and 9 or 8.5 m and may result from an extended sea ice cover along with a more pronounced haline stratification of the water column (but not necessarily at the position of the core) which may have existed during and for some time after the Allerød warm stage, when large amounts of melt water were discharged into the Norwegian Sea. Comparatively light $\delta^{13}C$ values were also observed on *G. pachyderma* (sin.) from Younger Dryas sediments in cores 31–33 and 25-09 from the S Norwegian Sea (Sejrup et al. 1984). The aeration of the Skagerrak deep water improved in the cooler phase of the late Younger Dryas (about 7.5 ±0.3 m) and deteriorated again in the next period of strong melt water run-off, i.e., the Preboreal. With the disintegration of the ice caps coming to an end, aeration improved in late Preboreal and Boreal time. Further on in the Atlanticum, the ^{13}C variations could reflect a more or less developed thermocline in the Norwegian Sea. In the core sections following, where sample density is smaller than below, the $\delta^{13}C$ record is rather noisy, but a general trend of increasing $\delta^{13}C$ is evident. Heavier $^{13}C/^{12}C$ ratios seem to prevail in the late Subboreal and Subatlantic zone and, more pronounced, coincide with the most recent cold epoch culminating in the Little Ice Age.

In summary, the correspondence between the ^{13}C profile and the NW European climatic history seems consistent, but the basic processes bringing about this response have still to be worked out in detail.

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Lithostratigraphic and biostratigraphic studies

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Lithostratigraphic and biostratigraphic studies: Summary core GIK 15530-4

A wide spectrum of micropaleontological and sedimentological studies of sample material from core GIK 15530-4 was carried out. We hoped that these studies would enable us 1) to establish a detailed litho- and biostratigraphic zonation, and 2) to reconstruct the depositional environment of the Upper Quaternary Skagerrak basin.

Main results

From the bio- and lithostratigraphical studies the core can be subdivided into two main units (Fig. 1): a lower unit from the base to 700 cm, and an upper unit from 650 cm to the top of the core. The lower unit is characterized by a cold water (Arctic) flora and fauna of low diversity, and the sediment contains abundant coarse terrigenous clastic particles. The upper unit has a highly diverse microfossil assemblage which indicates temperate water conditions. The sediment coarse fraction is largely dominated by biogenic particles.

The lower unit of the core with Arctic type microfossils can be defined as a *Cassidulina reniforme*/*Amphimelissa setosa* assemblage zone with planktonic diatoms. Planktonic foraminifers are rare throughout this unit, and only reworked coccoliths have been found. Trace fossils are tube burrows. The lower unit has not been subdivided, but it should be noted that 'brown crusts' of iron oxides are restricted to the lowermost 2 m

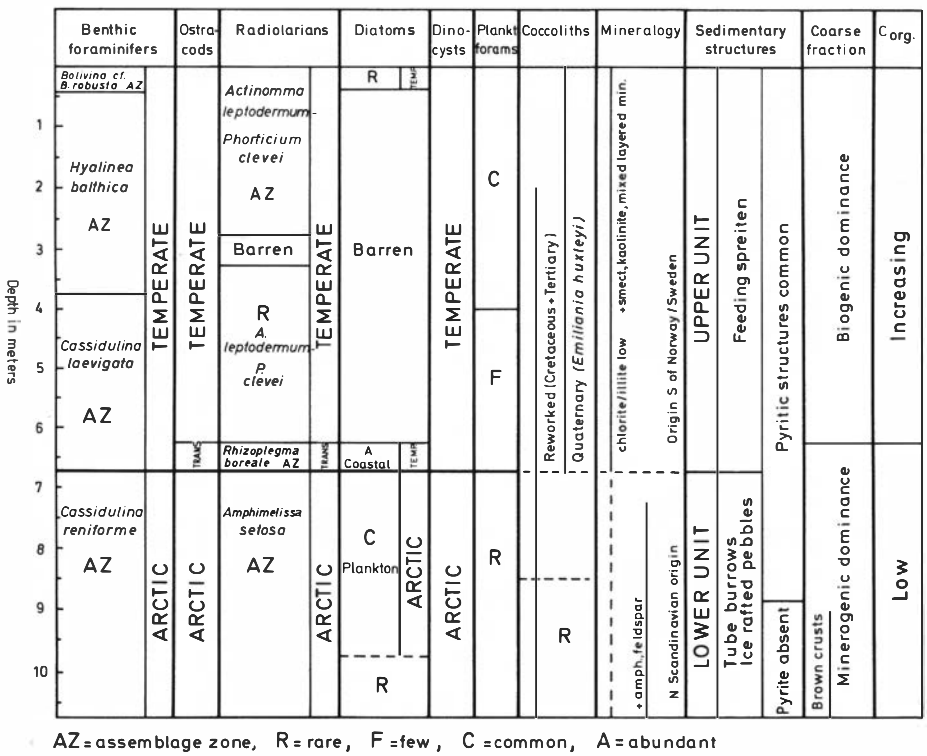

Fig. 1. Diagram showing the main biostratigraphic and lithostratigraphic results.

of the core, which corresponds to the section where pyritic structures are absent.

The upper unit, which is dominated by microfossils that indicate temperate conditions, starts with a transitional zone. This zone is defined by a *Paralia sulcata – Rhizoplegma boreale – Sarsicytheridea bradii* assemblage. The lower boundary of this zone corresponds to the immigration of temperate type microfossils, the upper boundary to the shift from a minerogenic to a biogenic dominance in the coarse fraction. The transitional zone still contains elements of the Arctic assemblage of the lower unit, but indicates a major change in the circulation pattern with influx of warm Atlantic water.

The upper unit is characterized by the occurrence of planktonic foraminifers and Quaternary coccoliths (*Emiliania huxleyi*). The preservation of siliceous material is poor; diatoms are absent in most of this unit. It can be subdivided into three different zones using benthic foraminiferal assemblages. The lower part (up to 400 cm) is characterized by an assemblage dominated by *Cassidulina laevigata*. The interval 350–50 cm is defined as a *Hyalinea balthica* assemblage zone, which also corresponds to an increase in the frequency of planktonic foraminifers and radiolarians (*Actinomma leptodermum – Phorticium clevei*). The upper 50 cm is characterized as a *Bolivina* cf. *B. robusta* assemblage zone, and indicates present conditions. The trace fossils are mainly of feeding spreiten type throughout the upper unit. The amount of organic carbon is rather low (<0.75%) in the lower part of the core, but increases from 6 m and upwards.

The mineralogical analyses have shown that the lower unit (below 700 cm) contains material which is interpreted to be mainly of northern (Scandinavian) origin and thus indicates a transport with meltwater from the north. The upper unit is characterized by clay minerals that point to an origin south of Norway and Sweden, and thus a different transport direction. The biogenic material found in the lower unit has a conspicuous amount of reworked material, especially coccoliths and pollen.

On the relationship between shear strength and effective overburden pressure in Upper Quaternary marine Skagerrak deposits

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Rosenqvist, I. Th. & Pederstad, K.: On the relationship between shear strength and effective overburden pressure in Upper Quaternary marine Skagerrak deposits. *Norsk Geologisk Tidsskrift,* Vol. 65, pp. 63–64. Oslo 1985. ISSN 0029-196X.

Measurements of physical properties of a core through Upper Quaternary marine sediments of the Skagerrak reveal major changes with age and depth. Water contents decrease from approximately 150% (in per cent of dry sediment weight) close to the surface (in Subatlantic sediments) to 60–70% below the 4 m-interval; in the lowermost 6 meters of the core (deposited during the time span from Younger Dryas to Atlantic) water contents fluctuate between 60 and 80%. Shear strengths increase from < 50 KN/m^2 close to the surface to > 200 KN/m^2 around the 5 m interval, but decrease slightly in the section below this interval.

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Upper Quaternary sediments originating from marine depositional environments cover wide regions of the Skagerrak as well as of the remainder of the Norwegian continental margin. A detailed knowledge of their physical properties becomes increasingly more important because of the construction activities of the petroleum industry. Such sediments are also found in many parts of Norway in areas which once were covered by Late Weichselian and Holocene seas, but which since then have fallen dry due to the isostatic uplift after the deglaciation; studies of physical properties of such sediments thus have long traditions in Norway.

In this paper we present data of physical properties from core GIK 15530-4 which have been collected in the outer Skagerrak (Stabell et al. this volume). The mineralogy (Rosenqvist, this volume) and grain size (Stabell et al. this volume) of the core are relatively similar throughout the core. The bulk mineralogy, within experimental error, is qualitatively similar throughout the core. The mineral composition is semiquantitatively determined in some samples; thus the exact relation between inactive minerals (Skempton 1953) like quartz-feldspar etc. and the clay minerals illite, smectite and kaolinite is not determined accurately. Grain-size distribution was not studied in detail by us.

Physical properties

This core has been investigated with respect to shear strength and water content (Fig. 1). These data show two different shear strength regression lines respectively with core depth and, more clearly, with the effective overburden stress (cf. Richards 1976).

There is a marked change in shear strength and water content between 500 and 700 cm. In the upper part the water content (% dry weight) decreases from 150% to 70% at 500 cm depth. From 700 to 1050 cm the water content is relatively constant between 60 and 70%.

To calculate the effective overburden stress, the density of the clay minerals is assumed to be 2.7 g/cm^3. Wet sediment densities are as follows: 150% water content = 1.34 g/cm^3, 100% water content = 1.45 g/cm^3, 80% water content = 1.54 g/cm^3 and 70% water content = 1.58 g/cm^3. At 100 cm depth the effective consolidation stress is calculated to be 4.2 KN/m^2 and from 100 to 200, 200 to 300 and 300 to 500 cm respectively the data show an increase of 5 KN/m^2, 5 KN/m^2 and 11.6 KN/m^2. Thus the consolidation pressure at 500 cm depth has been 25.6 KN/m^2, but the shear strength regression line is about 16.5 KN/m^2 at this level and 3 KN/m^2 at the surface.

At 1050 cm depth the calculated consolidation pressure is 59 KN/m^2. The shear strength at 1050

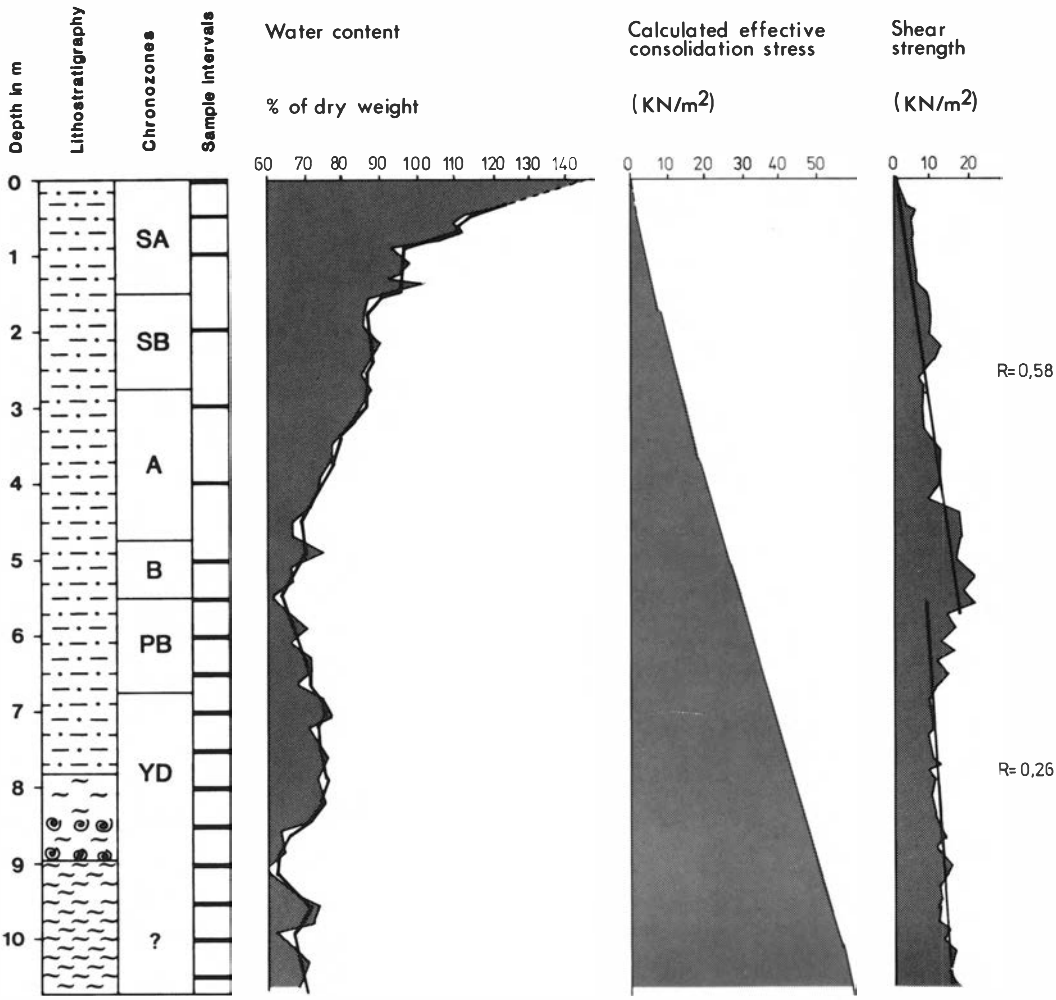

Fig. 1. Distribution of water content (in per cent of dry weight of sediment, also showing three point running average), calculated effective consolidation stress and shear strength in core GIK 15530-4. R = ratio shear strength/effective consolidation stress.

cm depth is 16 KN/m^2 and at the surface 3 KN/m^2.

The relationship between the increase in shear strength and effective overburden stress is thus at the upper 500 cm about 0.58, and at the lower part 0.26.

Discussion

The Norwegian Geotechnical Institute and several institutions in foreign countries have reported that the increase in shear strength with respect to effective consolidation stress depends on ionic state, clay minerals and grain size. These data from the deeper part of the core are in accordance with similar data from Scandinavian clays, but the upper 500 cm can be related to more active clays (Jörgensen et al. 1981).

Smectite has been identified throughout the entire core and the observations may indicate that the deeper part has more inactive clay minerals due to coarser particles or greater relationship between inactive and active minerals.

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Sedimentary structures and the record of trace fossils in Upper Quaternary marine Skagerrak deposits

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Analysis of sedimentary structures as displayed in radiographs leads to a lithostratigraphic subdivision of a 10 m sediment core from the Skagerrak mainly in terms of ichnofacies. The first order boundary between the Upper and Lower Units separates sections with feeding-spreiten and tube burrows (*Lophoctenium, Chondrites*) and with biodeformational structures in the Upper Unit from sections with characteristic types of small hollow tubes and partly extensive pyrite structures accompanying ice-rafted material in the Lower Unit. The coincidence of this subdivision with other stratigraphic events indicates that the ichnofacies represents a significant record of conditions of the paleoenvironment.

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The investigation of sedimentary structures of core 15 530-4 from the Upper Quaternary sediments of the Skagerrak may contribute in

- defining litho-stratigraphic boundaries;
- giving information on some processes influencing the distribution patterns of sediment components;
- reconstructing the depositional paleoenvironment;
- estimating degrees of vertical mixing of sediments and thereby indicating the resolution of stratigraphic signals.

Similar to cores from the southern Skagerrak slope studied earlier (Jørgensen et al. 1981), the sedimentary structures of this core are dominated by bioturbation. Therefore, the present study mainly deals with the analysis of biogenic structures. For this, palichnological criteria are used. Some palichnological analyses of biogenic structures in young oceanic sediments have been carried out, including papers by Berger et al. (1979), Wetzel (1981), and Werner & Wetzel (1982). In the Skagerrak the sedimentary and biogenic structures in two cores from the Danish slope have been studied by Jørgensen et al. (1981). Progress in ecological interpretation of trace fossils is expected from regional comparison of biogenic structures analysed on this basis (Werner & Wetzel 1982). The present paper is understood as a contribution in this respect.

Methods

X-ray radiographs were made from 8 mm thick sediment slabs throughout the core. Although this thickness is an acceptable compromise between satisfying resolution of details and a desirable volumetric integration of fabric elements, it should be kept in mind that not all of the observed structures are statistically representative, and that some of the fabric elements, even when considered as being typical features, are just above the limit of perceptibility. For exposure, a voltage of 30 kV has been applied.

Results

On the whole, the structures document the undisturbed recovery of the piston core. The top layer shows some characteristic features which plead for the preservation of the real sediment surface (see also Erlenkeuser, this volume). A crashing of the core liner during the coring process caused only minor disturbances in the upper part of the core.

The log record of the observed structures led to the subdivision of the core profile (Fig. 1) into several units. At 6.8 m below sediment surface, a first order boundary may be defined, while two other boundaries within both the upper and the lower sections respectively should be considered as of second order.

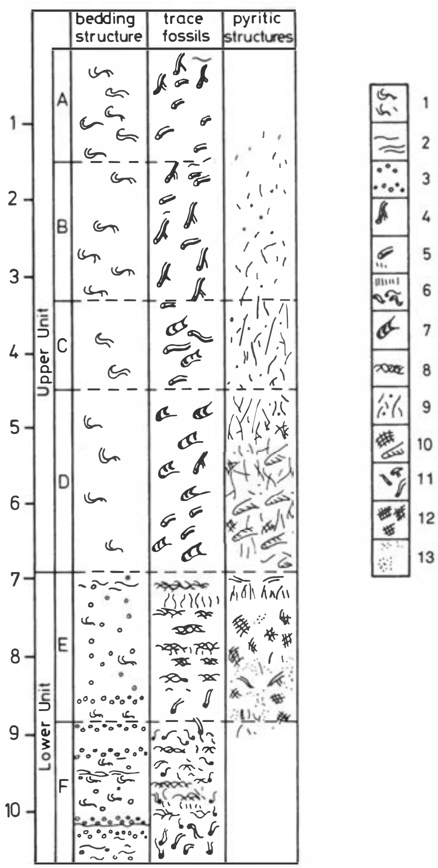

Fig. 1. Sequence of sedimentary structures in core GIK 15530-4. 1 = biodeformational structure, more horizontally or more vertically oriented. 2 = bedding structure. 3 = ice-rafted coarse grains, irregularly scattered or in layers. 4 = branched tubes of *Chondrites* type. 5 = Tube burrows, sediment-filled, *Planolites* type and others. 6 = Open tubes in lower core sections of different sizes and types. 7 = Feeding burrows (cf. *Lophoctenium*). 8 = Tube clusters of *Helminthopsis*. 9 = Mostly unbranched pyrite 'threads', partly *Trichichnus*. Dots: Aggregates of pyrite framboids. 10 = Clusters of small pyrite threads partly arranged within feeding burrows. 11 = bended type of hollow tubes. 12 = Dense clusters of myceloid pyrite structures. 13 = Disseminated pyrite framboids. Columns on left hand: lithostratigraphic subdivision as defined in the text.

1) First order boundary

The two sections separated by the first order boundary are characterized by the data compiled in Table 1. According to the stratigraphic results of other papers of this volume (see synthesis of Bjørklund et al.), this boundary coincides remarkably well with the Pleistocene-Holocene boundary. The boundary is reflected both by primary and by biogenic structures.

2) Upper Unit (Holocene)

The structural inventory, summarized in Fig. 1, allows a subdivision into several sections which are characterized in the following:

Section A (0–1.5 m)

- Biodeformational structures of a mainly wavy-flaser character are dominant. The contrasts in the radiographs are low (Fig. 2b).
- Tubes probably made by polychaetes are rare and poorly defined and more frequently found in the uppermost 30 cm, particularly in the top layer 0–5 cm (Fig. 2a). They are filled with sediment of higher water content than the host sediment; no wall lining can be observed. They are mainly vertically or sub-vertically oriented.
- Occasionally, vertical spreiten burrows of the *Teichichnus* type occur (Fig. 2c); the contrasts in the spreiten lamellae are low.
- Pyritic structures begin to occur sparsely about 35 cm from top, remaining rare and of small individual size throughout the section. They are mainly present as small, thin (ca. 0.1 mm or less) and irregularly formed threads which hardly can be attributed to trace fossil types, their pre-existence as tubes being doubtful.

Section B (1.5–3.2 m)

- Tube burrows are of two types. The first type forms a dense network of short, partly branched, partly defined tubes of ca. 2 mm diameter without wall linings, often concentrated in layers of a few centimetres thickness. The filling is less dense than the host sediment, contrasting very little with the latter. They are defined as *Chondrites*. When compared with the different *Chondrites* types described by Wetzel (1981), there exists much similarity with his type B. The second type refers to tubes of somewhat larger diameters (ca. 3 mm) and longer, mostly straight-lined extensions, horizontally or at low angles obliquely oriented. The tubes occur isolated. They may be put in the *Planolites* group (type D of Wetzel, 1981).
- Biodeformational structures are definitely less frequent than in the previous section, the wavy-flaser character having nearly disappeared.
- Occasionally, large biogenic structures occur which are related to the *feeding-spreiten* burrows as described in the following section.

Table 1. Characterization of first order boundary based on radiographs.

Descriptive category	Upper unit (Holocene)	Lower unit (Pleistocene)
Lithology	Homogenous clayey silt	Clay alternating with coarse-grained material (layers or interspersed)
Stratification	Missing	Layers of coarse material (coarse sand or gravel size)
Biodeformational structures	Partially abundant	Little or missing
Definite biogenic structures	Solitary and dense nets of tube burrows (Chondrites) and dominant feeding spreiten (Lophoctenium)	Only small hollow tubes interspersing the otherwise intact sediment
Pyritic structures	Abundant; mostly long tread-like structures dominant	In parts totally absent; if present, different from upper section, mostly "myceloid" structures

- *Pyritic structures* are more common than in the previous section, although the 'threads' are still very thin and only occasionally trace pre-existent tubes (the *Chondrites* tubes are not pyritized). The threads are either straight, several centimetres long and oriented vertically or sub-vertically and can then be attributed to *Trichichnus* (Häntzschel 1975) or, short and bended, then comparable to structures described as 'Mycellia' by Blanpied & Bellaiche (1981). Aggregates of pyrite framboids are partially abundant, forming light dots in the radiographs (Fig. 2d, e).

Section C (3.2–4.4 m)

This part of the core represents a zone of transition between the previous and the following section, joining elements of both of them. Pyrite structures increase both in frequency and in length of single threads which occasionally attain lengths of up to 10 cm (Fig. 2e).

Section D (4.4–6.8 m)

The dominating biogenic structures in this section are burrows which cross the core profile mostly at a low angle to the horizontal (Fig. 2f, 3a), giving the appearance of bands. The burrows are filled with sediment of slightly less density than the host sediment, which forms a halo of condensed material. The filling appears mostly as homogenous in the radiographs, but occasionally a faint crescentic backfill structure can be observed (Fig. 2b). This feature and the complete absence of corresponding circular or elliptic cross sections supports the interpretation as a sheet-like feeding-spreiten burrow. Following Jørgensen et al. (1981), who also describe such structures from the eastern Skagerrak, they may be interpreted as ichnogenus *Lophoctenium.* Off West Africa, Wetzel (1981) described burrows of a similar structure from the continental slope as *Lophoctenium.* Very similar burrows are found also in muddy sediments of the deeper continen-

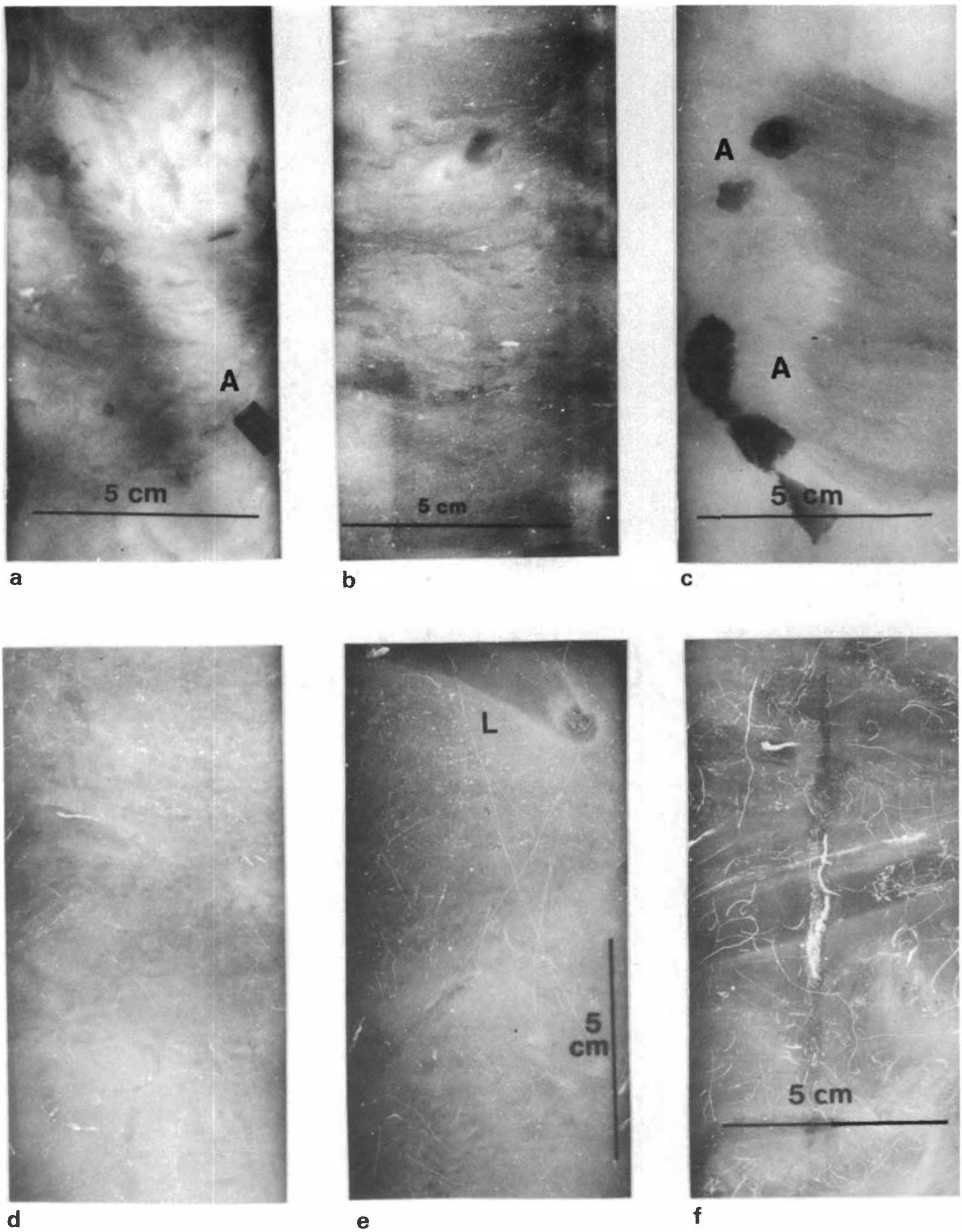

Fig. 2. Prints of radiograph negatives. a) 1–12 cm sediment depth. Branched tubes filled with soft sediment and biodeformational structure with predominantly horizontal orientation. b) 74–86 cm wavy-flaser biodeformational structure, tube burrows. c) 131–143 cm. Vertical spreiten burrows of *Teichichnus* type. A = artifacts d) 250.5–262.5 cm. Small branched tubes filled with very soft sediment, probably mostly *Chondrites* burrows; small pyrite structures. e) 375–389 cm. Feeding-spreiten burrows (L) and *Chondrites* (CH). Pyrite 'threads' (*Trichichnus*) with thigmotactic relation to feeding burrows. Light dots are aggregates of pyrite framboids. f) 448–460 cm. Section dominated by (cf.) *Lophoctenium* feeding-spreiten and intense pyritization. Within feeding burrows and vertical tube myceloid pyrite clusters (m). All prints are in the same scale.

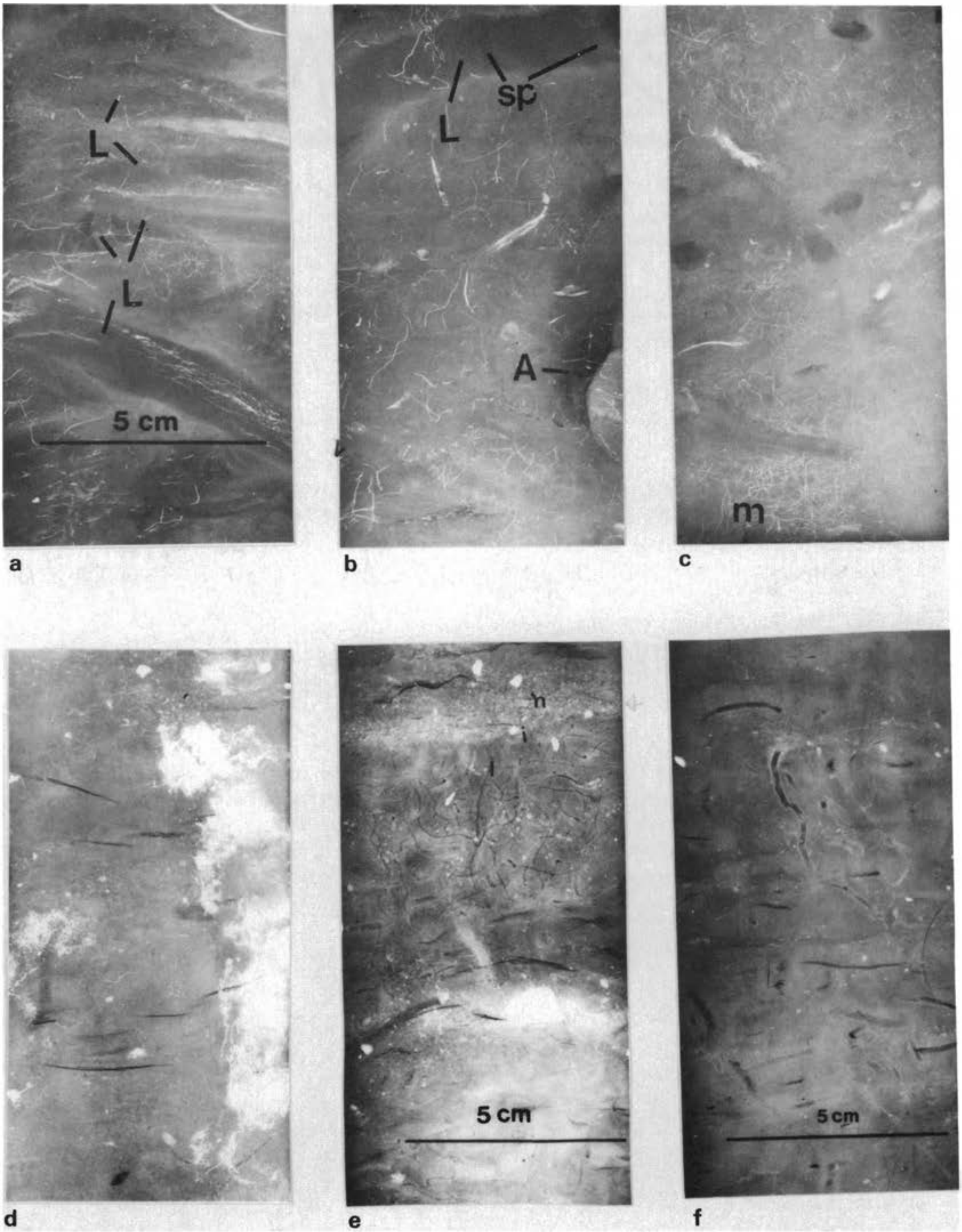

Fig. 3. Prints of radiograph negatives (cont.). a) 564.5–558.5 cm. Feeding-spreiten burrows (*Lophoctenium*), partly crossing each other. Concentration of pyrite clusters in burrows. Spreiten structure (sp.). b) 602–614 cm. Feeding-spreiten burrow (L) with spreiten structure (sp.), indistinct tube burrows (*Chondrites?*). A = artifact. c) 617–629 cm. Indistinct small tubes of (?) *Chondrites* type and *Planolites* tubes with wall thickening (halo). Dense, myceloid clusters of pyrite threads (m). d) 746–758 cm. Extensive pyrite clusters in nearly structureless sediment with dropstones. e) 1004–1016 cm. Two layers with coarse ice-rafted material, the upper one with erosion surface. Small hollow tubes with wall lining, partly branched; dropstones. f) 1052–1064 cm. Hollow tubes of different types and size, partly with wall lining. All prints are on the same scale.

tal shelf off northern Norway (E. Thomsen, Tromsø, pers. comm.).

