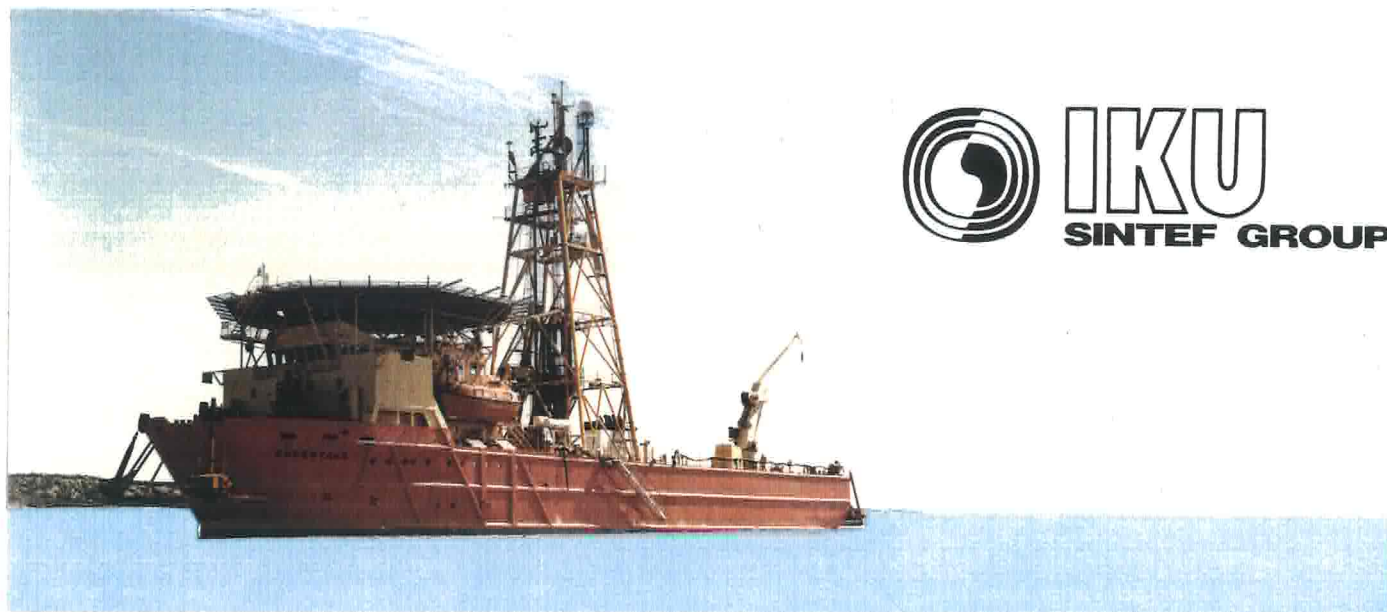
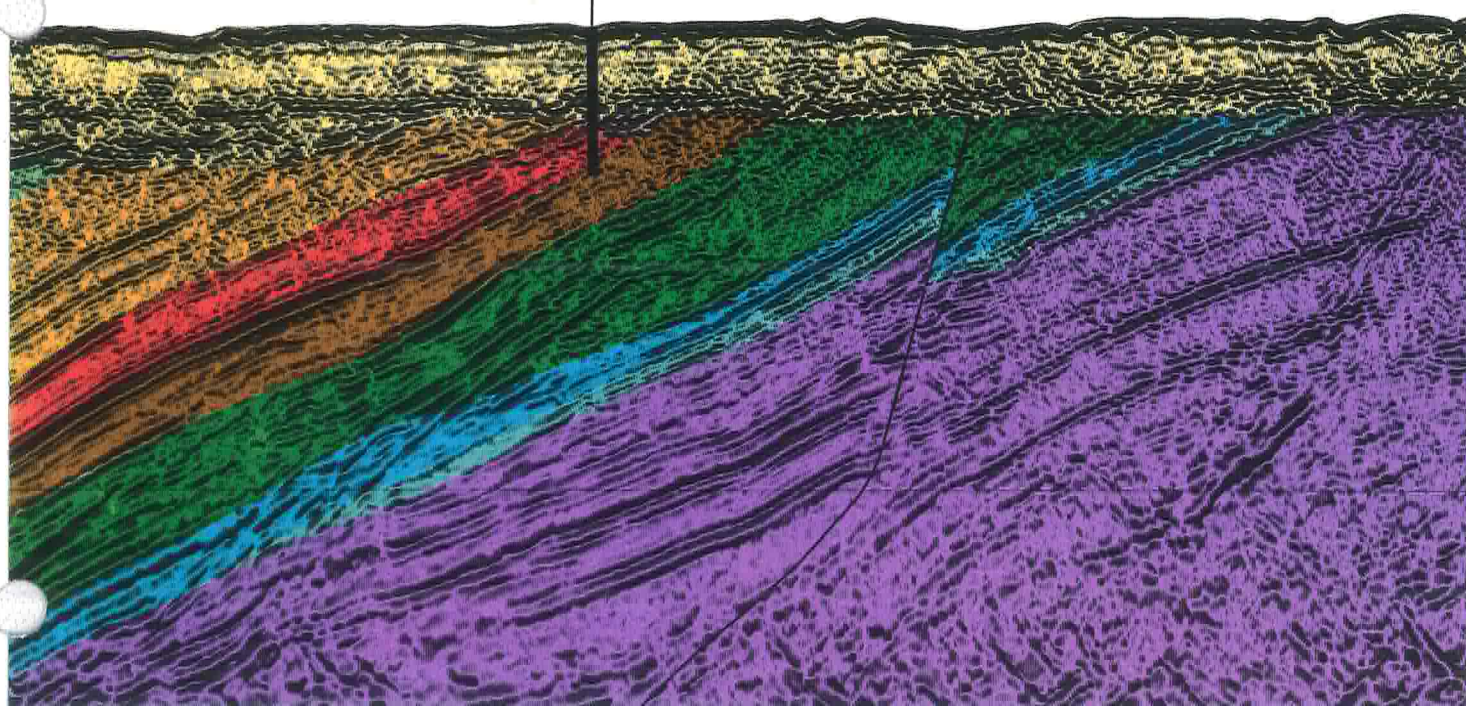




Shallow Drilling Bjørnøya West 1989

Main Report



REPORT

REG. NO.:	ACCESSIBILITY:
90.024	Restricted

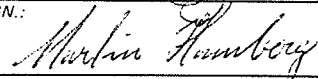
REPORT TITLE:

SHALLOW DRILLING BJØRNØYA WEST 1989
 MAIN REPORT

REPORT NO.: 21.3465.00/02/90

AUTHORS:

Tom Bugge, Stig Bakke, Geir Elvebakk, Stein Fanavoll, Robert M. Goll, Leslie Leith, Atle Mørk, Mai Britt E. Mørk, Morten Smelror, Joar Sættem, Jacob G. Verdenius

DATE: 20-FEB-90	NO. OF PAGES: 202	NO. OF APPENDICES: 15	PROJECT MANAGER: Tom Bugge	SIGN.: 
CLIENT: Amoco Norway Oil Company, A/S Norske Shell, Conoco Norway Inc., Elf Aquitaine Norge A.S., Mobil Exploration Norway Inc., Norsk Hydro a.s, Oljedirektoratet, Saga Petroleum a.s, Statoil, Total Marine Norsk A/S,			APPROVED: Martin Hamborg	SIGN.: 

SUMMARY:

The project comprises drilling of four shallow cores and acquisition of 550 km high resolution seismic in the Bjørnøya West area. The four bedrock cores are Tertiary in age and represent the pre-rift to post-rift depositional episodes. The cores extended 108-225 m below seabed. Totally 355 m of core was obtained from a cored interval of 415 m (85% core recovery). Downhole petrophysical logging was performed in all holes. The project has led to revision of the Tertiary stratigraphy and provided new information about the Tertiary uplift and erosional history of the Barents Sea. The data will also be of great help for age dating of Tertiary sediments as well as for understanding of the various erosional agents involved in the Cenozoic and volcanism.

The report has four appendix volumes following this main text volume; Volume 1: Field reports, consultant reports, core logs etc. Volume 2: Quaternary sediments and glacial geology, Volume 3: Organic geochemistry data, Volume 4: Core photos.

KEY WORDS: Shallow drilling	Tertiary
Barents Sea	Volcanics
Bjørnøya West	Erosion

CONTENTS

	Page
Preface	4
1. INTRODUCTION.....	6
2. SUMMARY AND CONCLUSIONS.....	9
3. RESULTS AND INTERPRETATION.....	18
3.1 Core 7317/10-U-01.....	18
3.1.1 Summary.....	18
3.1.2 Seismic interpretation.....	19
3.1.3 Sedimentology.....	22
3.1.4 Petrophysical logs and core measurements.....	26
3.1.5 Paleontology.....	29
3.1.6 Organic geochemistry.....	33
3.1.7 Depositional environment.....	40
3.2 Core 7316/06-U-02.....	42
3.2.1 Summary.....	42
3.2.2 Seismic interpretation.....	43
3.2.3 Sedimentology.....	45
3.2.4 Petrophysical logs and core measurements.....	49
3.2.5 Paleontology.....	51
3.2.6 Organic geochemistry.....	62
3.2.7 Depositional environment.....	69
3.3 Core 7316/06-U-01.....	71
3.3.1 Summary.....	71
3.3.2 Seismic interpretation.....	72
3.3.3 Sedimentology.....	74
3.3.4 Petrophysical logs and core measurements.....	81
3.3.5 Paleontology.....	83
3.3.6 Organic geochemistry.....	88
3.3.7 Depositional environment.....	88
3.3.8 Cretaceous rocks cored in a Quaternary glaciotectonic megablock	91
3.4 Core 7316/03-U-01.....	99
3.4.1 Summary.....	99
3.4.2 Seismic interpretation.....	101
3.4.3 Sedimentology.....	103
3.4.4 Volcanology.....	118
3.4.5 Petrophysical logs and core measurements.....	140
3.4.6 Paleontology.....	142
3.4.7 Organic geochemistry.....	148
3.4.8 Depositional environment.....	157

	Page
4. DISCUSSION.....	162
4.1 Seismic interpretation.....	162
4.2 Glacial erosion.....	174
4.2.1 The record of Cenozoic climate and glaciations.....	174
4.2.2 Observations in the present study.....	174
4.2.3 Cenozoic glacial erosion rates.....	178
4.3 Evolution of the Bjørnøya West area.....	181
4.3.1 The pre-rift phase.....	181
4.3.2 Early Eocene-Early Oligocene transpressional shear phase.....	182
4.3.3 Late Oligocene-Pliocene post-rift phase.....	183
4.4 Burial depth.....	188
5. REFERENCES.....	195

APPENDICES

In this volume

- A.1 Summary sheets of each core (scale 1:500)
- A.2 Organic geochemical summary of all core

In Appendix volume 1

- A.3 Field work - seismic survey and drilling operation
- A.4 Navigation
- A.5 Analytical methods
- A.6 Oedometer test
- A.7 Trace fossils
- A.8 Apatite fission track analysis
- A.9 Thermal conductivity
- A.10 Discussion of petrophysical core measurements
and wireline logs
- A.11.1 Range chart 7316/06-U-02
- A.11.2 Range chart 7316/03-U-01
- A.12 Core logs (scale 1:100)

In Appendix volume 2

- A.13 Quaternary sediments and glacial geology

In Appendix volume 3

- A.14 Organic geochemistry data

In Appendix volume 4

- A.15 Core photos

CORRIGENDUM

Redating of core **7316/03-U-01** by Mørk & Duncan in 1993 gave a **Late Pliocene** age (2.2 – 2.4 Ma).

Please consider this when reading the summary, the core description (Chapter 3.4), the discussion (Chapter 4) and the summary sheets (Appendices A.1.4 and A.2).

Reference:

Mørk, M.B.E. & Duncan, R.A., 1993: Late Pliocene basaltic volcanism on the western Barents Shelf margin: implications from petrology and ^{40}Ar - ^{39}Ar dating of volcanoclastic debris from a shallow drill core. Norsk Geologisk Tidsskrift 73, 209-225.

PREFACE

This project was supported by:

Amoco Norway Oil Company,
A/S Norske Shell,
Conoco Norway Inc.,
Elf Aquitaine Norge A.S.,
Mobil Exploration Norway Inc.,
Norsk Hydro a.s.,
Saga Petroleum a.s.,
Statoil,
Total Marine Norsk A/S
Norwegian Petroleum Directorate (NPD)

Their participation and cooperation throughout are gratefully acknowledged.

Tom Bugge has been project manager. Mette A. Sagmo has been project secretary and also been responsible for word processing and editing of the report. The report is based on contributions from the following persons:

Field work:

Siv Andersen, Tony Boassen, Tom Bugge, Geir Elvebakk, Stein Fanavoll, Svein Fløttum, Jørgen Hansen, Martin Hamborg, Leslie Leith, Gunn Mangerud, Rune Mogensen, Atle Mørk, Leif Rise, Unni M. Wang, Dag A. Westgaard.

Seismic interpretation:

Tom Bugge, Stein Fanavoll, Stephen J. Lippard

Sedimentology:

Geir Elvebakk (cores 7316/06-U-02 and 7317/10-U-01),
Atle Mørk, Mai Britt E. Mørk (cores 7316/03-U-01 and 7316/06-U-01)

XRD interpretation:

Reidar Bøe, Hans Rendall, Vidar Fjerdingsstad

Volcanology:

Mai Britt E. Mørk

Micropaleontology:

Robert M. Goll, Jacob G. Verdenius

Technology:

Richard G. Bromley (Ichnos, Denmark)

Palynology:

Morten Smelror, Atle Scholze (Univ. of Oslo)

Organic geochemistry:

Leslie Leith, May Britt Myhr, Liv Schou

Quaternary samples and glacial geology:

Joar Sættem, Dag Arne Westgaard, Kari Lise Ellingsen (University of Bergen), David Poole (University of Tromsø), Hans Petter Sejrup (University of Bergen), Rolf Sandveen (SINTEF Div. of Geotechnical Engineering).

Oedometer tests:

Stein O. Christensen (Sintef Div. of Geotechnical Engineering)

Laboratory work:

Wenche Andersen, Tony Boassen, Svein Fløttum, Grete Haugen, Øistein How, Inghild Kaarstad, Kristin Lind, Astrid Løvø, Hans Rendall, Øystein Rossing, Sølvi Stene, Torunn Vinge, Unni M. Wang, Knut Wicks

Computing:

Stig Bakke, Hermann M. Weiss, Dag A. Westgaard.

Drawing:

Rolf Angell-Petersen, Inger L. S. Berg, Ingrid Brandslet, Berit Fossum, Kåre Holst, Jan H. Johansen

Word processing:

Mette A. Sagmo, Ellen M. Solberg, Ann Merete Espås

Printing:

Geir Herje

Special gratitude is directed to the crew on board the seismic vessel M/S "Geo Surveyor" and the drilling vessel M/S "Bucentaur" and to the personnel from British Plaster Board (BPB) Ltd., Fugro, Geoteam and Dia-Team for excellent work during the drilling operation. Stephen Lippard is thanked for valuable comments of the manuscript.

1. INTRODUCTION

This report gives the results from bedrock coring at four sites in the Bjørnøya West area in the Barents Sea and is the main report in the project "Shallow drilling Bjørnøya West 1989".

Previously an interim data report has been submitted. Since then two of the cores (7316/03-U-01 and 7316/06-U-01) have been re-dated to significantly younger ages. Documentation and discussion of this and all information in the data report are given in the present Main report.

There is a relatively dense grid of conventional seismic data in the area, and the project was planned on the basis of a selection of these lines. The closest exploratory wells through Tertiary sediments are on the Senja Ridge, about 250 km further south, but are hard to tie into the Bjørnøya West area. The only bedrock core available in the area was the 1985 IKU shallow core from 7317/02-U-01 (D85-V) of early-middle Eocene age. The age estimates that could be applied to the seismic interpretation were therefore uncertain, and a main objective of the project was to date sequences and major sequence boundaries in addition to check potential source-rock intervals.

The first part of the field work was a shallow seismic survey carried out in May/June 1989 (Figure 1.1). It comprised 550 km of digital seismic (120 channels, 1500 m streamer, 12.5 m shot point interval, 2.5 s recording, 2 ms sampling interval). Analog mini-airgun data with single channel recording were collected simultaneously with the digital data on most of the lines. Side-scan sonar and deep towed boomer were also used at the same time. The drilling operation lasted for about 24 days in the period 4 - 28 August. More details of the field work are given in Appendix A.3.

The four cores were drilled in loose and poorly consolidated sediments which caused stability problems for some of the boreholes. It also hampered the petrophysical logging program, and only one hole (7317/10-U-01) could be logged without the drill string acting as casing.

Hammer samples of the Quaternary overburden were collected at all sites and continuous diamond coring was done in some intervals at sites 7316/03-U-01 and 7317/10-U-01. Some extra sampling in the latter borehole was done as part of a separate IKU research project. A huge block of glacially transported Cretaceous rock was found within the Quaternary sediments at

site 7316/06-U-01. This is described in Section 3.3.8, while the Quaternary sediment samples in general are described in Appendix Volume 2.

The present report is organized in the same way as our recent shallow drilling reports, with the data presentation core by core in Chapter 3, starting with a short summary of each core and followed by a discussion of the local tie to the seismic data, core description, datings, organic geochemistry results and a description of the depositional environment.

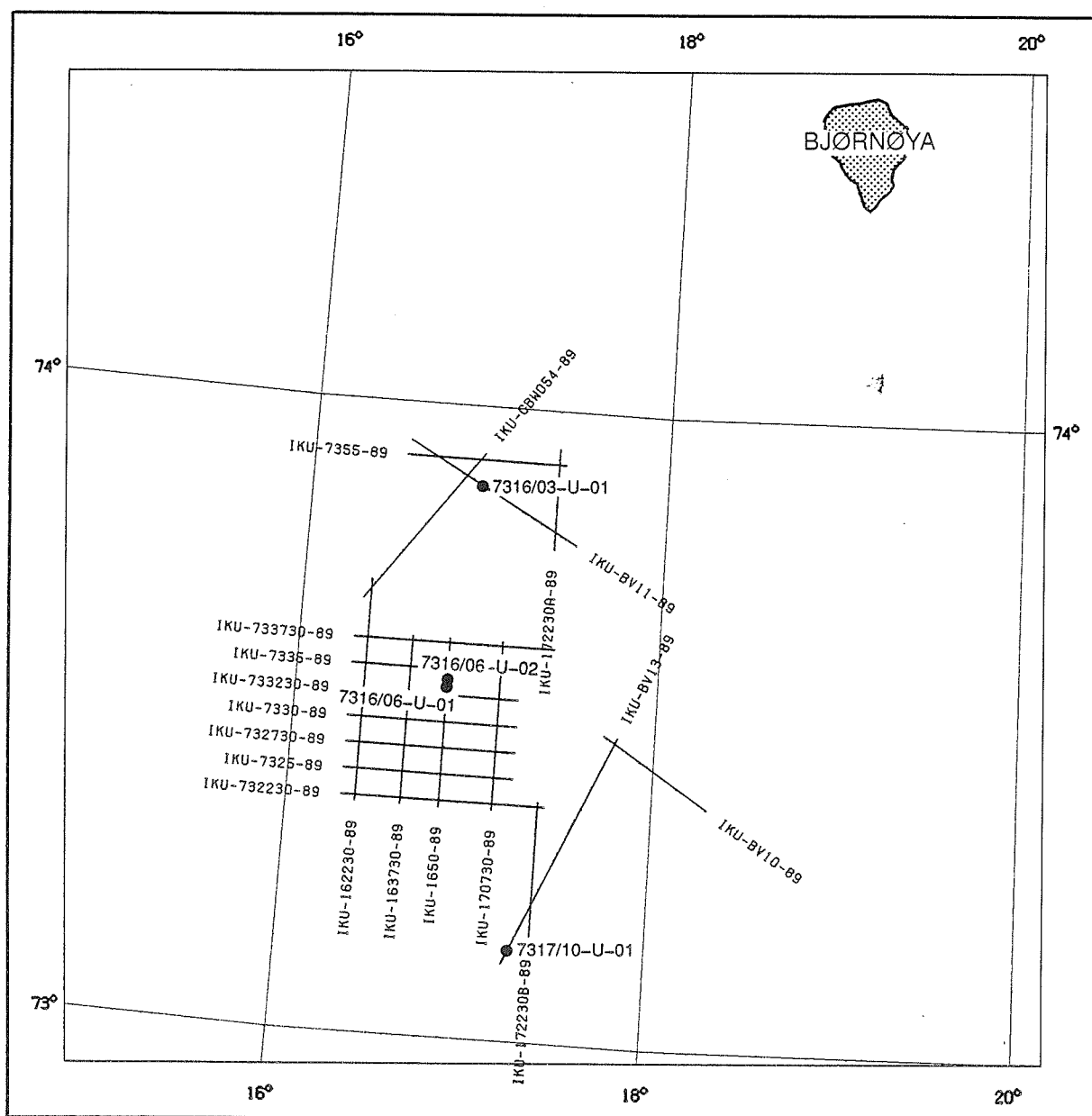


Figure 1.1 Four shallow cores were drilled and 550 km of shallow seismic data were acquired in the Bjørnøya West area.

In Chapter 4 we present a seismic interpretation of the area, with subcrop map and seismic sections showing tentative correlations of the drilled seismic units. The geological evolution of the area is discussed and revised models are suggested. Summary discussions of the glacial geology and erosion of the area are also given.

Background data and much of the analytical results are presented in appendices, organised in this way:

Appendices in the main volume:

- Summary sheets of each core (scale 1:500)
- Organic geochemical summary of all cores

Appendix volume 1:

- Field work - seismic survey and drilling operation
- Navigation
- Analytical methods
- Oedometer test
- Trace fossils
- Apatite fission track analysis
- Thermal conductivity
- Petrophysical core measurements and wireline logs
- Range charts cores 7316/06-U-02 and 7316/03-U-01
- Core logs scale 1:100

Appendix volume 2:

- Quaternary sediments and glacial geology

Appendix volume 3:

- Organic geochemistry data

Appendix volume 4:

- Core photos

2. SUMMARY AND CONCLUSIONS

During ten days in May/June 1989 about 550 km of high resolution shallow seismic data were shot in the Bjørnøya West area (Figure 1.1). In August four bedrock cores were drilled to depths of between 43 m and 160 m into bedrock (Tables 2.1 and 2.2). Two of the cores were drilled approximately to the limit of 200 m below seabed, while one was drilled on dispensation to 225 m. This drill site had 143 m of Quaternary overburden, which is the thickest overburden penetrated by shallow drilling so far.

Table 2.1 Key data of the core holes.

Core no	Seismic line	Shotpoint	Water depth	Overburden (Top of core)	Penetration in bedrock	Core length	Core recovery	Tot. penetr. below seabed (Base of core)
7316/03-U-01	IKU-BV11-89	~1077	297 m	31.9 m	160.8 m	137.7 m	85.6%	192.7 m
7316/06-U-01	IKU-1650-89	~1688	328 m	65.0 m	43.0 m	9.8 m	22.7%	108.0 m
7316/06-U-02	IKU-1650-89	~1814	421 m	68.5 m	131.0 m	128.8 m	98.4%	199.5 m
7317/10-U-01	IKU-BV13-89	~3344	465 m	143.2 m	81.4 m	78.4 m	96.3%	224.6 m
			Av. 378 m	308.6 m	416.2 m	354.7 m	85.2%	Av. 181.2 m
			Av. 77.1 m Av. 104.0 m					

Table 2.2 Final positions (ED50 and UTM zone 34 with central meridian 21°E).

Core no.	Latitude	Longitude	Easting	Northing
7316/03-U-01	73° 52' 55.8"N	16° 54' 28.9"E	373285	8203472
7316/06-U-01	73° 33' 16.5"N	16° 49' 59.6"E	368422	8167166
7316/06-U-02	73° 34' 7.7"N	16° 50' 0.9"E	368544	8168746
7317/10-U-01	73° 8' 58.5"N	17° 16' 13.0"E	379389	8121151

The four cores were all drilled southwest of the island Bjørnøya on the Tertiary clastic wedge that has been built out along the western margin of the Barents Sea. Very little exact age information was available from this area and it was a general opinion that the main part of the wedge was of Eocene age and that younger clastic sediments made only a minor part below the cover of glacial sediments deposited during Plio-Pleistocene glaciations in the last 2.5 million years.

Dating of the cores shows, however, that a significant part of the wedge is Miocene and younger and indicates that there probably is a major hiatus from late Eocene/earliest Oligocene to early Miocene represented by an angular unconformity between cores 7316/06-U-01 and -02 and below core 7316/03-U-01 (Figures 2.1 and 2.2). The latter core has been dated both by isotope analysis of volcanic clasts and by fossils. It consists dominantly of turbidites and debris flow deposits. Much of the Miocene and possibly Pliocene parts of the wedge represents reworked Eocene sediments derived from the outer part of the paleoshelf. A short summary of each core is presented below.

1

Core 7317/10-U-01 penetrated 81.4 m of siltstones and silty shales deposited in Late Paleocene to ?Early Eocene. The lower 75 m of the core (Subunit A) comprise a major coarsening upward succession interrupted by thinner upward coarsening successions, possibly deposited in an inner, storm-influenced shelf during Late Paleocene time. The unconformably overlying Subunit B has an abrupt transition to slightly coarser sediments containing very thin beds of polymict conglomerates. The transition correlates to a possible seismic sequence boundary characterized by down-lap. The siltstones of Subunit B were deposited in inner, storm-influenced shelf environment. The sediments of the core are characterized by abundant Late Cretaceous fossils, probably sourced from subsequent uplift and erosion of the Stappen High.

Core 7316/06-U-02 penetrated 131 m of silty shales, siltstones and very fine-grained sandstones of Early and Middle Eocene age. The sediments form an overall but composite coarsening upward grain size pattern, reflecting deposition during progradation of an inner, storm-influenced shelf. The core contains abundant reworked Late Cretaceous fossils possibly sourced from subsequent erosion of the Stappen High.

Core 7316/06-U-01 penetrated 43 m of sediment deposited in Late Oligocene - Early Miocene. The sediments contain reworked fossils indicating that

Cretaceous and Eocene sediments sourced the area. The lower part of the core (Unit A) consists of immature, very fine-grained sandstone, possibly deposited in shelf break or slope environment by mass flow mechanisms. Large parts of the cored interval were not sampled due to drilling problems, but in the upper part a polymict conglomerate is overlain by a sandstone consisting of well rounded grains. On the seismic record a channel is correlated to this level of the core, and the polymict conglomerate may represent lag deposits at the channel bottom, while the channel may have been filled by the well rounded sand originated from a higher energy source area.

Core 7316/03-U-01 penetrated 160.8 m of bedrock of Late Miocene age. The core comprises a major coarsening upward sequence interrupted by shaly intervals. Abundant fossils of Cretaceous and Eocene age indicate that the sediments have undergone extensive reworking and redeposition. The lower 39 m thick Subunit A consists of very fine-grained glauconitic, immature sandstone probably deposited in channels and by turbidity currents on a continental slope. The succeeding interval (Subunit B) is 60 m thick and consists of similar sediments, but with several beds of mixed volcanic debris and sandstone interpreted to have been deposited by gravity flows.

A silty shale interval (Subunits C and E) is 18 m thick but is interrupted by a 4 m thick interval of fining upward volcanic debris/sandstone beds (Subunit D). Abundant siliceous fossils in Subunits C and E indicate reworking of Eocene siliceous ooze sediment. Units C, D and E may all represent gravity flow deposited sediment, although hemipelagic sedimentation of the shales cannot be ruled out. The volcanic debris, however, show fining upward and chaotic textures indicative of debris flows.

The upper part of the core (Subunit F) is 40 m thick and consists of fine to medium grained sandstone. Abundant beds with volcanic debris occur in the lower part of the interval and are present throughout the subunit although in reduced number. This part of the core was probably deposited by both debris flows/turbidity currents and as channel fill, but in shallower conditions than the underlying sediments.

The volcanics consist of basaltic glass (ash, lapilli) and a few blocks of more massive olivine basalt, all within a narrow compositional range (sub-alkaline to transitional alkaline).



Figure 2.1 Summary figure of the cores. See also Figure 2.2.

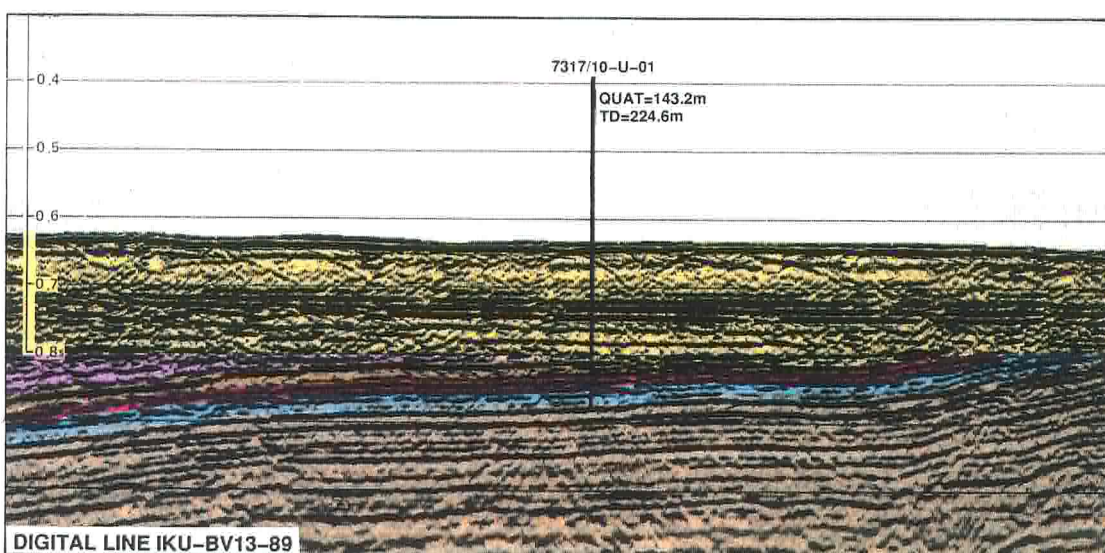
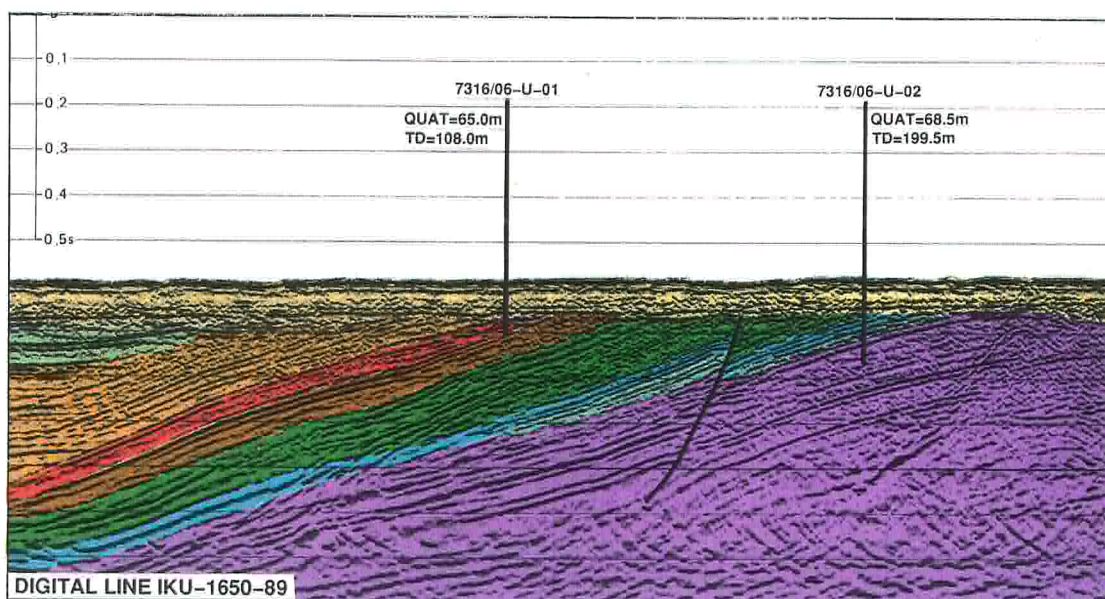
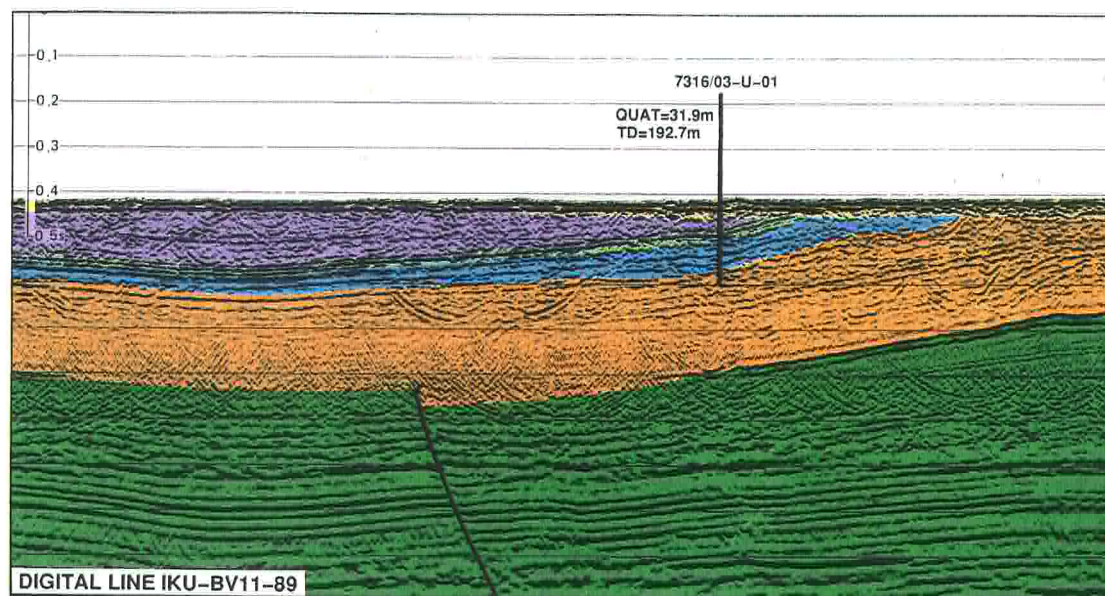


Figure 2.2 Shallow seismic sections of the coring sites. Coloured units correspond to Figure 2.1.

Dating by the K-Ar method has given results between 10 Ma and 2.5 Ma, without any systematic variation through the core. Nd- and Sr-isotopic compositions of the basalts indicate that the magma originated in an off ridge setting and interacted with continental crust. The eruption took place in the vicinity of the sedimentary basin.

The majority of the organic material in the four cores is characterized by poor to fair amounts of terrestrially-influenced, largely inert kerogen of insignificant hydrocarbon generation potential. An exception to this occurs in Subunit E of core 7316/03-U-01, where a thin siltstone interval contains significant amounts of reasonably well-preserved terrestrial organic matter. If laterally and vertically developed, this interval could have some potential for gas/condensate generation with increased thermal maturity. All of the samples examined were found to contain thermally immature kerogen.

Based on the information obtained from this project, we believe that the evolutionary history of this area could be improved and also partly revised. In short, the evolution of this part of the wedge could be as follows:

In late Cretaceous time the western Barents Sea including the Stappen High area (Figure 2.3) was in general a shelf area with marine deposition. The Bjørnøya and Sørvestsnaget basins were the main depocentres. In the late Cretaceous/early Tertiary an uplift started as a forerunner of the later rifting in the Norwegian Sea. This led to an erosion and deposition on the slope evolving in the west. The Stappen High experienced its first phase of uplift. The main sediment source was rocks of Cretaceous age, now recognized by the content of reworked Cretaceous fossils in the Eocene and Miocene sediments.

As the rifting started and the entire Barents Shelf was uplifted, mainly in Eocene time, large volumes of sediments were eroded and deposited on the western slope. The Stappen High became a high and was exposed to erosion. This situation continued through the Eocene and possibly into earliest Oligocene. Earliest Oligocene to Miocene sediments are probably lacking and the time period is only represented by the above mentioned angular unconformity (green in Figure 2.2).

From other data in the Barents Sea it is supposed that a second episode of uplift and erosion occurred in the late Tertiary (Miocene). The same is

also observed off Mid Norway, where the same hiatus occurred on the Vøring Plateau.

Because of sparse core data and very poor dating, a parallel history was not until recently suggested for the Barents Sea wedge. The Early/Middle Eocene dating of core 7316/06-U-02 and Miocene dating with reworked earliest Oligocene fossils of core 7316/06-U-01 1,5 km away, and the unconformity observed between the cores, are evidences of a similar history. Core 7316/03-U-01 about 30 km farther to the north was dated as Late Miocene. A major unconformity observed 0.2 s below the base of the core is suggested to represent the same hiatus.

The sediments in the two Miocene cores seem to have been derived from Eocene strata, and we find the most probable source area to be the Eocene sediments already deposited on the wedge. Volcanic clasts found as sorted layers in core 7316/03-U-01 are believed to come from volcanoes that erupted through a continental crust and clastic sediments, partly in a subaerial environment and partly in shallow water conditions. The isotope dates of the clasts (2.5 - 10 Ma) coincide with the biostratigraphical dating of the core. We interpret the core to represent rapid deposition partly by turbidity currents/debris flows or other types of sediment gravity flows. The mass flows were probably triggered by earthquakes related to volcanic eruptions and/or seismic activity on the faults along the western margin of the Stappen High.

Paleomagnetic analysis and amino acid chronology of the oldest Quaternary units shows that at least 800 m of sediments were deposited on the wedge during the last approximately 700 000 years. The Quaternary sediments are found on top of the Tertiary sediments separated by an erosional unconformity. From the ODP Leg 104 on the Vøring Plateau it is known that major glaciers there reached the shelf about 2.5 Ma. Minor input of glacial drop in the sediments shows that some glaciers may have reached the coasts of the Norwegian-Greenland Sea as early as 5.5 Ma. In land areas bordering the Baffin Bay glaciations started about 8 million years ago.

Extrapolated to the Barents Shelf this could infer that elevated areas such as Bjørnøya and the Stappen High could have been glaciated at least 5 million years ago, probably even earlier. This could explain part of the rapid erosion and high sediment supply out to the shelf edge in

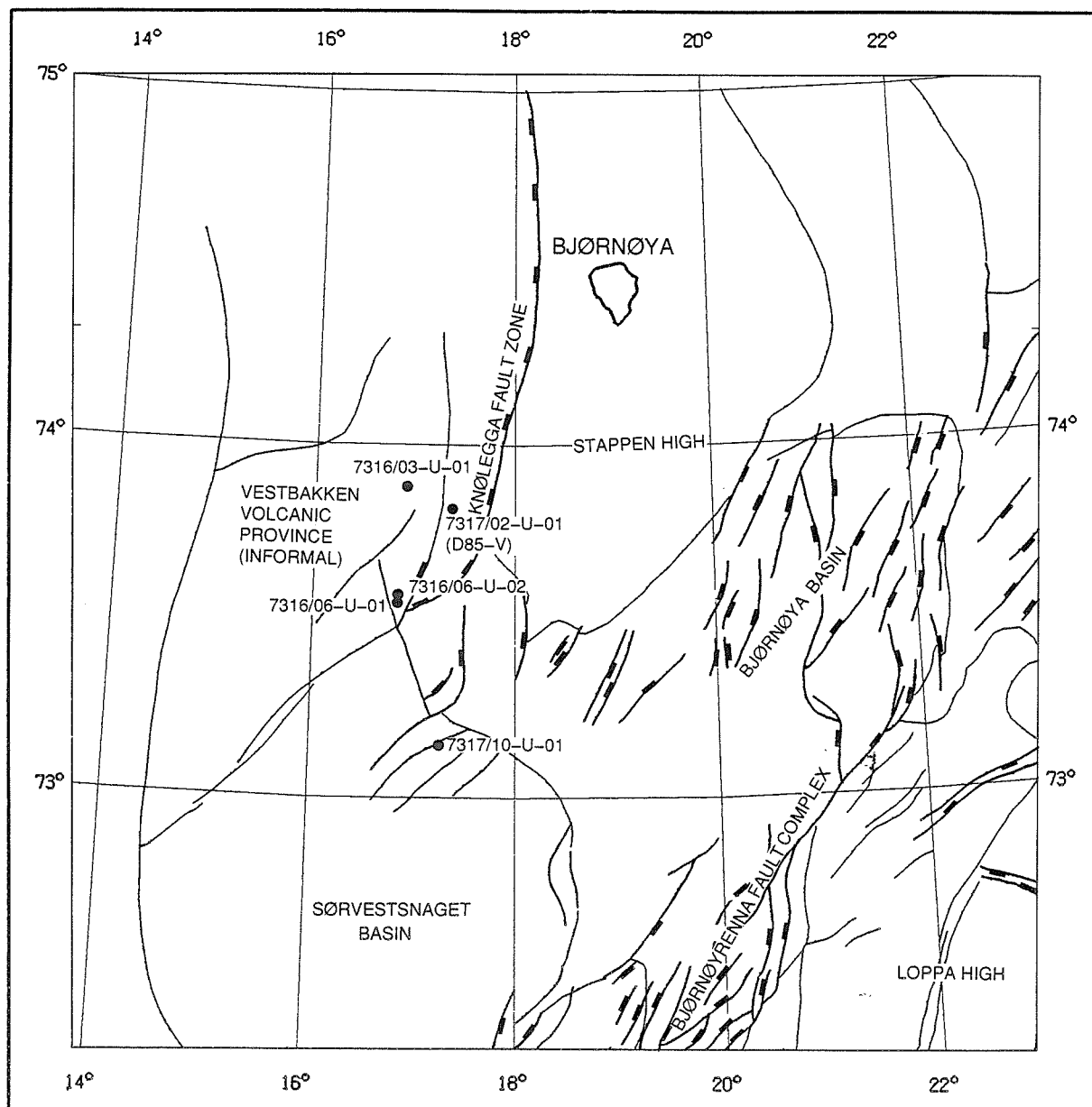


Figure 2.3 Main structural elements (from Gabrielsen et al. in press.).

Miocene/Pliocene time. It also could imply that glaciation of the shelf started early and left only a time period of the order of 5 million years for deposition of the Miocene/?Pliocene part of the wedge, below the glacial sediments.

Essential conclusions are:

- A major part of the erosion in the Barents Sea and deposition on the clastic wedge occurred in the Miocene/Pliocene.

- Glacial erosion was a significant part of the total Cenozoic erosion in the Barents Sea and may have started as early as 5-8 million years ago.
- A probably significant uplift occurred in the Miocene, although heavy glacial erosion can occur without any uplift at all. Glacial erosion over a wide area is, however, accompanied by uplift due to isostatic rebound.

Other points to remember for future exploration as well as scientific research, are:

- Extremely rapid reworking of Tertiary sediments, as observed for instance in core 7316/03-U-01, can easily lead to wrong age dating, because the content of fossils representing the correct age can be only a few percent. When dealing only with cuttings and sidewall cores, as is often the case in an exploration well, age dating should therefore be handled with the utmost care.
- Isotopic dating of volcanic clasts has proved a potential for supplementing biostratigraphic dating and may be combined with biostratigraphy to identify reworking of fossils.
- This study has also revealed the existence of until now unknown basalt volcanism in Late Miocene time to the east of the Eocene Vestbakken volcanic province, involving continental crust.
- The existence of the large glaciotectonic megablock comprising Cretaceous rocks found within the Quaternary sediments of core 7316/06-U-01, also shows the dangers of misinterpreting of cutting and sidewall core samples. The Cretaceous section contained pure Cretaceous rocks with no intermixing of younger fossils. Samples from below such a slab containing mixture of fossil assemblages, could there be interpreted as having contamination from caving, sampling procedures etc. Our recent studies show that the occurrence of such allochthonous rock bodies could be relatively frequent.

3. RESULTS AND CONCLUSIONS

3.1 Core 7317/10-U-01

3.1.1 Summary

Core 7317/10-U-01 was drilled in the southern part of the survey area and was aimed at penetrating sediments characterized by bands of seismic reflections. The core penetrated 81.4 of bedrock (Figure 3.1.1).

Based on lithology and seismic data the core has been subdivided in two subunits, A and B, with the transition at 149.4 m below seabed. Subunit A consists of coarsening upward medium dark grey silty shales to siltstones with thin very fine-grained sandstone beds. The subunit was deposited in regressive shelf environment in Late Paleocene time. A minor unconformity, separating subunits A and B, is indicated both from seismic data (onlapping reflectors) and sedimentological data. The beds of Subunit B were deposited in a shallow shelf environment in Late Paleocene to ?Early Eocene time.

The core showed little variation in organic geochemistry. The rocks contain fair amounts of thermally immature, terrestrially-dominated, type III/IV kerogen. This material has no significant hydrocarbon generation potential.

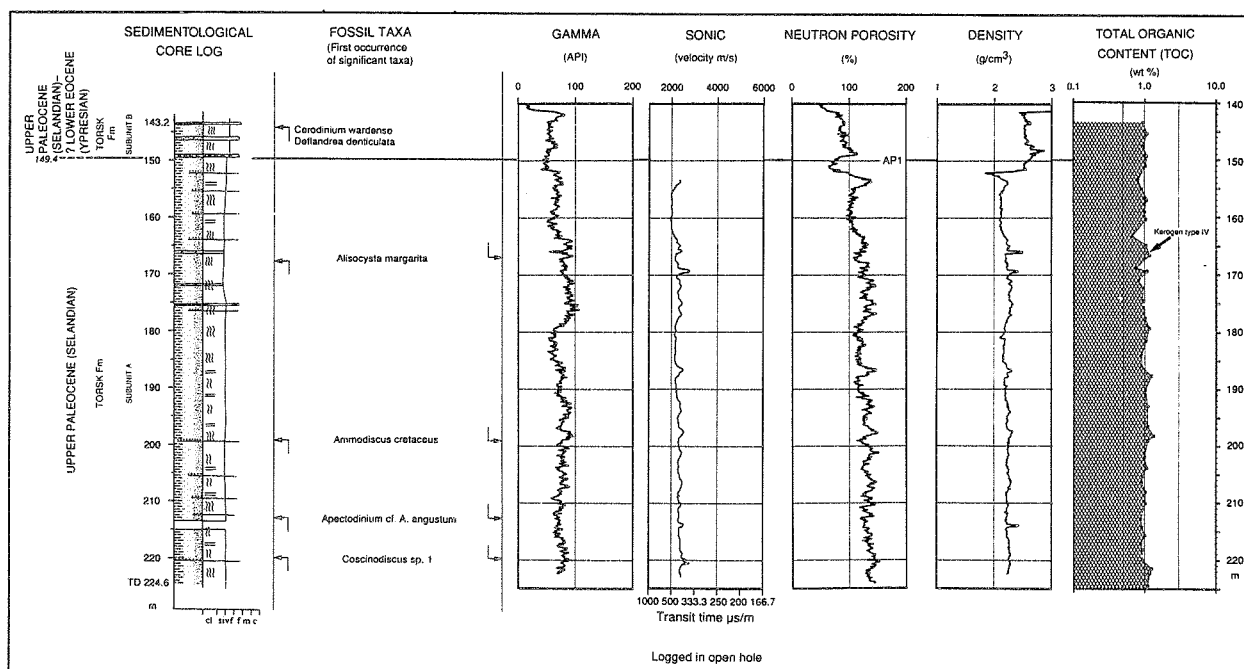


Figure 3.1.1 Summary of main data from core 7317/10-U-01.

Trace amounts of asphaltene-rich bitumen were found in some samples which appears to share a common source dominated by terrestrial organic matter. An indigenous source for this material is considered unlikely.

3.1.2 Seismic interpretation

Core 7317/10-U-01 was drilled on seismic line IKU-BV13-89, sp 3344. The overburden was 143.2 m thick.

The objective for this core was to penetrate a sediment sequence characterized by a band of seismic reflections. In the seismic data the top of this band seems to subcrop about 12 shotpoints (150 m) to the southwest of the drill site but, due to limited vertical resolution in the seismic data, the actual subcrop point is probably slightly to the northeast of the drill site, and the top of the band was penetrated at 149.4 m (4 - 5 m below rockhead). Nearly 40 m were cored below the reflection band (Figure 3.1.2, Table 3.1.1).

Table 3.1.1 Correlation of seismic data and core 7317/10-U-01. Interval sound velocities noted "seismic" appear from this seismic-core correlation.

Depth below seabed (m) (msec)	Core unit	Thickness of seismic unit (m) (msec)	Interval sound velocity (m/sec)		
			Seismic	Core	Log
143.2 159 166 179 186 197 198 207 220 226 TD224.6 230	Quaternary	143.2 159	1800	var. 1675-1860	
	Unit B/Unit A	22.8 20	2280	1970	2125
	Unit A	20 18	2220	1900	2270
	—	12 10	2400	2020	2320
	"	22 19	2300	2100	2370
	"	4.6 4	est2300	2210	2370
Water depth: 465 m (635 msec) Depth below sea level: 689.6 m (865 msec)					

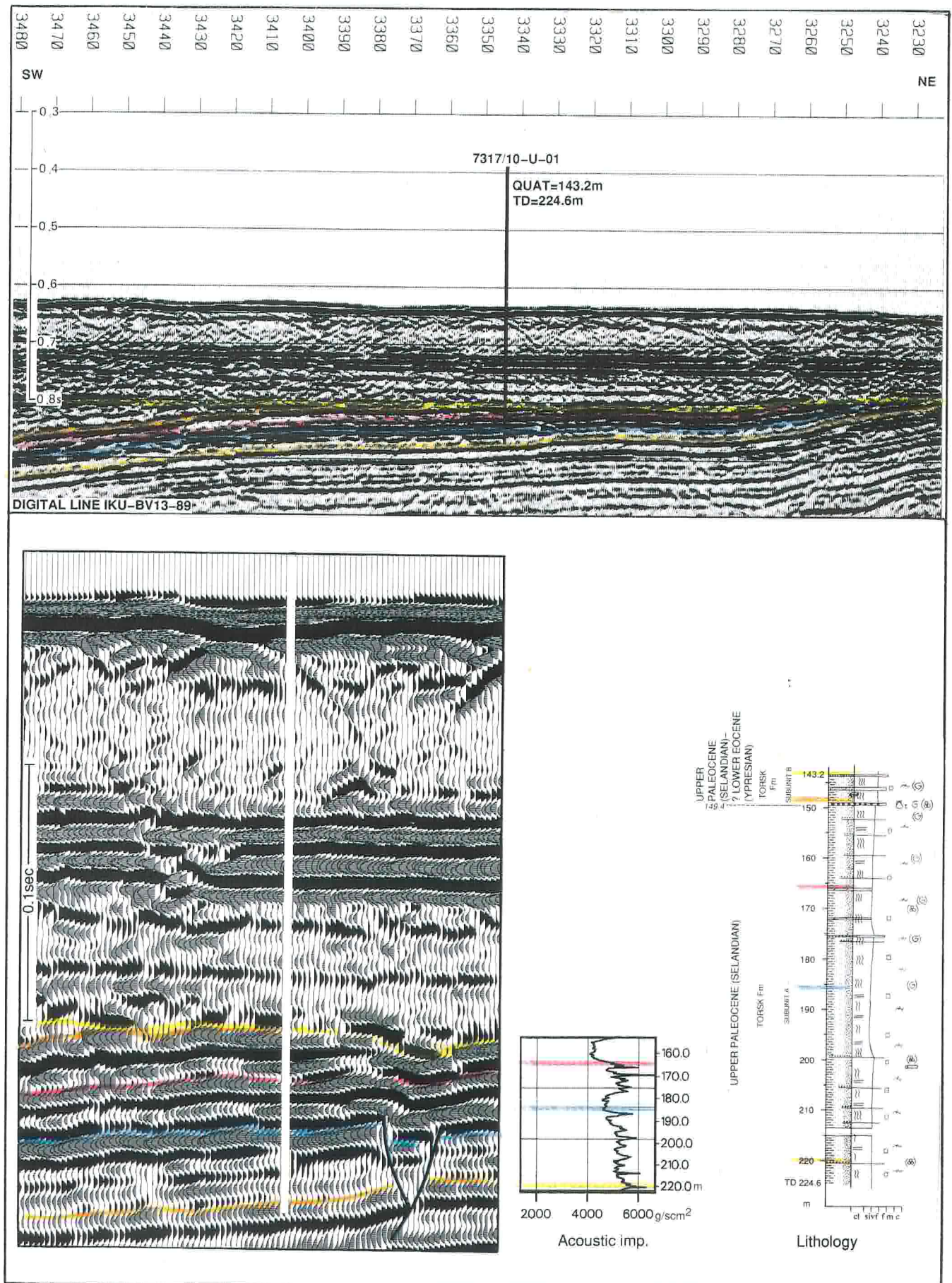


Figure 3.1.2 Seismic correlation of core 7317/10-U-01.

The lowermost about 5 m (TD - 220 m) of the core were from a homogeneous looking, 20 - 25 m thick unit between two relatively strong and continuous reflections. The succeeding interval 220 - 186 m is also below the above mentioned reflection band. Two weak reflections are correlated to possible flooding surfaces seen in the core at about 208 m and 198 m. In the neutron porosity log (Figure 3.1.1, Appendix A.1) this corresponds to three intervals with upward increasing porosity. The reflection correlated to the 186 m level in the core, is strong and continuous. The three downhole logs (density, porosity and sonic) all show positive peaks at this level.

The interval 186 - 149.4 m comprises the mentioned reflection band. A semi-continuous reflection is correlated to a siderite layer at 166 m. This reflection seems to fade out downdip together with a thinning of the reflection band. As suggested above, the uppermost 4 - 5 m of the core probably belong to the sediments above the reflection band.

3.1.3 Sedimentology

224.6 (TD) - 143.2 m: Torsk Formation, Late Paleocene - ?Early Eocene

A total of 81.4 m of bedrock has been cored between 143.2 m and 224.6 m below sea floor.

The entire core consists of shales and siltstones, and has been subdivided in Subunit A below 149.4 m and Subunit B above (Figure 3.1.3). The abrupt transition takes place below a polymict conglomerate bed.

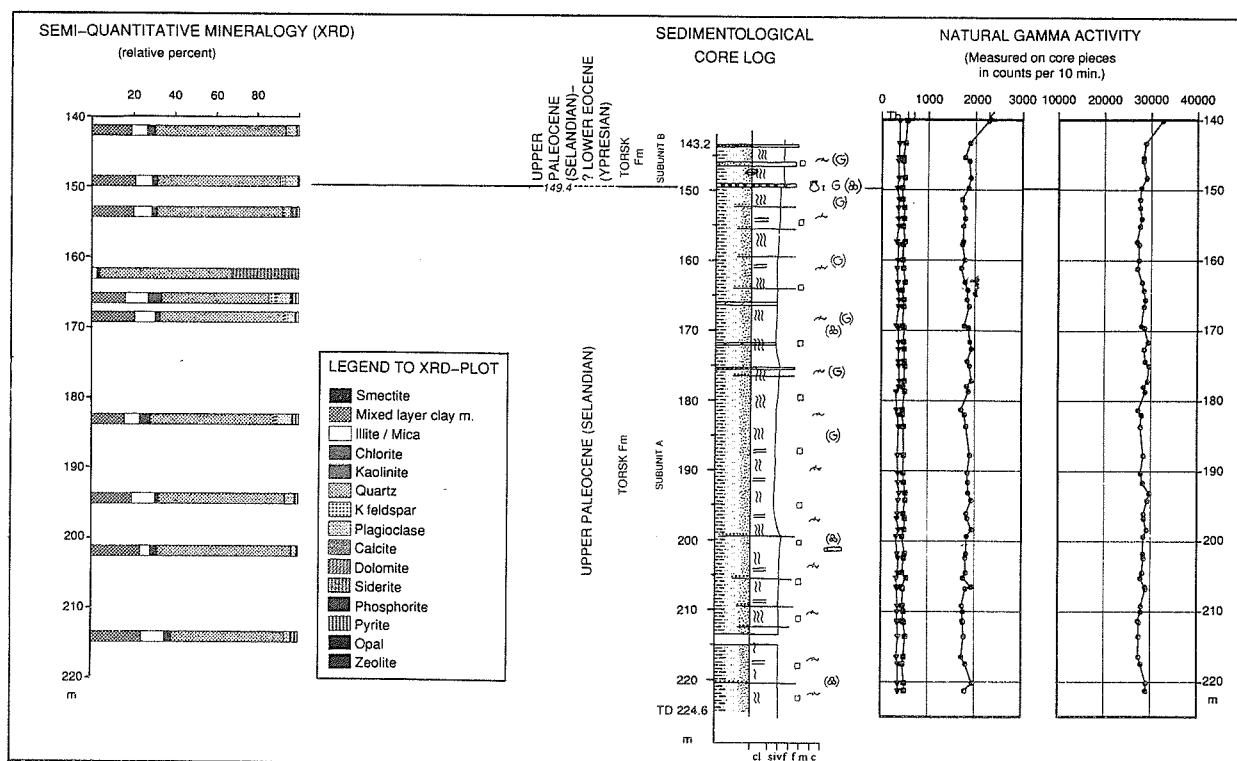


Figure 3.1.3 Lithological column, XRD mineralogical analysis and natural gamma radiation of core 7317/10-U-01.

Subunit A (224.6 - 149.4 m) contains three minor coarsening upward intervals (gradual transitions at about 200 m and 166 m) together forming an overall coarsening upward stacking pattern. The coarsening upward intervals are also detected on the neutron porosity log (Figure 3.1.1).

The lithology of Subunit A consists of clayey siltstones to siltstones grading upward into sandy-clayey siltstones. The coarsening upward patterns

are accompanied by an increase in number and thickness of very thin and fine-grained sandstone and siltstone beds. In the lower part of two lower coarsening upward intervals (below 166 m) the beds are of mm scale but thicker beds (cm scale) are occasionally observed. The beds are fining upward and have sharp bases. Most beds are massive, however, horizontal laminations and/or low-angle ripple laminations are present, overlying massive parts of beds or may occur in individual beds.

A few brownish coloured siderite concretions/beds are observed in the siltstones. The siderite may contain minor inclusions of pyrite. Pyrite is also occasionally present as cement partly filling some of the burrows.

The intensity of bioturbation varies, but is in general incomplete. The intensity follows the grain size showing in general little bio-reworking in the clayey siltstones and increasing in the sandy clayey siltstones. Here, the basal part of thicker sandstone beds are the only primary sedimentary features left.

Small calcareous fossil fragments are sporadically encountered; some representing bivalve and foraminifera fragments. In addition minor plant debris are observed.

Subunit B (149.4 – 143.2 m). The transition between subunits A and B is schematically illustrated in Appendix A.12. An oblique surface with a dip of about 20 degrees cuts the horizontal laminae of the underlying Subunit A. Burrows with a diameter of about 0.5 cm are present along the oblique surface. A few centimetres thick, bioturbated, very fine-grained sandstone bed lies concordantly above the oblique surface and fill the burrows along the surface. This sandstone bed is discordantly overlain by moderately bioturbated sandy and pebbly siltstones with sandstone beds. The bedding in the sediment in Subunit B is concordant with the bedding in Subunit A. An about 3 cm thick, polymict conglomerate bed is present about 50 cm above the subunit transition. The conglomerate bed contains pebbles in a sandy matrix consisting of clastic grains, glauconite grains and *Inoceramus* fragments. The pebbles consist of greenish coloured glaucony, dark grey to black chert fragments containing fragments of sponge spicules, quartzite pebbles, medium dark grey silty shales, and dark grey to black phosphorite pebbles. The pebbles are up to 1.5 cm in diameter and are in general subrounded.

The medium dark grey sandy-clayey siltstone of Subunit B contains a few 5 cm thick fine to medium grained sandstone beds containing black, 0.5 cm large pebbles and grey shale clasts. The beds have sharp basal boundaries overlain by massive and horizontally laminated sandstones. One large burrow (3 cm in diameter) is filled with a sandy matrix with pebbles.

The sediments in this unit are moderately bioturbated. A few calcareous fossil fragments have been found. One early diagenetic (pre-compactional) calcite concretion is observed.

Petrography

Seven thin sections and ten X-ray diffraction samples have been studied from core 7317/10-U-01.

Table 3.1.2 and Figure 3.1.3 show a semi-quantitative mineralogical composition of the shales and siltstones based on XRD analysis. Quartz contributes with 50-65% of the rocks; highest in the more silty and sandy deposits. Clay minerals include illite and chlorite, and mixed layer clay minerals contribute with a few to almost 40% of the rocks, with the highest content in the finest grained rock. In addition, alkali feldspars, plagioclase, siderite and pyrite are among the other minerals present in small amounts. One sample of a concretion contains 32% siderite.

Thin sections have been studied from clayey sandy siltstones, a siderite concretion (172.37 m) and the conglomerate bed at 148.95 m.

Thin sections from 175.91 m, 163.65 m and 222.20 m were from matrix-supported clayey sandy siltstones. A faint bedding consisting of variation in sand/silt content vs. clay, was present. Burrows are common and show a slight enrichment of authigenic pyrite. Other places burrows were filled by grain-supported very fine-grained sandstones. Sand grains consist mostly of monocrystalline quartz grains, often showing undulatory extinction. In addition chert, alkali feldspars, plagioclase, muscovite, biotite, siderite and glauconite pelloids are present in small amounts. Brown to black plant debris were also present.

The concretion bed at 172.37 m contains abundant siderite pelloids cemented by microcrystalline siderite cement. The bedding is deformed around the concretions. None of the pelloids show any sign of squeezing, indicating

pre-compactional diagenesis. Inclusions of clastic grains and glauconite pelloids are present within the concretion. The concretions have been formed during early (pre-compactional) diagenesis.

The conglomerate bed at 148.95 m contains subangular to subrounded pebbles less than 1.5 cm in diameter in a sandy matrix. The clasts consist of glauconite, shale, phosphorite and chert. The chert contains sponge spicules and minor glauconite, which is characteristic, although not unique, for the Upper Permian Kapp Starostin Formation from Spitsbergen and for the Middle Carboniferous Efuglvika Member of the Kapp Kåre Formation from Bjørnøya. In addition, quartzite grains with interlobate grain boundaries are present.

The groundmass of the conglomerates is dominated by dull to bright green glaucony. In addition, some clastic grains are present. The glaucony is represented by pelloids as well as zoned, spherical to elliptically shaped and more irregular shaped masses of glaucony. Growth fractures are common in pelloidal glaucony. Bored prisms of the bivalve *Inoceramus* are also common. In addition, a few foraminifer fragments are present.

Table 3.1.2 *Semi-quantitative mineral composition of the sediments based on X-ray diffraction analysis of core 7317/10-U-01.*

Depth	Mix	Mi/Il	Chl	Qtz	KFsp	Plag	Sid	Pyr
152.02	18.8	7.3	4.2	63.0	0.6	4.8	0.7	0.8
159.20	20.7	7.7	2.8	59.6	2.1	6.2	0.9	0.0
163.66	19.9	8.5	2.7	60.8	0.7	3.2	3.2	1.1
172.36	0.0	2.2	1.3	63.6	0.0	0.6	32.3	0.0
175.81	16.1	10.7	6.7	52.0	3.6	6.8	0.9	3.2
178.51	20.6	9.5	2.7	60.0	1.9	3.4	0.5	1.4
193.13	15.6	7.1	5.4	59.7	1.4	7.3	0.9	2.7
204.46	19.4	10.9	1.9	61.1	0.2	4.4	0.0	2.0
211.90	23.5	4.5	3.6	65.0	0.2	2.3	0.7	0.2
224.10	24.1	10.9	3.6	54.6	0.1	2.9	0.6	3.2

3.1.4 Petrophysical logs and core measurements

Downhole logs

Gamma, sonic, density, caliper and neutron logs were run in open hole from TD to 153 m and inside the drill string from 153 m to the seabed. The caliper log shows no cavings in the wellbore, which means that the quality of the logs is good. The gamma log values between 153 m and 143.2 m are recalculated to open hole values based on a regression analysis from a previous shallow drilling survey. Due to very low variance in the gamma log values it was impossible to calculate any depth shift correction (see Appendix A.10). The logs are shown in Figure 3.1.1 and in Figures 3.1.3 and 3.1.4 together with the core measurements. Petrophysical properties calculated from the logs are shown on Figure 3.1.5. Logs and petrophysical properties are also shown together in the summary sheet in Appendix A.1.

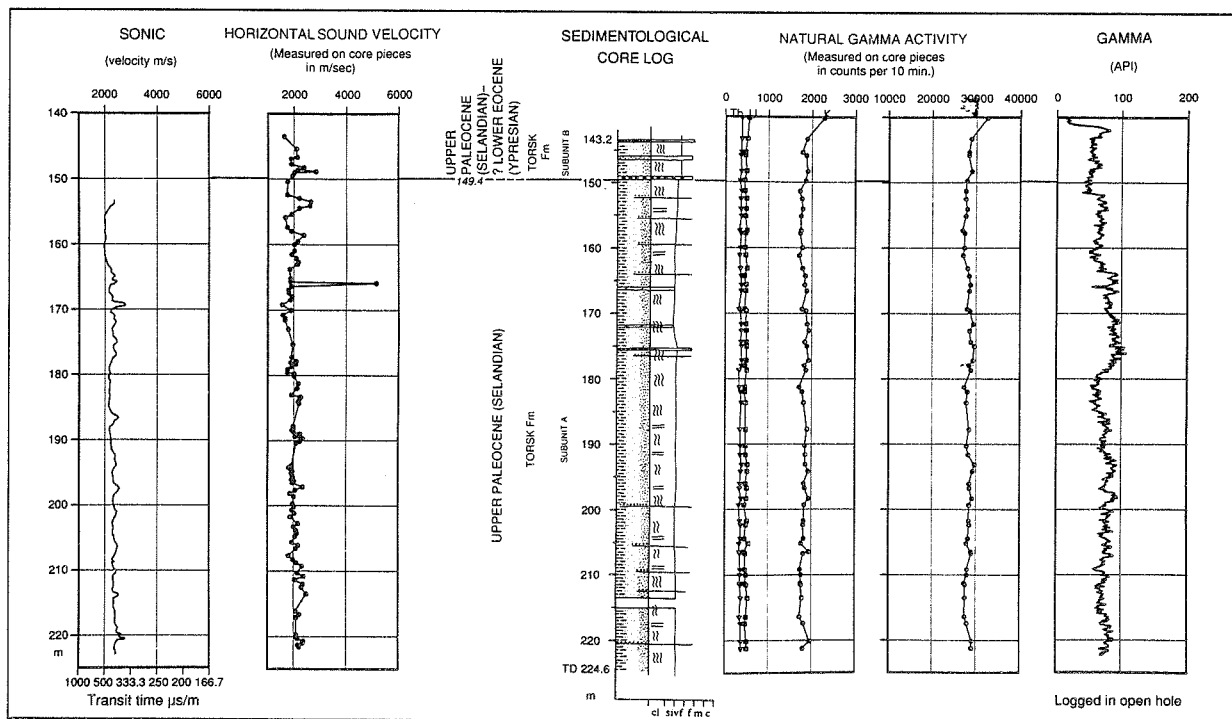


Figure 3.1.4 Lithology log and laboratory measurements of spectral gamma and horizontal sound velocity plotted together with downhole gamma and sonic logs of core 7317/10-U-01.

224.6 (TD) - 149.4 m: Subunit A

The gamma log in this subunit reflects the four stacked coarsening upward sequences with transition zones at about 208 m, 175 m and 165.8 m which

appears as upward decreasing trends. The relatively high gamma values between 60 and 100 API are caused by clay contents, K-feldspar contents up to 10% and mica contents between 5% and 10%. The interval between 177 m and 163 m shows the highest gamma values.

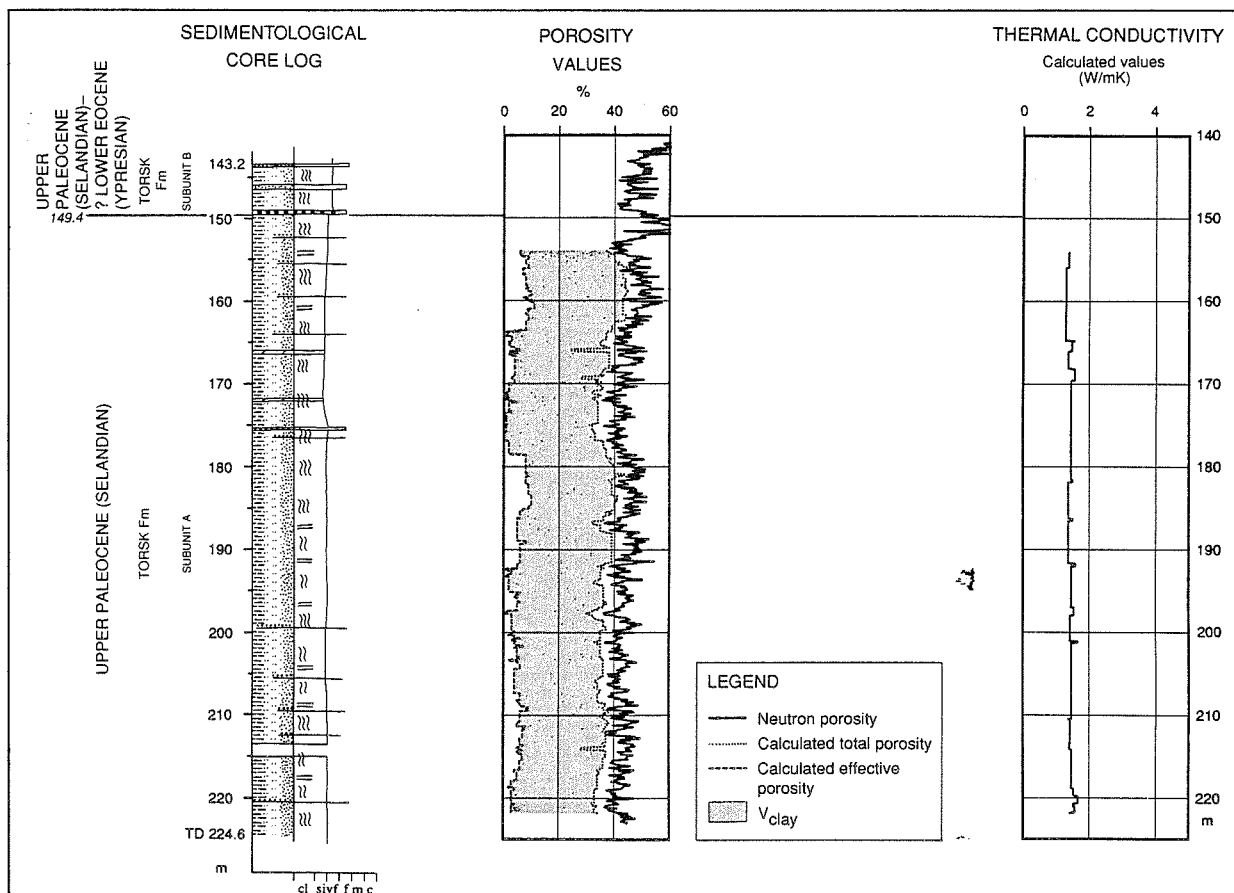


Figure 3.1.5 Lithology log, neutron porosity with calculated total porosity, effective porosity and V_{shale} , and calculated thermal conductivity of core 7317/10-U-01.

The sonic log shows relatively stable velocities around 2100 m/s with a slightly increasing trend downwards probably due to compaction, and small peaks at cemented zones. The density log shows bulk density values around 2.2 g/cm³ with small peaks at cemented zones. The neutron porosity log shows values just above 40%, total porosity just below 40% and effective porosity values between 0% and 10% due to high V_{clay} values. The above mentioned coarsening upward trends are expressed clearer on the neutron porosity log than on the gamma log. There seems to be six coarsening upward sequences with transitions at 208 m, 198 m, 186 m, 175 m and 165.8 m.

149.4 - 143.2 m: Subunit B

The recalculated gamma log shows no break into Subunit B. The gamma log values vary between 70 and 100 API with a small peak at 147.5 m. The attenuation of the HR string disturbs the other logs in this subunit.

Sound velocity and gamma measurements on core

Figure 3.1.4 shows the sonic and gamma wireline logs together with the horizontal sound velocity and gamma radiation values measured on the core. The trends of the gamma laboratory measurements correlate visually well with the downhole gamma log values. The K-plot on the spectral gamma laboratory measurements shows that K is the main source for the gamma log response. The U and Th values vary only very little from the background level.

Also the trend and peaks of the horizontal sound velocity values measured on the core correlate visually well with the log derived sound velocity values. The small differences in velocity levels are most probably due to velocity anisotropy, stress release in the cores and different measuring tools, see Appendix A.10.

Thermal conductivity

The calculated thermal conductivity values in Figure 3.1.5 are between 1.3 and 1.6 W/mK. See Appendix A.9.

3.1.5 Paleontology

Core 7317/10-U-01 contains common, but moderately to poorly preserved marine microfloras. The diversity of dinoflagellate cysts is generally low, most common forms being representatives of the genera *Cerodinium* and *Cometodinium*. The stratigraphic distribution of dinoflagellate cysts in the core is shown in Figures 3.1.6 and 3.1.7. The occurrence of *Cerodinium wardenense* and *Deflandrea denticulata* through the upper part of the core suggests an age not younger than Early Eocene. The presence of *Alisocysta margarita* at 168.97 m suggests a Late Paleocene age at this level. The very rare arenaceous foraminifera that occur at two levels in Subunit A are characteristic for the Paleocene. The common occurrence of the diatom *Coscinodiscus* sp. 1 near to the base of the core is evidence of a Late Paleocene age.

Based on the combined palynological and micropaleontological evidence, a general Late Paleocene to earliest Eocene age is concluded for core 7317/10-U-01.

Common reworked Cretaceous (Aptian - Cenomanian?) dinoflagellate cysts occur through the core.

224.6 (TD) - 149.4 m: Subunit A, Upper Paleocene (Selandian)

Subunit A contains low diversity dinoflagellate cyst assemblages, but with relatively common *Cerodinium* spp. in some samples. Although the sediments contain characteristic Lower Paleogene marine microfloras, few good marker fossils are recorded; the exceptions being *Alisocysta margarita* at 168.97 m and possibly *Apectodinium* sp. cf. *A. angustum* at 214.10 m. The latter species is possibly an indication of a pre-Eocene age. According to Powell (1988) *Apectodinium angustum* ranges to the top of the Upper Eocene series in the Central North Sea Area.

A modified dinoflagellate cyst biozonation for the uppermost Paleocene and lowermost Eocene sediments from the Central North Sea has recently been published by Powell (1988). Powell used the first downhole appearance (i.e. the extinction) of *Alisocysta margarita* to define the top of his *Alisocysta margarita*

Interval Biozone. The zone was indirectly calibrated to the calcareous nannoplankton biozones NP8 (pars) to NP7? (pars?). According to Powell (1988), *A. margarita* has its earliest occurrence in the underlying *Palaeoperidinium pyrophorum* Interval Biozone, which was indirectly calibrated with the NP biozones 7? (pars.) to 4. If the range of *Alisocysta margarita* is the same in the Central North Sea area and the western Barents Sea area, the presence of this species at 168.97 m in core 7317/10-U-01 indicates that Subunit A is of Late Paleocene age.

The presence of *Deflandrea oebisfeldensis* at 161.28 m and the earliest appearance of *Deflandrea denticulata* at 199.03 m are also of biostratigraphic significance. These species are generally restricted to the Upper Paleocene - Lower Eocene (Williams & Bujak 1985). The co-occurrence of *Deflandrea oebisfeldensis* and *D. denticulata* and the two species mentioned in the previous paragraph, support a Late Paleocene age for Subunit A.

The arenaceous foraminifera *Cyclammina amplexans* and *Ammodiscus cretaceus* occur at 199.80 m and 169.25 m respectively. It has been observed that the latter taxon can be quite abundant in the Paleocene of the North Sea and Norwegian Sea, although it is not restricted to this period. At 219.82 m, *Coscinodiscus* sp. 1 is common, which allows a correlation to the Late Paleocene Zone NSP 4 (King 1983). Based on this restricted evidence, a Late Paleocene age is most probable for this subunit in the core 7317/10-U-01.

149.4 - 143.2 m: Subunit B, Upper Paleocene (Selandian) - ?Lower Eocene (Ypresian)

The occurrence of *Cerodinium wardenense* through Subunit B and the presence of *Deflandrea denticulata* at the top of this subunit, suggest an age not younger than Early Eocene. In overall character the dinoflagellate cyst assemblages from Subunit B appear similar to those recovered from Subunit A, and a Late Paleocene age cannot be excluded.

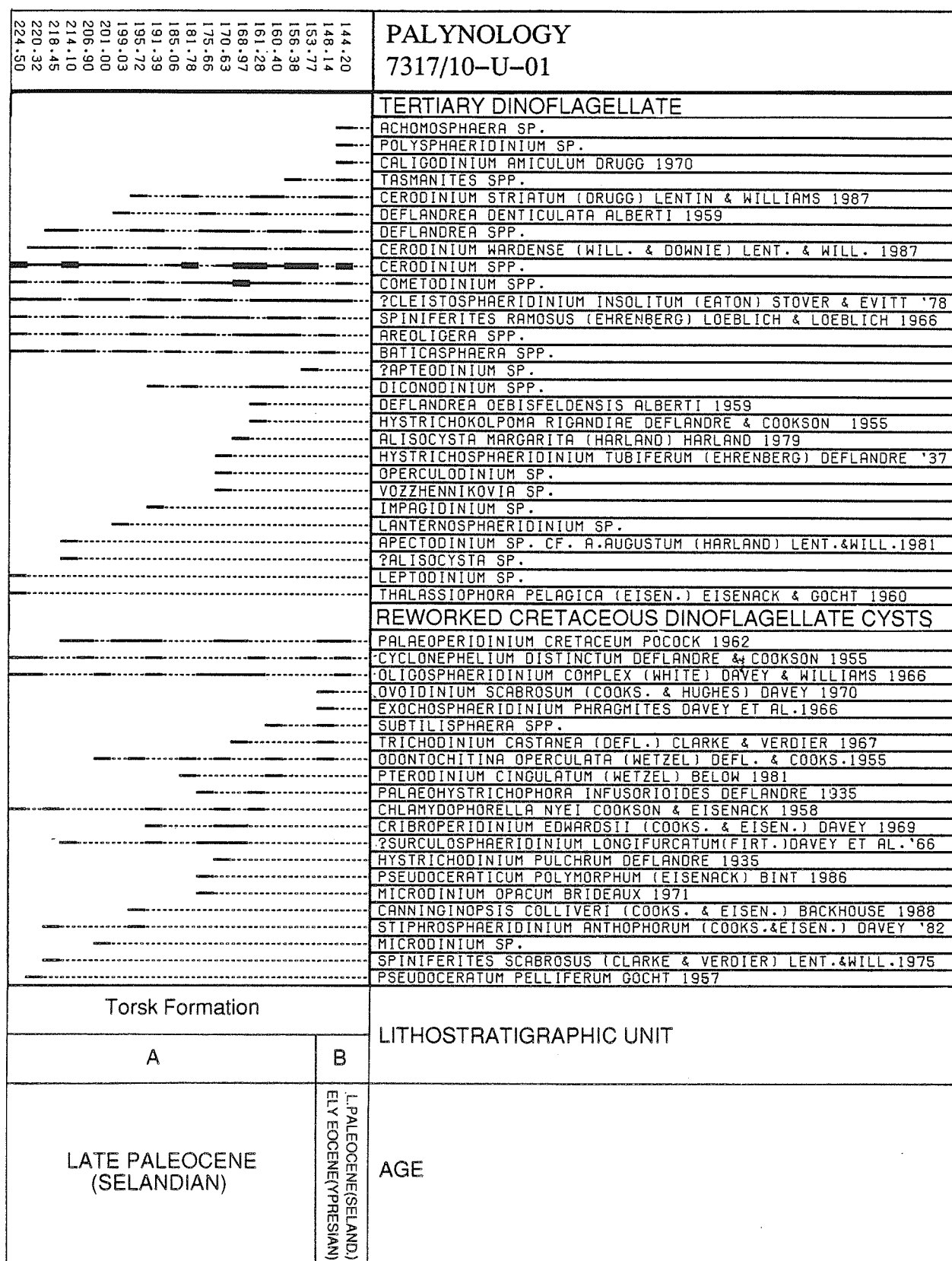


Figure 3.1.6 Distribution chart of dinoflagellate cysts in core 7317/10-U-01, sorted according to earliest occurrence.

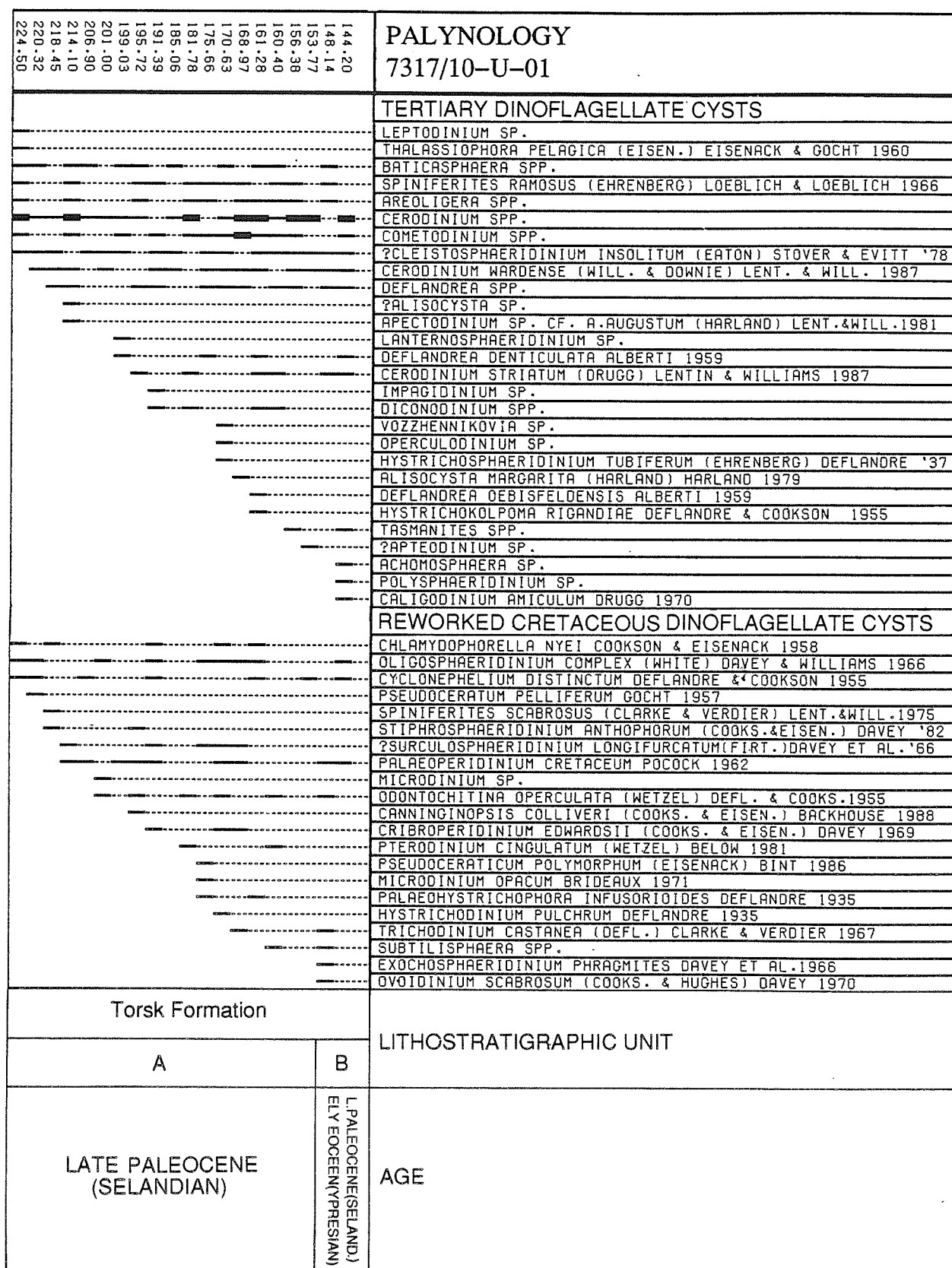


Figure 3.1.7 Distribution chart of dinoflagellate cysts in core 7317/10-U-01, sorted according to latest occurrence.

3.1.6 Organic geochemistry

A total of eight canned samples from this core was analysed for headspace and occluded gas composition. Methane was the dominant gaseous hydrocarbon, particularly in the headspace gas fraction where it was the only gaseous hydrocarbon present (Figure 3.1.8). Trace amounts of C₂ hydrocarbons were present in the occluded gas fraction of all the samples. C₃, C₄ and C₅₊ hydrocarbon compounds were only present in the occluded gas fraction of the five samples below 177.60 m. There was no obvious feature of the cores which coincided with the occurrence of the higher molecular weight gaseous hydrocarbons.

Carbon isotope data were obtained for methane and ethane from two samples at 149.40 m and 187.66 m. The isotopic composition of these two compounds suggests derivation from thermally mature oil- or condensate-prone source rocks, although nothing can be said regarding the age or location of this source. There is also some evidence for an input of biogenically-derived material in these two samples.

Forty-four core chip samples were analysed for total organic carbon (TOC) content (Figure 3.1.9). The TOC content of these samples was found to show relatively little variation, with most samples having values of between 0.8 wt% and 1.1 wt%, and an average TOC content of 1.0 wt%. Rock-Eval pyrolysis data from twenty-four selected samples also show relatively little variation (Figure 3.1.9). Low S₁ concentrations of typically less than 0.05 mg/g suggest that only trace amounts of extractable organic matter (EOM) are present in the samples. The low S₂ concentrations (0.24-0.61 mg/g rock) and hydrogen index values (32-49 mg/g TOC), together with Tmax values of between 422°C and 435°C, suggest that the kerogen present in the analysed rocks is thermally immature and retains little or no hydrocarbon generation potential.

This conclusion is further supported by pyrolysis-gas chromatography data, which show the kerogen pyrolysate in five selected samples to be dominated by aromatic compounds. Aliphatic peaks are present in all of the pyrolysis-gas chromatograms, but are not significantly developed beyond the C₁₁ peak. Microscopic examination of the kerogen indicates that the kerogen consists mainly of woody and coaly debris, as might be expected from the pyrolysis data. Significant amounts of liptinitic debris in four of the five samples examined probably represent reworked material.

Vitrinite reflectance data obtained from a total of eight samples gave mean values of between 0.38% and 0.43% Ro. Taken together with the yellow to yellow-orange sporinite fluorescence and thermal alteration index values which vary from below 2 to 2.5, these data indicate that the kerogen present in the analysed succession is thermally immature.

Relatively low extractable organic matter (EOM) concentrations were recovered from five samples at 147.55 m, 177.60 m, 199.04 m, 206.36 m and 214.30 m (Figure 3.1.10). The lowest EOM concentration of 99 ppm was recovered from the uppermost sample, while higher EOM concentrations of between 300 and 400 ppm were recovered from the remaining four samples. Asphaltene compounds accounted for between 40% and 70% of the total EOM, where sample quantity permitted accurate determination. Hydrocarbon compounds typically accounted for less than 20% of the total EOM recovered from the samples. Low recoveries for the saturate and aromatic hydrocarbon fractions prevented acquisition of meaningful carbon isotope data.

Gas chromatograms of the saturation hydrocarbon fraction of the five sample extracts show similar n-alkane profiles with a marked odd-carbon number predominance in the nC_{25} - nC_{31} peak range. This is reflected in the relatively high carbon preference index values of 1.8 to 2.4 (Figure 3.1.11). In addition, the remaining n-alkane peaks show a tendency to reach a maximum in the area around the nC_{19} - nC_{21} peaks, an exception to this being the samples at 177.60 m and 199.04 m where this subsidiary mode occurs towards the nC_{24} peak. The samples have similar pristane/phytane ratios of between 1.1 and 1.4, with exception of the uppermost sample at 147.53 m which has a pristane/Phytane ratio of 2.2. Although this similarity may suggest a common source for the bitumen, it could also reflect evaporative loss from the bitumen.

The gas chromatograms of the aromatic hydrocarbon fraction are dominated by a number of prominent, unidentified peaks, together with a prominent "hump" of unresolved material.

Pyrolysis-gas chromatograms of two asphaltene samples at 177.60 m and 206.36 m show similar well-developed homologies of aliphatic compounds up to C_{25} , together with marked peaks representing the pristenes, (m+p)-xylene and an unidentified compound(?) which elutes prior to (m+p)-xylene. These pyrolysis-gas chromatograms are similar to those produced by the majority of the asphaltene samples analysed from the other cores, suggesting a common source for these bitumens.

The saturate fractions obtained from the above two samples were further analysed using combined gas chromatography-mass spectrometry. The terpane (m/z 191) mass fragmentograms from these two samples both show rather immature hopane distributions. These are characterised by high relative abundances of the 17 β (H),21 β (H)-hopanes and the C₃₁₊ 22R hopanes. In addition, a peak representing norhop- $\Delta^{13(18)}$ -ene was recognised in the m/z 191 mass fragmentogram of the sample at 206.36 m. These features are usually associated with thermal maturities equivalent to a vitrinite reflectance of 0.3-0.4% Ro. This is similar to vitrinite reflectances measured from the kerogen in this core.

The sterane (m/z 217,218) mass fragmentograms of the two samples showed rather unusual peak distributions which could not be correlated with a particular maturity level at this time. The prominence of the C₂₉ 20R sterane peak in these fragmentograms may indicate an input from terrestrial organic matter, although coelution of an unusual triterpene seen in samples from other cores could also explain this.

In conclusion, the analysed bed-rock succession from this well is characterised by a fair content of thermally immature, largely inert type IV kerogen consisting mostly of reworked terrestrial organic debris. Traces of immature, asphaltene-rich bitumen are similar in composition to bitumen present in cores 7316/03-U-01 and 7316/06-U-02. This material is thought to be derived either by migration from a local source containing terrestrial organic matter, or by reworking of material eroded from a bitumen-rich horizon.

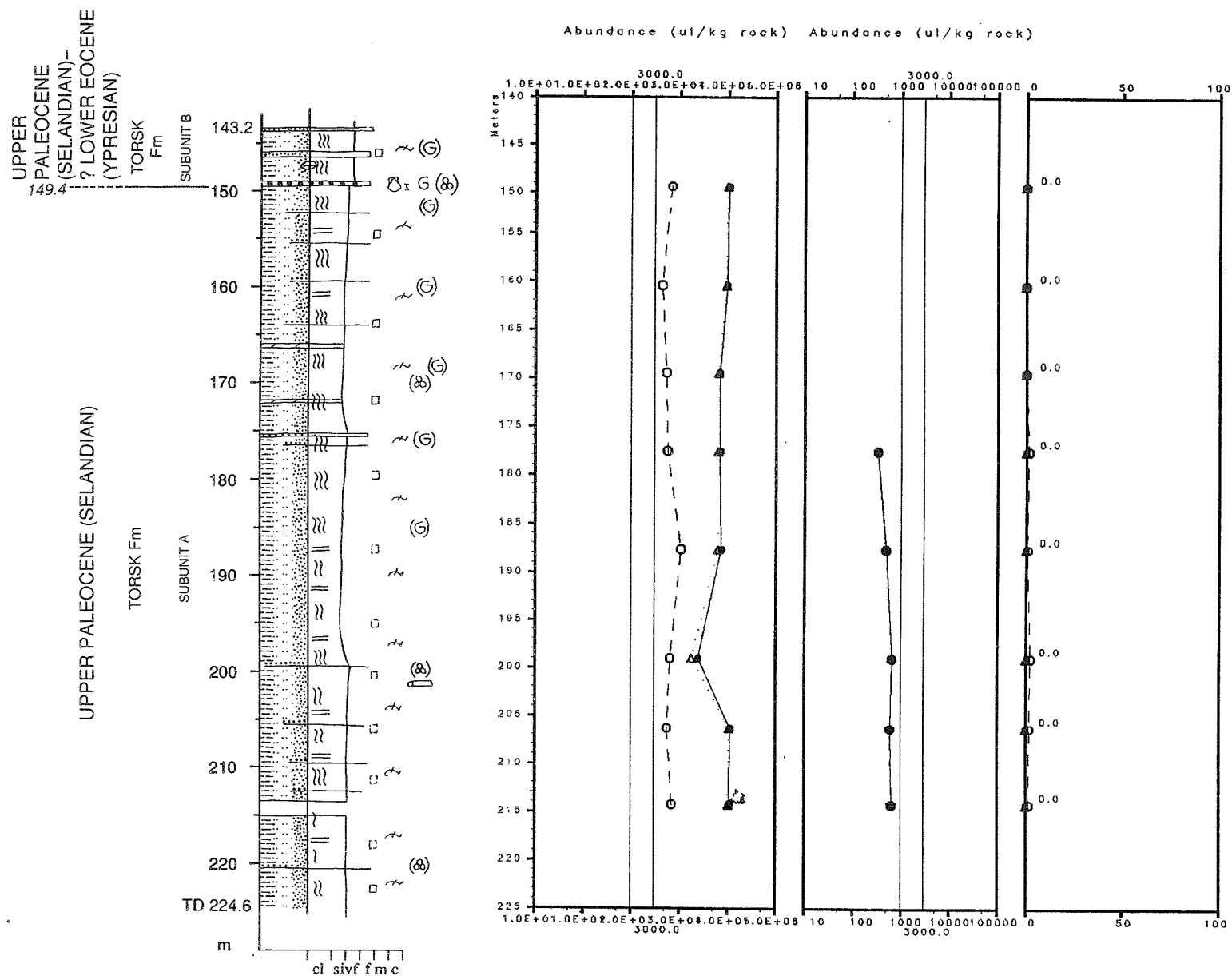


Figure 3.1.8

Yield and composition of headspace/occluded gas of core 7317/10-U-01.

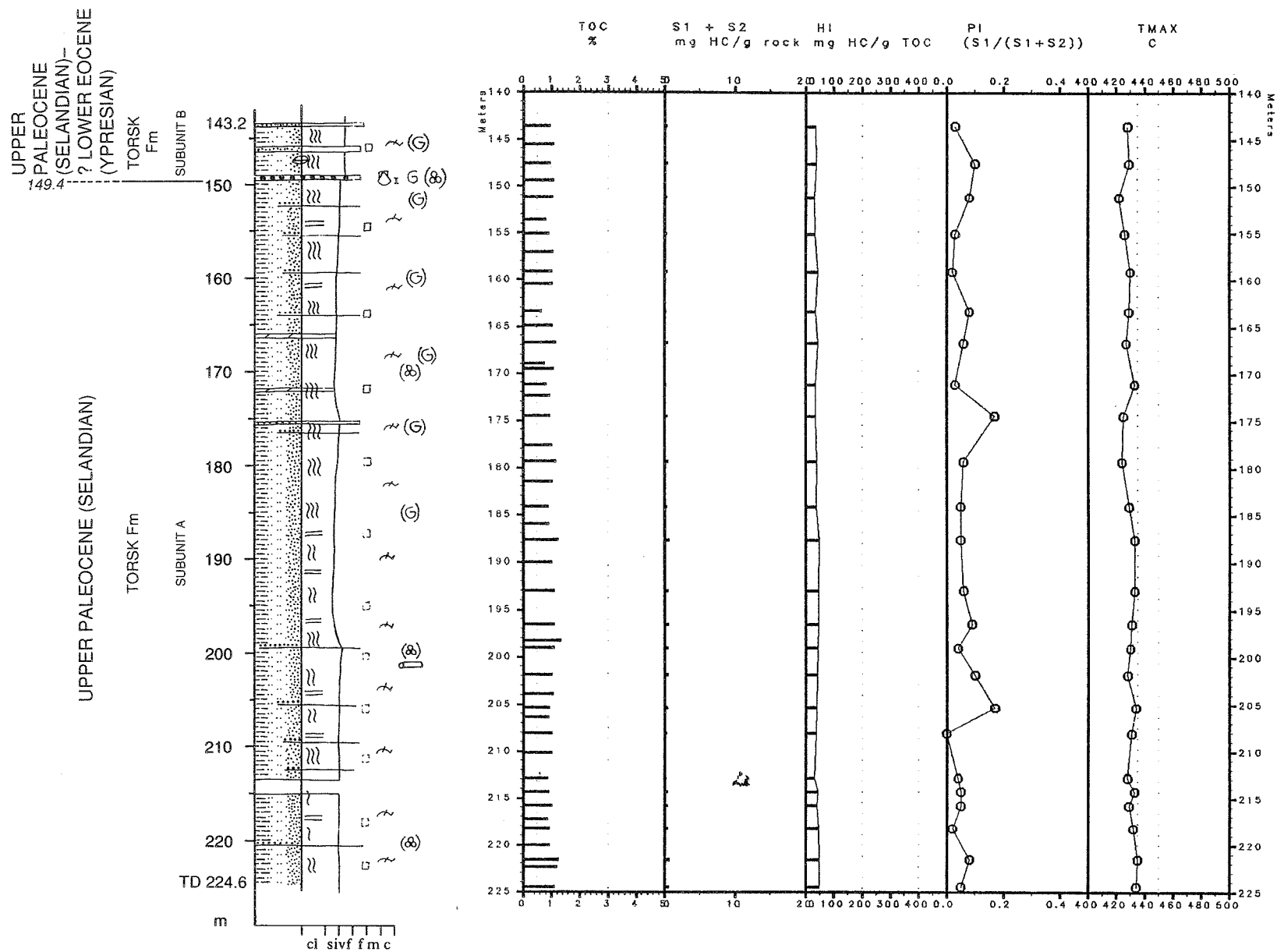


Figure 3.1.9

Total organic carbon content and Rock-Eval data of core 7317/10-U-01.

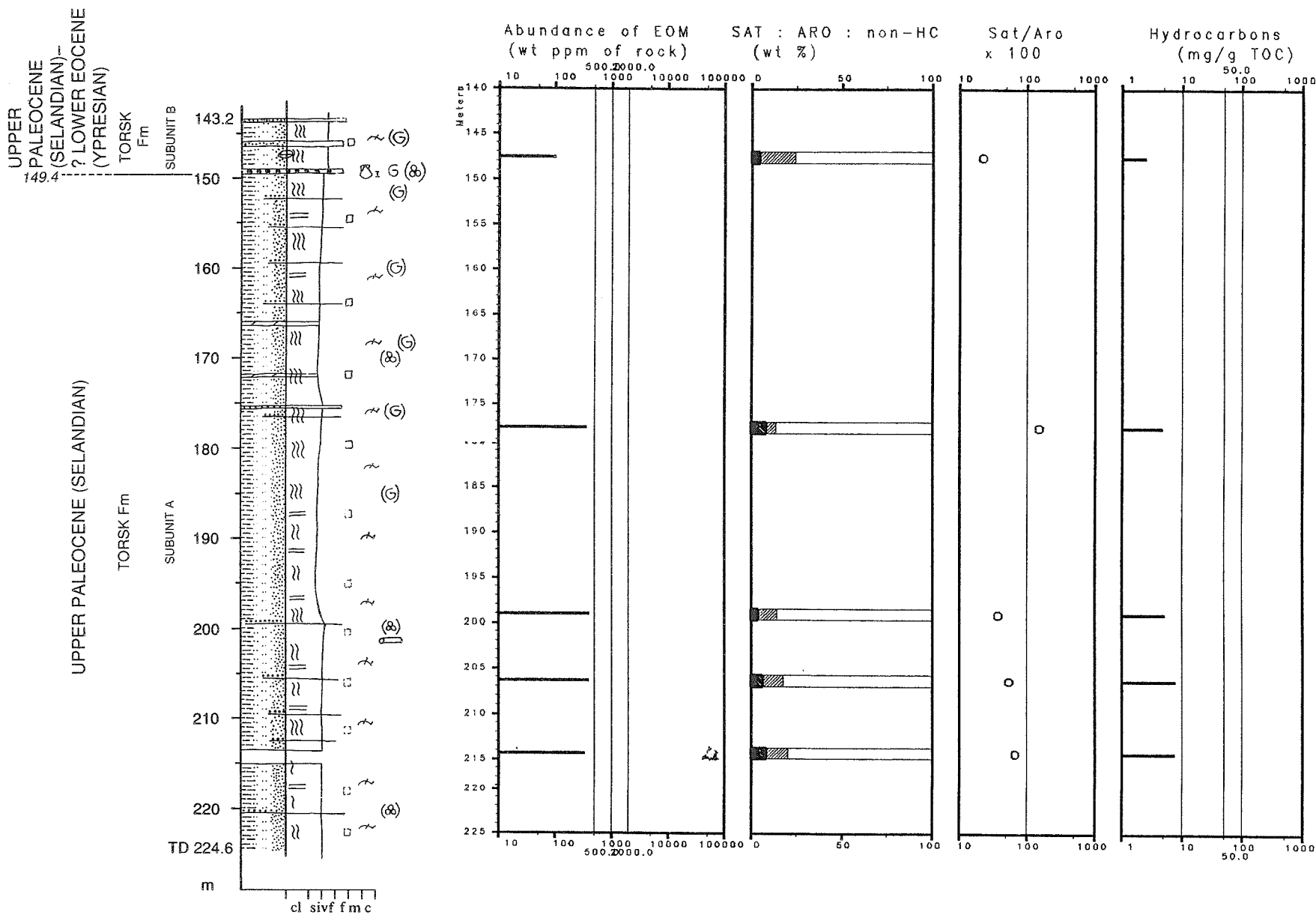


Figure 3.1.10

Extract data for core 7317/10-U-01.

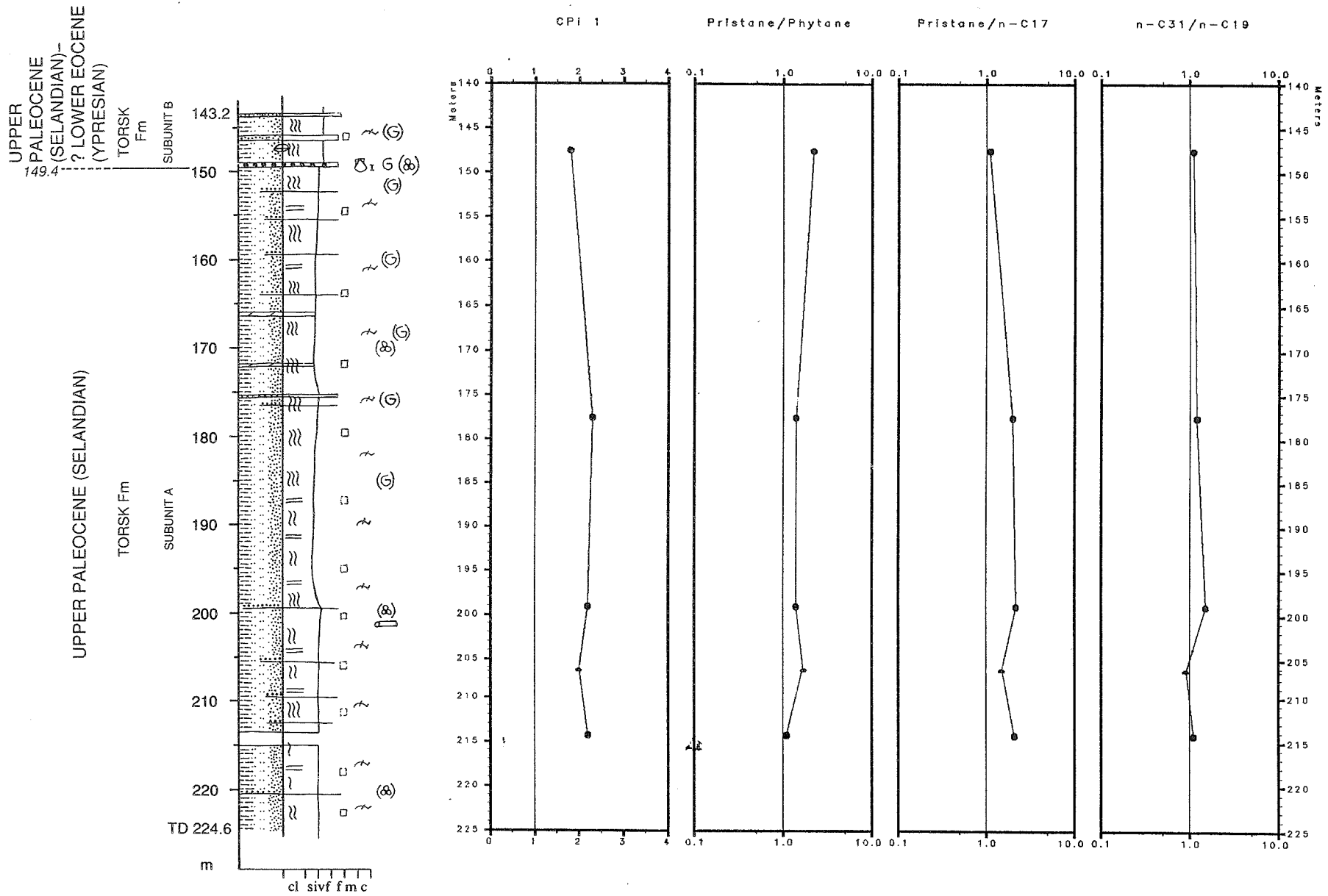


Figure 3.1.11

Compound ratios from gas chromatograms of saturated hydrocarbon fractions from core 7317/10-U-01.

3.1.7 Depositional environment

The siltstones and shales of both subunits in core 7317/10-U-01 contain marine palynomorphs and foraminifer fragments confirming a marine depositional environment.

The dominant clayey and silty lithology indicates deposition predominantly from suspension. However, thin sandstone beds are also observed. The sharp bases of the beds indicate pulses of higher energy. The hierarchy of structures indicates deposition during waning flow energies. A rapid deposition of the sands is also suggested by the ichnofauna, where the upper part of some thicker beds are totally reworked, while in the lower part the primary physical sedimentary structures are preserved. The lithology and sedimentary structures within subunits A and B are commonly reported from shelf type of deposits. The sandstone beds may represent storm events where the sands may have been supplied from nearby shorelines. Walker (1984) indicated that this type of sedimentary facies represents deposition below storm wave base in an inner shelf type of setting.

Shelf depositional models described by Walker (1984) and Bouma et al. (1982) indicate that the thickness of the sandstone beds should increase toward the shoreline. A regressive shelf deposit should, in this context display a coarsening upward grain size pattern. Subunit A of the core shows an overall, but composite coarsening upward grain size pattern. Each of the coarsening upward intervals may represent shoaling upward successions in a general upward shoaling succession. The regressive stacking pattern is characteristic for high-stand deposits (Posamentier et al. 1988).

The transition between subunits A and B shows a change from siderite/pyrite concretions/beds to calcite, reflecting more oxygenated conditions in Subunit B than in A. The transition also represents a shift to thicker sandstone beds containing both intra- and extrabasinal clasts. The inclined surface representing the transition shows burrows filled by sands from the subsequent deposited sandstone beds; i.e. a minor time gap is thus indicated. The unconformity most likely represents a minor shoaling event.

A bioturbated sandstone bed is lying concordantly on the inclined surface. It is overlain by horizontally bedded rocks of Subunit B. The inclined surface may have formed as a response to tectonic movements, creating a small relief on the sea floor. The minor uplift may have caused an

increased input of Late Mesozoic (?Late Cretaceous) and also Late Paleozoic sediments, indicated by the abundance of *Inoceramus* prisms and probably also glauconite and cherts.

Lithology and age assesement of the core indicate that the rocks are equivalents to the Torsk Formation of the Sotbakken Group of Worsley et al. (1988).

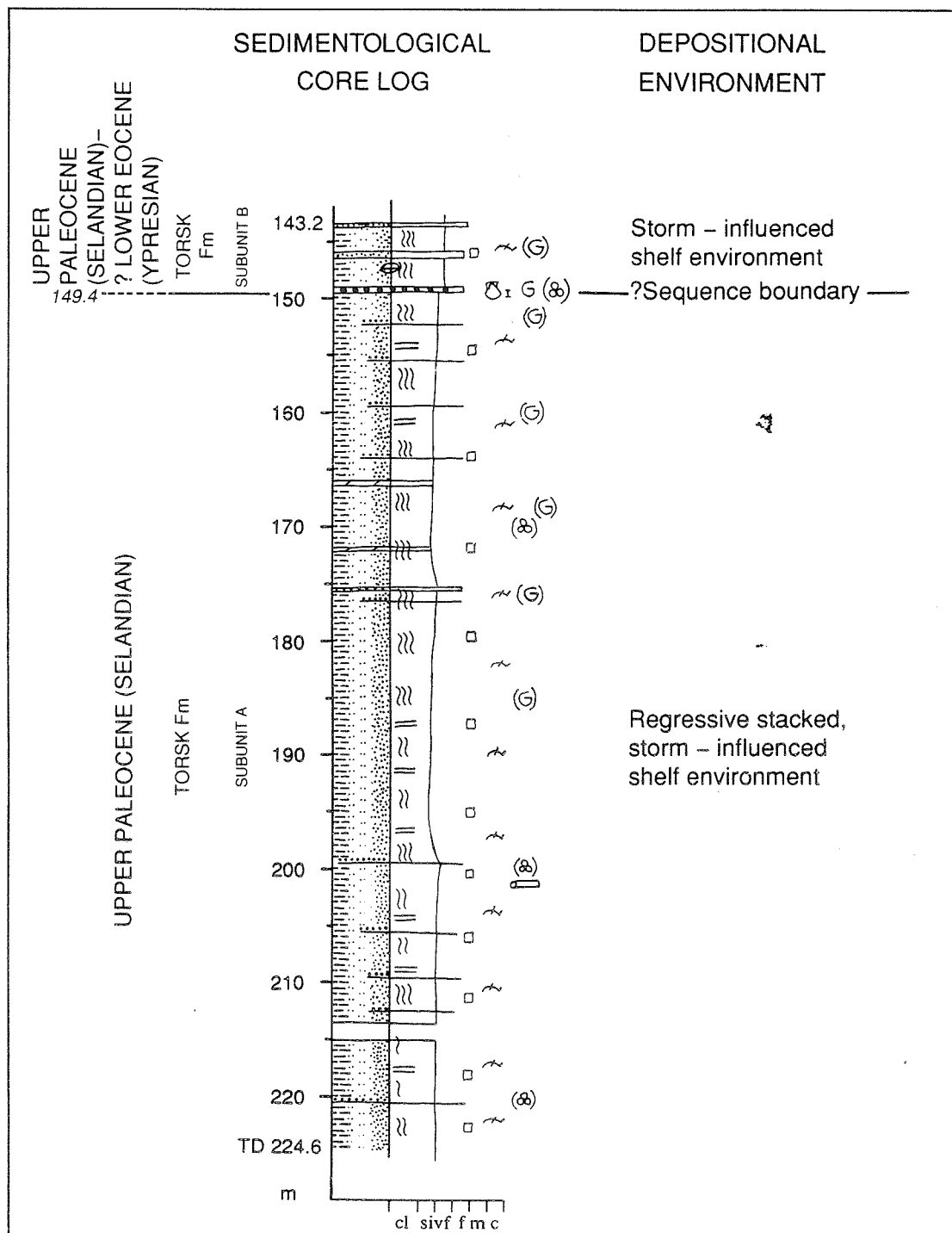


Figure 3.1.12 Interpreted depositional environment of core 7317/10-U-01.

3.2 Core 7316/06-U-02

3.2.1 Summary

Core 7316/06-U-02 was drilled about 1.5 km north of core 7316/06-U-01. The cores are separated by a significant angular unconformity representing a major hiatus. Based on seismic data a unit below about 136 m is nearly acoustically transparent with no continuous reflectors. It is overlain by a seismic unit with two strong reflectors where one is correlated to a flooding surface at 102.8 m in the core.

The core represents an overall coarsening upward stacking pattern of medium dark grey silty claystone to silty sandstones with thin graded sandstone beds (Figure 3.2.1). The sediments were probably deposited in regressive, storm-influenced inner shelf environments in ?Early/Middle Eocene time.

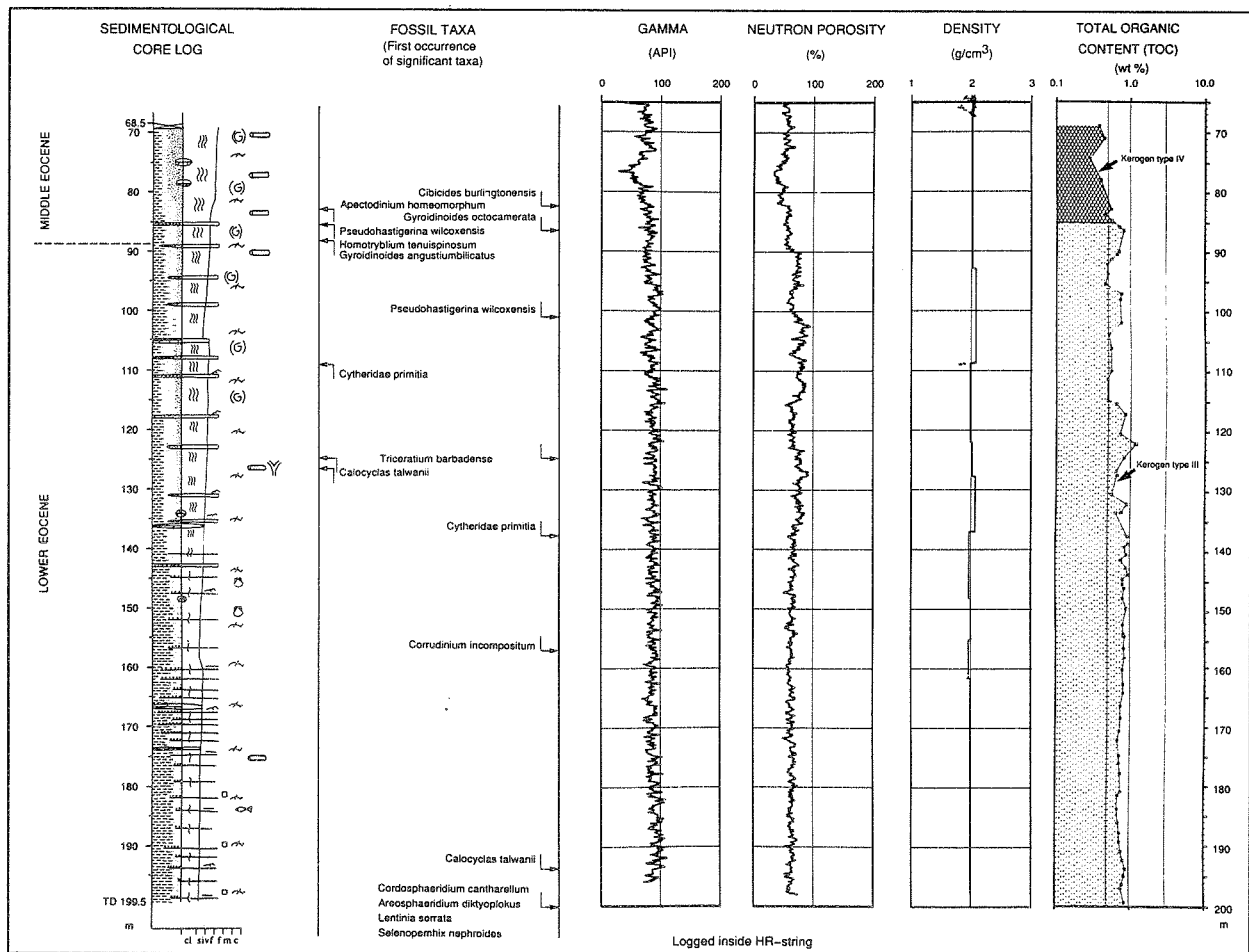


Figure 3.2.1 Summary of main data from core 7316/06-U-02.

The rocks analysed from this core contained moderate amounts of thermally immature, type III/IV kerogen with little or no hydrocarbon generation potential. Traces of asphaltene-rich bitumen in the core appear to share a common source dominated by terrestrial organic matter. This bitumen is similar in composition to that described from core 7317/10-U-01, and an indigenous source is considered unlikely.

3.2.2 Seismic interpretation

Core 7316/06-U-02 was drilled about 1.5 km to the north of core 7316/06-U-01 (seismic line IKU-1650-89, sp 1814). There is a significant angular unconformity between the two sites, representing a major hiatus. This is discussed further in Section 4.1, where the local tie to the seismic is presented.

Table 3.2.1 Correlation of seismic data and core 7316/06-U-02. Interval sound velocities noted "seismic" appear from this seismic-core correlation.

Depth below seabed (m) (msec)	Core unit	Thickness of seismic unit (m) (msec)		Interval sound velocity (m/sec)		
				Seismic	Core	Log
68.5 75	Quaternary	68.5	75	1820	1725	-
83 89	-	14.5	14	2070	-	-
102 107	-	19	18	2100	-	-
137 142	-	35	35	2000	1940	-
TD199.5 197	-	62.5	55	est.2250	2260	-
Water depth: 421 m (512 msec) Depth below sea level: 620.5 m (709 msec)						

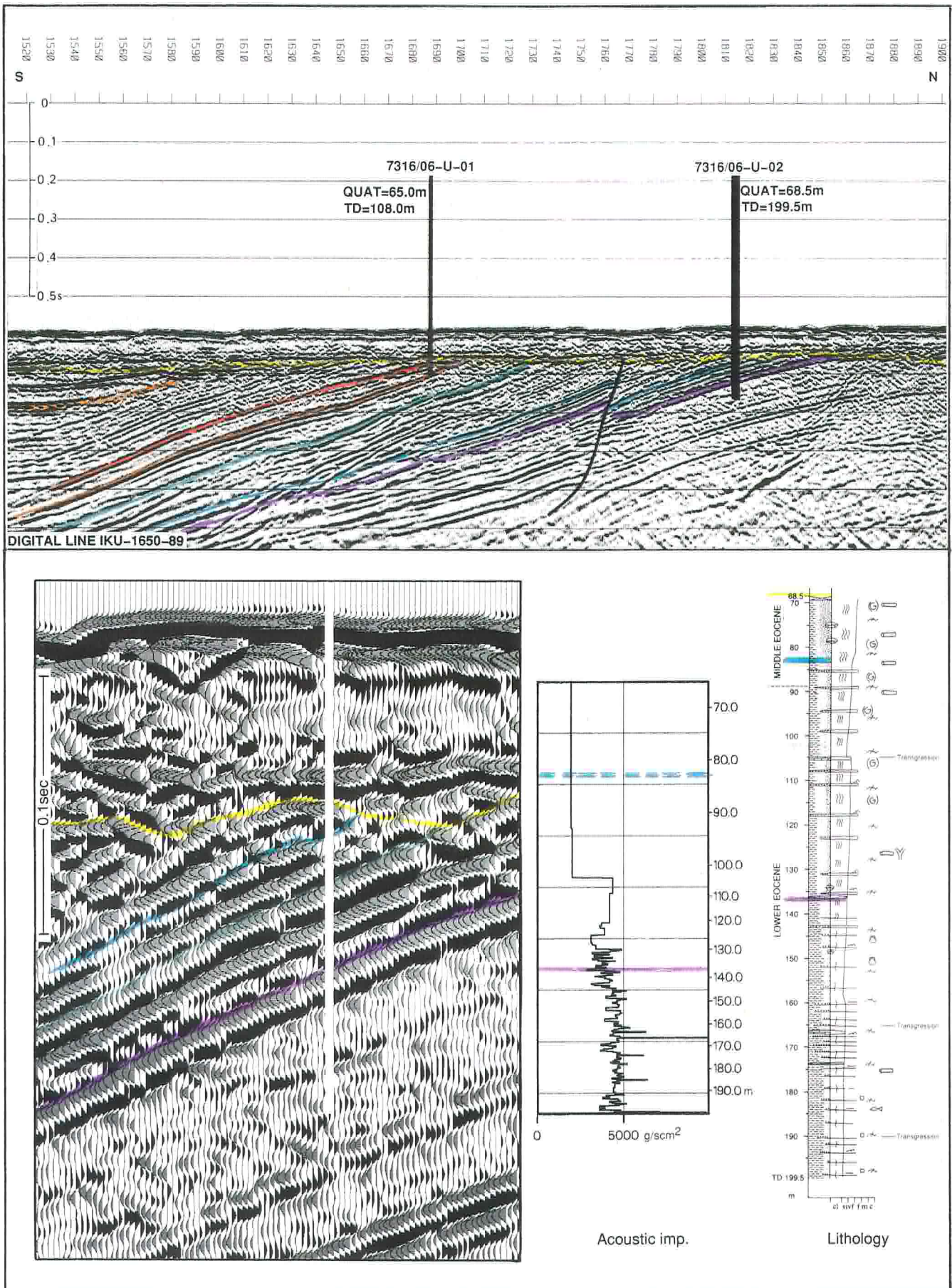


Figure 3.2.2 Seismic correlation of core 7316/06-U-02.

The core has not been formally divided into lithological subunits, but from the seismic and the log data the core seems to comprise at least two units. The lowermost one extends up to about 136 m. Here the sound velocity measured on the core (Figure 3.2.1 and Appendix A.1.2) drops from an average of about 2250 m/s to about 1900 m/s. There is also a change in the pattern of the neutron log from a straight line with only small peaks to a decreasing trend with more variations. The log suggests another change at about 102 m.

In the seismic data the lowermost interval (TD - 136 m, Figure 3.2.2, Table 3.2.1) appears nearly acoustically transparent with no continuous reflections, which seems to be in agreement with an interpretation of deposition in open marine shelf environment. The core data shows that the boundary at 136 m represents a gradual transition to a shallower environment.

The interval 136 - 83 m comprises two relatively strong reflections. One may coincide with an increase in the measured neutron porosity at about 115 m and the other with a transgression recognized at 102.8 m in the core. In the seismic data the interval 136 - 83 m represents a band of strong and parallel reflections.

The reflection pattern above 83 m is more irregular and it may seem that the reflections downlap on the reflector at 83 m. In the core this is represented by a rapid coarsening upward trend at 83 m.

3.2.3 Sedimentology

Core description

199.5 (TD) - 68.5 m: Torsk Formation, ?Early/Middle Eocene.

At this site 131 m of bedrock were cored between 68.5 m and 199.5 m below sea floor. The core contains silty shales, siltstones and very fine-grained sandstones forming an overall coarsening upward grain size pattern. The major coarsening upward trend is composite and contains four distinct coarsening upward sequences with the transitions at 189.5 m, 166 m and 102.8 m.

Silty claystones/shales with a few graded thin sandstone beds/lenses dominate the lithology below 139 m. Clayey and sandy-siltstones dominate between 139 m and 83 m and silty sandstones above 83 m. The sandstone beds have sharp bases, overlain by horizontally laminated and/or rippled laminae.