Pyritic structures are abundant throughout the section, in character partly similar to those of section C, although the pyritization is generally intensified. Frequently pyritic 'threads' forming clusters of a 'mycelic' character are observed (Fig. 3b, c). Such structures are often also tracing the *Lophoctenium* burrows (Fig. 3a). Partially, the pyritic thread-like structures are branching in the same pattern as *Chondrites* (Fig. 2f) although no pre-existing tubes of this type are observed. Very sparsely, other trace fossils occur: *Chondrites, Teichichnus,* and little characteristic tubes.

Based on the interpretation of the coarse material in the lower part of the core as ice rafted, the first order boundary may be considered as separating Postglacial from Glacial deposits (Holtedahl & Bjerkli 1982, Vorren et al. 1984). This general division in a 'homogeneous' upper and a layered lower section is in harmony with the acoustostratigraphic character of the profile in the sediment echogram (Stabell et al., this volume, Fig. 3). The boundary between a homogeneous upper part and the layered lower section there corresponds with the sediment depth of the first order boundary of the core profile.

Lower unit (Pleistocene)

The core profile below the first order boundary can also be subdivided into several sections according to the structure inventory. The most conspicuous feature is the presence of more or less densely scattered pebbles which must be interpreted as ice-rafted material ('dropstones', Holtedahl & Bjerkli 1982, Vorren et al. 1984, Fig. 3d-f, this paper).

Section E (6.8–8.85 m)

- Ice-rafted pebbles are occurring mainly as scattered grains instead of layers.
- None of the types of trace fossils from the upper unit is found here, except for some pyritic structures.
- At least three other types of tube burrows can be distinguished. The first type forms hollow, mainly vertical-orientated tubes of about 0.2 mm in diameter, cf. ichnogenus *Trichichnus* (Frey 1970). In shallow water, similar tubes have been attributed to polychaete genera like *Magelona* and *Scolecolepides* (Frey 1978). This reference should be understood, however, only as one case example of many possibilities. The second type of hollow tubes is larger in diameter (ca. 0.7 mm) and bended. The halo (wall thickening) of denser material, large in relation to the tube diameter, is a characteristic feature (Fig. 3e). The third type is represented by small, lined tubes filled with low-density sediment, forming dense networks arranged in layers, and can be related to *Helminthopsis* (Wetzel 1981). Mostly these layers are 1–3 cm thick and consist exclusively of the tubes. The maximum thickness observed was 6 cm.
- *Pyrite structures* are abundant throughout the section. Three main types can be distinguished. 1) Clusters or lumps of small pyrite 'threads' again similar to mycels may sometimes attain considerable extensions (Fig. 3d). 2) Thicker filaments of irregular shape which are interpreted as pyritized tubes, although no transition to non-pyritized forms is observed. 3). Isolated 'threads' as occurring in the upper unit (cf. *Trichichnus*).

Section F (8.85–10.75 m)

- Ice-rafted pebbles are abundant and dominantly arranged in layers (Fig. 3e). One of them shows a sharp, erosive lower boundary.
- As biogenic structures, hollow tubes of the larger types of the above section are more frequent than those of the smaller type, although these are occasionally abundant in layers (Fig. 3e).
- No pyritization is found in this section.

Discussion

The suggested stratigraphic subdivision based on the radiograph analysis is subjected to certain technical restrictions due to contrast and resolution of the radiography, and to available cross-section size of the core. It does not appear reasonable, therefore, to try a more detailed subdivision. On the other hand, in this core a wide spectrum of different log parameters offers a chance for better interpretation of the results by comparison of biogenic structures with other environmental data. It should be noted that the suggested subdivision has been made unbiased from the other core data as elaborated in this volume.

The degree of harmony between structural boundaries and other data may be checked by the synoptic diagram of Bjørklund et al. (this volume, Fig. 3). While the first order boundary at 6.8 m sediment depths is found in nearly all data logs, the other boundaries are only partly reflected by other data. Good agreement exists for the section boundary C/D at 4.4 m as regards petrographic and faunistic criteria and the sedimentation rate, which jumps markedly to lower values at the end of section D. Also the upper end of section C coincides exactly with a change in the sedimentation rate. In terms of biogenic structures, these changes refer mainly to the occurrence of feeding-spreiten burrows (cf. *Lophoctenium*) and to the frequency of single tubes or tube-like feeding burrows (*Chondrites*). These relationships agree well with previous results. On the West African continental slope, Wetzel (1981) was able to show that feeding-spreiten associations with *Lophoctenium* (?) are characteristic for areas of medium sedimentation rates (about 0.4 mm y^{-1}; i.e. similar values as calculated for section D in our core). Sediments with higher rates are there characterized either by coarser grain size (and *Scolicia* burrows) or by biodeformational structures, replacing feeding burrows and other trace fossils. In their study of two cores from the central Skagerrak, Jørgensen et al. (1981) also found burrows similar to *Lophoctenium* in those parts of the cores that were characterized by lower sedimentation rates, rather than in parts with increased grain size and/or biodeformational structures. It should be noticed, however, that *Lophoctenium* (Häntzschel 1975), introduced by Wetzel (1981) into the trace fossil inventory of modern sediments, is still problematic in the terminological sense. The nature of these somewhat irregularly constructed feeding burrows is not yet well studied in their three-dimensional shape and in their microstructure.

The drastic change in the ichnological content together with the first order boundary at 6.8 m sediment depth can easily be understood as a response of burrowing fauna to the presence of dropstones and coarse sand layers. The trace fossil distribution off West Africa shows many examples of the sensitivity of burrowing organisms to coarse sediment grains (Wetzel 1981, 1983). Moreover, the lower water temperatures indicated for the Lower Unit of the Skagerrak core by several faunistic parameters described in the present volume, may have had an influence upon the infauna. However, very little is known about the causes of these relationships.

The pyritic structures in the Upper Unit are very similar in character to those found in the deeper core (450 m water depth) of Jørgensen et al. (1981). The pyrite structures trace a secondary ichnofauna within the *Lophoctenium* spreiten too, but more forms resembling *Chondrites* than in the present case are reported there. Although little is known about the ecological meaning of pyritized trace fossils, the coincidence of the three section boundaries defined by pyritized as well as by non-pyritized biogenic structures may possibly be significant for the use of pyritic structures in environmental analysis.

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Mineralogy of material from the Upper Quaternary Skagerrak sediment core GIK 15530-4

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In core GIK 15530-4 the following minerals have been identified: K-feldspar, plagioclase, amphibolite, (pyroxene?), pyrite, calcite, dolomite (siderite?), illite, smectite, chlorite, vermiculite, kaolinite and mixed layered minerals. The ratio between non-clay and clay minerals did not vary much throughout the core, although maximum values for non-clay materials seem to be concentrated in the deepest part of the core.

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The mineralogy and distributional patterns of components of upper Quaternary marine Skagerrak deposits are not well known. Only few detailed surveys have been performed hitherto (e.g. Jørgensen et al. 1981), and especially differences in the sediment composition of the upper Quaternary deposits on the Danish and Norwegian sides of the Skagerrak are poorly known. In conjunction with the detailed investigations of core GIK 15530-4, it therefore seemed important to determine the main mineralogical composition of the sediments penetrated by the core.

Methods

The sediment material from 13 different depths in the core has been examined by X-ray diffraction. The preparation for X-ray diffraction was carried out on inverted filtrate on millipore paper prepared from suspensions. In both sets of samples the following technique has been used:

a) Untreated sample; goniometer velocity 1° per minute, paper velocity 10 mm per minute. In the interval 2° 2θ to 45° 2θ.
b) Slow scan; goniometer velocity 1/8° per minute and paper velocity 25 mm per minute. In the interval 23° 2θ to 27° 2θ.
c) Treatment with ethylene glycole; ordinary velocities in the interval 2° 2θ to 20° 2θ.
d) Heat treatment of the samples to 450° for two hours and directly run at ordinary velocities. In the interval 2° 2θ to 20° 2θ.
e) In samples from four different levels the samples have also been treated with 1 normal HCl. These samples were run at normal velocity in the interval 2° 2θ to 20° 2θ and slow scan in the interval 23° 2θ to 27° 2θ. This was done in order to get a better relative determination of kaoline and chlorite.

In the fine grained fraction samples, the following runs were performed: 1) untreated samples, 2) ethylene glycole treated samples and 3) heat treated samples.

Results and discussion

The following minerals have been identified in all 13 samples: K- feldspar, plagioclase, amphibole (pyroxene?), pyrite, calcite, dolomite (siderite?), illite, smectite, chlorite, vermiculite, kaolinite and mixed layered minerals. From the bulk samples we may conclude that the ratio between non-clay minerals and clay minerals has not differed very much throughout the core. Probably non-clay minerals make up from 40 to 60 % of the total material. Maximum values for non-clay minerals seem to be concentrated in the three deeper samples 905 to 910 cm, 955 to 960 cm and 1055 to 1060 cm. In this case X-ray determination is a very uncertain method and should be followed up with chemical and optical methods. In all samples quartz and carbonates (calcite + dolomite) are the dominating non-clay minerals. Amphibole is more common in the four deepest samples 755 to 760 cm, 905 to 910 cm, 955 to 960 cm, 1055 to 1060 cm compared to

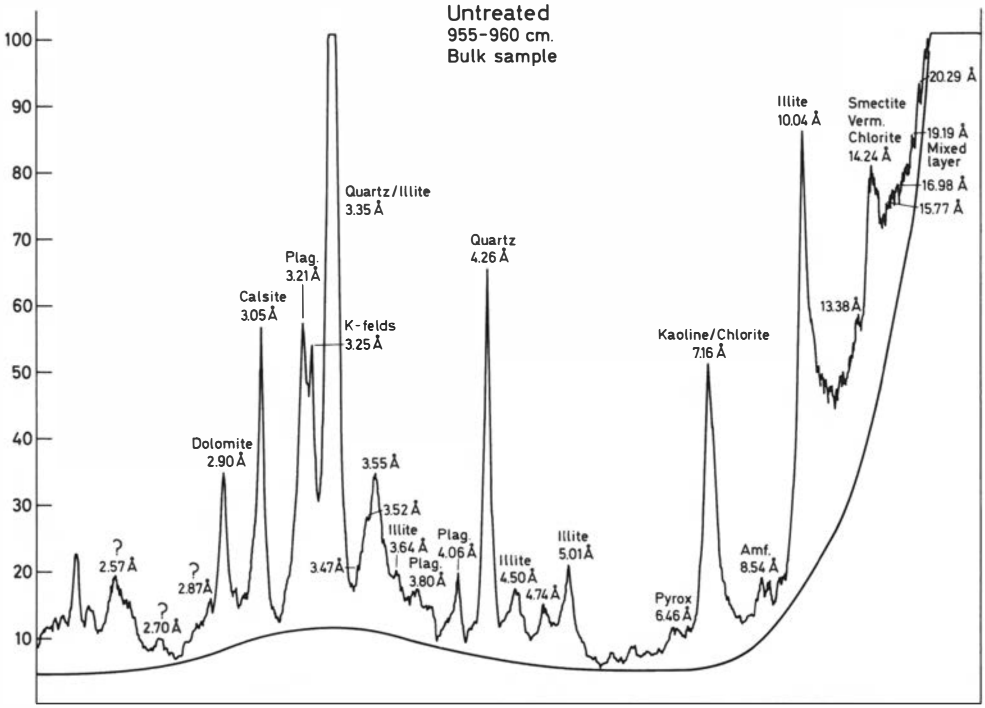

Fig. 1. Example of X-ray diffractogram of typical Skagerrak sediments (core GIK 15530-4, 955–960 cm) showing the minerals discussed.

the nine samples from 5 cm to 710 cm. The ratio plagioclase to K- feldspar seems to be fairly constant from 905 cm up to 205 cm. In the upper two samples from 5 to 35 cm the relative amount of plagioclase seems to be somewhat higher. This may be due to the break-down of plagioclase during diagenesis that has not reached quasi equilibrium in the youngest material. If this explanation is correct, it also indicates that the silt fraction of the sediment originally had a higher than average plagioclase/K- feldspar ratio in the oldest samples.

If we compare the mineralogy of this long Skagerrak core with the Norwegian and Swedish clay sediments above the present shoreline (Jörgensen 1965, Berry & Jörgensen 1971), we will notice several important differences:

1) The quartz/feldspar ratio is higher in the long core than normally found in the Scandinavian glaciomarine sediments of approximately the same grain-size distribution.
2) The ratio of chlorite to illite is lower in the core than normally found in Norwegian clays. In the Norwegian glaciomarine sediments this ratio is usually high (up to 0.5). The presence of amphibole, chlorite and to some extent feldspar may thus be taken as indicators of Scandinavian crystalline rock source, whereas the appearance of kaolinite and smectite indicates a transport from pre-Quaternary sediments.

The feldspar/quartz relationship with the presence of plagioclase and K-feldspar and amphibole indicates *glacial erosion* of crystalline bedrock, whereas smectite and kaolinite and to some extent the mixed layered minerals point to chemical weathering and a source of origin south of the Scandinavian peninsula.

The first conclusion is therefore that all the 13 samples examined have a mixed origin, partly derived from the older sediments and partly derived from Pleistocene tills and Scandinavian derived material.

A total examination of all diagrams necessary for real quantitative or semi-quantitative variations has not been carried out yet. Nevertheless qualitative to semi-qualitative examinations indicate that the amount of material of the 'Northern' type decreases from 755 cm to 710 cm and remains at a lower value up to the top of the core. This may indicate fluvioglacial conditions at the time of deposition of the deepest part of the core with high amounts of melt water which brought more Scandinavian mud out in the water masses of Skagerrak than after the time of deposition of the 755 cm to 760 cm sample. The total picture of the core indicates, however, that the greatest part of the minerals are not derived from Scandinavia and that most of the transport must have been from the North Sea or the Baltic.

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Uranium concentrations in Upper Quaternary Skagerrak deposits

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The vertical distribution of uranium in core GIK 15530-4 is determined by neutron activation analysis. The previously reported distribution of ^{210}Po is normalized according to uranium in order to obtain a stable background level for the determination of sedimentation rates during recent environmental conditions.

The relatively high content of uranium in the bottom part of the core (8–10 m) indicates that the minerals mainly originate from the Scandinavian peninsula. In the upper part of the core (0–4 m), however, a predominance of minerals derived from southern areas may account for the significantly lower values of uranium observed. This is also consistent with mineralogical and geotechnical observations.

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The concentration of uranium in sediments such as those of core GIK 15530-4 from the Skagerrak may vary according to mineralogy and origin of deposits. The distribution of daughter nuclides, including ^{210}Pb and ^{210}Po, will vary accordingly, provided radioactive equilibrium is maintained throughout the chain and no loss of ^{222}Rn occurs. When the vertical distribution of daughter nuclides in the core is used for the determination of sedimentation rates, the contribution from naturally occurring uranium should be distinguished.

In the present paper the vertical distribution of uranium in the core is determined using neutron activation analysis. Furthermore the vertical distribution of ^{210}Po (Erlenkeuser, this volume) is normalized according to uranium. Thus a stable background level for the determination of sedimentation rates under recent environmental conditions is established.

Finally the distribution of uranium in the core is discussed with respect to origin of deposits.

Methods

Samples and standards

Finely crushed oven-dried (105°C) sediment samples of about 200 mg each were accurately weighed and wrapped in aluminium foils. As standards 200 mg accurately weighed samples of BCR-1 (U.S. Geological Standard basalt) were used. In addition standards of uranium were prepared by evaporating 30 μl of a $UO_2\ (NO_3)_2$ standard solution on separate 2×2 cm sheets of aluminium foil.

Irradiation

Samples and standards were irradiated for 3 days in the Jeep II reactor, Kjeller, Norway, at a neutron flux of about 1.5×10^{13} n cm^{-2} s^{-1} and stored for 10 days before transferring into counting vials.

Measurements

A Canberra Ge (Li) detector (20% efficiency and a resolution of 1.89 keV for the 1332.5 keV γ from ^{60}Co) connected to a Canberra 8100 multichannel analyzer was used. Peak location and calculation of peak areas were performed according to the program GAMANL (Gunnik et al. 1967) as described in the work of Salbu et al. (1975).

In the quantitative determination of uranium, γ-energies from ^{239}Np and fission products (^{140}Ba, ^{140}La, ^{141}Ce, ^{152}Eu) were used. The uranium content in the BCR-1 standard was chosen to be 1.7 ppm (Brunfelt & Steinnes 1978) as scattered results (1.2–2.2 ppm) are seen in literature (Morrison et al. 1969).

Table 1. Concentration of uranium in core GIK 15530-4.

LEVEL	U (conc.) ppm
5-10 cm	2.9 ± 0.3
40-45 cm	3.1 ± 0.3
75-80 cm	2.9 ± 0.3
110-115 cm	3.3 ± 0.3
145-150 cm	3.4 ± 0.3
180-185 cm	3.3 ± 0.3
215-220 cm	3.1 ± 0.3
240-245 cm	3.1 ± 0.3
275-280 cm	3.0 ± 0.3
310-315 cm	3.2 ± 0.3
345-350 cm	3.2 ± 0.3
380-385 cm	3.6 ± 0.3
415-420 cm	3.4 ± 0.3
440-445 cm	3.8 ± 0.3
475-480 cm	3.3 ± 0.3
510-515 cm	3.8 ± 0.3
545-550 cm	3.9 ± 0.3
580-585 cm	4.0 ± 0.3
615-620 cm	4.1 ± 0.4
640-645 cm	4.4 ± 0.4
675-680 cm	3.6 ± 0.3
710-715 cm	3.9 ± 0.3
745-750 cm	3.9 ± 0.3
780-785 cm	5.6 ± 0.5
815-820 cm	6.3 ± 0.6
840-845 cm	5.7 ± 0.5
875-880 cm	5.7 ± 0.5
910-915 cm	5.7 ± 0.5
945-950 cm	5.9 ± 0.5
1015-1020 cm	5.9 ± 0.5
1040-1045 cm	5.9 ± 0.5

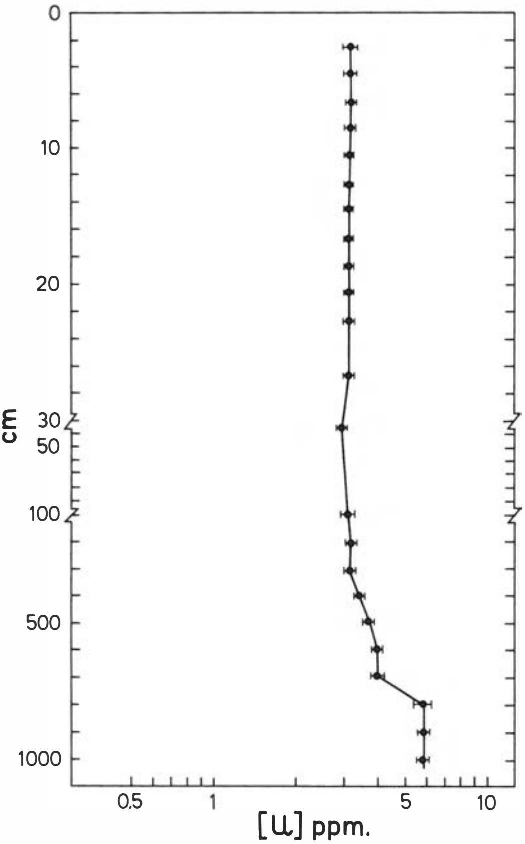

Fig. 1. Distribution of uranium with sample depth.

Results and discussion

The concentration of uranium in the samples investigated is given in Table 1 and distribution of uranium with sample depth is illustrated in Fig. 1. The standard deviations (about 10%) shown are based on counting statistics.

As seen from the figure, the uranium content is about constant (3 ppm) from the surface down to 4 m's depth. Then an increase towards depth is seen (4–8 m), until reaching a constant uranium level of 6 ppm in the bottom of the core (8–10 m).

When the distribution of ^{210}Po-counts/min/5 g d.m. (Fig. 2a) is normalized according to the uranium content (Fig. 2b) the observed increase of ^{210}Po towards depth is significantly reduced.

Thus a more stable background level can be established in order to determine the sedimentation rate under recent environmental conditions.

As the uranium content is a factor of 2 higher in the lower part of the core (8–10 m) compared to the upper part (0–4 m), a change of the source areas of the sediments may be indicated.

In the Fennoscandian Precambrian granitic areas the uranium content is generally higher than average of earth crust (4 ppm, according to Mason 1952). Based on ca. 1000 samples of granites, gneisses and pegmatites from southern Norway, uranium contents of 10 to 40 ppm have been reported (Rosenqvist 1947). In samples of upper Cambrian carbon- and sulphur-rich alunshales from southern Norway and Sweden, the uranium content amounts from 10 to more than 200 ppm (op.cit.). However, the uranium content in the sedimentary areas south of the Scandinavian peninsula, Denmark and Germany, is normally significantly lower (Mason 1952, Day 1963). Sedimentary argillites of low carbon content and limestones have uranium contents from less than 1 ppm up to crust average. As demonstrated by Wassmann (this volume) the lower part of the core with the highest uranium content has the lowest content of TOC.

The distribution of uranium may therefore indicate that a major fraction of minerals in the lower part of the core (8–10 m) may originate from the uranium-rich rocks of the Scandinavian peninsula, while a contribution from southern areas should be predominant in the upper (0–4 m) part of the core. This is also consistent with the mineralogical and geotechnical observations.

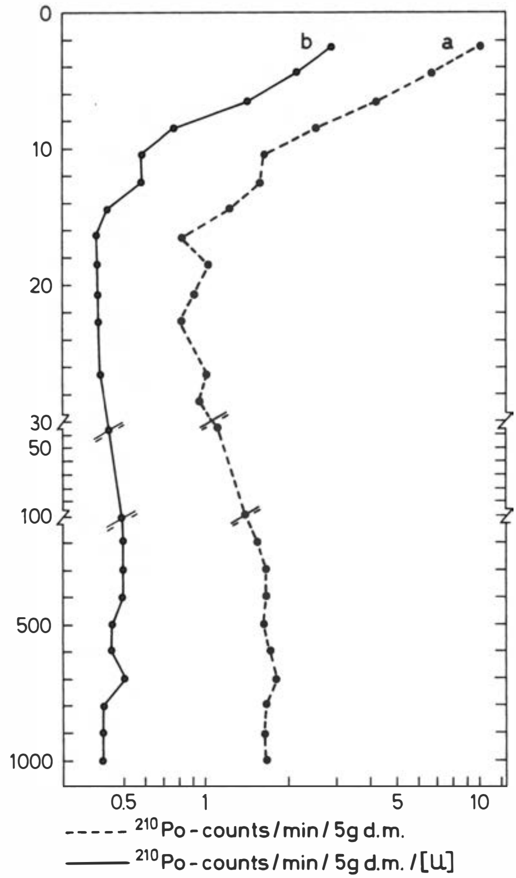

Fig. 2. Distribution of ^{210}Po (a) normalized according to the uranium content (b).

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Coarse sediment components in Upper Quaternary marine Skagerrak deposits

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The coarse fractions of the Upper Quaternary marine sediment sections of core GIK 15530-4 consist of biogenic, terrigenous detrital, and authigenic components. In general, the core can be subdivided into an upper unit whose sand fractions are dominated by biogenic (mainly calcareous benthic foraminifers, some planktonic foraminifers, bivalves, pteropods, radiolarians, sponge spicules, diatoms), whereas a lower unit is largely dominated by terrigenous clastic components (frequently rock fragments). The boundary between these two units is situated at the 600–650 cm level (close to the boundary between Younger Dryas and Preboreal).

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Studies of the composition of coarse fractions provide a simple, but effective tool to describe aspects of the depositional environment and of the sources from where these sediments originate. An analysis of the coarse fractions of the sediments penetrated by core GIK 15530-4 from the outer Skagerrak has therefore been carried out. These sediments can be traced across a wide area of the Skagerrak using shallow seismic reflection profiles (Stabell et al., this volume, Fig. 3), see also van Weering (1982). The core which has been analysed for this study also penetrated the uppermost clear seismic reflector having any regional significance.

Although the core appeared macroscopically rather homogenous, both with respect to sedimentary structures and textural properties, the composition of the sand-sized material changed drastically downcore. Therefore it seemed desirable to quantify these changes and to locate the position of any boundary. A simplified version of the coarse fraction analysis method developed by Sarnthein (1971) has been applied here.

Methods

To assess the composition of the sand fractions, a set of 18 wet bulk sediment samples was washed through a 0.063 mm-sieve after drying and weighing the bulk sample. After determining the weight of the material larger than 0.063 mm (Table 1), the sand-sized material was split into three size fractions: >1.0 mm, 1.0–0.125 mm and 0.125–0.063 mm. The size of the largest terrigenous clasts (Table 1) was determined measuring grains from the two coarse fractions (see Stabell et al., Fig. 6, this volume), and the composition of the planktonic foraminiferal faunas

Table 1. Proportion of sand-sized (> 0.063 mm) material and diameter of largest terrigenous detrital fragments in core GIK 15530-4.

Core level (in cm below sediment surface)	Sand-sized material (in weight percent of total sediment)	Diameter of largest grain (mm)
5 - 10	0.65	0.85
55 60	0.87	0.40
105 - 110	0.30	0.75
205 - 210	0.39	1.25
305 - 310	0.32	0.25
405 - 410	0.79	0.35
505 - 510	1.09	0.90
555 - 560	1.72	1.00
605 - 610	1.49	2.00
655 - 660	3.84	3.25
705 - 710	3.84	4.00
710 - 715	5.67	n.m.[x]
720 - 725	4.75	n.m.[x]
730 - 735	2.20	n.m.[x]
740 - 745	4.08	n.m.[x]
755 - 760	2.95	3.40
805 - 810	4.04	3.00
855 - 860	11.29	5.50
905 - 910	3.74	3.15
955 - 960	2.58	1.25
1005 - 1010	5.76	3.25
1055 - 1060	1.30	1.15

[x] n.m. = not measured

Table 2. Distribution of grain categories in the coarse fraction (> 0.063 mm) of core GIK 15530-4.

Level in cm	Nos of grains counted	Arenaceous forams	Arenaceous forams fragm.	Calc. benthic forams	Calc. benthic forams fragm.	Planktonic forams	Pterodpods	Benthic gastropods (juv.)	Bivalves	Bivalves, juv.	Bivalves, fragm.	Bryozoans
5- 10	307	1		34		.3	.3		.3			
50- 66	343		1	36	26	.3			.3			
105-110	226		1	27	20	.5	1	5	.5	5		
205-210	242			31	11	1	.5		7	1		
305-310	224			45	23	.5	.5		4	.5	10	
405-410	256			56	20	.3	.7				5	
505-510	177		1	35	51	1					2	
555-560	178			23	11	1	1			1	1	
605-610	34			33	11							
655-660	256			22	9							
705-710	196			5	1							
755-760	235			9	3							
805-810	205	.5		8	.5	.5						
855-860	316	.3		2								
905-910	245			3	4							
955-960	351			3								
1005-1010	229			3		.4						
1055-1060	187			1								

Level in cm	Echinoderms	Calc. "pipes"	Ostracods	Other crustaceans	Radiolarians	Diatoms	Sponge spicules	Fish bones	Dinoflagellate cysts	Org. material	Fibres	Pellets	Dolomite rhombs	Pyrite	Volc. glass	Quartz	Mica	Coal ash	Glauconite	Rock fragm.	Brown crusts
5- 10	3		2				.3		.6	.6	.6	53		1		6		.3			
50- 66	1		.3		.3					.3		29		3		4	.3				
105-110	2		2				3			4		27		.5	5	9	1				
205-210	.5	.5	.5				.5			3			5	1		42	5			5	5
305-310	.5		.5				5	.5		4		7		2		2					
405-410	.3	3	4				1			3		1		4		3				1	
505-510	2	1	2				1		1	3		21		11		2	1				
555-560	4	1	1				1	1		6		30		6		8					2
605-610	1		5							17		6		24		4					
655-660	2		.4			.2	.4	.4		2		2		18		39	2			1	
705-710			1			19	1			2				23		45			1	3	
755-760	.3					30	.3			3				.6		57			.3	1	.3
805-810	2					9				1				.5	.5	75	.5		5	1	
855-860						5				1				3		89	3			.3	
905-910						27				1		4		6		41				2	20
955-960	3					1	.3			1						15				1	81
1005-1010	.4		.4			.4				1					.4	61	.4		.4	7	25
1055-1060	.5					.5				3		1				18	1				75

was counted in the two fine sand fractions (Thiede, this volume).

The following grain types were discerned (Table 2):

1. *Biogenic components:*

1.1. Calcareous material: Benthic foraminifers, fragments of benthic foraminifers, planktonic foraminifers, pteropods, benthic gastropods (larvae), bivalves, bivalve larvae, bivalve fragments, bryozoans, echinoderms, calcareous 'pipes' of unknown origin, ostracods.

1.2. Opaline material: radiolarians, diatoms, sponge spicules.

1.3. Organic-walled components: dinoflagellate cysts, organic 'brown' components, fibers.

1.4. Other biogenic material: Arenaceous foraminifers, fragments of arenaceous foraminifers, crustaceans (other than ostracods), fish bones, fecal pellets.

2. *Terrigenous detrital components:*

quartz, mica, rock fragments.

3. *Authigenic and other components:*

dolomite rhombs, pyrite, glauconite, 'brown crusts', volcanic glass, coal ash.

The distribution of coarse fraction components in core GIK 15530-4

The distribution of the quantitatively most important components (Table 2) is shown in Fig. 1. The core can be subdivided into an upper unit whose sand fractions are dominated by biogenic components, whereas a lower unit is largely dominated by terrigenous clastic components (rock fragments). The boundary between these two units is situated at the 600–650 cm level.