Siderite concretions/beds are occasionally observed in the rocks as shown in Figure 3.2.3. One of them occurs at the transition between two coarsening upward subunits at 105 m. Pyrite is also present in small amounts. It is most common as cement in burrows, but may also be present as small euhedral crystals in the siltstones.

Small fossil fragments occur in small amounts throughout the core, together with foraminifer fragments and a few bivalve and wood fragments.

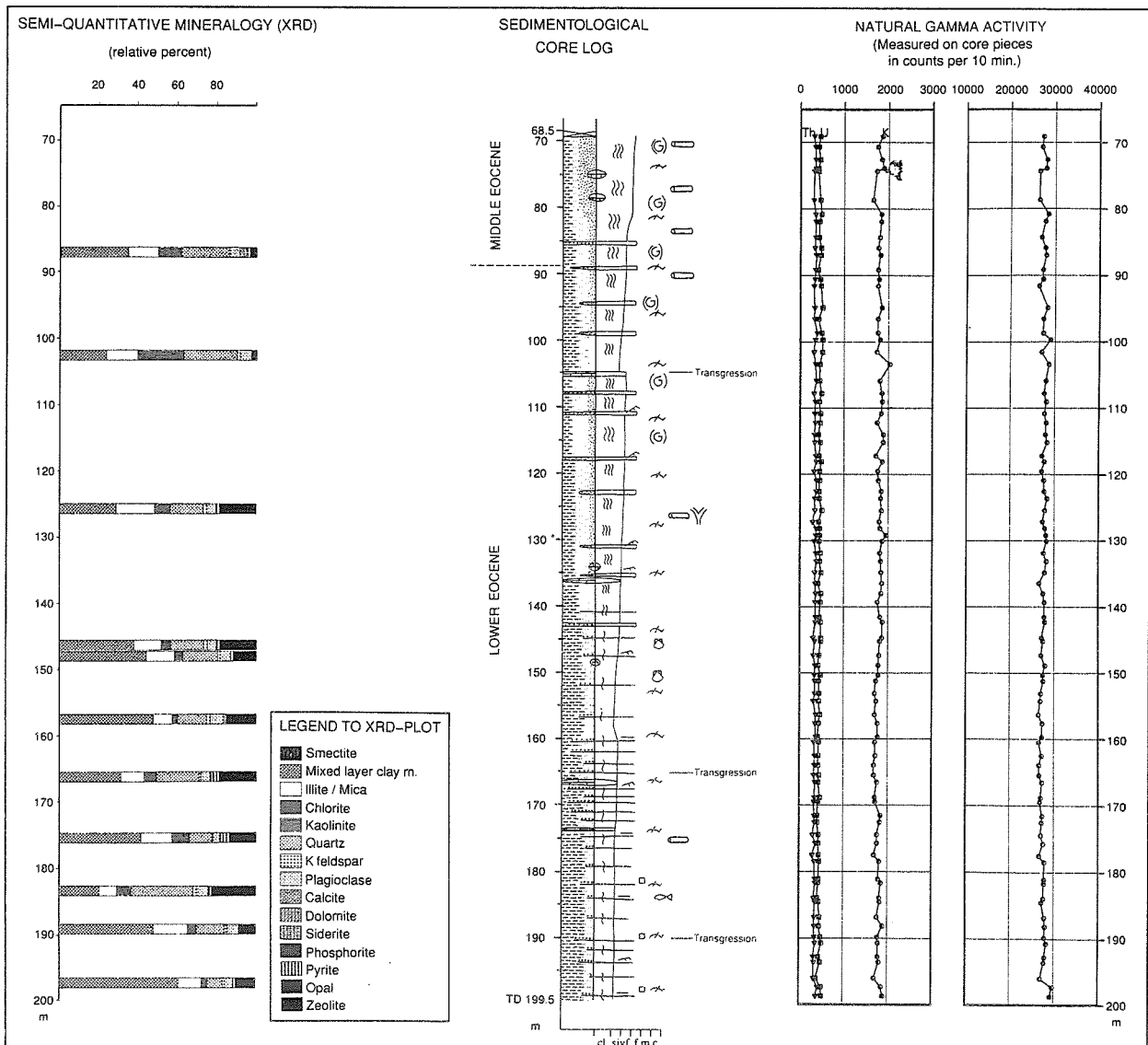


Figure 3.2.3 Lithological column, XRD mineralogical analysis and natural gamma radiation of core 7316/06-U-02.

Bioturbation occurs throughout the core (Appendix A.7). The ichnodensity is controlled by the lithology. In silty claystones/shales below 139 m the biogenic structures are small and filled by coarser, more sandy sediments than the surrounding rocks. The bioturbation is incomplete by low diversity sediment-feeder communities. Above 139 m the diversity and burrow size increase. In certain beds the diversity is 6-8, showing four- to five-tier successions. In general, the diversity is slightly lower. Back-fill structures (1 cm in diameter) are observed together with smaller fills. *Helmintopsis* burrows are commonly seen above 140 m. In addition, several size classes of *Planolites* as well as *Palaeophycus* are present, while large burrows of *Thalassinoides* are present above 128.6 m.

Petrography

Eleven X-ray diffraction (XRD) and eight thin sections have been analysed (see Table 3.2.2).

Table 3.2.2 *Semi-quantitative mineral composition of the sediments based on X-ray diffraction analysis of core 7316/06-U-02.*

Depth	Mix	Mi/Il	Chl	Qtz	KFsp	Plag	Calc	Sid	Pyr	Opal	Zeo
87.11	35.1	15.0	11.9	24.0	0.6	4.7	2.1	2.1	1.5	1.6	1.4
102.60	23.7	15.7	23.6	27.0	1.5	5.6	0.0	0.6	1.4	0.0	0.7
125.75	29.1	19.4	8.0	16.6	2.1	4.3	0.0	0.5	1.6	18.4	0.0
146.32	37.9	13.6	4.9	16.6	2.1	4.3	0.0	0.5	1.6	18.5	0.0
148.03	44.3	14.0	4.4	17.8	1.0	5.3	0.0	0.4	1.1	11.8	0.0
157.57	47.9	9.5	2.9	14.8	1.9	6.5	0.0	0.9	1.0	14.8	0.0
166.27	31.4	11.7	6.3	21.4	1.1	4.6	0.0	0.5	4.0	18.3	0.7
175.40	41.7	15.6	8.9	11.5	0.5	3.4	0.0	0.7	4.3	13.5	0.0
183.50	20.2	8.9	7.4	31.7	2.2	5.3	0.0	0.4	1.4	21.1	1.4
189.30	48.1	17.4	4.4	14.5	1.5	5.6	0.0	0.2	1.0	7.5	0.0
197.48	60.9	11.6	3.0	8.1	0.7	4.3	0.0	0.3	1.3	9.5	0.4

The XRD analysis shows that clay minerals are the most important mineral group represented in the sediments. Mixed layer clays form the most important group and contribute with 20 - 60% of the samples. Also illite and chlorite are important clay minerals in the sediments. Quartz is the second most important mineral with the highest content in siltstones and sandstones. It contributes with 20 - 60% of the rock. In addition alkali feldspars, plagioclase, calcite, siderite and pyrite are present in small amounts.

Opal-CT (0-21%) and clinoptilolite (0-1.5%) are more exotic minerals. Clinoptilolite and opal-CT are commonly associated with supply of volcanic sediments, and a diagenetic alteration of volcanic ash may be a likely source for the two minerals (Kastner et al. 1977).

Radiolaria tests have been recrystallized from Opal-A to Opal-CT. The mineral transformation from Opal-A to Opal-CT takes place at temperatures of 40-50°C. Further opal-CT transformed to quartz at low temperatures. Roaldset (pers. comm.) and Gjeldsvik (in press) indicate that the transition to quartz takes place at temperatures of about 50°C, while Bjørlykke (1989) indicated slightly higher temperatures (60-80°C). A maximum burial temperature of about 50°C is therefore likely.

Thin section analysis of 6 of the 8 slides were from very fine-grained sandstones.

The arenites contain subangular to angular grains. In addition, a few percent of spherical to elliptical pelloids and mica flakes are present. Quartz dominates the sandstones, and other important grains are cherts, alkali feldspars, plagioclase, muscovite, biotite, pelloids and glauconite pelloids. Trace amounts of rutile, zircons, apatite, tourmaline, siliceous microfossils and black to dark brown plant fragments are also present.

Mechanical compaction has played a major role in the diagenesis of the sandstone and is indicated by a dense packing of the grains. Minor amounts of pyrite cement are also present. Alteration of biotite is indicated by expanded edges and common inclusions of brownish mineral, probably anatase. Alkali feldspars and basic plagioclase are incompletely altered to clay minerals.

The major difference between the arenites and the greywacke sandstones is that clay mineral matrix fills all the porosity of the greywacke sandstones. In addition, they also contain more mica than the arenitic sandstones.

Thin sections from siderite and calcite concretions have been studied from 136.09 m and 102.85 m. They show many textural similarities; both consist of microcrystalline masses of carbonate with inclusions of terrigenous grains, glauconite and pyrite. However, a major difference in texture is the presence of radial fibrous calcite cement on the surface of mica grains at 102.85 m in the calcite concretion/bed. The cement shows similarities with fibrous aragonite cement and may be calcite replacement after marine aragonite cement. This suggests that the growth of the calcite concretion started at the sea floor.

3.2.4 Petrophysical logs and core measurements

Downhole logs

The gamma and neutron logs of core 7316/06-U-02 were run inside the HR string from TD to the seabed. The gamma log values logged through only the HR string are recalculated to open hole values based on a regression analysis from a previous shallow drilling survey. Due to very low variance in the gamma log values it was impossible to calculate any depth shift correction (see Appendix A.10). The logs are shown on Figure 3.2.1 and on Figure 3.2.4 together with the core measurements. The gamma log and petrophysical properties are also shown together in the summary sheet in Appendix A.1.2.

The open marine deposited fine-grained sediments in this core show mainly statistical variations from a straight line around 90 API for the gamma log. The lower three coarsening upward units below 149 m, 166.5 m and 102.8 m have not caused any noticeable trends in the gamma log. Above 102.8 m a general upward decreasing trend reflects the coarsening upward unit. A negative peak occurs at 76.5 m.

The neutron porosity log shows only relative differences in the neutron signal. This log shows stable readings below 136 m, more spiky readings between 136 m and 103 m and an upward decreasing trend above 103 m, which

may be caused by the coarsening upward trend. Density measured on core pieces (Figure 3.2.1) shows stable values around 2.0 g/cm^3 .

Sound velocity and gamma measurements on core

Figure 3.2.4 shows the gamma wireline log together with the spectral gamma values and the horizontal sound velocity measured on the core. The trends of the gamma laboratory measurements correlate well with the gamma log values. The K-plot on the spectral gamma laboratory measurements shows that K is the main source for the gamma log response. The U and Th values vary

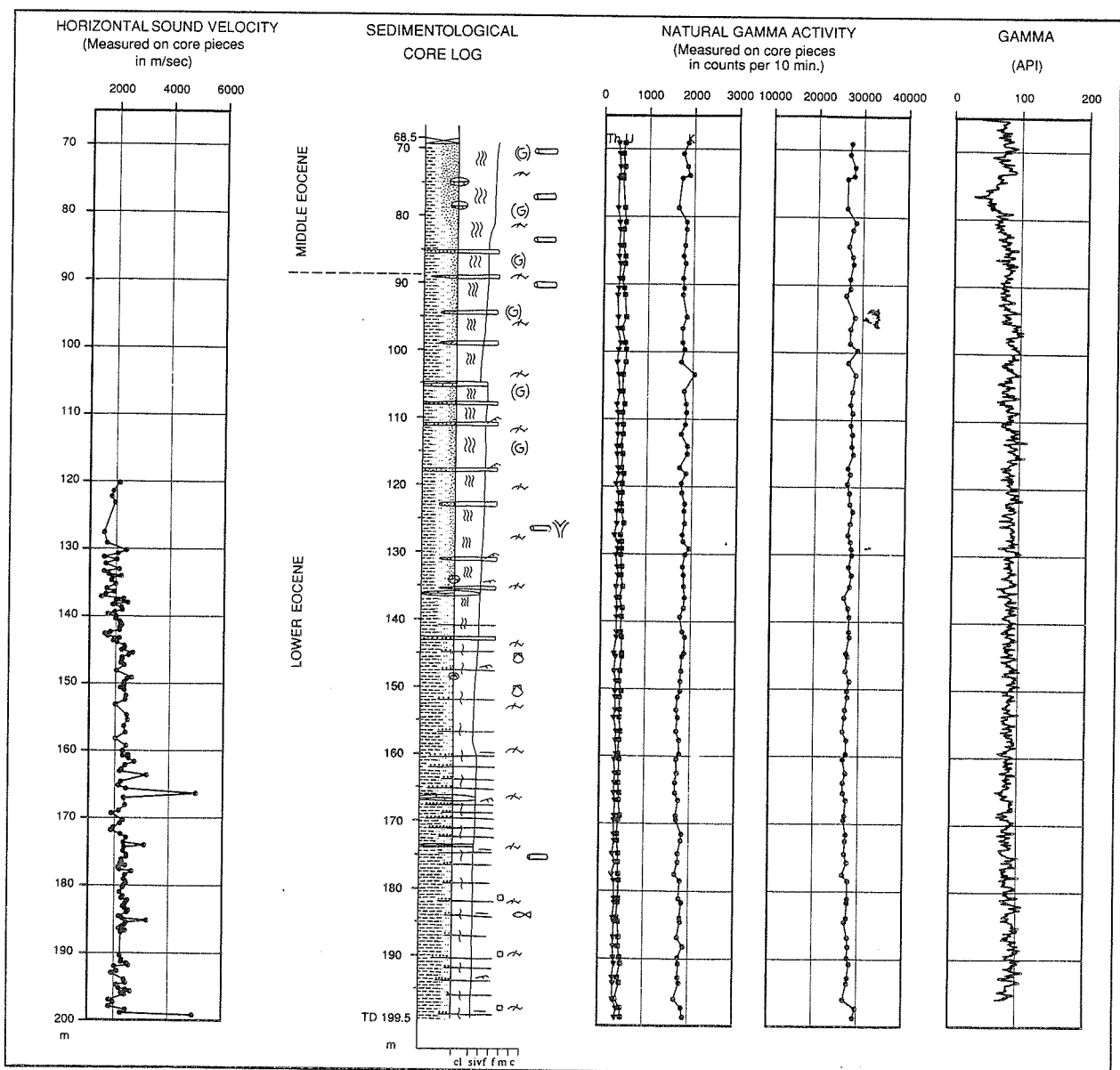


Figure 3.2.4 Lithology log and laboratory measurements of spectral gamma and horizontal sound velocity of core 7316/06-U-02 plotted together with downhole gamma log.

only very little from the background level. The horizontal sound velocity measurements show mainly values of 2000-2200 m/s with peaks exceeding 4000 m/s in cemented zones.

Thermal conductivity

See Appendix A.9.

3.2.5 Paleontology

Core 7316/06-U-02 contains foraminifera, dinoflagellate cysts, Prasino-phycean algae, ostracodes, radiolaria, diatoms, sponge spicules and echinoid spines.

Intervals of the core containing calcareous foraminifera were also examined for calcareous nannofossils, but none were found. Dinoflagellates and foraminifera are the most numerous and diverse microfossils present in this core, and the foraminiferal assemblages offer definitive indication of an Early and Middle Eocene age. Radiolaria, diatoms and ostracodes are less numerous and diverse, and they yield only a general biostratigraphic signal that is consistent with the foraminiferal biostratigraphy.

Foraminifera

The stratigraphic distribution of foraminifera ostracods, diatoms, radiolaria and other microfossils in core 7316/06-U-02 is shown in Figures 3.2.7 and 3.2.8. A total of 46 species of foraminifera was identified in 38 samples from the bedrock section of core 7316/06-U-02 (199.5 - 69.03 m), and the observed assemblages vary widely in terms of both species composition and frequency of occurrence. The samples at 69.03 m, 153.00 m, 162.50 m, 165.00 m, and 166.68 m were barren of foraminifera, and numerous other samples contained only single specimens of one or a few species. In the majority of the samples, arenaceous foraminifera have higher relative frequencies than the calcareous foraminifera. In the richer samples, both arenaceous and calcareous foraminifera are more numerous and diverse. No arenaceous foraminifera were observed above 84.00 m. Based on foraminiferal biostratigraphy, the core can be subdivided into an Early Eocene interval from 197.76 m to 88.76 m, and a Middle Eocene interval from 86.00 m to the highest fossiliferous sample at 72.75 m.

Foraminiferal species of the assemblages recorded between 199.5 m and 88.76 m have been reported in Paleogene deposits from western and central Europe. *Gyroidinoides angustiumbilicatus* is restricted to the Early Eocene (Ten Dam 1944, Murray et al. 1981). *Pseudohastigerina wilcoxensis* has most of its range in the same age interval, just reaching into the Middle Eocene (King 1983). *Vaginulinopsis decorata* occurs in the Early and Middle Eocene and is mentioned by many workers in the Northwest European sedimentary basins, e.g. Murray et al. (1981), Doppert and Neele (1983) and King (1983). *Spiroplectammina spectabilis* and *Cyclammina amplexans* are conspicuous components of the regional Early Eocene foraminiferal association, although they also occur in older and younger strata. On the base of the foraminiferal assemblages, this interval in the core can be correlated to the North Sea zones NSB 3-4 or NSP 5-6 (King 1983), or to zone FI by Doppert and Neele (1983), that was erected on material from the Netherlands and Belgium.

Calcareous foraminifera are very rare in the Eocene of exploration wells from the British and Norwegian continental shelves. At site 338 of DSDP Leg 38 located just west of the Vøring Escarpment, however, a calcareous foraminiferal association was met that is quite similar to the one that occurs in the above mentioned interval of core 7316/06-U-02 (Hulsbos et al. 1986). At site 338, this association occurs in an Early Eocene interval that has been dated with planktonic foraminifera and calcareous nannofossils (zones NP 11-12).

The foraminifera found between 86.00 m and 72.75 m are characteristic for Middle Eocene sediments from South England, the Netherlands and Belgium. *Vaginulinopsis decorata* is present in all samples. This taxon occurs in the Early and Middle Eocene, and the same is true for *Cibicides proprius*. A characteristic Middle Eocene element is *Cibicides burlingtonensis* at 82.83 m. *Gyroidinoides octocamerata* occurs in the Middle and Late Eocene. Biostratigraphic evidence about these taxa has been derived from Murray et al. (1981) and Doppert and Neele (1983). *Elphidium hiltermanni*, that is considered as a characteristic element of the Late Oligocene in the North Sea basinal area, occurs as early as the Paleocene on the British Isles (Murray et al. 1981).

In North Sea zonation this interval of the 7316/06-U-02 core is situated in zones NSB 5-6 (King 1983). In the Netherlands - Belgium zonation the

interval correlates to zone FH2, that is considered approximately coeval to the calcareous nannoplankton zone NP 14 (Doppert & Neele 1983).

Dinoflagellate cysts

Core 7316/06-U-02 contains abundant and well-preserved dinoflagellate cysts (Figures 3.2.5 and 3.2.6). A total of 71 species has been identified in 35 samples from the bedrock section (199.5 - 60.03 m). The assemblages include 9 species of displaced Cretaceous (Aptian-Cenomanian?) dinoflagellates indicating older sediment reworking in the intervals 182.90 - 177.00 m, 137.40 - 116.15 m, and 69.03 m. The remaining Paleogene flora indicates an Early Eocene to Middle Eocene age, depending on the species criteria employed and the biostratigraphic range information presented by various authors, as discussed below.

The dinoflagellate cyst assemblages in core 7316/06-U-02 include species in common with other assemblages reported from contemporaneous deposits elsewhere in western Europe (Eaton 1976, Bujak et al. 1980), offshore West Greenland (Toxwénius 1986), the Norwegian-Greenland Sea (Manum 1976, Manum et al. 1989) and West Siberia (Kul'kova 1988).

Bujak et al. (1980) subdivided the Eocene Series of southern England into thirteen dinoflagellate cyst assemblage zones. *Cordosphaeridium cantharellum* and *Selenopemphix nephroides* have their first occurrences at the base of their *Heteraulacacysta porosa* Assemblage Zone (BAR-1; correlative with calcareous nannofossil zone NP16; ca. 45 Ma). *C. cantharellum* occurs in 6 of 29 samples in core 7316/06-U-02 in the interval 199.5 - 98.03 m, and *S. nephroides* occurs in 14 of the 35 samples from the throughout core. The single occurrence of *Heteraulacacysta* sp. cf. *H. porosa* at 146.17 m may further indicate a correlation to the *H. porosa* Assemblage Zone. *Lentinia serrata* also has its first occurrence in the *H. porosa* Assemblage Zone in southern England, and this species is present in 24 of the 35 samples from core 7316/06-U-02. Assemblage zones can be time transgressive when correlated over long geographic distances, however, and this is especially true when significant latitudinal distances are involved.

Toxwénius (1986) used the concurrent occurrences of *Eatonicysta ursulae*, *Glaphyrocysta vicina*, *Dracodinium varielongitutum*, *Wilsonidinium tabulatum*, *Phthanoperidinium comatum* and *Areosphaeridium diktyoplokus* to recognize the Early Eocene succession (zone

D9) in deep wells offshore western Greenland. With the exception of *D. varielongitudum* and *A. diktyoplopus*, none of these species occur in core 7316/06-U-02. *D. varielongitudum* has a single occurrence at 119.0 m, and *A. diktyoplopus* occurs in 23 of the 35 samples from throughout the bedrock section. Williams and Bujak (1985) indicate that *A. diktyoplopus* ranges throughout most of the Eocene (zones NP 13 to NP 20; ca. 26-43 Ma), but *D. varielongitudum* is regarded as restricted to a shorter range (zones NP 11 to NP 14; ca. 50-56 Ma). Thus, occurrences reported by Toxwinius (1986) (<50 Ma) are at variance with those of Bujak et al. (1980) (< ca. 45 Ma), if the single occurrence of *D. varielongitudum* in core 7316/06-U-02 is regarded as biostratigraphically significant. Toxwinius (1986) recognized the base of the Middle Eocene (zone D10; correlative with mid-zone NP 14; 52 Ma) on the basis of the first occurrence of *Cordosphaeridium cantharellum*. According to Williams and Bujak (1985), *Cordosphaeridium cantharellum* has its first occurrence in upper Middle Eocene sediments (zone NP 16), as previously discussed.

In a recent study of the Eocene to Miocene palynology of the ODP Leg 104 on the Vøring Plateau, Manum et al. (1989) defined four dinoflagellate cyst zones within the Middle and Late Eocene. Although none of the eponymous species used in the definition of their zonation were recognized in core 7316/06-U-02, there are some species in common. *Areosphaeridium diktyoplopus* and *Corrudinium incompositum* have their first occurrences at the base of and within (respectively) the *Adnatosphaeridium vittatum* Zone of Manum et al. (1989). The distribution of *A. diktyoplopus* in core 7316/06-U-02 has already been described. *C. incompositum* has a discontinuous distribution in 3 of 22 samples within the interval 156.27 - 71.95 m, and the Barents Sea obviously represents the margin of the zoogeographic province of this species where its stratigraphic range may have little chronostratigraphic significance. According to Manum et al. (1989) the *A. vittatum* Zone is of Middle Eocene age. However, the chronostratigraphic position of the *A. vittatum* Zone was disputed by Goll (1989), who suggested that the zone may also include upper Upper Eocene sediments.

On the Vøring Plateau, *Araneosphaera araneosa* and *Cordosphaeridium cantharellum* have their first occurrences at the base and within the lower part (respectively) of the overlying *Deflandrea* sp. B Zone of Manum, et al.

(1989)(Middle Eocene). Elsewhere, *A. araneosa* is known to have a much longer stratigraphic range (Middle Paleocene to late Eocene, according to Williams & Bujak 1985). In core 7316/06-U-02, *A. araneosa* occurs only at 88.08 m, and it is apparent that the distribution of this species in the Norwegian and Barents seas is not representative of its stratigraphic range as published in the literature.

The above cited literature suggests that core 7316/06-U-02 could be as old as late Early Eocene but is probably no younger than Middle Eocene. Contrary to the published stratigraphic range data, however, IKU's in-house data base indicates that the stratigraphic ranges of both *Cordosphaeridium cantharellum* and *Corrudinium incompositum* extend down into the Early Eocene, and we conclude that the stratigraphic ranges of the dinoflagellate cysts in core 7316/06-U-02 are equally consistent with an Early Eocene age for the base of the core.

The presence of *Areosphaeridium dictyoplokus* and *Cribroperidinium giuseppi* in the upper part of the core further suggest that the core is not younger than Late Eocene (Matsuoka & Bujak 1988, Williams & Bujak 1985). The presence of *Homotrybilum tenuispinosum* in 14 of 31 samples in the interval 199.5 - 88.98 m and *Apectodinium homeomorphum* in 11 of 32 samples in the interval 199.5 - 83.66 m suggests that the core 7316/06-U-02 is not younger than Middle Eocene. According to Williams and Bujak (1985), these two species do not range above zone NP 17 (40 Ma). IKU's in-house data base indicates that the samples between 93.00 m and 71.95 m are no younger than 47-49 Ma, i.e. early Middle Eocene - latest Early Eocene.

Ostracods, diatoms, and radiolarians

The stratigraphic distribution of ostracods, diatoms and radiolaria in core 7316/06-U-02 is shown in Figures 3.2.7 and 3.2.8. Two species of ostracods are recorded in 4 of the 38 samples from the bedrock section of core 7316/06-U-02 (197.76 - 69.03 m), and three species or undescribed morphs of diatoms are recorded in 4 of these 38 samples. Five species or undescribed morphs of radiolaria are recorded in 24 of these samples. Essentially all of the siliceous microfossils have undergone recrystallization from the original opal-A to christobalite, and only large specimens are preserved. The size fractions <125 µm contain no siliceous microfossils. This state of

preservation indicates that these sediments were formerly buried to a much greater depth than at present.

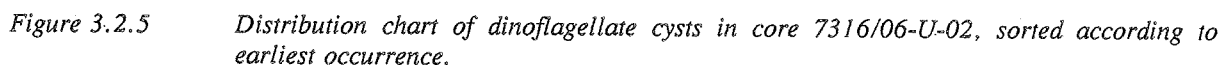
Recrystallized Eocene radiolaria are sporadically common to abundant in this core, but numerous intervals are barren of these microfossils or contain only rare specimens. Specimens previously designated in open nomenclature (*Cenosphaera* sp.) are here identified as *Cenosphaera cristata* on the basis of gross morphological similarity to the latter species in DSDP Leg 38 sediments studied by Dzinoridze et al. (1978). Specimens assigned to this species occurring in sediments below 116 m are tuberculate or have remnants of spines, whereas specimens occurring above 106 m are smooth walled and may merit taxonomic distinction. Large specimens of *Spongodiscus* sp. and *Phacodiscus testatus* also occur in variable frequencies in sediments below 137.40 m and 117.13 m, respectively, and the ranges of these taxa may have future biostratigraphic significance. The most important member of this assemblage is *Calocyclus taiwanii*, which is sporadically present in low frequencies in the interval 193.00 - 126.18 m. Bjørklund (1976) reported the occurrence of *C. taiwanii* as ranging from the *Acronocalpis tumulosa* Zone to the *C. taiwanii* Zone (42-54 Ma, according to Goll 1989). ¶

The diatom, *Triceratium barbadense*, was found in a single sample at 124.28 m. This well-established Middle Eocene marker occurs in DSDP Leg 38 sediments assigned to the *Craspedodiscus oblongus* and *Triceratium barbadense* Zones of Schrader and Fenner (1976), but *T. barbadense* was also observed in Lower Eocene sediments at Site 343 by (Dzinoridze et al. 1978). Thus the range of this species is approximately coeval with that of *C. taiwanii*. Thin, circular, pyritized plates bearing a central depression that occur in the samples from 118.38 m and 142.00 m are here inferred to represent valves of *Craspedodiscus coscinodiscus*, which was recorded in the *C. oblongus* Zone by Dzinoridze et al. (1978). Two specimens of a pyritized diatom in the samples from 142.00 m and 100.74 m bear the general morphology of *Coscinodiscus* sp. 1 of Bettenstaedt (1962), but these specimens are regarded to be too small to merit assignment to this species.

Ostracods are sparse and sporadically distributed in this core. Rare to few specimens of *Cytheridae primitiva* occur in the samples at 137.40 m, 126.18 m

and 109.97 m. Keen (1978) reported the range of this species as restricted to planktonic foraminiferal zones P6 to P11 (Early Eocene to early Middle Eocene). A single ostracod specimen here identified as *Haplocytheridea* sp. cf. *H. debilis* was found in the sample from 88.76 m. Keen (1978) reported the range of the species as restricted to the early Late Eocene, but this identification is too tenuous to warrant such an age assignment for core 7316/06-U-02.

On the basis of these lines of evidence, the interval below 109.97 m is inferred to be no older than Early Eocene (52-58 Ma) and no younger than early Middle Eocene (46-52 Ma). Ostracods and siliceous microfossils offer no reliable age indications for the sediments in core 7316/06-U-02 above this depth.



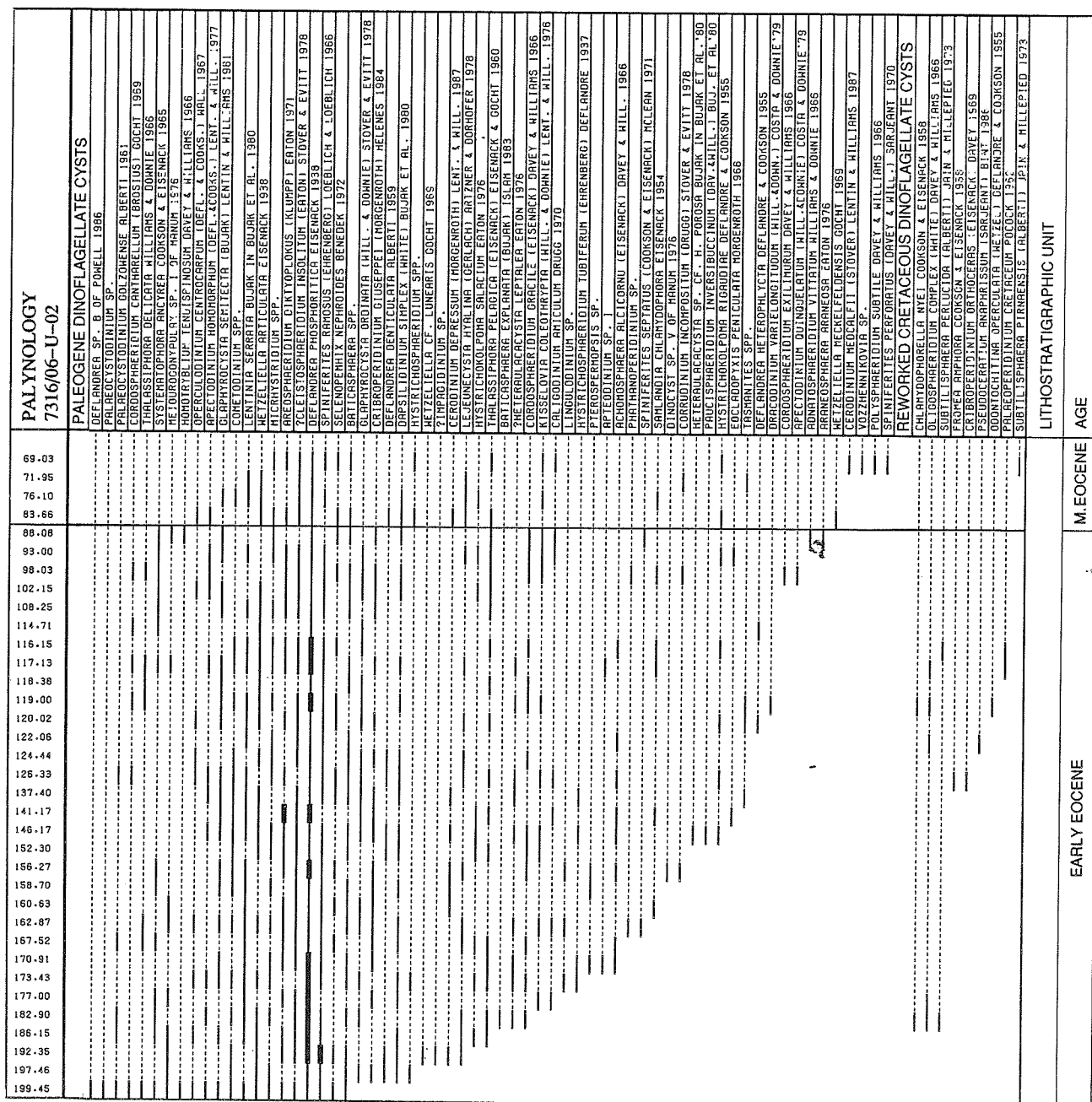


Figure 3.2.6 Distribution chart of dinoflagellate cysts in core 7316/06-U-02, sorted according to latest occurrence.

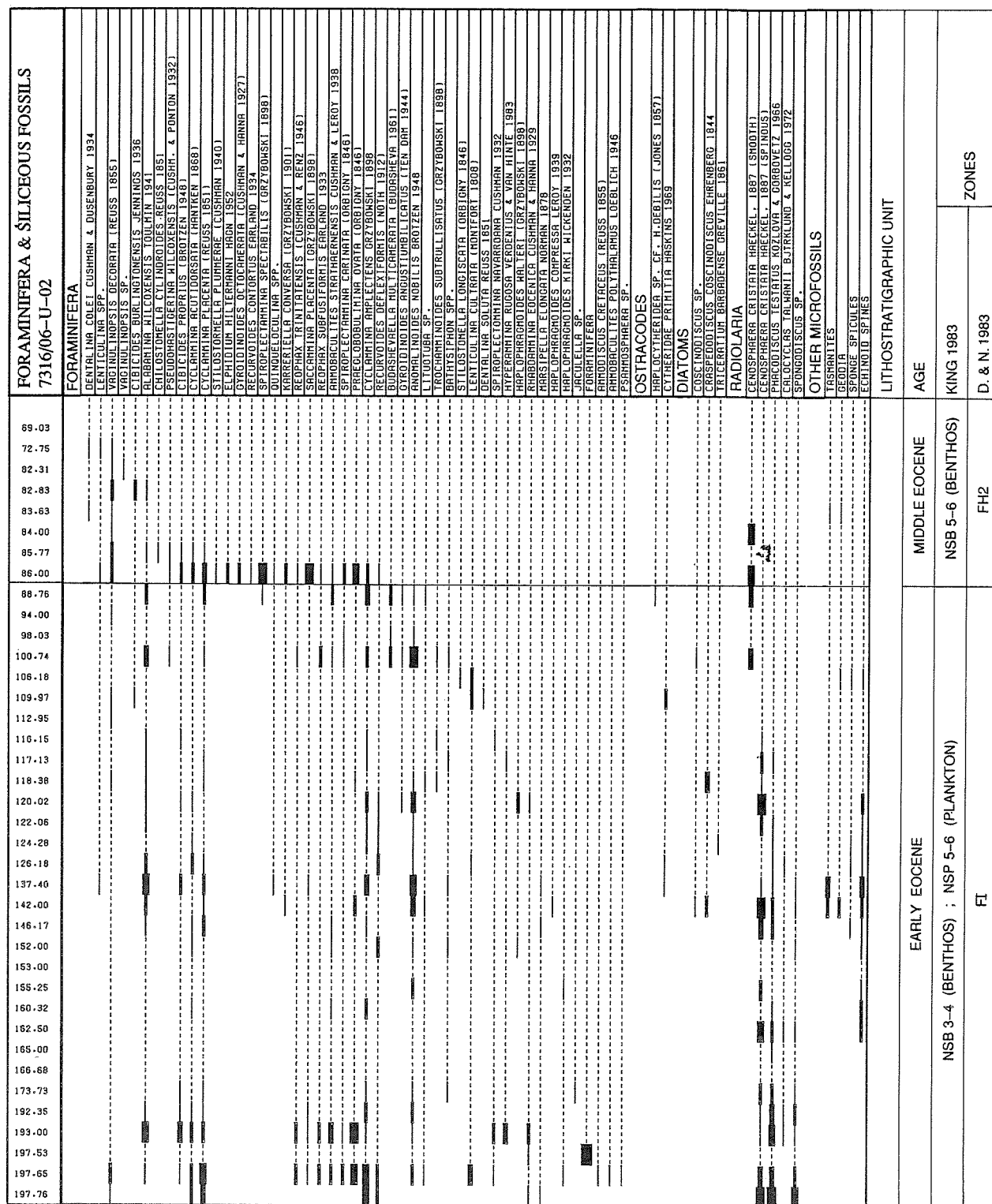
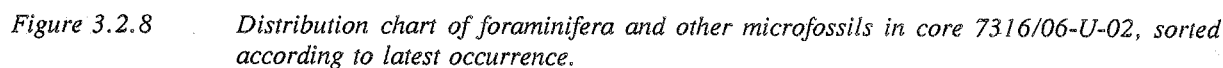


Figure 3.2.7 Distribution chart of foraminifera and other microfossils in core 7316/06-U-02, sorted according to earliest occurrence.



3.2.6 Organic geochemistry

Headspace and occluded gas data were obtained from a total of fourteen canned core samples collected of core 7316/06-U-02 at intervals of 6-10 m between 68.5 m and 199.5 m (Figure 3.2.9). Methane was the dominant gaseous hydrocarbon present in all of the samples, particularly in the headspace gas fraction. The higher molecular weight gaseous hydrocarbons were represented by trace amounts of ethane and propane in the samples above ca. 120 m. Below this, increased concentrations of ethane and propane were accompanied by traces of butane and C₅₊ gaseous hydrocarbons. The highest concentrations of higher molecular compounds tended to occur in the lower, more argillaceous half of the cored succession.

Methane carbon isotope data were obtained from two samples at 85.58 m and 169.12 m. Data could not be obtained for $\delta^{13}\text{C}_{\text{ethane}}$ and $\delta\text{D}_{\text{methane}}$ due to insufficient material. The $\delta^{13}\text{C}$ data tend to indicate a thermogenic origin for the methane, although contamination by, or oxidation of biogenic methane cannot be excluded as an explanation for the observed data.

A total of seventy-four core chip samples taken between 69.03 m and 199.31 m were analysed for total organic carbon (TOC) content (Figure 3.2.10). TOC contents were found to vary between 0.29 wt% and 1.19 wt%, with most values falling between 0.5 wt% and 0.9 wt%. The lowest TOC contents (i.e. 0.29-0.56 wt%) occur over the interval at 69-84 m, in conjunction with the occurrence of slightly coarser sediments.

Rock-Eval pyrolysis data were obtained for twenty-nine samples and suggest that there is little variation in the nature of the kerogen present in the cored succession (Figure 3.2.10). The low S₁ peak concentrations suggest that no significant amounts of soluble hydrocarbons are present in these rocks. In addition, relatively similar S₂ peak values and hydrogen indices of between 38 and 83 mg/g TOC, together with low Tmax values of 415 to 429 °C indicate that the kerogen in the rocks has no hydrocarbon generation potential and may be characterised as thermally immature type III to III/IV kerogen.

Pyrolysis-gas chromatograms of eight selected samples were similar in appearance. These were dominated by peaks representing aromatic compounds, with aliphatic peaks being poorly developed. The appearance of these pyrolysis-gas chromatograms is typical of degraded and inert kerogen.

The Rock-Eval and pyrolysis-gas chromatography data are supported by the dominance of woody and coaly debris in the kerogen. A slightly increased content of liptinitic material in the two samples at 156.23 m and 199.45 m is not reflected in Rock-Eval data from adjacent samples, which suggests that this material may be either oxidised or only of local importance. Evidence for recycling of the kerogen from microscopic observations suggests that the poor quality of the kerogen may be due to degradation during erosion and transport.

For the most part, the nature of the kerogen in the cored rocks caused some problems in obtaining reliable vitrinite reflectance data. However, there is relatively little spread in the mean values which vary between 0.35% and 0.40%. These values, together with the yellow to yellow-orange sporinitite fluorescence observed under ultra-violet light and thermal alteration index values of less than 2.0, suggest that the kerogen in these rocks is thermally immature.

Extraction of five samples selected from the cored bed-rock succession yielded rather poor concentrations of extractable organic matter (EOM). The concentration of total EOM obtained from the samples varied from 199 ppm to 434 ppm, of which the major part was represented by asphaltene (30-50%) and non-hydrocarbon (19-56%) compounds (Figure 3.2.11). With the exception of two samples at 86.38 m (35%) and 154.23 m (60%), hydrocarbon compounds account for less than 20% of the total EOM.

The saturated and aromatic hydrocarbon gas chromatograms of the sample at 86.38 m indicate that the hydrocarbons from this sample consist almost totally of higher molecular weight unresolved material. On the saturated hydrocarbon gas chromatograms of the remaining four samples at 144.59 m, 154.23 m, 193.81 m and 194.95 m, the n-alkane peaks show a marked bias towards higher molecular weight compounds and a pronounced odd-carbon number predominance. These two features are reflected in low nC_{17}/nC_{27} ratios and high carbon preference indices, respectively (Figure 3.2.12). The pristane/phytane ratios from the samples are uniformly low (0.5 - 0.8), although this could reflect evaporative loss of material during analysis rather than source input. Almost no recognisable peaks were present on the gas chromatograms of the aromatic hydrocarbon fraction.

Samples of the asphaltene fractions from two samples at 193.81 m and 194.95 m were further analysed using pyrolysis-gas chromatography. The pyrolysis-gas chromatograms of the two samples show similar, well-developed

homologies of aliphatic compounds which extend beyond C_{25} , together with a prist-1-ene peak and a prominent peak(?) representing a currently unidentified compound which elutes prior to the (M+P)-exylene peak. These two pyrolysis-gas chromatograms are similar to those of most of the other asphaltene samples analysed from other Bjørnøya West cores, suggesting a common source for these bitumens.

Three extract samples at 86.38 m, 144.59 m and 193.81 m were analysed using combined gas chromatography-mass spectrometry. The shallowest of the three samples contains both sterane and hopane distributions typical of early "oil window" thermal maturity, i.e. equivalent to a vitrinite reflectance of around 0.5%. The other two samples contain less mature hopane distributions. A high relative abundance of $17\beta(H), 21\beta(H)$ -hopanes and of the 22R isomer of the C_{31+} hopanes may indicate a thermal maturity equivalent to a vitrinite reflectance of 0.3-0.4%, i.e. similar to that measured from the kerogen in this core. The lack of C_{30} hop- $\Delta^{13(18)}$ -ene suggests that the peak which co-elutes with the C_{29} $17\alpha(H), 21\beta(H)$ -hopane is not the C_{29} hop- $\Delta^{13(18)}$ -ene, but is in fact another unidentified compound.

4

Comparison with the mass fragmentograms of the samples from core 7316/03-U-01 suggests that the two samples at 144.59 m and 193.81 m also contain the two unusual triterpanes described from this core. At least one of these compounds is also thought to be present in the sample at 86.38 m where the prominence of the C_{29} 20R sterane peak is thought to be partly due to coelution of one of these unusual triterpanes. These compounds are thought to represent an input from terrestrial organic matter.

In conclusion, the analysed bedrock succession is characterized by a poor to fair content of thermally immature, type IV kerogen with no hydrocarbon generation potential. Traces of rather immature bitumen occur in the core which show evidence of being derived from a source containing terrestrial organic matter. This bitumen is similar in composition to that described from cores 7316/03-U-01 and 7317/10-U-01. The origin of the bitumen is unlikely to be indigenous and the bitumen is therefore considered to be either migrated or reworked from an eroded bituminous horizon.

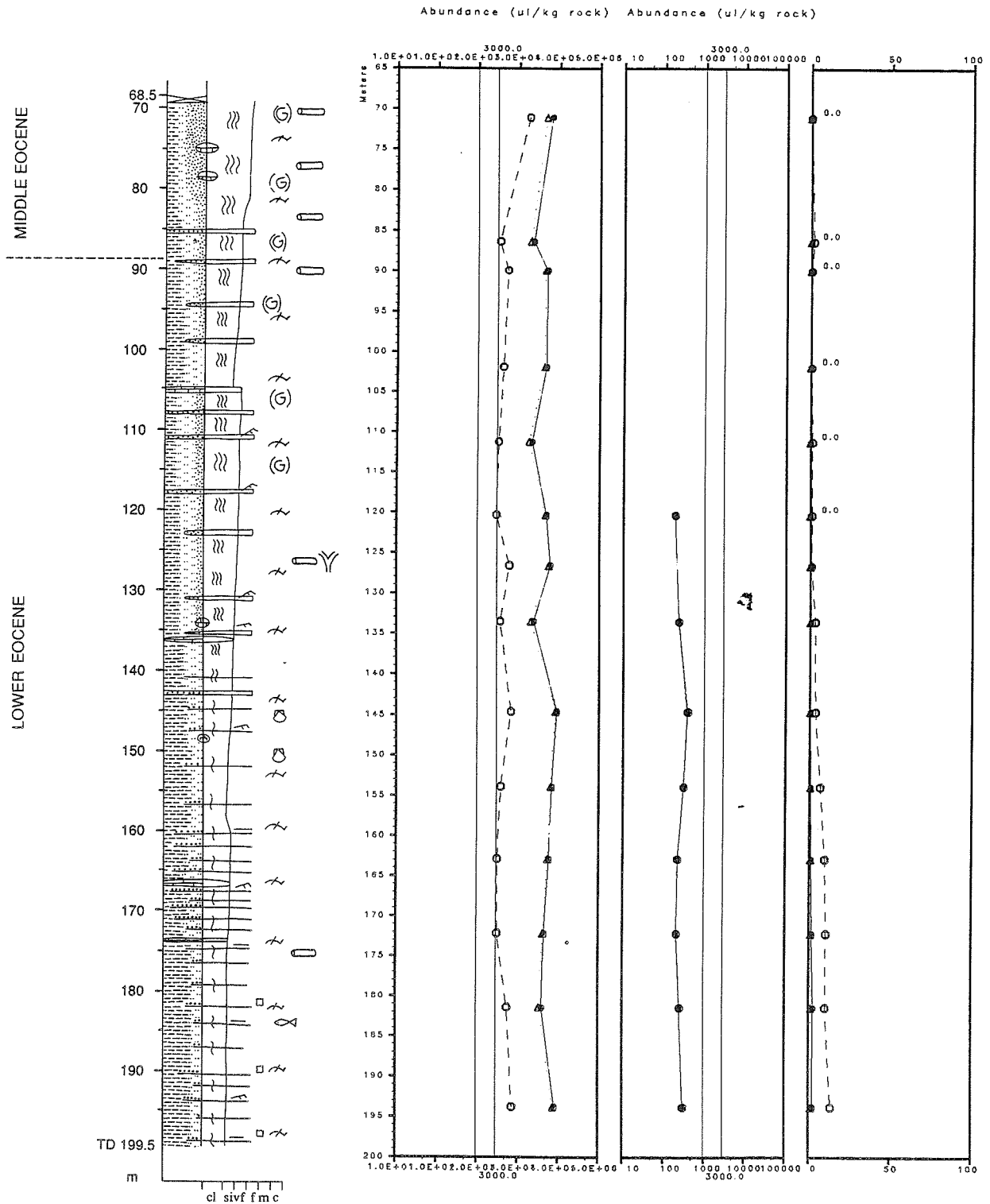


Figure 3.2.9 Yield and composition of headspace/occluded gas of core 7316/06-U-02.

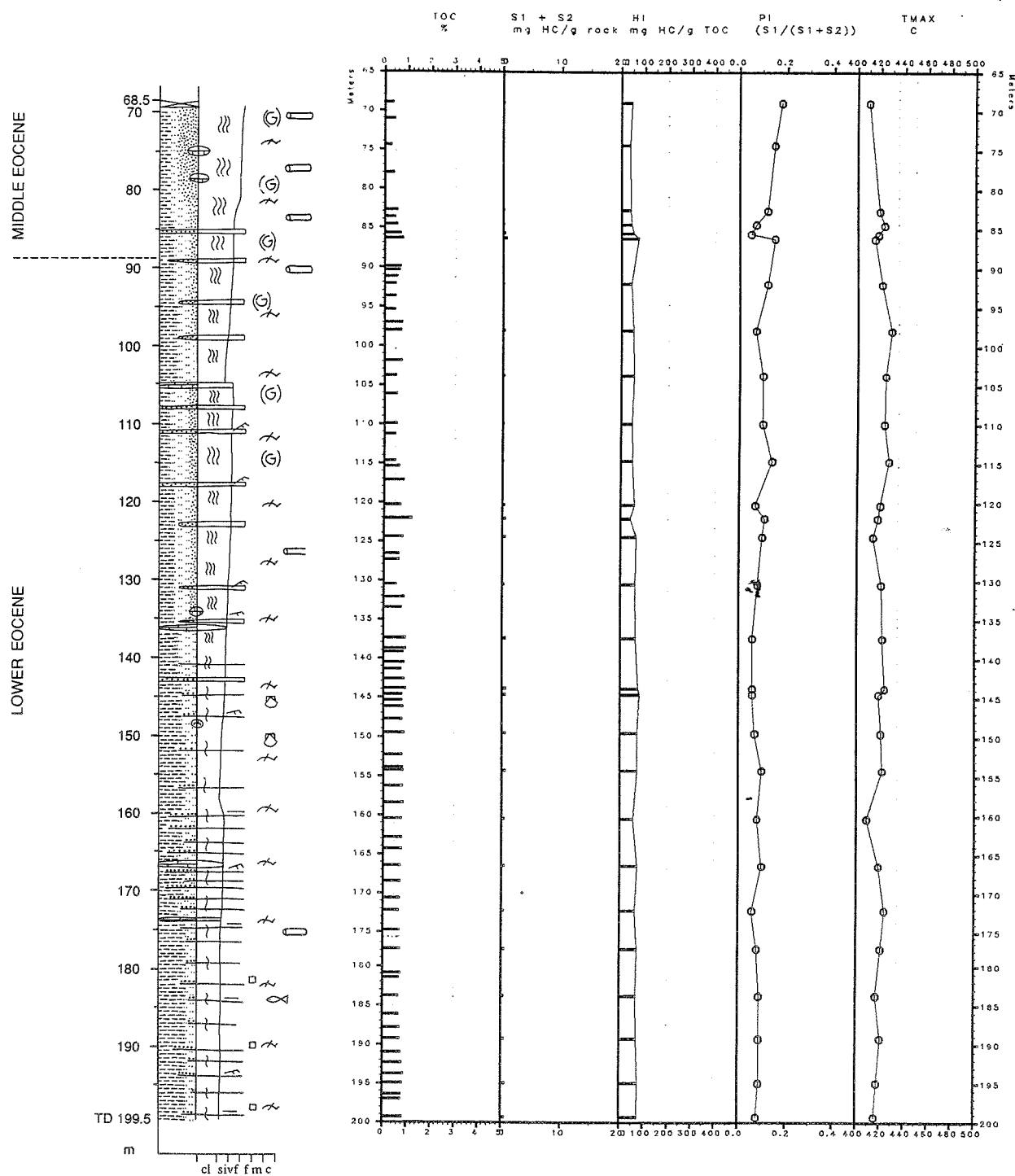


Figure 3.2.10 Total organic carbon content and Rock-Eval data of core 7316/06-U-02.

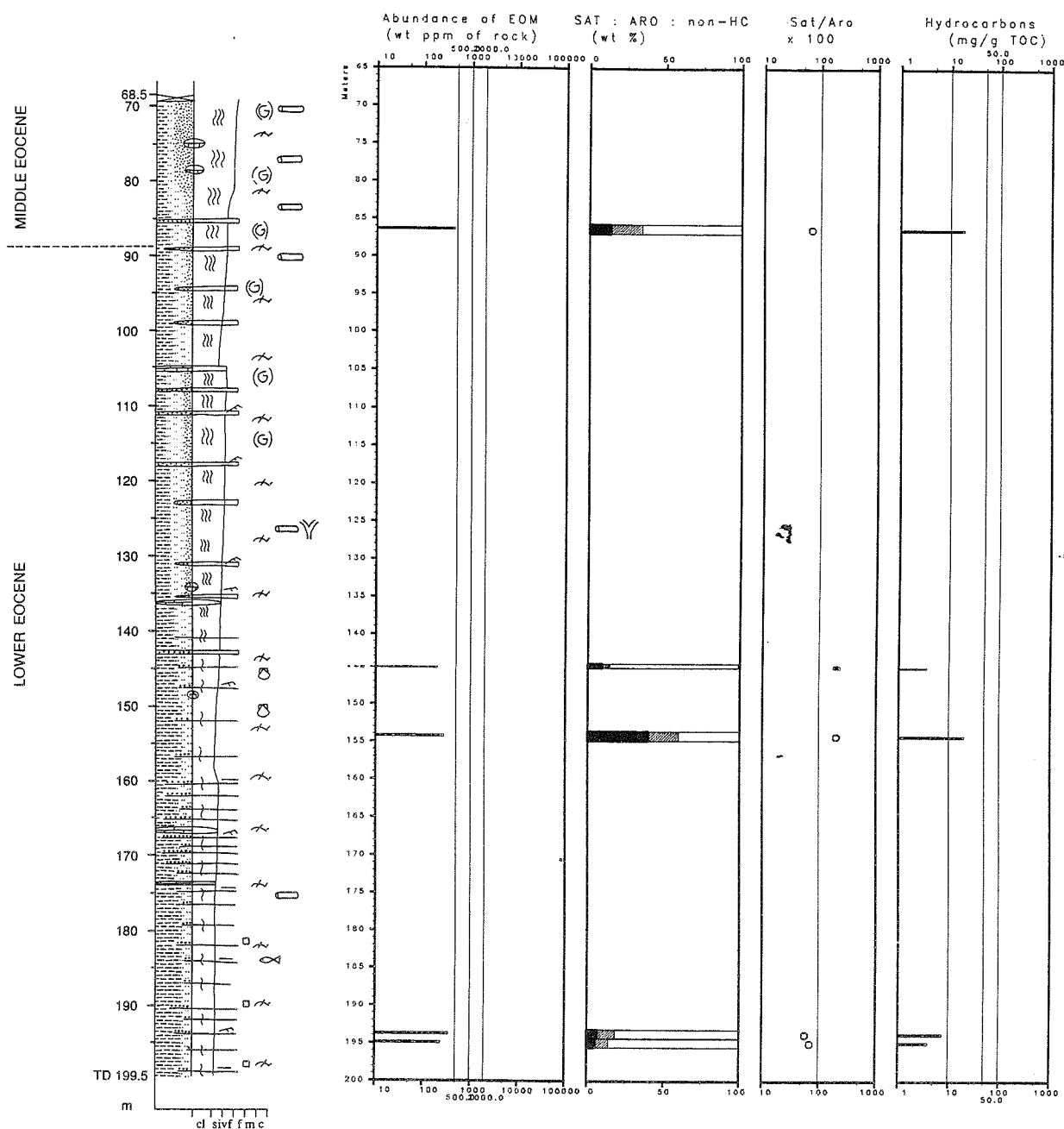


Figure 3.2.11 Extract data for core 7316/06-U-02.

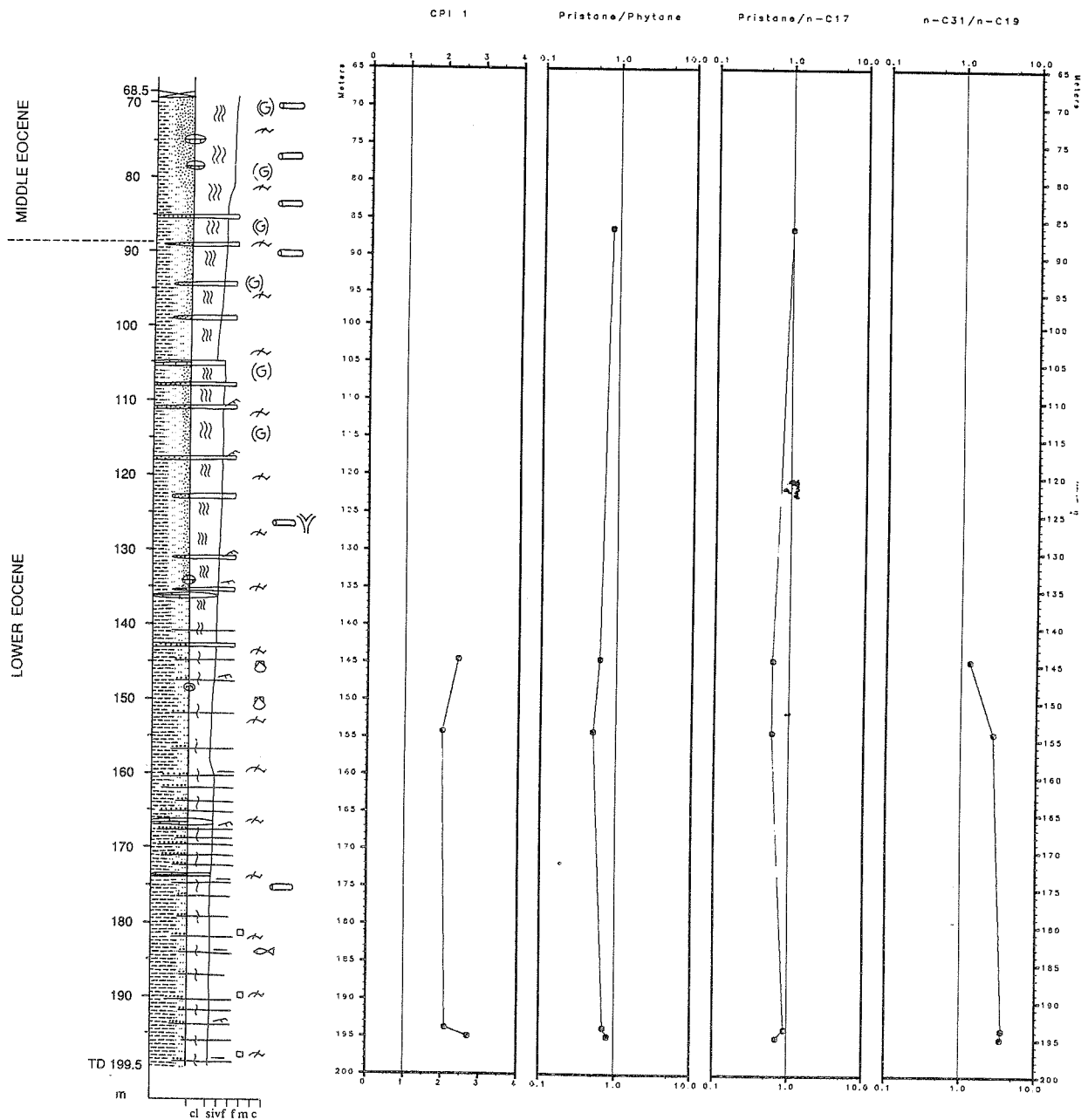


Figure 3.2.12 Compound ratios from gas chromatograms of saturated hydrocarbon fractions from core 7316/06-U-02.

3.2.7 Depositional environment

The siltstones and shales/claystones of core 7316/06-U-02 contain marine fossils indicating an open marine depositional environment (Figure 3.2.13). In addition, the ichnological data support an open marine setting for the sediments (see Appendix A.7).

The lithology of the core is dominated by clayey and silty sediments, reflecting deposition predominantly from suspension. In addition, thin beds of very fine-grained sandstone characterized by sharp bases and horizontal and/or ripple laminae occur throughout. The sandstone beds reflect pulses of higher energy events and deposition during waning flows. Walker (1984) and Bouma et al. (1983) indicated that sedimentary facies comparable to the core are commonly found in inner shelf areas. The sandstone beds may represent storm deposits. Odin and Fullagar (1988) indicated that glauconitization takes place in water deeper than 60 m. The pelloidal glaucony in the sandstone beds of the core may, if they formed in situ, thus indicate water depths greater than 60 m.

7

Walker (op cit.) and Bouma et al. (op cit.) indicated that in an inner shelf environment the thickness of sandstone beds should increase in a shoreward direction, i.e. regressive deposits should form coarsening upward succession. The core shows an overall, but composite coarsening upward grain size pattern. Each of the four coarsening upward intervals may represent shoaling upward successions in an overall upward shoaling trend. The regressive interpretation is supported by the ichnological data with low diversity and density ichnofabric below about 140 m. Above this, a trend is seen toward high ichnodiversity dominated by *Helminthopsis* burrows. These represent a muddy shelf bottom deposit feeder community. Regressive stacking pattern is characteristic for high-stand deposits (Posamentier et al. 1988).

The lithology and fossil dating of the sediments of the core indicate that the rocks could be equivalents to the Torsk Formation of the Sotbakken Group of Worsley et al. (1988).

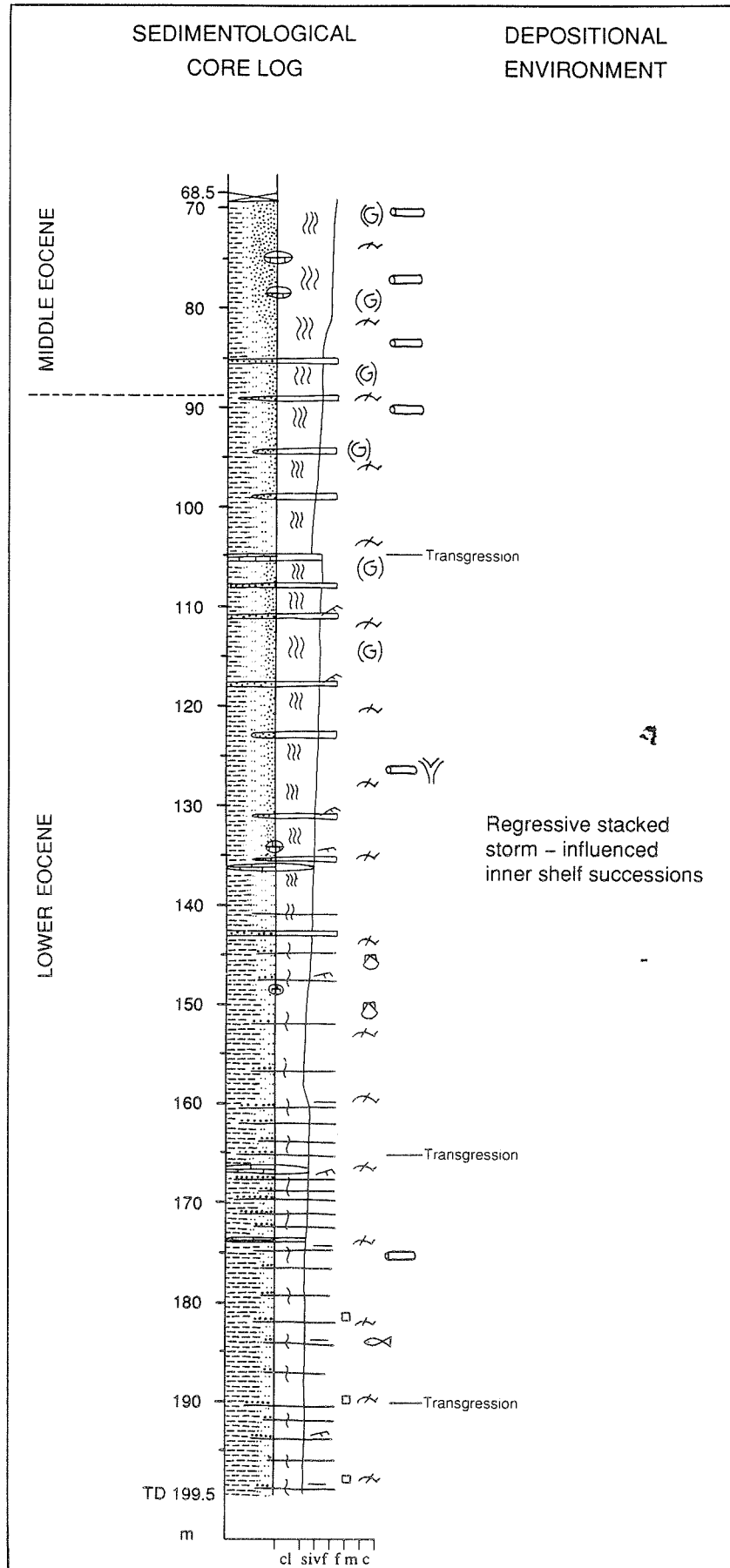


Figure 3.2.13 Interpreted depositional environment of core 7316/06-U-02.

3.3 Core 7316/06-U-01

3.3.1 Summary

Core 7316/06-U-01 was drilled about 1.5 km south of core 7316/06-U-02. These cores are separated by an angular unconformity possibly representing a major hiatus from ?Early Oligocene to Early Miocene.

Within the Quaternary sediments a 15 m thick section of Cretaceous rock was cored (Figure 3.3.1). This represents a huge block that has been lifted off the bedrock somewhere east of the drill site and transported by a glacier to its present position. It comprises 80 km², but could be discountuous.

Due to drilling problems, the core recovery in the bedrock was only 23%. The core is subdivided into three units based on lithology and log response as well as correlation with the seismic record.

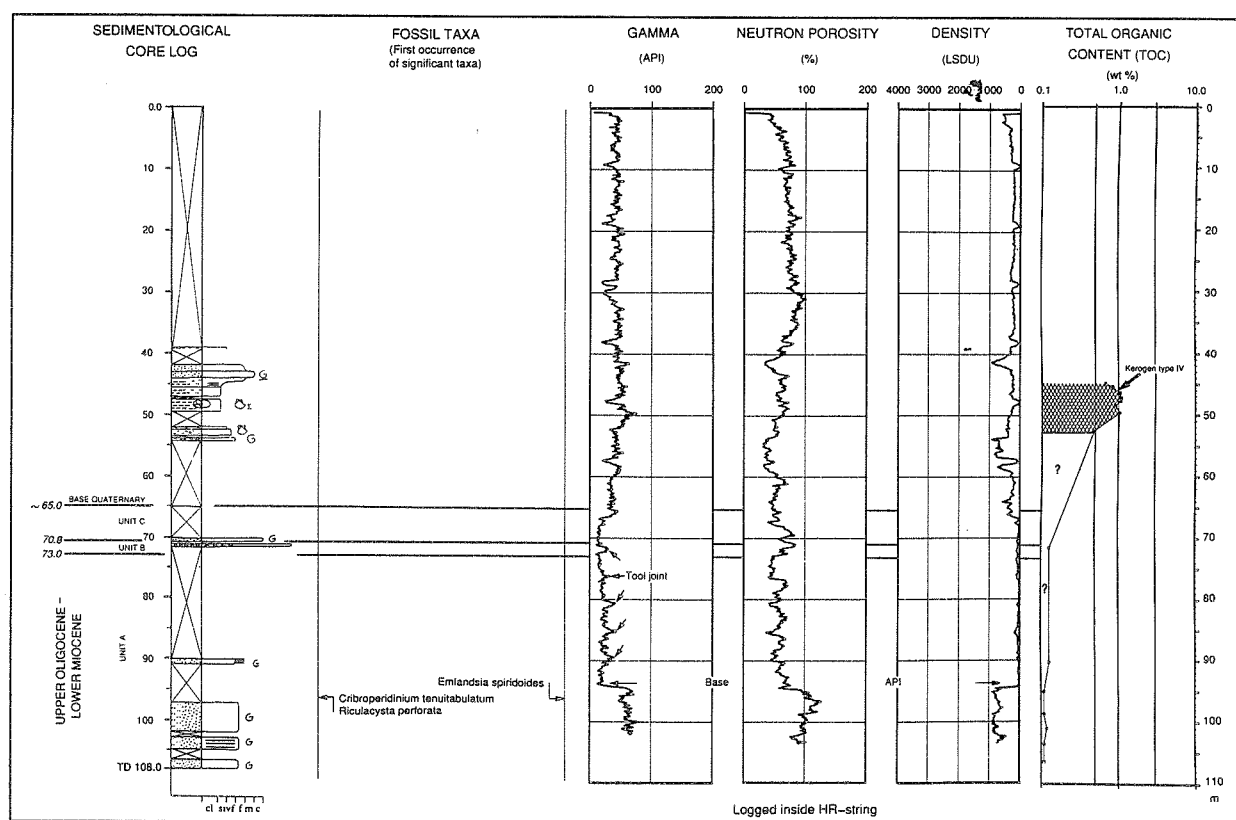


Figure 3.3.1 Summary of main data of core 7316/06-U-01.

The lower Unit A consists of immature grey-green very fine-grained sandstone with a minor glauconite content. The sandstone is not consolidated and only faint lamination is observed. This unit is interpreted to continue up to the base of a thin polymict conglomerate bed at 73.0 m that constitutes Unit B. The conglomerate may form the basal deposits of a wide channel which has been filled by coarse, well rounded sandstone (Unit C). Only 4 cm were recovered of Unit B and 48 cm of Unit C, however the log responses indicate that Unit B may have a thickness of about two metres, while Unit C may be several metres thick; the channel is 10-12 m deep based on the seismic interpretation.

In situ fossils which are only found in Unit A are of ?Middle Oligocene to Middle Miocene age. Deposition in marine environments in an upper slope setting seems most probable.

The glaciotectionic megablock is 15 m thick at the drill site and consists of a marly shale in the lower part, succeeded by light grey calcareous silty/sandy claystone with *Inoceramus* prisms and glauconitic sandstone with a high clay content. Both foraminifera and palynomorphs date the sediments to Late Albian - Early Cenomanian.

Samples from the coarse-grained bedrock of Tertiary age proved to contain almost no organic matter at all. In contrast, rock samples from the glacially-emplaced slab of Cretaceous material contained fair amounts of organic matter. However, this organic matter proved to be thermally immature and inert type IV kerogen.

3.3.2 Seismic interpretation

Cores 7316/06-U-01 and -02 were located in the middle part of the surveyed area, about 1.5 km from each other on seismic line IKU-1650-89 (Figure 3.3.2). The glacial deposits were 65 m thick and contained a large megablock of Late Cretaceous sediments transported by an advancing glacier. The allochthonous rock body appears on the analog seismic data as shadow-like acoustic reflections. More details are given in Section 3.3.8 and in Appendix 13, Appendix Volume 2, which deals with the Quaternary deposits.

The uppermost part of the pre-Quaternary sediments seems to be slightly disturbed. There might be a small channel, about 250 m wide and 10-12 m deep, that was penetrated at the top of the core. The conglomerate

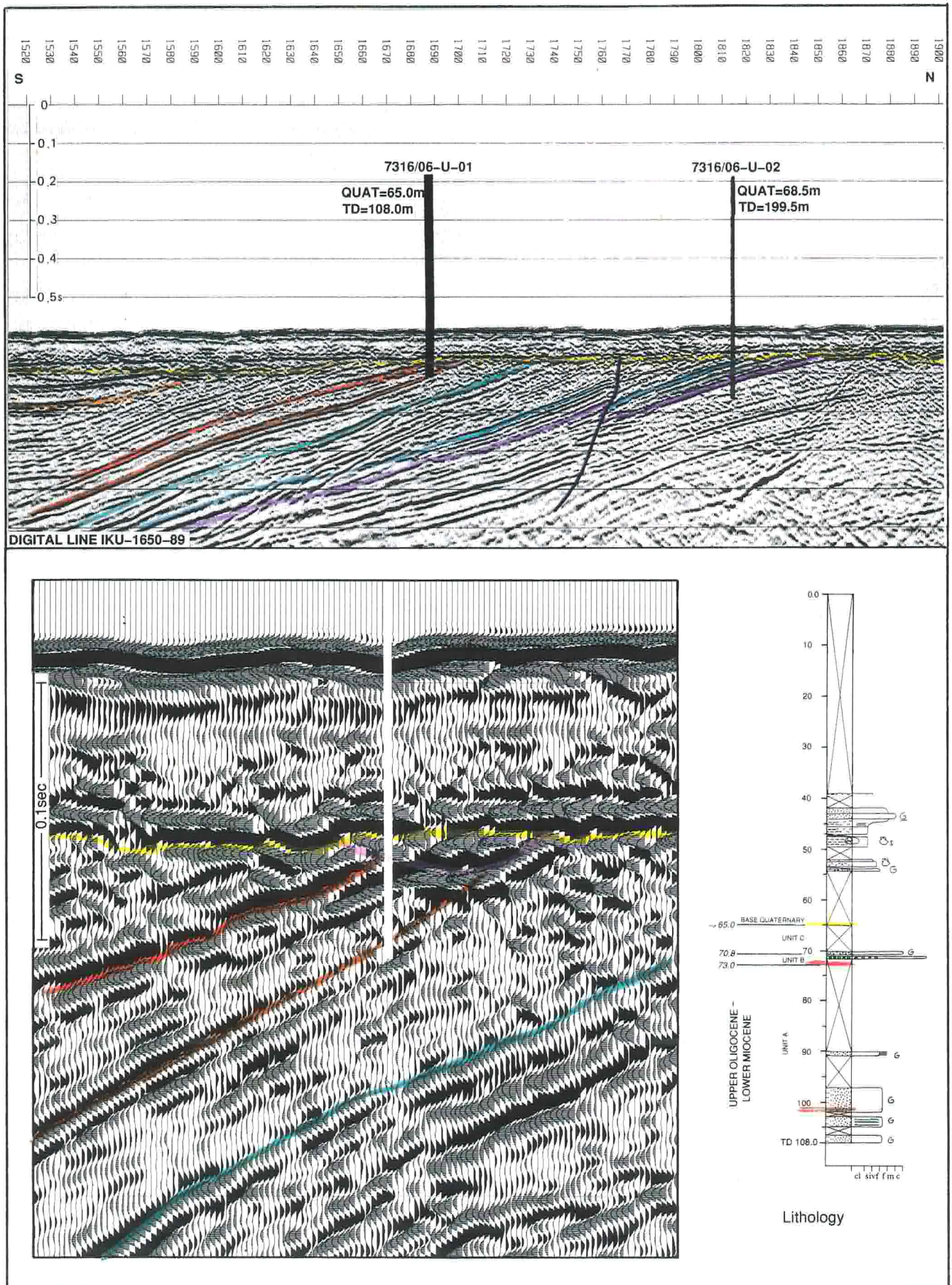


Figure 3.3.2 Seismic correlation of core 7316/06-U-01.

recovered at about 71.6 m seems from the logs to extend down to about 73 m. The sand recovered just below 70 m could be a channel fill, covering the interval 71.6 - 65 m of the borehole. Unit A of the core is probably found immediately below the channel and seems to comprise a boundary giving a relatively strong seismic reflection (at about 101 m in Figure 3.3.2). The two seismic units above and below both have an irregular reflection pattern, indicating rapid deposition as sediment mass flow or related processes. They also show some thickness variations.

Drill sites 7316/06-U-01 and -02 lie close to a northerly trending fault. The tie across this fault and a more regional seismic interpretation are presented in the discussion chapter, Section 4.1.

Table 3.3.1 Correlation of seismic data and core 7316/06-U-01. Interval sound velocities noted "seismic" appear from this seismic-core correlation.

Depth below seabed (m) (msec)	Core unit	Thickness of seismic unit (m) (msec)		Interval sound velocity (m/sec)		
				Seismic	Core	Log
65.0 75	Quaternary	65	75	1730	1730	-
73 84.5	B + C	8	9.5	1800	-	-
101 115.5	A	28	31	1800	-	-
TD108 123	A	7	7.5	est1800	-	-
Water depth: 427 m (583 msec) Depth below sea level: 535 m (706 msec)						

3.3.3 Sedimentology

A total of 43 m of bedrock was penetrated between 65.0 m and 108.0 m below seabed, and additionally a 15 m thick megablock of sedimentary rocks of Cretaceous age occurs as a large (800 km²) transported slice complex of rock within Quaternary sediments. Core from more than half of the megablock was recovered (see Section 3.3.8).

Due to drilling problems the core recovery in the Tertiary bedrock was only 23%. The recovered sediments are subdivided into three units. The basal Unit A consists of dark grey green very fine-grained sandstone and is sampled below 90.2 m; this unit extends upwards to the overlying conglomerate probably at about 73 m. Unit B is represented by a 4 cm long core piece of polymict conglomerate, collected at 71.6 m. Uppermost in the core, Unit C consists of coarse-grained sandstone that was sampled between 70.31 and 70.79 m.

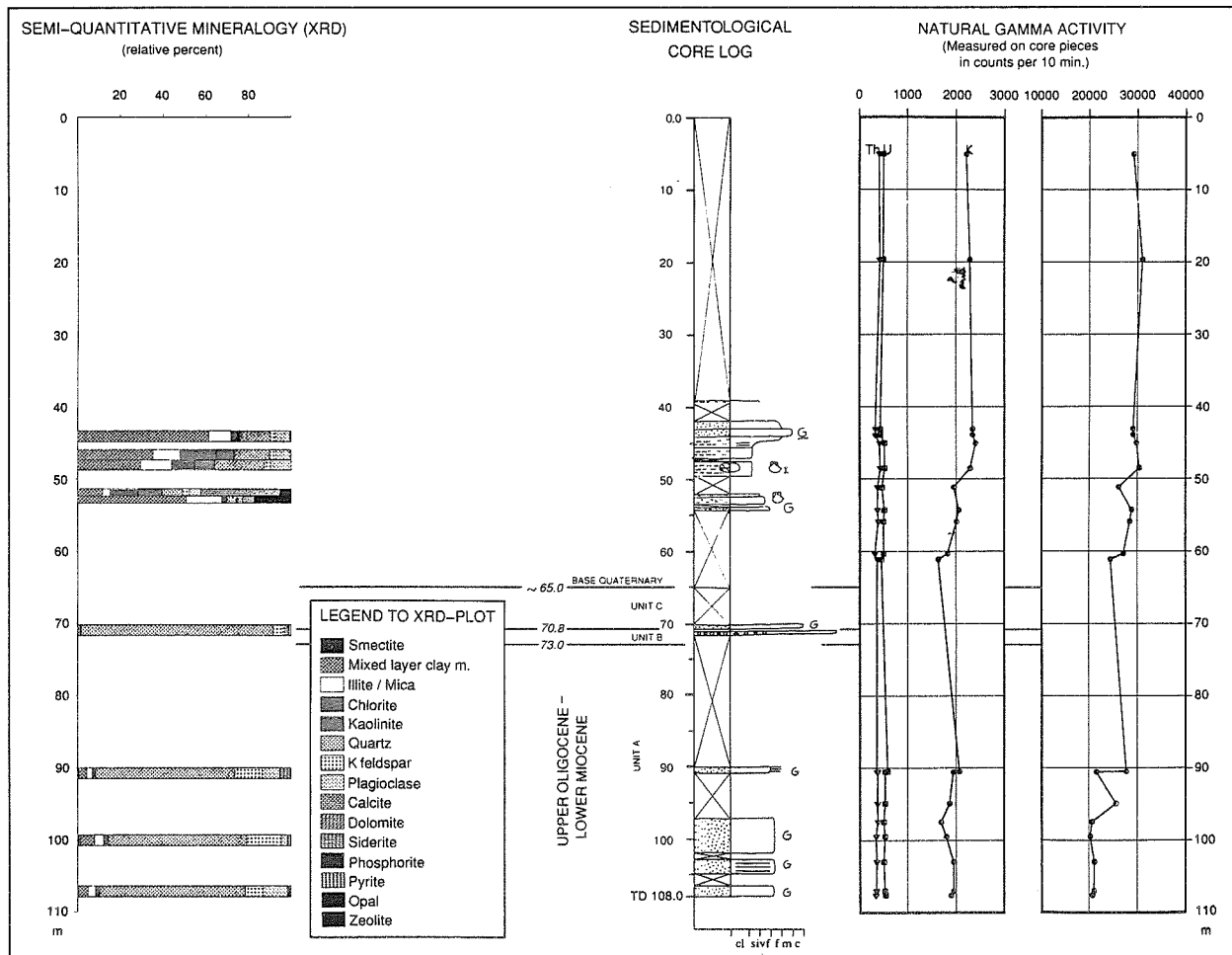


Figure 3.3.3 Lithologic column, mineralogical analysis and laboratory measurement of spectral gamma values of core 7316/06-U-01.

Unit A is sampled from 108.0 m (TD) to 90.2 m, but the unit probably continues somewhat higher. The weak decrease in gamma radiation upwards (although measured inside the drill string) indicates coarsening upwards successions. The neutron porosity that is also measured inside the drill string shows a marked reduction at the 76 m level followed by an increase at 73 m. Unit A may extend up to either the log response change at 76 m or 73 m and also further downward from the base of the drilling.

The unit consists of dark grey-green silty, very fine-grained sandstone with a minor content of glauconite (Figure 3.3.4 a). The only sedimentary structure observed is faint lamination.

Unit B is only represented by a 4 cm long core piece of polymict conglomerate. Due to drilling problems this was the only recovered piece of the unit. As discussed for the underlying unit the transition between these units is highly uncertain. On the log the increased reading in neutron porosity at 73 m is followed by a further increase at ca 71.6 m and decrease at 70.8 m. Given the uncertainties in the tie between the drilled core and the logs the interval between 73 m and 70.8 m may represent the cored conglomerate.

The polymict conglomerate consists of rounded clasts of quartz, sedimentary rock fragments and glauconite. The sample is dolomite cemented.

Unit C is sampled between 70.31 m and 70.79 m and consists of coarse grained sandstone (Figures 3.3.4 b and c). The sand grains are well rounded, and the sandstone also contains rock fragments with glauconite.

Grain size distribution analyses of three samples from Unit A and one from Unit C are presented in Figure 3.3.5. Only the fraction larger than 45 μ m has been analysed. Unit A shows dominance of coarse silt to very fine-grained sandstone, while a clear difference is seen to Unit C that is totally dominated by coarse sand without any content of fine-grained material.

Petrography

Unit A of core 7316/06-U-01 contains very fine-grained, loosely consolidated sandstones of sublitharenite composition, showing moderate sorting, with subrounded to subangular grains shapes (Figure 3.3.4 a). The sandstone has a high porosity. The clastic minerals are quartz, plagioclase, alkali-

feldspar (microcline), glaucony and accessory contents of white mica, chlorite, epidote - clinozoisite, zircon, tourmaline, rutile, sphene, opaques. Lithic fragments include chert (fibrous chalcedony).

The rounded shape of the glaucony grains may reflect both primary growth features and later sedimentary rounding. Glaucony is interpreted as clastic in this core both from the scattered appearance, and from the data from Unit C described below.

The mineralogy of Unit A suggests that the sediments represent both metamorphic and sedimentary source rocks. The rounded appearance of accessory heavy minerals also indicates multistage sedimentary evolution histories (recycling).

The polymict conglomerate of Unit B contains coarse sand and pebbles up to 2 cm in diameter and all pebbles are well rounded. The phosphate pebbles occur as oval shaped up to several cm long clasts, and rounded clasts. Chert pebbles are rounded with fibrous structures within the grains. Some contain faint fossils (the lithology resembling the Permian Kapp Starostin Formation). Centimetre-sized rounded quartz pebbles also occur, including both aggregates of coarse grains with seriate boundaries and single grains with strong undulatory extinction. The latter type shows evidence of strong deformation with shearing by the orientation of domains with low-angle boundaries and the appearance of fractures with trains of minute subgrains.

The matrix sand is quartz rich medium to coarse grained with subrounded and rounded grains. The quartz grains show undulating extinction patterns, some have developed low angle boundaries, indicating different degrees of recovery following deformation. Other clastic grains are glaucony (brownish heterogeneous and green mature grains), microcline (crosshatched twinning), plagioclase, subhedral K-feldspar without twinning (orthoclase?), epidote, white mica, opaques and tourmaline (?amphibole). The conglomerate is completely cemented by dolomite (or ankerite) characterized by rhomb shaped grains.

The coarse-grained sand (Unit C) is sedimentologically mature both with respect to texture and composition (quartz-rich sublitharenite). This nearly unconsolidated rock is characterized by rounded and well sorted grains and has a high porosity (Table 3.3.2, Figure 3.3.4 b). The main clastics are quartz, various lithic fragments and feldspar (microcline). The quartz grains show evidence of strong pre-depositional deformation by

undulating extinction and low-angle boundaries. The heavy minerals staurolite and epidote have been detected in the one thin section examined (70.75 m). Different types of composite clasts occur: chert, polygranular quartz, feldspar and quartz, and carbonate cemented aggregates of quartz and glaucony and sometimes feldspar and epidote (Figure 3.3.4 c).

Table 3.3.2 Modal analysis of sandstones from core 7316/06-U-01.

SAMPLE NO	Grain-size								Q + F + L = 100%		
		% Quartz	% K-fsp	% Plagioclase	% Glaucony	% Div. lithics	% Matrix	% Porosity	Q	F	L
7316/06-U-01											
70.75 m	C	52.0	1.3	0.2	-	6.0	0.2	40.4	87	3	10
90.19 m	F	37.1	7.2	1.9	2.7	11.5	1.6	37.9	61	15	24
97.50 m	VF	42.6	5.7	1.8	2.1	13.3	0.3	34.1	65	11	24

Abbreviations:

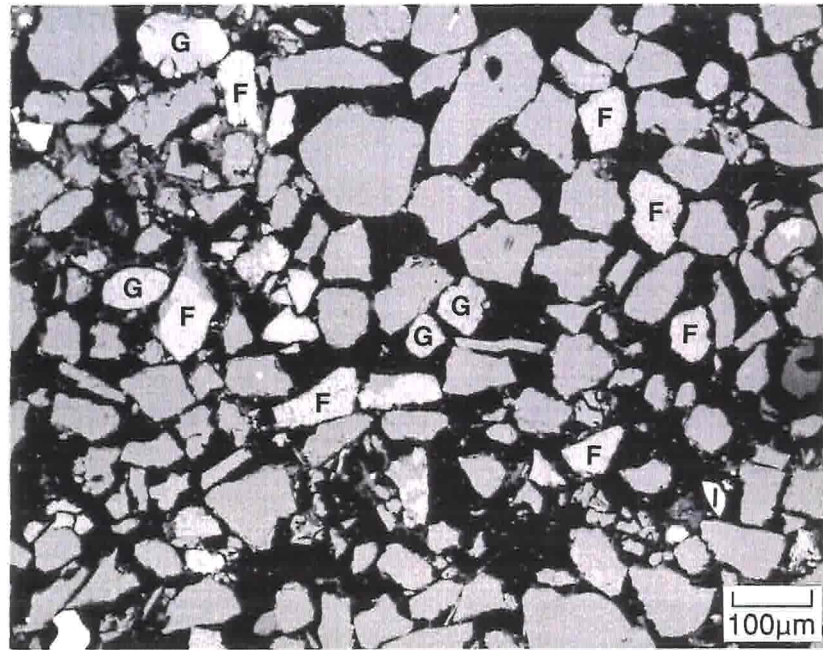
C = coarse
F = fine
VF = very fine

Q = quartz F = feldspar L = lithics

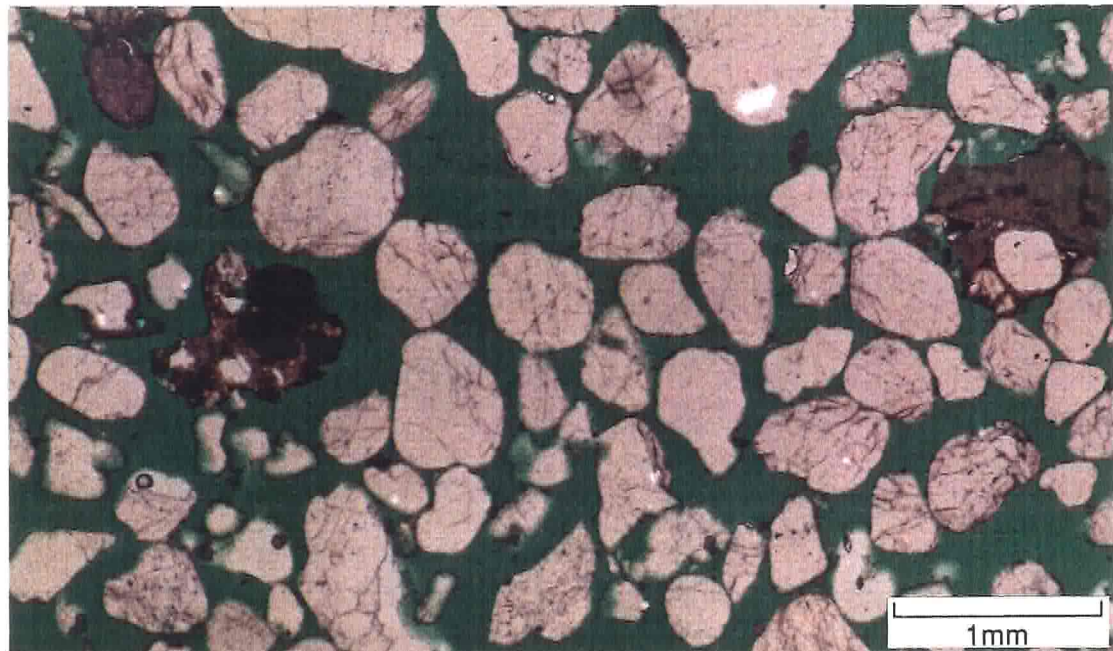
The glaucony in this sample originates from an older carbonate cemented sedimentary rock (resembling Cretaceous glaucony bearing sediments elsewhere on the Barents Shelf, e.g., the Cretaceous megablock in the present core) and it was not formed in situ at the present location. Other evidences of sedimentary source rocks include chert which resembles lithologies of the Permian Kapp Starostin Formation. Quartz- feldspar clasts and some of the polygranular coarse quartz grains indicate the presence of metamorphic source rocks, e.g. gneiss, quartzite and schists. This is supported by the presence of staurolite which is typical for medium grade metamorphic schists and by the presence of epidote.

Figure 3.3.4 Thin section photos from the sandstones of core 7316/06-U-01.

a)
90.19 m. Back scattered electron image showing sandstone texture from Unit A. Grey grains are quartz. F marks the K-feldspar and G marks the glauconite grains.



b)
70.75 m.
Microphoto, plane polarized light, of coarse-grained sandstone, Unit C. The high porosity is shown by green coloured epoxy. Light grains mainly represent quartz. Two glaucony-containing lithic fragments are dark (see c)).



c)
Close up of lithic fragment with rounded glaucony (dark), quartz and feldspar (white) in a carbonate cemented matrix.



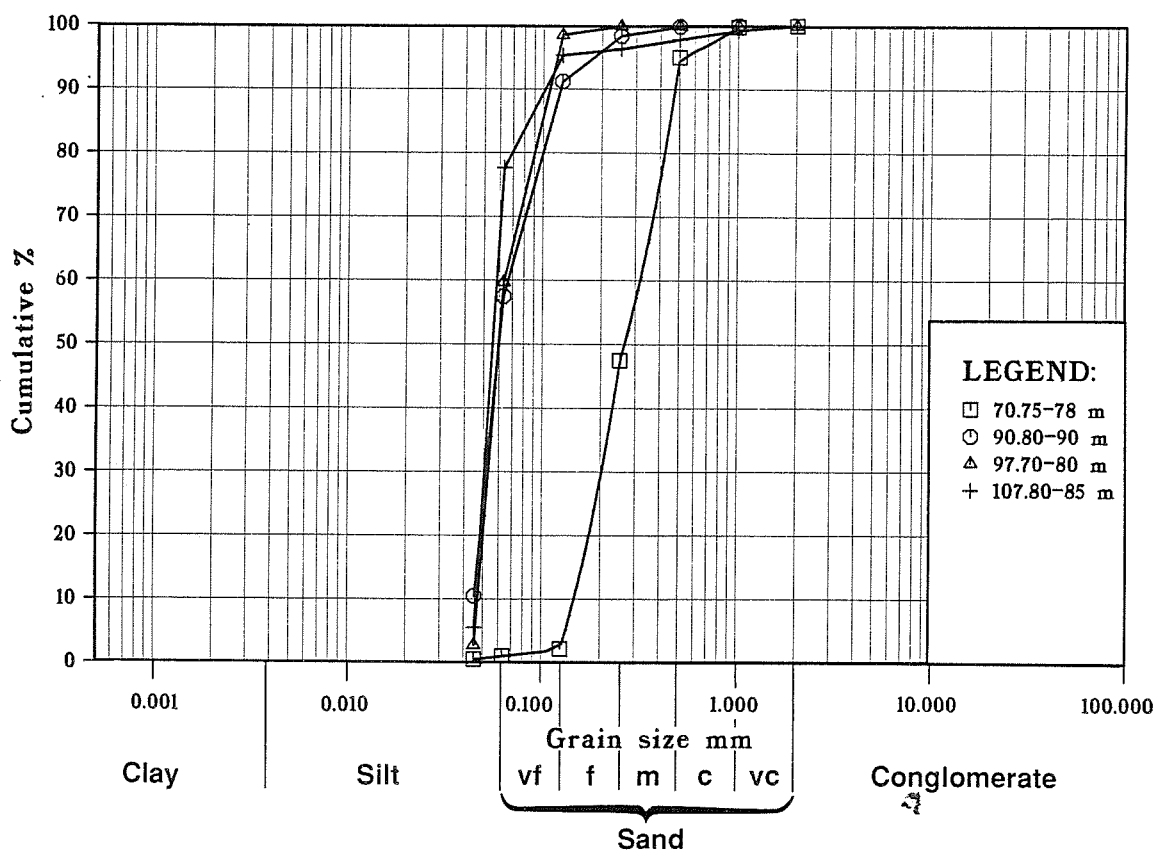


Table 3.3.3 *Semi-quantitative mineral composition of the sediments from core 7316/06-U-01 based on X-ray diffraction analysis.*

Depth	Smec	Mix	Mi/Ill	Chl	Kaol	Qtz	KFsp	Plag	Sid	Pyr	Zeo	Amphi
44.15	0.0	61.5	10.1	3.4	0.9	14.4	2.3	6.7	0.0	0.0	0.0	0.0
46.70	0.0	35.3	12.5	17.0	8.5	16.5	3.0	6.6	0.0	0.0	0.0	0.0
48.00	0.0	29.6	14.3	10.8	9.5	23.3	3.9	9.7	0.0	0.0	0.0	0.0
52.17	0.0	11.7	3.5	13.0	11.5	9.4	3.2	5.3	37.1	0.0	0.0	0.0
52.65	0.0	50.9	16.5	2.6	0.0	4.5	0.7	2.0	5.4	0.0	0.0	0.0
70.91	0.0	1.2	0.5	0.0	0.0	89.9	4.5	0.9	2.1	0.9	0.0	0.0
90.78	0.4	3.9	2.5	1.4	0.2	65.5	12.7	8.1	5.1	0.4	0.0	0.0
100.00	1.3	6.7	4.1	2.4	0.0	64.8	16.3	2.9	0.4	0.9	0.3	0.0
107.10	0.3	4.6	3.4	2.2	0.0	68.2	8.5	11.2	0.6	0.0	0.4	0.6

Low diagenetic conditions are also shown by the apatite fission track data which indicate that the maximum diagenetic temperature was always lower than 50°C, but further limitations are not possible from these data. (Appendix A.8).

3.3.4 Petrophysical logs and core measurements

Downhole logs

The gamma, density and neutron logs were run inside the HR string from TD to the seabed and also inside the API drill string from 94 m. The gamma log values logged through only the HR string are recalculated to open hole values below the API bit, based on a regression analysis from a previous shallow drilling survey. Due to very low variance in the gamma log values it was impossible to calculate any depth shift correction (see Appendix A.10). The logs are shown in Figure 3.3.1. The gamma log and petrophysical properties are also shown together in the summary sheet in Appendix A.1.

108.0 (TD) – 73.0 m: Unit A

The fine-grained sandstone shows gamma log values around 60 API below the API drill bit. Above the API bit the gamma log is attenuated by the weight collars of the API string up to 66 m and of the API drill string above 66 m. Peaks with 4.5 m spacing in the weight collars and 9.5 m in the API string are caused by tool joints.

The neutron porosity and density logs give very little information, but an increase on the neutron porosity log was suggested to define the boundary between units A and B at 73 m (see Section 3.3.3).

73.0 - 65.0 m: units B and C

The signals from these units are disturbed by the attenuation of the two drill strings. The high reading of the neutron porosity log at 70.8 m is correlated to the proposed boundary between units B and C.

54 - 39 m: Cretaceous megablock within Quaternary sediments

The gamma log shows a somewhat higher level in this interval than below 65 m. The density log shows lower values, but this effect may be caused by cavings. The neutron porosity raw values show also some changed values in this interval, but the neutrons are very sensitive to wellbore effects, (or drill string effects) when they have passed twice through a drillstring.

Gamma measurements on core

Figure 3.3.3 shows the spectral gamma values measured on the core from Unit A and the Cretaceous megablock. The K-plot on the spectral gamma laboratory measurements shows that K is the main source for the gamma log response, except at 90.53 m where both K and U show higher values.

The Cretaceous glacial megablock between 54 m and 39 m shows relatively high gamma values, mainly caused by K with an increasing trend upward.

Due to the poor consolidation of the rock there was not possible to measure any horizontal sound velocity values.

Thermal conductivity

See Appendix A.9.

3.3.5 Paleontology

Core 7316/06-U-01 is extremely poor in *in situ* Neogene dinoflagellate cysts, but contains well preserved and fairly common reworked Paleogene and Cretaceous (Aptian - Cenomanian?) species. The stratigraphic distribution of dinoflagellate cysts in this core is shown in Figures 3.3.6 and 3.3.7. No foraminifera or siliceous fossils have been found in this core.

Based on the limited evidences given by the marine microfloras, a general Late Oligocene to Early Miocene age is suggested for the cored unit between 108.0 m and 73.0 m.

From 43.26 m to 52.40 m in the Quaternary sediments at this site, a probably uncontaminated Cretaceous fossil assemblage has been found with dinoflagellate cysts, foraminifera and calcareous nannoplankton. The distribution of foraminifera and calcareous nannoplankton is shown in Figures 3.3.8 and 3.3.9. The biostratigraphic results give evidence of a Late Albian - Early Cenomanian, probably Early Cenomanian age for these assemblages (see Section 3.3.8).

108.0 (TD) - 73.0(?)m, Unit A: Upper Oligocene - Lower Miocene

The samples at 107.20 m and 99.10 m contain *Cribroperidinium tenuitabulatum*. This species ranges from Early Miocene to Middle Miocene on the Vøring Plateau ODP Leg 104 (Manum et al. 1989), but is also known to range down into the Middle Oligocene in southern areas. The sample at 99.10 m also contains *Riculacysta perforata* and *Emslandia spiridoides*. *Riculacysta perforata* is previously reported from the Middle and Late Oligocene (Stover 1977), while the latter species in ODP Leg 104 was found within the Late Oligocene - Early Miocene (Manum et al. 1989).

Compared to ODP Leg 104, *C. tenuitabulatum* indicates an Early - Middle Miocene age at 99.10 m, but a Late Oligocene age cannot positively be neglected. The samples at 103.4 m, 101.10 m and 97.15 m have no age diagnostic Neogene species.

73.0 - 65.0 m, units B and C: Indeterminate.

No fossils have been found in these units.

97.15 99.10 101.10 103.40 107.20	PALYNOLOGY 7316/06-U-01
	OLIGOCENE - MIOCENE DINOFLAGELLATE CYSTS
	SPINIFERITES SPP.
	IMPAGIDINIUM SPP.
	CRIBROPERIDINIUM TENUITABULATUM (GERLACH) HELENES 1984
	EMLANDSIA SPIRIDOIDES BENEDEK 1972
	RICULACYSTA PERFORATA STOVER 1977
	CALIGODINIUM AMICULUM DRUGG 1970
	OPERCULODINIUM CENTROCARPUM (DEFL. & COOKS.) WALL '67
	REWORKED PALEOCENE DINOFLAGELLATE CYSTS
	DAPSILIDINIUM SIMPLEX (WHITE) BUJAK ET AL. 1980
	HYSTRICHOSTROGYLON MEMBRANIPHORUM AGELOPOULOS 1961
	CORRUDINIUM INCOMPOSITUM (DRUGG) STOVER & EVITT 1978
	APTEITODINIUM HOMEOMORPHUM (DEFL. & COOKS.) LENT. & WILL. '77
	?IMPAGIDINIUM SIMPLICIUM (COOKS. & EISEN.) STOVER & EVITT '78
	KISSELOVIA COLEOTHRYPTA (WILL. & DOWNIE) LENT. & WILL. 1976
	DEFLANDREA PHOSPHORITICA EISENACK 1938
	APTEODINIUM SP. I
	SPINIFERITES RAMOSUS (EHRENBERG) LOEBLICH & LOEBLICH 1966
	GLAPHYROCYSTA SEMITECTA (BUJAK) LENTIN & WILLIAMS 1981
	LENTINIA SERRATA BUJAK IN BUJAK ET AL. 1980
	THALASSIOPHORA PELAGICA (EISENACK) EISENACK & GOCHT 1960
	ACHOMOSPHERA ALCICORNU (EISENACK) DAVEY & WILL. 1966
	PALAEOCYSTODINIUM GOLZOWENSE ALBERTI 1961
	LEJEUNECYSTA HYALINA (GERLACH) ARTZNER & DORHOFFER 1978
	?CLEISTOSPHERIDIUM INSOLITUM (EATON) STOVER & EVITT 1978
	HOMOTRYBILUM TENUISPINOSUM DAVEY & WILLIAMS 1966
	PHANOPERIDINIUM AMOENUM DRUGG & LOEBLICH 1967
	GLAPHYROCYSTA ORDINATA (WILL. & DOWNIE) STOVER & EVITT 1978
	CRIBROPERIDINIUM GIUSEPPI (MORGENROTH) HELENES 1984
	WETZELIELLA ARTICULATA EISENACK 1938
	BATICASPHERA SPP.
	DRACODINIUM VARIELONGITUDUM (WILL. & DOWNIE) COSTA & DOWNIE '74
	MEIOUROGONYALAX SP. I MANUM 1976
	HYSTRICHOKOLPOMA SALICA EATON 1976
	LEPTODINIUM SP.
	COMETODINIUM SP.
	REWORKED CRETACEOUS DINOFLAGELLATE CYSTS
	CRIBROPERIDINIUM EDWARDSII (COOKS. & EISEN.) DAVEY 1969
	PTERODINIUM CINGULATUM (WETZEL) BELOW 1981
	RHOMBODELLA PAUCISPINA (ALBERTI) DUXBURY 1980
	CYCLONEPHELICUM DISTINCTUM DEFLANDRE & COOKSON 1955
	OLIGOSPHERIDIUM COMPLEX (WHITE) DAVEY & WILLIAMS 1966
	SPINIFERITES SCABROSUS (CLARKE & VERDIER) LENT. & WILL. 1975
	TRICHODINIUM CASTANEA (DEFLANDRE) CLARKE & VERDIER 1967
	PALAEOPERIDINIUM CRETACEUM POCKOCK 1962
	LEBERIDOCYSTA CHLAMYDATA (COOKS. & EISEN.) STOVER & EVITT '78
	PSEUDOCERATIUM PELLIFENUM GOCHT 1957
A	LITHOSTRATIGRAPHIC UNIT
LTE OLIG.- ELY MIO.	AGE

Figure 3.3.6

Distribution chart of dinoflagellate cysts in core 7316/06-U-01, sorted according to earliest occurrence.

97.15 99.10 101.10 103.40 107.20	PALYNOLOGY 7316/06-U-01
	OLIGOCENE - MIOCENE DINOFLAGELLATE CYSTS
	IMPAGIDINIUM SPP.
	OPERCULODINIUM CENTROCARPUM (DEFL. & COOKS.) WALL '67
	EMLANDSIA SPIRIDOIDES BENEDEK 1972
	RICULACYSTA PERFORATA STOVER 1977
	CALICODINIUM AMICULUM DRUGG 1970
	CRIBROPERIDINIUM TENUITABULATUM (GERLACH) HELENES 1984
	SPINIFERITES SPP.
	REWORKED PALEOCENE DINOFLAGELLATE CYSTS
	CRIBROPERIDINIUM GIUSEPPI (MORGENROTH) HELENES 1984
	GLAPHYROCISTA ORDINATA (WILL.&DOWNIE) STOVER & EVITT 1978
	WETZELIELLA ARTICULATA EISENACK 1938
	DEFLANDREA PHOSPHORITICA EISENACK 1938
	APTEODINIUM SP.I
	COMETODINIUM SP.
	MEIOUROGONYAULAX SP.I MANUM 1976
	DRACODINIUM VARIELONGITUDUM (WILL.&DOWNIE) COSTA&DOWNIE '74
	HYSTRICHOKOLPOMA SALICA EATON 1976
	LEPTODINIUM SP.
	BATICASPHAERA SPP.
	HOMOTRYBILUM TENUISPINOSUM DAVEY & WILLIAMS 1966
	LEJEUNECYSTA HYALINA (GERLACH) ARTZNER & DORHOFFER 1978
	PHANOPERIDINIUM AMOENUM DRUGG & LOEBLICH 1967
	PALAEOCYSTODINIUM GOLZOWENSE ALBERTI 1961
	?CLEISTOSPHAERIDIUM INSOLITUM (EATON) STOVER & EVITT 1978
	ACHOMOSPHAERA ALCICORNU (EISENACK) DAVEY & WILL. '1966
	KISSELOVIA COLEOTHRYPTA (WILL.&DOWNIE) LENT. & WILL. 1976
	SPINIFERITES RAMOSUS (EHRENBERG) LOEBLICH & LOEBLICH 1966
	GLAPHYROCISTA SEMITECTA (BUJAK) LENTIN & WILLIAMS 1981
	LENTINIA SERRATA BUJAK IN BUJAK ET AL. 1980
	THALASSIOPHORA PELAGICA (EISENACK) EISENACK & GOCHT 1960
	?IMPAGIDINIUM SIMPLICIUM (COOKS.&EISEN.) STOVER & EVITT '78
	APTEITODINIUM HOMEOMORPHUM (DEFL.&COOKS.) LENT. & WILL. '77
	HYSTRICHOSTROGYLON MEMBRANIPHORUM AGELOPOULOS 1961
	CORRUDINIUM INCOMPOSITUM (DRUGG) STOVER & EVITT 1978
	DAPSILIDINIUM SIMPLEX (WHITE) BUJAK ET AL. 1980
	REWORKED CRETACEOUS DINOFLAGELLATE CYSTS
	OLIGOSPHAERIDIUM COMPLEX (WHITE) DAVEY & WILLIAMS 1966
	PSEUDOCERATIUM PELLIFENUM GOCHT 1957
	LEBERIDOCYSTA CHLAMYDATA (COOKS.&EISEN.) STOVER & EVITT '78
	PALAEOPERIDINIUM CRETACEUM POCKOCK 1962
	SPINIFERITES SCABROSUS (CLARKE&VERDIER) LENT. & WILL. 1975
	TRICHODINIUM CASTANEA (DEFLANDRE) CLARKE & VERDIER 1967
	CYCLONEPHELICUM DISTINCTUM DEFLANDRE & COOKSON 1955
	CRIBROPERIDINIUM EDWARDSII (COOKS.&EISEN.) DAVEY 1969
	PTERODINIUM CINGULATUM (WETZEL) BELOW 1981
	RHOMBODELLA PAUCISPINA (ALBERTI) DUXBURY 1980
A	LITHOSTRATIGRAPHIC UNIT
LTE OLIG.- ELY MIO.	AGE

Figure 3.3.7

Distribution chart of dinoflagellate cysts in core 7316/06-U-01, sorted according to latest occurrence.

43.26 48.40 49.25 52.40	FORAMINIFERA, CALC. NANNOPL. 7316/06-U-01
	FORAMINIFERA
	CIBICIDES CF. PROPRIUS (BROTZEN 1948)
	GYROIDINOIDES SP.
	VALVULINERIA ?SP.
	HETEROHELIX SP.
	HEDBERGELLA CF. DELRIOENSIS (CARSEY 1926)
	LENTICULINA CULTRATA (MONTFORT 1808)
	HAPLOPHRAGMOIDES AFF. GIGAS MINOR NAUSS 1947
	RECURVOIDES SP.
	HYPERAMMINA RUGOSA VERDENIUS & VAN HINTE 1983
	AMMOLAGENA CLAVATA (JONES & PARKER 1860)
	TROCHAMMINA SPP.
	KALAMOPSIS ?SP.
	GAUDRYINA FILIFORMIS BERTHELIN 1880
	AMMODISCUS CRETACEUS (REUSS 1855)
	GLOMOSPIRA GORDIALIS (JONES & PARKER 1860)
	RECURVOIDES AFF. IMPERFECTUS HANZLIKOVÁ 1953
	HAPLOPHRAGMOIDES SP.
	CALCAREOUS NANNOPLANKTON
	STAUROLITHITES MATALOSUS STOVER 1966
	GARTNERAGO OBLIQUUM (STRADNER 1963)
	WATZNAUERIA BIPORTA BURKY 1969
	TRANOLITHUS SALILLUM (NOEL 1965)
	AHMUELLERELLA OCTORADIATA (GORKA 1957)
	EPROLITHUS FLORALIS (STRADNER 1962)
	CHIASTOZYGUS LITTERARIUS (GORKA 1957)
	HELICOLITHUS TRABECULATUS (GORKA 1957)
	VEKSHINELLA STRADNERI ROOD 1971
	GARTNERAGO NANUM THIERSTEIN 1974
	EIFFELLITHUS TURRISEIFFELLII (DEFLANDRE 1954)
	TEGUMENTUM STRADNERI (THIERSTEIN 1972)
	REPAGALUM PARVIDENTATUM (DEFLANDRE & FERT 1954)
	MISCELLANEOUS
	INOCERAMUS PRISMS
	MOLLUSE DEBRIS
	FISH TEETH
	LITHOSTRATIGRAPHIC UNIT
EARLY CENOMANIAN	AGE

Figure 3.3.8

Distribution chart of foraminifera and calcareous nannofossils in an allochthonous megablock over core 7316/06-U-01, sorted to lowest occurrence.

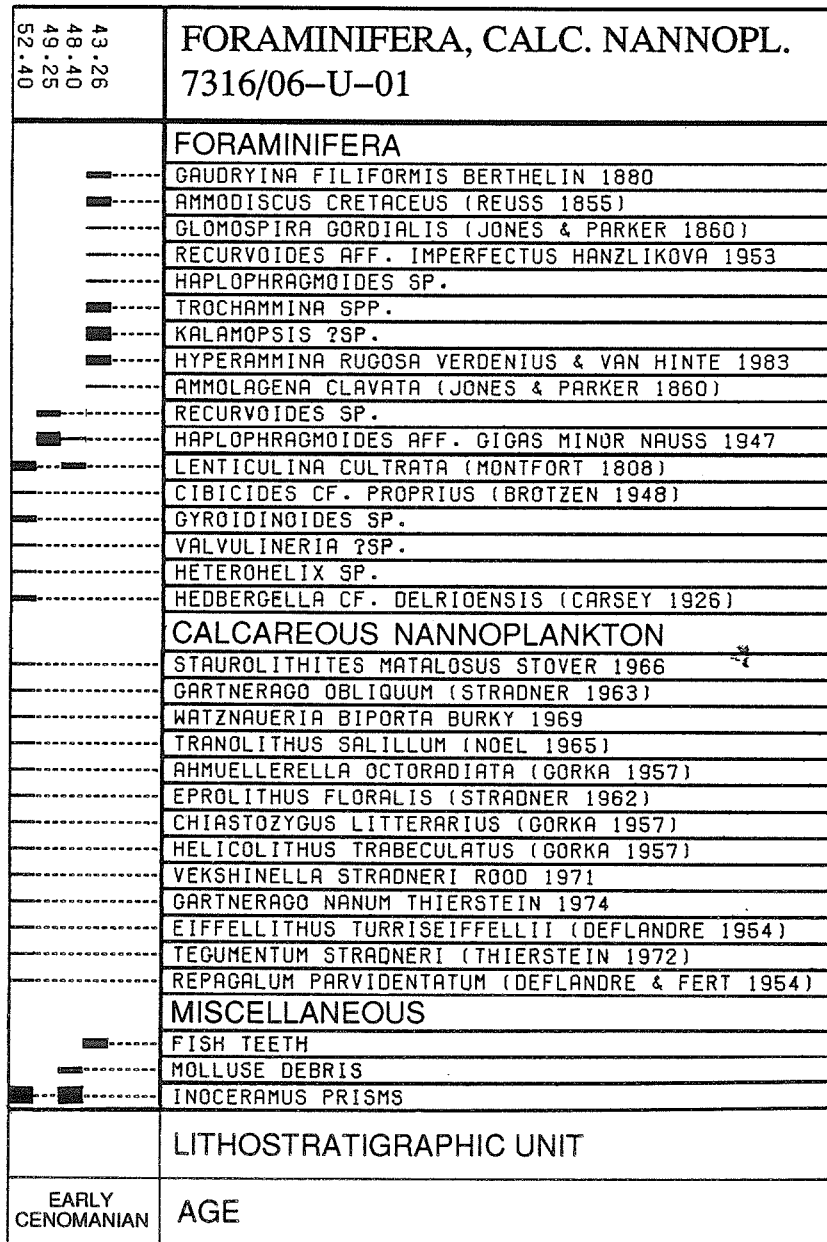


Figure 3.3.9

Distribution chart of foraminifera and calcareous nannofossils in an allochthonous megablock over core 7316/06-U-01, sorted to highest occurrence.

3.3.6 Organic geochemistry

Consideration of the organic geochemistry of the rocks sampled in core 7316/06-U-01 can be divided into two parts, i.e. a glacially-emplaced Cretaceous slab overlying the bedrock and the bedrock itself. The data obtained from the Cretaceous strata are considered in Section 3.3.8.

The seven samples taken from cored bedrock consisted of poorly-cemented sandstone. Poor sample quality prevented material being collected for analysis of gaseous hydrocarbons. The samples were found to have very similar total organic carbon contents of between 0.11 wt% and 0.13 wt% (Figure 3.3.10). The low content of organic matter in the samples and the absence of any evidence for hydrocarbon staining suggested that further analysis would be pointless.

It was not possible to obtain any vitrinite reflectance measurements from this part of the core due to an absence of suitable vitrinite particles.

The organic geochemical data obtained from this core indicate that the traces of kerogen present in these rocks have no hydrocarbon generation potential.

3.3.7 Depositional environment

The depositional environment (Figure 3.3.11) is difficult to interpret from the very sparse core recovery, sparse paleontological information, and log responses hampered by being measured inside the drill strings.

Unit A with Late Oligocene to Middle Miocene palynomorphs is clearly of marine origin containing abundant reworked Cretaceous and Eocene palynomorphs. From the seismic data it is indicated that the unit is deposited by rapid deposition and/or gravity mass flows. From the sedimentological parameters this interpretation can neither be verified nor rejected, but the similarities with the sandstones of subunits A and B in core 7316/03-U-01 are striking indicating similar depositional mechanisms. The abundant reworked Eocene fossils may indicate redeposition of Eocene shelf sediments, probably on the upper slope.

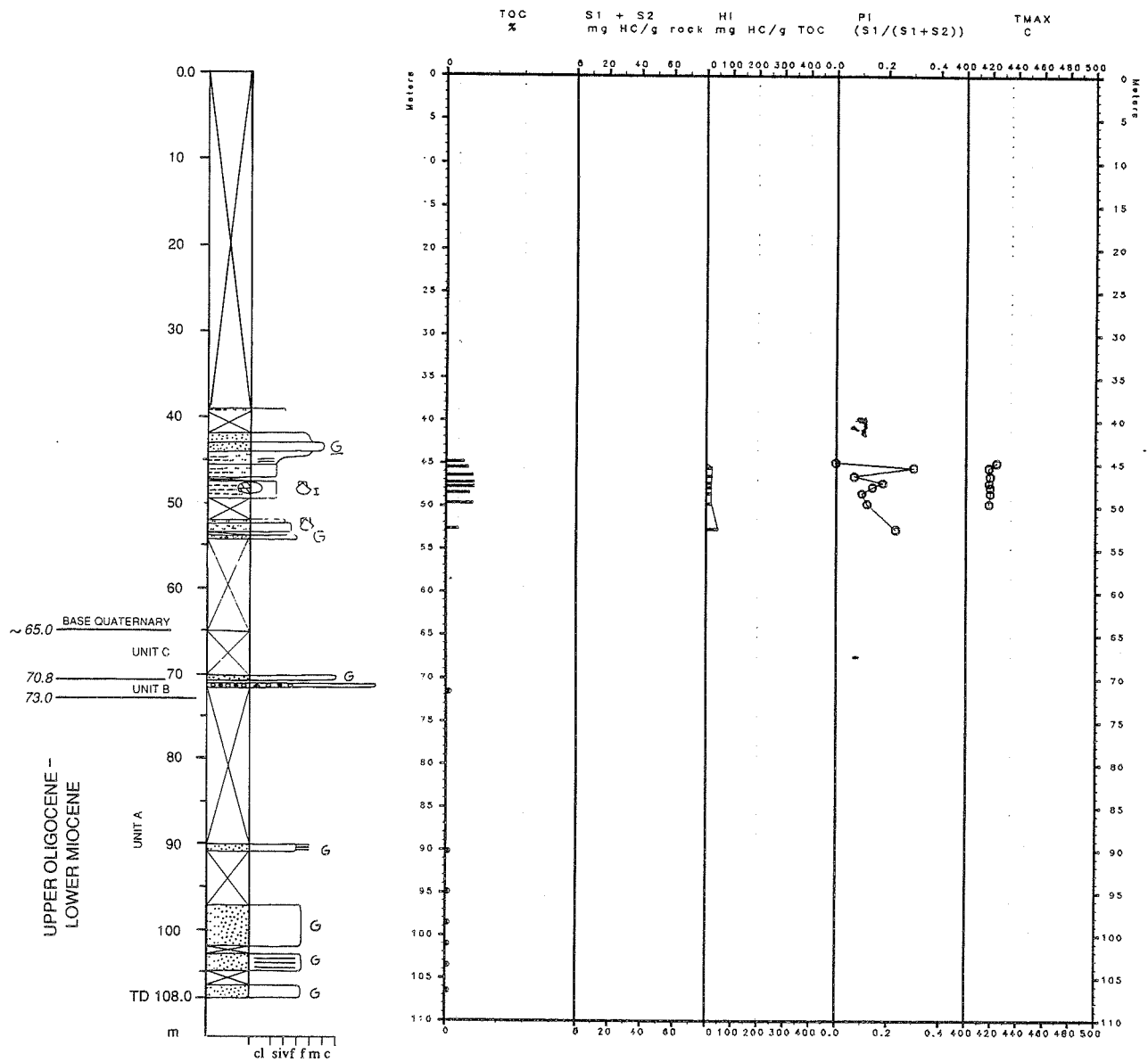


Figure 3.3.10 Total organic carbon content and Rock-Eval data of core 7316/06-U-01.

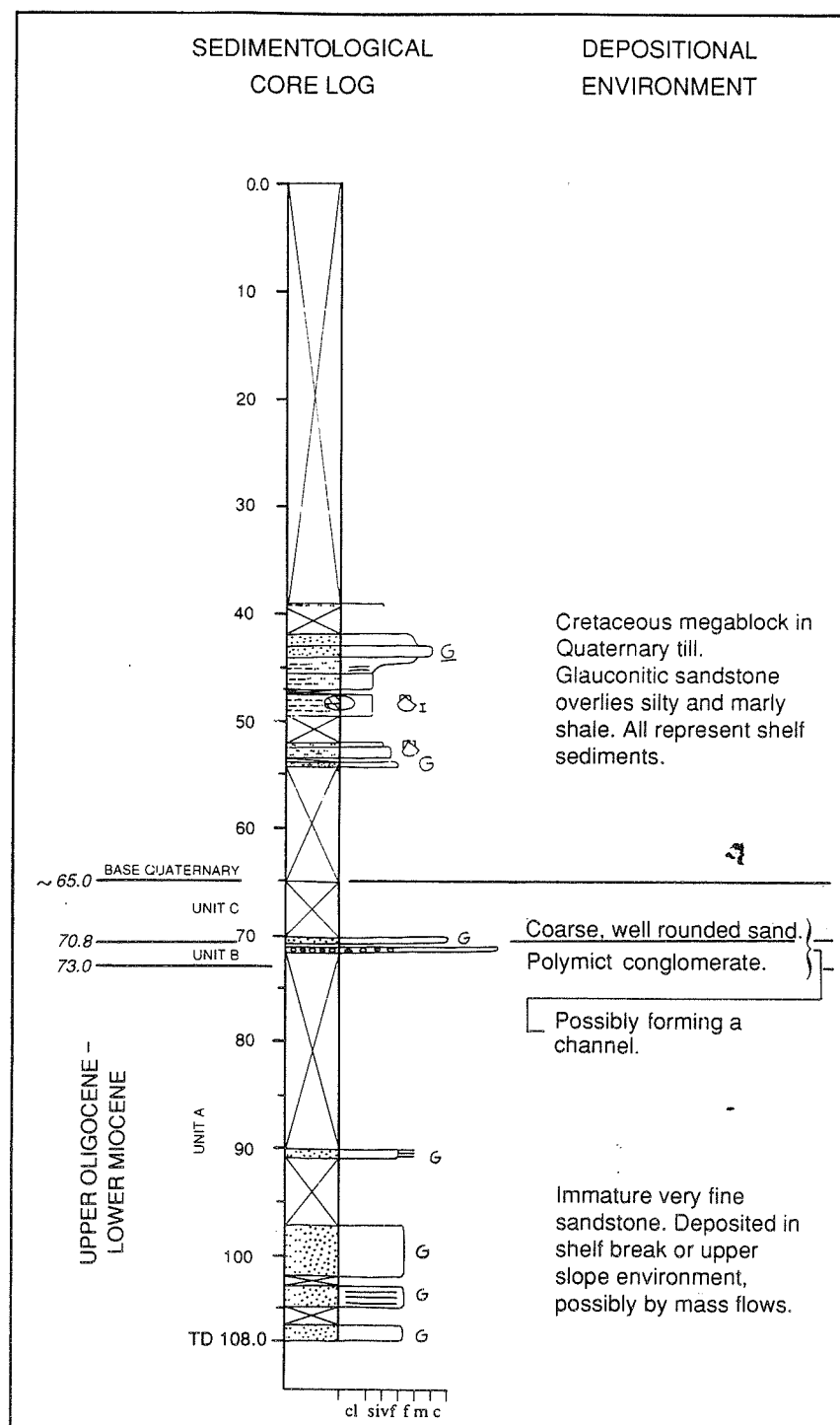


Figure 3.3.11 Interpreted depositional environment for core 7316/06-U-01.

In the seismic interpretation it was indicated that units B and C may form a channel ca. 250 m wide and 10-12 m deep. The sedimentological study may support such an interpretation. The polymict conglomerate may thus represent a basal conglomerate of the channel that has been filled with well rounded sand. The rounding of sand grains in Unit C indicate deposition in high energy environment or possibly that such sediment has been transported, in the channel, and redeposited at the present locality.