The most important biogenic components are calcareous benthic foraminifers (Nagy & Qvale, this volume) which down to a level of 650 cm represent more than 20 % of the sand fractions,

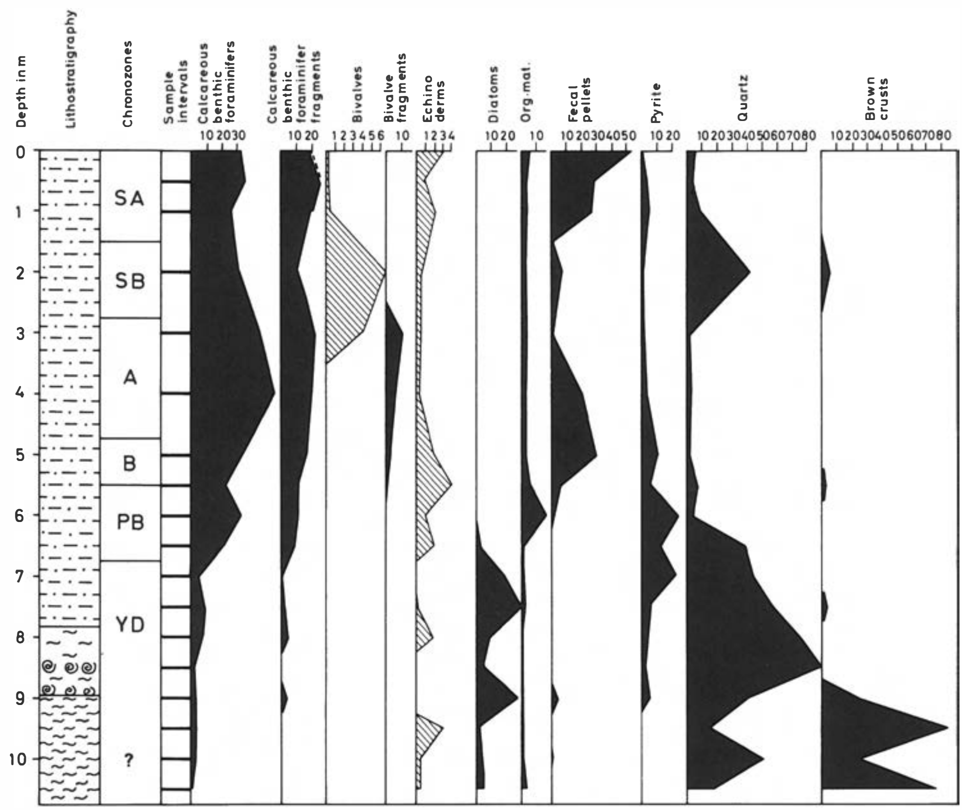

Fig. 1. Distribution of main coarse fraction (0.125–1.0 mm) components in core GIK 15530-4 (in per cent of total grain assemblages). Note differences in scales.

whereas their proportion decreases to a few per cent of the sand fractions in the lowermost 3.5 m of the core. Their maximum proportion has been observed around the 400 cm level. 10–20 % of the grain assemblages consist of fragments of calcareous benthic foraminifers. The preservation of this calcareous shell material is best in the core levels below 7–8 m beneath the sediment surface. It seems questionable if this fragmentation is due to dissolution (Alexandersson 1978), because virtually no fragments have been observed in the deeper intervals.

Other benthic remains observed in this core are bivalves with a maximum at the 200 cm level, bivalve fragments with a maximum at the 300 cm level, echinoderms with a maximum at the 550 cm level, and a number of other components which occur in minor quantities throughout the core or in certain intervals (Table 2).

Planktonic organisms with calcareous shells comprise Quaternary planktonic foraminifers (Thiede, this volume), occasionally reworked upper Cretaceous and Paleogene planktonic foraminifers, and very rare modern pteropods. The pteropod shells belong to the species *Limacina lesseurii* and *L. trochiformis*.

Radiolarians and sponge spicules (Bjørklund, this volume) are relatively rare opaline sand fraction components, but benthic and planktonic diatoms (Stabell, this volume) occur in high abundances in the 650–950 cm interval where they reach up to 30 % of the total grain assemblages. In the sediments above this interval they have been found only in minor quantities and usually poorly preserved. Organic 'brown' fibre-like substances occur most frequently at the 600 cm level, where also the abundance of pyrite grains is highest, although these have been observed throughout the entire core in minor quantities. Sand-sized oval-shaped fecal pellets occur quite frequently above the 600 cm level (Table 2) and with upwards increasing frequency in the shallowest metre of the core. Quartz and diverse assemblage of rock fragments are rare in the upper 6 m of the core, but they are very characteristic components of the coarse fractions in the lower 4 m. It is clear from their distribution that they are not evenly distributed throughout the entire interval, but that they are enriched in horizons, although they seem to 'float' in a matrix of fine-grained sediment (see also Werner, this volume). Most of the rock fragments consist of crystalline or metamorphic rocks of a gneissic or granitoid composition, but some have obviously originated from mafic rocks.

A very conspicuous component of the coarse fraction grain assemblages are broken-up 'brown crusts' of iron oxides which are entirely restricted to the lowermost 2 m of the core. Enrichments of iron oxides in horizons intercalated into sediments deposited towards the end of the last Glacial have also been observed in the Norwegian-Greenland Sea (Jansen 1981) and in the Atlantic Ocean (McGeary & Damuth 1973), and are sometimes found to mask hiatuses. Although it is has yet to be proven that they document an isochronous paleooceanographic event, it is interesting to note that they occur in sediments deposited in the same stratigraphic setting. In core GIK 15530-4 this interval can be dated to a time during the early part of Younger Dryas and lasting for <1000 years.

Conclusions

1. The presence of well preserved though scarce foraminiferal faunas down to the deepest core level documents together with the occurrence of other marine fossils that the Skagerrak has been a marine basin since the earliest time span represented by the sediments of this core, which cover an age interval from Younger Dryas to Subatlantic times.
2. The coarse fractions of these sediments are composed of diverse grain assemblages of biogenic, terrigenous and authigenic origin.
3. A major change in sedimentation during the later Younger Dryas is clearly reflected in the coarse fractions. The lower part of the core of Younger Dryas and Alleröd age contains coarse fractions which are dominated by clastic detrital components and diatoms. The coarse fractions of the upper part of the core of Preboreal to Subatlantic age comprise dominantly biogenic benthic calcareous components.
4. Conspicuous 'brown crusts' have been found frequently in the lowermost 2 m of the core of Younger Dryas age. They can possibly be related to enrichments of iron oxides in sediments which were deposited towards the end of the Last Glacial.

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Acid resistant components of organic matter in Upper Quaternary Skagerrak sediments

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The dominantly terrigenous fine-grained marine Upper Quaternary Skagerrak deposits contain acid-resistant organic matter whose composition and distribution has been studied in core GIK 15530-4. Inert, woody, undifferentiated and amorphous materials have been discerned and their downcore distribution recorded.

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The Upper Quaternary marine sediments of the Skagerrak are known to contain varying amounts of organic matter, both of land-derived materials and of marine origin. They have been described in some detail by Combaz et al. (1977). In general the approach has been to investigate the composition of this material with methods of visual kerogen analysis which have been developed mainly by petroleum geologists. In this study we present observations on such materials which have been collected from core GIK 15530-4 (Stabell et al., this volume).

Methods

The analyses were made on palynological preparations (HCl and HF dissolution of mineral matter); strew mounts of total components and mounts of the fraction >20 μm were studied in transmitted light under the microscope.

Results

Acid resistant components were differentiated in the following categories:

Inert material: 'Fossil charcoal' or fusinite particles, entirely opaque with irregular angularity (differentiation from pyrite usually unproblematic).

Woody material: Transparent or at least light transmitting particles possessing wood structure.

Undifferentiated material: Light transmitting fragments of cuticle, digested wood etc., 'resinous' lumps. Particles were usually too small to make further differentiation of this category a worthwhile excercise.

Amorphous material: Colourless and structureless particles, either finely dispersed or aggregated. Origin and nature of this matter are uncertain. UV-excitation produces only weak fluorescence of pale yellow-greenish colour which seems to exclude a significant contribution of algal or liptinitic material. Most probably it is composed of other finely disseminated organic material and acid resistant mineral compounds in a colloidal association.

Total organic material (TOM) preparations from all samples are completely dominated by amorphous material. The tendency to aggregation, however, varies considerably. Down to 300 cm, 'lumpy' aggregates are prominent, from 300 cm down to 600 cm dispersed material is predominant; below 600 cm there is a marked tendency to variation from sample to sample – some have prominent amounts of lumps whose shape is suggestive of fecal pellets. TOM retained on 20 μm mesh has been particle counted for 9 samples and the results are shown in a percentage frequency diagram (Fig. 1).

Discussion

In the terrigenous components and palynomorph frequency curves of this core the only features

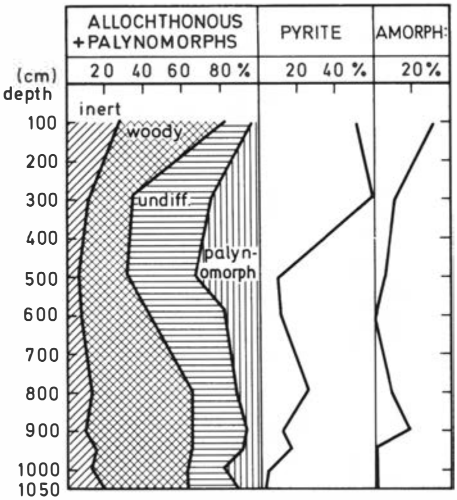

Fig. 1. Distribution of acid resistant particles >20 μm in core GIK 15530-4. The individual counts of terrigenous (allochthonous) components (inert, woody and undifferentiated) and the total palynomorphs (pollen, spores and dinoflagellate cysts) are presented as percentages of total terrigenous plus palynomorphs. Pyrite and amorphous are recorded as percentages of total terrigenous plus palynomorphs.

worth noting are the relatively low woody and the high palynomorph counts between 300 and 500 cm. The palynomorph high is due to higher cyst frequencies (1 1/2 to 2 times the pollen, whilst usually only 1/2 the pollen frequency). It is perhaps interesting to note that this woody low and palynomorph high coincide with the interval of higher accumulation rates during Atlantic time. (Thiede, this vol., Fig. 1).

A striking feature throughout the core is the variation in pyrite, which occurs in single crystals as well as aggregates. Their frequency varies, as well as the shape of the aggregates which is often suggestive of having been formed in association with organic structures; frequently aggregates occur inside pollen grains, cysts, foraminifer tests etc. The observed variations in the pyrite, except for the lesser amounts above 500 cm, make no systematic pattern with the present degree of sampling resolution.

Reference

Combaz, A., Bellet, J., Poulain, D., Caratini, Cl. & Tissot, C. 1977: Étude microscopique de la matière organique de sédiments quaternaires de la mer de Norvège. ORGON I Mer de Norvège – Aug. 1974, 139–173.

Late Quaternary evolution of the Skagerrak area as mirrored by calcareous nannoplankton

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Mikkelsen, N.: Late Quaternary evolution of the Skagerrak area as mirrored by calcareous nannoplankton. *Norsk Geologisk Tidsskrift,* Vol. 65, pp. 87–90. Oslo 1985. ISSN 0029-196X.

Calcareous nannoplankton in the 10.7 m long Skagerrak core GIK 15530-4 from the southern slope of the Norwegian Channel provides information on the Late Weichselian and Holocene climatic evolution of the area. A very rare occurrence of coccoliths in the lower part of the core is parallelled by an absence of Quaternary forms in the poorly preserved assemblages of reworked Upper Cretaceous and mid-Tertiary species. The coccolith assemblages in this lower part of the core may correspond to the period of the Baltic Ice Lake, from where terrestrial material in great quantities was transported into the restricted Skagerrak area. With the formation of the Yoldia Sea, more open-marine conditions were established as indicated by the first notable appearance of the Quaternary coccolith species *Emiliania huxleyi* in the core section between 700 and 650 cm. Stabilization of the oceanographic environment to present day conditions is reflected by the coccolith assemblage in the upper two metres of the core. Here primary productivity of a present day composition and magnitude is mirrored by a well-preserved Quaternary coccolith assemblage.

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The sedimentary sequences of the deeper parts of the Skagerrak area contain abundant information on the Late Weichselian and Holocene evolution of the Skagerrak area (Holtedahl & Bjerkli 1975, van Weering 1975, Jansen et el. 1979).

Coccoliths are common constituents of marine sediments. Algae producing coccoliths are living in the euphotic zone of the oceans and do not thrive in brackish- or fresh water environments. As the species composition of recent nannoplankton assemblages is closely related to the sea surface temperature (McIntyre & Bé 1967), fossil assemblages provide useful information on past climatic conditions.

The Pleistocene-Holocene evolution of the North Atlantic and the North Sea adjacent to Skagerrak has been delineated by numerous studies (e.g. Flinn 1967, Jansen 1976, Kellogg 1976) and also illustrated by studies of the coccoliths in deep-sea cores (McIntyre 1967). The Pleistocene-Holocene transition in the Kattegat area as mirrored by coccoliths was recently studied in cores from the Swedish west coast (Mikkelsen & Perch-Nielsen 1982).

The information derived from coccolith studies, especially of near-coast sediments may, however, be hampered by reworking. The minute size of the coccoliths highly facilitates reworking of the calcareous plates together with other sedimentary components. The reworked coccoliths nevertheless provide useful information on, for example, the source area for the transported material and the direction of transport.

The present paper provides some information on the Pleistocene-Holocene evolution of the Skagerrak area as indicated by Quaternary and reworked older coccoliths in core GIK 15530-4.

Material and methods

From the 10.7 m long Skagerrak core GIK 15530-4, retrieved from a waterdepth of 325 m at the southern slope of the Norwegian Channel, 18 samples have been investigated. Smear slides were prepared from all samples by smearing a small piece of sample in a water drop on a glass slide. The dried slides were mounted with cover slips containing Canadabalsam, and cooked on a hot plate for about half a minute to harden the balsam. The samples were studied with a Leitz polarizing microscope at a magnification of 1400 times.

Observations were made on the assemblage composition and preservation of the coccoliths, and semi-quantitative estimates were made on the abundance of coccoliths.

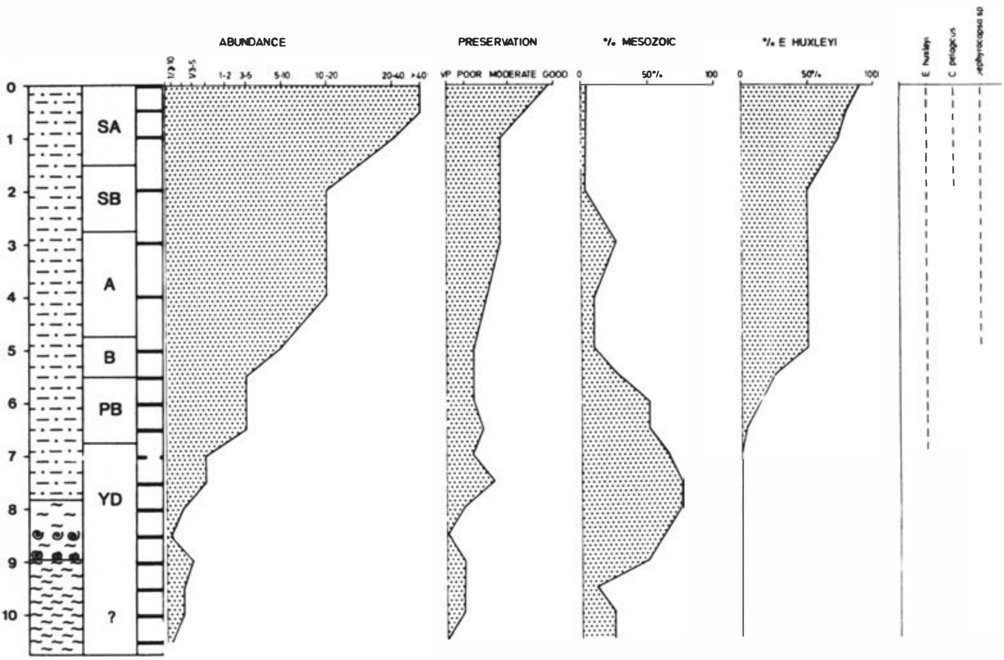

Fig. 1. Distribution of calcareous nannofossils in the Skagerrak core. Coccolith abundance is given as number of specimens on a smear slide per field of view at a magnification of 1400 times. The reworked Mesozoic coccoliths are recorded as percentage of the total assemblage, which is composed of Mesozoic, Tertiary and Quaternary species. The abundance of *E. huxleyi,* which is totally dominating the Quaternary assemblage, is likewise given as percentage of the total assemblage. VP = Very poor.

Results and discussion

The downcore variations in assemblage composition provide some distinct indications of the changing conditions in the Skagerrak area during the Pleistocene-Holocene transition. The coccolith species recorded in the 10.7 m long core represent three distinct geological periods: the Cretaceous (e.g. *Reinhardtites anthophorus, R. levis, Arkhangelskiella cymbiformis, Lithraphidites quadratus* from the Campanian and Maastrichtian), the Tertiary (e.g. *Neococcolithes dubius, Chiasmolithus* sp., *Reticulofenestra umbilica* from the Eocene-Oligocene), and the Quaternary (e.g. *Gephyrocapsa* sp., *Emiliania huxleyi*). The fairly consistent occurrence of Cretaceous and Tertiary coccoliths in the lower part of the core points to pronounced reworking of older strata. The geology of pre-Quaternary sub-sea outcrops in Kattegat and Skagerrak remains almost uninvestigated at present, and outcrops of the above-mentioned Cretaceous and Tertiary ages are presently known only from on-shore Denmark and southern Sweden (e.g. Forchheimer 1972, Perch-Nielsen 1979) (Fig. 1, Bjørklund et al., this volume). It seems likely, however, that most of the reworked coccoliths have an origin south and south-east of the Scandinavian peninsula.

The ratio between the abundance of Mesozoic (Cretaceous) and Cenozoic (Tertiary and Quaternary) coccoliths shows a distinct dominance of Mesozoic forms in the lower part of the core. The degree of reworking decreases remarkably above a core depth of 550 cm, and is negligible in the upper two metres (Fig. 1).

The notable difference in the degree of reworking may be interpreted as a result of the evolution of the hydrographic conditions of the Skagerrak. The lower part of the core, which is totally dominated by reworked coccoliths, may be evidence of rather restricted marine conditions, where the influx of terrigenous and coastal material (including reworked coccoliths) has dominated the sedimentary components.

Stabilization of the oceanic environment to present-day conditions is reflected by the assemblage composition of the coccoliths in the upper two metres of the core. The Quaternary species *Emiliania huxleyi,* which evolved during the

Quaternary and is living in present-day oceans in an optimal temperature range of approximately 2 to 14°C (McIntyre & Bé 1967), thus provides the best signal of the changing marine conditions in the Skagerrak area. The marked increase of this species from 200 cm to the top of the core is presumably a function of the transition to present-day conditions of primary productivity and the hydrography. *Emiliania huxleyi* was not observed in samples below 700 cm. The apparent lack may be attributed to a highly reduced or missing production of this species, and thus to a masking effect by reworked forms which were transported into the area from the surroundings by, for example, melt-water rivers.

Small specimens of *Gephyrocapsa* sp. make their first appearance at a core level of 500 cm and are consistently present to the top. *Coccolithus pelagicus*, a species presently living in a temperature range of 8–14°C (McIntyre & Bé 1967), provides another paleoclimatic signal by its fairly consistent occurrence in the uppermost 2 metres of the core. The representation of the species may thus point to a cooling trend and simultaneously signal the first consistent incursion of Atlantic water into Skagerrak.

The abundance of coccoliths in a sedimentary sequence is highly influenced by a number of factors as, for example, primary productivity, dissolution, and dilution. The coccolith abundance in the Skagerrak core shows a distinct increase from the bottom to the top of the core (Fig. 1). A conspicuous increase takes place between 700 and 650 cm. The extremely low abundance of coccoliths in the interval around 850 cm may be attributed to various events. It is likely that it is related to heavy dilution caused by, for example, the drainage of the Baltic Ice Lake and the adjacent ice shield, although post-burial dissolution cannot be excluded. The steady increase from poorly preserved coccoliths in the core bottom to well-preserved coccoliths in the upper part thus points to the impact of carbonate dissolution on the coccolith abundance. Shallow-water carbonate dissolution (Alexandersson 1975) and post burial dissolution undoubtedly altered the original abundance and species composition of the coccolith assemblages by removing more solution-prone species. However, dissolution cannot account for the total lack of Quaternary species in the bottom part of the core. The steadily increasing abundance of Quaternary coccoliths in the upper part of the core indicates the combined effect of increasing primary productivity and the reduced influx of reworked coccoliths.

Conclusion

Core GIK 15530-4 from the southern slope of the Norwegian Channel provides an excellent opportunity to study the impact of the terrestrial environment on the marine realm caused by the climatically induced ice retreat during Late Quaternary.

The coccoliths, being a part of the marine sedimentary components, mirror on a gross scale the changing physical conditions in the Skagerrak area.

Some distinct events are notable in the core. In the lower part, and especially around a core level of 850 cm, coccoliths are very rare and only reworked Cretaceous and Tertiary forms are present. This coccolith event may be a reflection of the drainage of the Baltic Ice lake, as abundant terrestrial material during this period was transported into the Skagerrak area, where coccolith production as a consequence was masked. A distinct increase in coccolith abundance takes place between 700 and 650 cm in the core. This increase is related not only to a drop in the accumulation rate, but also to the changing hydrographic conditions. As it coincides with the appearance of the Quaternary species *Emiliania huxleyi*, this point may represent the formation of the Yoldia Sea, where a closed-bay situation with heavy fresh water run-off from the Baltic Ice Lake is replaced by more open-marine conditions.

The transgression and the following regression connected with the formation of the Ancylus Lake apparently did not leave any notable coccolith imprint in the core. The increased coccolith abundance and the prominent population of *Emiliania huxleyi* may on the other hand mirror the Holocene transgression and the formation of the Littorina Sea.

Stabilization to present-day conditions of the marine environment in the Skagerrak is marked by the total dominance of *E. huxleyi* in the upper part of the core. Also, a change in the dominant current direction over time from an easterly south-easterly to a more westerly direction from the Atlantic into the Skagerrak may be deduced both from the general species composition of the Quaternary assemblage in the upper part of the core, and from the distinct drop in reworked Cretaceous and Tertiary coccoliths.

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Diatoms in Upper Quaternary Skagerrak sediments

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Diatoms occur only in the lower 5 m and in the uppermost sample of the core. The lower, diatomaceous part of the core can be divided into two main zones: a lower (975-675 cm) zone characterized by polar, planktonic species (*Thalassiosira antarctica, Nitzschia cylindrus* and *Actinocyclus* cf. *alienus* var. *arctica*) and an upper (675–625 cm) zone characterized by coastal, temperate species. The upper zone is dominated by *Paralia sulcata.* The absence of diatoms between 525 cm and 25 cm is interpreted as being due to opal dissolution.

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Diatoms are photosynthetic, single-celled, siliceous algae. They are thereby restricted to the euphotic zone, where they can be found both as planktonic and benthic organisms. They are important constituents in both marine and lacustrine aquatic environments, and the various diatom species are more or less exclusively bound to specific salinity ranges. Diatoms are used as a stratigraphic tool to render information on paleoclimate, paleosalinities, etc.

Earlier diatom investigations from the Skagerrak area have been limited to studies of living marine diatoms and near-shore diatom successions (mainly for shore-displacement investigations).

Methods

The eighteen selected samples from core GIK 15530-4 were analysed for their diatom content. The samples were prepared according to the procedure described in Kaland & Stabell (1981), using *Lycopodium* spores as indicators. Approximately 0.2 g of sample was oxidized in 30% H_2O_2, the sample was diluted with distilled water, and 2 *Lycopodium* capsules (Stabell & Henningsmoen 1981), each containing 38 000 spores, were added. The samples were then repeatedly washed in distilled water to remove clay particles and colloids. Slides were prepared using a few drops of concentrated sample which were spread over a coverglass, dried and embedded in a strongly refractive embedding medium (Hyrax = 1.66). Analysis was carried out at maximum magnification of the light microscope (about 1250×) with oil immersion and phase contrast. When necessary, a scanning electron microscope was used to obtain a higher magnification. The identification of the different diatom taxa was carried out mainly with reference to Hustedt (1930–66) and Cleve-Euler (1951–55). Information concerning the ecological requirements of the different taxa was obtained from the above references and from Simonsen (1962) and Hendey (1964). To obtain a statistically valid result at least 200 diatom valves should be counted (Miller 1964). As the diatom valve frequency varied greatly in the different samples, the number of diatom valves counted varies from 27 to 537.

Ecological grouping and diagram construction

All the taxa identified are listed with their relative frequencies in Table 1. A diagram with absolute number of valves per g sediment, with separate curves displaying absolute frequency of the most characteristic diatoms and with relative percentages of planktonic and benthic valves, are shown in Fig. 1. The absolute frequencies of *Thalassiothrix longissima* are based on fragments of this taxon. Due to the shape of this taxon which is long (1–4 mm) and narrow (3–4 μm), it breaks easily and is only to be found as fragments in the samples. *Actinocyclus* sp. is relatively small, and the specimens could only be identified as belonging to this genus by the use of the Scanning Electron Microscope (Fig. 2). It resembles most closely *A. alienus* var. *arctica* Grunow and is therefore included among the polar planktonic species.

Table 1. Percentage distribution of all diatom taxa encountered in core GIK 15530-4. Relative frequency of Thalassiothrix longissima shown separately.

Core level (cm)	Cymbella ventricosa	Paralia ornata	Pleurosigma sp.	Opephora gemmata	Coscinodiscus oculus-iridis	Thalassionema niztschioides	Raphoneis surirella	Thalassiosira antarctica	Paralia sulcata	Nitzschia cylindrus	Actinocyclus sp.	Navicula directa	Nitzschia sp.	Licmophora sp.	Thalassiosira spp.	Navicula sp.	Coscinodiscus spp.	Diploneis smithii	Cocconeis spp.	Melosira islandica	Grammatophora angulosa	Actinoptychus undulatus	Biddulphia sinensis	Cocconeis scutellum	Hyalodiscus scoticus	Diploneis sp.	Cymatosira belgica	Fragilaria sp.	Delphineis penelliptica	Unidentified	Thalassiothrix longissima
0,5							6		70										3			+					3	3	3	9	
50-55									+																	+	+				
100-105																															
200-205	BARREN																														
300-305																															
400-405																															
500-505									+																						
550-555									86																10	2		2			
600-605							21	+	64			1					4					7			+	+		+			0.1
650-655					+	2	18	8	53		1					+	+		4			8	1	+	+	+	2	2		+	0.3
700-705						1	7	44	21	7	11	1			4	2	1	1			1	1									0.8
750-755					1		4	71	5	2	14																				1.6
800-805					1		8	27	1	15	37			1			6	1		1	1										3.8
850-855				1	1		2	66	2	7	6				14			1	1												0.9
900-905				1			1	37	3	37	3			1	16																4.4
950-955				6			3	29	1	26	22	2	1	1	4	1	1														4.0
1000-1005				10			10	30	15	18	15	2																		2	1.0
1050-1055	4	4	8	12	4	8	15	44	4																					2	0.2

Diatom distribution throughout the core

Fig. 1 and Table 1 illustrate the diatom distribution in the sediments. Five diatom assemblage zones were distinguished.

Zone A (1050–975 cm). Diatoms are rare ($<5 \times 10^5$ valves/g sediment). Mixed diatom floras comprise planktonic and benthic taxa. The assemblage is dominated by the polar planktonic taxon *Thalassiosira antarctica,* mixed with Tertiary, reworked taxa (*Paralia ornata, Opephora gemmata*). The diatoms in this zone ought to be characterized as reworked.

Zone B (975–675 cm). Diatoms are common ($5–25 \times 10^5$ valves/g sediment). The zone is characterized by a mainly polar planktonic assemblage (resting spores of *Thalassiosira antarctica*, *Nitzschia cylindrus, Actinocyclus* cf. *alienus* var. *arctica*). Both *T. antarctica* and *N. cylindrus* are planktonic, cold water and bipolar species, belonging to extreme inshore waters and/or ice (Hasle 1976). *A. alienus* var. *arctica* lives in the Arctic ocean (Hustedt 1930–66). These three species make up 76–87 % of the assemblage from 975 cm to 725 cm in zone B. The upper part of this zone, sample 700 cm, is transitional, the polar planktonic species are reduced to 62 % and there is an increase of *Paralia sulcata* to 21 %.

Zone C (675–625 cm). Diatoms are abundant ($>25 \times 10^5$ valves/g sediment). The zone is characterized by only one sample, 650 cm. The assemblage is characterized by coastal, temperate taxa. *Paralia sulcata* dominates the assemblage and there are high frequencies of *Raphoneis surirella* and *Actinoptychus undulatus*.

Zone D (625–25 cm). The zone starts with a transitional section, 625–525, rare in diatoms, and dominated by *Paralia sulcata*. There is evidence of dissolution of the diatom frustules in this

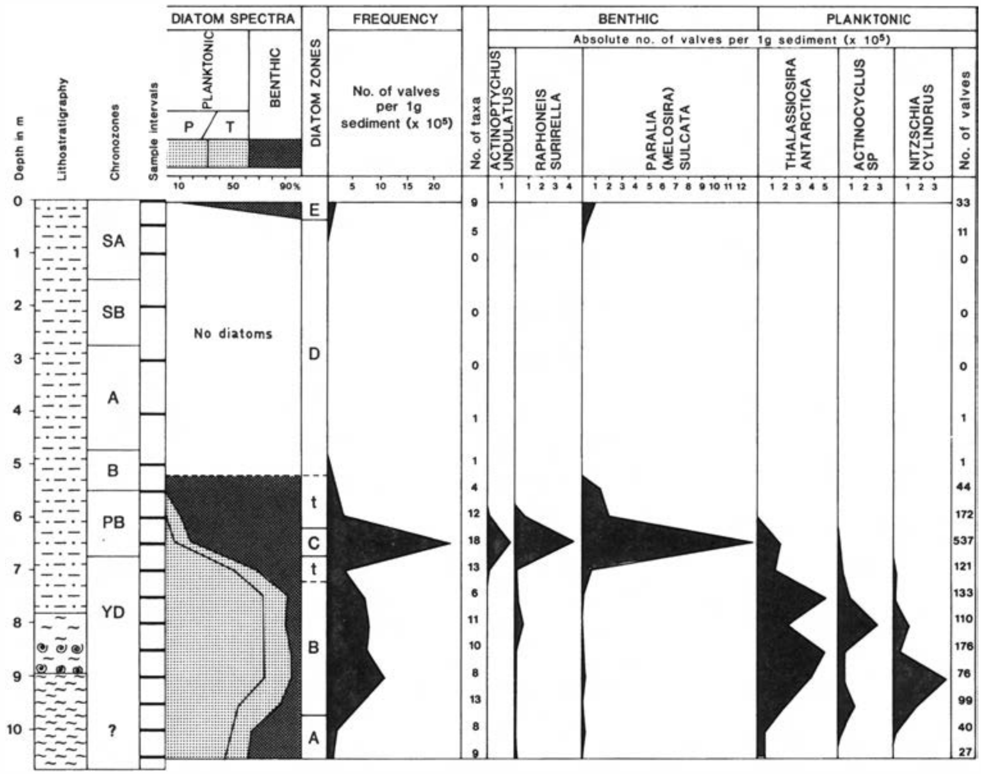


Fig. 1. Diatom distribution (spectra) in core GIK 15530-4 (P = polar, T = temperate) with diatom frequency, number of taxa and separate curves of characteristic species (absolute frequency). Diatom zones in alphabetical order (A–E), t = transitional. No. of values (column to the right) indicate numbers of diatoms counted per sample. These numbers are the basis for the calculations of no. of valves per 1 g sediment.

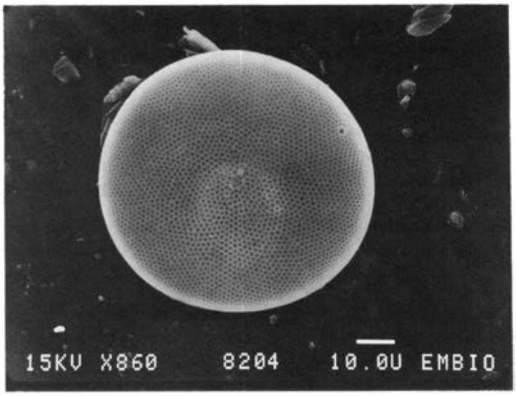
transitional zone; the absence of diatoms from 525 to 25 cm is therefore interpreted as being due to opal dissolution.