3.3.8 Cretaceous rocks cored in a Quaternary glaciotectionic megablock

The glaciotectionic structure

In core 7316/06-U-01 bedrock was met at about 65 m (75 ms) below the seabed. The drilling did, however, reveal well cemented sedimentary rocks from 39 m to about 54 m below the seabed. These rocks superposed a till unit in the interval 65 - 54.1 m, which in turn rested on pre- Quaternary sedimentary rocks. The boundary at 54.1 m is sharp, but fragments of the overlying rock are clearly sheared into the fine-grained underlying till (Figure 3.3.13). The core interval between 39 m and 54.13 m is interpreted as part of an allochthonous body of bedrock which is glaciotectionically displaced by a grounded ice sheet from its original position further to the east. This type of structure is classified as a glaciotectionic megablock (Aber et al. 1989).

Amino acid measurements assign a late Pleistocene age to the underlying till, and the glaciotectionic event must have occurred during the Late Pleistocene. Seismic correlations and amino acid data from site 7317/10-U-01 further suggest that it happened during the last glaciation in the area, which probably occurred about 30 000 - 13 000 radiocarbon years ago. Details of this megablock complex and its genesis are given in Appendix A.13, but a few remarks are repeated here as a background to understand the structure and to point out possible source areas for the cored allochthonous rocks.

The megablock complex is mapped seismically to cover about 800 square km. The structure is, however, poorly defined on the seismic records and may have a wider occurrence. The location of its northern boundary between sites 7316/06-U-01 and -U-02 is confirmed by these two boreholes (Figure 3.3.12). The thickness of the block varies, but is typically a few tens of metres, and the dimensions classify this as one of the largest of its kind in the world. (To our knowledge the only larger example is a 1000 km² wide glaciotectionic megablock consisting of Cretaceous shale in eastern Saskatchewan, Canada, Sauer 1978, Aber et al. 1989).

The entire megablock complex does not necessarily consist of Cretaceous rocks, but the data does not allow further evaluation of this.

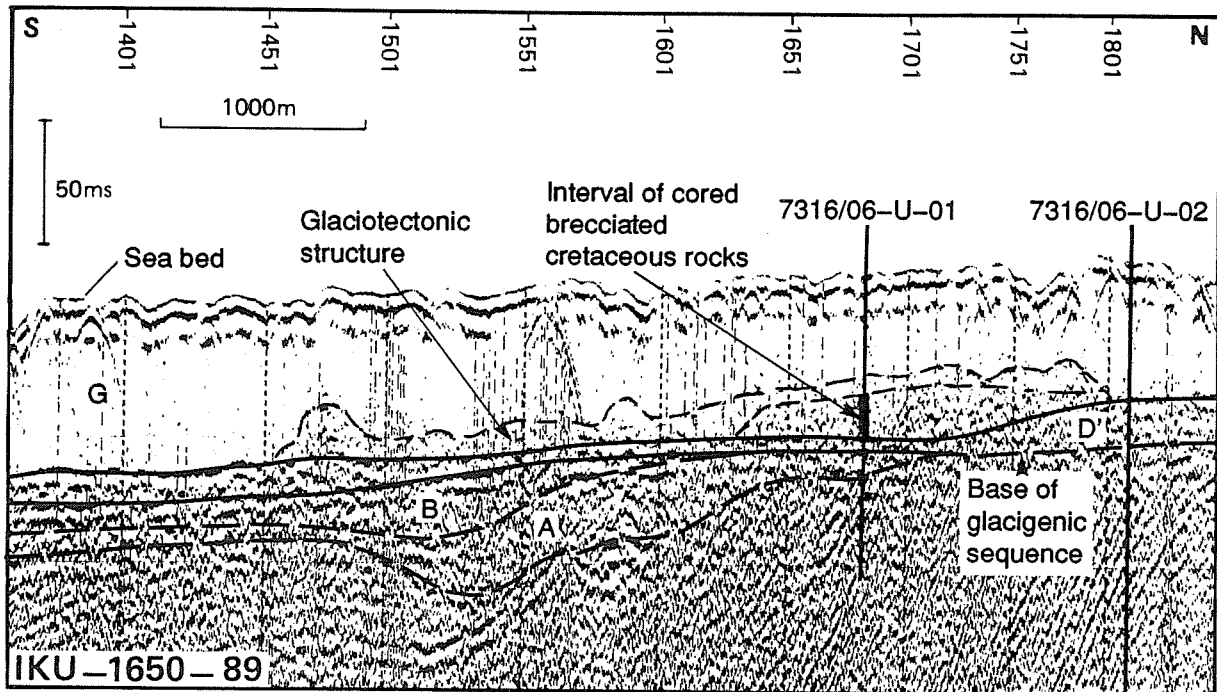


Figure 3.3.12 Seismic section across sites 7316/06-U-01 and -U-02. Cretaceous sedimentary rocks embedded in Pleistocene glaciogenic sediments were penetrated in borehole 7316/06-U-01 but not in borehole 7316/06-U-02. These rocks are interpreted as part of a large (800 km²) glaciotectionic megablock complex probably formed about 10 000 - 30 000 radiocarbon years ago. The seismic section shows part of the structure. Capital letters denote mapped seismic units within the glaciogenic sediment sequence. (See Appendix A.13).

Figure 3.3.14 shows the extent of the megablock complex. The figure also shows the occurrence of Cretaceous sedimentary rocks as well as the areal extent of Quaternary sediments older than the formation of the megablock complex. (That is, areas where the bedrock was protected from glacial activity when the structure was formed). The megablock complex was probably transported as one or more large, sheet-like rock fragments frozen to the base of a grounded ice sheet, which probably flowed in a roughly westerly direction in this area. Glacial transport distances of such structures are known to be up to 300 km or more, and the source area for the structure must therefore be searched for in a large area. The source for the megablock complex should be in the intersection between these two areas, unless the displacement happened in several phases.

7316/06-U-01

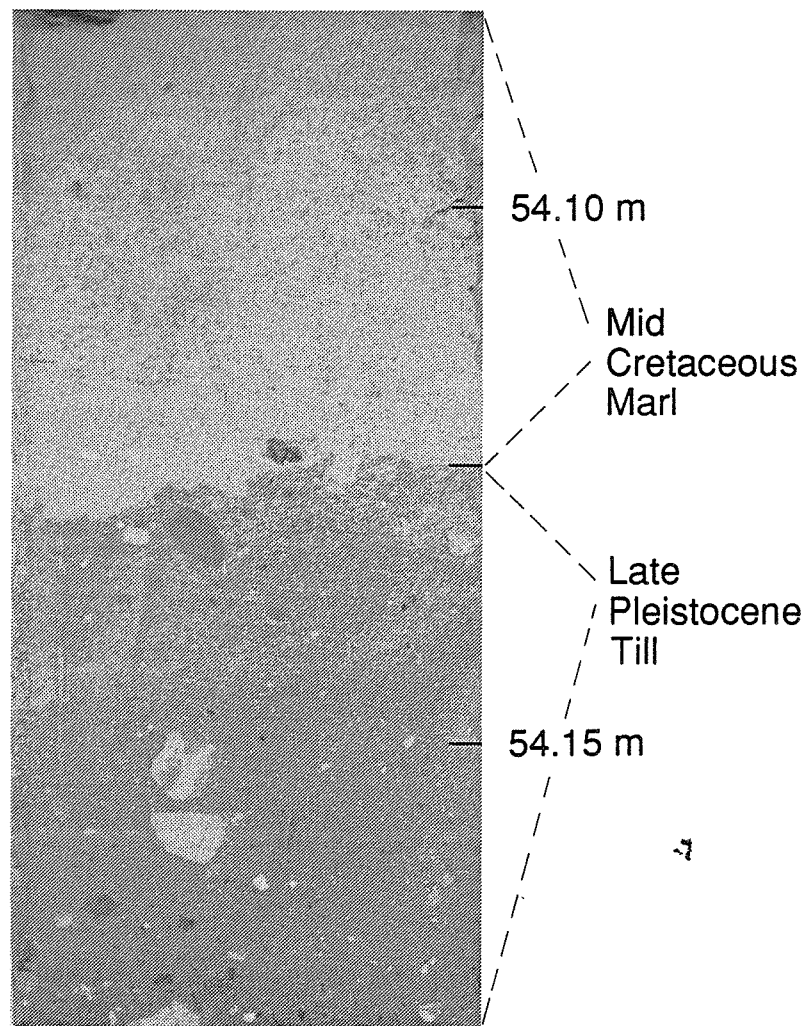


Figure 3.3.13 *Sharp boundary between the base of the glaciotectionic structure consisting of mid Cretaceous rocks and underlying Late Pleistocene till.*

Conclusive evidence for a source area is not found, but in Figure 3.3.14 three candidates are indicated from interpretation of the base Quaternary unconformity. Area 1 is characterized by elongate depressions on the unconformity at the base of the glaciogenic sediment sequence. The megablock complex may originate from rocks displaced from these depressions. As the glaciotectionic structure is mapped it does, however itself cover the most pronounced (and closest) depression. This suggests that at least part of the structure has another and more distant source. Areas 2 and 3 are characterized by a "blocky and tectonized" unconformity at the base of the glaciogenic sediment sequence. This may be taken as indicative of glaciotectionism on this surface in these areas. Areas 2 and 3 are more comparable in size to the megablock complex. These areas, in contrast to area 1, also have generally flatlying bedding which is considered favourable for release of large sheets of rock. Areas 2 and 3 may therefore be preferred as source

areas. This implies a transport length of about 3 - 4 times the width of the structure.

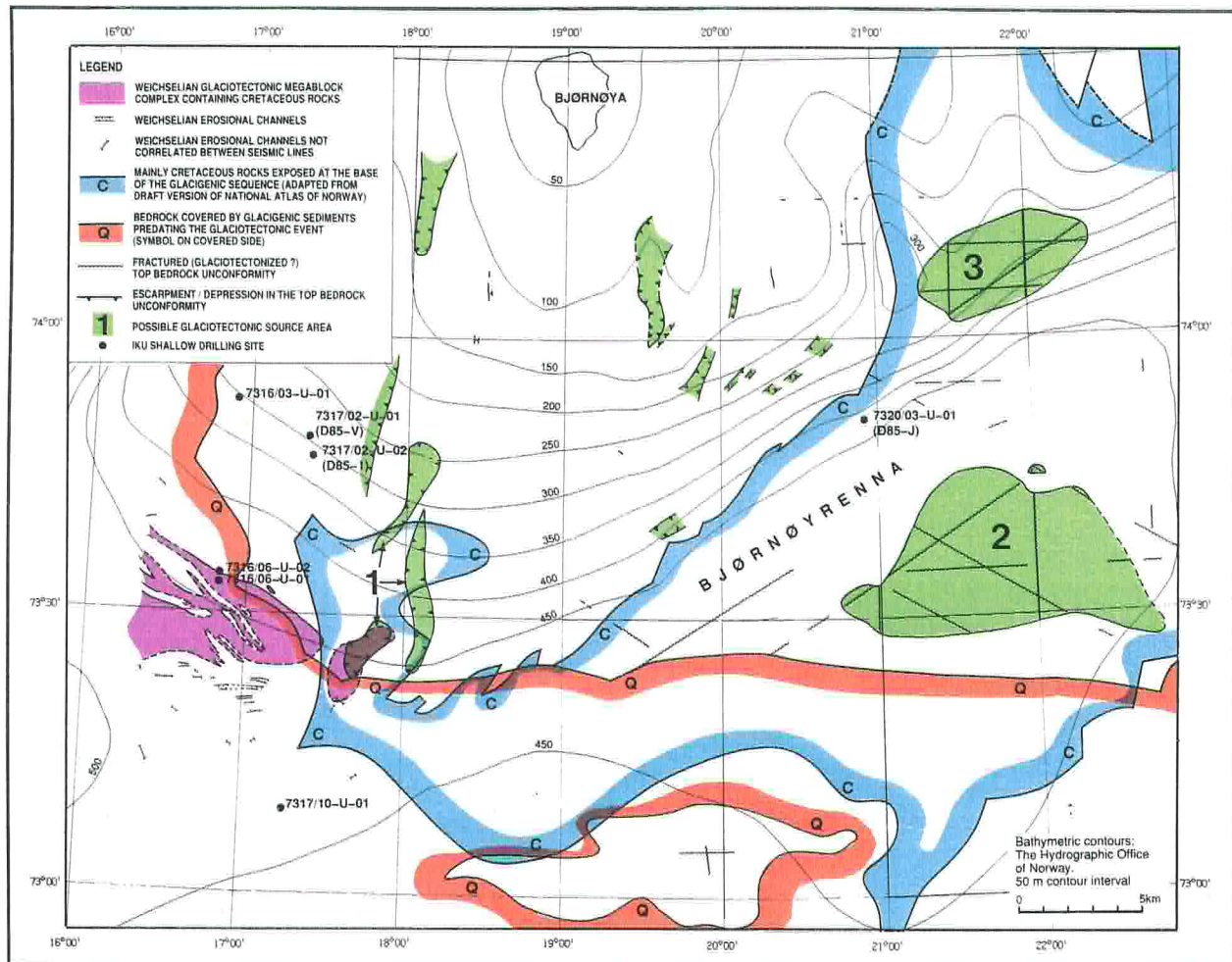


Figure 3.3.14 Location of a glaciotectionic megablock complex in western Bjørnøyrenna penetrated in borehole 7316/06-U-01. The cored rocks were dated to mid Cretaceous, but were probably transported to their present position during the late Weichselian (13 000 - 30 000 years ago). Possible source areas for the displaced rocks must be located in areas where Cretaceous rocks were exposed to the ice sheet at the time of formation. Candidate source areas denoted 1, 2 and 3 are discussed in the text.

Core description

The top of the megablock complex was probably penetrated at 29.3 m in borehole 7316/06-U-01. A sand sample was obtained at this level and both comments from the driller as well as petrophysical log data indicate a lithological boundary at this level. Rock cores where, however, not obtained between 29.3 m and 39 m and the following chapter addresses the geology of the cored interval below 39 m. The core of Cretaceous sediments at site 7316/06-U-01 comprises two lithologies, sharply overlying the Quaternary till (Figures 3.3.3 and 3.3.13). From 54.1 m, light grey-grey calcareous

silty/sandy claystone continues up to 44.2 m where it is overlain by a sandstone with a very high content of glauconite grains. This unit is cored up to 42.8 m, however a core piece at 39.0 - 39.2 m of similar lithology, although more fine-grained, indicates upward extension of the lithology to at least this level. Note, however, that there are several breaks in the recovery of the Cretaceous rocks. These rocks have been transported horizontally for a long distance and may not be correctly stratigraphically positioned or oriented. As slickenside surface was noted on the lithologic boundary at 44.2 m, indicating that this rather may be a tectonic boundary.

The lower fine-grained interval (54.1 - 52.2 m) consists of light grey sandy-silty claystone. The content of calcite (?mud) is very high in the light grey lower part (49% calcite by XRD at 52.17 m) and the lithology may be regarded as a marl. The core interval 52.8 - 52.4 m is particularly strongly brecciated. A thin section from this level (52.65 m) shows well rounded glauconite clasts and very fine-grained sand grains. The quartz grains are subangular. The texture is partly grain supported and partly matrix supported. Prisms of *Inoceramus* are observed.

~1

The upper part of the fine-grained interval consists of silty shale and contains less calcite than below. The sediment is dark grey to brownish grey. The rock is partly laminated and bioturbated in the uppermost part. At 48.2 m there is a brown 12 cm thick carbonate "bed" interpreted to represent a siderite concretion. Fragments and prisms of the bivalve *Inoceramus* are abundant in the lower part of this interval.

The upper part of the core consists of sandstone mostly composed of abundant well rounded glaucony grains with a minor content of fine-grained sand(stone) and clay. Lamination is present, and at the base there are some claystone pebbles. The sandstone is multicoloured varying from greenish in the lower part through a mottled green-red interval to mottled light greyish in the upper part. The core piece at 39.2 - 39.0 m has a similar lithology although more fine-grained, and is dark green in colour.

X-ray diffraction analysis has been undertaken on five samples between 55.65 m and 44.15 m (Table 3.3.3 and Figure 3.3.3).

The mixed layer clay mineral is a disordered smectite-illite. Only whole rock analyses are performed and glycol saturation is not undertaken. Therefore, the proportion of illite/smectite is not calculated. What can be concluded, however, is that the amount of expandable layers is greater than

about 30% since the (001)-reflection of smectite disappears with a higher proportion of illite. Illite is present in amounts between 10% and 16% in most of the samples, the only exception is a marl from 52.17 m. The illite probably reflects the presence of glauconite as seen in thin sections. The mixed layer illite-smectite may also be due to glaucony, reflecting an immature stage of the development of glauconite.

Opal-CT is detected in one sample only. The same sample contains about 9% zeolite (clinoptilolite). Opal-CT is indicative of low temperature as it recrystallizes to quartz at 50-70°C. Both opal-CT and clinoptilolite are commonly associated with material derived from volcanic activity. Opal-CT is also formed from recrystallization of silica producing plankton.

The two lithologies form together a coarsening upward succession. The high content of well rounded glauconite grains and the *Inoceramus* fragments suggest deposition in shelf conditions at a depth permitting abrasion by wave activity.

Biostratigraphic results

At 43.26 m, an arenaceous foraminiferal fauna (Figures 3.3.8 and 3.3.9) of some diversity was met, whereas at 44.25 m, a nearly monospecific assemblage of *Haplophragmoides* aff. *gigas minor* occurs. The genera *Hyperammina* and *Kalamopsis* that occur at 43.26 m, have regional ranges from the Early Cretaceous into the Early Tertiary. Several arenaceous taxa with a more compound structure allow a narrower stratigraphic attribution, but their presence is also related to specific environmental characters. Similar foraminiferal association to those found in the samples at 43.26 m and 44.25 m occurs in the outer Carpathians in Poland (Geroch & Nowak 1984). There *Gaudryina filiformis* occurs from the Aptian to the Santonian, whereas *Recurvoides imperfectus* and *Haplophragmoides gigas minor* are characteristic for the *Plectorecurvoides alternans* Zone, that straddles the Albian - Cenomanian boundary. In our material occur forms that strongly resemble these two taxa, and an age in the Late Albian - Early Cenomanian time interval is considered very probable.

The marine microfloras recovered at 47.20 m, 47.78 m and 54.07 m contain a rich and distinct Late Albian - Early Cenomanian dinoflagellate assemblage, including characteristic species like *Dorocysta litotes*, *Exochosphaeridium phragmites*,

Litosphaeridium siphonophorum, *Epilidosphaeridia spinosa*, *Subtilisphaera piranensis*, *Stephodinium coronatum*, *Subtilisphaera perlucida*, *Rhombodella natans*, *Palaeoperidinium cretaceum* and *Palaeohystrichophora infusorioides*.

At 52.40 m, calcareous foraminifera and calcareous nannoplankton occur. Among the foraminifera are rare *Heterohelix* sp. and *Hedbergella* sp. and possibly *Hedbergella delrioensis* that ranges from the Middle Albian to the Late Turonian. The calcareous nannoplankton assemblage is characteristic for the Early Cenomanian because of the co-occurrence of *Gartnerago obliquum* and *Gartnerago nanum*.

In conclusion, the combined evidences from the arenaceous foraminiferal fauna, the nannoplankton and dinoflagellates favour a Late Albian - Early Cenomanian age for the Cretaceous rocks cored within the Quaternary sediments.

Organic geochemistry

The eight rock samples taken from the Cretaceous strata between 44 m and 53 m had a range of total organic carbon contents of between 0.48 wt% at 52.70 m and 1.08/1.09 wt% at 47.29/47.78 m (Figure 3.3.10). Although the total organic carbon content of the samples was fair, the Rock Eval data indicate that the organic matter is typically type IV kerogen associated with low Tmax values of between 418°C and 424°C.

Vitrinite reflectance data were obtained for two samples and values of 0.30% and 0.33% suggest that the kerogen in these rocks is thermally immature. This agrees reasonably well with the Tmax data from these rocks.

The organic geochemical data obtained from the Cretaceous samples indicate that the kerogen in these rocks has no hydrocarbon generation potential.

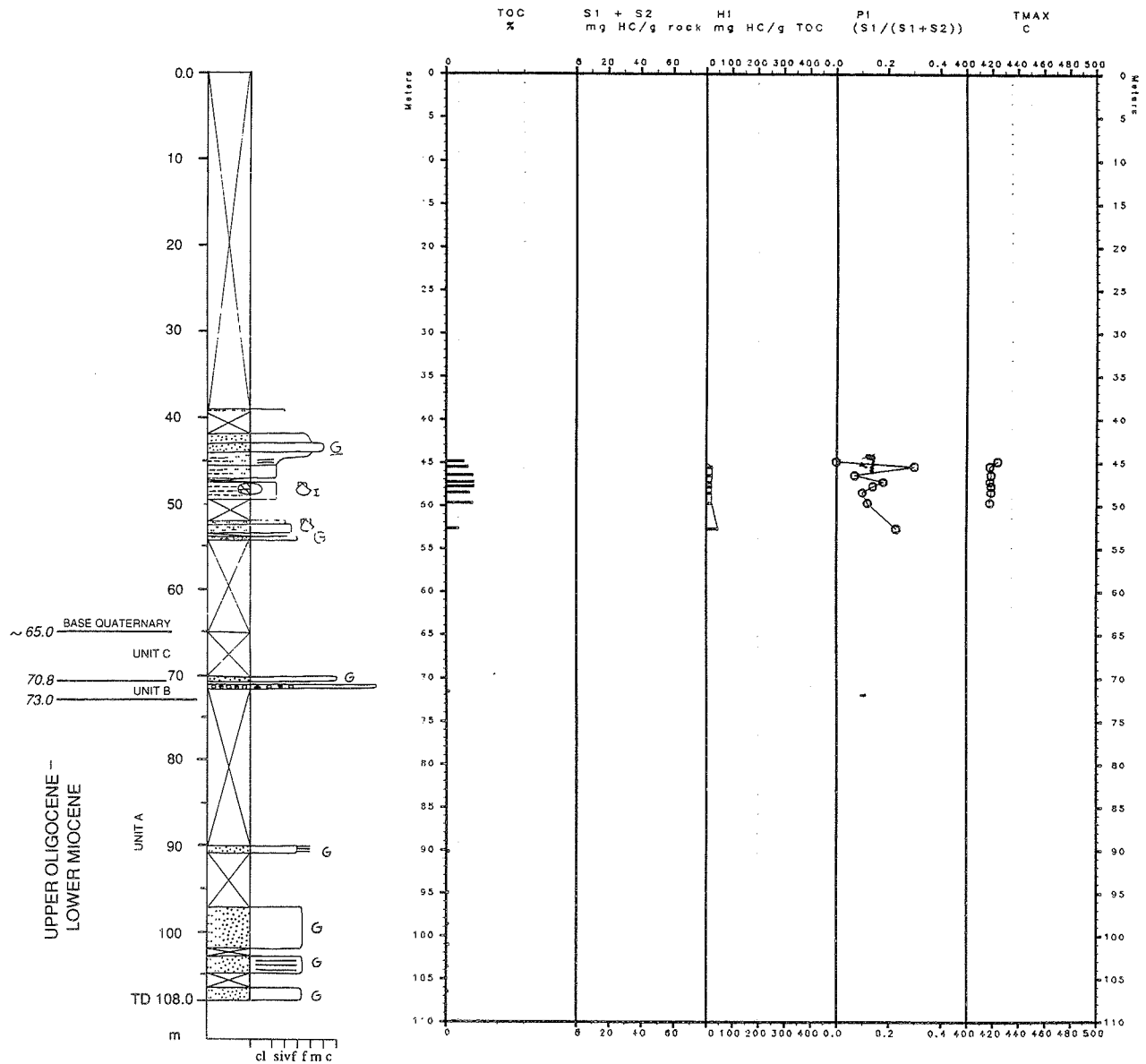


Figure 3.3.15 Total organic carbon content and Rock-Eval data of Cretaceous megablock of core 7316/06-U-01.

3.4 Core 7316/03-U-01

3.4.1 Summary

Core 7316/03-U-01 was the northernmost core drilled in the survey area and penetrated 160.8 m of bedrock.

The core (Figure 3.4.1) represents an overall coarsening upward trend, and based on lithology and seismic data it has been subdivided into six subunits. The lower two subunits (A & B) mainly consist of very fine-grained, glauconitic sandstone. They are overlain by a shaly interval (C & E) interrupted by volcanic debris beds mixed with sandstones (Subunit D). The upper part of the core is dominated by sandstones with abundant volcanic debris throughout (Subunit F).

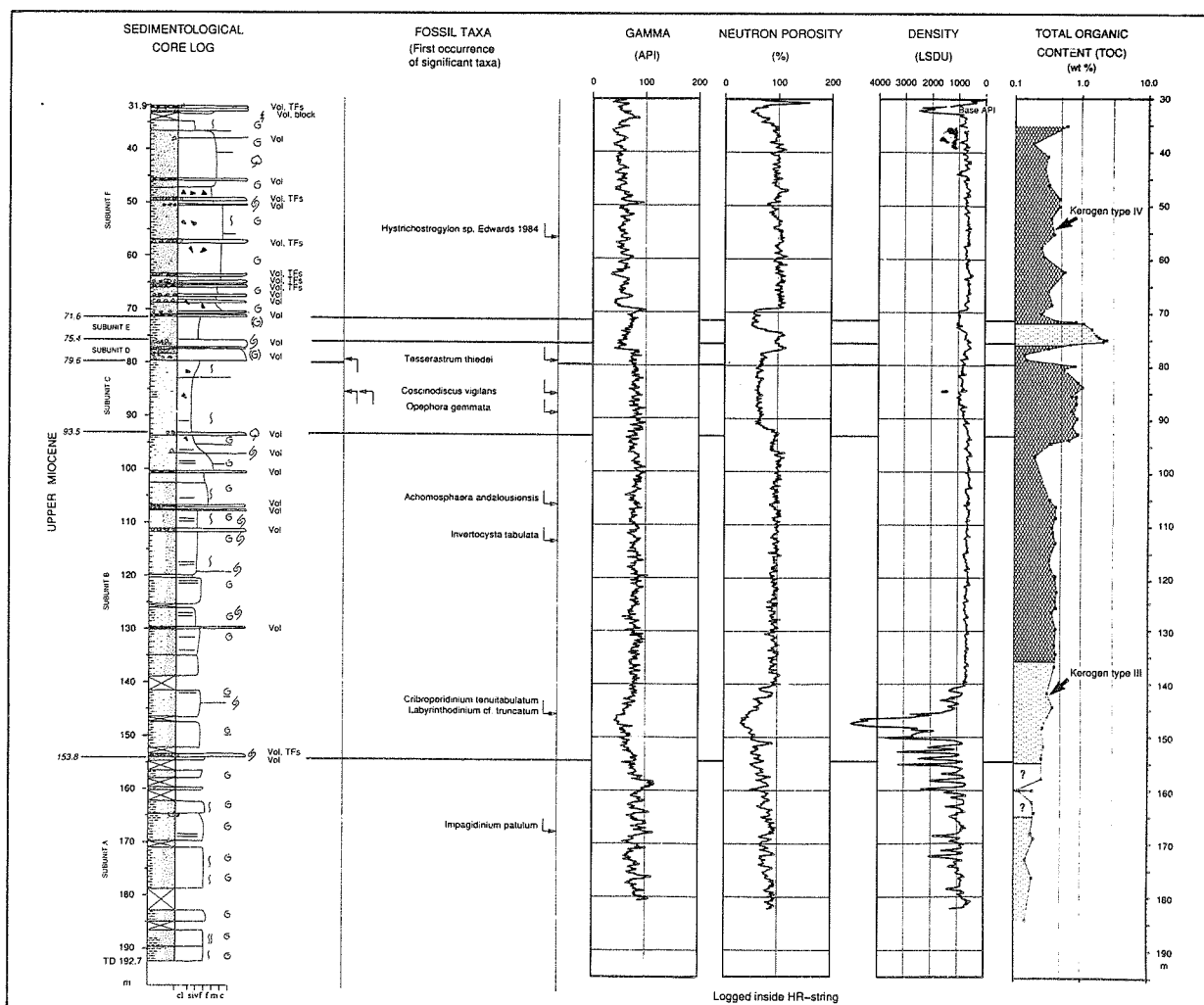


Figure 3.4.1 Summary of main data of core 7316/03-U-01.

The volcanic material occurs from Subunit B upwards as debris of ash, lapilli and blocks, all of basaltic composition. The volcanic material occurs as resedimented clasts embedded in sandstones. These beds always sharply overlie the underlying sediment and often fine upward. Chemical and isotopic composition indicate that the basaltic magma was modified by interaction with continental crust material. The eruption took place through a sedimentary basin involving both subaqueous and subaerial phases.

The sediment contains reworked palynomorphs of early Cretaceous age, abundant palynomorphs of Eocene age and a sparse palynoassemblage of Middle to Late Miocene age. K-Ar dating of volcanic clasts gave ages between 10 Ma and 2.5 Ma. A late Miocene age is concluded for the core.

From seismic data it is suggested that the lower Subunit A is dominated by channel fill deposits. Major parts of subunits B-E (ca. 80 m thick) are interpreted as debris flow/turbidite deposits. On seismic data the reflection pattern of subunits B-E is fairly chaotic on the updip side, but becomes more regular with parallel reflections downdip (NW on seismic line IKU-BV11-89). This, together with a slight distal thinning of the sequence, could indicate a gradual change from dominance of debris flows to turbidity currents downslope. The environmental setting was probably on the upper slope with sediment supply from the outer part of the shelf and slides originating on the shelf edge.

The present dip measured along the seismic line is 6.5° updip of the drill site with a decrease to about 2.5° downdip. Considering the subsidence since the Late Miocene, there was probably still a significant dip at the time of deposition. Knowing that submarine slides frequently occur at dip angles of 1°-2° or even less (e.g. Bugge 1983, table 1), the necessary relief for initiation of debris flows probably existed.

3.4.2 Seismic interpretation

The core was located in the northern part of the surveyed area, on seismic line IKU-BV11-89, sp 1077. From the available seismic data it was planned to penetrate a regional unconformity. From the final processed sections it appeared, however, that the seismic reflector was conformable with the units below and above (Figure 3.4.2).

The Quaternary overburden was 32 m thick and was mainly a glacial till. From the seismic data the cored bedrock section can be subdivided into three units; a lower unit from TD up to about 153 m, a middle unit at 153 - 71 m and an upper unit at 71 - 32 m. The lower seismic unit seems to continue about 0.2 s down to a very strong seismic reflection representing a regional unconformity. This unit is characterized by a fairly chaotic reflection pattern which could indicate channel erosion and infill and some other disturbances that could be interpreted to represent slide deposits and slide disturbed sediments.

The transition to the middle seismic unit is not definite, but could be defined at the top level of a channel, seen to be about 800 m wide and 60 m deep on seismic line IKU-BV11-89 (sp 960 - 1025).

The middle seismic unit is about 80 m thick at the drill site and the reflection pattern at the drill site and updip could indicate dominance of debris flow deposits. From the drill site and out to about shot point 950 the seismic character changes to a less disturbed pattern with parallel reflections. The whole unit shows a slight thinning along seismic line IKU-BV11-89, heading northwestwards. This could be interpreted to represent a gradual change from debris flow deposits to deposition by turbidity currents. Turbidites and possible debris flow deposits in the core seem to support this interpretation. The shaly interval in the core at about 95 - 80 m indicates that episodes of normal marine deposition could have occurred.

The upper seismic unit has a very chaotic seismic reflection pattern, which could represent debris flows and other variations of gravity mass flow, or an environment of general rapid deposition.

The lateral extent of the seismic units encountered is uncertain, as they can be followed on seismic data only where they are above the water bottom multiple. The absolute minimum extent is then approximately 8 km.



bsmp90/REPORT/AA2=kap.3/102/mas 19-FEB-90

Because of the very soft and unstable nature of the cored sediments, there are no sonic log or sound velocity data available. This makes exact correlation difficult, but the most probable correlation between the core and the seismic data gives sound velocities of 1900 - 2000 m/s (Table 3.4.1), which we find reasonable for these poorly consolidated sediments.

Table 3.4.1 Correlation of seismic data and core 7316/03-U-01. Interval sound velocities noted "seismic" appear from this seismic-core correlation.

Depth below seabed (m) (msec)	Core unit	Thickness of seismic unit (m) (msec)	Interval sound velocity (m/sec)		
			Seismic	Core	Log
31.9 37.5	Quaternary	31.9 37.5	1700	-	-
71 74.5	F	39.1 37	1900	-	-
92 95.5	C/D/E	21 21	2000	-	-
154 157	B	62 61.5	2000	-	-
TD192.7 196	A	38.7 39	2000	-	-
Water depth: 297 m (405 msec) Depth below sea level: 489.7 m (601 msec)					

3.4.3 Sedimentology

Introduction

Totally 160.8 m of bedrock was drilled between 31.9 m and 192.7 m below seabed. The core consists of sandstones and minor silty claystones forming an overall coarsening upward trend. The core contains beds with volcanic debris above 153.8 m and a middle clay-rich interval interrupted by volcanic debris beds enabling the core to be subdivided into six units (Figures 3.4.1 and 3.4.3):

Subunit F	31.9	-	71.6 m	thickness =	39.7 m+
Subunit E	71.6	-	75.4 m	"	3.8 m
Subunit D	75.4	-	79.6 m	"	4.2 m
Subunit C	79.6	-	93.5 m	"	13.9 m
Subunit B	93.5	-	153.8 m	"	60.3 m
Subunit A	153.8	-	192.7 m	"	38.9 m+

Minor parts of the core were lost below 125 m and in the uppermost 4 m, nevertheless the core seems to represent a continuous succession.

The core is dominated by sandstone rich in glauconite grains with a variable content of clay and silt. Lamination is sparse throughout the core, mostly due to bioturbation but partly also due to disturbance of the sediment during drilling.

Volcanic material (see Section 3.4.4) occurs at several (ca. 20) different levels within the core between 154 m and 32 m (top) (subunits B, D and F), and its occurrence varies from scattered clasts in the sandstone to distinct conglomerate beds dominated by volcanic debris. 7

Lithology

Subunit A: 192.7 (TD) - 153.8 m. This 38.9+ m thick subunit consists of very fine-grained sandstone with a grey-green colour. It comprises the lower part of the core up to the volcanic debris bed marking the transition to the overlying subunit. This transition at -153 m is also recognized on the seismic record. The sandstone is poorly to not consolidated. The sandstone is quartz dominated, but also contains glauconite, feldspar and mica grains as well as rock fragments. The sandstone contains a fair amount of silt grains and a minor clay content. Sparse bioturbation is observed in most parts of the unit.

Subunit B: 153.8 - 93.5 m. The subunit is 60.3 m thick and mainly consists of glaucony rich very fine-grained sandstone as in the unit below. At the base of the unit two thin beds rich in volcanic debris rest with a loaded contact on the underlying unit. Similar beds with volcanic debris mixed with clastic sediment occur at ten levels throughout the unit. Most of these beds fine upwards, and some of them have a loaded contact at the base. Lamination is seen in parts of the unit. In the upper seven metres the core becomes silty with some mud flakes close to the top. A few well polished subangular pebbles are observed.

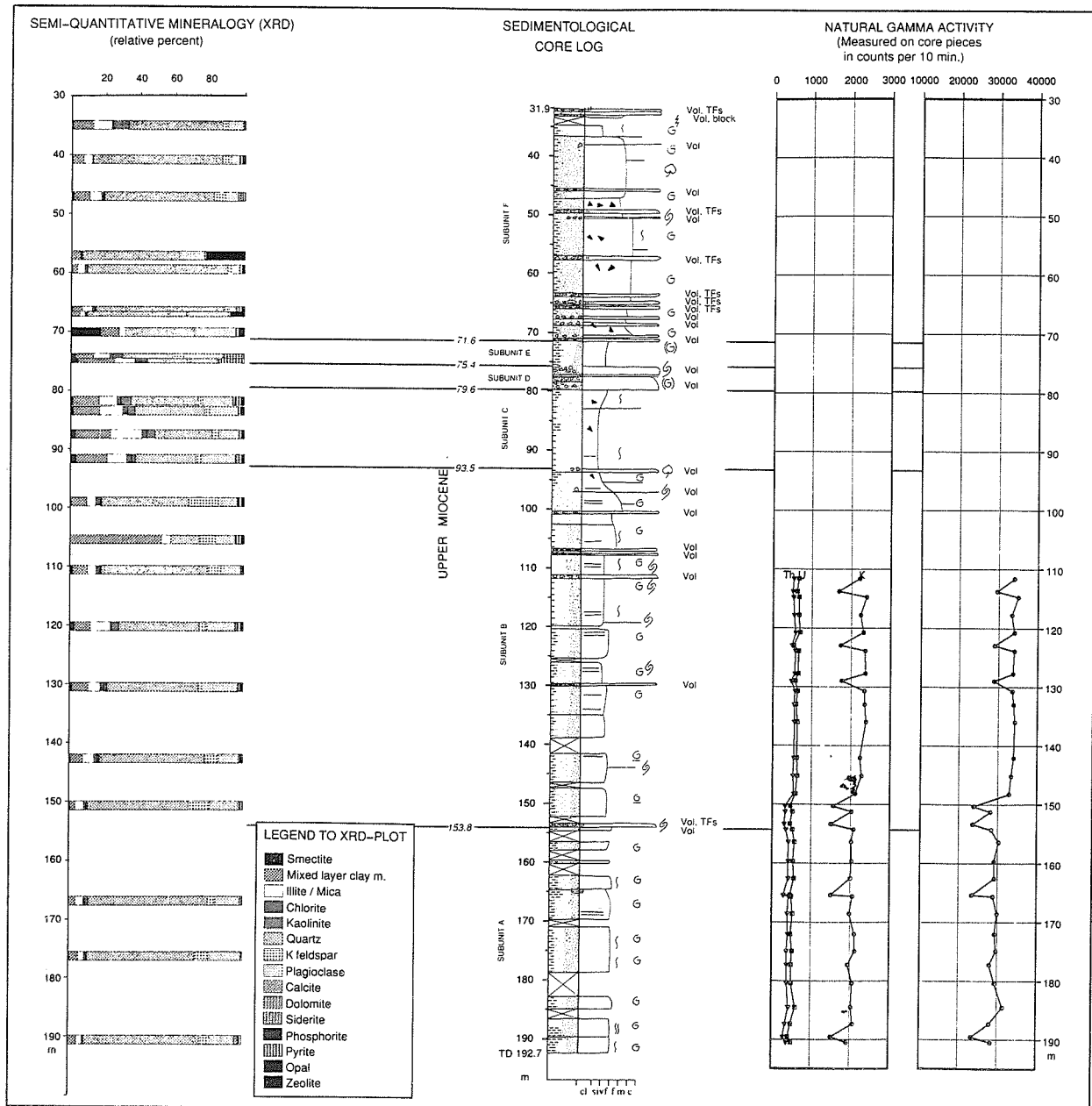


Figure 3.4.3 Lithological column, gamma measurements and XRD data of core 7316/03-U-01.

Subunit C: 93.5 - 79.6 m. The subunit is 13.9 m thick and is dominated by silty dark grey green claystone. The base is defined on top of the uppermost volcanic bed of the underlying unit. However, a gradational reduction in grain size taking place in the upper part of the underlying unit continues into the silty claystone of Subunit C. The silty claystone is partly laminated and shows some bioturbation in the lower part. Mudflakes occur throughout the unit. One well rounded subangular pebble is seen at 86.4 m. It should be noted that from the micropaleontological study it is concluded that a large proportion of the silty claystone may be composed of retransported siliceous ooze (Middle Eocene age).

Subunit D from 79.6 m to 75.4 m is only 4.2 m thick and consists of a series of beds of mixed volcanic debris and sandstone. The beds mainly show poor sorting, and some are fining upwards by reduction of volcanic material as the sand content increases. Some glauconite is present.

Subunit E from 75.4 m to 71.6 m is only 3.8 m thick and consists of poorly laminated grey clayey siltstones that have a minor content of glauconite grains. As in Subunit C micropaleontological studies have indicated that a large proportion of the sediment may consist of siliceous ooze.

Subunit F from 71.6 m to the top of the core (39.7+ m thick) is dominated by fine to medium grained sandstone, glauconitic and with volcanic clasts, occurring both dispersed in the sediment and as beds. The majority of the sand grains are fine, but with a significant content of medium grained sand. The base of the unit is defined where fine to medium grained sandstone overlies siltstone of Subunit E. The lower six metres are dominated by volcanic fragments. Towards the top of the core a clayey siltstone with a disturbed texture has been affected either by the drilling process or by glacial processes. Volcanic basalt blocks or bombs occur near the top of the core. Several well rounded subangular pebbles, partly with a triangular shape, and with small surface marks occur throughout the unit and may represent extrabasinal origin. The surface marks could be due to glacial impact.

Sedimentological description

Sedimentary structures are sparse and mostly restricted to faint lamination. The absence of structures is due to bioturbation and to the very loose nature of the sediment, which therefore has been disturbed physically during the drilling process. Slumping and soft sediment deformation structures are abundant in parts of the core, often below beds with volcanic debris.

Parts of the core are bioturbated (see Appendix A.7), and both infauna and dominichnid traces are observed. Although difficult to see, the ichno diversity and density are often high. While a generally open marine assemblage dominates in the lower part of the core, traces of stationary (?suspension feeders) have been recognized in the upper sub-unit.

Macrofossils have not been recognized in the core except for some undeterminable shell remains at 73 m and a few plant remains.

Mudclasts are present in the top of Subunit B and are abundant in subunits C and F. Grain size distribution analysis has been performed on these poorly consolidated sediments (Figure 3.4.4). Although the analytical results are partly biased as the finer grain sizes (clay and silt) may have

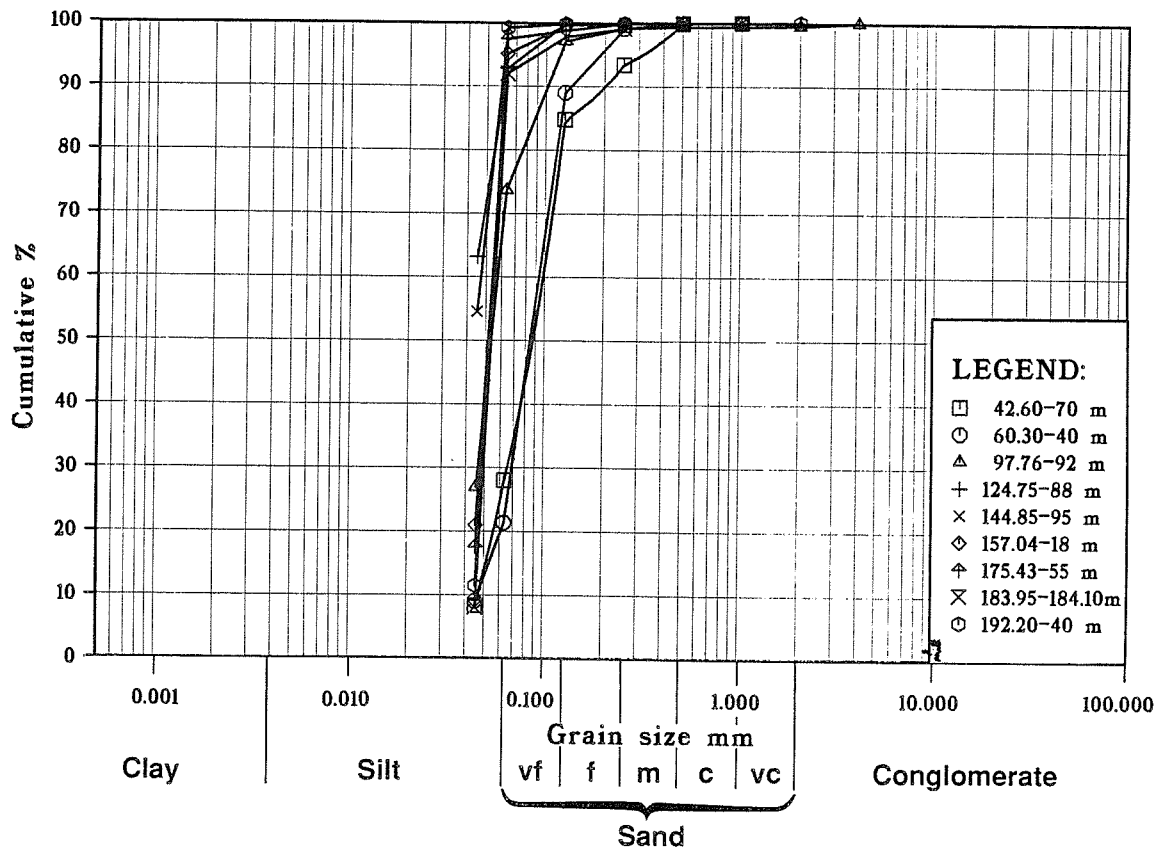


Figure 3.4.4 Grain size distributions of core 7316/03-U-01.

formed aggregates that could not be disintegrated, the analyses clearly show that the lower part of the core (subunits A and B) contains mixed fine-grained sandstone and siltstone, while the uppermost Subunit F contains well sorted sandstone of fine to medium grain size.

The most characteristic feature of this core is the many beds with volcanic debris. The volcanology and geochemistry will be discussed in Section 3.4.4; here the textural and sedimentological relationships will be presented. Altogether 20 beds with volcanic debris have been recognized. At 153.5 m two thin beds with volcanic clasts and tuff represent the first occurrence of volcanic material (Figures 3.4.3 and 3.4.5). Volcanic clasts up to a size of 2 cm are present, and the bed is fining upwards. The volcanoclastic bed is steeply dipping due to slumping.

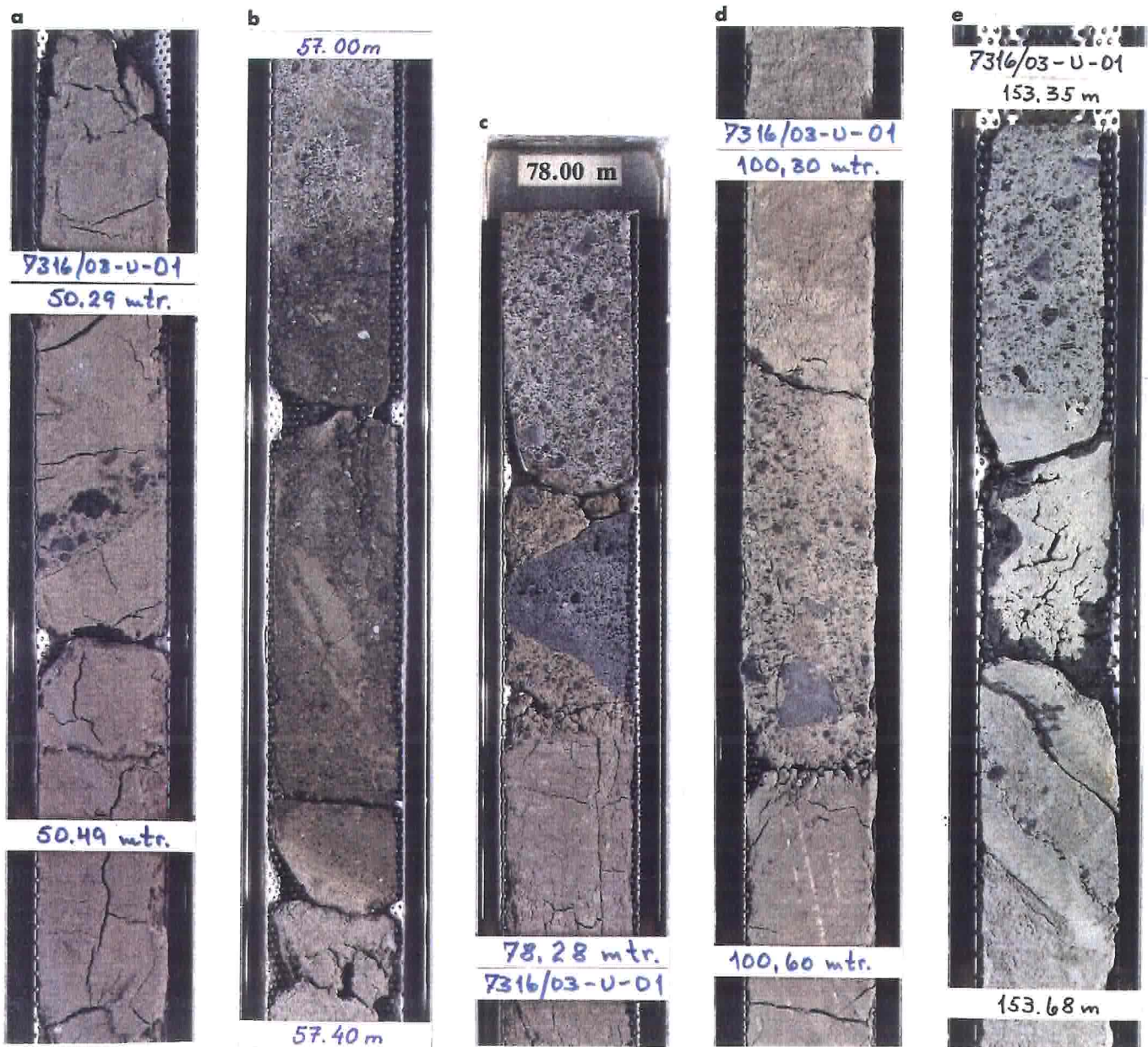


Figure 3.4.5

Photos of core-slabs demonstrating the occurrence of volcanic material in core 7316/03-U-01.

- A thin bed with volcanic debris in sandstone. The oblique bedding is due to loading.
- Above an oblique contact with underlying sandstone a 35 cm thick bed shows increasing content of volcanic material upwards. The light coloured upper part is better cemented than below.
- Volcanic bomb and overlying light coloured tuffaceous lens within volcanic debris rich sandstone. The lower part is silt with a trace fossil, see sketch in Figure 3.4.7.
- A bed with fining upward texture of volcanic debris in a siltstone.
- The lowermost occurrence of volcanic material in the core (base Subunit B). A cemented sandstone bed that contains volcanic debris rests with a loaded contact on poorly cemented sandstone. At top of the photo another debris bed shows a weakly loaded contact (see insinking of volcanic fragment in matrix of sandstone on the right side of the core).

In the upper part of Subunit B several beds with volcanic debris are present and most of them show a fining upward trend. Volcanic material is not present in the sediment between the debris beds.

The intervals 80 - 76 m and 72 - 63 m show a high content of volcanic material and a detailed log is presented in Figure 3.4.6. In these intervals twelve beds are recognized. Most of the beds show fining upwards textures defined by decrease in the coarse volcanic material and increase of the amount of sandstone. Some beds, however, show large pebbles of volcanic material also in the upper part of the beds. Lamination is seldom recognized in the ash-lapilli-sandstone beds, however, the sediments immediately below the volcanics often show planar lamination. Below volcanic beds slump structures are often present. In the interval with volcanic debris at 76.7 - 75.8 m fine-grained clastics show slump structures.

The volcanic debris bed between 78.2 m and 77.7 m (Figure 3.4.7) demonstrates the different structures involved. Underlying the volcanic bed sandy siltstone shows horizontal lamination, and a trace fossil penetrates 5 cm down into the sediment. This siltstone is cut by an erosional contact and sharply overlain by a sandstone that is rich in volcanic debris (ash, lapilli and coarser material, Figure 3.4.7). A volcanic "boulder" (bomb?) larger than the core diameter occurs a few centimetres above the contact in the debris bed. Above this there is a 12 cm thick interval with a high density of volcanic debris that is much better cemented (tuffaceous lapilli sandstone, Figures 3.4.5, 3.4.6 and 3.4.8). The upper part of the bed consists of sandstone rich in volcanic debris and with several large vesicular volcanic clasts several centimetres in diameter. This volcanic rich bed is overlain by a horizontally laminated siltstone that again is overlain, with loaded contact, by another bed rich in volcanic debris.

Most of the volcanic beds in the core have sharp lower boundaries, and are sometimes overlying horizontally laminated sediments. The volcanic beds may, however, show oblique lamination and a disturbed lower boundary due to loading. Several of the beds show graded bedding as the amount of sandstone and silt/clay increases upwards. Nevertheless, large volcanic clasts may occur throughout the beds and occasionally be concentrated in the upper part of beds. The upper contact from the volcanic rich beds to the overlying sediment is mostly gradational.

7316/03-U-01

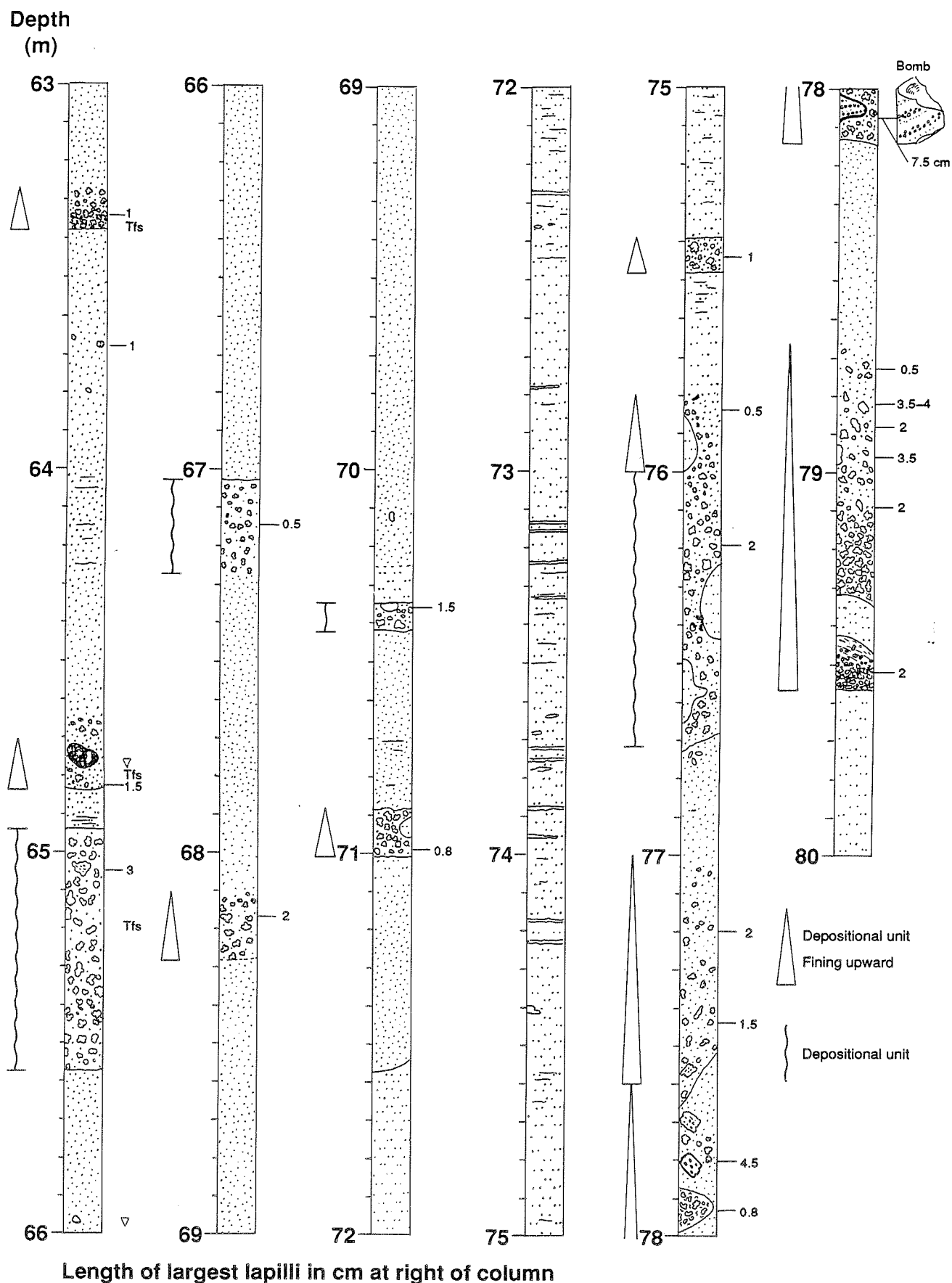


Figure 3.4.6 Occurrences of volcanic debris bed between 80.0 m and 63.0 m alternating with sand and silt without volcanic material.

The sandstones and most of the sandy layers with volcanic debris are poorly consolidated. Some, but not all, of the layers with a high density of volcanic debris are better consolidated (tuffaceous lapilli sandstone). Blocks of consolidated tuffaceous lapilli sandstone have also been detected within less consolidated lapilli sandstone (Figures 3.4.5 and 3.4.7). The possibility that some of the cementation took place in the present position is indicated by the downward gradual increase in cementation degree in the volcanic debris layer at 57 m (Figure 3.4.5), although some of the blocks may have been cemented in another setting before the final deposition.

Petrography

The sandstones are very fine to medium-grained with moderate to well sorting and very poorly consolidated (Figure 3.4.8). Most of the clastic grains are subrounded. The samples are characterized by a high porosity (Table 3.4.2). Matrix is absent or present in small amounts in most of the sandstones studied. There is very low or no cementation in most samples. The sandstones are mainly classified as sublitharenite following Pettijohn et al. (1973).

The major clastic minerals are quartz, plagioclase, alkali-feldspar, white mica, glauconite and chert. Other clastic minerals are biotite, epidote, chlorite, zircon, rutile, garnet, tourmaline, sphene, carbonate and opaques including ilmenite and pyrite (Table 3.4.2). Amphibole (hornblende) appears in sample 43.55 m.

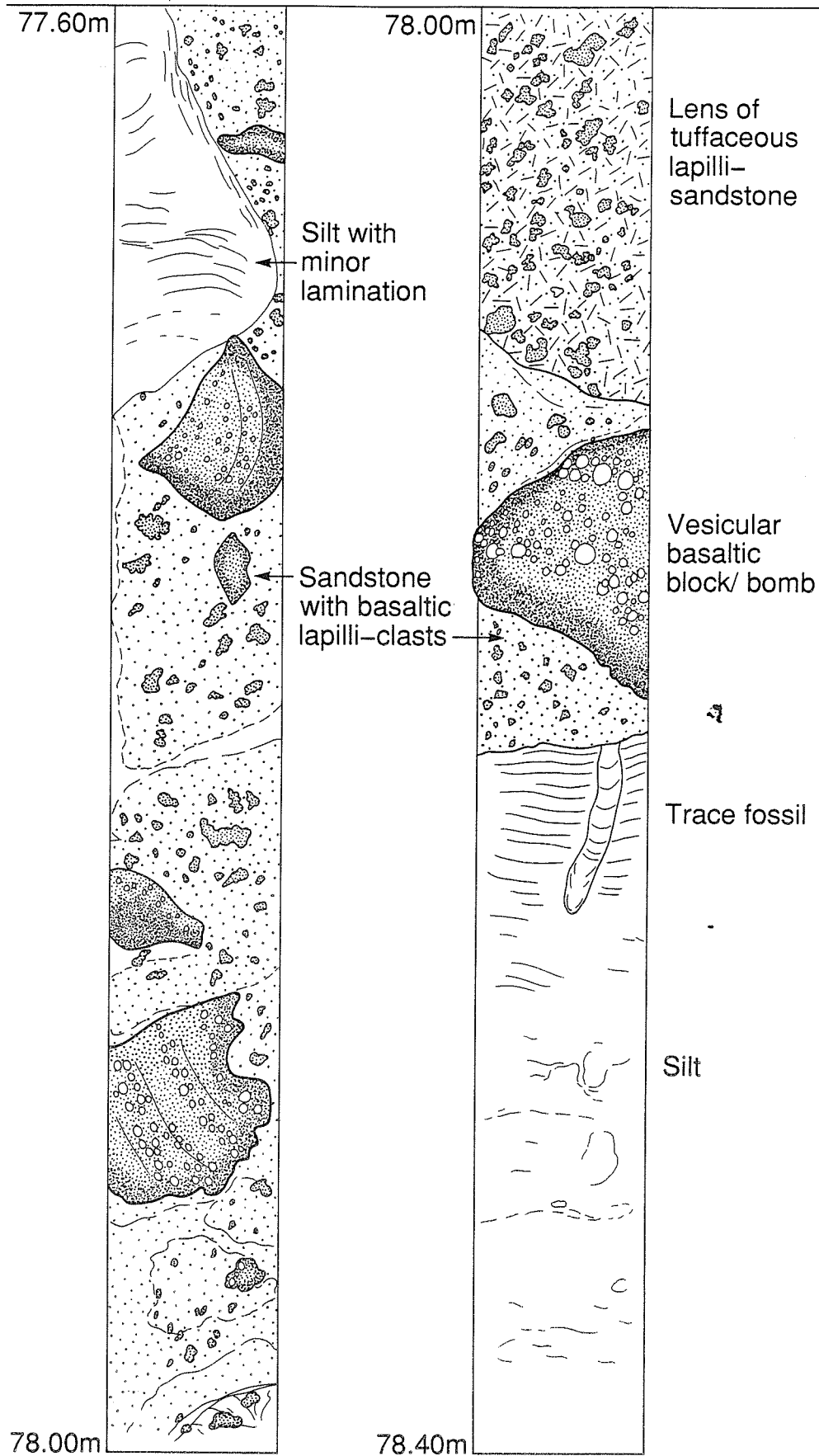


Figure 3.4.7 Details between 77.60 m and 78.40 m of a volcanic debris bed with silt below. The core width is 4.7 cm. See also photo in Figure 3.4.5.

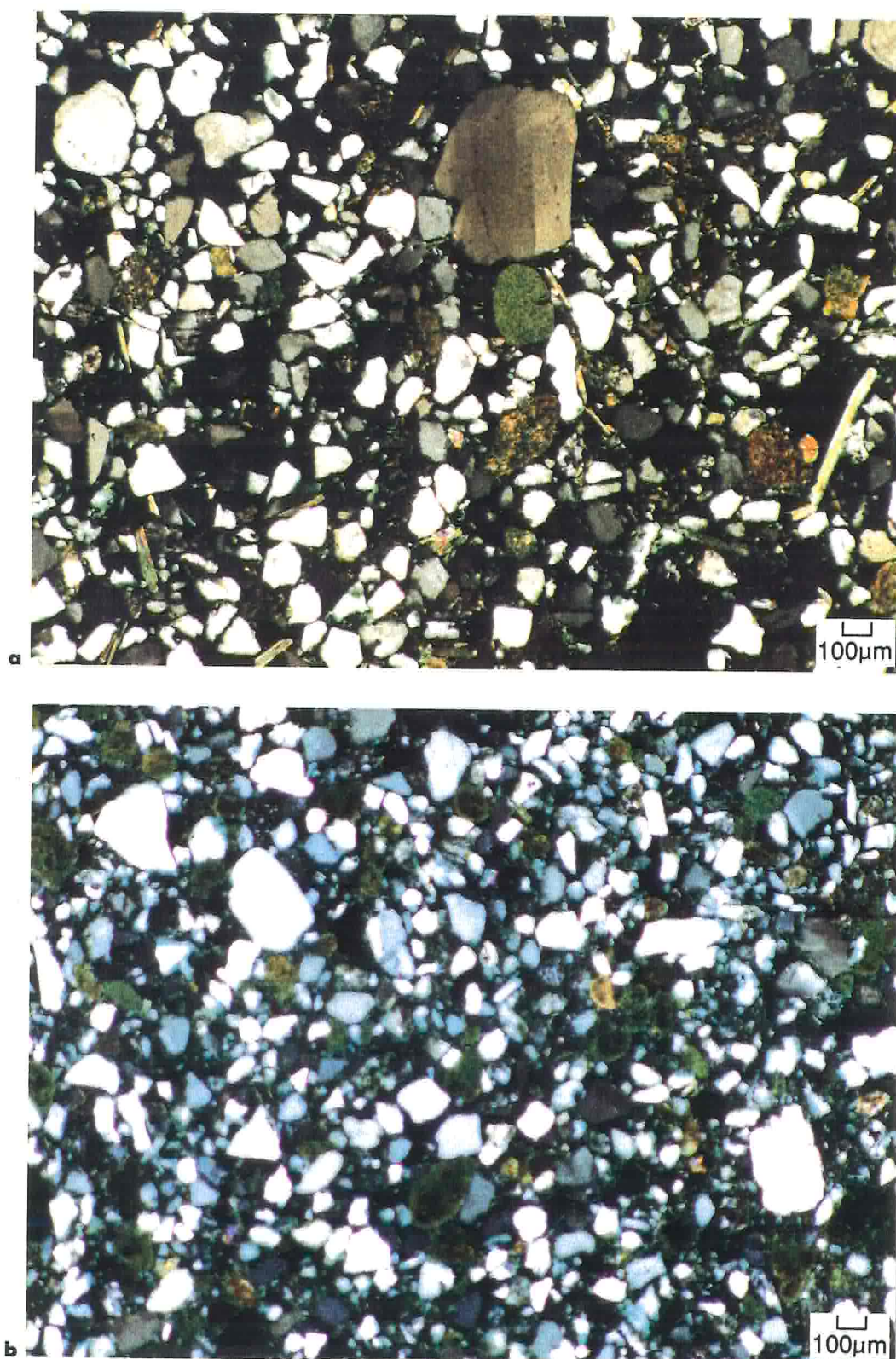


Figure 3.4.8 Optical microphotos of sandstones from core 7316/03-U-01 samples at a) 60.05 m and b) 97.75 m using crossed polars. Clastic quartz (white, grey), glaucony (green, yellowish). Porosity appears black.

Table 3.4.2 Mineral content, petrographic features and some modal analyses of sandstones from core 7316/03-U-01.

SAMPLE NO.	CLASTIC MINERALS																PETROGRAPHY				MODAL ANALYSES											
	MAJOR						ACCESSORY																									
7316/03 -U-01	Quartz	Plagioclase	K-feldspar	Glaucopy	Chert	White mica	Biotite	Chlorite	Epidote	Garnet	Amphibole	Tourmaline	Zircon	Rutile	Sphene	Carbonate	Opakes	Grain-size	Sorting	Rounding	Matrix	% Quartz	% K-fsp	% Plag	% Glaucopy	% Div. lithics	% Matrix	% Porosity	Q + F + L = 100%			
																													Q	F	L	
43.55 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	F	M	SR	(+)	45	3	1	8	5	8	30	72	7	21	
60.05 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	F	M	SR	(+)											
82.80 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	VF/S	P	SA	(+)											
97.75 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	VF/F	M	SR	(+)	39	5	1	2	11	5	27	57	10	33	
124.80 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	VF/S	M	SA	+											
145.00 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	S	M	SA	(+)											
157.20 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	VF/F	W	SR	(+)											
174.40 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	VF	W	SR	(+)											
190.35 m	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	VF	W	SR	(+)	29	9	1	2	11	2	46	56	18	25	

+ = present (positive identification), (+) = rare

Abbreviations:

F = fine
VF = very fine
S = silt
Q = quartz
W = well
M = moderate
P = poor
SR = subrounded
SA = subangular
L = lithics, including glaucony

The quartz grains show different appearances optically including single grains with plane extinction, undulose extinction or low angle boundaries. This indicates that the quartz grains reflect different stages of deformation and annealing. There are also examples of recycled sedimentary quartz grains indicated by rounded quartz overgrowths on rounded quartz cores.

The alkali feldspar is microcline with cross hatched twinning. Clastic plagioclase with distinct polysynthethic twins is common. Two microprobe analyses (An_{15} , An_{31}) of clastic plagioclase in sandstone associated with volcanics have lower calcium content than the basaltic microphenocrysts of the volcanics, indicating a different origin.

Of the clastic heavy minerals, epidote (subhedral) is particularly common in all the samples studied. Zircon is rounded or anhedral indicating a complex sedimentological history.

The glaucony (Figure 3.4.8) is glauconite sensu stricto, with dusty inclusion patterns in central parts after the clastic host grains. Several grains have rounded shapes indistinguishable from sedimentological rounding. Lobate grains with cracks indicate typical mature glaucony growth structures. The glauconite is interpreted to be reworked.

The matrix consists of mixtures of illite, chlorite and partly dissolved biotite, and some quartz. Silica (opal) and carbonate cements appear to be more common in some of the "volcanic" layers than in the sandstones, with highest concentrations of silica in the light coloured tuffaceous layers.

Twenty-four samples have been analysed for bulk mineralogy by X-ray diffraction (Table 3.4.3, Figure 3.4.3). Generally the mineral distribution does not vary much throughout the core, but some interesting features are observed. The amount of clay minerals in Subunit A is small compared with what should be expected from the core description and grain size analyses.

The major clay minerals are mixed layer smectite-illite, mixed layer mica and illite. Smectite is abundant in the 70.18 m sample, a sample collected immediately above a bed rich in volcanic debris. Opal-CT is present in three samples in Subunit F within or above volcanic debris beds.

Trace amounts of amphiboles occur in the coarser grained samples while in the finer grained samples small amounts of zeolite (clinoptilolite) is present. The clinoptilolite has also been identified by SEM (Figure 3.4.9).

Grain size analyses (of grains $>45\text{ }\mu\text{m}$) have been performed on nine samples from subunits A, B and F. Subunits A and B show an increasing content of material finer than $45\text{ }\mu\text{m}$ upwards and the content is never below 8%. The sorting is, however, good with the major part of the grains in the coarse silt to very fine sand fraction. Subunit F on the other hand has most of the grains in the fine to medium fraction and the sand seems not to be as well sorted as below.

Table 3.4.3 *Semi-quantitative mineral composition of the sediments based on X-ray diffraction analysis.*

Depth	Smec	Mix	Mi/Ill	Chl	Kaol	Qtz	KFsp	Plag	Calc	Dol	Sid	Pyr	OpCT	Zeo	amphi	Goeth
35.02	0.9	11.6	10.3	1.0	9.1	53.9	2.6	9.0	0.0	0.0	0.6	0.5	0.0	0.0	0.4	0.0
40.85	0.6	6.4	4.7	0.7	0.0	74.1	5.0	4.8	0.0	0.0	0.6	1.1	0.0	0.6	1.4	0.0
47.15	1.2	9.3	6.4	1.4	0.7	62.9	6.3	7.7	1.4	0.0	0.6	1.8	0.0	0.4	0.0	0.0
57.15	0.0	5.2	0.0	1.0	0.0	55.9	0.0	14.5	0.0	0.0	0.0	0.8	22.7	0.0	0.0	0.0
59.40	0.4	3.0	4.1	1.7	0.0	80.9	1.7	5.1	0.0	0.0	0.6	1.3	0.0	0.7	0.6	0.0
66.50	0.6	5.5	5.6	1.7	0.8	73.7	3.2	4.0	0.2	0.0	0.9	2.6	0.0	0.9	0.5	0.0
66.75	0.0	4.8	2.8	1.4	0.0	57.1	1.0	23.1	0.0	0.0	0.7	1.2	6.8	0.9	0.2	0.0
70.18	16.7	10.8	3.8	0.0	0.0	39.9	0.0	23.8	0.0	0.0	0.0	1.8	3.3	0.0	0.0	0.0
74.50	2.2	10.7	9.1	2.1	5.6	28.8	6.6	22.5	0.6	0.0	0.7	10.7	0.0	0.5	0.0	0.0
74.60	3.8	21.3	11.5	7.5	0.0	22.9	2.0	15.1	0.9	0.0	0.6	14.4	0.0	0.0	0.0	0.0
81.81	1.0	15.4	10.2	2.3	5.9	38.8	2.4	17.4	0.3	0.0	1.1	3.3	0.0	2.0	0.0	0.0
83.58	1.2	15.6	13.1	3.0	4.4	40.0	2.7	16.4	0.3	0.0	1.4	0.4	0.0	1.6	0.0	0.0
87.54	2.7	20.3	18.1	3.0	4.5	32.7	3.9	11.5	0.2	0.0	0.8	0.8	0.0	1.4	0.0	0.0
91.75	3.2	18.1	11.1	3.2	2.1	34.6	3.0	19.5	0.3	0.0	0.8	2.6	0.0	1.6	0.0	0.0
99.01	0.7	9.0	5.0	3.2	0.4	50.3	16.6	10.9	0.2	0.0	0.8	2.2	0.0	0.7	0.3	0.0
105.43	0.0	52.7	5.2	0.0	0.0	16.5	8.9	11.4	0.2	0.0	0.5	3.3	0.0	0.0	0.0	1.4
110.62	1.5	8.9	4.7	2.5	0.7	60.9	7.6	11.2	0.0	0.4	0.7	0.0	0.0	0.7	0.3	0.0
120.20	2.5	9.5	11.5	0.9	4.2	46.3	4.2	16.1	0.4	0.0	0.9	2.7	0.0	0.9	0.0	0.0
130.58	1.2	9.8	6.8	2.3	1.9	52.5	1.8	20.6	0.2	0.0	0.7	1.3	0.0	0.9	0.0	0.0
142.75	1.0	6.8	6.5	2.1	1.1	60.4	7.9	11.6	0.0	0.0	0.9	0.9	0.0	0.8	0.0	0.0
150.75	0.4	3.1	4.8	1.4	1.0	58.0	12.1	16.5	0.0	0.0	1.1	1.3	0.0	0.4	0.0	0.0
166.90	0.2	3.8	4.3	1.1	2.0	68.6	4.2	14.2	0.0	0.0	1.1	0.5	0.0	0.0	0.0	0.0
176.35	0.7	5.1	3.1	1.2	0.7	61.8	8.0	18.5	0.0	0.0	0.7	0.3	0.0	0.0	0.0	0.0
190.67	0.1	4.5	3.1	1.2	0.0	66.8	5.6	14.5	2.0	0.0	0.9	1.4	0.0	0.0	0.0	0.0

-3

Diagenesis

The observed cementation is mainly restricted to carbonate and opal cementation in some of the volcanic layers (Section 3.4.4). This association is a common late alteration product associated with palagonitization of tuffs (cf. Fisher & Schmincke 1984). Opal may also appear early during alteration of glassy rocks (Jakobsson 1972).

Clinoptilolite is a common low-temperature zeolite in marine sediments where it may have formed by dissolution of volcanic glass or via diagenesis of biogenic siliceous skeletons (Fisher & Schmincke 1984). Kirsch (1983) gives an example of clinoptilolite formation at burial depths of 0.9 - 2 km in Neogene pyroclastics elsewhere.

A low degree of diagenesis is indicated in general by the excellent preservation of volcanic glass (Section 3.4.4), by the preservation of labile clastics such as amphibole and plagioclase and by the very low degree of cementation. Another evidence is the preservation of opal-CT (XRD data) which indicates diagenetic temperatures less than 60-70°C (Bjørlykke et al. 1989). In addition, non-recrystallized siliceous microfossils (Opal-A) are reported from this core. In the absence of impurities these would indicate

temperatures below ca. 40°C, but it is also possible that they were preserved metastably during the burial. Low diagenetic temperatures are also supported by the apatite fission track data which concludes that the samples have not been subjected to temperatures in excess of 50°C at any time since deposition, but the method does not allow more detailed estimates at the low temperature of interest (Appendix A.8).

Provenance

The clastic mineralogy indicates that the sediments have been derived from metamorphic source rocks. The presence of garnet, biotite, white mica, chlorite, oligoclase, epidote, quartz and hornblende suggests that the sedimentary source areas typically contained metamorphic rocks of medium grade, while low grade metamorphic rocks are also likely to be represented. Recycling of the sediments is evidenced from the frequent appearance of reworked fossils, the redeposited sand grains and the high content of glauconite.

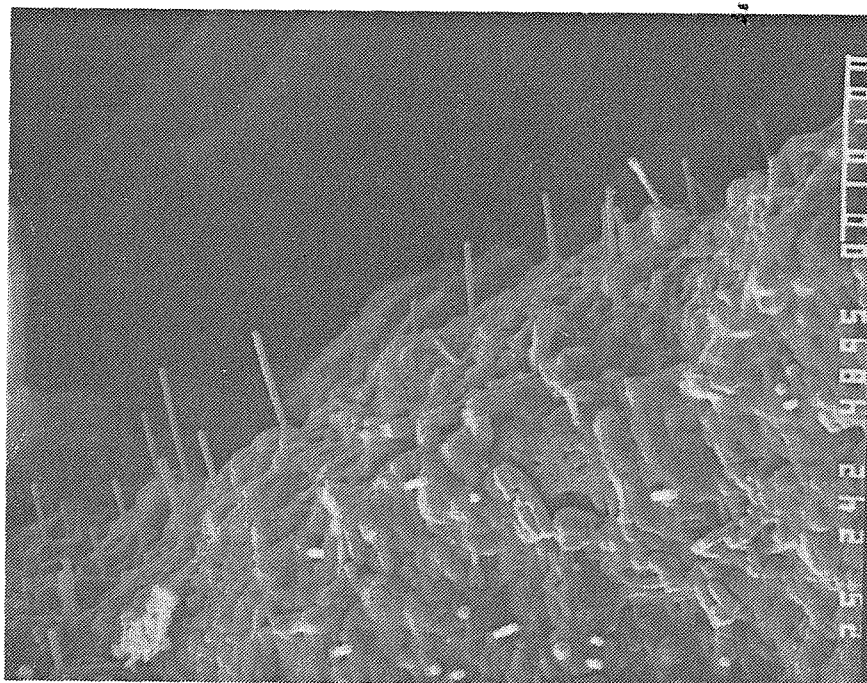


Figure 3.4.9 SEM image of clinoptilolite needles growing on clastics in core 7316/03-U-01, 53.9 m.

The volcanic debris layers give evidence for the existence of Late Tertiary igneous activity in the vicinity of the sedimentary basin (Section 3.4.4). Outside the volcanic debris layers the core does not contain original igneous mineral clasts. However, metamorphosed, or altered igneous rocks of

unknown age may be represented by some of the heavy minerals (e.g. amphiboles, epidote, zircon) within the sandstone layers.

Apatite fission track age data contain components of two different age groups pointed out elsewhere (Appendix A.8); 80 - 100 Ma and 200 - 400 Ma. The older group can be interpreted in terms of Caledonian - Hercynian basement rocks. The report by Green & Marshallsea in Appendix A.8 indicates that the younger ages can reflect Late Cretaceous or Early Tertiary uplift or volcanism. This hypothesis is supported by common reworked Late Cretaceous and Early Tertiary fossils into the sediments of core 7316/03-U-01.

3.4.4 Volcanics

Occurrence: The occurrence and distribution of the volcanics have been described in Section 3.4.3 and are illustrated in the core logs in Appendices A.12 and A.2 and Figures 3.4.3, 3.4.5, 3.4.6 and 3.4.7. The volcanic material is basaltic and mainly of pyroclastic origin. Vesicular lapilli, ash fragments and a few lava blocks or bombs appear, intermixed with sediment matrix in the interval between 154 m and 32 m depth. The distribution varies from the occurrence of a few isolated lapilli clasts within sandstone to distinct beds with a high density of volcanic material. The bed thickness is typically one or a few decimetres, the thickest interval with high density of volcanics being the four metres thick section in Subunit D between 79.5 m and 75.5 m (Figure 3.4.6).

The size of the volcanic material ranges from a few μm ash fragments up to several cm sized lapilli clasts. The largest volcanics may represent a lava block (16 cm thick, 32 m level) and volcanic bomb or block (>7 cm, 78 m). The blocks or bombs are vesicular with relicts of chilled margins.

The most consolidated layers with volcanic detritus have been classified as tuffaceous lapilli-rich sandstone. These are characterized by a high proportion of volcanic clasts and matrix glasses (ash) relative to the sand matrix. The consolidated appearance is in part related to carbonate and silica cements (Figures 3.4.5, 3.4.10 b). It is also possible that fine-grained volcanogenic material can have been replaced by alteration products.

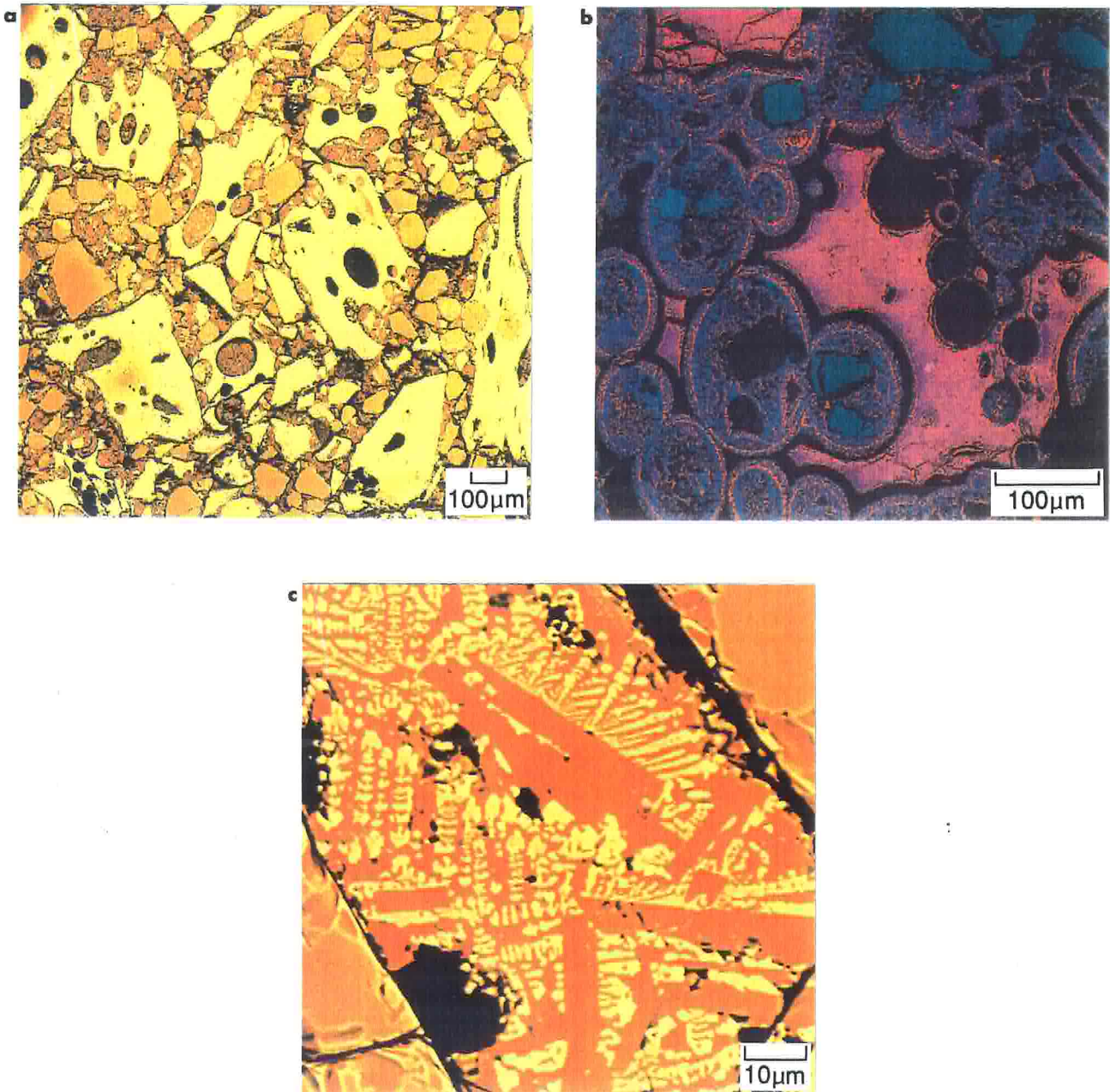


Figure 3.4.10 Back scattered electron image of volcanics in core 7316/03-U-01.

- a) Homogeneous, vesicular glasses (light yellow) of ash to lapilli size situated in sediment with quartz grains (dark yellow). Sample 57.15 m (subarea 1).
- b) Tuffaceous lens with vesicular glass fragments (reddish). Clastic quartz and feldspar grains (blue) and fine silica cements (also blue) are filling the pore spaces. Black domains are empty spaces. Material has been lost around the glass rims. This could possibly indicate that alteration products (palagonitization?) have been present earlier.
- c) Detail of very fine-grained lava matrix between two olivine phenocrysts in basalt (32.05 m) showing an intergrowth of plagioclase microlites (orange) and clinopyroxene (yellow). While olivine formed at depths in the magma chamber, the plagioclase and pyroxene intergrowth formed at a later stage by quenching during eruption.

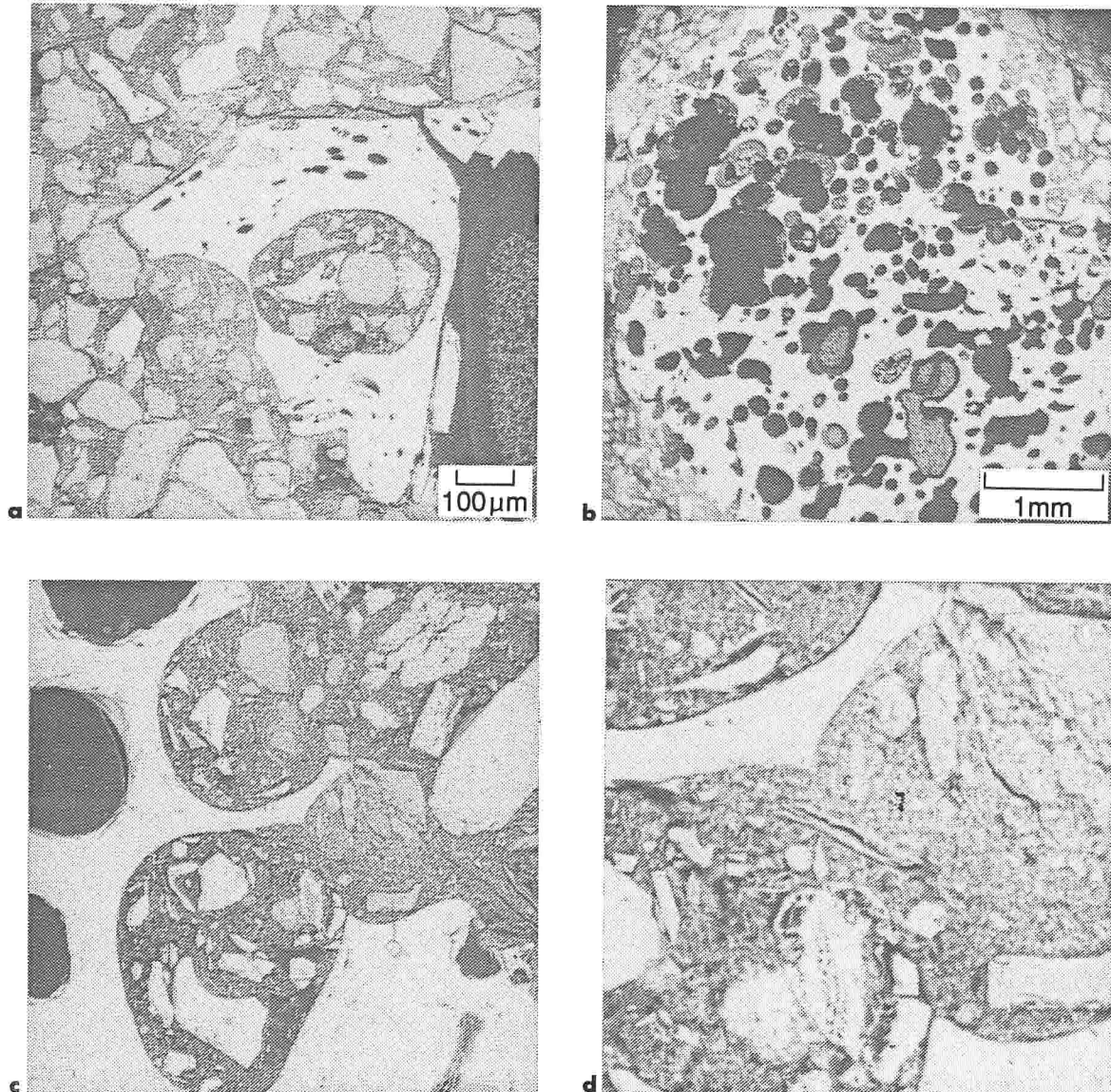


Figure 3.4.11 Back scattered electron images of the smaller volcanic fragments in core 7316/03-U-01.

- a) Vesicular basaltic glass (light) in sediment matrix (49.57 m). Note the presence of sedimentary quartz grains (grey) and small glass fragments (light) in the sediment inclusions.*
- b) Vesicular basaltic glass (light) in sediment (49.57 m). Empty pore spaces appear black. (Some have been filled with epoxy; appearing greyish).*
- c) Detail of vesicular basaltic glass (light) illustrating the sediment glass relation (49.57 m). Sediment - glass have been intermixed after the quenching. Small light fragments also show the presence of fine glass particles in the sediment filling the pore spaces.*
- d) Close-up of central part of c showing thin (a few μm thick) glass threads (light) in sediment matrix.*

Matrix sediment appears also within disrupted and open pore spaces in vesicular volcanic clasts (Figures 3.4.11 and 3.4.12). There is some evidence for physical mixing of sediment into the melt when still being hot. This is shown by the appearance of sediment xenocrysts in some of the glasses, particularly at 77.10 m where quench crystals in the glass have formed at the contact with a xenocryst of rounded clastic quartz (Figure 3.4.12). However, most of the textural sediment - glass relations have been established after the solidification of the glass.

The volcanic clasts have generally been excellently preserved from secondary alterations, unaltered glasses being preserved throughout the core.

Petrography

Basalt lava blocks: The basalt (32.05 m) is vesicular and aphanitic with olivine phenocrysts. The groundmass appears black and oxidised. Plagioclase microlites (100 - 200 μm) show a random distribution in the matrix. Very fine-grained matrix clinopyroxene has been detected using back scattered electron images; plagioclase and clinopyroxene forming a branching network of quench crystals (Figure 3.4.10c). Olivine (70 μm - 1.4 mm) shows different habits occurring as euhedral and subhedral phenocrysts and glomerocrysts, and also as embayed and skeletal phenocrysts intergrown with matrix.

Volcanic clasts: The largest lapilli clasts (2 - 6 cm) and bombs are vesicular with a black oxidised matrix. Homogeneous light brown glass has, however, been preserved as rare isolated preserved pockets adjacent to glomerocryst aggregates in some cases (cf. 106.73 m, 107.55 m).

Olivine is abundant as phenocrysts and glomerocrysts occurring as euhedral, subhedral or embayed and skeletal grains. Euhedral Cr-spinel has been detected as inclusions in olivine. Plagioclase microlites are abundant in a few clasts only (78.08 m, 111.45 m) showing a small grain size (60 μm - 1 mm, 60 μm - 0.1 mm). Very fine-grained interstitial clinopyroxene at 78.08 m has a similar appearance as in the above described basalt (32.05 m).

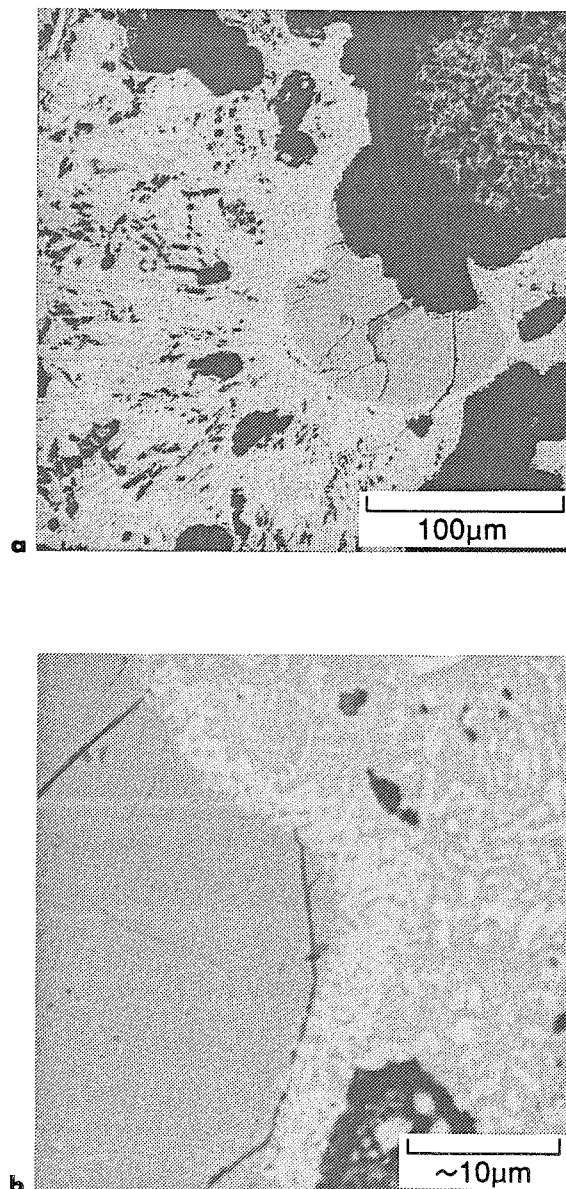


Figure 3.4.12 Back scattered electron images of quartz xenocryst in lapilli in core 7316/03-U-01 (77.10 m).

- a) Rounded clastic quartz grains (dark grey) in lava matrix containing microlites of plagioclase (light grey) and clinopyroxene (light). Note radiating fabric of microlites around the quartz.
- b) Close-up of (a) showing a rim of clinopyroxene crystallites which have nucleated on the clastic quartz. This shows that the quartz grain was incorporated into the basalt magma when the melt was hot suggesting magma eruption through sediments.

The smaller clasts show a range of sizes from 0.01 mm thin threads to 2 mm sized clasts and several mm to cm sized clasts. Most of the clasts are vesicular, but many of the smaller clasts appear less vesicular. This may reflect that they are derived from larger vesicular lapilli, but splintering of glassy material due to rapid cooling under water or ice are also possible mechanisms.

The lapilli consist of excellently preserved basaltic glass. Light brown vesicular homogeneous glass is the dominant type. Dark brown smaller fragments and vesicular black (oxidised) fragments have a much lower abundance, but they are represented by a few fragments in the samples from all parts of the core (Table 3.4.4).

Table 3.4.4 Petrographic characterization of volcanics in core 7316/03-U-01 (+ present, * dominating).

SAMPLE		SIZE OF CLASTS			MATRIX, GLASSES			CRYSTAL CONTENT			MINERALS			SEDIM.		CEMENT		OLIVINE phenocrysts	PLAGIOCLASE microclites			
THIN SECTION	SAMPLE CHARACTERIZATION	<0.2 cm ASH	0.2-2 cm LAPILLI	BOMB? BLOCK	Light brown	Dark brown	Black, oxidized	Crystal-free	Crystal-bearing Microlites	Very fine microl.	OLIVINE pheno.	PLAG. microl.	CPX microl.	Cr-Spl.	XENOCRYST	SEDIM. in vesic.	SEDIM. MATRIX			CARBONACEOUS	OPAL/SILICA	
7316/03-U-01																						
32.05	Lava	-		+			*		+	+	+	+	+	+							85,82-78	57
32.10	Lava	-																				
38.13	Clasts in sed.	-	+		+		*		+	+	+	+				+	+				85	
49.22	Tuffaceous sed.	+	+		+	+	+	+	+	+	+	+									81-84	65
49.45	layer	+	*		+	+	+	+	+	+	+	+			+		+	+	+		84	62-64
49.57		+	+		+	+	+	+	+	+	+	+			+		+	+	+		83	
57.15	Tuffaceous	+	+		*	+	+	+	+	+	+	+							+		84	
64.70	Tuffaceous	+	+	?	+	+	*	+	+	+	+	+				+	+	+	+			
65.20	Tuffaceous	+	+		*	+	+	+	+	+	+	+				+	+	+	+		83-84	
70.18	Tuffaceous	+	+		*	+	+	+	+	+	+	+				+	+	+	+			
71.00	Sed. w. lapilli clasts	+	+		*	+	+	+	+	+	+	+				+	+	+	+			
75.50	Sed. w. lapilli clasts	+	+		*	+	+	+	+	+	+	+				+	+	+	+			
77.10	Sed. w. lapilli clasts	+	+		+	+	*		+	+	+	+	+			+	+	+	(+)			
77.83	Sed. w. lapilli clasts	+	+		+	+	+		+	+	+	+	+			+	+	+	+			49-57
77.97	Tuffaceous, light col.	+	+		+	+	+	+	+	+	+	+	+			+	+	+	+			55
78.05	Vesicular bomb	-	+	+		+	*	+	+	+	+	+	+			+	+	+	+	+	85-86	50-58
79.55	Sed. w. lapilli clasts	+	+		*	+	+	+	+	+	+	+						+	+			
106.73	Lapilli clast	-	+		(+)	+	*	+	+	+	+	+	+			+	+	+	+		84-85	61
107.55	Lapilli clast	-	+		(+)	+	*	+	+	+	+	+	?			+	+	+	+		84-85	
111.45	Lapilli clast	-	+			+	*	+	+	+	+	+									85	
129.90	Lapilli clast	+	+		*	+		+	+	+	+	+			+	+	+	+	+		84-86	49-60
153.35	Lapilli in sediment	+	+		+	+	+	+	+	+	+	+				+	+	+	+		84-86	

Glass with olivine phenocrysts has been detected in all the samples, but the modal content is usually very low, and in the group of ash sized glass

fragments crystal free glasses dominate. Olivine has a similar appearance (Figure 3.4.13), but showing a lower abundance than in the larger clasts. Very fine-grained plagioclase microlites and matrix clinopyroxene have been detected in a few clasts (Table 3.4.4).

Phenocryst compositions: Most of the olivine phenocryst analyses fall within a limited compositional range Fo_{83} - Fo_{86} with no systematic variation with core depth (Table 3.4.4). Very fine-grained matrix olivine in the basalt at 32.05 m has slightly lower Fo-content (Fo_{78-82}).

The fine-grained plagioclase microlites are mostly labradorite ranging between An_{49} and An_{65} (Table 3.4.4). The compositional differences are related to size differences which may reflect formation at different stages of cooling.

Alteration: Most of the smaller glass fragments show no sign of alteration. Some of the smaller glasses and most of the largest several cm sized clasts have a dark and oxidised appearance. The ash to fine lapilli sized fragments within the tuffaceous lens at 77.97 m show the most extensive transformation into homogeneous black opaque material (red in the colour-print, Figure 3.4.10 b). The pore spaces have been filled with a mixture of dark material, opal and sediments. The structure resembles palagonite (hydrated, altered glass), - the strongly altered outer rim has been dissolved. Palagonitization may occur at fast rates in hydrothermal environments, but it is also a common alteration process during low-temperature weathering and alteration of volcanic glass (cf. Fisher & Schmincke 1984). Minor carbonate cements appearing within pore spaces and sporadically seen to replace olivine also record alteration of volcanic material. It is also likely that clinoptilolite and smectite recorded in the sandstones (Section 3.4.3) are alteration products of glass or palagonite.

Geochemistry

Chemical variation: Quantitative electron microprobe analyses (IKU Super-probe) of major element of the glassy volcanics have been performed to determine the compositional variation. The chemical composition of the homogeneous and unaltered glass is representative for the basaltic melt at the stage of eruption. Volcanic samples from 18 different levels have been

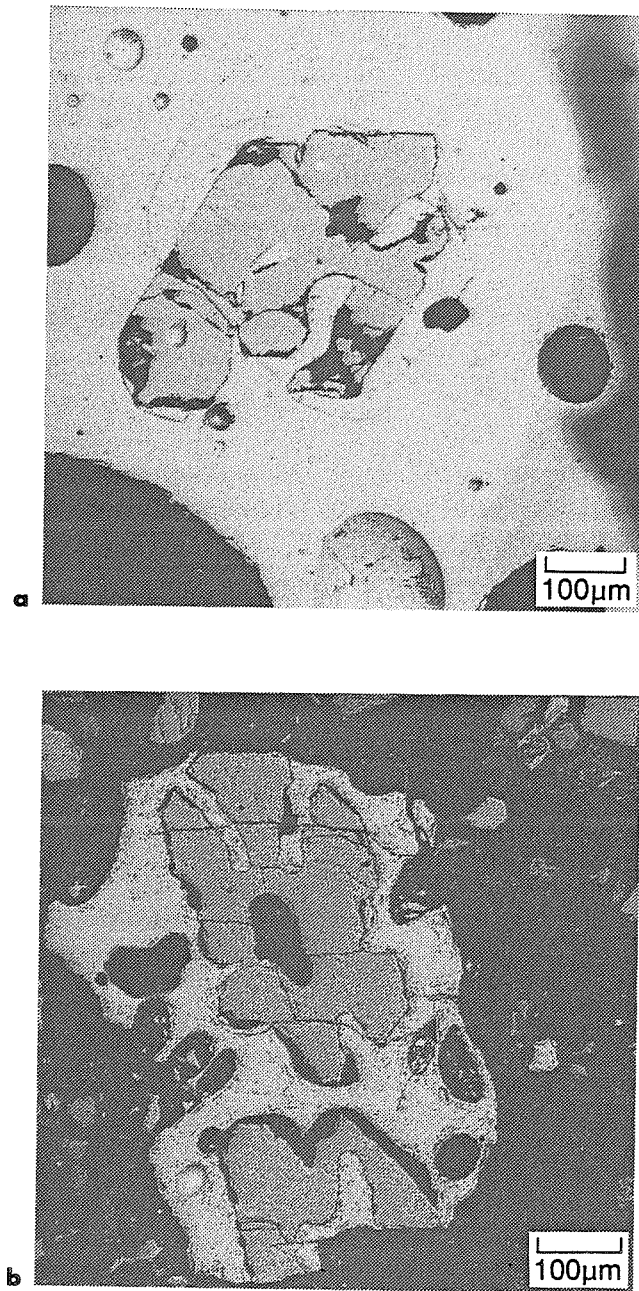


Figure 3.4.13

Back scattered electron images of olivine phenocrysts (dark grey) in vesicular basaltic glasses (light grey). a) Lapilli from sample 129.90 m. Olivine shows embayed appearance with glass inclusions. Circular cracks have formed around the phenocryst during cooling. b) Ash fragment from sample 153.35 m with branching olivine. The olivine structures indicate rapid growth of the phenocrysts in the magma chamber.

analysed. The electron microprobe analyses of glasses have been performed using a slightly defocused beam (10 μm) in order to prevent loss of alkali elements during the analyses. Single analyses are listed at the end of Section 3.4 (Table 3.4.8). XRF whole-rock analyses of major and trace-elements and ferrous iron titration and H_2O analyses have been performed on two massive samples at Washington State University and at the Norwegian Institute of Technology (NTH) in collaboration with Professor Tore Prestvik (Tables 3.4.5 and 3.4.8).

Table 3.4.5 Chemical composition of volcanics represented by mean values of electron microprobe analyses of glasses (a,b,c) with standard deviations shown in brackets and XRF-whole-rock analyses of two basaltic blocks (d,e). Ferric/ferrous iron ratios analysed for the two whole-rocks have been used as a basis to calculate ferric iron for the glasses using $\text{Fe}_2\text{O}_3/\text{FeO}=0.25$.

	a	b	c	d	e
	Light brown glasses	Dark brown glasses	Black oxidised fragments	32.05 whole rock	78.05 whole rock
SiO_2	51.33 (0.77)	51.42 (0.96)	51.99 (0.66)	50.55	51.18
Al_2O_3	15.86 (0.36)	15.89 (0.19)	16.14 (0.88)	15.08	16.15
TiO_2	1.46 (0.16)	1.46 (0.18)	1.58 (0.29)	1.44	1.53
Fe_2O_3	1.91	1.94	1.85	1.49	1.97
FeO	7.62 (0.54)	7.76	7.42 (1.30)	7.67	6.41
MnO	0.13 (0.05)	0.15 (0.06)	0.15 (0.05)	0.14	0.13
MgO	6.21 (0.53)	6.30 (0.22)	5.08 (1.06)	8.95	7.20
CaO	9.03 (0.55)	9.02 (0.23)	9.44 (1.28)	8.95	9.50
Na_2O	3.67 (0.36)	3.44 (0.58)	3.47 (1.09)	3.63	3.92
K_2O	0.93 (0.12)	0.85 (0.16)	0.95 (0.22)	1.03	1.06
P_2O_5	0.25 (0.07)	0.24 (0.05)	0.30 (0.09)	0.22	0.23
L.O.I.	n.a.	n.a.		0.72	1.11
Sum	98.40 n= 111	98.47 n=17	98.37 n=12	99.87	100.39

The glasses are olivine and hypersthene normative basalts. Subalkaline to transitional alkaline basaltic affinities are indicated using the classification schemes of Irvine and Baragar (1971) and Miyashiro (1978) (Figure 3.4.14). The relatively high silica content combined with high Mg/Fe ratios and high alkali content is notable.

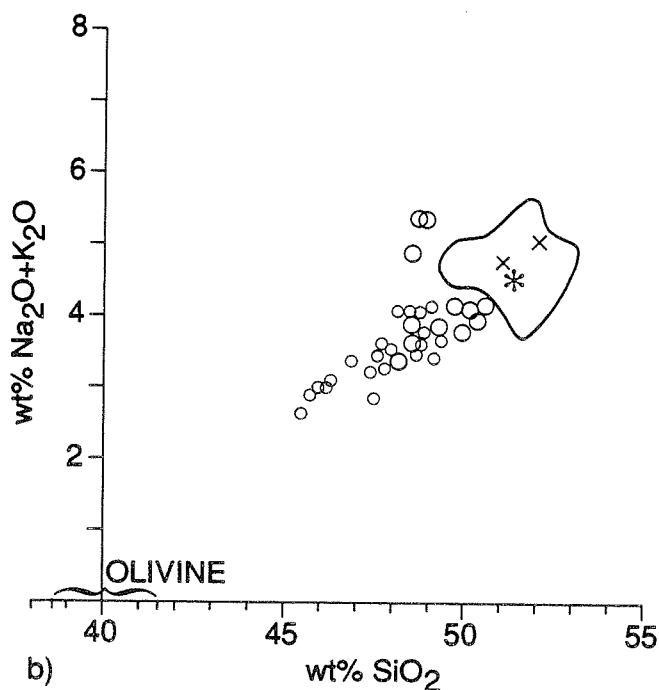
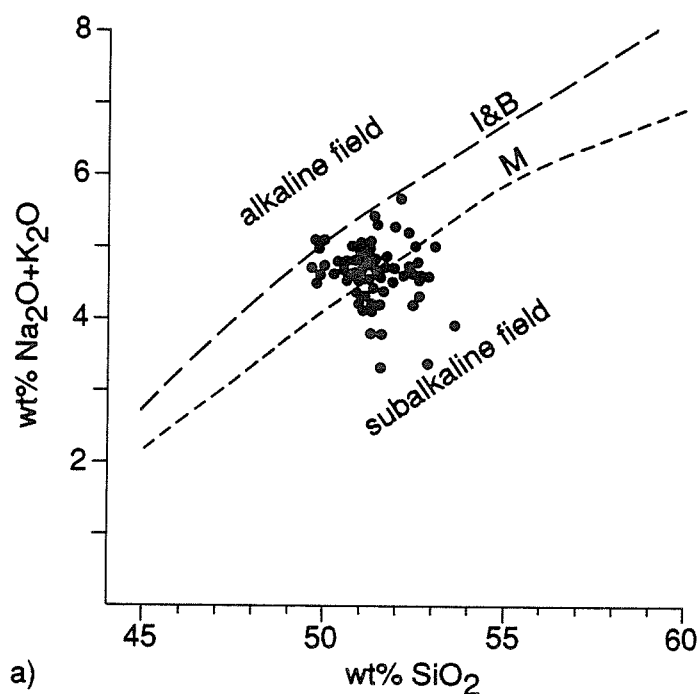


Figure 3.4.14 Alkali - silica plot for volcanics in core 7316/03-U-01.

- a) All glass analyses are shown. Dividing lines between subalkaline and alkaline fields are after Miyashiro (1978) (M) and Irvine & Baragar (1971) (I&B). The samples show a subalkaline to transitional alkaline appearance.
- b) Comparison of the main Bjørnøya West basaltic compositions (encircled area) from this study with Cenozoic basalts from Spitsbergen using data from Hagen (1988) (small circles) and from Prestvik (1978) and unpublished (large circles). This study: Crosses = whole rock analyses, * = mean value of light brown glasses.

In order to determine if different magma types or different phases of eruptions are present in the core, chemical compositions have been compared for the different colour types and as a function of core depth.

The mean values of the major element compositions for homogeneous light brown glasses, dark brown glasses and black oxidised glasses are fairly similar (Table 3.4.5), the main difference being that the latter group shows a larger standard deviation particularly for iron, magnesium, calcium and sodium. However, the chemical overlap with the other two groups may indicate that some of the scattering is due to minor secondary alteration within the group of oxidised clasts.

As seen in Figure 3.4.15 there is no systematic variation of the glass compositions with core depth. Combined with the relatively small compositional variation in general, this may suggest that only one magma type and one stage of igneous evolution is represented by the volcanics.

XRF whole-rock data overlap with the results from the glass analyses for all oxides except MgO. The higher MgO values of the whole-rock analyses can be explained by the presence of Fe-rich olivines which are not included in the pure glass analyses. Trace element analyses of whole rock basalt samples from two different levels (32.05 m and 78.05 m) give very similar values (Table 3.4.8).

Rb-Sr and Sm-Nd isotopic data: In general, the rubidium - strontium (Rb-Sr) and samarium - neodymium (Sm-Nd) isotopic systems have applications both for dating and for geochemical interpretations. However, due to the limited chemical variation of these volcanics, isochron dating was not possible.

Chemical separation and mass spectrometry analyses for calculation of the isotopic ratios $^{87}\text{Rb}/^{86}\text{Sr}$, $^{87}\text{Sr}/^{86}\text{Sr}$, $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ have been performed at the Geological Museum, University of Oslo using standard techniques. Concentration values of the trace elements Rb, Sr, Sm and Nd are derived using the isotopic dilution method.

The massive basalt near the top of the core (32.05 m) and a fraction of 1-2 mm sized homogeneous basaltic glasses (49.57 m) have been selected for comparative studies. The millimetre sized fragments were separated by careful hand-picking under the microscope to get the most optimal glasses.

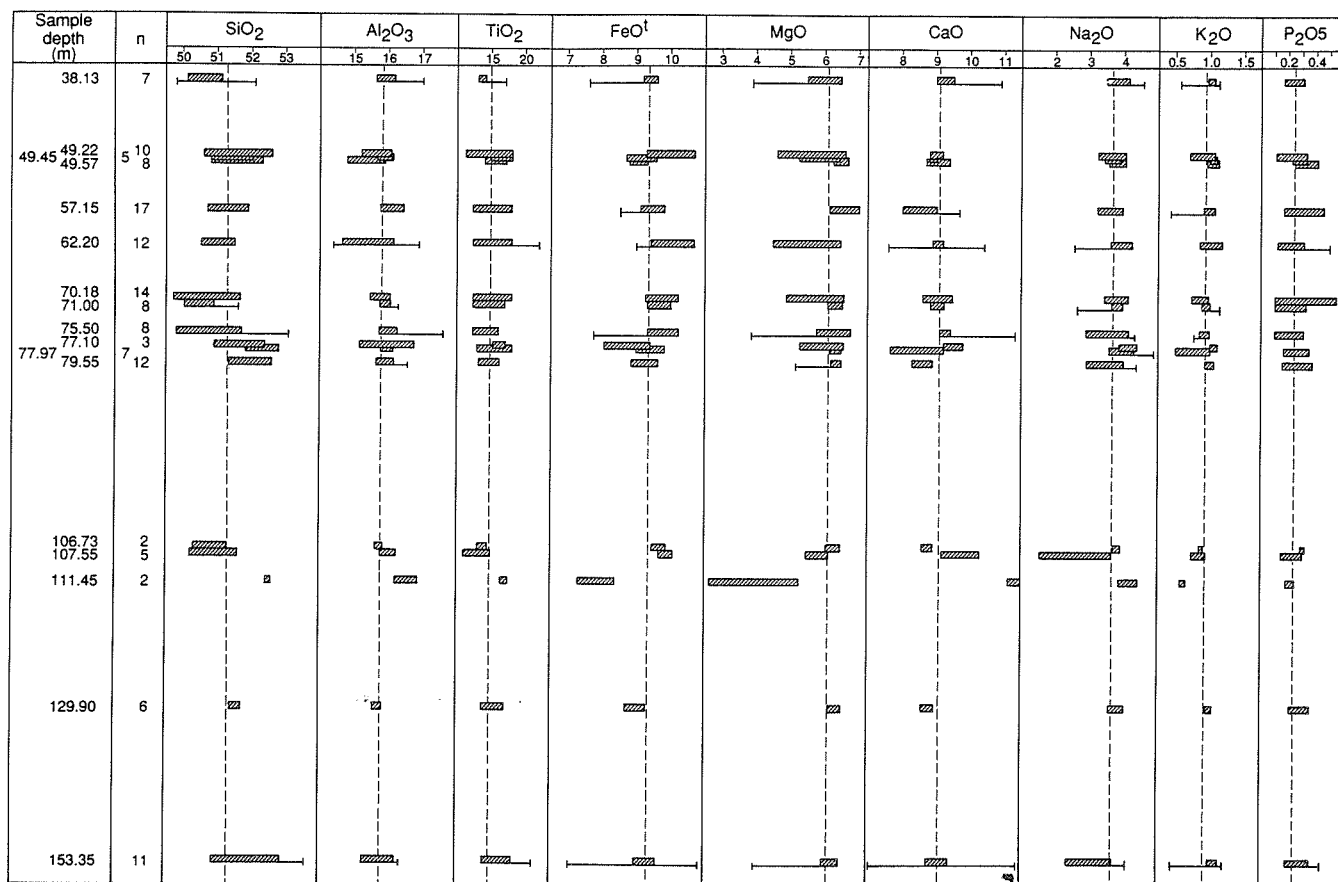


Figure 3.4.15 Range of glass compositions (oxide %, electron microprobe analyses) for each sample arranged after sample depth in core 7316/03-U-01. Broad lines give compositional ranges for the homogeneous light brown and dark brown glasses, while the thin lines also include the black oxidised/alters clasts. *n* gives the number of analyses. Dashed vertical lines refer to the mean composition of the light brown glasses.

Table 3.4.6 Sm-Nd and Rb-Sr isotopic analyses of volcanics from core 7316/03-U-01. Sample 32.05 m represents a lava block while sample 49.57 m represents a hand-picked separate of homogeneous glass fragments.

Sample	ppm Sm	ppm Nd	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd
32.05 m	4.54	20.3	0.1359	0.512325 ± 0.000006
49.57 m	4.71	21.0	0.1366	0.512324 ± 0.000012
Sample	ppm Rb	ppm Sr	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr
32.05	22.8	337	0.1957	0.705306 ± 0.000008
49.57	25.0	352	0.2052	0.705288 ± 0.000006

The isotopic and concentration data are identical for the two samples (Table 3.4.6). This supports the interpretation that the massive basalt and the small glass shards have a similar origin and that they are derived from the same basaltic magma.

The $^{143}\text{Nd}/^{144}\text{Nd}$ value is low compared with mid ocean ridge basalts and more similar to the most enriched continental basalts. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio overlaps with the enriched examples of basalt from ocean islands and continental basalt. The isotopic data show that the basaltic magma differs from the majority of mid-ocean ridge basalts, ocean island basalts and ocean floor basalts that formed by melting of a depleted mantle. The samples fall below the established mantle array defined by volcanic rocks in the $^{143}\text{Nd}/^{144}\text{Nd}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ diagram (Figure 3.4.16 a, b).

A large negative epsilon value for Nd (-6) and a positive epsilon value for Sr (+12) suggest that these volcanics have either formed by extensive mixing of a crustal component into the basaltic magma, or alternatively that the magma formed from an anomalously enriched mantle. Comparisons with mixing lines calculated elsewhere indicate that the isotopic composition can be explained by mixing of "mantle array basalt" with granulitic crustal components rather than shallow granitic crust (Faure 1986, Figure 3.4.16b). This would mean that the contamination has occurred at depths corresponding to the lower part of a continental type crust. The compositionally homogeneous nature of the glasses also supports that the mixing may have occurred at some depth in the crust prior to the eruption event.

Isotopic dating of volcanics

K-Ar dates: The K-Ar method has been widely used to date young basaltic rocks. The method gives reliable cooling ages for unaltered basalt and glass assuming that no argon was present in the rock originally and that the radiogenic argon has been retained in the sample. Secondary alteration and cements (devitrified glass, calcite, zeolite, clay minerals) would result in a lowering of K-Ar dates. On the other hand, xenoliths and xenocrysts are likely to contain excess radiogenic ^{40}Ar whose presence in a sample will increase the measured K-Ar date (Faure 1986).

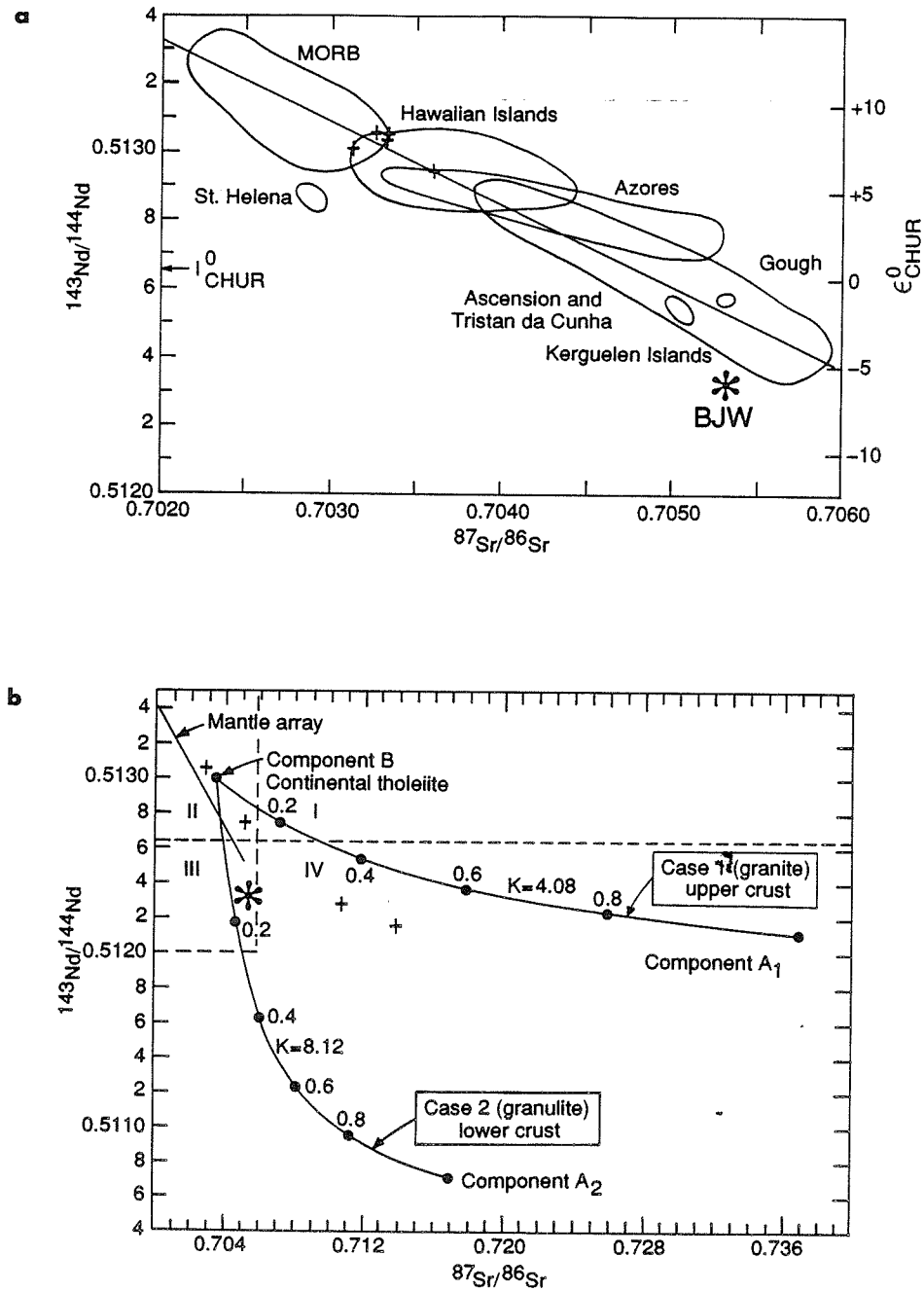


Figure 3.4.16

- a) $^{143}\text{Nd}/^{144}\text{Nd}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic composition of basaltic samples from Bjørnøya West (BJW; asterisk) compared with data from mid ocean ridge basalts (MORB) and selected oceanic islands as compiled by Faure (1986). Early Tertiary tholeiites (series 2) from the Vøring Plateau (Taylor & Morton 1989) are shown by crosses for comparison. Note the negative value of the BJW samples on the right scale giving the E-normalization to chondritic uniform reservoir (CHUR) at the present time ($t=0$).
- b) Mixing hyperbolae formed by combining rocks from continental crust with continental tholeiite basalt (numbers give proportions) from Faure (1986) using two different types of crustal material. The present (BJW) samples (asterisk) are close to the mixing curve with granulite (lower crust) material. Another mixing trend given by crosses represents contaminated andesitic series 1 of Early Tertiary material from the Vøring Plateau (Taylor & Morton 1989). Inset in the upper left part of the diagrams shows the position of the diagram above (a).

Five samples of volcanics were sent for K-Ar dating at the Oregon State University, USA, in collaboration with Professor Robert A. Duncan. The samples were selected to represent the material best suitable for K-Ar dating using preferentially the largest clasts and clasts from different stratigraphic levels (Table 3.4.7).

Table 3.4.7 *K-Ar dates and sample information.
The analyses have been performed at the Oregon State University, College of Oceanography.*

Sample	Type of material	Weight	%K	% radiogenic ^{40}Ar	Age (Ma)
7316/03-U-01					
32.05 m	Lava block	~104 g	0.910	33.2	7.35 ± 0.15
32.10 m	Lava block	~ 40 g	0.906	34.0	3.19 ± 0.05
77.70 m	4 cm lapilli	~ 35 g	0.909	39.0	2.34 ± 0.08
78.05 m	Bomb/block	~ 75 g	0.943	45.9	10.20 ± 0.15
111.45 m	3 cm lapilli	~ 23 g	0.899	24.4	4.39 ± 0.10

7

The K-Ar dates give a spread from 2.34 ± 0.08 Ma to 10.20 ± 0.15 Ma indicating Miocene - Pliocene ages for the volcanics. There is no systematic variation between the ages and the sample depths. The large spread in the data is questionable when considering the homogeneous appearance of the volcanics chemically and isotopically. Parallel samples from one piece of massive basaltic rock (at 32.05 m and 32.10-32.25 m) showed ages of 7.35 ± 0.15 and 3.19 ± 0.05 Ma, respectively. A possible explanation is that the youngest age has been artificially produced by contamination with cements. The vesicular lapilli-clasts gave young and intermediate dates, while the massive bomb at 78.05 m gave the oldest date of 10.20 ± 0.15 Ma. Altogether, this may indicate that vesicular samples with the largest possibility of cement inclusions yield too young ages, and that the volcanics have a Miocene age. A Late Miocene age is also supported by fossil evidence. The present results should be tested by other methods. One possibility is to use the ^{40}Ar - ^{39}Ar technique which can be used to see through the effect of excess argon. Such analyses will be attempted in the near future.