Zone E (0–25 cm). Diatoms are rare. The assemblage can be characterized as coastal and temperate, with a dominance of *Paralia sulcata.*

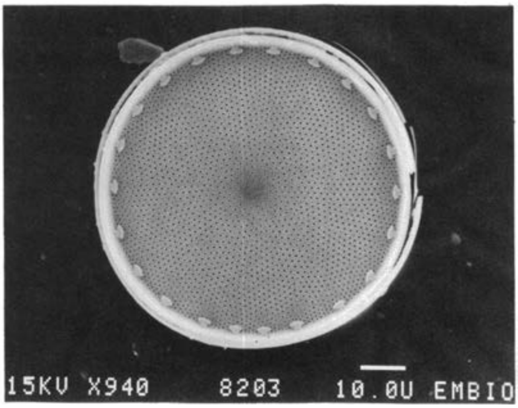
Environmental changes inferred from the diatom assemblages

Zone A has a diatom assemblage which can be characterized as reworked. This is in accordance with the other microfossil investigations from this core, where a conspicuous amount of redeposited material of especially coccoliths and pollen has been found (Mikkelsen, Henningsmoen & Höeg, all this volume). But, contrary to the results from coccolith and pollen analysis, where reworked material is found throughout the entire lower unit (1050–700 cm), the diatoms mainly show reworked (Tertiary) material in the lower 1 m of the core. Above 950 cm only 1 % of the diatom assemblage consists of reworked species.

Zone B is characterized by polar planktonic species. The diatom assemblage is similar to the one found in the Bohuslän cores from western Sweden (Miller 1982). The assemblage of zone B is interpreted as characterizing a situation when the Norwegian Channel was a polar fjord with a steep rise to a land area to the south and to the icefront to the north. The eastern border of the fjord sloped more gently, and the Bohuslän assemblage is believed to have been deposited at a shallow water depth at a distance of about 70–100 km from the icefront. The Bohuslän flora has an influx of freshwater plankton while no freshwater species were found in core GIK 15530-4. It is interesting to note the similarity in the two assemblages despite the large differences in wa-



A



B

Fig. 2. Scanning electron microscope photograph of *Actinocyclus* cf. *alienus* var. *arctica* ? A: outside view, B: inside view.

ter depth. The water depth when zone B was deposited is estimated to have been about 275–300 m at the site of core GIK 15530-4.

Zone C has a totally different flora from the one found in zone B. There is a marked increase in diatom frequency, mainly caused by a dominance of *Paralia sulcata*. The assemblage is characterized as coastal and temperate. The information on ecological requirements of *Paralia sulcata* varies with the different authors (Miller 1964). This is caused by it being tychoplanktonic (von Stosch 1956), a bottom dwelling form (benthic) which in turbulent water is brought up into the water mass and therefore is commonly found among the coastal plankton. It should be noted that *P. sulcata* is characterized as a coastal, temperate planktonic species by Miller (1982). Miller interprets the change from her Arctic to coastal temperate plankton to be caused by the rapid decrease of water depth at the end of the Younger Dryas Interstadial. The large occurrence and total dominance of *P. sulcata* in the Norwegian Channel is more probably caused by a change in the circulation pattern, while the more general change from a polar to a temperate flora was caused by a climatic amelioration triggering the northward movement of the Subarctic Convergence (Polar Front). The change in circulation pattern may have resulted in a supply of allochthonous diatom valves brought in from the shallow areas south of the coring site. This would explain the change to a more benthic assemblage than the one found in zone B. The assemblage found in zone C resembles the flora found in the surface sediments in the area today (Stabell, unpubl.). The increased diatom frequency may be explained by an increased productivity in the shallow areas (which were dry land during the deposition of the zone B assemblage) or by the sudden decrease in accumulation rates (Thiede, this volume). Diatom zone C may therefore be regarded as a transitional zone between zone B (when the Norwegian Channel was a polar fjord) and zone C (with a stable situation, resembling the present). The situation in the Norwegian Channel, with no important water depth change, would be very different from the one at the Swedish west coast, where great changes in water depth occurred at this time. The end of the Younger Dryas is also marked by a large meltwater discharge (drainage of the Baltic Ice Lake). This is clearly reflected in the cores from Bohuslän. The fresh water was probably forced north and westwards along the Norwegian coast and would therefore not affect the conditions at the core site (south side of Norwegian Channel).

Zone D, which can be characterized as mainly barren, demonstrates a stable situation with mainly oceanic water undersaturated in silica.

Zone E shows the present situation with a coastal, temperate assemblage dominated by *P. sulcata*. It should be noted that the fragile, thin valved planktonic species dissolve easily and are normally not found in the sediments. The assemblage in Zone E does not therefore represent the present diatom flora in the water mass at the site.

Comparison with other investigations of the core

From the above discussion it is evident that a clear change from a polar planktonic to a transitional (littoral, temperate) diatom assemblage occurs between 650 and 700 cm. This is in good accordance with the other microfossil investigations. The deposition of diatom zone A and B is unanimously interpreted as occurring in a glacially influenced cold environment. The dating of the core (cf. Henningsmoen & Höeg, Schönharting, Stabell, all this volume) places diatom zones A and B in (late Alleröd (?) and) Younger Dryas chronozones. The deposition of diatom zone C is just as unanimously characterized as temperate and, by the ostracod and radiolarian investigations (Qvale, Bjørklund, both this volume), also as transitional. Diatom zone C was deposited in early Preboreal; the boundary between diatom zone B and C thereby coincides with the Pleistocene/Holocene boundary. Diatom zones D and E are almost void of diatoms and are therefore not suited for comparison. The two zones cover the main part of the Holocene.

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Dinoflagellate cyst analysis of Upper Quaternary sediments in core GIK 15530-4 from the Skagerrak

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Dinoflagellate cyst analysis (cysts/g dry weight of sediment, and percentage composition of the assemblages (suggests 3 distinct assemblage zones. Zone 3 (7–10.7 m) with a low diversity assemblage dominated by *Protoperidinium* cysts is believed to contain Younger Dryas sediments representing a cold water, ice dominated environment. Zone 2 (5.5–7 m) with a dramatic shift to an *Operculodinium centrocarpum* dominated assemblage and a peak of *Bitectatodinium tepikiense* is believed to represent a Preboreal intrusion of warming Atlantic Ocean waters. Zone 1 (5.5–0 m) with a similar assemblage to the present-day (dominant *O. centrocarpum* and higher percentages of *Spiniferites spp* and *Peridinium faeroense* than that previously) represents development of conditions similar to those of the southern and western Norwegian coast today.

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Dinoflagellate cysts (referred to hereafter simply as cysts, for brevity) are potentially very useful microfossils for Quaternary biostratigraphy and paleoenvironmental interpretation. Dinoflagellates are a major group of phytoplankton found in almost all aquatic environments today, and their resting cysts have already proved to be important microfossils, particularly in Mesozoic and Tertiary biostratigraphy. Calcareous and siliceous forms are known, but most cysts are palynomorphs composed of sporopollenin-like material, and they are therefore not subject to dissolution problems sometimes affecting other microfossils such as foraminifers, coccoliths and diatoms.

The use of Quaternary cysts may be compared in many ways with pollen analysis. However, the potential for using cysts as 'marine pollen' is still not fully developed, though Rossignol (1962) first explored this possibility over 20 years ago. The main reason for this, as Reid (1982) pointed out, is insufficient understanding of the ecological distribution of living cysts today. Biologists interested in dinoflagellates have almost exclusively studied the non-fossilizable motile stages commonly found in phytoplankton samples, and until recently they were largely unaware of the morphologically different encysted stage in the life cycle. The cysts were 'discovered' by paleontologists looking for living equivalents for the fossils (Evitt & Davidson 1964, Wall 1965, Wall & Dale 1966, 1967). Many cysts have not yet been correlated with their corresponding motile stages and therefore little of the ecological information produced by biologists is directly applicable to Quaternary dinoflagellate paleontology.

Living cysts have been studied from several regions, but ironically Quaternary cyst studies so far have usually been attempted from other regions with little or no direct back-up information concerning living cysts. For example, living cyst information is almost exclusively from just a few coastal regions (e.g. USA East Coast, Wall & Dale 1968; Bermuda and Puerto Rico, Wall & Dale 1970) while Quaternary cyst studies are mostly from outer neritic or oceanic regions (e.g. N.E. Atlantic Ocean, Turon 1980; North Sea, Holmes 1977; Red Sea, Wall & Warren 1969; off West Africa, Rossignol-Strick & Duzer 1979). As an alternative to living cyst information, several investigations suggest that cyst distribution patterns in Recent bottom sediments accurately reflect overlying water masses and thus may provide useful background information for interpreting Quaternary paleoenvironments (Reid 1972, Reid & Harland 1977, Wall et al. 1977, Williams 1971). However, while this certainly appears true for coastal regions, cyst distribution patterns described so far for Recent sediments in the deeper ocean probably reflect mainly large scale transport of the fine silt component of oceanic sediments rather than plankton from overlying waters (Dale, in press).

Obviously, for cysts to be fully utilized as 'ma-

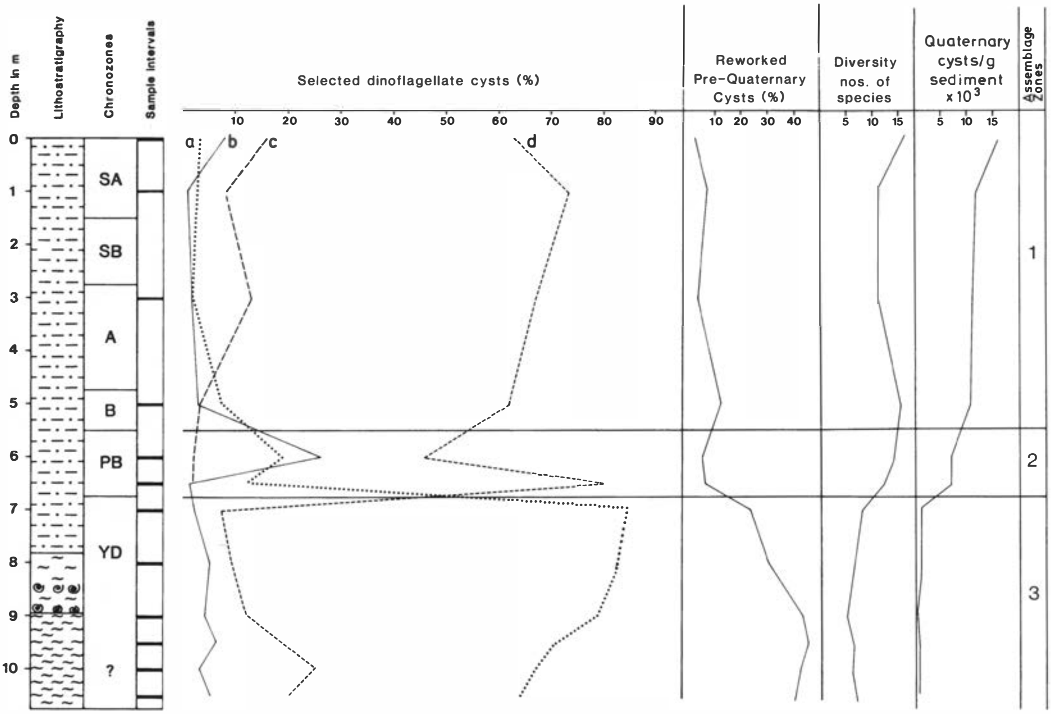

Fig. 1. Distribution of selected dinoflatellate cysts, reworked pre-Quaternary cysts, species diversity, cysts/g dry weight of sediment, and assemblage zones. a = spherical brown *Protoperidinium* cysts, b = *Bitectatodinium tepikiense,* c = *Peridinium faeroense* and d = *Operculodinium centrocarpum.*

rine pollen' in Quaternary studies will necessitate developing relevant ecological information comparable with present-day plant biogeography used by pollen analysts, preferably starting with detailed regional studies. The work reported here is part of one such study: a detailed investigation of living, Recent and Quaternary cysts from Norway and the surrounding region. The main aim is to document the biogeographical distribution of living and Recent cysts from the region as background information for interpreting Quaternary sequences. So far the author has studied living cysts from several areas spanning the whole Norwegian coast from Oslofjord to Varangerfjord (Dale 1983, and in prep.), Recent cyst distribution for several areas including the Faeroe Islands and the Baltic (Dale 1976, and in prep.), and Quaternary sequences from Fjøsanger, near Bergen, and from a number of boreholes in the North Sea (Dale, in prep.). Two student theses supervised by the author provide detailed information of Recent cyst distribution in bottom sediments in the Oslofjord/Skagerrak area (Bakken 1983, and Konieczny 1983), and Harland (1982) has documented cysts from several Quaternary boreholes from S.W. Sweden. Also relevant for the region as a whole are studies of Recent cyst distribution in coastal sediments round the British Isles (Reid 1972) and in shelf sediments from the British Isles (Harland 1977).

Materials and methods

12 samples from core GIK 15530-4 were examined for cysts (see Fig. 1 for sample details). After oven-drying at 40°C and weighing, samples were prepared using standard palynological methods involving HCl, HF and wet sieving to give >25 μm fraction of the remaining organic residue. Microscope slides were prepared representing a known fraction of this, allowing both quantitative analysis (cysts/g dry sediment) and qualitative analysis (% composition) of the cyst assemblages counted.

Total cyst counts included in Table 1 show that numbers of cysts which could be counted within a

TABLE 1. Composition of cyst assemblages (in percent) from Core GIK 15530-4, Skagerrak (+=<1%)

Zones	*Cyst species / Samples (Core depth cm)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	Cysts counted
1	0	3			3		+		1	+	16	+	63	8	5		+		+		+		1	1	+	+	+		413
	100	3			3		+		+		8	+	73	1	8				1				1	2			1		153
	300	2			2				+		13	+	67	2	10				2		1			1	1	1		1	173
	500	7	+	+	7				+		4		62	3	8	1	+		2		1	2	2	5	1	+	+	+	312
2	600	17	2	+	19		+	+			2		46	26	1	+		+	1		+	+	+	1	+	1		1	515
	650	11	1	+	12		+	1			2		80	1	1	+			+						+	+	+	1	1003
3	700	79	3	2	84								7	2				+	2	+				1				5	191
	800	74	6	1	81								9	5				+		+				1				4	364
	900	65	9		74								12	4						3				1				7	104
	950	63	6		69								19	6					1	1				1				3	144
	1000	60	5		65	1							24	3					1	1				2				4	254
	1050	61	1		62	3					1		20	5					2	3				1				3	164

* 1. Unidentified spherical brown Protoperidinium cysts. 2. Protoperidinium conicoides. 3. Protoperidinium denticulatum. 4. Total spherical brown Protoperidinium cysts. 5. Diplopeltopsis minor. 6. Protoperidinium conicum. 7. Protoperidinium compressum. 8. Protoperidinium pentagonum. 9. Cyst sp A. 10. Peridinium faeroense. 11. Lingulodinium machaerophorum. 12. Operculodinium centrocarpum. 13. Bitectatodinium tepikiense. 14. Spiniferites bulloideus. 15. Spiniferites cf S. bulloideus. 16. Spiniferites belerius. 17. Spiniferites delicatus. 18. Spiniferites elongatus. 19. Spiniferites ramosus. 20. Spiniferites lazus. 21. Spiniferites membranaceus. 22. Spiniferites mirabilis. 23. Spiniferites spp indet. 24. Nematosphaeropsis labyrinthus. 25. Planinosphaeridium choanum. 26. Polykrikos schwartzii. 27. ? Multispinula minuta.

reasonable time varied according to sample productivity (the lower 6 samples were particularly low) and amounts of background organic material encountered. Many counts fall short of the 400 minimum considered ideal for palynological assemblages, but the major trends in data presented here agree with those found elsewhere for the region and would not be expected to change appreciably if counts were increased to the ideal level.

Results

All 12 samples yielded usable cyst assemblages. The cysts recovered were either reworked pre-Quaternary forms or Quaternary forms, almost all of which also live in the general region today. Pre-Quaternary cysts identified were Cretaceous or Tertiary forms. They are expressed as percentages of the total cyst count (see Fig. 1). All other cyst data refer only to Quaternary forms. 24 Quaternary cyst species were identified, and percentage abundances of these are shown in Table 1. Changes in percentage abundances, absolute abundances and number of species recorded allowed recognition of three main assemblage zones within the core, numbered 1 to 3 from the top (see Fig. 1).

Zone 3

This zone is represented by the lower 6 samples studied (1050-700 cm). It is characterized by relatively higher percentages of pre-Quaternary cysts (up to 45 %), low amounts of Quaternary cysts (mostly between 1000–2000/g sediment) and relatively low numbers of Quaternary cyst species (5–8). The cyst assemblage is heavily dominated by spherical brown protoperdinioid cysts (probably including many *P. conicoides*) that gradually increase their percentage abundances from 62 % in the bottom sample to 84 % in the upper sample. Other numerically important cysts include *Operculodinium centrocarpum* (7–24 %),? *Multispinula minuta* and *Bitectatodinium tepikiense* (2–6 %). Two cyst types were restricted to zone 3, cysts of cf. *Diplopeltopsis minor* and of cf. *Spiniferites ramosus,* but neither is previously known and their significance therefore remains unclear.

Zone 2

This is represented by samples at 600 cm and 650 cm. It is characterized by relatively low percentages of pre-Quaternary cysts (5–6 %), moderate amounts of Quaternary cysts (6611/g sediment at 650 cm and 6473/g sediment at 600 cm) and increased numbers of Quaternary cyst species recovered (12 and 14). The cyst assemblages are heavily dominated by *O. centrocarpum* (80% and 46%), with important percentages of *Protoperidinium* cysts (12% and 19%) and a significant peak of *B. tepikiense* (up to 26%). ? *Multispinula minuta* remains present, but at lower percentage values (1%). Otherwise, many forms which persist up through the remainder of the core are first recovered in zone 2 (including the outer neritic form *Nematosphaeropsis labyrinthus*). Cysts of *Protoperidinium compressum* were only recorded from zone 2 (1% and <1%).

Zone 1

This is represented by the upper 4 samples studied down to 500 cm. It is characterized by relatively low percentages of pre-Quaternary cysts (2–12%), relatively high concentrations of Quaternary cysts (increasing steadily from 10058/g sediment at 500 cm to 11642/g sediment at 100 cm with a more rapid increase to almost 16000/g sediment in the top sample) and similar numbers of Quaternary cyst species recorded to those of zone 2 (11–16). The cyst assemblage is heavily dominated by *O. centrocarpum* (62–73%), with important percentages of total *Spiniferites* cysts (around 10–20%) and *Peridinium faeroense* cysts (4–16%). ? *Multispinula minuta* was not recorded from the upper two samples.

Paleoenvironmental interpretations

Zone 3

This assemblage is considered typical for colder water, ice dominated marine environments of the region; the interglacial sequence at Fjøsanger is bracketed by two such zones representing glaciomarine conditions before and after the main climatic warming event, and sediment from this zone in the Skagerrak core contains significant amounts of ice-rafted material and is therefore also considered glaciomarine. So far, living cysts have not been studied from comparable environments near permanent sea ice. However, one of the main components of the *Protoperidinium* dominated assemblage described here is probably *P. conicoides,* which was first described in plankton collected in between floating ice in the Denmark Straight off the east coast of Greenland (the distinguishing features for identifying spherical *Protoperidinium* cysts are often obscured, but when they could be identified in zone 3 they were usually *P. conicoides*). Dale (1983, p. 126) has suggested that such *Protoperidinium* species are non-photosynthetic and may therefore dominate cyst assemblages such as those in zone 3 by their independence from the unstable light regime offered by ice dominated waters. This may also explain the more dramatic shift in cyst assemblages than has been noted for other microfossils between samples at 650 cm and 700 cm. Cysts may reflect a rapid but fundamental environmental shift from the difficult light conditions for swimming phytoplankton posed by dominant sea ice as this melted, while many other groups presumably reflected primarily a more gradual shift in sea water temperature. Consistent presence of ?*Multispinula minuta* throughout this zone (including a maximum of 7% as in the lowermost glaciomarine samples at Fjøsanger) confirms this as an important cold water indicator species.

The low number of cysts/g sediment in zone 3 is interpreted as reflecting two main possibilities. On the one hand, overall dinoflagellate productivity (at least for the many photosynthetic types) was probably much lower than during subsequent times, due to the unstable light regime discussed above. In addition, larger amounts of sand and coarser material in the sediments of zone 3 would also serve to 'dilute' the cysts. The relatively high percentage of reworked pre-Quaternary cysts also probably reflects increased amounts of coarser sediment (some of which are probably fragments of older sediments containing palynomorphs) together with drop stones and the general increased transport of pre-Quaternary 'debris' associated with melting glaciers. But lower production of Quaternary cysts would also in effect increase the percentage of pre-Quaternary forms.

Lower numbers of cyst species recorded from zone 3 are interpreted as reflecting genuine lower diversity typical for adverse environmental conditions. 'Adverse' conditions in this context probably refer mainly to difficult light conditions which would presumably affect many photosynthetic dinoflagellates represented in zone 2 and 3 (as discussed previously), but changing salinity probably also contributed. Certainly both subsidiary cyst species typical for zone 3 tolerate large reductions in salinity from normal marine; *O. centrocarpum* is probably the most cosmopolitan cyst type known and for example is reported from cool temperate, brackish stratified estuaries as is also *B. tepikiense* (Wall et al. 1977).

Zone 2

Both the shift from a *Protoperidinium* dominated assemblage to an *O. centrocarpum* dominated assemblage and an associated peak of *B. tepikiense* are considered typical in this region for mixing of colder polar front waters with warmer Atlantic waters. For example, in the Fjøsanger sequence the transition from the lower glacial event into the interglacial is accompanied by just such assemblage shifts, which then are reversed going from the interglacial into the next glacial event. The extent of the *B. tepikiense* peak in core GIK 15530-4 (26%) is not as large as those at Fjøsanger (33% and 44%) or some North Sea sequences (up to more than 90%, Harland 1977). This may possibly reflect less mixing of the two water masses in the more restricted Skagerrak or simply that the relatively few samples examined here did not reveal the full peak.

Zone 2 is interpreted as representing a transition from the colder, ice dominated water of zone 1 towards the more temperate conditions developed to the present day. This was almost certainly caused by penetration of warmer Atlantic water into the area in the time interval represented between samples at 700 cm and 650 cm in core GIK 15530-4. First records in the core, already at 650 cm, of typical cold temperate forms (e.g. *P. faeroense* and *Planinosphaeridium choanum*), which then persisted through to the present day, testify to the establishment of significantly warmer waters, while the presence of *N. labyrinthus* strongly suggests Atlantic influence. *N. labyrinthus* is a form typical for outer shelf to oceanic environments (Wall et al. 1977, Reid & Harland 1977) and its distribution in Recent sediments of the Skagerrak corresponds to penetration of denser Atlantic waters into the deeper basin today (R. Konieczny, pers, comm.).

The two cyst assemblages of zone 2 probably represent different phases of transition. The assemblage at 650 cm shows a 'spike' of *O. centrocarpum* (80%). Similar spikes have been observed elsewhere (e.g. in the transition from glaciomarine to interglacial at Fjøsanger), and they probably reflect an extremely cosmopolitan species with broad tolerance for wide ranges of temperature and salinity exploiting the rapidly shifting environment caused by the convergence of two very different water masses. The assemblage at 600 cm with its peak of *B. tepikiense* seems to represent a different phase, the significance of which is as yet unclear. Interestingly, in the Fjøsanger sequence the *B. tepikiense* peak immediately precedes the *O. centrocarpum* spike, whereas in core GIK 15530-4 the *B. tepikiense* peak follows that of *O. centrocarpum*.

Records of *P. compressum* only in zone 2 are interesting, since Harland (1982) also recorded this species from fairly restricted horizons in several cores from the Swedish side of the Skagerrak. However, occasional records of this species from the general region today as yet offer no particular explanations for this.

Zone 1

Assemblages from zone 1 are generally comparable with present-day assemblages from coastal waters of southern and western Norway and probably reflect environmental conditions similar to those of today throughout the period covered. However, two trends in assemblages from zone 1 probably reflect development of these conditions. The fact that *Lingulodinium machaerophorum* was only recorded from the upper 3 samples studied probably reflects warmer summer waters during this interval than previously in the core. *L. machaerophorum* is a cosmopolitan species with a present-day distribution along the Norwegian coast restricted to southern and western Norway. It is typically found today in the innermost branches of fjords in warmest summer waters with less than normal marine salinities. Similarly, *P. faeroense* is typically found today in sheltered bays and fjords but seems to tolerate lower temperatures than *L. marchaerophorum*. It is presently distributed along the whole Norwegian Coast and occurs in early spring plankton in the Oslofjord (Dale 1977). Overall percentage increases of *P. faeroense* from zone 3 to the present day probably reflect establishment of shallower, more sheltered bays and their associated plankton along the fjord margin. Both *P. faeroense* and *L. machaerophorum* are most likely transported out into the Skagerrak from nearer shore. Recorded absence of the cold water species ?*Multispinula minuta* in the upper two samples probably also reflects warming conditions.

Comparison with previous studies in the region

The main trends in cyst assemblages described here from core GIK 15530-4 agree closely with other Quaternary sequences from the Norwegian Coast (e.g. Fjøsanger; Dale, in prep.). They offer a model for cyst-based paleoenvironmental interpretations at least for coastal regions of North temperate regions. Unfortunately, the only other Quaternary sequences studied so far from the Skagerrak area (Harland 1982) yielded such poor assemblages that comparison with the present work is very limited. While similarities can be found to the three main assemblages described here, no clear paleoenvironmental trends are resolved.

Comparison between the coastal north-temperate paleoenvironmental trends proposed here with North Sea Quaternary sequences suggests both interesting similarities and differences. For North Sea boreholes, Harland (e.g. in Holmes 1977) and Dale (in prep.) have recovered similar cyst assemblage trends to those described here. However, whereas *B. tepikiense* only occurs as relatively sharp peaks in 'transition zones' in coastal areas, it sometimes *dominates* large parts of cold sequences in the North Sea. According to Turon (1980), this species also dominates Recent assemblages north of 60°N in the Norwegian Sea, together with very common *Protoperidinium* cysts, which is consistent with paleoenvironmental interpretations presented here. The little available evidence so far suggests that *B. tepikiense* may be a more important indicator of colder climatic conditions in more offshore sequences, while *Protoperidinium* spp. are more important in coastal regions such as the Skagerrak.

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Upper Weichselian – Holocene radiolarian stratigraphy in the Skagerrak (NE North Sea)

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Three radiolarian assemblage zones were defined in Upper Weichselian-Holocene sediments from the Skagerrak (NE North Sea), the *Amphimelissa setosa* Assemblage Zone, the *Rhizoplegma boreale* Assemblage Zone and the *Actinomma leptodermum- Phorticium clevei* Assemblage Zone. The three assemblage zones represent a hydrographical evolution from arctic, via transitional to modern conditions respectively. The biogenic opal is generally poorly preserved in the upper part of the core, improving downcore. A maximum of volcanic ash shards was found at 895–898 cm, representing middle Younger Dryas ca. 10,600 years B.P., and can be correlated throughout the NE North Sea.

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Plankton investigations from the epicontinental North Sea complex were carried out during the years 1902–1908 and resulted in a special report on the radiolarians by Mielck (1913). 31 species of polycystine radiolarians (14 Spumellaria and 17 Nassellaria) were reported, the highest abundance of species and specimens being found in the Norwegian Channel and Skagerrak. Mielck (1913) also concluded that radiolarians occupied the surface waters during the winter season, while descending to deeper water during the summer season. Even if most of the species are in low quantities, they occur throughout the year. From this observation Mielck (1913) concluded that radiolarians lived and reproduced in the area, and were not a drift fauna originating from the North Atlantic Current.

The author has investigated the uppermost parts of several cores from the North Sea, Norwegian Channel and Skagerrak with respect to radiolarians. Generally, North Sea sediments are barren or have only traces of radiolarians, while in the Norwegian Channel and the Skagerrak radiolarians were found in modest numbers (less than 1000 radiolarians per 1 g dry bulk sediment).

This report is the first attempt to utilize radiolarians for biostratigraphical purposes in this distal position of the epicontinental North Sea complex.

Material and methods

The slide preparation for radiolarians follows the procedure described by Goll & Bjørklund (1974). Due to the rather rare occurrence of radiolarians in this core, slides for quantitative estimates of radiolarians downcore were not made.

Results

Due to the rather low concentrations of radiolarians only fauna slides were made from the sediment samples. The distribution and relative occurrence in per cent should therefore be taken as qualitative estimates rather than absolute values (Fig. 1).

Despite the low number of radiolarians, three different assemblages could be recognized from the core as follows:

0–600 cm: *Actinomma leptodermum – Phorticium clevei* Assemblage Zone (1)

The base of this zone is defined by the sudden decrease in *P. clevei,* the first occurrence of *A. boreale* and *A. leptodermum* and a marked increase in *Rhizoplegma boreale.* The 300 cm level in this core is free of biogenic opal, while the interval 300–600 cm generally can be de-

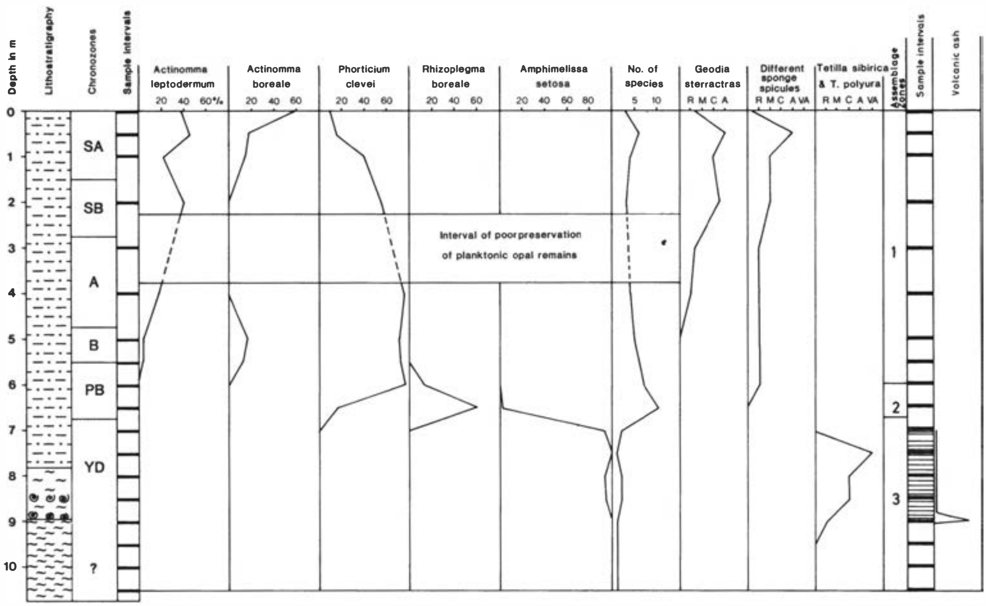

Fig. 1. Relative occurrence of radiolarians and sponges in core GIK 15530-4, with proposed biozonation.

scribed as a level of poor opal preservation. This assemblage zone has a low species diversity, but it indicates warm water conditions.

600–675 cm: *Rhizoplegma boreale* Assemblage Zone (2)

The top of this zone is coincident with the base of the *A. leptodermum – P. clevei* Assemblage Zone. The base is defined by the first occurrence of *R. boreale* together with a sudden decrease in *Amphimelissa setosa.* This assemblage zone is characterized by its high species diversity, and is of a transitional nature, as both warm- and cold-water species are present.

675–1050 cm: *Amphimelisssa setosa* Assemblage Zone (3)

The top of this zone is coincident with the base of the *R. boreale* assemblage zone, while the base is not seen in this core. It is an assemblage dominated by *A. setosa,* indicating cold water.

The general evolution of the species assemblages and the three assemblages which have been recognized can be seen in Fig. 1.