Fission track method

The possibility to date the volcanics by apatite and zircon fission track analyses was examined. However, inspection of heavy mineral separates

showed the absence of euhedral zircons typically found in volcanic rocks. Apatite fission track data carried out at Geotrack International also concluded with absence of volcanic apatite (Appendix A.8). A third possibility, to study fission tracks in homogeneous glass was not recommended by Geotrack International. The negative results from these efforts are, however, important for discussion of the volcanism and sediments in general.

Discussion of origin of the volcanics

Eruption type: The glass textures and vesicularity are typical for basaltic tephra, resembling phreatomagmatic eruptions from Iceland. Tephra is the result of explosive eruptions. The explosivity may record boiling of the magma during ascent to the surface due to the reduced solubility of volatiles in the magma during the pressure decrease. Another mechanism is boiling caused by interaction of hot magma and water at shallow water depths (e.g. groundwater, shallow marine water, lakes) during eruption. This latter mechanism may cause explosive phases during eruption even in tholeiitic lavas with low volatile contents, as documented in Iceland.

The volcanics in this study are redeposited in the present setting, but the shapes and excellent preservation of the clasts indicate very little alteration by depositional and diagenetic processes. Some solid blocks with detritus may have been cemented (opal) before redeposition.

The intermediate sized lapilli fragments are highly vesicular, while the smaller are more massive and equant. The vesicular clasts are typical for air falls (lapilli, bombs, blocks or parts of lava fountains). The more solid and smallest clasts may also resemble the products of rapid quenching such as in contact with water or ice. From the general appearance of the volcanics and the sediment inclusions, and comparisons with recent eruptions from Iceland, the model of explosive eruption through water (wet sediments?) is preferred. Such a mechanism may explain the occurrence of sediment xenocrysts in glass and the vesicularity of the main clasts. Transition from submarine to subaerial volcanism has been documented in the literature (Fisher & Schmincke 1984) and has also been observed in recent volcanism in Iceland.

Petrogenesis: The chemical composition of the volcanics is different from tholeiitic basalt formed at active spreading ridges, indicating an off ridge icelandic or continental (?shelf) environment. As suggested above,

the magma may have assimilated material from continental crust prior to the eruption event. This indicates presence of continental crust in the area of the eruption site.

Extensive formation of olivine phenocrysts occurred at depths within the magma chamber. Chrome spinel formed simultaneously. From the olivine growth features this may reflect undercooling within the magma chamber. It is possible that this is related to extensive assimilation of deep crustal material into the magma followed by homogenization of the contaminated melt.

The compositional similarity of all the volcanics in the core suggests that they have a common origin. The eruption was triggered by ascent of magma to shallower levels after the olivine crystallization. Microphenocrysts, mainly of plagioclase, formed late during the eruption of lava flows or larger clasts. Near to or at the surface, the magma erupted through and was physically mixed with unconsolidated sediments. A possible model is that the volcano was built up through a shallow Miocene sedimentary basin or at the edge of it. The unstable mixture of sediments and pyroclastic products was subjected to rapid erosion, with redeposition into deeper water environments. The sediments could partly have been transported by submarine land-slides and debris flows that may have been triggered by earthquake activity following the volcanism.

Comparison with North Atlantic Cenozoic Volcanism

The North Atlantic borderlands (north, east and west Greenland, the Faeroe Islands, Rockall Plateau and Rockall Trough, Vøring Plateau and Faeroe-Shetland Escarpment, Britain and Ireland) were the sites of extensive volcanism in Early Tertiary with a climax around 60 to 50 Ma (cf. Upton 1988). This volcanism was related to the evolution from rifting of the continental Laurasian lithosphere to sea-floor spreading and generation of oceanic crust.

The majority of the North Atlantic basalts are tholeiites (Fe-Ti rich basalt and depleted mid-ocean ridge basalts). In some of the borderland areas, igneous activity also continued into Oligocene time with a tendency of more alkaline or silicic compositions (Upton 1988). The youngest lavas known from North Greenland are Miocene (?10-12 Ma) alkaline - peralkaline basalts (A.K. Pedersen pers. comm. 1990). Late Tertiary and Quaternary volcanism involves Iceland, Jan Mayen, North Spitsbergen and various mid-

ocean ridge segments, that nearest to Bjørnøya West being the Mohn - Knipovich ridge.

Late Tertiary ashes from Icelandic eruptions are widespread, and the low-K Icelandic ash deposits have been found on the Vøring plateau indicating extensive volcanism around 8.0 to 8.5 Ma (Goll 1989). Cenozoic basalts from North Spitsbergen give K-Ar dates of 10.4 ± 1.1 Ma and 11.5 ± 1.2 Ma (Prestvik 1978), while more recent work gives K-Ar ages of 2.5 to 4.5 Ma and an Ar-Ar age of 6.8 Ma (Skjelkvåle 1987). The Late Tertiary basalts from Spitsbergen have been classified as olivine tholeiite (Hagen 1988), some being characterized as transitional to alkaline (Prestvik 1978). This basalt volcanism was followed by Quaternary volcanism that erupted highly alkaline magmas (nepheline basanite to hawaiite; Skjelkvåle 1987, Skjelkvåle et al. 1988).

Recent basalt volcanism in Iceland covers wide compositional ranges reflecting different tectonic regimes, different degrees of interaction with older crust, and different melt forming histories (cf. Oskarsson et al. 1982). Thus both low-K tholeiites and alkaline basalt compositions are produced. Jan Mayen consists of potassic alkaline volcanics younger than 0.7 Ma (Imsland 1978, 1986). From the Jan Mayen Platform there is a transition from alkaline towards more tholeiitic composition along the Mohn's Ridge in the direction towards Bjørnøya (Neumann & Schilling 1984).

While the olivine basaltic rocks from Bjørnøya West (7316/03-U-01) are subalkaline to transitional alkaline with high contents of both K_2O (ca 1%) and Na_2O (3-4%) and SiO_2 (close to 51%), and with enriched Nd and Sr isotopic signatures, the majority of the Early Tertiary basalts, including both continental flood basalts and basalts produced during advanced stages of rifting and seafloor spreading are characterized by low-K tholeiitic compositions. The basaltic compositions from Bjørnøya West thus differ both from the common continental tholeiites and the mid ocean ridge basalts by their high K (and Na) contents. They also differ compositionally from the more strongly alkaline rocks that are related to episodes with rejuvenation of volcanism in off-rift settings.

The compositions that are closest to the Bjørnøya West volcanics are the Late Tertiary basalts from North Spitsbergen (south of Woodfjorden), which are also close in age. In the alkali - silica plot (Figure 3.4.14) the Spitsbergen basalts define a trend from compositions with lower silica towards the relatively silica-rich Bjørnøya West basalts. The $^{143}Nd/^{144}Nd$

and $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic compositions of the latter indicate that the magma could have been modified by interaction with deep continental crust (granulite). In the basalt area on Spitsbergen, xenoliths of granulite have been brought up by the young alkaline volcanism (Amundsen et al. 1987), indicating the presence of granulite bearing crust underneath the Late Tertiary volcanic province on Spitsbergen. Thus, comparisons with the Late Tertiary Spitsbergen basalts may form a starting point for further modelling of the evolution of the present basalts.

Geological implications

The present location of the Bjørnøya West volcanics at 7316/03-U-01 is on the continental margin to the west of the Stappen High, and not far from the inferred Early Tertiary Vestbakken volcanic province defined to the west on the basis of seismic data (the Bjørnøya Marginal High of Eldholm et al. 1987 and Faleide et al. 1988). The main volcanic events in that area are regarded as Eocene basalts related to the initial opening of the southern Greenland Sea (Faleide et al. 1988). Volcanic peaks are also identified on the northwestern flank of the Vestbakken volcanic province. Faleide et al. (1988) related them to a rejuvenation of volcanic activity resulting from a major change in plate motions in Oligocene time. The present study suggests them to be coeval with the Miocene volcanism giving the volcanics in core 7316/03-U-01.

The present study has recognized the presence of volcanism in Late Miocene (? Early Pliocene) times to the east of the older volcanic province. The geochemistry of the volcanics combined with sedimentology and seismic interpretation suggest that the volcanism was related to a continental type crust.

Table 3.4.8a *Electron microprobe analyses of glasses from core 7316/03-U- 01 arranged after sample depth (*=black oxidized (?) clasts).*

Sample	Point	SiO2	Al2O3	TiO2	FeO	MnO	MgO	CaO	Na2O	K2O	P2O5	Sum
38.13	1.10	50.42	15.71	1.34	9.17	0.14	6.35	9.19	3.61	1.00	0.16	97.14
38.13	2.20	51.08	15.84	1.35	9.36	0.13	5.53	8.92	3.53	1.03	0.29	97.12
38.13	3.20	49.83	16.90	1.67	8.54	0.15	5.20	10.92	3.98	0.56	0.25	98.02*
38.13	4.10	50.12	16.00	1.41	9.64	0.17	5.98	9.45	4.05	1.00	0.30	98.17
38.13	5.20	51.04	16.06	1.44	9.58	0.09	6.47	9.05	4.02	1.03	0.29	99.12
38.13	6.10	50.33	15.83	1.43	9.54	0.22	6.34	9.32	3.66	0.96	0.24	97.90
38.13	6.20	52.14	17.04	1.59	7.62	0.07	3.94	9.67	4.56	1.10	0.21	97.97*
49.22	1.10	50.93	16.03	1.45	9.61	0.13	6.29	8.81	3.57	0.82	0.32	98.00
49.22	3.10	50.98	15.86	1.31	9.43	0.15	6.26	9.19	3.67	0.87	0.21	97.97
49.22	3.20	50.96	15.88	1.06	9.66	0.13	6.41	8.88	3.52	0.70	0.20	97.44
49.22	3.30	50.97	15.70	1.15	9.91	0.14	6.72	9.13	3.46	0.80	0.22	98.23
49.22	3.50	52.58	15.21	1.76	10.71	0.22	4.56	8.78	3.24	1.06	0.26	98.43
49.22	4.10	51.04	15.83	1.40	9.68	0.15	6.32	8.90	4.04	0.78	0.23	98.43
49.22	5.10	50.93	16.11	1.17	9.47	0.12	6.35	9.02	3.77	0.85	0.10	97.94
49.22	6.10	50.73	15.86	1.19	9.25	0.07	6.28	9.07	3.77	0.85	0.26	97.36
49.22	6.10	50.73	15.86	1.19	9.25	0.07	6.28	9.07	3.77	0.85	0.26	97.36
49.22	7.20	51.33	15.89	1.49	10.04	0.11	6.51	8.99	3.99	0.77	0.14	99.32
49.45	1.10	51.70	15.66	1.49	8.92	0.13	6.37	8.82	3.44	1.00	0.30	97.87
49.45	1.20	51.97	15.97	1.61	8.72	0.10	6.42	8.80	3.72	1.01	0.22	98.59
49.45	2.10	50.84	16.05	1.77	9.63	0.15	5.69	8.91	3.90	1.09	0.27	98.36
49.45	2.70	50.92	15.82	1.53	9.50	0.19	5.32	8.87	3.73	1.03	0.28	97.24
49.45	4.10	51.34	16.10	1.69	9.28	0.13	5.71	8.95	3.88	1.06	0.32	98.50
49.57	3.00	50.83	15.66	1.67	8.79	0.13	6.59	8.95	3.79	1.00	0.40	97.87
49.57	3.00	51.55	15.64	1.51	9.28	0.19	6.19	9.01	3.59	1.01	0.22	98.22
49.57	4.00	51.16	14.80	1.51	9.29	0.20	6.26	9.40	3.91	1.04	0.22	97.86
49.57	7.00	51.49	15.68	1.46	8.97	0.10	6.62	9.06	3.78	0.96	0.24	98.40
49.57	9.00	51.43	15.55	1.57	9.10	0.22	6.70	8.70	3.95	1.11	0.22	98.58
49.57	20.00	50.95	15.90	1.39	9.25	0.15	6.17	9.13	3.67	1.06	0.32	98.05
49.57	21.00	52.25	15.92	1.56	9.17	0.16	6.27	9.30	3.52	1.06	0.20	99.45
49.57	25.00	51.76	15.79	1.61	9.20	0.10	6.21	8.92	3.77	1.06	0.27	98.74
57.15	1.10	51.82	16.01	1.70	9.75	0.07	6.45	8.60	3.82	1.02	0.18	99.48
57.15	1.20	51.36	15.77	1.27	9.08	0.05	6.36	8.92	3.15	1.09	0.19	97.30
57.15	1.30	51.73	16.01	1.20	9.40	0.13	6.47	8.83	3.61	1.02	0.22	98.67
57.15	1.40	50.77	15.84	1.42	9.38	0.13	6.66	8.74	3.30	1.06	0.30	97.64
57.15	1.60	50.66	16.32	1.59	9.77	0.22	6.14	8.62	3.48	0.98	0.45	98.27
57.15	1.70	50.99	16.09	1.59	9.48	0.10	6.79	8.47	3.50	0.93	0.26	98.24
57.15	1.80	50.83	16.18	1.51	9.49	0.17	6.42	8.79	3.67	1.05	0.29	98.46
57.15	2.10	51.90	15.88	1.55	9.58	0.14	6.91	8.74	3.51	0.98	0.25	99.47
57.15	3.10	50.75	16.08	1.51	9.28	0.09	6.70	8.55	3.87	0.99	0.26	98.13
57.15	5.10	51.82	16.24	1.76	9.21	0.01	6.52	8.66	3.70	0.99	0.16	99.09
57.15	6.10	50.96	15.90	1.51	9.17	0.19	6.68	8.58	3.68	1.03	0.31	98.05
57.15	7.10	51.25	16.49	1.34	8.51	0.15	6.83	9.67	3.64	0.41	0.22	98.56*
57.15	8.10	51.16	15.79	1.21	9.46	0.16	6.79	8.62	3.63	0.89	0.28	98.04
57.15	8.20	51.13	16.23	1.46	9.53	0.12	6.74	8.56	3.57	0.95	0.32	98.66
57.15	9.10	51.12	15.98	1.54	9.14	0.23	6.95	8.58	3.68	1.04	0.33	98.63
57.15	10.20	51.56	15.80	1.57	9.41	0.16	6.36	8.29	3.29	0.92	0.34	97.75
57.15	11.10	51.06	16.37	1.36	9.59	0.13	6.98	7.94	3.49	0.98	0.27	98.20

Table 3.4.8a cont.

65.20	1.40	51.04	15.71	1.38	9.58	0.14	6.22	9.24	3.80	0.97	0.20	98.31
65.20	2.10	51.30	16.02	1.28	9.43	0.17	6.19	9.22	3.80	0.87	0.23	98.56
65.20	3.10	51.24	16.85	1.54	9.04	0.20	4.28	10.28	3.93	1.08	0.28	98.77*
65.20	4.10	51.52	16.05	1.55	9.22	0.20	6.31	9.02	3.79	0.88	0.14	98.73
65.20	4.20	51.49	14.92	1.76	9.03	0.12	5.70	10.36	3.47	0.89	0.37	98.16*
65.20	5.10	50.93	15.95	1.38	9.72	0.17	6.54	9.13	3.68	0.87	0.28	98.70
65.20	6.20	50.45	15.65	1.31	10.10	0.16	6.50	8.93	3.86	0.87	0.30	98.18
65.20	6.30	50.88	15.89	1.36	9.59	0.11	6.36	9.08	3.71	0.83	0.10	97.95
65.20	7.10	51.50	14.61	1.64	10.66	0.09	4.47	8.90	4.19	1.17	0.30	97.56
65.20	7.30	50.66	15.47	1.66	9.65	0.14	6.48	8.96	3.71	0.81	0.19	97.78
65.20	8.20	50.75	15.81	1.26	9.47	0.12	6.05	9.20	3.62	0.90	0.31	97.53
70.18	0.00	49.71	15.96	1.29	9.20	0.15	6.16	9.46	3.69	0.97	0.54	97.17
70.18	0.00	50.22	15.66	1.78	9.46	0.10	6.28	9.35	3.73	0.87	0.15	97.64
70.18	1.10	51.05	15.61	1.28	9.83	0.11	6.31	8.83	3.77	0.76	0.20	97.74
70.18	1.20	50.76	15.52	1.41	9.61	0.07	6.60	8.80	3.89	0.73	0.24	97.67
70.18	1.30	51.25	15.50	1.15	9.58	0.16	6.32	8.75	3.96	0.74	0.08	97.52
70.18	2.10	49.87	16.03	1.45	9.61	0.11	6.34	9.14	4.06	0.91	0.21	97.76
70.18	3.10	51.11	15.78	1.38	10.22	0.15	5.63	9.27	3.31	0.78	0.23	97.89
70.18	3.20	51.66	15.51	1.30	9.63	0.17	6.26	8.58	3.87	0.84	0.22	98.09
70.18	4.10	51.36	15.65	1.43	10.04	0.06	4.93	8.82	3.80	0.86	0.13	97.16
70.18	6.10	50.87	15.92	1.52	9.91	0.08	6.36	8.76	3.75	0.74	0.18	98.14
70.18	7.10	50.86	15.90	1.33	9.71	0.14	6.44	8.79	3.86	0.71	0.32	98.10
70.18	7.20	51.13	15.77	1.53	9.27	0.21	5.87	9.45	3.75	0.95	0.23	98.22
70.18	8.10	50.98	15.91	1.47	9.86	0.16	6.23	9.32	3.71	0.97	0.18	98.86
70.18	8.20	50.63	15.77	1.49	9.42	0.16	6.14	9.40	3.92	0.92	0.27	98.17
71.00	5.10	50.04	15.82	1.68	9.46	0.19	6.25	9.15	3.79	0.88	0.21	97.52
71.00	5.20	50.51	15.92	1.63	9.86	0.12	6.36	9.09	3.76	1.02	0.17	98.44
71.00	6.10	50.40	15.93	1.38	9.80	0.14	6.46	9.01	3.92	0.90	0.15	98.15
71.00	7.10	50.86	16.03	1.39	9.63	0.07	6.12	9.07	3.57	0.88	0.09	97.76
71.00	8.10	50.93	15.84	1.15	9.27	0.07	6.18	8.85	3.82	0.84	0.32	97.31
71.00	9.10	50.71	15.82	1.47	9.77	0.14	6.15	8.75	3.76	0.80	0.23	97.65
71.00	9.20	50.49	15.77	1.37	9.38	0.18	6.31	8.82	3.81	0.89	0.18	97.25
71.00	9.30	51.59	16.31	1.26	9.74	0.23	6.36	8.93	2.63	1.14	0.20	98.44*
75.50	1.10	50.86	15.99	1.41	9.34	0.10	6.75	9.16	3.95	0.84	0.07	98.52
75.50	2.10	51.31	15.88	1.17	10.19	0.13	6.16	9.37	2.79	0.97	0.19	98.21
75.50	3.10	51.65	16.18	1.43	9.45	0.08	5.82	9.23	3.68	0.98	0.13	98.64
75.50	4.10	53.07	17.55	1.22	7.63	0.12	3.92	11.25	4.18	0.77	0.28	100.03*
75.50	5.10	51.10	15.89	1.60	9.87	0.13	6.28	9.26	3.83	0.91	0.30	99.21
75.50	7.10	49.83	15.78	1.49	9.96	0.16	6.28	9.26	4.12	0.93	0.23	98.09
75.50	7.20	51.25	15.82	1.51	9.40	0.16	6.14	9.18	3.91	0.99	0.20	98.61
75.50	9.10	51.01	15.99	1.24	9.61	0.13	6.25	9.05	3.91	0.92	0.28	98.44
77.10	3.00	50.85	15.15	1.55	9.41	0.12	6.61	9.22	3.84	1.01	0.26	98.06
77.10	10.00	51.36	16.02	1.71	9.26	0.11	6.34	9.24	3.85	1.10	0.17	99.20
77.10	12.00	52.35	16.81	1.53	7.99	0.14	5.28	9.75	4.26	0.98	0.27	99.36
77.97	1.10	51.83	16.03	1.33	9.08	0.11	6.46	8.94	3.86	0.75	0.20	98.64
77.97	2.20	52.50	15.93	1.78	9.24	0.16	6.21	7.64	4.00	0.96	0.22	98.68
77.97	3.10	52.59	15.94	1.44	9.68	0.10	6.37	8.58	4.21	0.58	0.15	99.69
77.97	4.10	52.90	15.90	1.46	9.79	0.12	6.40	9.24	4.08	0.49	0.25	100.67
77.97	4.20	52.63	16.09	1.80	9.19	0.27	6.54	8.99	3.63	0.89	0.34	100.42
77.97	5.10	52.41	16.02	1.61	8.96	0.11	6.31	8.83	3.83	0.89	0.24	99.26
77.97	6.00	52.82	15.85	1.74	8.86	0.20	6.37	9.30	3.54	1.01	0.22	99.95

79.55	1.10	51.67	15.99	1.42	8.82	0.16	6.36	8.68	3.63	0.99	0.23	98.00
79.55	1.20	51.68	15.76	1.29	8.89	0.03	6.48	8.63	3.72	1.02	0.36	97.91
79.55	1.30	51.54	16.09	1.33	8.85	0.12	6.43	8.74	2.79	1.05	0.28	97.26
79.55	1.40	51.98	16.45	1.53	9.58	0.09	5.15	8.21	4.26	0.99	0.25	98.53*
79.55	2.10	52.41	15.94	1.52	9.12	0.19	6.30	8.77	3.74	0.99	0.13	99.16
79.55	2.20	52.25	15.78	1.60	9.10	0.17	6.24	8.73	3.65	0.98	0.28	98.83
79.55	2.40	52.55	16.07	1.56	9.61	0.13	6.36	8.92	3.29	1.01	0.23	99.76
79.55	3.10	51.31	15.83	1.39	8.76	0.21	6.39	8.42	3.85	0.95	0.33	97.48
79.55	4.10	51.27	15.66	1.36	8.88	0.10	6.29	8.65	3.69	0.99	0.19	97.13
79.55	4.20	51.48	15.87	1.77	9.00	0.16	6.36	8.61	3.72	0.92	0.29	98.23
79.55	5.10	52.42	15.75	1.27	8.84	0.21	6.18	8.39	3.55	1.01	0.29	97.95
79.55	6.20	52.55	15.89	1.55	8.95	0.09	6.52	8.28	3.85	0.95	0.22	98.91
106.73	1.10	51.30	15.73	1.41	9.81	0.13	6.03	8.91	3.57	0.88	0.29	98.11
106.73	2.10	50.33	15.81	1.31	9.38	0.07	6.46	8.63	3.71	0.86	0.31	96.94
107.55	1.10	51.34	16.00	1.14	9.62	0.09	5.67	9.78	2.00	0.83	0.20	96.72
107.55	1.20	51.23	15.84	1.39	9.77	0.17	5.89	9.35	3.57	0.89	0.22	98.36
107.55	2.20	51.63	16.04	1.46	9.99	0.12	5.51	9.61	2.40	0.94	0.15	97.89
107.55	2.30	50.79	16.18	1.24	9.56	0.23	5.80	10.32	1.40	0.81	0.30	96.68
107.55	4.20	50.19	15.88	1.24	9.88	0.14	6.05	9.16	3.25	0.73	0.24	96.80
111.45	8.00	52.47	16.20	1.68	7.21	0.10	5.33	11.53	3.65	0.57	0.19	98.96
111.45	15.00	52.46	16.89	1.71	8.31	0.06	2.73	11.08	4.34	0.64	0.17	98.42
129.90	1.10	51.44	15.54	1.44	9.12	0.12	6.24	8.80	3.57	1.04	0.28	97.64
129.90	1.30	51.69	15.82	1.48	8.63	0.10	6.48	8.63	3.41	0.97	0.28	97.56
129.90	2.30	51.62	15.62	1.56	8.66	0.10	6.21	8.79	3.65	0.94	0.29	97.49
129.90	3.10	51.40	15.65	1.64	8.83	0.09	6.42	8.89	3.63	0.98	0.21	97.78
129.90	4.10	51.47	15.45	1.44	8.73	0.14	6.38	8.75	3.80	0.92	0.36	97.49
129.90	5.10	51.52	15.56	1.70	9.15	0.15	6.45	8.73	3.54	1.00	0.32	98.17
153.35	1.10	52.85	16.23	1.47	9.60	0.10	6.53	8.85	2.38	0.99	0.28	99.31
153.35	2.10	51.37	15.75	1.46	10.82	0.20	6.04	7.01	4.10	1.21	0.23	98.22*
153.35	3.20	51.95	15.78	1.78	9.27	0.10	6.37	9.13	3.56	1.07	0.27	99.32
153.35	3.30	52.00	15.38	1.41	9.26	0.20	6.01	9.27	3.35	1.09	0.34	98.35
153.35	4.10	53.61	16.43	1.53	6.85	0.09	6.54	11.38	3.44	0.46	0.32	100.70*
153.35	5.10	50.94	15.40	1.57	8.98	0.12	6.03	9.25	3.70	1.15	0.36	97.56
153.35	6.10	51.44	15.42	1.45	9.42	0.11	5.98	9.38	3.70	1.07	0.23	98.26
153.35	6.20	51.75	15.27	1.47	9.37	0.19	6.31	9.10	3.54	1.01	0.32	98.34
153.35	6.30	51.48	15.70	1.58	9.14	0.10	6.03	8.82	3.68	0.99	0.18	97.76
153.35	7.20	52.02	15.59	1.43	7.90	0.17	5.73	10.27	3.60	0.88	0.31	97.95*
153.35	8.00	52.70	16.17	2.05	8.19	0.15	4.01	10.62	3.40	1.20	0.44	98.96*

Table 3.4.8b XRF whole-rock trace-element analyses (ppm) for two basaltic samples from core 7316/03-U-01.

SAMPLE	Ni	Cr	Sc	V	Ba	Rb	Sr	Zr	Y	Nb	Ga	Cu	Zn	Pb	La	Ce	Th
32.05	252	314	25	189	352	23	340	138	24	22.5	20	81	98	3	21	53	3
78.05	275	313	26	183	375	23	369	143	24	22.8	20	89	104	3	33	50	2

3.4.5 Petrophysical logs and core measurements

Downhole logs

The gamma, density and neutron logs were run inside the HR string from TD to the seabed. The gamma log values are recalculated to open hole values based on a regression analysis from a previous shallow drilling survey. Due to very low variance in the gamma log values and because gamma laboratory measurements could be measured only below 111 m, it was impossible to calculate any depth shift correction (see Appendix A.10). The small peaks on the density log with 3.05 m spacing are caused by joints in the HR string. Some of the log responses measured through the HR string may be due to wellbore effects like cavings. The logs are shown on Figure 3.4.1. The gamma log and petrophysical properties are also shown together in the summary sheet in Appendix A.1.4.

192.7 (TD) - 153.8 m: Subunit A

The glauconitic fine-grained sandstones show mainly gamma values just below 100 API with a decreasing trend above 158 m. The relatively high gamma values are due to K-feldspar, mica and glauconite. The density log shows increasing number and size of negative peaks upward probably due to cavings. The neutron porosity values show a slightly upward decreasing trend which also may be due to cavings.

153.8 - 93.5 m: Subunit B

In this subunit which consists of glauconitic fine-grained sandstone with some volcanic material, the gamma log drops to 40 API at 147 m and then increases to a stable level around 80 API from 141 m and upward. The relatively high gamma values are due to clay minerals, K-feldspar, mica and glauconite. The density log shows a similar trend as the gamma log with a negative peak at 147 m, increasing to a higher stable level above 140 m. The negative peak is probably due to cavings. The neutron log shows a similar trend with a drop at 147 m and a higher stable level above 140 m.

93.5 - 79.6 m: Subunit C

The gamma log shows a very small break into this more shaly subunit with a slightly increasing API level and a somewhat more spiky appearance. The density log values drop gradually to a slightly lower level and the neutron log values show a significant drop just above the boundary between subunits B and C at 93.5 m.

79.6 - 75.4 m: Subunit D

There is no major break in the gamma log into this subunit which consists of volcanic debris in sandstone, but at 77 m the API value drops to about 50 API and increases a little upward to 75.4 m. The density log shows a slightly increasing trend in this subunit. The neutron log is low at 79 m and increases at 77 m.

75.4 - 71.6 m: Subunit E

The upward increasing trend of the gamma log continues into this laminated silty claystone subunit with gamma values reaching 80 API. With reservation for wellbore effects this indicates a fining-upward trend which corresponds relatively well with the sedimentology log. The density log shows a decreasing trend in this subunit and the neutron log signal drops at 73.5 m.

71.6 - 31.9 m: Subunit F

In this medium grained sandstone the upward increasing trend continues to 69.3 m where the gamma log drops to a stable, but spiky, level around 60 API. The high gamma level between 71.6 m and 69.3 m is probably caused by high glauconite and K-feldspar contents. The density log shows a break at 69.3 m to a somewhat higher level which is stable up to a caving just below the API drill bit. The neutron log shows a similar break at 69.3 m up to a higher level which continues up to the caving just below the API drill bit.

Sound velocity and gamma measurements on core

Appendix A.1.4 shows the gamma wireline log together with the spectral gamma values measured on the core. The trends of the gamma laboratory measurements correlate well with the gamma log values. The K-plot on the spectral gamma laboratory measurements shows that K is the main source for

the gamma log response (Figure 3.4.3). The U and Th values vary only very little from the background level.

Two horizontal sound velocity measurements at 186 m show values around 1900 m/s.

Thermal conductivity

See Appendix A.9.

3.4.6 Paleontology

192.7 (TD) - 31.9 m: Upper Miocene

Core 7316/03-U-01 contains abundant, diverse and well preserved reworked Cretaceous (Aptian-Cenomanian ?) and Paleogene marine microfloras in the 33 examined core samples from the lithological subunits A, B, C, E and F, but very few in situ dinoflagellate cysts. No samples have been studied from Subunit D. The stratigraphic distribution of dinoflagellate cysts in core 7316/03-U-01 is shown in Appendix A.11. *Impagidinium patulum*, *Achomosphaera andalousiensis*, *Invertocysta tabulata*, *Dapsilidinium pastielsii*, *Tuberculodinium vancampoe* and *Hystrihostrogylon* sp. of Edwards (1984) are characteristic Neogene species. The co-occurrence of these species favours a Middle Miocene to Late Miocene age for the cored unit.

A few of the dinoflagellate cysts recorded in the core are also known to range into the Pliocene (i.e. *Impagidinium patulum*, *Dapsilidinium pastielsii* and *Achomosphaera andalousiensis*), but a Pliocene age cannot be confidently confirmed on the basis of the dinoflagellate cysts.

Siliceous microfossils in core 7316/03-U-01 are largely restricted to subunits C and E. Their stratigraphic distribution is shown in Figures 3.4.17 and 3.4.18. The radiolarian *Tesserastrum thiedei* gives evidence of a Late Miocene age. All other recognized siliceous microfossils are considered to be reworked.

Based on evidence from all available evidence from in situ fossils, a Late Miocene age is most probable for all subunits in this core. Subunits B and C are of Late Miocene age because of direct, age-restricted fossil

evidence. Subunits D to F are of the same age because of the subsequent occurrence of several fossils or the superposition of lithostratigraphic subunits. Subunit A yielded no unequivocal Late Miocene age evidence. This subunit is not fundamentally different from the overlying one however, and there is no reason to assume a different age.

Dinoflagellate cysts

The lowermost sample with age-diagnostic Neogene species at 168.15 m (Appendix A.11) contains *Baticasphaera sphaerica* and *Impagidinium patulum*, neither of which are observed below the Early Miocene in the ODP Leg 104 (Manum et al. 1989). These authors used the earliest appearance of the latter species to define the base of their *Impagidinium patulum* Zone on the Vøring Plateau. The base of this zone was also characterized by the earliest appearance of *Labyrinthodinium* cf. *truncatum*. In core 7316/03-U-01 *Labyrinthodinium* cf. *truncatum* is found at 141.95 m. In the Vøring Plateau both *I. patulum* and *L.* cf. *truncatum* first appear within the Early Miocene, but both species range into the Pliocene.

Cribroperidinium tenuitabulatum and *Spiniferites pseudofurcatus* are also found at 141.95 m. *C. tenuitabulatum* ranges up to 85.57 m in core 7316/03-U-01, while *S. pseudofurcatus* is found up to 56.55 m. On the Vøring Plateau both species are restricted to the Miocene, and the first species is not recorded above the Middle Miocene.

The inception of *Invertocystia tabulata* is noted at 114.55 m. The earliest appearance of this species (which is found up to 56.55 m) is generally regarded to mark the Middle Miocene. On the Vøring Plateau this species ranges into the Late Miocene, but was not found in younger deposits.

The presence of *Achomosphaera andalusiensis* at 107.30 m and 102.99 m is also of biostratigraphic significance. Manum et al. (1989) used the earliest appearance of this species to define the base of their *Achomosphaera andalusiensis* Zone on the Vøring Plateau. This zone was related to the Late Miocene, but the top of the zone was not defined. *Achomosphaera andalusiensis* is elsewhere observed in Upper Miocene to recent deposits.

Hystrihostrogylon sp. of Edwards (1984), which in core 7316/03-U-01 is found from 56.55 m to 43.07 m, previously has been reported from the Miocene (Edwards 1984).

The uppermost part of core 7316/03-U-01 contains no age-diagnostic Neogene dinoflagellate cysts. *Tuberculodinium vancampoe* found at 51.95 m is usually a good Early Miocene marker (Matsuoka & Bujak 1988, Costa & Downie 1979), but this species is also known from recent deposits (Williams & Bujak 1985). According to Wall et al. (1977) and Harland (1982) *Tuberculodinium vancampoe* is a subtropical to warm-temperate species in the North Atlantic.

Siliceous microfossils

Diatoms, radiolaria, ebridians, silicoflagellates, archaeomonads, *Geodia*, and sponge spicules are present in this core, and 41 taxa or informal morphs have been identified in 30 samples from the bedrock section of 7316/03-U-01 (Figures 3.4.17 and 3.4.18). These occurrences are largely restricted to lithologic subunits C and E.

With only two exceptions of no biostratigraphic importance, siliceous fossils are essentially absent in Subunit B. Large pyritized diatom specimens similar in gross morphology to *Coscinodiscus* sp. 1 of Bettenstaedt (1962) occur in the sample at 116.15 m, and this species is commonly regarded as a latest Paleocene-earliest Eocene marker in the North Sea (King 1983). Because these specimens lack certain diagnostic characteristics of *Coscinodiscus* sp. 1, we are reluctant to assign biostratigraphic significance to this occurrence. Small diatoms (?*Stephanopyxis turris*) are abundant in the sample at 103.05 m, but taxonomic identification is obstructed by dense manganese encrustations.

Siliceous microfossils are generally abundant in the samples from the claystones and clayey siltstones (91.45 m and 80.44 - 74.11 m). The abundance of biosiliceous skeletal debris increases upwardly in the lower claystone interval. These microfossils are particularly numerous in the size fraction <45 µm., but some radiolaria and large diatoms are present in the >45 µm fraction. Cristobalite replacements of large spumellarian radiolaria are rare to common throughout subunits C and E, and contrast to well-preserved opal-A Paleogene specimens that compose the majority of the assemblages. Present among these "Cenosphaera" ghosts are specimens of the Cretaceous genus *Dictyomitra*, which indicates that most or all of these replacements are of an older redeposited assemblage. Of the well-preserved microfossils, the diatoms *Melosira sulcata* and *Stephanopyxis turris* are the most abundant, but these long-ranging species are not biostratigraphically significant.

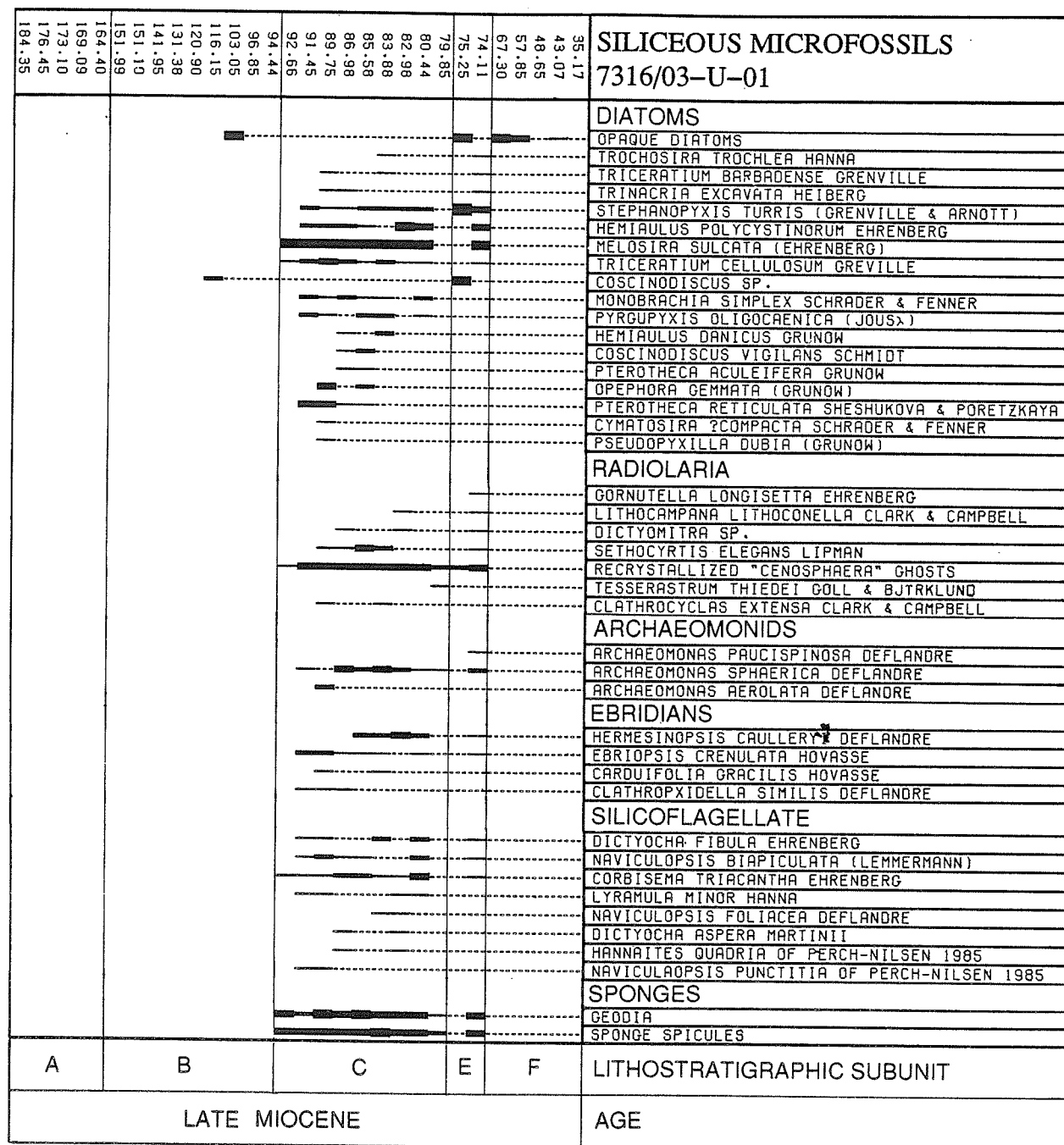


Figure 3.4.17 Distribution chart of siliceous microfossils in core 7316/03-U-01, sorted according to earliest occurrence.

In order to determine the age of deposition of the lapilli beds in Subunit D, two samples were analysed at 79.85 m and 75.25 m. Very few siliceous microfossils were found in these samples. The sample from 79.85 m contains rare specimens of *Archaeomonas sphaerica*, which may be reworked from older deposits, and badly recrystallized radiolaria including *Tesserastrum thiedei*. These recrystallized specimens are regarded as in situ. *T. thiedei* was reported by Goll and Bjørklund (1989) to be abundant in upper Upper Miocene sediments of the Vøring Plateau, and its age range is restricted to 5.3–6.1 Ma. This time span is regarded as representative of the time of emplacement of the lapilli beds.

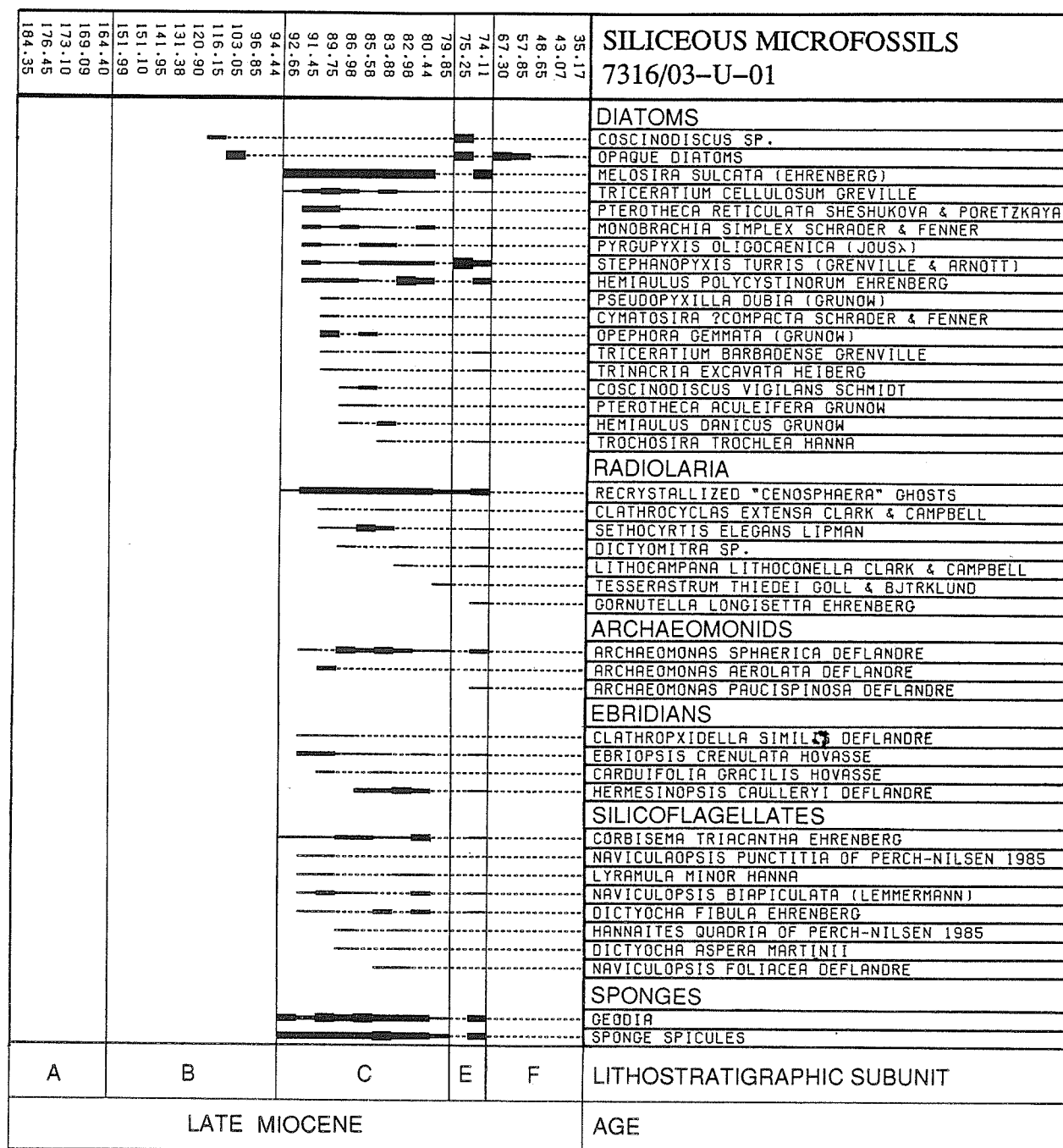


Figure 3.4.18 Distribution chart of siliceous microfossils in core 7316/03-U-01, sorted according to latest occurrence.

Many of the other diatoms and radiolaria are most probably reworked. Several of the taxa in subunits C and E have been reported from the Middle Eocene siliceous oozes of the Vøring Plateau by Schrader and Fenner (1976) and Dzinoridze et al. (1978), and these faunas and floras have affinities to those reported for Siberia. The persistent occurrence of *Corbisema triacantha* indicates an age that is not older than the *Naviculopsis lata* Zone of Martini and Müller (1976). The base of this zone is assigned an age of 28 Ma by Goll (1989). The other silicoflagellates are more consistent with a Middle Eocene age, however. There is also some evidence of a still later age in

this reworked assemblage. Rare to few specimens of *Coscinodiscus vigilans* occur in the samples at 86.98 m and 85.58 m. Schrader and Fenner (1976) reported the occurrence of this species as ranging from the *Pseudodimerogramma filiformis* Zone to the *Nitzschia maleinterpretaria* Zone (Late Oligocene to Early Miocene; 19-29 Ma, according to Goll 1989). *Opephora gemmata* also occurs in the samples at 85.58 m and 89.75 m. Schrader and Fenner (1976) reported this species as ranging from the *Synedra jouseana* Zone to the *Rhizosolenia bulbosa* Zone (15-21 Ma according to Goll 1989). These two latter species that represent only a very minor percentage of the entire siliceous microfossil assemblage, might indicate an additional younger source of redeposited sediment. It is also possible that bioturbation has caused some of the biostratigraphic anomalies in the assemblage of subunits C and E.

The only siliceous microfossils occurring in the four samples from Subunit F (67.30 - 35.17 m) are rare to abundant fragments of opaque diatoms similar to those in the sample at 75.25 m and in the samples from Subunit B.

Foraminifera

3

There is no straightforward explanation for the absence of in situ, i.e. Late Miocene foraminifera. Assuming a marine environment because of the presence of dinoflagellate cysts and siliceous microfossils, it is possible that seabottom conditions could have been unfavourable for living foraminifera or for the preservation of their tests.

In other fossil groups, reworked specimens from several earlier Tertiary and Cretaceous sediment sources have been recognized. Similarly reworked foraminifera fossils are extremely rare in this core. This is in clear contrast to the very abundant reworking in another depositional frame, i.e. the nearby Pliocene-Pleistocene subaqueous fans, which biostratigraphy has been described in NPD Contribution 27 (Eidvin & Riis 1989). In general, the dominant Paleocene arenaceous foraminiferal assemblages in the offshore basins of North Europe, very well supported erosion and resedimentation processes. Their virtual absence in core 7316/03-U-01 in an assemblage of reworked Paleogene dinoflagellate cysts and siliceous microfossils, could imply that they have been rare in their Paleogene source sediments too. This may confirm the observations in two other cores of this survey, where a very poor Paleocene arenaceous foraminiferal assemblage was found in core

7317/10-U-01, whereas the foraminiferal assemblage of core 7316/06-U-02 was quite different from the usual Eocene assemblage in wells from the Hammerfest Basin.

3.4.7 Organic geochemistry

Hydrocarbon composition data were obtained for the headspace and occluded gas fractions from fifteen canned core chip samples. Methane was the dominant gaseous hydrocarbon in the samples and was the only hydrocarbon present in the headspace fraction (Figure 3.4.19). Trace amounts of C₂ and C₃ hydrocarbons were present in all of the samples analysed, although C₄ and C₅₊ hydrocarbon traces were present in the lower three samples at 160.15 m, 169.12 m and 173.06 m. The highest methane concentrations occurred in and adjacent to the siltstones in Subunit E, while the lowest methane concentrations were observed in the uppermost (40.80 m) and lowermost samples (169.12 m and 173.06 m).

Methane carbon isotope data were obtained from two samples at 85.58 m and 169.12 m, poor sample concentrations preventing any data being acquired for ethane. The $\delta^{13}\text{C}_{\text{methane}}$ ratios of -47.5 o/oo and -30.4 o/oo are consistent with a thermogenic origin for the methane, although the possibility of a biogenic input cannot be excluded without additional data.

Subunit A, the lowermost unit in the core, may be characterised as organic-lean based on total organic carbon contents of between 0.12 wt% and 0.26 wt% (Figure 3.4.20). These low TOC values are accompanied by equally unpromising Rock-Eval S2 peak concentrations of 0.15 mg/g rock, and less. These equate to a largely inert type III/IV kerogen. The rocks of the overlying unit, Subunit B, are slightly more organic-rich with TOC contents of around 0.4 wt%. However, low S2 concentrations and hydrogen indices indicate that there is no significant change in the chemical character of the kerogen between subunits A and B.

The change to the claystone lithology of Subunit C is accompanied by a substantial increase in organic carbon content to a mean value of 0.8 wt%. As was noted in Subunit B, the increase in TOC content is not matched by an improvement in kerogen quality. The boundary with the overlying volcanic/sandstone interval which comprises Subunit D coincides with a marked decrease in TOC content, to values of between 0.14 wt% and 0.32 wt%. The TOC content of the uppermost interval from this unit shows an increase to 0.67 wt%, immediately before the sharp increase of TOC content which marks the base of Subunit E.

Subunit E is the most organic-rich unit in the core, with a mean TOC content of ca. 1.6 wt%. The organic carbon content of the rocks in this unit shows a tendency to decrease up the unit from a maximum of ca. 2 wt% at the base of the unit, to a minimum of ca. 0.6 wt% at the top. This tendency is mirrored by the hydrogen indices of the samples which decrease rapidly from a maximum values of 125 mg/gTOC at the base of the unit, to the background level for this core (i.e. less than 50 mg/gTOC) at around 73 m. The boundary between Subunit E and the overlying Subunit F appears almost gradational from the bulk geochemical data, although the boundary is rather more obvious from the lithology of the two units. Subunit F is characterised by lower TOC contents than Subunit E, with a mean TOC content of ca. 0.4 wt%. In addition, Rock-Eval data from Subunit F are more typical of this core than Subunit E.

Where reliable Tmax data exist from this core, values of between 415°C and 432°C suggest that the kerogen in the rocks is thermally immature. This indication is supported by vitrinite reflectance values of between 0.31% and 0.42%, low thermal alteration indices of less than 2.0 and yellow sporinite fluorescence colours.

7

It is interesting to note that thin, relatively organic-rich shales of Eocene age are associated with volcanic strata in the Hammerfest Basin (Bjørøy et al. 1983) and in the North Sea (Muré 1987). Whether the occurrence of these shales represents some influence of the volcanic episode which occurred in this area during the Tertiary, or whether these represent isolated events cannot be properly determined without further investigation.

The largely inert nature of the kerogen throughout most of this core is supported by pyrolysis-gas chromatography data. Most of the pyrolysis-gas chromatograms are dominated by aromatic peaks and lower molecular weight aliphatic peaks. In addition, a number of prominent unidentified peaks occur in the samples at 40.80 m, 110.12 m and 136.87 m which may represent contaminants. An expanded aliphatic peak homology is seen in the pyrolysis-gas chromatograms of samples from Subunit E, particularly that of the sample at 75.30 m. This is consistent with the increased hydrogen indices of these samples.

Microscopic examination of kerogen concentrates from selected samples indicates a dominance of woody and coaly debris (60-70%), the remainder of the kerogen consisting of amorphous and liptinitic debris. Traces of tasmanitid algae were observed at 168.15 m, while traces of Botryococcus algae were observed at 74.11 m. Given the lack of geological evidence for fresh/brackish water conditions in Subunit E, this material is likely to be reworked. Evidence of recycled Cretaceous material is present in a number of samples. The kerogen in samples from Subunit E differs from that in other samples from this core in a higher liptinite content of 35-50%. Algal debris is also present in the three samples between 74.11 m and 75.30 m.

The extractable organic matter (EOM) from a total of seven samples was evaluated (Figure 3.4.21), Subunit E being the most heavily sampled. The sample at 160.15 m was selected in order to further investigate the occurrence of C₁₅+ hydrocarbons in the occluded gas fraction. The samples from Subunits A and B contained very low EOM concentrations of less than 200 ppm. The two siltstone samples from Subunit C contained higher concentrations of between 322 and 240 ppm of EOM. The highest EOM concentrations, however, occur in the organic-rich siltstones of Subunit E, decreasing from a maximum of 1264 ppm at the base of the unit to a minimum of ca. 500 ppm at the top of the unit. These variations are not reflected in the sediment gas samples.

The asphaltene fraction accounts for between 24% and 80% of the total EOM, the most asphaltene-rich EOM being recovered from the sample at 136.87 m in Subunit B. The variation in the content of polar compounds roughly follows that of the asphaltenes, both fractions accounting for 60% to 80% of the EOM in total. The hydrocarbon fractions typically account for 30% or less of the total EOM, the lowest contents being found at the top of Subunit C and the base of Subunit E. This feature may reflect leaching of the more mobile elements of the EOM into a more porous volcanic-arenaceous Subunit D. Similar features have been reported from bitumen-rich horizons which are situated next to porous sandstones. However, in this case, this explanation is purely speculative without further study.

The gas chromatograms of the saturated hydrocarbons can be used to divide the seven samples analysed into three groups. The lowermost group consists of the two samples from Subunits A and B. Both of these gas chromatograms show marked "humps" of unresolved material, together with an n-alkane peak mode around nC₂₆. The four prominent peaks in the gas chromatogram of the sample at 136.87 m are very unusual and appear to reflect the co-elution of

a number of unidentified compounds with the C_{18} , C_{20} , C_{22} and C_{24} alkanes. The lower gas chromatogram at 160.15 m differs from the previous one in a marked n-alkane mode at nC_{17} - nC_{18} . Both samples have similar pristane/phytane ratio and carbon preference index (CPI) values (Figure 3.4.22).

The second group consists of the two samples from the lower siltstone interval in Subunit C, i.e. 79.85 m and 85.58 m. Both gas chromatograms show similar "humps" of high molecular weight unresolved compounds and an n-alkane mode around nC_{27} . However, the upper sample differs from the lower sample in the occurrence of a prominent n-alkane mode at nC_{17} - nC_{18} . Both samples have pristane/phytane ratios similar to those observed in the samples from Subunit A, although CPI values from the two samples at 79.85 m and 85.58 m were significantly higher.

The final group consists of three samples from the siltstone interval of Subunit E, i.e. 71.65 m, 71.88 m and 75.00 m. The gas chromatograms of the lower two samples were particularly similar with a marked n-alkane mode around nC_{29} , a strong odd-carbon number preference and the elution of a series of terpanoid and steroid peaks. The gas chromatogram of the uppermost sample shares the marked higher molecular weight n-alkane mode with the other two samples, but contains a less well-developed series of terpanoid and steroid peaks and has a more marked odd-carbon number preference.

With the exception of the extract sample at 75.00 m, pristane/phytane ratios from the seven samples were less than 1, although this could reflect loss of sample material rather than differences in source input. Carbon preference indices all indicate an odd-carbon number preference which is more marked in the samples from Subunit C and the top of Subunit E, than in those from Subunit A, B or E.

Gas chromatograms obtained for the aromatic fraction of these samples showed no identifiable peaks, but often showed a marked "hump" of unresolved compounds. The composition of the aromatic hydrocarbon fraction would be consistent with either low thermal maturity or a residual bitumen.

Pyrolysis-gas chromatograms were obtained for the asphaltene EOM fractions of the samples at 71.88 m, 75.00 m and 79.85 m. The two samples at 71.88 m and 79.85 m show similar, well-developed aliphatic homologies and can only be properly differentiated by the height of a currently unidentified peak which elutes prior to the (m+p)-xylene peak.

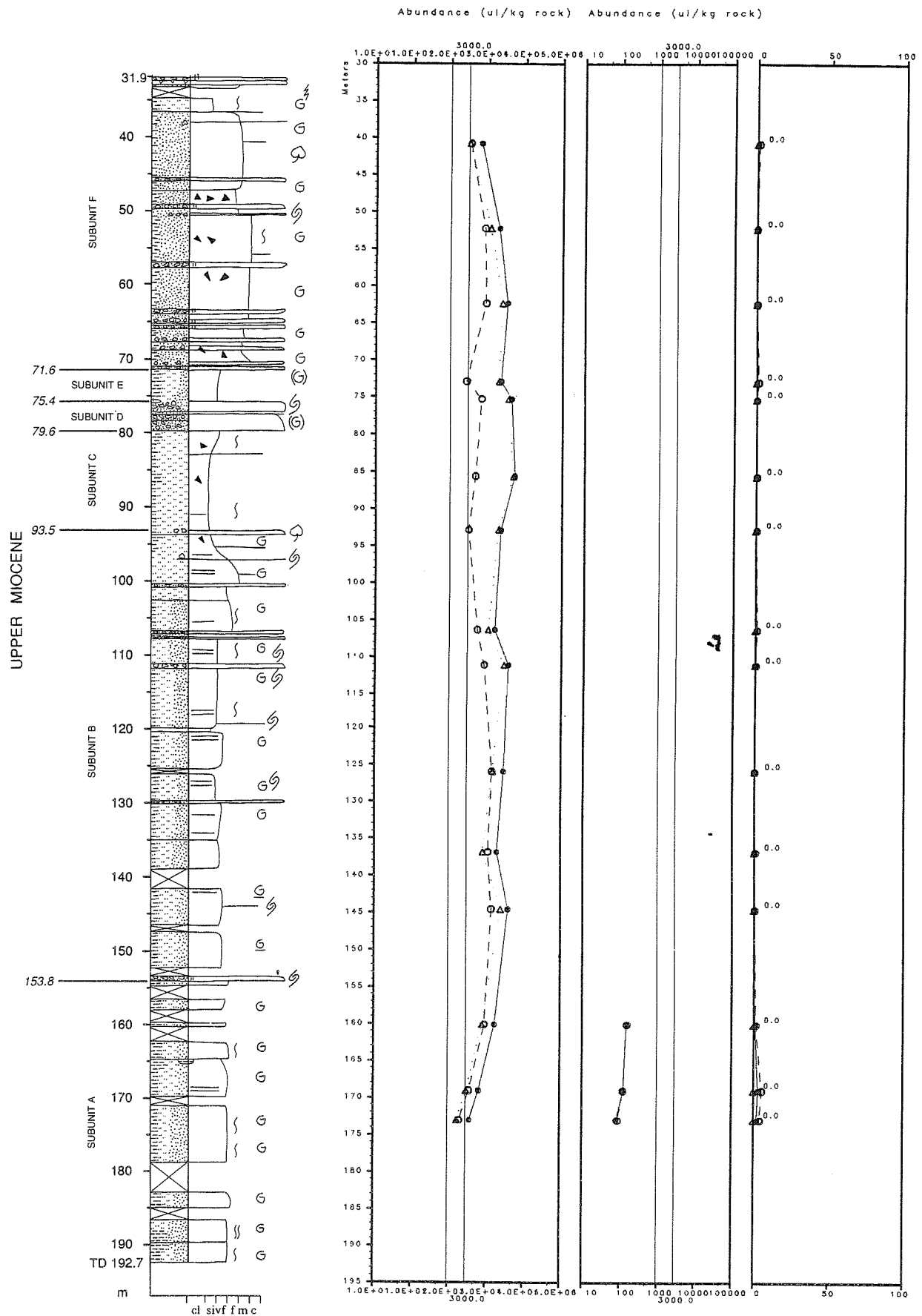
This peak also occurs in certain of the poorer whole rock samples where it may reflect the presence of bitumen. The asphaltenes from 75.00 m differ from the other two samples in the greater development of aliphatic peaks which show a multi-modal distribution, and the prominence of the prist-1-ene peak. These differences may reflect contamination of the bitumen by the more abundant liptinite-derived material at 75.00 m. The variation in asphaltene composition could also indicate different sources for the two groups of samples.

The steroid and hopanoid composition of three samples at 71.88 m, 75.00 m and 160.15 m was investigated using combined gas chromatography-mass spectrometry (GC-MS). The sample at 160.15 m was distinguished by a near total lack of response on the m/z 217 mass fragmentogram and the occurrence of a homology of unidentified compounds on the m/z 191 mass fragmentogram. In contrast, the mass fragmentograms of the other two samples showed very similar distributions of compounds.

The terpane (m/z 191) mass fragmentograms showed typically immature distributions of compounds with high relative abundances of 17 β (H), 21 β (H)-hopanes and 22R hopanes.

The sterane (m/z 217, 218) mass fragmentograms of the two upper samples are dominated by two prominent peaks which are believed to represent pentacyclic triterpenes and which have been associated with land-plant material. There are no other recognizable peaks on the m/z 217 mass fragmentogram. The dominance of these two peaks is suggestive of a low thermal maturity (Ten Haven & Rullkötter 1988).

In conclusion, the rock succession penetrated by this core is typified by rocks containing relatively small amounts of thermally immature, inert (Type IV) kerogen. More promising type III kerogen is restricted to a thin siltstone interval in Subunit E which overlies the volcanic ash/sandstone-dominated Subunit D. Traces of heavy bitumen which occur in the cored succession show evidence of having received a significant input from terrestrially-derived organic matter. These bitumens appear to share a common source, although changes may have occurred to the bitumen since generation. It is considered unlikely that this material has been sourced by the cored material, and the bitumen may therefore represent either migrated or reworked material.