The *Amphimelissa setosa* Assemblage Zone

A. setosa has been reported in plankton samples only from cold waters and does not belong to the present assemblage along the west coast of Norway (Jørgensen 1905). *A. setosa* comprises up to 75 % of the total fauna in surface sediment samples from the Iceland Plateau, ca. 30 % in the Norway Basin and 1 % at the continental slope of Møre at a depth of 786 m (V 27–93). This is taken as evidence for its cold-water affinity. *A. setosa* is interpreted to occupy the cold water which underlies the warm Norwegian Current. This is supported by the work of Bernstein (1934), who reported *A. setosa* only from the deeper part of the Kara Sea, in water below the warmer surface water of Atlantic origin.

The environment during the deposition of the *A. setosa* assemblage zone is interpreted to represent cold Arctic water. This is also supported by the abundance of sponge spicules between

750–900 cm, belonging to the species *Tetilla polyura* (Koltun 1966, Fig. 31,5) and *Tetilla sibirica* (Koltun 1966, Fig. 33,6). Their Holocene distribution is restricted to the Barents, Kara, Laptev and Greenland seas.

On the radiolarian slides volcanic ash was observed so frequently as to suggest that there must be an ash horizon somewhere in the core. Particle countings were carried out on the 44–125 μm fraction for every 10 cm between 700 cm and 900 cm. Ash was found throughout the interval with about 1 % of the particle assemblage, but samples 885–887.5 cm and 895–898 cm contained 7.3 % and 14.3 % ash respectively. Ash has not been observed deeper in the core. Mangerud et al. (1984) described a volcanic ash layer from several localities at Sunnmøre. It has been dated with ^{14}C to have been deposited between 10,300 and 10,600 years B.P., which is a good marker for the middle Younger Dryas. The same ash layer has been found in Sognesjøen (Seland 1981) and in the Norwegian Channel (Godvik 1981). It is therefore assumed that the ash layer found in this core represents the same ash layer as reported further north in the North Sea complex. According to radiocarbon dating in this core on bivalve fragments from the core interval 850–885 cm, an age of 10,260 ± 280 years B.P. was obtained (Stabell, this volume). An age of the volcanic ash peak of about 10,600 years B.P. at 895–898 cm is therefore in good agreement with the radiocarbon age.

The *Rhizoplegma boreale* Assemblage Zone

The deposition of the *Rhizoplegma boreale* assemblage reflects a transitional assemblage with both cold- and warm-water species. It parallels a peak of benthic diatoms and there is a high diversity of well-preserved biogenic opaline components.

This transitional zone is of interest as the zone as defined by radiolarians is clearly indicated by the peak occurrence of *R. boreale*. The same zone can be recognized from Korsfjorden (Aarseth et al. 1975), where the frequency of *R. boreale* also reaches a peak. The associated diatom peak is, however, composed of different assemblages. This indicates different hydrographical conditions in the Skagerrak and in the west Norwegian fjords.

The *Actinomma leptodermum-Phorticium clevei* Assemblage Zone

The *Actinomma leptodermum – Phorticium clevei* assemblage zone is difficult to interpret as the opal preservation is poor and the number of radiolarians is low.

Mielck (1913) reported that *A. leptodermum* may occur with 8,000 specimens for every 1 m^2 of surface waters. This species occurs associated with *P. clevei* and *A. boreale* in this assemblage zone. Mielck (1913) reported 31 polycystine radiolarians in the plankton, but only 8 species (6 Spumellaria, 2 Nassellaria) were separated from the sediment in this zone. All members have a dense test, indicating that selective dissolution may be responsible for the rare occurrence of radiolarians. The tests which are present are heavily abraded, also indicating that hydrodynamic separation by bottom (?) currents might have been active during the deposition of this assemblage. Supporting this is the rather frequent occurrence of the *sterractra* spicule type, diagnostic for the sponge genus *Geodia.* These spicules do not show any sign of mechanical breakage or chemical dissolution. As the *Geodia* sterractras are solid opal spheres roughly 80 μm in diameter, they are believed to have been transported to the depositional site. Koltun (1966) regards the *Geodia* as hard-bottom indicators. The *Geodia* sterractras are abundant to common from 50 cm to 300 cm, rare at 300–400 cm, but absent below this depth (Fig. 1).

A similar distribution is observed for the foraminifers in the 63–125 μm size fraction. Both planktonic and benthic foraminifers have their highest frequency in the upper 300 cm.

A further suggestion of an increase in bottom (?) currents during deposition of the upper 300 cm of the core, is a sudden change in the occurrence of plant fragments. These are abundant from 400 cm to the bottom of the core, while they are absent in the upper 300 cm. This indicates a change in the environment, either an increased current activity or a sudden stop in the discharge of plant remains from land.

Conclusions and summary

1) The whole core represents open and fully marine conditions.
2) Three radiolarian assemblage zones could be identified, each representing characteristic

environmental stages in the hydrographical evolution in the upper Quaternary North Sea.

3) Upper Weichselian (equiv. to 1050–675 cm). The radiolarians and the sponge fauna clearly indicate cold water conditions during the sedimentation of this interval. A middle Younger Dryas peak of volcanic ash shards is present in this zone at 895–898 cm, representing ca. 10,600 years B.P.
4) Early Holocene (equiv. to 675–600 cm). This interval represents a transitional assemblage.
5) Middle-Late Holocene (600–0 cm). This interval is characterized by its warm water fauna. Grain-size data in the 63–125 μm fraction suggest that some changes in the hydrography and the sedimentological regime took place between 400 and 300 cm. Small planktonic foraminifers occur above 400 cm, lacking below, while there is a marked increase in the small benthonic foraminifers – up to 60 % at 300 cm and less than 20 % at 400 cm. On the radiolarian slides there is a clear change in the presence of plant remains in the fraction larger than 44 μm, as plant remains are practically not present above 400 cm. These observations lead to the conclusion that the present day hydrographical situation was established between 400 and 300 cm, representing late Atlantic.

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Benthic foraminifers in Upper Quaternary Skagerrak deposits

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Benthic foraminifers occur abundantly in all samples. The core can be divided into two distinct zones; the lower part (from base to 7.0 m) is characterized by a cold water assemblage, the upper 6.5 m by temperate (boreal) species assemblages.

Cassidulina reniforme, Elphidium excavatum, Nonion labradoricum and *Islandiella* spp. are the main constituents of the cold water assemblage. The upper part of the core can be further subdivided into three assemblage zones (AZ) named after its dominant species: 6.5–4 m, *Cassidulina laevigata* AZ; 3.5–0.5 m, *Hyalinea balthica* AZ; and 0.4–0 m, *Bolivina* cf. *B. robusta* AZ.

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Benthic foraminifers are widely used as a stratigraphical tool and are among the best indicators of environmental parameters. Although our knowledge about their ecology is still very limited, it has been recognized for more than a decade that distributions of benthic foraminifers are controlled directly or indirectly by water mass properties.

Previous studies of foraminifers in Upper Quaternary (Holocene) deposits in the Skagerrak have been carried out by Lange (1956), Kihle (1971) and Jørgensen et al. (1981).

Methods

For the study of benthic foraminifers 21 samples of core GIK 15530-4 have been used. In addition to the 18 'standard' samples we included three samples (0.35–0.40 m, 3.50–3.55 m and 4.50–4.55 m) to improve the zonation. The samples were dried at 70°C and then weighed. The bulk dry weight varied between 15 g and 75 g. After disintegration in hot water the samples were wet sieved through 1 mm, 125 μm and 63 μm sieves and then dried. For the foraminiferal analysis we used the fraction > 125 μm. The fraction > 1 mm contained no foraminifers. When counting the foraminifers, we always used a known split of each sample.

All benthic foraminifers observed were identified, but only the most important species are mentioned in the text and illustrations. The relative abundances of the most important species are shown in Fig. 1.

The faunal diversities were calculated in accordance with Walton (1964) (Fig. 2), while for the Fisher α-indices we used the graphic method described by Murray (1973) (Fig. 3).

For synonymy and information on the geographical distribution of the species discussed, the reader is referred to Feyling-Hanssen et al. (1971). Exceptions are *Cassidulina reniforme, Islandiella helenae* and *Melonis barleeanum*, for which we refer to Sejrup & Guilbault (1980), Feyling-Hanssen & Buzas (1976) and Corliss (1979), respectively.

Main faunal features

Foraminifers occur abundantly in all samples studied, though being less numerous in the lower part of the core. The number of specimens per g dry bulk sediment varies between 158 in the uppermost to 2 in the lowermost part of the core (Fig. 2).

Benthic foraminifers predominate throughout the core. Planktonic species are rare, and never exceed 1.5 % of the total foraminiferal assemblage in the fraction studied, while in the smaller fraction (63–125 μm) they are quite abundant (Thiede, this volume).

The faunal diversity (Fig. 2) is low in the lowermost part of the core. Maxima occur between 4 and 6 m and around 1 m.

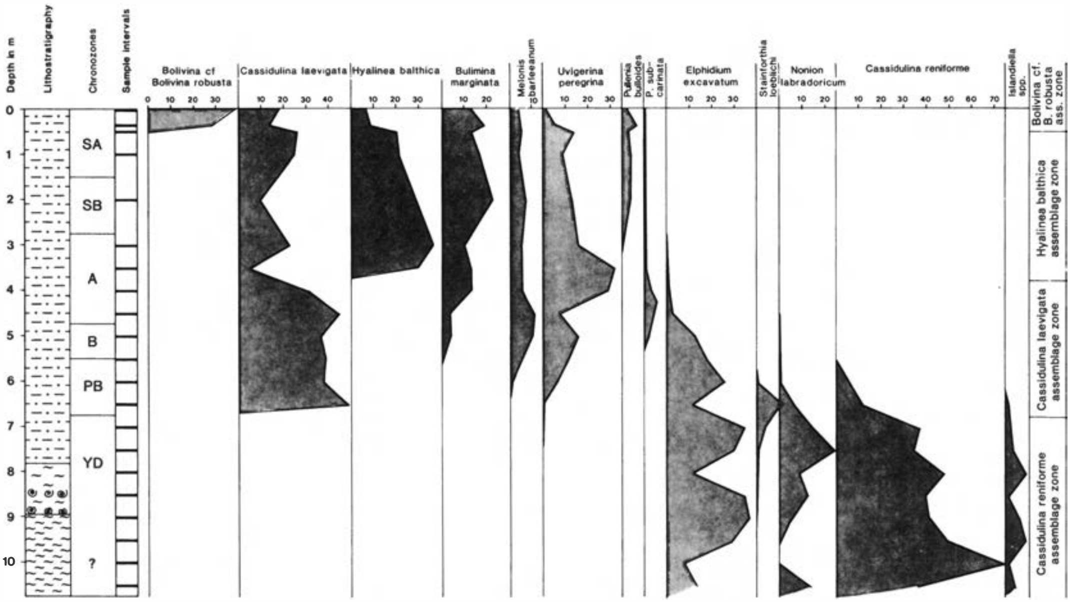


Fig. 1. Relative abundances (in % of the total benthic foraminiferal assemblage) of the most significant species. The benthic foraminiferal assemblage zones established are indicated on the right side of the diagram.

The most striking feature in the foraminiferal zonation in core GIK 15530-4 is the rather abrupt change from a cold water assemblage in the lower 3.5 m of the core to a more temperate fauna above 7.0 m. This broad faunal trend is illustrated in Fig. 3, which clearly shows the distribution of two main faunal components. The cold water fauna includes *Astrononion gallowayi, Cassidulina reniforme, Elphidium excavatum, Islandiella helenae, I. norcrossi, Nonion labradoricum, Pyrgo williamsoni, Stainforthia loeblichi* and *S. schreibersiana.*

The temperate fauna includes *Bolivina* cf. *B. robusta, Bulimina marginata, Cassidulina laevigata, Eggerella scabra, Globobulimina turgida, Hyalinea balthica, Melonis barleeanum, Nonionella turgida, Pullenia bulloides, P. subcarinata, Trifarina angulosa* and *Uvigerina peregrina.* A large number of species which are represented by only one or two specimens cannot be differentiated.

The foraminiferal assemblages occurring in the core samples between 4.5 and 6.5 m included both cold and temperate elements (Fig. 2) and may be characterized as transitional.

Benthic foraminiferal assemblage zones

The core can be divided into four different zones, each designated by its dominant species (Fig. 1).

The *Cassidulina reniforme* Assemblage Zone

The lower part of the core (below 6.5 m) is characterized by a cold water assemblage (see above) dominated by *Cassidulina reniforme, Elphidium excavatum, Nonion labradoricum* and *Islandiella* spp. *Stainforthia loeblichi* is also found but only in small numbers. The faunal diversity is low (Fig. 2) and Fisher α-indices are generally below 3 (Fig. 3). The assemblages are very similar to those occurring near glacier fronts today, e.g. in the inner part of Kongsfjorden in Spitsbergen (Elverhøi et al. 1980). Similar assemblages have also been described by Feyling-Hanssen (1980, 1982) from two cores through Upper Quaternary deposits in the northern North Sea. According to Feyling-Hanssen (op.cit.) these assemblages reflect Arctic to high-Arctic conditions during a 'stadial age of a glacial

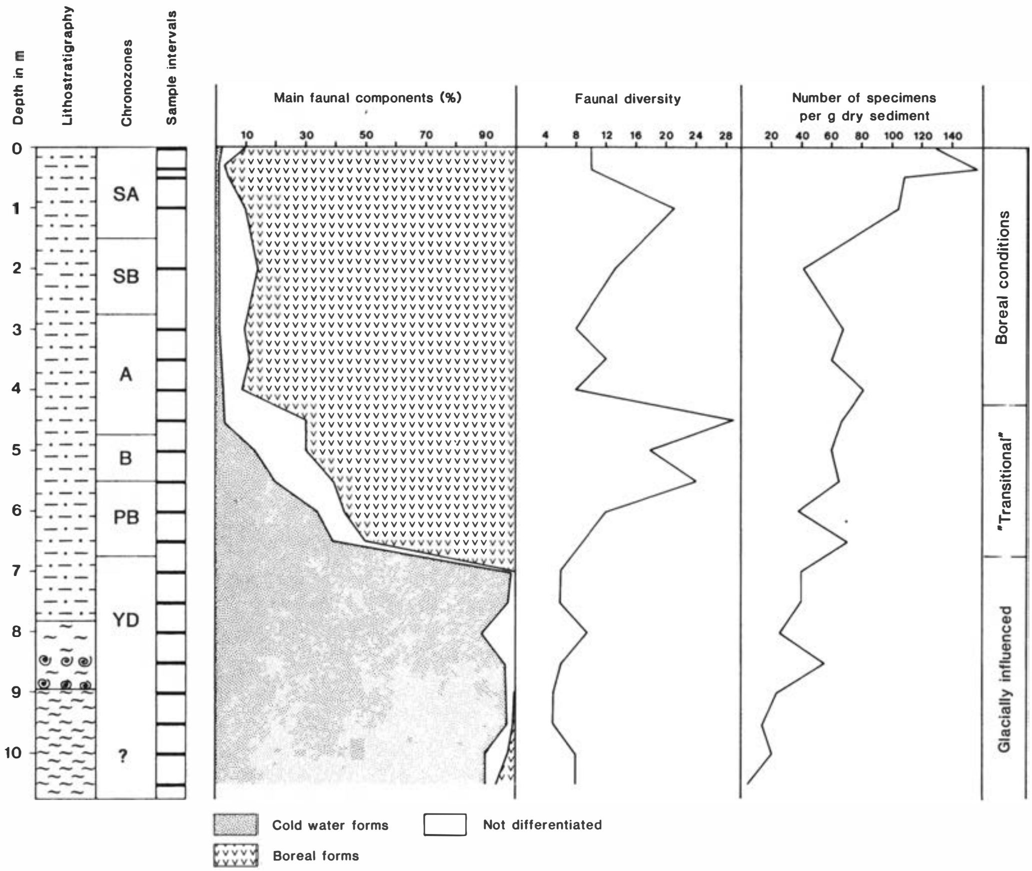


Fig. 2. Main faunal components, faunal diversity and number of benthic foraminifers in one gram dry bulk sediment. The faunal diversity are calculated in accordance with Walton (1964).

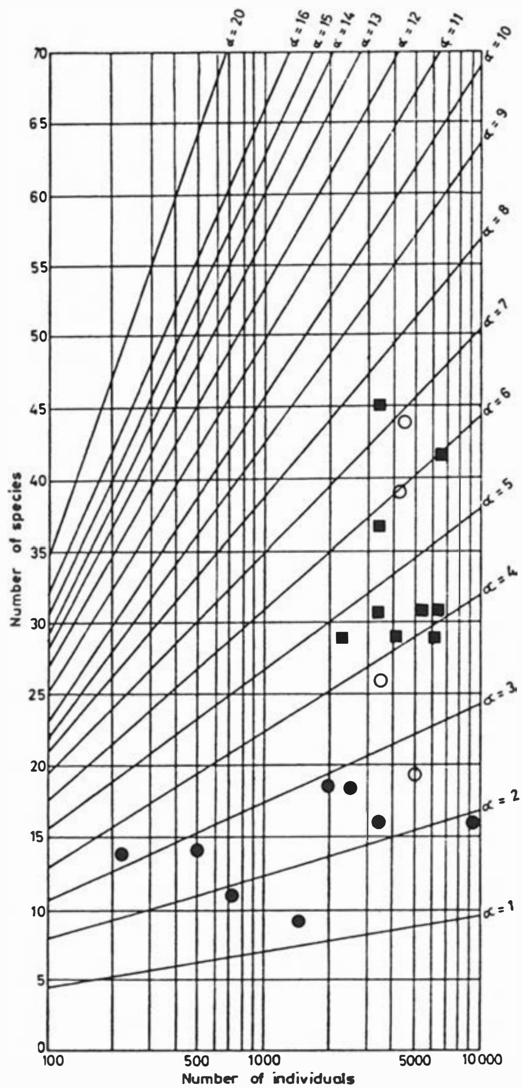
stage' in that area. He considers the salinity of the water close to normal marine.

The *C. reniforme* assemblage zone corresponds well to the assemblage zone defined in the lower part of three cores from southwestern Sweden (Knudsen 1982). These cores were studied in an attempt to find a stratotype locality for the Holocene/Pleistocene boundary (Olaussen 1982). The high-Arctic faunas in these cores were followed by more temperate assemblages, which differ, due to the different depositional environment (near shore, shallow water), considerably from the assemblages in the upper part of our core.

The *Cassidulina laevigata* Assemblage Zone

The core interval between 6.5 and 4.0 m is characterized by *Cassidulina laevigata* associated with *E. excavatum* in the lower part, and *Uvigerina peregrina, Melonis barleeanum, Bulimina marginata* and to some extent *Pullenia subcarinata* in the upper part (Fig. 1). The lower part of the *C. laevigata* zone is also characterized by a rapid decrease in the frequency of the cold water species. The faunal diversity is variable but reaches high numbers (Fig. 2) as do the Fisher α-values. *C. laevigata* suddenly appears in the sample taken at 6.50–6.55 m, where it constitutes 50 % of the total benthic foraminiferal assemblage. From the lower part of the core only two specimens were observed.

C. laevigata seems to be able to live under unstable hydrographic conditions. Its main occurrence in the Norwegian Channel today is found along the southern and western margins of the basin at depths which roughly correspond to the boundary between the seasonally changing



■ Samples from the upper 4m of the core

○ **Samples between 4.5 and 6.5m of the core**

● Samples from the lower (7.0–10.0m) part of the core

Fig. 3. Diagram showing the Fisher α-values of the assemblages. For this diagram the graphic method described by Murray (1973) was used.

surface water masses and the stable bottom water mass (van Weering & Qvale 1983, Qvale & van Weering, in prep.). On the slope off western Norway *C. laevigata* is most abundant below the boundary zone between the Norwegian Sea intermediate water and the Norwegian Sea deep water (Skarbø 1980). In the western Barents Sea an assemblage dominated by *C. laevigata* occurs near the boundary zone between Arctic and Atlantic water (Østby & Nagy 1982).

U. peregrina, which increases in numbers towards the top of this zone is known to tolerate low oxygen content of the water (Pflum & Frerichs 1976, Lohmann 1978, Streeter & Shackleton

1979, Schnitker 1979, 1980). *U. peregrina* is the most abundant species found in surface sediments from the deepest part of the Norwegian Channel off western Norway (Qvale & van Weering, in prep.).

The immigration of *C. laevigata* (and later other temperate water species) reflects a major environmental change which must be due to an influx of warm, saline Atlantic water into the Skagerrak. The Atlantic water thus entered the Skagerrak much later (about 3000 years) than the Norwegian Sea, where the immigration of *C. laevigata* (Skarbø 1980) has been dated to about 13,000 years B.P. (Mangerud 1977, Kellogg et al. 1978).

Regarding *C. laevigata* as an indicator of unstable hydrographic conditions, the Skagerrak area remained hydrographically unstable for about 2000 years (according to our chronozones) after the first influx of Atlantic water.

The *Hyalinea balthica* Assemblage Zone

An assemblage dominated by *Hyalinea balthica* associated with *B. marginata, U. peregrina, M. barleeanum* and a few percent of *Pullenia bulloides* characterizes the samples from 3.5 to 0.5 m in the core (Fig. 1). The faunal diversity is variable, with a maximum around 1 m (Fig. 2). The Fisher α-values are generally between 4 and 8 (Fig. 3).

H. balthica makes up 10–15 % (max. 25 %) of the benthic foraminiferal assemblage in surface sediments from most parts of Skagerrak (van Weering & Qvale 1983).

H. balthica occurs frequently in interglacial (?Eemian) deposits from the North Sea (Feyling-Hanssen 1980), while it is absent in sediments of Weichselian age of the same area. This indicates that its northern boundary of distribution was much further south than it is today. *H. balthica* is characterized as a boreal-lusitanian form (Nørvang 1945), and the immigration of this species into the Skagerrak may therefore indicate an amelioration of the climate. As for *C. laevigata* the immigration seems to take place over a rather short time interval, and it is likely that it has been connected to a change in the circulation pattern. The immigration of *H. balthica* into Skagerrak corresponds roughly to the opening of the English Channel, about 7,800 years B.P. (Jelgersma 1979), and which represents one possible immigration route.

The *Bolivina* cf. *B. robusta* Assemblage Zone

The upper 40 cm of the core are characterized by an assemblage dominated by *Bolivina* cf. *B. robusta* (Fig. 1). In the sample taken at 0.50–0.55 m in the core only a few specimens are observed, while in the interval 0.35–0.40 m *B.* cf. *B. robusta* constitutes nearly 30 % of the total assemblage. This sudden immigration of *B.* cf. *B. robusta* has been observed by Lange (1956) and Jørgensen et al. (1981) in cores from the central part of Skagerrak. In cores studied by Fält (1977, 1982) from the Swedish west coast, *B.* cf. *B. robusta* occurs only in the upper part, but due to the shallow water depths during the deposition of these cores it is only a minor constituent of the total assemblage. Today *B.* cf. *B. robusta* characterizes the assemblages occurring under the stable deep water mass of Skagerrak (van Weering & Qvale 1983).

Lange (1956) and Jørgensen et al. (1981) observed three distinct peaks in frequency of *B.* cf. *B. robusta.* Kihle (1971) also reported fluctuations in the occurrence at different sediment levels, but these are not further described or illustrated in his paper. Since the *B.* cf. *B. robusta* zone is very compressed in our core, greater sample density was required to study this. Nigam (unpubl. data) analysed samples every 5 cm of the upper metre of the core, and found three distinct peaks in the upper 40 cm. We are presently not able to explain these fluctuations.

Based on the foraminiferal zonation of his core Lange (1956) suggested an age of about 2000 years B.P. for the immigration of *B.* cf. *B. robusta.* ^{14}C-datings carried out on organic carbon by Jørgensen et al. (1981) gave an estimated age of 1000 years B.P. Pollen zonation (Henningsmoen & Høeg, this volume) indicated an age of about 1300 years B.P., maybe somewhat younger, between 0.75–0.5 m in our core, and thus corresponds well to the ^{14}C-datings by Jørgensen et al. (1981). We assume that the immigration is close to isochronous over the entire Skagerrak.

As mentioned earlier, *B.* cf. *B. robusta* lives under the stable bottom water masses in Skagerrak today. The deep water is formed by cascading of Atlantic water and is especially pronounced during extremely cold winters (Lee 1980). The deterioration of the climate about 900 years B.P. might have initiated the formation of the Skagerrak deep water and thus the establishment of Recent hydrographic conditions. We are, however, not able to explain from where *B.* cf. *B. robusta* came into Skagerrak, as this species is known with certainty only from late Holocene sediments from the Norwegian continental shelf and Norwegian fjords.

The *H. balthica* and *B.* cf. *B. robusta* assemblage zones of our core are very similar to the assemblages in Lange's (1956) 9 m long core from the central part of the Skagerrak. Due to the much higher sedimentation rates in the central and inner part of the Skagerrak (Jørgensen et al. 1981), Lange's 9 m long core corresponds at maximum to the upper 3.5 m of our core.

Conclusions

1. The benthic foraminiferal assemblages indicate two main depositional environments for the sediment sequence in core GIK 15530-4: 1) the lower part (below 7 m) being deposited under polar, probably glacially influenced, conditions, and 2) the uppermost part (above 6.5 m) being deposited under temperate (postglacial) conditions.
2. The upper part of the core can be subdivided into three different assemblage zones; the lower boundary of each defined by a sudden immigration of *Cassidulina laevigata, Hyalinea balthica* and *Bolivina* cf. *B. robusta,* respectively (Fig. 1).
3. The hydrographic conditions have not been stable during the deposition of core GIK 15530-4. The immigration of *C. laevigata* (and other temperate water species) corresponds to the influx of Atlantic water into Skagerrak.
4. The immigration of *H. balthica* may reflect a climatic amelioration and/or a change in the circulation pattern. The immigration of *B.* cf. *B. robusta* indicates that the Skagerrak had reached about the same hydrographic conditions as today, and may reflect the formation of the Skagerrak deep water. The pronounced fluctuations in the frequency of *B.* cf. *B. robusta* may be due to minor changes in environment. Detailed studies of the sediment and other fossil groups must be carried out before we may be able to solve this problem.

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Planktonic foraminifers in Upper Quaternary marine Skagerrak sediments

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Thiede J.: Planktonic foraminifers in Upper Quaternary marine Skagerrak sediments. *Norsk Geologisk Tidsskrift*, Vol. 65, pp. 115–118. Oslo 1985. ISSN 0029-196X.

Planktonic foraminiferal faunas in Holocene Skagerrak deposits consist of six different species: *Globigerina bulloides, G.falconensis, G. quinqueloba,* dextral and sinistral *Globoquadrina pachyderma, Globigerinita glutinata* and *G. uvula.* The oldest (Preboreal and Boreal) specimens are found in a size fraction 1.0–0.125 mm, whereas the uppermost deposits (Atlantic to Subatlantic) also contain frequent very small specimens. The appearance of planktonic foraminifers coincided with a sharp decrease of coarse ice-rafted terrigenous debris in the sediments indicating the presence of Atlantic waters. The appearance of numerous small specimens might indicate the development of a Norwegian Coastal Current regime since Atlantic time.

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Planktonic foraminifers are commonly inhabitants of open ocean surface water masses. When they are found as fossils in neritic sediments they indicate the advection of oceanic into coastal water masses. Such planktonic foraminiferal shell assemblages are often characterized by frequent kummerforms (Berger 1971) and anomalous species compositions. Along the Norwegian continental margin, planktonic foraminiferal shells are rare to common constituents of the coarse fractions of Holocene surface sediments. They are known to enter the Skagerrak with the water masses from the North Sea and they have therefore been observed in low concentrations even in the inner Skagerrak (Qvale & Thiede 1980). Planktonic foraminifers have also been found in core GIK 15530-4 although they occur less frequently than anticipated if compared to other cores from the outer Skagerrak (Kihl 1971).

Methods

Planktonic foraminiferal shells have been indentified and counted in samples which have been taken from core GIK 15530-4 collected in the outer Skagerrak (Stabell et al., this volume). The samples have been washed through 0.063 mm sieves and the sand-sized material has then been split into three size fractions: > 1.0 mm, 1.0–0.125 mm, 0.125–0.063 mm. Planktonic foraminiferal shells have only been found in the two smaller size fractions. They have been identified in general following Bé 1977. The results of these counts from core GIK 15530-4 are presented in Table 1 and Fig. 1.

Distribution of planktonic foraminifers in core GIK 15530-4

In the size fraction 1.0–0.125 mm, planktonic foraminiferal shells occur too sporadically to enable reconstruction of the faunal changes throughout the core. However, the core can be subdivided into two intervals, one from 655 to 1060 cm which is virtually devoid of Quaternary planktonic foraminifers. Only in sample 1005–1010 cm were two shells of sinistral *Globoquadrina pachyderma* found, and in sample 805–810 cm one shell of *Globigerinita glutinata.* Samples from the upper part of the core contain regularly a few planktonic foraminiferal shells of *Globigerina bulloides, G. falconensis, G. quinqueloba,* dextral *Globoquadrina pachyderma* and *Globigerinita glutinata.* The most frequent specimens in this size fraction belong to the species *G. bulloides* and dextral *G. pachyderma.*

The occurrences of planktonic foraminiferal shells in the finer size fraction (0.063–0.125 mm) allowed the core to be subdivided also into

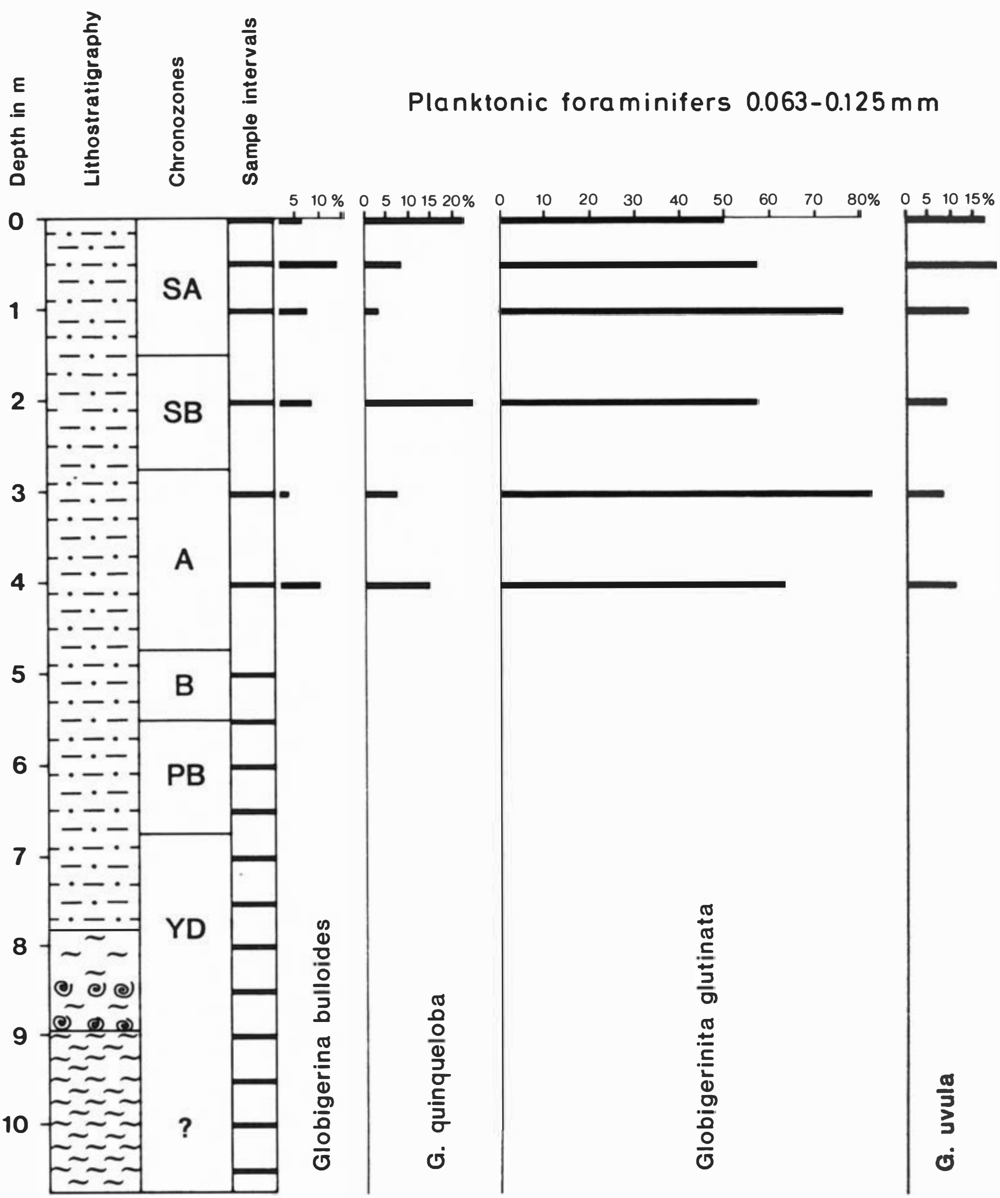

Fig. 1. Distribution of planktonic foraminifers (0.063 mm–0.125 mm) in core GIK 15530-4 from the outer Skagerrak.

two halfs. Whereas virtually no Quaternary planktonic foraminifers in this size fraction have been observed below the sample at 405–410 cm, they are frequent components of the particle assemblage in the fine-sand size fractions of the upper four metres of the core. It is not quite clear if they are too diluted because of high fluxes of sand-sized terrigenous detrital material or if they once lived too sparsely in the surface water masses of the Skagerrak. The planktonic foraminiferal faunas in the fine-sand fractions of the upper part of the core consist of six different species: *Globigerina bulloides*, *G. falconensis*, *G. quinqueloba*, dextral and sinistral *Globoquadrina pachyderma*, *Globigerinita glutinata* and *G. uvula*. However, of these six species only *G.*

Table 1. Distribution of planktonic foraminifers in GIK 15530-4 (Size fraction 0.063 - 0.125 mm)

Depth in core (cm)	Total no. counted (only Quaternary forms)	% G. bulloides	% G. falconensis	% G. pachyderma (R)	% G. pachyderma (L)	% G. quinqueloba	% G. glutinata	% G. uvula	Cretaceous and Tertiary reworked planktonic foraminifers (no. of observations during the count)
5- 10	18	6		6		22	50	17	3
50- 60	201	13	1	1		8	57	20	1
105-110	297	7	1	3		3	76	13	1
205-210	111	8	1	1		24	57	9	3
305-310	286	3	3		3	7	81	8	5
405-410	63	10	2	2		14	62	11	21
505-510		No Quaternary planktonic foraminifers found further down core							1

bulloides, G. quinqueloba, G. glutinata and *G. uvula* occur frequently; the latter three species have typical small shells and are characteristic for the abnormal faunas observed under the Norwegian Coastal Current (Bjørklund et al. 1979, Qvale & Thiede 1980).

Reworked globotruncanid and heterohelicid foraminiferal shells of Late Cretaceous age and some lower Tertiary planktonic foraminifers are rare in almost all samples of the fine-grained sand fraction from the upper four metres of the core (cf. Table 1).

Paleoceanographic significance of planktonic foraminifers in Upper Quaternary Skagerrak sediments

It is not clear how modern planktonic foraminifers reach the area of the Skagerrak. Jarke (1961) has mapped planktonic foraminiferal distributions in North Sea surface sediments. Their frequency distribution suggests that Atlantic water masses entering the North Sea north of England carry planktonic foraminifers which are then sedimented in the epicontinental sea. They would then enter the Skagerrak carried by the Jutland Current. However, in the region of the Norwegian Channel, subsurface waters originating in the Norwegian Sea have been observed which also might carry planktonic foraminifers.

The subdivision of the core into two intervals virtually with and without Quaternary planktonic foraminifers suggests a major change of the surface water hydrography close to the boundary between Younger Dryas and Preboreal, approx. 10,000 years ago. During the Late Weichselian and the early part of the Holocene, watermasses with planktonic foraminifers obviously did not reach the area of the coring site, and Quaternary planktonic foraminiferal shells therefore only occasionally occurred as stray forms embedded into the sediments. It seems at present impossible to judge if the occurrence of these specimens might be due to reworking of older Quaternary foraminifer bearing deposits.

The appearance of planktonic foraminifers as well as other temperate microfossil groups coincide with a sharp decrease of the concentrations of coarse probably ice-rafted terrigenous debris in the sediment (Stabell et al., this volume). It seems remarkable that the oldest sediments which contain relatively frequently planktonic foraminiferal shells are of Preboreal and Boreal age. These early planktonic foraminiferal faunas comprise dominantly large specimens and seem to originate from Northeast Atlantic waters which enter the Skagerrak from the North Sea and the Norwegian Sea.

During the later part of the Holocene (Atlantic and younger), however, the planktonic foraminiferal faunas are dominated by very small forms. Such faunas are today characteristically found in areas under the Norwegian Coastal Current which is different from Norwegian Current water masses because of reduced salinities. If this correlation holds, the establishment of the Norwegian Coastal Current can be traced back to Atlantic time, when a circulation regime of the surface water in the Skagerrak of approximately the same pattern as we know from today was established.

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Ostracods in Upper Quaternary Skagerrak deposits

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The core GIK 15530-4 from the outer Skagerrak can be divided into three distinct ostracod assemblage zones. The interval below 7 m is characterized by a cold water assemblage with *Cytheropteron paralatissimum* and *Normanicythere leioderma*. The 6.5 m level is transitional and dominated by *Sarsicytheridea bradii*. The upper 6 m is characterized by a boreal (temperate water) assemblage dominated by *Krithe praetexta, Muellerina abyssicola, Echinocythereis echinata, Cytheropteron alatum* and *C. testudo*. This change is interpreted as indicating the influx of warmer Atlantic water into the Skagerrak.

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Ostracods inhabit all aquatic environments. They are important both in the benthos and as marine zooplankton, but only benthic ones have been found as fossils in sediments. During recent years ostracods have become increasingly useful for recognizing climatically induced shifts in near-shore ocean water temperatures (Cronin 1981, Hazel 1968, among others).

The distribution patterns and ecological preferences for most of the species observed in core GIK 15530-4 are fairly well known, due to extensive work by Sars (1866, 1928) off Norway, by Elofson (1941) off Sweden and in the Skagerrak, by Rosenfeld (1977) in the Baltic Sea, and by Neale & Howe (1975) in the Barents Sea.

The study of the ostracods was based on the same samples and fraction as for the benthic foraminifers (see Nagy & Qvale, this volume). All specimens were picked out and transferred to a separate slide for later identification and counting.

The ostracods were generally well preserved, but occurred, unfortunately, mostly as isolated valves, and only a few articulated carapaces have been found. The counts therefore represent single valves (Fig. 1). The lack of preserved appendages made the identification of the smooth shelled forms difficult. Many of the juvenile instars, which formed a large part of the assemblages, were also often difficult to identify. Most of the ostracods observed were Quaternary (Recent) forms, and only a few undoubtly reworked (? Tertiary) specimens were recognized.

Species distribution

Ostracods occur in all samples studied, but in very low numbers below 8 m (Fig. 1). All species identified are benthic, marine forms. The number of (single) valves in one sample varies from 1 (in sample 10.50–10.55 m) to 479 (in sample 4.50–4.55 m). The number of species is rather low; in total 43 species have been identified from the entire core. The highest diversity occurs between 1 m and 6 m in the core. Eight species dominate the assemblages, while the remaining 35 species are represented by only a few valves. A list of the eight species and selected references to their synonomy are given in Table 1.

The core can be divided into three different assemblage zones. The lower part of the core (from 7.0 m and down) is characterized by a very poor assemblage dominated by Arctic (cold water) species, while the upper 6.0 m are dominated by boreal (temperate water) species. One level (6.5) m is characterized as transitional, and is dominated by species which tolerate a wide range of temperatures.

The lowermost zone is characterized by *Cytheropteron paralatissimum* and *Normanicythere leioderma*. These species have been reported from Upper Pleistocene, near-shore marine deposits in Alaska (Swain 1963), Canada (Cronin 1981) and England (Neale 1959), and from the Recent sediments in the Barents Sea (Neale & Howe 1975), and thus indicate polar to subpolar conditions. The minor species of this lower unit, *Heterocyprideis sorbyana* and *Acanthocythereis dunelmensis*, are also cold water species which seem to tolerate reduced salinities (Rosenfeld 1977).

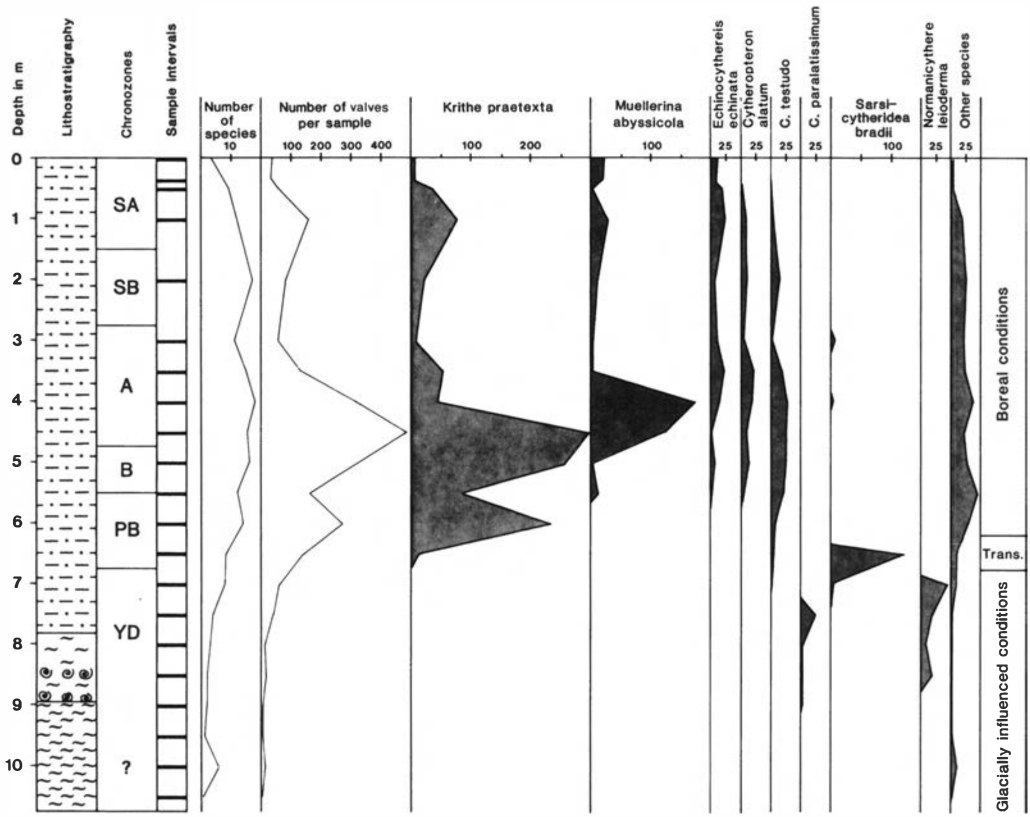

Fig. 1. Distribution of ostracods in core GIK 15530-4. Due to the similarity in sample size (about 60 g) throughout the core, and the usually low number of valves per sample, the abundance of each species is given as total number of valves and not as relative frequency. Please note the difference in scale.

Sarsicytheridea bradii is found in great numbers in one sample (6.50–6.55 m) and characterizes the transitional zone. *S. bradii* tolerates higher temperatures and has a more widespread distribution than the typical Arctic species. It is abundant around the entire Norwegian coast at moderate depths (Sars 1928) and occurs around the British Isles, Iceland, Arctic Canada and in the Barents Sea. The occurrence of *S. bradii* is interpreted to indicate an influx of more temperate water.

The upper 6 m of the core are dominated by boreal (temperate water) species (Fig. 1) and they also indicate normal marine salinities (> 30‰). The dominant species are *Krithe praetexta* and *Muellerina abyssicola, Cytheropteron alatum* and *C. testudo*. Also the minor species are Boreal forms, like *Argilloecia conoidea, Polycope* spp., *Cyprideis torosa,* among others. These species are quite common in Norwegian fjords and along the coast up to Lofoten (Sars 1928), and in the Skagerrak (Elofson 1941). The same species that characterize the upper 6.5 m of the core 15530-4 were also the most important in a 9 m long core from the central part of Skagerrak, studied by Lange (1956). Due to the much higher sedimentation rate in the inner Skagerrak (Jørgensen et al. 1981), Lange's core can be correlated with the upper 4 m (maximum) of our core (Nagy & Qvale, this volume).

Lord (1982) studied the ostracods in three cores from south-western Sweden collected in search of a stratotype locality for the Holcoene/ Pleistocene boundary (Olausson 1982). These cores penetrated sediments deposited in very shallow nearshore environments, and, consequently, the ostracod assemblages were documented by species different from those recorded in the present study.

Table 1. Synonymy (selected references) of the most important species in core GIK 15530-4

Cytheropteron alatum Sars
Cytheropteron alatum Sars, 1866, p. 81.

C. paralatissimum Swain
Cytheropteron paralatissimum Swain, 1963, p. 817., pl. 95, fig. 12, text-fig. 8b.

C. testudo Sars
Cytheropteron testudo Sars, 1869, p. 173.

Echinocythereis echinata (Sars)
Cythereis echinata Sars, 1866, p. 44; *Echinocythereis echinata* (Sars) van Morkhoven, 1963, p. 173.

Krithe praetexta (Sars)
Ilyobates praetexta Sars, 1866, p. 60; *Ilyobates bartonensis* (Jones) Brady, 1868, p. 432, pl. 34 figs. 11–14, pl. 40, fig. 5; *Krithe bartonensis* (Jones) Brady, Crosskey & Robertson, 1874, p. 182, pl. 40, figs. 22–26 (non *Cythere (Cytherideis) bartonensis* Jones, 1857 = *Dentokrithe bartonensis* (Jones) Khosla & Haskins, 1980); *K. bartonensis* (Jones) Sars, 1928, p. 165, p. 76; *K. bartonensis* (Jones) Lange, 1956, p. 81, Pl. 9, fig. 22, pl. 10, fig. 22; *K. praetexta* (Sars) Athersuch, 1982, p. 242, pl. 7, figs. 6, pl. 8, figs. 5–8.

Muellerina abyssicola (Sars)
Cythereis abyssicola Sars, 1866, p. 43; *Hemicythere latimarginata* (Speyer) Sars, 1928, p. 188, Pl. 86, Fig. 3; *Cythereis (Paracythereis) latimarginata* (Speyer) Lange, 1956, p. 82, pl. 9, fig. 20, pl. 10, fig. 20; *Muellerina abyssicola* (Sars) Bassiouni, 1965, p. 510, pl. 1, figs. 3–6.

Normanicythere leioderma (Norman)
Cythere leioderma Norman, 1869, pp. 255, 291; *Cythereis leioderma* (Norman) Blake, 1933, p. 239; *Normanicythere leioderma* (Norman) Neale, 1959, p. 78, pl. 13, figs. 1–2, pl. 14, figs. 1–8; *N. concinella* Swain, 1963, p. 827, pl. 95, figs. 18a-d, text-fig. 11b.

Sarsicytheridea bradii (Norman)
Cythere bradii Normann, 1865, p. 192; *Cypredeis bairdii* Sars, 1866, p. 52; *Cytheridea papillosa* Bosquet Sars, 1928, p. 159, pl. 73, pl. 74, fig. 1
Eucytheridea bairdii (Sars) Rosenfeld, 1977, p. 21, pl. 4, figs. 53–56;
Sarsicytheridea bradii (Norman) Athersuch, 1982, p. 241, pl. 7, figs. 2, 4, pl. 8, figs. 1–4; figs. 7a, 8c-e.

Conclusions

Three distinct assemblage zones can be established: 1) a lower (below 7 m) zone characterized by Arctic (shallow water) species and low diversity, 2) a transitional zone characterized by *Sarsicytheridea bradii,* and 3) the upper (above 6 m) zone characterized by a boreal (temperate) assemblage and higher diversity. The faunal shift indicates influx of warmer, more saline water, probably from the Atlantic at the beginning of this time represented by zone 2. The very high number of specimens in the samples between 6 and 4 m in the core must be due to either very favourable ecological conditions and/or reduced sedimentation rates.

It is not possible based on the ostracod assemblages to make a further subdivision of the upper 6 m of the core.

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Evaluation of depositional environment

Upper Quaternary accumulation rates of marine outer Skagerrak sediments: Core GIK 15530-4

JÖRNTHIEDE

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The flux of the sediments at the location of core GIK 15530-4 in the outer Skagerrak has been quantified in terms of accumulation rates (weight per area and time unit) for the time span of the past 11,000 years. Bulk sediment accumulation rates are approximately 50–100 g $cm^{-2}10^{-3}y^{-1}$ during the entire time span, except during the Younger Dryas (10,000–11,000 years B.P.) when they rose to > 800 g $cm^{-2}10^{-3}y^{-1}$. The dominant portion of these sediments consists of clayey and silty material except during the Younger Dryas when material coarser than 0.063 mm, in part pebble sized probably ice rafted material accumulated at the coring site with rates of up to 80 g $cm^{-2}10^{-3}y^{-1}$ (for example also documented by sand-sized quartz grains). Biogenic components are quantitatively rare in these sediments, but during the Younger Dryas both opaline and calcareous ones accumulate many times faster than usual.

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Accumulation rates and their importance for the understanding of a depositional environment

The speed at which sediments accumulate is an important variable in attempting to understand a depositional environment. This speed can be expressed in various ways. To do so, usually linear *sedimentation rates* are used because they can be calculated based on sediment thickness between stratigraphic fix points (in length per time unit). However, this variable does not take into account variations in pore space or pore water contents, mainly because in most sediment studies these variables have not been measured. However, *accumulation rates* which define this speed in terms of weight per area and time units are a much better variable to express quantitatively the speed of sedimentation because the sedimentation rates have been corrected for the usually downcore decreasing pore space. Once accumulations for the bulk sediment have been calculated, they can also be apportioned according to sediment composition and one is able to quantify the particle flux of individual grain categories through time. Procedures for these calculations have been developed mainly based on deep-sea sediments (van Andel et al. 1975).

In this study, then, accumulation rates will be described for core GIK 15530-4, which penetrated marine Upper Quaternary deposits in the outer Skagerrak (Stabell et al., this volume). These sediments blanket wide areas of the Skagerrak, and their lithology appears to be rather homogenous, although their stratigraphy clearly reveals the response of their depositional environment to changes of the paleohydrography of the Skagerrak and of the Upper Quaternary paleoclimate of NW Europe. Up to now estimates of accumulation rates have only been made for very young Holocene deposits (Suess & Erlenkeuser 1975) in time scales of a few tens to a few hundred years (see also Müller & Irion 1984). With the data of this core we are able to present evidence (although on a less detailed scale) which extends back into the end of the Last Glacial, probably to the early Younger Dryas (approx. 11,000 years B.P.). These data will allow us to quantify precisely the flux of the bulk sediment and of its various components through time at the locality of this core (Stabell et al., this volume), and a few examples of such component fluxes will be presented.

The determination of accumulation rates

The determination of bulk accumulation rates requires detailed data about age, sediment thickness and water content of the section under in-

Table 1. Core GIK 15530-4: Stratigraphic fix points, methods, age and linear sedimentation rates.

Core level (cm below surface)	Age (years B.P.)	Method *	Sedim.-rate (cm $10^{-3}y^{-1}$)
2.5	17.5	Pb	143
6.5	33.0	Pb	258
8.5	55.0	Pb	91
10.5	85	Pb	67
12.5	90	Pb	400
14.5	120	Pb	67
16.5	160	Pb	50
75	1200	P	56
275	5000	P	53
425	7000	P	75
500	8200	P	63
575	9400	P	63
[670	10000	O	158] 1)
675	10200	P	125
[850-885	10500	C	250 (min)-367 (max)] 1)
895	10600	(C)	100 (min)-450 (max)

* P = Pollen (from Henningsmoen & Höeg, this volume)

Pb = ^{210}Pb (from Erlenkeuser, this volume)

C = ^{14}C (from Stabell, this volume), (C) = inferred Radiocarbon age (from Bjørklund, this volume)

O = ^{18}O (from Erlenkeuser, this volume)

1) Datum has not been used because of the uncertainty of its precise relationship to the pollen-datum/ash layer datum just below. For the calculation of the sedimentation rates prior to 10200 years B.P. we have used the lowermost pollen-datum and the datum obtained from the ash layer; the resulting value is 550 cm $10^{-3}y^{-1}$, the extra polated age for the base of the core is 10930 years B.P. For discussion of possible errors see also above-mentioned papers.

vestigation (van Andel et al. 1975). Compositional data of the sediments can then be used to apportion the bulk sediment accumulation rates according to the proportion of individual sediment components. The accumulation rates in this paper have been expressed in g $cm^{-2}10^{-3}y^{-1}$. Because coarse fraction compositions have been determined by grain counts, they do not exactly correspond to weight percent.

The age data of core GIK 15530-4 used for the establishment of the temporal framework have been compiled in Table 1. They have been taken from various contributions in this volume, where also detailed discussions about discrepancies be-

Table 2. Core GIK 15530-4: Sedimentation rates, water contents (from Rosenqvist & Pederstad, this volume) and calculated accumulation rate data.

Core level (cm below surface)	Sedimentation Rates (cm $10^{-3}y^{-1}$)	Water Content (% of dry sed.)	Acc. Rates ($gcm^{-2}10^{-3}y^{-1}$)
10	153	147	162
25	102	127	117
50	50	111	61
75	56	104	72
100	53	96	71
125	53	96	71
150	53	90	73
175	53	86	75
200	53	87	73
225	53	88	73
250	53	86	75
275	53	86	75
300	75	86	106
325	75	80	110
350	75	78	110
375	75	77	112
400	75	72	114
425	75	71	114
450	63	68	99
475	63	69	97
500	63	69	97
525	63	65	100
550	63	63	100
575	63	64	100
600	125	68	196
625	125	70	192
650	125	71	189
675	125	71	189
700	550	75	820
725	550	73	830
750	550	73	830
775	550	74	825
800	550	75	820
825	550	74	825
850	550	71	840
875	550	63	881
900	550	62	886
925	550	66	865
950	550	71	840
975	550	70	844
1000	550	68	855
1025	550	70	844
1050	550	70	844

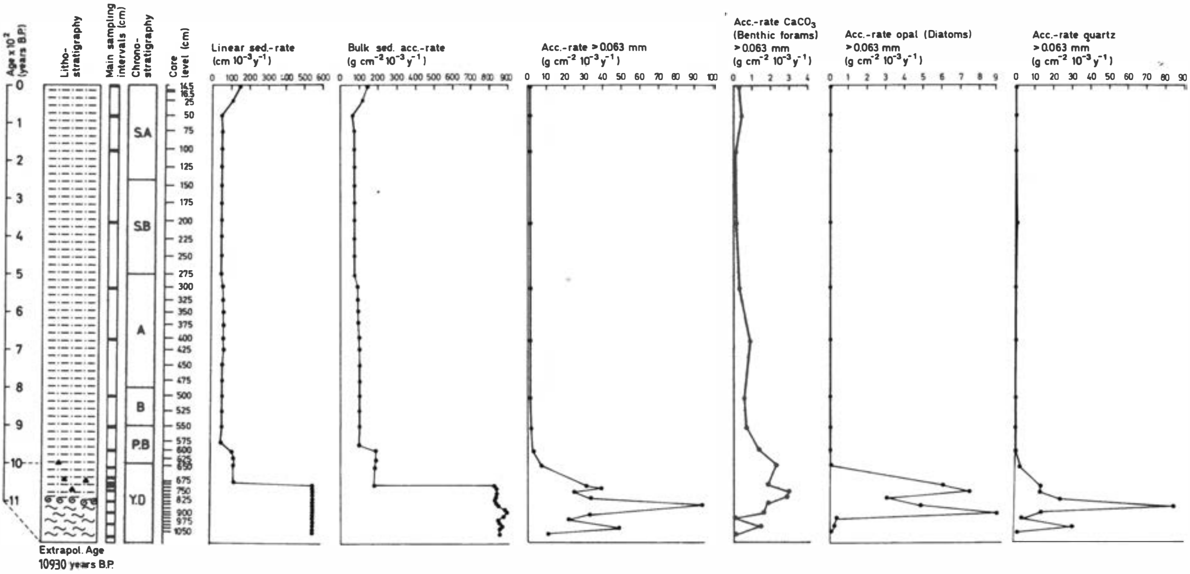

Fig. 1. Accumulation rates (g $cm^{-2}10^{-3}y^{-1}$) at site GIK 15530-4. The chronostratigraphy of the late Quaternary follows Mangerud et al. (1974). The time levels have been chosen based on ^{210}Pb datings, pollen, radiocarbon and oxygen isotope data (Table 1). The ages in the lowermost part of the core have been extrapolated assuming linear sedimentation rates. Coarse fraction data which were needed for the calculation of accumulation rates, foraminifers, diatoms and quartz are from Thiede (a, this volume).

tween individual methods and errors are discussed in detail. A systematic difference which is suggested between the ages based on magnetic data and on pollen data was most problematic. The discrepancies in placing stratigraphic boundaries do not exceed 1000 years, but are usually less. For the sake of consistency and because it did not appear solvable at the present time, I have followed the pollen datings of Henningsmoen & Høeg (this volume). Ages in between the stratigraphic fix points (Table 1) have been estimated based on constant and uninterrupted sedimentation between these points (linear sedimentation rates). It is interesting to note that the ages of the surface sediment layer based on ^{210}Pb dating (Erlenkeuser, this volume), suggest that the Recent surface sediments are completely or almost completely preserved in this core.

The water contents data have been taken from Rosenqvist & Pederstad (this volume), the grain size and coarse fraction data from Thiede (this volume). The density of the sediment has been assumed to be 2.61, and the calculations follow largely the procedure developed by van Andel et al. (1975). Sedimentation rates, water content and bulk sediment accumulation rate data are tabulated in Table 2 for each 25-cm interval. However, since most compositional data have only been determined for a rather coarse grid of samples, accumulation rates for these points have been determined by smoothing the data of Table 2 by 3-point-running average (weighted 1-2-1). Sedimentation rates, accumulation rates of the bulk sediment, of the total sand-sized material, of sand-sized foraminifers, diatoms, and quartz have been plotted against a time scale in Fig. 1. Considerations about accumulation rates of organic matter and some of its components can also be found in Wassmann (this volume).

Accumulation rates in core GIK 15530-4

The available stratigraphic fix points allow us to determine the age of sections of core GIK 15530-4 above the 900 cm level (Table 1, Fig. 1). Although the available points have a somewhat uneven spacing, that part of the core can be dated to the time span of earliest Younger Dryas (11,000 years B.P.) to late Subatlantic time. For the core section below the 9 m-interval no actual age data are available. However, if extrapolating linear sedimentation rates from directly above, one can estimate the age of the end of the core to be close to the boundary Alleröd/Younger Dryas (see Fig. 1).

The plot of linear sedimentation rates and of bulk sediment accumulation rates against time reveal that the flux of sediments except during Younger Dryas (10,000–11,000 years B.P.) has been rather even. Bulk sediment accumulation rates are around or slightly above 800 g $cm^{-2}10^{-3}$ y^{-1} during Younger Dryas to drop off to values of 50–150 g $cm^{-2}10^{-3}y^{-1}$ during late Younger Dryas. Minor fluctuations during the past 10,200 years (e.g. during Atlantic time) are probably reflecting imprecision of stratigraphic data rather than time changes of the sediment flux.

The anomaly of the sediment flux which developed during Younger Dryas is accompanied by important changes of the sediment composition, as can be demonstrated by the accumulation rates of material coarser than 0.063 mm (Fig. 1). The proportion of sand-sized components of the bulk sediment is in general < 2%. Only in the Younger Dryas section of the core do the percentages of sand-sized material rise to > 5–10%, whereas the remainder of the bulk sediment comprises mainly clayey, and some silty material. Accumulation rates of the sand-sized components are 10 g $cm^{-2}10^{-3}y^{-1}$ in the lowermost sample of the core; but rise to a peak of > 80 g cm^{-2}-$10^{-3}y^{-1}$ during Younger Dryas, in an interval where ice-transported coarse particles are occurring in the sediments (Stabell et al., this volume). After Younger Dryas the accumulation rates of sand-sized material drop rapidly to very small values. This change coincides with a major change in composition of the sand-sized material from a dominance of authigenic and terrigenous components in Younger Dryas sediments to mainly biogenic components in the Holocene deposits (Thiede, this volume).

The dominant proportions of the sediments of core GIK 15530-4 consist of clays and silt whose source areas are discussed elsewhere in this volume and which have come in part from the North Sea area and in part from Scandinavia. However, to describe additional characteristics of the depositional environment of this core, accumulation rates of calcareous benthic foraminiferal shell material, of the opaline frustules of diatom as they appear in the coarse sections, and of coarse quartz grains have been calculated (Fig. 1). The benthic foraminifers are believed to reflect the productivity of the sea floor environment. The peak of the fluxes of large diatom frustules coincides with a maximum of *Thalassiosira antarctica* (Stabell, this volume) which is a planktonic diatom and which henceforth is believed to reflect surface water productivity. Maxima of benthic foraminifers, of diatoms and of the highest fluxes of coarse quartz grains coincide. Accumulation rates of benthic foraminifers, diatoms and coarse quartz drop rapidly off towards the end of Younger Dryas and remain low throughout the remainder of the Holocene.

Sediment fluxes and the depositional environment in the outer Skagerrak during the past 11,000 years

Accumulation rates define the speed at which particle assemblages are deposited and preserved on the sea floor. In the case of this core from the outer Skagerrak it was most surprising that this variable, except during the earliest 1,000 years of the history documented by this core, did not change drastically, despite the observation of clear biostratigraphic boundaries which indicate important changes during the past 10,000 years, both as regards the characters of the bottom water as well as of the surface water masses. Based on the Holocene accumulation rates (both for fine-grained and coarse-grained terrigenous and for biogenic components) one must conclude that the Skagerrak depositional environment has not experienced drastic changes since Preboreal time. However, we know that the configuration of this basin under the influence of relative sea level fluctuations (Stabell & Thiede, this volume) changed considerably, that it acted as part of a seaway between the North and Baltic Seas and that its surroundings experienced the effects of important climatic changes. It therefore came as a surprise that sediment fluxes during roughly the past 10,000 years have been as stable as they appear in Fig. 1.

It is only during the Younger Dryas that the coring site experienced drastic changes of sediment fluxes. Both coarse and fine-grained sediment fluxes increased by a factor of 5–10 or even higher in comparison to the following stratigraphic intervals. The lithology of the sediments deposited during Younger Dryas is quite similar to the deposits above and below, with the exception, perhaps, of increased quantities of ice-rafted coarse terrigenous clasts (Thiede, this volume). One could argue that this increased sediment flux has to do with slumping or other local erosional/depositional processes. However, the fossil assemblages do not indicate abnormally high proportions of shallow water derived materials; the contrary seems to be the case. The peak of quartz fluxes coincides with the highest proportion of ice-rafted material and probably reflects a climatic cold phase during the Younger Dryas when large amounts of floating ice covered the Skagerrak. The peak in accumulation of both sand-sized material as well as sand-sized quartz decreased quickly but gently towards the end of Younger Dryas.

The maximal accumulation rates of both biogenic components (Fig. 1) coincide roughly with the quartz maximum. It is a fact that both planktonic living organisms (in this case large diatoms) which here are believed to represent a signal of surface water productivity, and benthic shell materials as a signal of sea floor productivity, experience maximum fluxes at the same time as the maximum influx of ice-rafted material occurred. If the accumulation rates of these two biogenic components, which reflect organisms of highly different ecologies, coincide as they appear to do, then the entire marine food web must have gone through a highly productive stage, which for example also documented itself in the distribution of trace fossils in this core (Werner, this volume). This interpretation does not stand unchallenged at present because estimates of paleoproductivity based on organic carbon distributions (Wassmann, this volume) seem to indicate a reduced productivity of the marine water masses during Younger Dryas to Preboreal, but an increased productivity during the past 8,000 years with a maximum in Subatlantic time.

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Accumulation of organic matter in core GIK 15530-4 and the Upper Quaternary paleo-productivity in the Skagerrak

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An analysis of the content of particulate organic carbon (POC), particulate organic nitrogen (PON) and δ^{13}C-values of organic matter in core GIK 15530-4 revealed that the accumulated material is of modern, marine origin in the upper part of the core, whereas relatively more terrestrial components reached the Skagerrak during the Younger Dryas and Allerød. Changes in the environmental conditions of the Skagerrak during the last 11,000 years are well reflected in marked changes in the accumulation rates of organic matter and the estimated paleo-productivity. The supply of large fresh-water volumes and terrigenous matter to the Skagerrak during the Late Weichselian gave rise to highly stratified, brackish surface water with high particle content. Suppressed primary production and increased accumulation rates during the Younger Dryas are the consequences of these environmental conditions.

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The content of particulate organic carbon (POC), particulate organic nitrogen (PON) and carbonate was analyzed in core GIK 15530-4 in order to answer the following questions:

– What is the origin of the deposited organic material during the Late Weichselian and Holocene history of the Skagerrak?

– Do changes in the accumulation rates of organic material reflect changes in the environmental conditions during the past 11,000 years?

– How has the primary production of the Skagerrak changed since the Younger Dryas?

Methods

Using a CHN-analyser (Carlo Erba Strumentatione, 1106) 22 fine grained sediment samples from different depths of the core GIK 15530-4 were analyzed in triplicate for total organic carbon and particulate organic nitrogen content. The carbonate and particulate organic carbon content was determined in triplicate after adding HCl.

The organic matter concentrations were transformed to accumulation rates according to Müller & Suess (1979):

$$C_A = \frac{C \cdot S}{10} p_s(1 - \varnothing), \qquad (1)$$

where C_A = accumulation rate (g C $m^{-2}y^{-1}$) of POC, PON and carbonate;
C = POC, PON and carbonate content (% dry wt);
S = sedimentation rate (cm $10^{-3}y^{-1}$);
p_s = dry sediment density = 2,71 g cm^{-3} dry sediment, and
$\varnothing$ = porosity (Calculated according to Berner 1971).

If terrestrial organic matter input is negligible or of minor significance the paleo-primary production rate can be estimated according to Müller & Suess (1979):

$$R = \frac{C \cdot p_s \cdot (1 - \varnothing)}{0.003 \cdot S^{0.3}}, \qquad (2)$$

where R = paleo-primary production rate (g C m^{-2} y^{-1});
C = POC content;
S = sedimentation rate;
p_s = dry density of solids, and
$\varnothing$ = porosity (for units see equation (1)).

Results

The organic matter concentrations decrease sharply between the surface of the core and 0.5 m, presumably due to diagenesis of fresh organic matter (Fig. 1a), but slowly down to the depth of 6 m, showing little variation below this depth. The POC content varied between 1.74 and 0.55%.

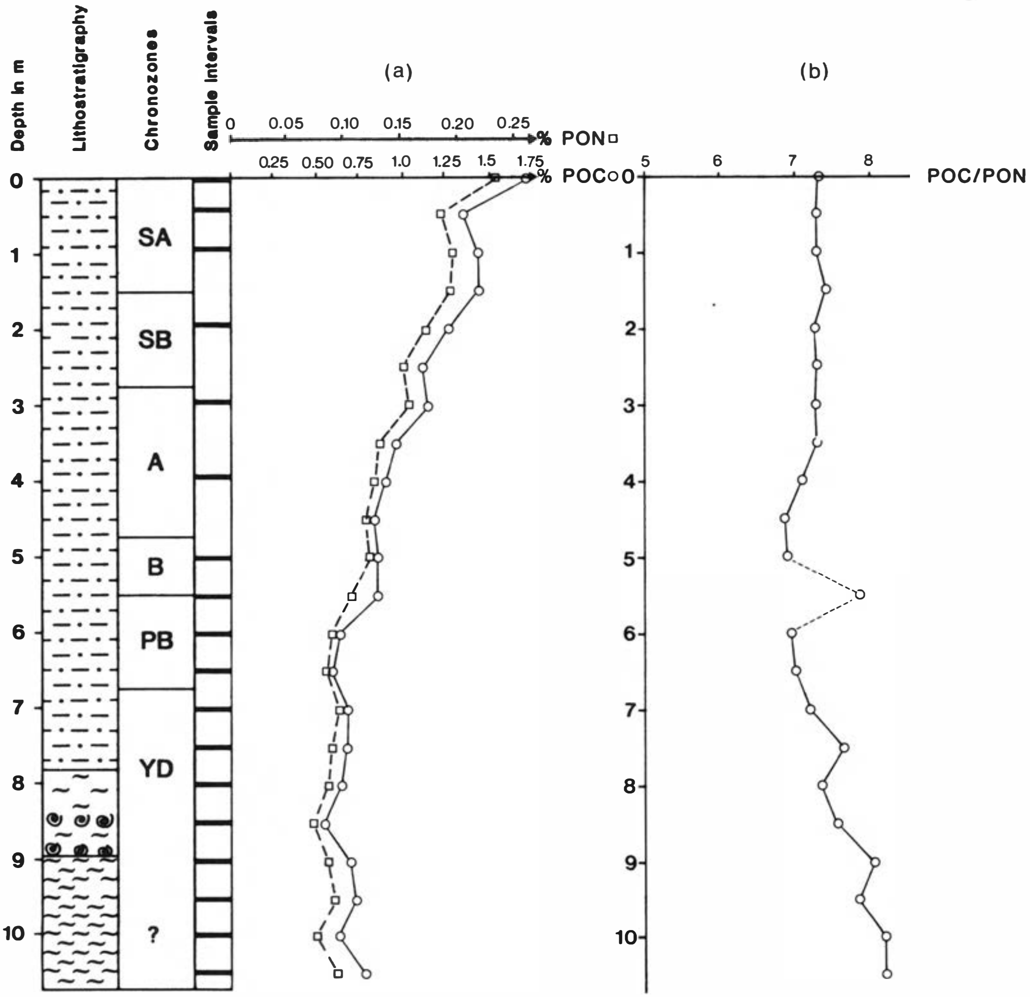


Fig. 1.
(a) Content of particulate organic carbon (POC) and particulate organic nitrogen (PON) (% dry weight) in relation to depth.
(b) POC/PON-ratio of organic material in relation to depth.

Besides one exceptionally high value at 5.5 m, the mean POC/PON-ration was about 7.3 down to 4.5 m and 6.8 from 4.5 m to 6.75 m (Fig. 1b). Below that depth, the POC/PON-ratios increased to 8.2.

Two intervals of increased POC and PON accumulation (Fig. 2) were found: in the upper 50 cm of the core and below 7 m. The same pattern was found for carbonate.

Today's primary production rate at the sampling site (90 g C m^{-2} y^{-1}) was used to predict the POC content of the sediment (Müller & Suess 1979). On the supposition that the production had not changed during the last 11,000 years, the POC content was calculated. Fig. 3b compares measured and calculated POC contents in the core. It seems to be evident that the primary production was higher between 1 and 2 m and significantly lower below 6 m than today. This is clearly demonstrated in Fig. 3c. Between 2.5 m and 4.5 m the primary production was in the same range as today (Fig. 3c).

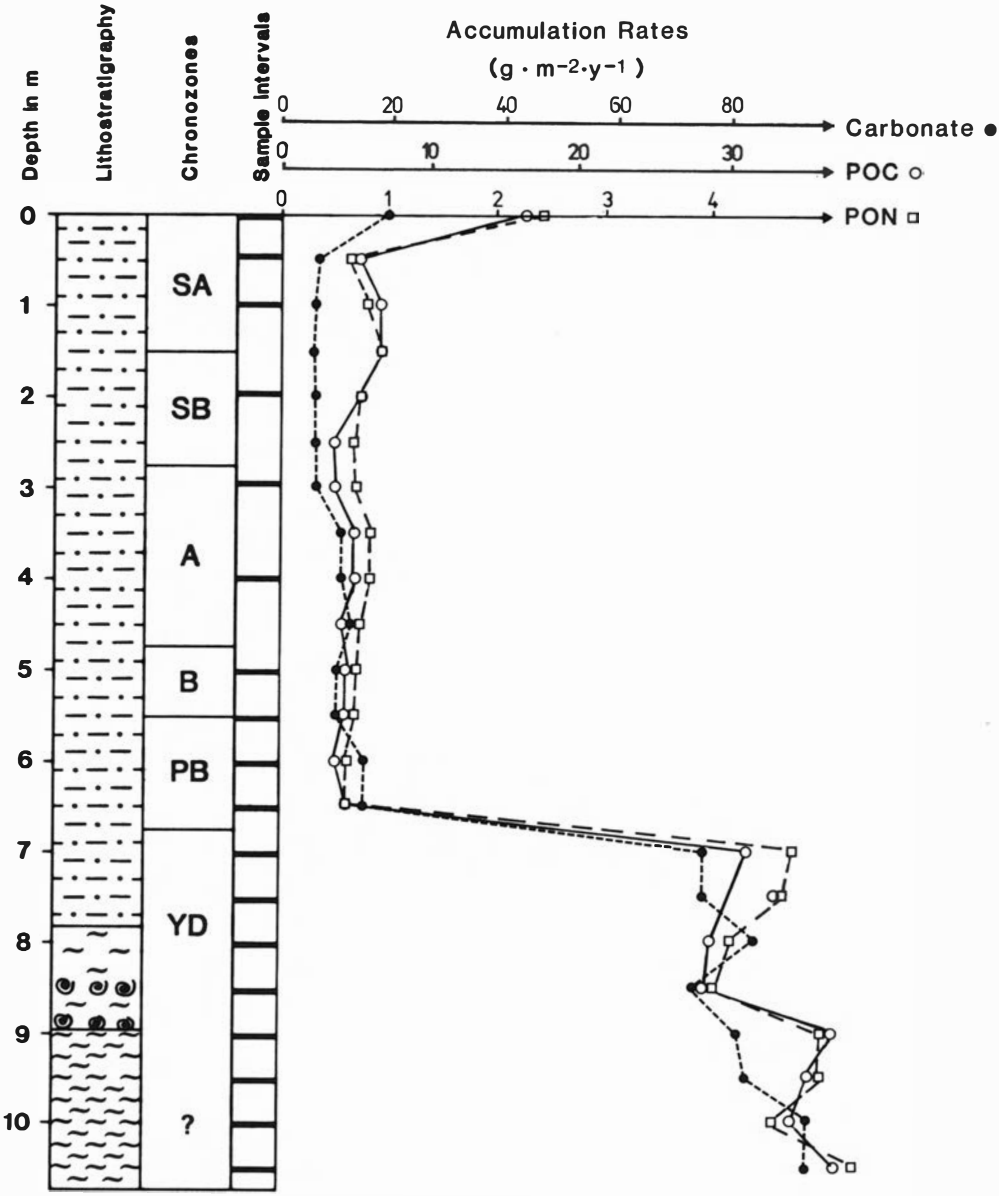
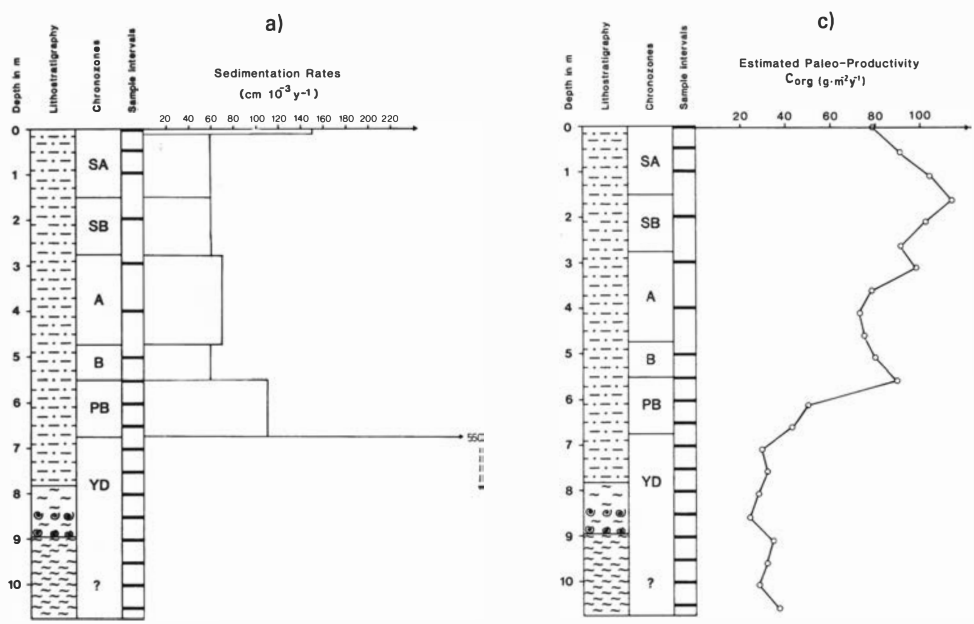


Fig. 2. Accumulation rates of particulate organic carbon (POC), particulate organic nitrogen (PON) and carbonate (g m^{-2} y^{-1}) in relation to depth.

Discussion

The POC concentrations found in the core represent the range of POC contents found in surface samples from the Norwegian Channel (Qvale, pers. comm.) and in surface sediments from other costal areas (Walsh 1981).

The POC/PON-ratio can be used as an index of areas of modern carbon deposition and possibly also of the fate of particles during transport and their origin (Walsh 1981). If the POC/PON-ratio of a sediment is about 6 and if the sediment is also rich in carbon, a marine origin might be assumed (Müller 1977). Terrestrial influenced sediments have mostly POC/PON-ratios > 10 (Müller 1977). The δ^{13}C-values (Erlenkeuser,



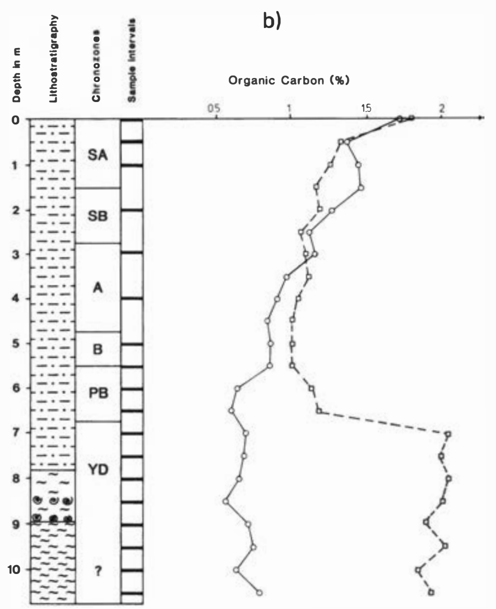


Fig. 3.
(a) Sedimentation rate (cm 10^{-3} y^{-1}) in relation to depth.
(b) Measures (circles) and calculated (squares) organic carbon content (% dry weight) in relation to depth.
(c) Estimated paleo-productivity (g C m^{-2} y^{-1}).

pers.comm.) of the organic carbon of the upper 6 m of the core ranged from −22.01 to −24.09 ‰ PDB (mean: −23.05‰; n = 22), but were significantly higher below (mean: −24.84 ± 0.56‰; n = 4). According to Chesselet et al. (1981), organic matter with $\delta^{13}C$-values of −21 ± 2‰ is considered to be of typical marine origin, but those of −26 ±2‰ are considered to be of terrestrial origin. Thus the organic matter accumulated below the depth of 6 m seems to be more influenced by terrestrial sources than the organic matter accumulated during later times. This is also reflected in increased POC/PON-ratios below 7.5 m. These results are supported by Tan & Strain (1979), who found an excellent negative overall correlation between $\delta^{13}C$- and C/N-values in recent sediments in the St. Lawrence Estuary and the Gulf of St. Lawrence. They concluded that the organic matter deposited in the Gulf of St. Lawrence (mean $\delta^{13}C$:22.43‰; mean POC content: 1.01; POC/PON-ratio: 6–8) is largely of marine origin. However, organic matter of totally marine origin is, according to Chesselet et al. (1981), not found in the core. On the other hand, Degens et al. (1968) found that phytoplankton-derived biochemical fractions had carbon isotope composition ($\delta^{13}C$ (‰)) ranging from −16 to −19‰ (pectines, amino acids, total sugars) to more than −26‰ (chloroform extractable lipids), while structural polymers like cellulose and lignin most closely approximated the isotope ratios found in the upper part of our core. Diagenetic processes which favour the mineralization of low molecular and more labile compounds change, therefore, the $\delta^{13}C$-values through time. Organic matter of marine origin will, during diagenesis, approach $\delta^{13}C$-values of the most refractive compounds, i.e. that of lignin and cellulose. $\delta^{13}C$-values are thus not only a function of the origin of organic matter, but also of its diagenetic history (Degens et al. 1968).

Although the POC/PON-ratios in the upper part of the core are > 6 and $\delta^{13}C$-val > −21‰ PDB, but significantly lower than 10 and −26‰, respectively, it is assumed that the accumulated organic material is of modern, marine origin. The increased POC/PON-ratios and $\delta^{13}C$-values in the lower part of the core seem to indicate that more nitrogen depleted, terrestrial influenced organic matter was deposited during the Late Weichselian than during the Holocene. The increased values might also reflect a higher degree of mineralization during the Late Weichselian than later on.

The two intervals of increased POC and PON accumulation (Fig. 2) are partly due to increased sedimentation rates (Fig. 3a) which favour the preservation of organic matter (Toth & Lermann 1977, Müller & Suess 1979). The supply of large fresh-water volumes and terrigenous matter when the ice was situated close to the shore line in the Younger Dryas (YD) implies high sedimentation rates and, therefore, high accumulation rates of organic matter in the lower part of the core. In addition the Skagerrak was much narrower during the Preboreal (PB) and YD than later on (Jelgersma 1979), and the sampling site was situated much closer to the former shoreline. The terrestrial influence was, therefore, more pronounced during those periods. However, most of the material deposited during these periods was inorganic since the glacially influenced and tundra-like areas close to the coring position were low-productive. The sedimentation rate below 6.5 m was almost nine times higher (about 550 cm 10^{-3} y^{-1}) than that in the upper 4 m of the core (about 60 cm 10^{-3} y^{-1}), while only six times as much POC and PON accumulated. It is concluded that the material accumulating below 7 m was more depleted for organic matter than the material accumulating above this depth.

The amount of POC which has accumulated in the sediment depends not only on the sedimentation rate but also on the primary production if allochthonous POC supply is of minor significance. Yearly estimates of the primary production of the Skagerrak are not known, but those of the nearby Kattegat (80–120 g C m^{-2} y^{-1}) are well documented (Gargas et al. 1980, Ærtebjerg-Nielsen et al. 1981). Today the coring position is situated in an area where the surface water is dominated by water from the Norwegian Coastal Current. The Baltic Current leaves the Kattegat on the east-side and follows the Norwegian coast at the northern part of the Skagerrak and the Jutland Current introduces North Sea water along the southern part of the Skagerrak. Both water masses give rise to the Norwegian Coastal Current and play a significant role for the upper part of the water column where primary production takes place (Dahl & Danielsen 1981).

The yearly primary production rates are greatly influenced by the degree of stratification and the light conditions of the water column. High stratification due to fresh-water run-off and decreased light penetration due to high supply of terrigenous material are supposed to have influenced the upper layers of the Skagerrak significantly up to 9,000 years B.P., when most of the land ice had disappeared. Comparable conditions are today, for example, found in the Arctic areas of the Atlantic Ocean. Relatively low primary production rates between 25–95 g C m^{-2} y^{-1} (mean: 57 g C m^{-2} y^{-1}; n = 7) have been found in those areas (Nemoto & Harrison 1981). Most of the fresh-water supply to the Skagerrak comes through the Baltic. The break-through of water from the Baltic Ice Lake when the ice retreated from the southern part of Sweden (about 10,000 years B.P.) and the highly brackish surface run-off from the Yoldia Sea influenced the Skagerrak surface strongly until the appearance of the Ancylus Lake and Littorina Sea (about 9,000 and 8,000 years B.P., respectively). After this period the influence of the Baltic on the Skagerrak became comparable to that of today. It is also assumed that the water leaving the Baltic up to the Yoldia Sea was low in nutrients. Suppressed primary production rates during the PB and YD (Figs. 2 and 3c) are, therefore, the necessary consequence of the hydrographic conditions during those periods. The relevance of equation (2) during the Late Weichselian, when terrestrial influenced organic matter was deposited, can be questioned. However, the ice-free areas surrounding the Skagerrak in the Late Weichselian were sparsely vegetated and mostly barren. The

TABLE 1 Mean primary production, mean sedimentation (according to Suess (1980)) and mean accumulation of organic carbon (g C m^{-2} y^{-1}) in core GIK 15530-4 during the Subatlantic, Subboreal, Atlantic, Boreal, Preboreal and Younger Dryas.

CHRONOZONES	PRIMARY PRODUCTION	SEDIMENTATION	ACCUMULATION
Subatlantic	98	13	9
Subboreal Atlantic	88	11	6
Boreal	81	10	5
Preboreal	48	7	6
Y. Dryas	32	5	35

allochthonous supply of organic matter was, therefore, small.

Indications for increased primary production rates due to the northward migration of the Polar Front (13,000–11,000 years P.B. and 10,000–9,000 years B.P.) commonly recognized in cores from the North Atlantic (Ruddiman & McIntyre 1981) were not found (Fig. 3c).

It is assumed that the high degree of stratification of the surface water of the Skagerrak (up to 9,000 years B.P.) prevented the upwelling of nutrient rich water from depth. This hydrographic phenomenon gives rise to the high production rates associated with the Polar Front, which is normally situated in the open ocean. The stratified water and the fjord-like appearance of the Skagerrak during the YD implies that the Polar Front was not present in the vicinity of the sampling site. The primary production in the area was more or less dominated by the water from the Baltic up to about 8,000 years B.P. (brackish water) although the sediments to the Skagerrak are supposed to integrate signals from several water masses entering the area, i.e. coastal water, Baltic water and Atlantic water.

The estimated recent primary production is in good agreement with measured primary production rates (^{14}C-uptake) from the Kattegat (Ærtebjerg-Nielsen et al. 1981). Table 1 shows estimates for paleo-primary production, sedimentation (Suess 1980) and accumulation of POC during different times. The mean accumulation rates of POC during the Subatlantic (SA), Subboreal (SB), Atlantic (A), Boreal (B) and Preboreal (PB) ranging from 5 to 9 g C m^{-2} y^{-1} are close to the Holocene accumulation rates found in the Baltic (Kögler & Larsen 1979) and in an open West-Norwegian fjord (Wassmann 1984).

The difference between the supply of organic matter to the sea bed and the accumulation rates indicates how much POC is mineralized by the sea-bed. In the Subatlantic (SA), Subboreal (SB), Atlantic (A) and Boreal (B) 4–5g C m^{-2} y^{-1} were mineralized. This rate is in the lower range of rates known from such depths (Pamatmat 1973, Jørgensen 1983). Compared to the estimated primary production much more POC was accumulated in PB and especially during the YD than during the SA. SB and A. It is clearly shown in Table 1 that the estimates during the last two periods do not balance. The sedimentation rate of POC must have been much higher to give rise to accumulation and benthic metabolism. This could be achieved by a primary production rate several times higher than the estimated production, which is not reasonable, or increased supply of organic matter to the sediment surface. The measurements used by Suess (1980) to calculate the average sedimentation at a given depth were mostly done in the open sea and at low latitudes (mean: 27°N; n = 26), where the pelagic food web plays a more significant role than in boreal and Arctic areas (Høpner-Petersen & Curtis 1979). It is therefore supposed that sedimentation rates based on Suess (1980) might underestimate the supply of POC to the sea-bed in boreal and especially in Arctic environments.

Conclusions

1. The organic material of the core GIK 15530-4 is of modern marine origin, but the accumulated material during the Younger Dryas seems to be more influenced by nitrogen depleted organic matter of terrestrial origin.

2. The accumulation rates of organic matter do reflect well the changes in the environmental conditions during the last 11,000 years. Marked changes in the accumulation rates between 6.5 and 7.0 m reflect clearly the environmental changes which took place between the Younger Dryas and the Preboreal.

3. The mean primary production rate of the Skagerrak at the sampling site varied between about 100 g C m^{-2} y^{-1} during the Subatlantic and 80–90 g C m^{-2} y^{-1} during the Subboreal, Atlantic and Boreal, but decreased significantly to 30–50 g C

m^{-2} y^{-1} during the Preboreal and Younger Dryas. Suppressed primary production rates during the Preboreal and Younger Dryas are supposed to be the consequence of stratified, brackish surface water low in dissolved nutrients (high fresh water supply) and poor light conditions in the upper part of the water column (high particle content).

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Evolution of the Upper Quaternary depositional environment in the Skagerrak: A synthesis

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The young Quaternary sediments covering the central and outer Skagerrak seafloor are layered and have preserved evidence for important changes of the depositional environment which has been marine for at least the past 11,000 years. In this study a summary of detailed investigations of sediment composition in one long piston core is presented. It can be shown that early on a major portion of the dominantly terrigenous fine-grained sediment originated from Fennoscandia, when the Skagerrak also received important quantities of ice-rafted material and when it was filled by brackish surface water and cold polar bottom water. Approximately 10,000 years B.P. more temperate water started to fill the Skagerrak and the influx of ice-rafted material and of Fennoscandian detrital components decreased, whereas a good portion of the sediment seemed to originate from regions south of the Skagerrak. This change is believed to be related to Atlantic water reaching the Skagerrak. Around 7,000 years B.P. the Norwegian Coastal Current regime came into existence. Since that time the depositional environment in the Skagerrak has remained temperate. Subtle changes of the pelagic and benthic Skagerrak environment did, however, also occur later on because new planktonic and benthic organisms appeared in irregular intervals, as documented, for example, by coccoliths and benthic foraminifers.

Introduction

Aim of the investigation

The Skagerrak comprises a major part of the seaway between the North and the Baltic seas, probably during the entire Quaternary. This seaway was not a permanent, but rather an intermittent feature, and the Baltic Sea was filled by marine waters only at times because of the nature of sections of this seaway and the influence of relative eustatic and isostatic sea level changes (for example the seaways across southern central Sweden or through the Danish Straits). However, except during times when the area was ice-covered, the Skagerrak should have had a marine depositional environment as part of the North Sea, or of a fjordlike depression which through the Norwegian Channel was connected to the southern Norwegian Sea.

In this paper we wish to summarize the results of a set of detailed studies which have been performed using samples from one long sediment core (Fig. 1) taken from the sediment cover of Upper Quaternary marine outer Skagerrak deposits. This core looked attractive for a detailed study because it penetrated sediments which have been deposited during the past 11,000 years after the Weichselian ice cap had retreated to the Norwegian side. The depositional history of this core does not seem to have been interrupted by hiatuses; the sediments contained a diverse group of sediment components sensitive for changes of the depositional environment, although the sediments appeared macroscopically to be very homogenous. Also, shallow seismic reflection lines document that the sampled sediment sequence is of widespread occurrence (van Weering 1982). Such core material has been studied rather frequently in the past (Lange 1956, Kihle 1971, Jørgensen et al. 1981, Fält 1982, and others), but most of the previous authors have not succeeded in establishing a well-documented chronostratigraphic framework for their interpretations. Only recently Olausson and a group of experts have described and dated in great detail some Upper Quaternary sediment sections from the Swedish west coast in an attempt to find a type locality for the Pleistocene–Holocene boundary (Olausson 1982). However, the sampling locations are located in very specialized locations which prevent a direct comparison to the open Skagerrak, and the scientific aim of this group has been oriented mainly towards stratigraphic problems, whereas it is our goal to reach an understanding of the depositional environment of the Skagerrak during the late Qua-

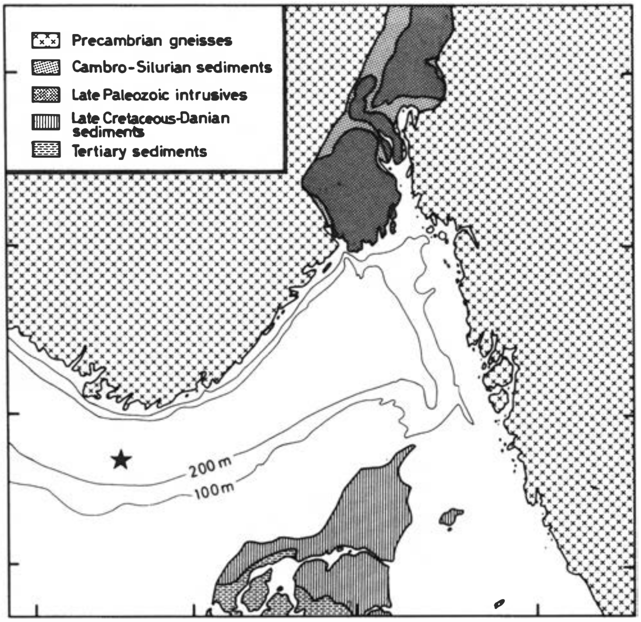

Fig. 1. Sampling location of core GIK 15530-4 in the outer Skagerrak. Bathymetric contours in the Skagerrak in metres. The geology of the surrounding land areas has been simplified and compiled from various sources.

ternary. We feel that we have an excellent and detailed time control on our core and we are therefore trying to describe this evolution in a temporal rather than in a depth-below-sediment-surface domain. The data compiled in this synthesis have therefore been plotted against an age scale rather than a depth scale as in most of the preceding studies.

Paleogeographic and -bathymetric framework

At about 11,000 years B.P., when the bottom part of the core was deposited, the water depth at the coring location was about 60 m below present sea level (Fig. 2). The Norwegian Channel, with the coring location situated at its southern flank, was a fjord-like depression connected to the southern Norwegian Sea. At that time this fjord was bordered by the ice-front to the north and by a land area to the south. This situation remained more or less unchanged until about 10,000 years B.P. (Stabell & Thiede, this volume, Fig. 1b, 1c). The sea level had risen about 20 m then, a rise of about 2 m/c (metre per century). During these first 1,000 years accumulation rates were high (Thiede c, this volume). This is explained by the paleogeographic setting, with a transport of sediments with meltwater from north and east, where the ice front was situated. A smaller amount of the sediment consists of reworked Cretaceous-Tertiary material (Mikkelsen, Henningsmoen & Høeg, both this volume), eroded from the southern flank of the fjord or from the adjacent North Sea area. During the late Younger Dryas the ice front withdrew from the northern coast, a large shallow area opened up to the east (western Sweden) and the Baltic Ice Lake drained into the Kattegat/Skagerrak area.

During the Preboreal, sea level rose about 13 m, a rise of 1.3 m/c, resulting in a transgression over the land area to the south. This caused a drastic paleogeographic change. The Norwegian Channel now represented a depression to the north in an otherwise shallow shelf sea. The accumulation rates decreased during this period, but were still high. The sediment deposited was of a more mixed origin than previously described. While there was still transport with meltwater from the north, there was also added a considerable amount of material with a southern origin. A circulation over the former land area to the south had then been established ('Jutland Current'), bringing in new material.

During the remaining part of the Holocene, the paleogeographic setting was more stable. By 8,000 years B.P. the ice had melted completely, resulting in a decrease of sediment transport from the north, the isostatic uplift decreased in

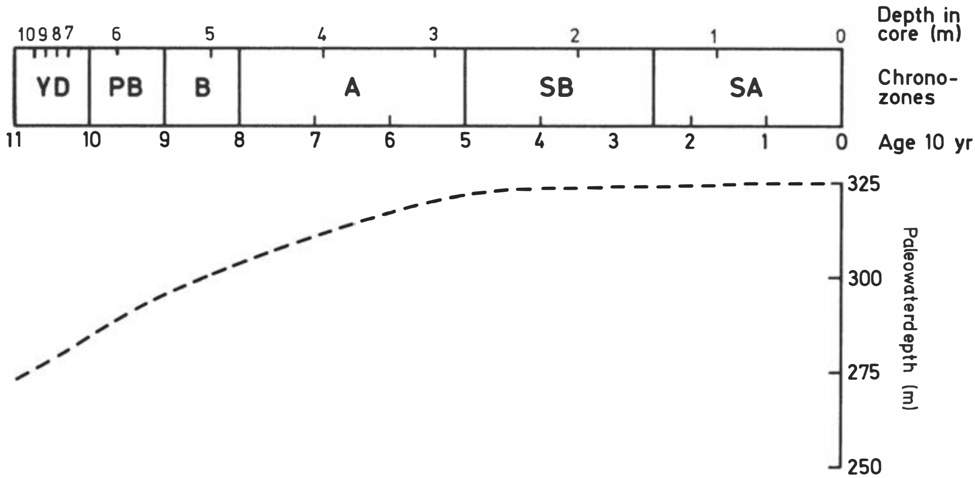

Fig. 2. Paleowater depth curve for the coring location plotted against an age scale.

speed and the coastline advanced slowly to its present position. The sea level (= eustatic rise at the coring location) rose about 20 m during Boreal and Atlantic times, a sea level rise of 0.7 m/c and 0.5 m/c respectively. The interactions between the change in speed between isostatic and eustatic rise during the period 10,000–5,000 years B.P. caused the changes in the Baltic Sea known as the Yoldia, Ancylus and Littorina stages and the Littorina (Tapes) transgression along the coasts of the Skagerrak and Kattegat. These changes do not seem to have had large effects at the coring location. During the past 5,000 years a eustatic sea level rise of maximum 5 m has occurred. The accumulation rate decreased further at the boundary Preboreal/Boreal and has remained low ever since. The clay minerals of this unit point to an origin south of Norway and Sweden.

Available data and material, approach to the problem

A description of the Quaternary depositional environment of the Skagerrak can be approached in a number of different ways. For this study we have chosen to collect data from one, carefully selected core and to study as many grain types and other parameters as possible from a limited number of samples. Because of the apparently homogenous nature of the sediments it was decided to take samples from the core at 1-m and 50-cm intervals from the > 10 m long core. Most studies have been carried out using a set of 18 samples, whereas only a few studies are documented by more densely spaced sampling intervals. It was also felt to be important that as many as possible of the variables to be measured are determined precisely in the same samples. This approach and the stratigraphy of the core resulted in a much better temporal coverage of the early part of the depositional history of the core than of the late one (Fig. 3).

Many of the stratigraphic studies in the region have been devoted to a single fossil group or selected aspects of the sedimentology of such core material. Only recently a group of colleagues studied Upper Quaternary core material from western Sweden (Olausson 1982) by collecting a comprehensive set of bio- and lithostratigraphic data from sections penetrating the Holocene-Pleistocene boundary. Our aim was to use the information contained in a diverse set of biogenic, terrigenous and authigenic particle assemblages of the core GIK 15530-4 to detect and to interpret changes of the Upper Quaternary marine depositional environment of the Skagerrak during the past 11,000 years. The data collected from this core are documented in detail in the preceding set of papers of this volume. This concluding paper, then, aims at a synthesis of the depositional environment of the Skagerrak which is a key area for understanding the Late Quaternary history of the entire NW Europe. We are discussing this environment with respect to its paleogeography and paleoclimate, the hydro-

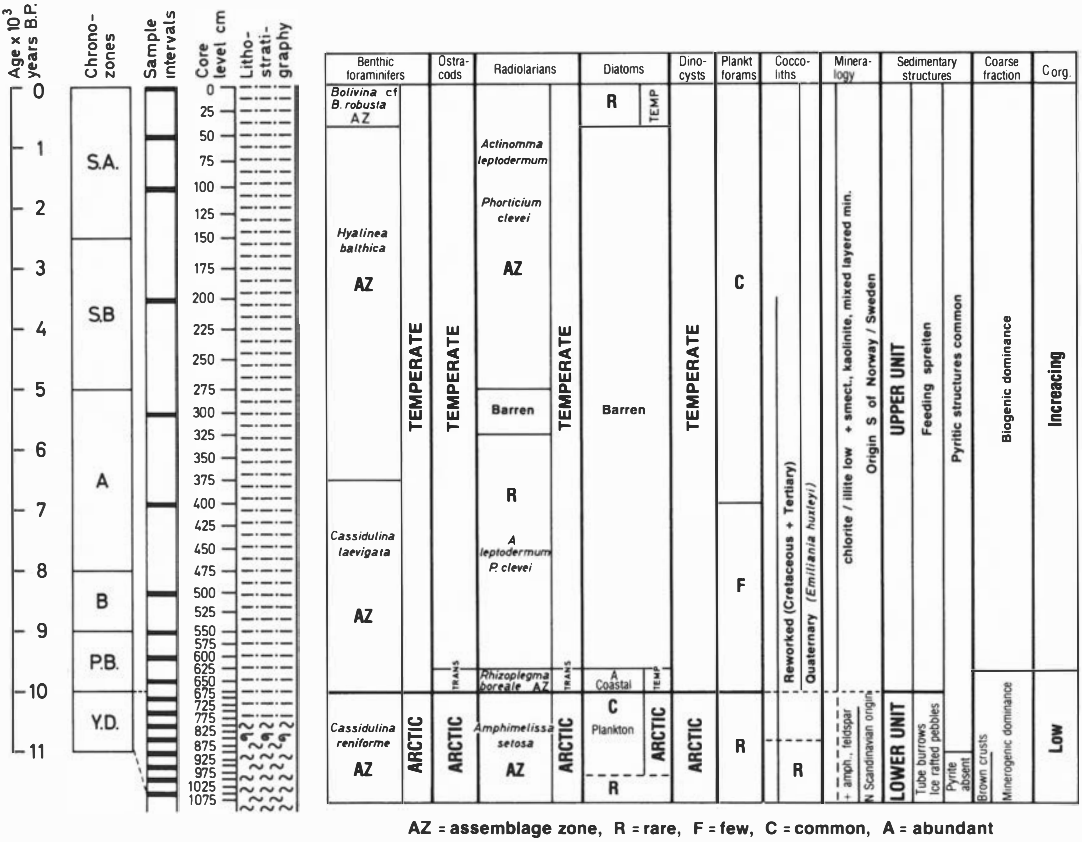

Fig. 3. Summary diagram of the stratigraphic data obtained from core GIK 15530-4. In this figure the stratigraphic data have been plotted against time rather than depth in core.

graphy of the pelagic and benthic watermasses and the source region for the dominantly terrigenous sediments contained in this Upper Quaternary sequence.

Temporal domain of studies

Core GIK 15530-4 has been dated (Fig. 3) by four different dating techniques; ^{210}Pb (Erlenkeuser, this volume), ^{14}C (Stabell, this volume), magnetostratigraphy (Schönharting, this volume) and pollen analysis (Henningsmoen & Høeg, this volume). We also have an oxygen isotope stratigraphy (Erlenkeuser, this volume) and a peak in volcanic glass (Bjørklund, this volume), which can be used to support the chronostratigraphic divisions.

The litho- and biostratigraphical investigations clearly divide the core into two main units, representing a lower cold phase and an upper warm phase. The boundary occurs between levels 650 cm and 700 cm and has tentatively been placed at 675 cm. This level is dated to 10,200 years B.P. by pollen analysis and to 9,500 years B.P. by magnetostratigraphy. We feel that this boundary represents the Younger Dryas/Preboreal (Pleistocene/Holocene) boundary. This boundary is at present defined at 10,000 ^{14}C years B.P. (Libby half-time), while waiting for an absolutely dated boundary in an approved stratotype locality (Cato et al. 1982, p. 253). A discrepancy of 200 years therefore exists between the biostratigraphically and chronostratigraphically defined boundaries in our core. The YD/PB boundary in core GIK 15530-4 (Fig. 4) is, however, based on the biostratigraphical change and *not* on the absolute chronology. It should be placed at a slightly higher level in the core if one were to follow a strict chronostratigraphical definition. Assuming a constant sedimentation rate during Younger Dryas, the bottom of the core coincides with the Allerød/Younger Dryas boundary close to 11,000 years B.P. The magnetostratigraphic datings based on comparison with the Lac de Joux record are substantially older than the datings presented in the chronostratigraphy column in Fig. 4. The magnetostratigraphic dating gives an age for the bottom of the core of about 15,000 years B.P., whereas we place it at about 11,000 years B.P. The age 11,000 years B.P. coincides with the redefined age of the bottom part of the Lac de Joux section by Mörner (pers.comm.).

The section in core GIK 15530-4 with a black sulfide zone as well as the section with scattered molluscs has been deposited during the Younger Dryas. The Younger Dryas/Preboreal boundary is not visible lithologically.

Almost 40 % of the sediment in our core was deposited during the first 1,000 years. We therefore achieved a very high stratigraphic resolution for this period. There is also a good resolution for the Preboreal, while the last 8,000 years are documented by only about 50 % of the sedimentary column of this core.

The temporal domain outlined in Fig. 4 will be used for compilation of the data presented below.

Evaluation of the paleoclimate over the past 11,000 years

Warming after the last ice age in the North Atlantic began at about 13,500 years B.P., with a reversal to cold conditions during Younger Dryas (Ruddiman et al. 1977). A similar change has not been detected in our Skagerrak core, where the Late Weichselian (Younger Dryas) is characterized by a cold water flora and fauna. During that time the Skagerrak was a fjord with the ice front along its northern and eastern boundary. It might have been very different from the open North Atlantic, as the area would have had a large input of meltwater from the retreating ice front.

The Holocene is divided into five chronozones. A drastic climatic amelioration started at the onset of the Holocene, with a Preboreal temperate climate. Summer temperatures were about the same as at present and probably little snow fell during the winters. A temperature change at the Younger Dryas/Preboreal bounda-

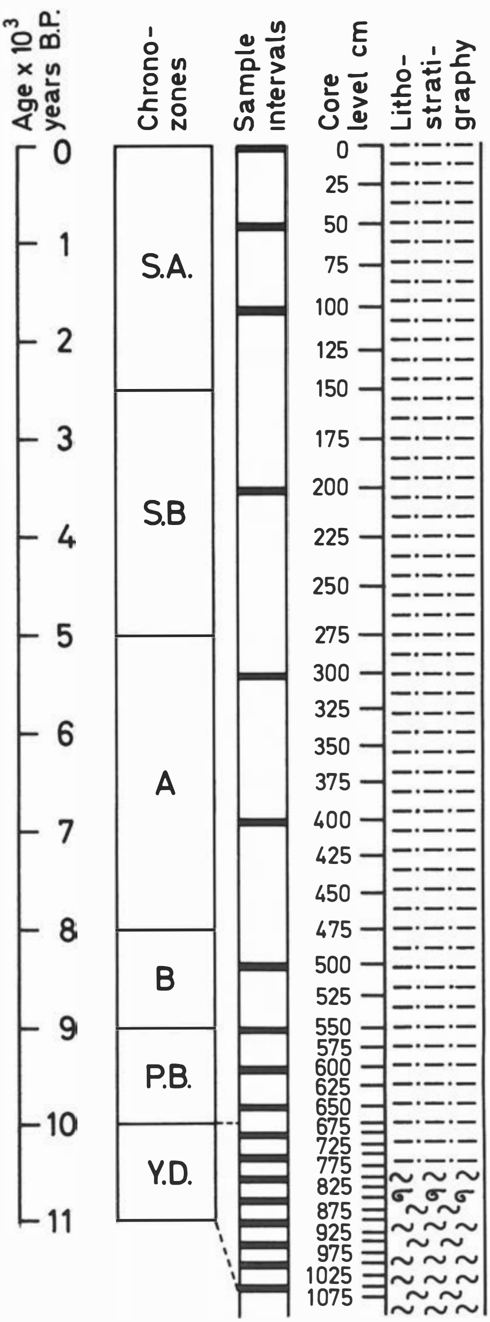

Fig. 4. Temporal domain of studies in core GIK 15530-4.

ry is also quite evident from the biostratigraphical investigations in core GIK 15530-4. The temperature continued to rise and reached a maximum during Atlantic and Subboreal times. The climate during Preboreal, Boreal and Atlantic

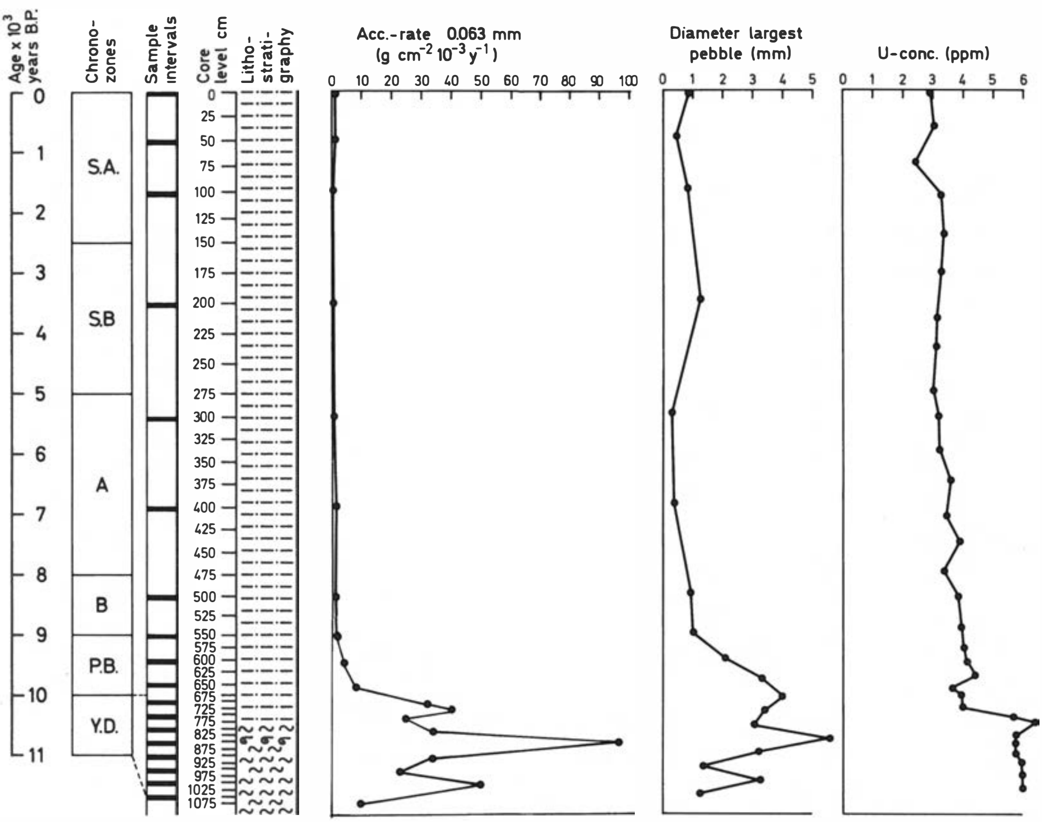

Fig. 5. Important changes of the flux of the terrigenous material to the coring site through time.

times can be characterized as coastal (warm and humid).

At the transition Atlantic/Subboreal times a clear change to a more inland-type climate (drier) took place. Finally, at the onset of the Subatlantic, a climatic deterioration to a colder and more humid climate took place. It is possible that some of the biostratigraphic changes seen in our core do reflect the climatic changes outlined during the Holocene, but so far a correlation has not been possible.

Major changes of the surface water masses (pelagic environment)

Changes of the hydrography of the surface water masses of the Skagerrak can be derived from two complete sets of data:

1) Distribution of coarse terrigenous particles which are interpreted as ice-rafted material (Fig. 5).

2) Distribution of fossils derived from phyto- and zooplankton groups which once inhabited the surface water masses and whose remains are now embedded in the Upper Quaternary sediments of the Skagerrak; we have data on pteropods, planktonic foraminifers, radiolarians, dinoflagellates, diatoms and coccoliths (Fig. 6).

The sediments of core GIK 15530-4 consist of dominantly silty and clayey material. Throughout the entire 11,000 years documented by this core they contain at least a modest amount of coarse material (with grains in general up to 0.5–1.0 mm maximum diameter). Except during an interval of max. 2,000 years in duration from Younger Dryas to Preboreal, sand-sized material

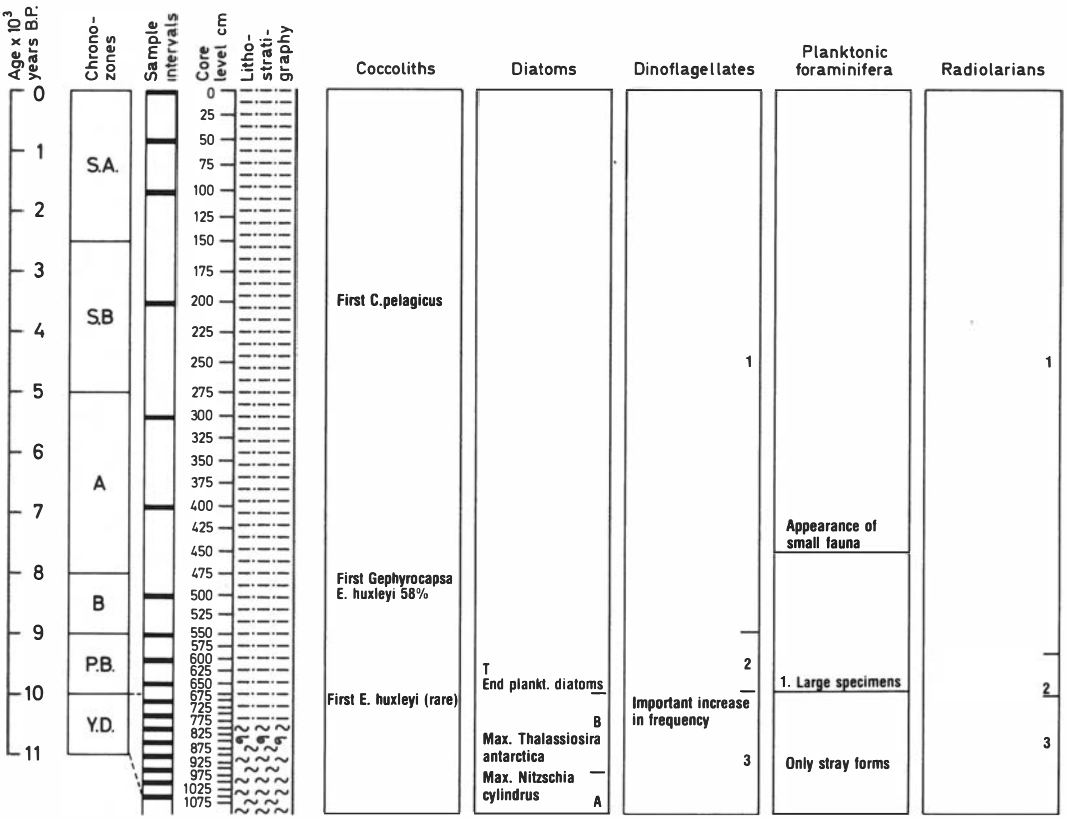

Fig. 6. Major changes of fossil assemblages produced by phyto- and zooplankton, which can be used to characterize the paleohydrography of surface water masses.

made up a very small proportion of the bulk sediment, and accumulated at very slow rates (Fig. 5). However, beginning in Younger Dryas times, with a peak in the middle Younger Dryas and then slowly dropping off to low values during late Younger Dryas and early Preboreal, relatively large amounts of coarse terrigenous components (max. > 5 mm in Younger Dryas sediments) reached the coring site. The dominant share of the coarse particles consists of fragments of crystalline and metamorphic rocks which are believed to have reached the site as ice-rafted material (dropstones because they are contained isolated in a fine grained matrix). It is not clear if they signal an intermittent or a largely perennial ice cover over the Skagerrak when the southern border of the Fennoscandian ice shield was located close to the present southern Norwegian coast line, or if they represent material carried out into the Skagerrak by icebergs. During that time the surface water must have been cold and brackish. From the distribution of the ice-rafted material in core GIK 15530-4 it is also clear that the ice-rafted material was transported to the Skagerrak in pulses of higher input rates alternating with periods of reduced input. The upper boundary of the occurrence of this material is not sharp, but rather transitional, signalling the disappearance of this cold environment over a period of about 1,000–2,000 years during late Younger Dryas and Preboral.

Significant changes of the pelagic environment which can be derived from the distribution of plankton groups, are compiled in Fig. 6. Most of the marine plankton described here have occurred more or less frequently in the Skagerrak throughout the past 11,000 years, although many of the species have changed. Pteropod shells, however, have been found only in Boreal and younger sediments, and always in small quantities.

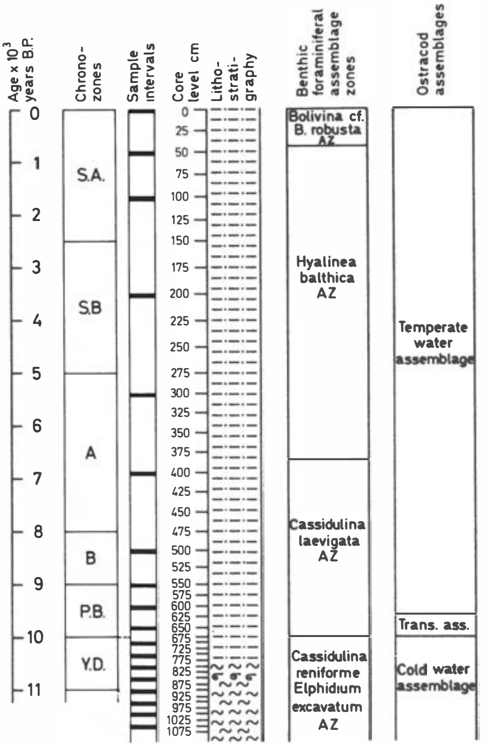

Fig. 7. Major changes of fossil assemblages of benthic foraminifers and ostracods, which can be used to characterize the paleohydrography of the bottom water masses.

Modern coccolith species have occurred in the Skagerrak only since late Younger Dryas-Preboreal time. The increase in frequency coincided with the appearance of *Emiliania huxleyi* and may mark an increase in salinity. It coincides with the first appearance of large Quaternary planktonic foraminifers and is believed to document the influx of Atlantic water of normal oceanic salinity into the Skagerrak and establishment of the Norwegian Coastal Current approximately since Preboreal times. Small specimens of *Gephyrocapsa* sp. make their first appearance in Late Preboreal times, whereas *Coccolithus pelagicus* with a modern temperature range of 8–14°C has only occurred during the past 3,000–4,000 years.

The most important changes of the diatom flora are restricted to the first 1,500 years of the history of this core. Planktonic diatoms disappeared after 9,500 years B.P. either because they were dissolved or because they did not inhabit the Skagerrak in large numbers. The planktonic diatoms of the Younger Dryas sediments of this core signal Arctic conditions (Diatom zone B), when the Norwegian Channel-Skagerrak was a fjordlike feature with a water depth of 250–300 m at the coring site.

The dinoflagellate assemblages can be subdivided into three zones. The earliest one (no. 3, Younger Dryas) is typical for cold water, ice dominated marine environments with a low dinoflagellate productivity causing low cyst frequencies in the sediment. Dinoflagellate zone 2 cysts (in Preboreal sediments) are typical in this region, where colder polar front water is mixed with warmer Atlantic water. The species of this zone seem to be extremely cosmopolitan with broad tolerances for wide ranges of temperature and salinity. Zone 1 cysts, which have occurred since Boreal times, reflect environmental conditons similar to those of today.

All radiolarians are typical for pelagic environments. Their stratigraphic occurrence has allowed us to define 3 different zones which in general coincide with those of the dinoflagellates. Radiolarian zone 3 is dominated by cold water species and restricted to Younger Dryas sediments. Radiolarian zone 2 is characterized by its high species diversity, and is of transitional nature, as both warm and cold water species are present (mainly in Preboreal sediments). Radiolarian zone 1 assemblages with a low species diversity indicate warm water conditions since late Preboreal times. They, as also diatoms, show a very poor state of preservation and signal the presence of silica-undersaturated oceanic water masses resulting in heavy opal remineralization.

Except for stray forms, Quaternary planktonic foraminifers do not appear before early Preboreal time. The large specimens in Preboreal and Boreal sediments signal the presence of Atlantic water (=Norwegian Current). Not until early Atlantic times did a small fauna appear which today lives in water masses of the Norwegian Coastal Current and which probably reflects the establishment of the modern current regime along the Norwegian continental margin.

Major changes in the bottom water masses

The changes in the hydrography of the bottom water masses have been derived from the study of two benthic animal groups, foraminifers and ostracods (Fig. 7). Benthic diatoms also occur in the upper 6.5 m of the core, but these have been transported from the surrounding shallow areas and are therefore not significant for the bottom water masses at our core site. Two major environments are indicated by the benthic foraminifers and ostracods: polar conditions during Younger Dryas, and temperate conditions during the Holocene.

The assemblages of foraminifers and ostracods found in the sediments from Younger Dryas have low diversity. The presence of *Elphidium excavatum* and low salinity tolerant ostracod species like *Cytheropteron paralatissimum* and *Normanicythere leioderma* may indicate a lowered salinity due to strong fresh water influence. The dominant benthic foraminiferal species in this zone is *Cassidulina reniforme*. This cold water species requires a normal marine salinity. The presence of *C. reniforme* indicates that the bottom water was probably little affected by the freshwater runoff.

The shift from a cold (polar) water assemblage in Younger Dryas to temperate water assemblages in Holocene is assumed to be the result of the influx of warmer, saline Atlantic water. The immigration of *Cassidulina laevigata* (which is also observed in our core) off western Norway is supposed to be connected to the influx of Atlantic water into the Norwegian Sea (Skarbø 1980), but took place about 3,000 years earlier (Mangerud 1977, Kellogg et al. 1978) than in the Skagerrak area.

The hydrography of the bottom water masses was by no means stable during the Holocene. The benthic foraminiferal assemblages, whose distribution is supposed to be controlled by mainly hydrographical properties, have changed considerably over the last 10,000 years.

Hyalinea balthica immigrated into the Skagerrak in Atlantic time. The occurrence may indicate a change in the circulation pattern or a slight increase in temperature. This species has a more southern distribution than the other species common in the Skagerrak. The opening of the English Channel (Jelgersma 1979) may have initiated both a change in the circulation pattern and an immigration route for *H. balthica*.

The immigration of *Bolivina* cf. *B. robusta* indicates that the Skagerrak reached the same hydrographical conditions as today, where *B.* cf. *B. robusta* lives under the stable deep water masses. At present we cannot explain why these changes took place, but they were perhaps initiated by the deterioration of climate before the 'Little Ice Age' and thus the formation of a stable bottom water mass in the Skagerrak.

Major sediment sources around the Skagerrak

Today the major quantity of sediments deposited in the Norwegian Channel is brought into the area with the Jutland Current, while only a small part comes from the Baltic and from Norwegian rivers and fjords. The suspended particles carried by the Jutland current partly originate from material brought into the southern North Sea from the continent by rivers, partly from erosion of the shallow areas along the Dutch, German and Danish west coasts.

The mineralogical analyses of core GIK 15530-4 have shown a mixed origin. During Holocene there is a dominance of material derived from south of Norway/Sweden. The sediments deposited during Holocene time are characterized by a low chlorite to illite ratio and a low uranium content (Rosenqvist, Bjørnstad et al., both this volume), which indicate a southern source.

In the sediments deposited during Younger Dryas there is a considerable amount of material derived from Fennoscandian Precambrian and Cambro-Silurian rocks, characterized by a high uranium content (Bjørnstad et al., this volume) and high chlorite to illite ratios (Rosenqvist, this volume). The large amount of material of northern origin indicates an increased transport with melt water in Younger Dryas.

Reworking of older sediment

The presence of pre-Quaternary and older Quaternary fossils in core GIK 15530-4 shows that older sediments have been redeposited. Reworked pollen, dinoflagellate cysts, diatoms, coccoliths, planktonic foraminifers and ostracods have been registered, but with the exception of coccoliths and dinoflagellate cysts, no quantitative work has been carried out. Reworked diatoms and ostracods are represented only by a few scattered or badly preserved specimens and are not further discussed.

Pre-Quaternary deposits

Reworked pre-Quaternary fossils have been found throughout the core, but are most frequent in the lower 4 m of the core. Coccoliths and dinoflagellate cysts of Late Cretaceous to mid-Tertiary age dominate in the sediments deposited during Younger Dryas. The pre-Quaternary coccoliths constitute the entire assemblage, while up to 45 % of the dinoflagellate cyst assemblage is composed of pre-Quaternary forms. Sediments of Late Cretaceous and Tertiary age occurs in northern and eastern Denmark, and may be the source area for the pre-Quaternary fossils. A few planktonic foraminifers of Late Cretaceous and Early Tertiary age have been found, but only in the upper 4 m of the core. The relative abundance of pre-Quaternary fossils decreases rapidly in the sediments of Holocene age.

Quaternary deposits

Also younger Quaternary deposits have contributed to the sediments in core GIK 15530-4. Although reworked Quaternary fossils are not easy to distinguish in this material, the presence of pollen from warmth-demanding plants in the lower parts of the core indicates rebedding of interglacial deposits.

Authigenic minerals

The sediments of this core contains two authigenic minerals with special significance for the depositional environment: 1) Pyrite which occurs throughout most of the core, and 2) very characteristic brown crusts which are restricted to the lowermost part of the core.

Pyritic structures begin to appear 35 cm below the core top, but remain rare in the upper 1.5 m of the core. Further down they are more common and in most cases they can be attributed to biodeformational structures, where walls have been pyritized once they reached a reducing environment due to increasing distance from the sediment surface.

Pyritic structures are lacking in the core section below 8.85 m, where characteristic brown crusts occur abundantly in the particle assemblages of the coarse fraction. Enrichments of iron oxides, although in much thinner horizons, have also been found intercalated into sediment deposited towards the end of the last Glacial in the Norwegian–Greenland Sea and in the Atlantic Ocean, but the paleoenvironmental significance is still debatable. It has been argued (Jansen et al. 1983) that their occurrence coincided with a change from high Glacial to low post-Glacial sedimentation rates. In our core such iron-oxide enrichment is also clearly found below the upper boundary of glacio-marine sediments, although sedimentation rates are higher by orders of magnitude than in the adjacent deep-sea area.

Conclusions

1. The Skagerrak retained a fjord-like shape until about 10,000 years B.P., when the land area which is today located under the North Sea was transgressed.
2. The microfossils demonstrate a paleoclimatic change from a cold water environment in the Late Weichselian to temperate water conditions in the Holocene.
3. Coarse terrigenous particles, interpreted as being ice-rafted material, reached the coring site in relatively large amounts during the first 1,000 years (Younger Dryas to Preboreal). The distribution of fossils derived from phyto- and zoo-plankton groups which once inhabited the surface water masses show a significant change interpreted as signalling the establishment of the Norwegian Coastal Current approximately since Preboreal times.
4. The changes of hydrography of the bottom water masses are indicated by benthic foraminifers and ostracods. The shift from cold to temperate water assemblages is assumed to be the result of the influx of warmer, saline Atlantic water.
5. During the Late Weichselian the source region of sediments was the Fennoscandian Precambrian and Cambro-Silurian rocks, while during the Holocene there is a dominance of material derived from south of Norway/Sweden.

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