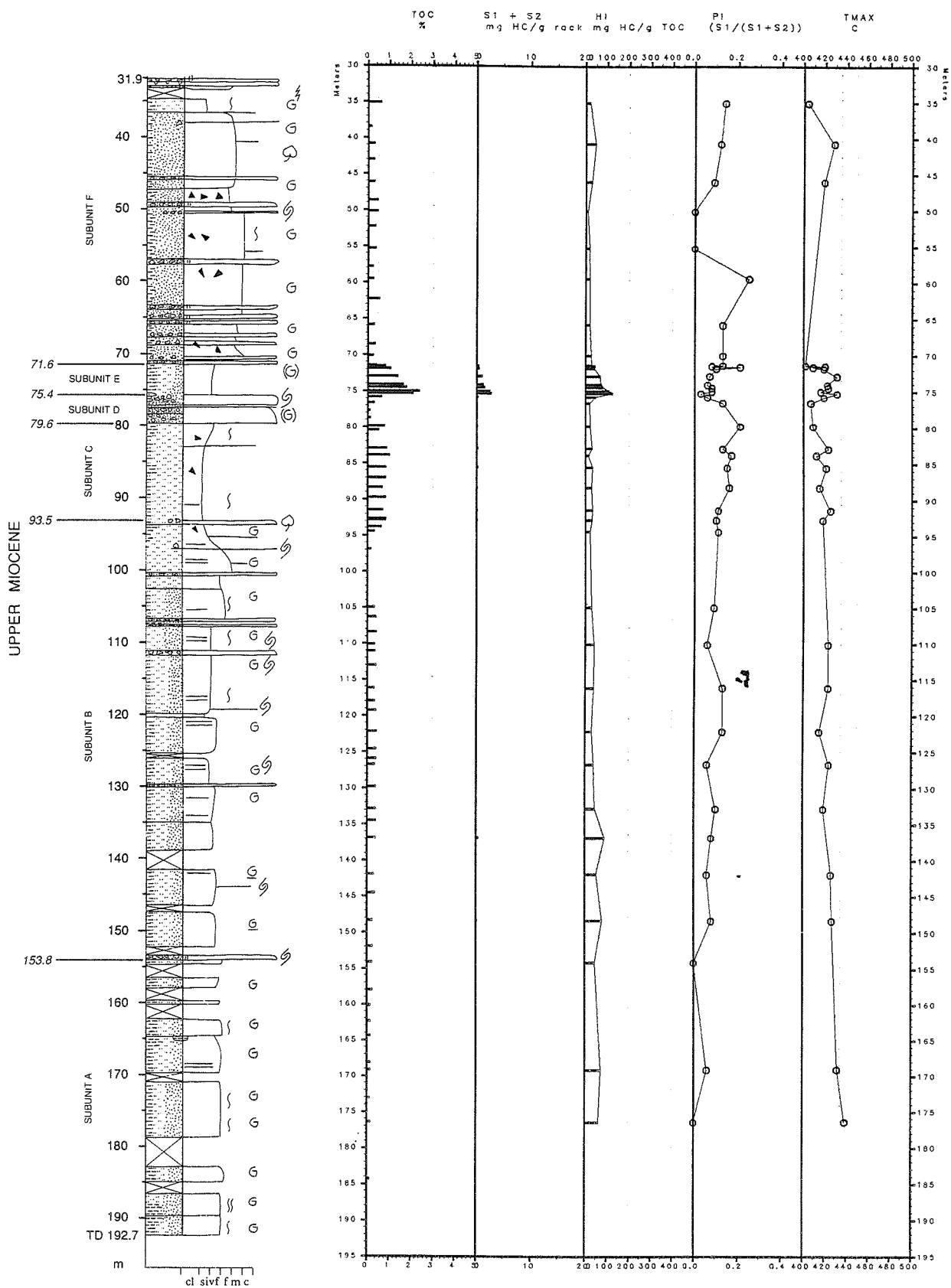


Figure 3.4.20 Total organic carbon content and Rock-Eval data of core 7316/03-U-01.

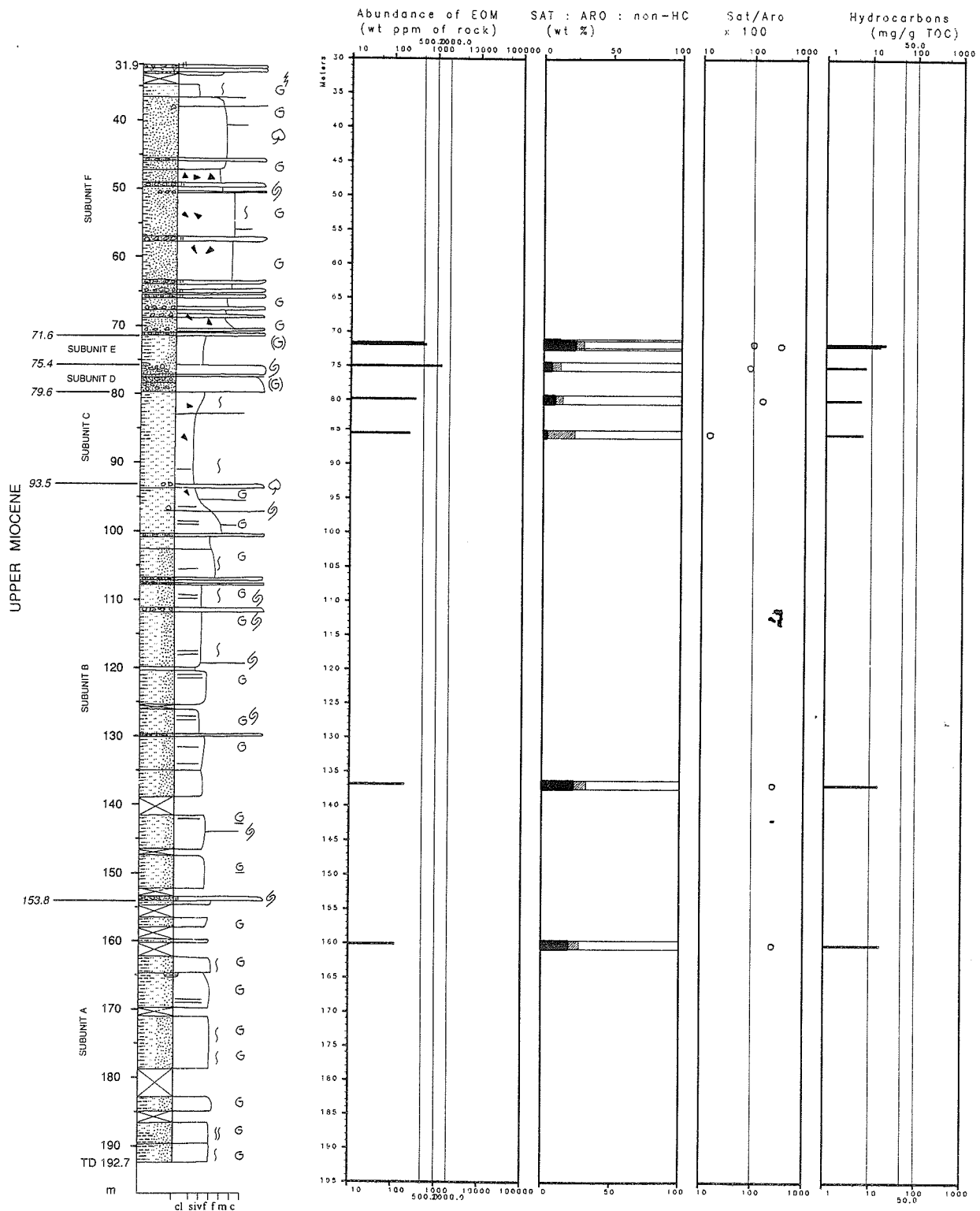


Figure 3.4.21 Extract data for core 7316/03-U-01.

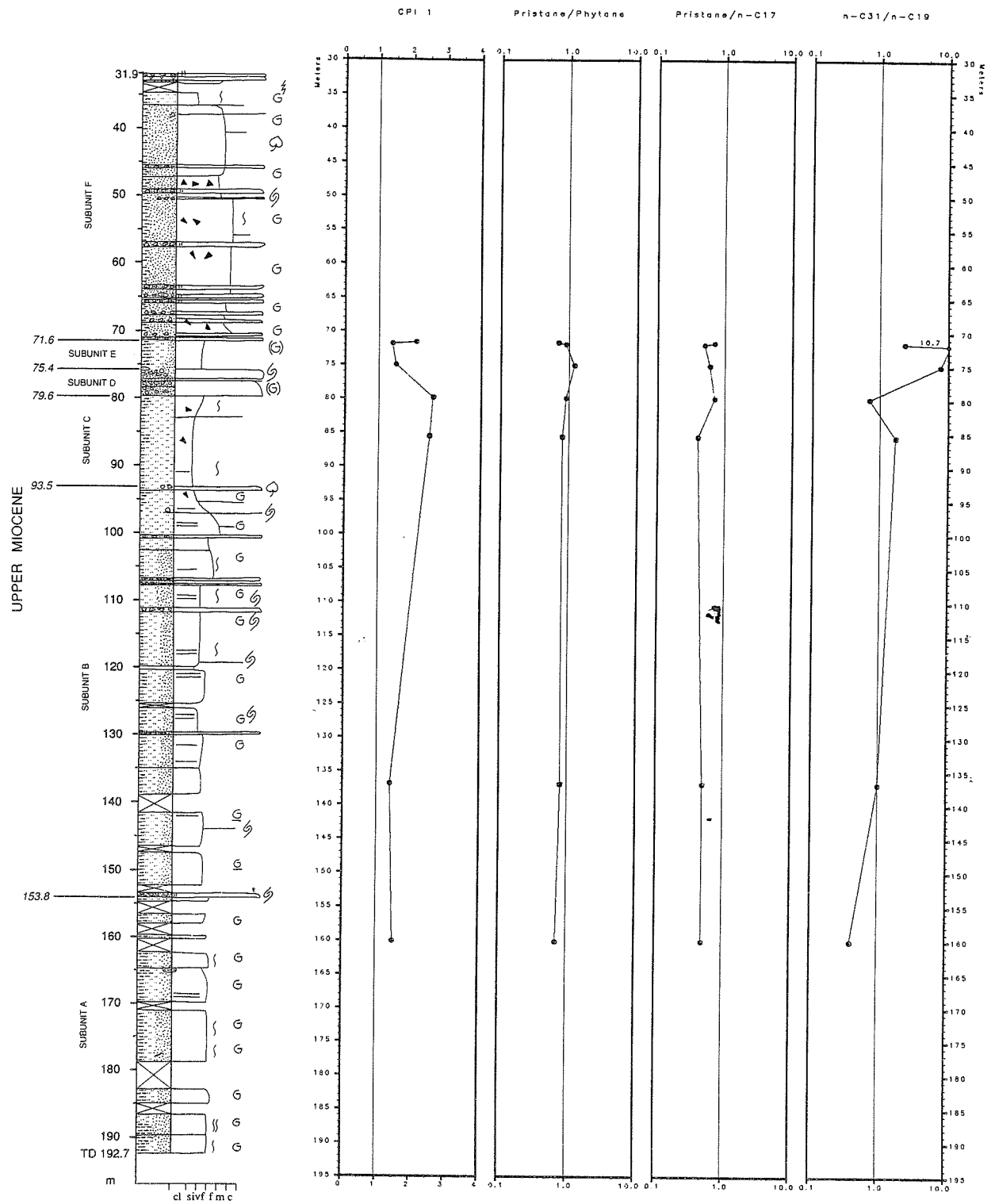


Figure 3.4.22 Compound ratios from gas chromatograms of saturated hydrocarbon fractions from core 7316/03-U-01.

3.4.8 Depositional environment

All the subunits (Figure 3.4.23) of the core are characterized by the high content of reworked fossils, mainly by Eocene forms but also including Cretaceous forms, and in situ fossils indicating a Late Miocene age. Glauconite (?reworked) is present in all sandstones. A large proportion of the silty clays and claystones of subunits C and E may be derived from Middle Eocene siliceous ooze. The whole core forms a major coarsening upward sequence, interrupted by the fine-grained material of subunits C and E. From the seismic interpretation it is also suggested that major parts of the core seem to be deposited by gravity flow mechanisms, although the lower subunit has a channel like pattern.

The volcanic debris occurs throughout most of the core from the base of Subunit B upwards. Volcanic material (Subunit D) also separates the two fine-grained intervals subunits C and E.

Beds with volcanic material occur at 20 levels in subunits B, D and F. Their similar appearance points towards a common mechanism⁷ of deposition. Each bed may be regarded as representing single depositional episodes; which probably occurred abruptly because the beds directly overlie horizontally laminated sediments and even rest directly on top of vertical trace fossils. The abundant loading structures at the base of the volcanic debris beds also indicate that the volcanoclastic debris was deposited rapidly on soft sediments with high water content. Many of the beds contain coarse volcanic material in the lower part and fine upwards. Some of the beds are unsorted as coarse material occurs also in the upper part of the beds. Gradational transitions from volcanic debris into the overlying sandstones or siltstones are common. The volcanic clasts are all basaltic with only small compositional variations, indicating a common origin.

Radiometric ages of the volcanic material varies from 10 Ma to 2.5 Ma. There is no systematic differentiation in the ages between the subunits. Combined with the similar compositions of the volcanic clasts this is taken as evidence that all erupted during the same major time interval. We therefore conclude that some of the scattering of the K-Ar dates are artifacts and that the age of the volcanics is Late Miocene (-?Early Pliocene).

Features that seem to be absent in the core are evidence for sub-aerial erosion, macrofossils, abundant bioturbation and abundant sedimentary structures. This indicates high sedimentation rate and deposition in an environment not associated with normal shelf deposition.

No other depth information is found than lack of surface related structures indicating that deposition took place below the depth where storms affect the sediment. The regional setting of the drilling locality between the Stappen High to the east and the spreading ridge further to the west may indicate that sedimentation took place outside the shelf escarpment, possibly on the slope.

All subunits contained reworked Cretaceous fossils, indicating that Cretaceous sediments had been exposed for erosion. The dominance of Eocene fossils indicates that this erosion and deposition occurred in Eocene time. Judging from the fauna and palaeoflora this ?Eocene deposition may have been in a shelf environment, and considering the position on the western margin of the recently elevated Stappen High, a prograding shelf environment would be probable. Glauconite occurs in noticeable amounts in all sandstones, but we hardly believe that it was formed during the rapid deposition in Miocene. Abundant glauconite is common in Cretaceous shales in Arctic areas, and such sediments may have sourced this core as well as core 7316/06-U-01. Glauconite may, however, have been formed on the Eocene shelf.

4

A common mechanism for sediment transport on the continental slope is by gravity driven mass flows. Grain flows, debris flows and turbidity currents vary by the slope gradient and the content of fine-grained and coarse-grained material available in the source area. Mixtures of these mechanisms and transitions from debris flows to turbidity currents often occur, and in areas with many sediment sources along the margin edge, they may form a slope-apron (also debris-apron) system. In the present core most of the deposition may have taken place by these mechanisms. The overall coarsening of the sandstones and thickening of the independent volcanic debris beds may indicate deposition in gradually shallower part of the slope. Lack of distinct feeding channels in the major part of the core may indicate that a slope-apron was developed in the area.

Subunit A shows no graded beds, but as the unit is indicated to contain abundant channels on the seismic record, the subunit may be interpreted to be dominated by slope channel fills. Such channeling is common in slope settings on the "middle" part of the slope. This unit is seen on the seismic data to extend downward to an important unconformity. Dating of our cores seems to limit the age of this unconformity to earliest Oligocene/ Early Miocene.

Subunit B: The transition from Subunit A to Subunit B is observed as the possible top of a channeled unit on the seismic record and marked by the first appearance of volcanic debris beds in the core. From the seismic record deposition by debris flows and turbidity currents is suggested for Subunit B. In the core there is little differences between these two lower sandstone units, except for the volcanics of Subunit B. The beds with volcanic debris may be regarded as deposited by turbidity currents or by debris flows. Several of the beds have loaded bases indicating rapid deposition. Sorting is mostly poor, although gradational upper transition to sandstone often occurs. The introduction of volcanic material shows that such material was present at the site of erosion. As the radiometric dating is consistent with the Late Miocene fossils of the sediment, it shows that the volcanic eruption was shortly followed by erosion and sedimentation. It was argued earlier that the eruption occurred in the vicinity of a sedimentary basin with explosive magma ascent through sediments and shallow water. Possibly the volcanic debris beds have been redeposited by debris flows triggered by earthquakes associated with the volcanic activity.

Subunit C represents the silty shale separating volcanic material. On the seismic record this interval occurs between two continuous reflections that may indicate that the interval represents a thick turbidite or normal marine deposit, generated in areas exposing beds with Eocene siliceous ooze. The sparse lamination and bioturbation observed at some levels show that the deposition probably represents several episodes. Hemipelagic deposition is also possible, however, as the unit is almost barren for Miocene fossils, this mechanism is not very likely.

Subunit D represents five fining upward beds with volcanic debris in sandstone and unsorted beds with mixtures of volcanic debris with larger boulders. The unit is the one showing the clearest evidence for deposition by debris flows. The concentration of debris beds in this thin interval may represent deposition during a period with high earthquake activity.

Subunit E: This silty shale unit is similar to Subunit C and was probably deposited by a similar mechanism; i.e. either by debris flows/turbidity currents or hemipelagic sedimentation.

Subunit F: Uppermost in the core the fine to medium grained sandstones of Subunit F contain twelve, mostly fining upward, volcanic debris beds. In this subunit glauconite is again abundant as well as mud flakes/pebbles. Mass flow deposition by debris flows and turbidity currents may have been

initiated in an area with exposures of medium grained Eocene sandstone. The volcanic material was probably exposed in the same area. Deposition on the uppermost part of the slope seems possible.

The palaeogeographic setting during deposition is not clear, however, the sediment may be derived from sandstones of Eocene age exposed at the shelf margin. A fault system occurs 4-5 km to the east of the drill site and may have been active. Miocene volcanic activity penetrating the wet sediments may have deposited volcanic material on the shelf. This and possibly related earthquakes may have triggered slides on the shelf edge. Abundance and thickness of the volcanic debris beds thus reflect the available amount of such material at the escarpment when the slides were triggered.

Some polished and extrabasinal clasts with possible glacially imposed features occur throughout most of the core. As discussed in Section 4.2 this may indicate that adjacent land areas were glaciated at the time of deposition of these sediments. This may imply that a glacier could have eroded the Stappen High and provided a high sediment input to the west. This could have supplied sediments to the drill site area directly or indirectly by feeding the sediment source area for core 7316/03-U-01. It was also mentioned in Section 3.4.4 that the shape of some of the smallest ash particles could reflect supercooling and splintering due to eruption through water or into ice, thus supporting the possibility of glacial environments in the area.

A Late Miocene age for the debris flow deposits could be correlated with a major sea level fall reported by Haq et al. (1987). The low sea level may have exposed sediment for erosion and aided redistribution of the sand from shelf setting to the slope.

The present model of gravity mass flow sedimentation on the (upper) slope explains the mixture of different ages and provenance of the material of the core. It is, however, not known whether both Cretaceous and Eocene (-?early Oligocene) and also metamorphic sources for the sediment were exposed on the shelf during the Miocene or, alternatively, if the slope sediments represent several cycles of redeposition. The low diagenesis and lack of any significant burial indicate a regional setting principally similar to that of today with a broad shelf to the east and a slope facing westward.

4. DISCUSSION

4.1 Seismic interpretation

Introduction

The data base for the interpretation has been the IKU shallow seismic (Figure 1.1) together with selected NPD lines in the Bjørnøya West area (Figure 4.1.1). For practical purposes we have divided the survey area into three parts, all containing at least one shallow well. These will be denoted as the northern, central and southern areas. The rectangular data grid in the central area is the basis for the outline of the unconformities and the sequences; we will later try to connect this interpretation to the observations we can make in the northern area. The southern area, which contains the oldest core (7317/10-U-01) is difficult to tie to the central area with our data base and will not be discussed in this section.

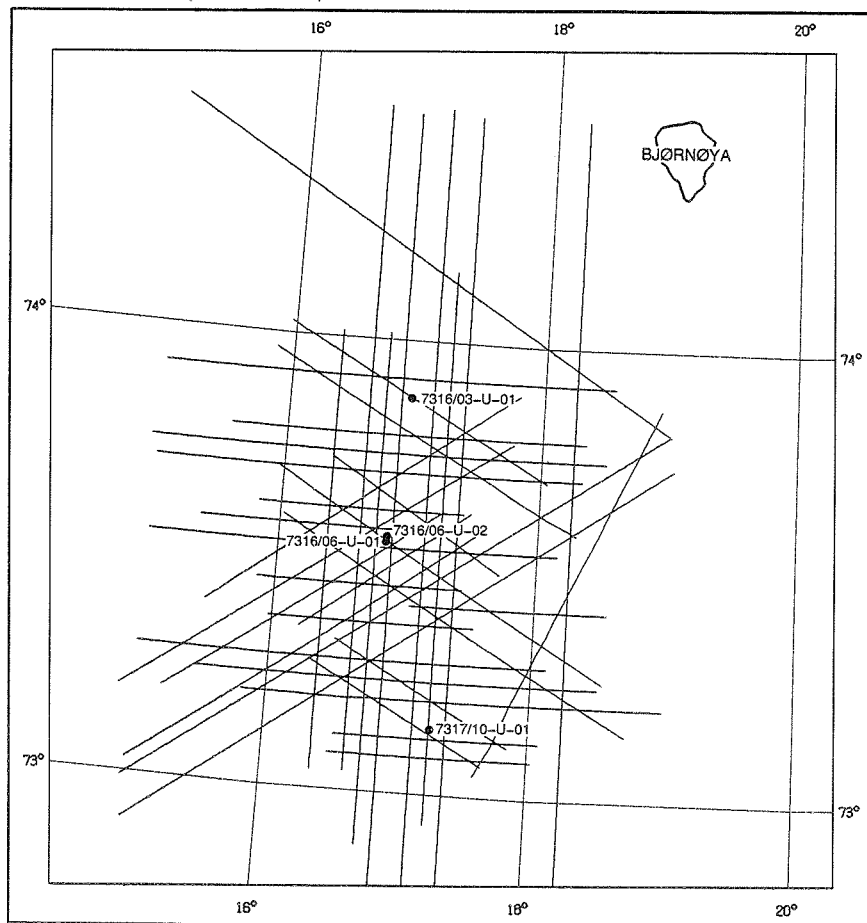


Figure 4.1.1 Seismic grid of NPD lines made available for this project.

Definition of main unconformities

Four unconformities have been selected as being important for the understanding of the evolution of the area. The uppermost unconformity that we will describe (Light Purple in Figure 4.1.2), is the dominating erosional surface seen on seismic data in this area, partially truncating all the underlying sequences that will be described below. This unconformity forms the base of a thick sequence of lobe-shaped sedimentary structures, which probably are some kind of fans, and considered to be of Pliocene age (Eidvin & Riis 1989). The succession below this surface consists of mainly parallel reflections. In some bands the reflections occur more frequently, this being an indication of layered, fine-grained material.

The next unconformity (Light Green) is also truncational and the sequence below varies considerably in thickness (300 msec at line 732230, sp 1100, to less than 50 msec at line 733230, sp 700). The sequence is characterized by some downlapping reflections, which may indicate that this is a prograding sequence, with the proximal part to the south. On the lines 732230 and 7325, the sequence actually gives the impression of being a prograding delta.

Below this sequence there is a surface (Orange) which is characterized by a series of channels (see also Figure 4.1.5). Also the sequence below this surface seems to be heavily channeled. The thickness of this sequence increases considerably in the basin separating the northern and the central areas. This sequence is interpreted to contain the 7316/06-U-01 core.

The last of the unconformities described in this chapter (the Green reflector) truncates the underlying sequence with increased erosion in a northern direction (Figure 4.1.3). This is different from the above unconformities, where the erosion increases westward. In the search for a possible Mid Oligocene unconformity, this unconformity will be a likely candidate. The underlying sequence consists of parallel to sub-parallel, high amplitude reflections. Within the sequence we have interpreted another reflector (Purple, see Figure 4.1.2), which can be correlated into well 7316/06-U-02. Further down (at approx. 3 sec. on Figure 4.1.3) some very high amplitude reflections can be observed in most of the study area. These are interpreted as a basaltic layer associated with volcanic activity at

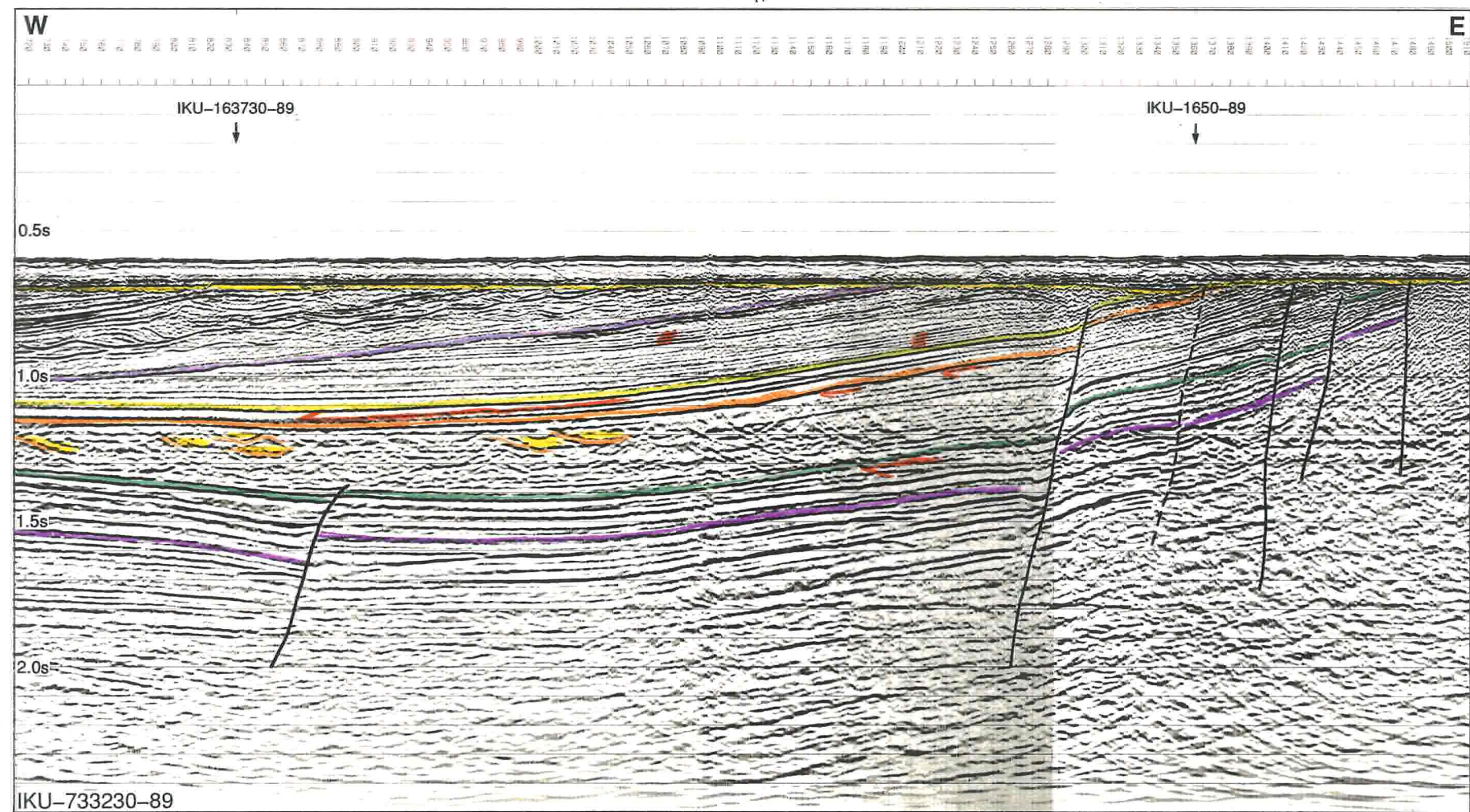


Figure 4.1.2 Part of shallow seismic line IKU-733230-89 showing the outline of the unconformities and sequences described in the text.

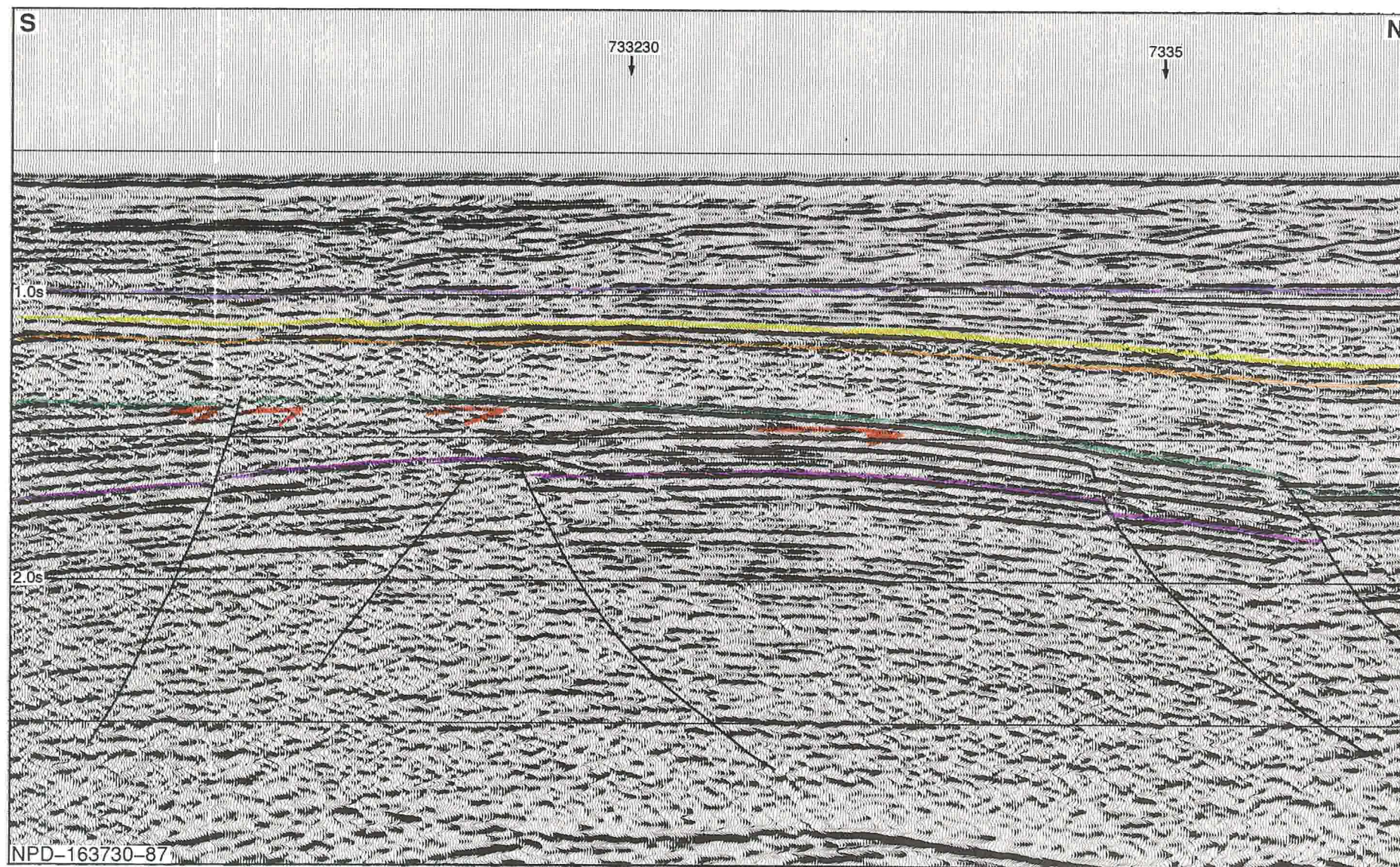


Figure 4.1.3 Part of seismic line NPD-163730-87 showing the "Green reflector" truncating underlying reflections.

the time of the opening of the Norwegian-Greenland Sea in early Eocene time.

Figure 4.1.4 shows the stratigraphic position of the four wells plotted into the regional NPD line 733230. Note that the well 7317/10-U-01 is put just below the basaltic layer, based on the datings and the assumption made above about the age of this layer.

Correlation and tectonic interpretation around wells 7316/06-U-01 and U-02

One of the criteria for selecting sites for shallow drilling, is the possibility of making ties to regional seismic and to carry the information into prospective areas with some degree of confidence. For the Eocene/Oligocene, which is an important interval in the Tertiary in this region, it is particularly difficult to fulfil this criterion. In fact, without moving the actual shallow seismic line from its original location to 16°50'E, we would have had much poorer alternatives for these targets. However, even on the line 1650 there are problems to understand the somewhat peculiar geometry and the tectonic movements that have occurred in the vicinity of the drill sites.

Whatever interpretation we decide upon, we believe that there must be out-of-plane reflections in the dipping part of the section that obscure the geometrical picture. Using the assumption that the reflection pattern commonly changes very little over short distances, it is possible to "cross-correlate" the reflection series on both sides of critical points. Figure 4.1.5a shows the most probable correlation across a possible fault with the resulting interpretation shown in Figure 4.1.5b, where, because the reflections may not come from the plane of paper it is difficult to draw the fault plane correctly.

Further updip we can observe two other faults, which by correlation have been interpreted to have reverse movements. One of them is more like a relatively low angle disturbance with only minor vertical displacement. This tectonic pattern may be explained as being part of a strike-slip regime creating compressional forces as the system makes a westward shift going from south to north. This will be further discussed later in this section (Figure 4.1.7). As a consequence of this interpretation, the large unconformity denoted as the 'Green reflector' will subcrop between the two wells. This unconformity seems to represent a major hiatus and accounts for the time gap from late Eocene/early Oligocene to early Miocene.

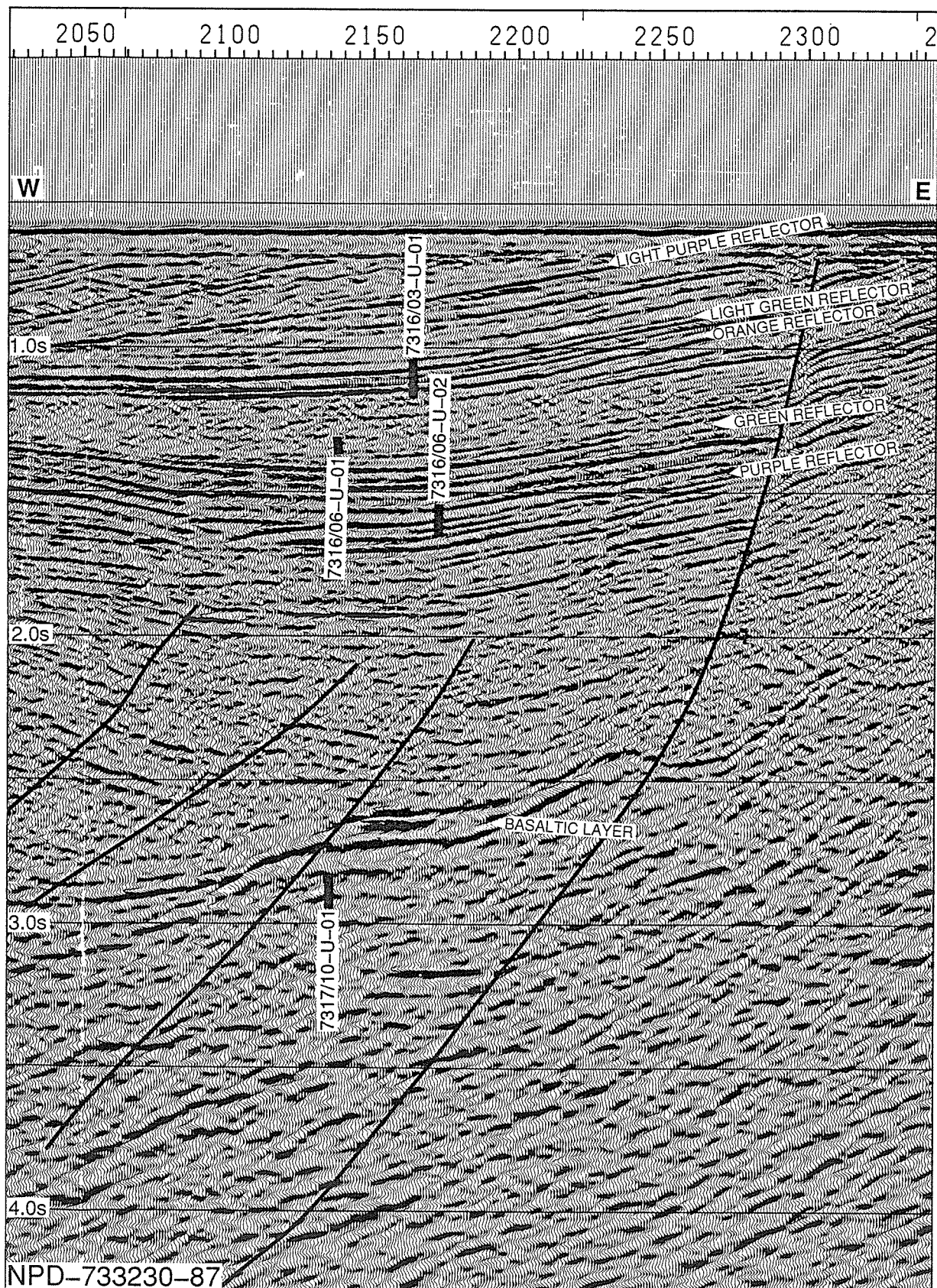


Figure 4.1.4 Stratigraphic position of the four 1989 shallow cores plotted on seismic line NPD-733230-87.

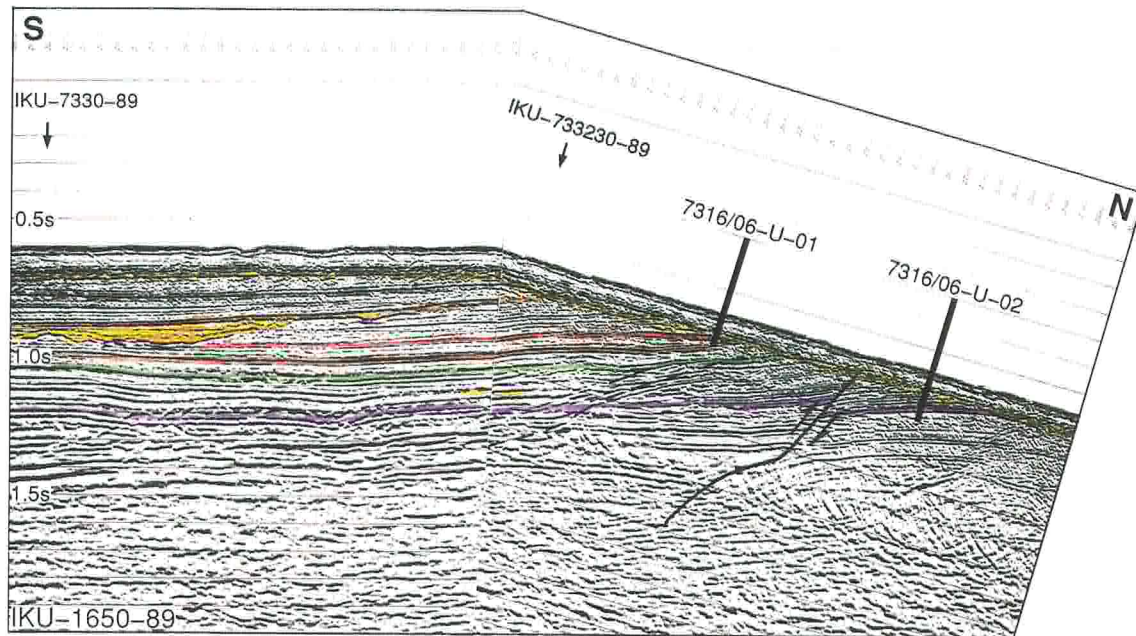


Figure 4.1.5a "Cross-correlation" on both sides of a difficult point on line IKU-1650-89, close to wells 7316/06-U-01 and 02.

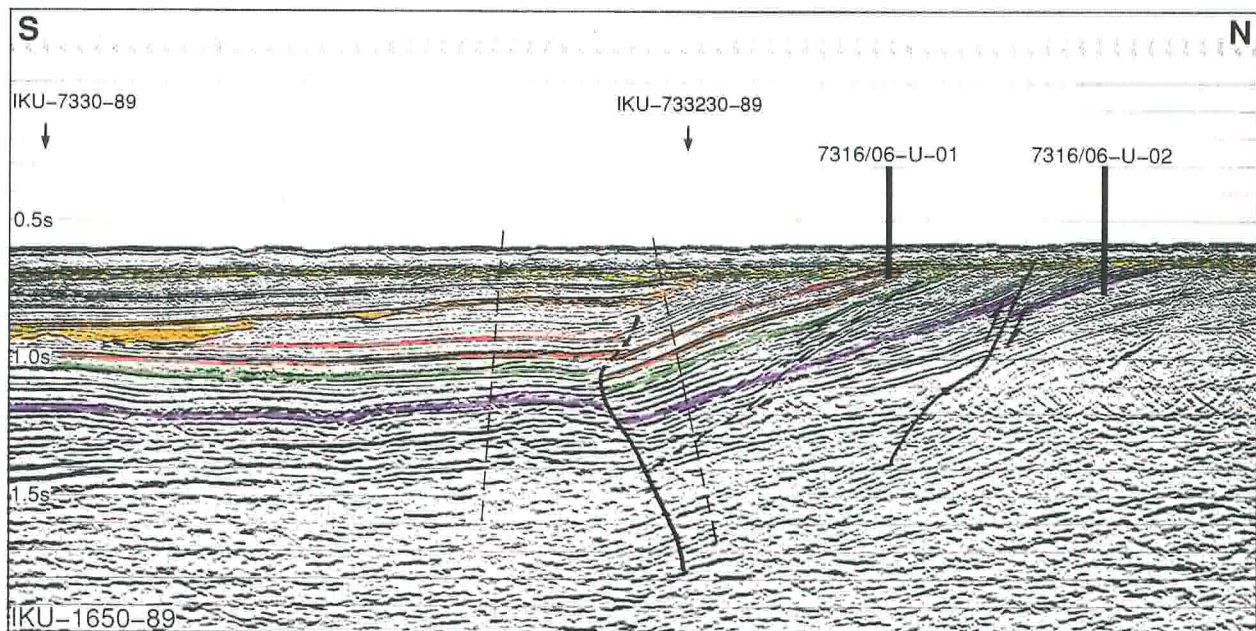


Figure 4.1.5b Interpretation of line IKU-1650-89 resulting from the "cross-correlation" above. The interpretation gives correlation of the cores to the seismic data in the western part of the study area.

Interpretation between the central and the northern areas

However important, the seismic correlation between the central and the northern areas has proven to be very uncertain, at least with the seismic data that have been available for this project. The two areas are separated by a deep basin, bounded to the south by a fault that is offsetting the Paleocene/Eocene basaltic layer which can be seen in most of the study area. The interpretation into this basin from the south is fairly simple, but at the northern end of the basin the reflections become obscure and difficult to follow. Disturbances that may be faults with unknown displacement will give different ties between the two areas.

Nevertheless, based on a "conservative" interpretation (regarding the faults) and a comparison of observations in the two areas, we have reached an interpretation which is reasonable and fits with the age datings in the three wells 7316/03-U-01, 7316/06-U-01 and 7316/06-U-02. Figure 4.1.6 shows the interpretation of line IKU-BV11-89.

Below the base of core 7316/03-U-01, at 0.8 sec, there is an erosional unconformity that marks the top of a sequence with high amplitude, parallel reflections. This is similar to, and is interpreted to be, the same unconformity as seen between cores 7316/06-U-01 and -02 (Green reflector). The sequence above this horizon has less internal reflections and is characterized by several channels. The upper part of this sequence may have been penetrated by the well 7316/03-U-01 (Subunit A). The turbidite deposits associated with the westward thinning band of reflections (Section 3.4.2) is probably not regionally widespread and cannot be correlated, but this interval may correspond to the prograding unit described earlier in this section. The base of the Pliocene fans is rather easy to correlate and subcrops at shotpoint 640.

The subcrop map

The subcrop map resulting from the interpretation (Figure 4.1.7) clearly shows the restricted areas for possible Eocene drill sites that can be correlated to regional seismic data with confidence. The western part, which is easiest to interpret, is dominated by Pliocene and Miocene sediments. To the north-east, a syncline controlled by a the large boundary fault separating the Tertiary region and the Stappen High can be observed. The sediments in this syncline can be correlated to the shallow well 7317/02-U-01 (D85-V). However, the seismic data do not allow a good

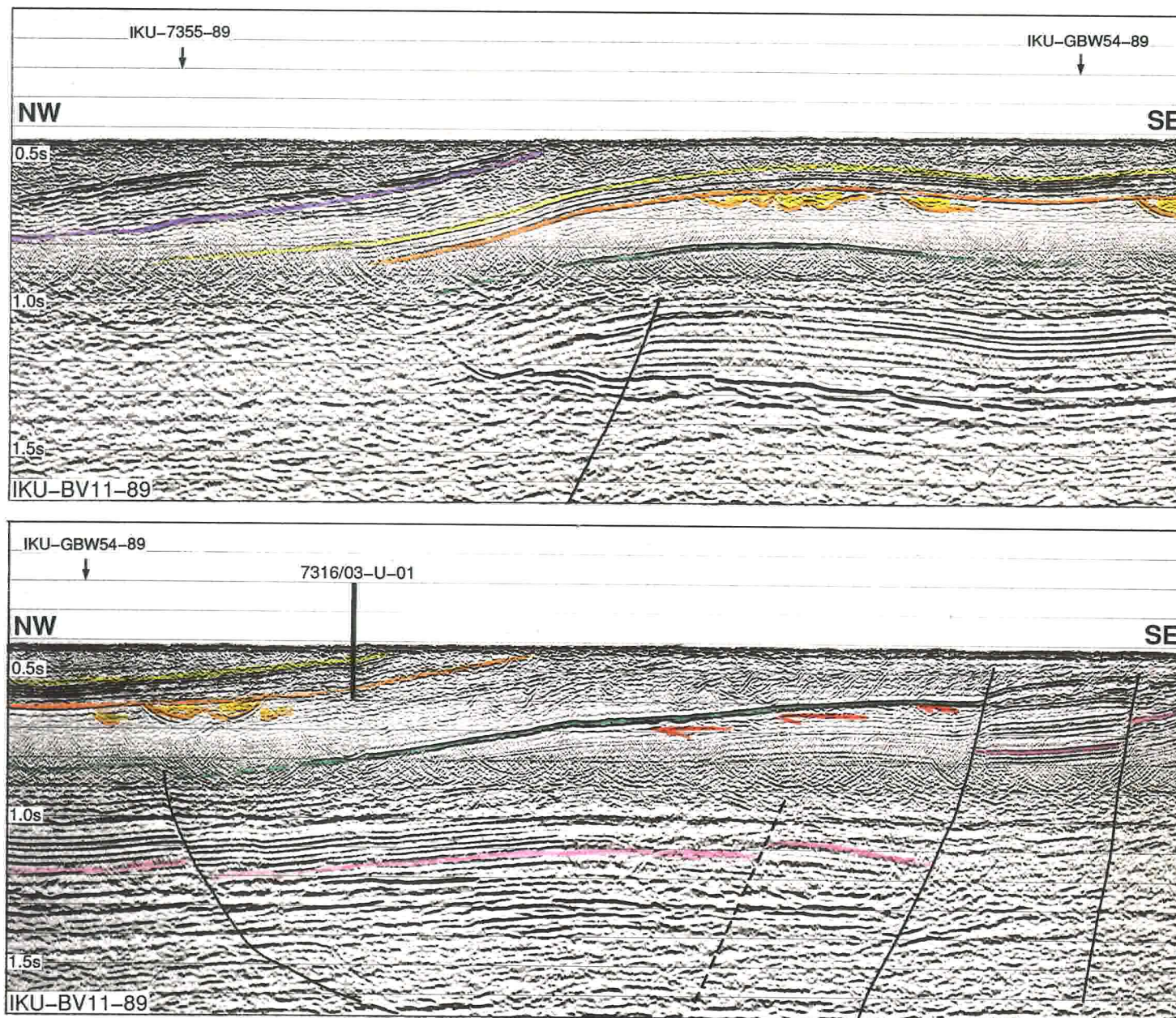


Figure 4.1.6 Interpretation of line IKU-BV11-89. The upper part of the figure is the western part of the line.

interpretation westwards. The pink horizon marked on Figure 4.1.6b is a tentative correlation to the interval penetrated by well 7317/02-U-01 (D85-V).

Another syncline containing evenly layered sediments is seen further south. However, these layers are also difficult to interpret across the fault zone. The best opportunity is probably along the line 7325, and a suggested interpretation is shown on the map. Normally the regional NPD lines would have been an aid to the interpretation of the faults, but unfortunately the survey containing this line has extremely poor quality compared to other 1986 and -87 data; consequently these data are of no help. This is also the case for line 7355, that could have been an aid to the interpretation in the northern area.

Apparent from the map are also some channel systems, cutting into the Pliocene fans and older sediments. These channels will affect the location of the subcrop lines, and also define areas unsuitable for shallow drilling.

7

Description of local tectonic features

The study area has undergone strong tectonic movements during several phases in the Tertiary. It is reasonable to assume that some fault systems have been reactivated in several of these phases. To the east, a large fault, known as the "Knølegga Fault", marks the boundary between the Stappen High and the study area (Figure 4.1.7). Another important fault zone is running north-south in the middle of the study area, and will be denoted as the "Central Fault Zone". To the west of this fault zone there has been less faulting activity, and only a few of these faults seem to have been active in the Miocene or later.

The area between the Central Fault Zone and the Knølegga Fault is strongly uplifted relative to the western part; this can best be seen by the large offset of the basaltic layer. This area can be separated into three parts. The first is the synclines described above which are associated with the Knølegga Fault. The second is an area with slightly rotated fault blocks and faint internal layering, which, as mentioned earlier, cannot be correlated across the Central Fault Zone. Between these two parts, there is a highly deformed area where the basaltic layer is broken up, rotated to

steep dips, and brought up to the surface. An area that has undergone such strong tectonic movements will naturally be a good candidate for the source area of the volcanic material found in core 7316/03-U-01, even if the seismic evidence for volcanic activity may have been removed by later erosion. The position of this area is more reasonable with respect to transport of the volcanic debris than the volcanoes that can be observed to the west (see Figure 4.1.7). Approximately 4-5 km east of well 7316/03-U-01 there is a fault (shown in Figure 4.1.6) that might be a controlling factor of the mass flows found in the core. The distance from the possible volcanic area to this fault is approximately 15 km.

The Central Fault Zone was apparently reactivated in different stress regimes. It gives the impression of being originally an extensional zone of normal faults. Some compressional features (for instance a "flower structure" on line 733730) and other faults cutting into the older normal faults indicate that it has been reactivated as a strike-slip fault zone (right lateral). The fault with the large "roll-over" structure may also have developed in this stress regime, dating the movements to Early Miocene. To the south, the north-south trend is taken over by an NE-SW trend. Compressional features at this location (around 73°15'N, 16°52'E) support the impression of strike-slip movements.

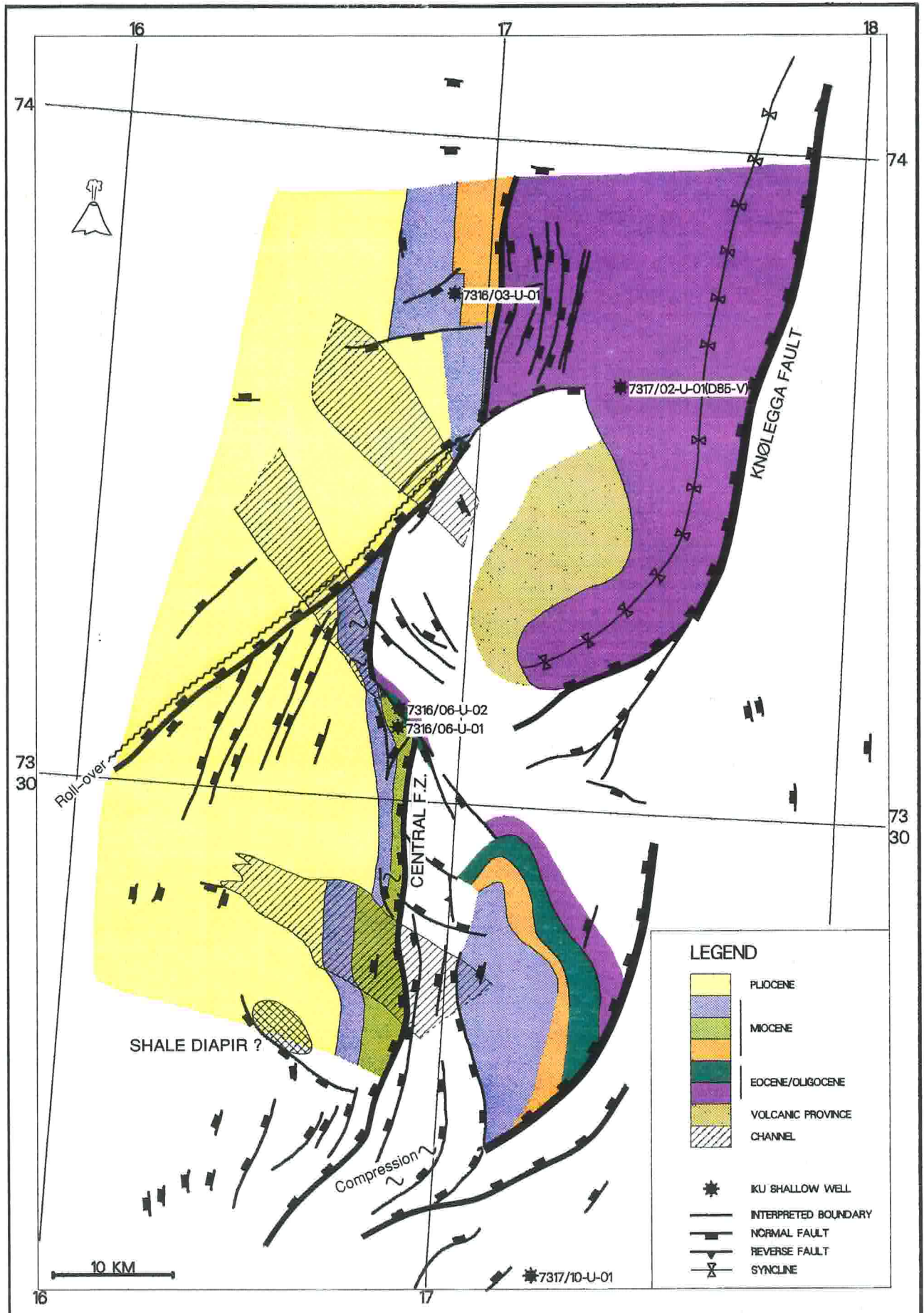


Figure 4.1.7 Subcrop and tectonic map of the Bjørnøya West area.

4.2 Glacial erosion

4.2.1 The record of Cenozoic climate and glaciations

Studies of ocean cores suggest that cooling of the Norwegian Greenland Sea started 13.6 Ma ago (Henrich et al. 1989). The earliest record of ice rafted debris (IRD) from the oceans is found in Baffin Bay, dated to 8 Ma (Arthur et al. 1986). Glaciers on land areas surrounding the Norwegian Greenland Sea were large enough to reach the coast 5.45 Ma ago, and have caused repeatedly IRD transport to the Vøring Plateau since then; with a major increase in IRD influx 2.56 Ma ago (Jansen et al. in prep.). In the Arctic Ocean IRD is recorded in sediments more than 4 Ma old (referred to by Henrich 1988: Aksu & Mudie 1985, Aksu 1985, Clarc et al. 1986), possibly as old as 4.5 Ma (Herman & Hopkins 1980).

There was a major increase in extent of glaciers advancing to the coast in Scandinavia 2.56 Ma ago, and the glaciations gradually became more severe during the period 1.2 - 0.6 Ma (Jansen et al. 1987, 1988). Still, some of the earlier glaciations were quite sizable, and a major glacial episode 5.1 - 5.0 Ma ago, for instance, lowered the sea level by 80 m. (The lowering caused by the last Pleistocene glaciation is loosely defined, but is somewhere between 80 m and 163 m (Matthews 1984)). The Antarctic ice sheet could, however, have played a major role in this event.

The oceanic record indicates that the first glaciers started to grow at high latitudes (Jansen et al. 1987, 1988, in prep.). It is further expected that glaciers existed in mountains and highlands in the north long before the first iceberg or sea ice transported IRD into the oceans. Significant on land glaciations on Svalbard may therefore have occurred prior to the first IRD record (5.5 Ma ago), possibly with some glacial activity as early as 8 Ma ago.

4.2.2 Observations in the present study

Site 7316/03-U-01: The bedrock core from this site contained a few scattered clasts of rounded quartzite, chert and one possible crystalline fragment. These clasts are found in subunits B, C and F, and are up to 2 x 1.2 cm in size, but normally less than 1 cm. They are rounded and polished but also have some faint fracture marks and impact pits. Angular -

subangular claystone clasts also occurred at several levels. In addition, irregular shaped vesicular clasts of volcanic material occur in abundance within several intervals. In particular, the co-occurrence of the rounded clasts with fracturing and the more angular clasts (the latter are mechanically weak) lead to the suspicion of glacial influence. The rounded clasts were studied by Lars Olsen at the Geological Survey of Norway. Although no clear striations were found, he believes that the clasts originated from a glacial environment, and considers the frequent occurrence of fractures, both fresh and later abraded, as indicative of glaciofluvial transport. The above indications, although far from conclusive, suggest that part of the sediments in core 7316/03-U-01 may have been sourced from a glacial environment.

Quaternary cores and seismostratigraphy: The results summarized below are preliminary as technical problems with amino acid measurements have delayed the work on the Quaternary sediments. Details and final results are given in Appendix A.13. The large glaciotectonic megablock complex, which was penetrated in borehole 7316/06-U-01, is described in Section 3.3.8.

7

A number of seismic units have been mapped on the outer shelf and upper slope. These have been tied to the boreholes and ages have been assigned to some units based on paleomagnetic polarity and amino acid measurements. The results are summarized in Figures 4.2.1 and 4.2.2. The paleomagnetic data show that all the sampled sediments were deposited during the Brunhes normal polarity chron (0 - 730 000 years BP). A few samples have a reversed polarity, but constraints from amino acid data show that they are younger than 730 000 years. We have not been able to assign any age to seismic unit A (Figure 4.2.2) which possibly was cored in borehole 7316/06-U-01.

Preliminary interpretation of the amino acid measurements on reworked shell fragments indicates that 350 m to more than 1000 m of sediments are deposited during the last approximately 200 000 years on the upper slope (at 73°30'N and 73°N respectively). This age is indicated from comparison with amino acid ratios in molluscs in a till deposited 130 - 200 000 BP in the North Sea, but must be considered preliminary. About 70% of a succession suggested by Vorren et al. (in prep.) to have been deposited during the last 0.8 Ma seems therefore to be less than 200 000 years old. Vorren et al. (in prep.) evaluate the source area for these sediments, and

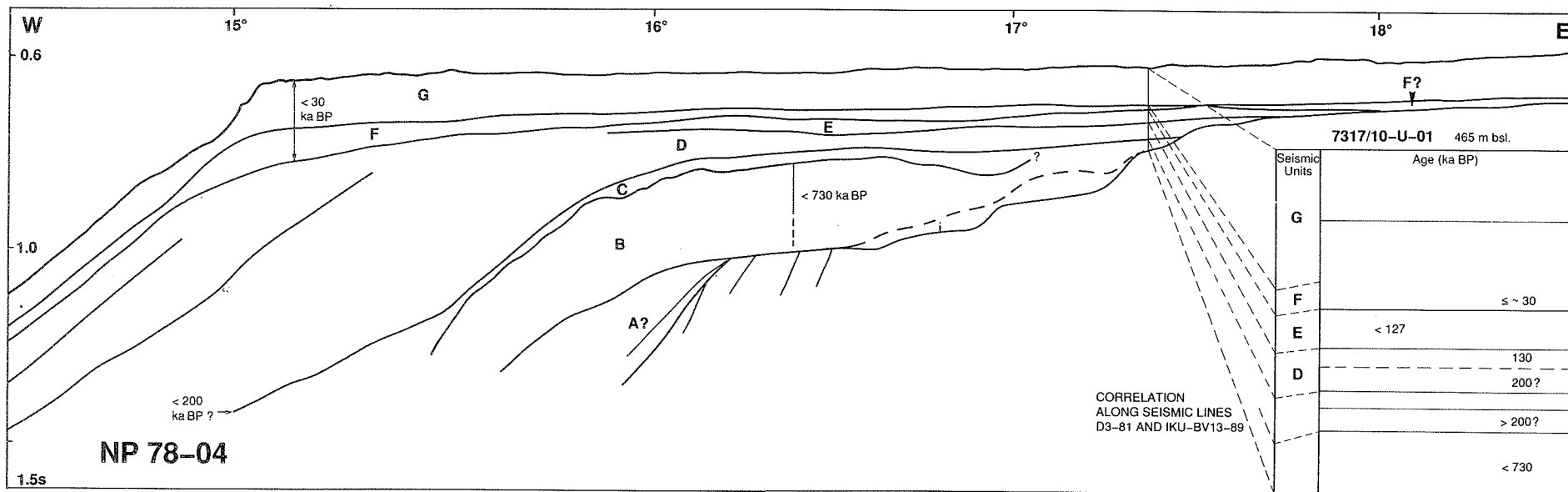


Figure 4.2.1 Interpretation of a seismic line at 73°N. Seismic units are correlated to site 7317/10-U-01. Ages of the seismic units are indicated. The results are preliminary.

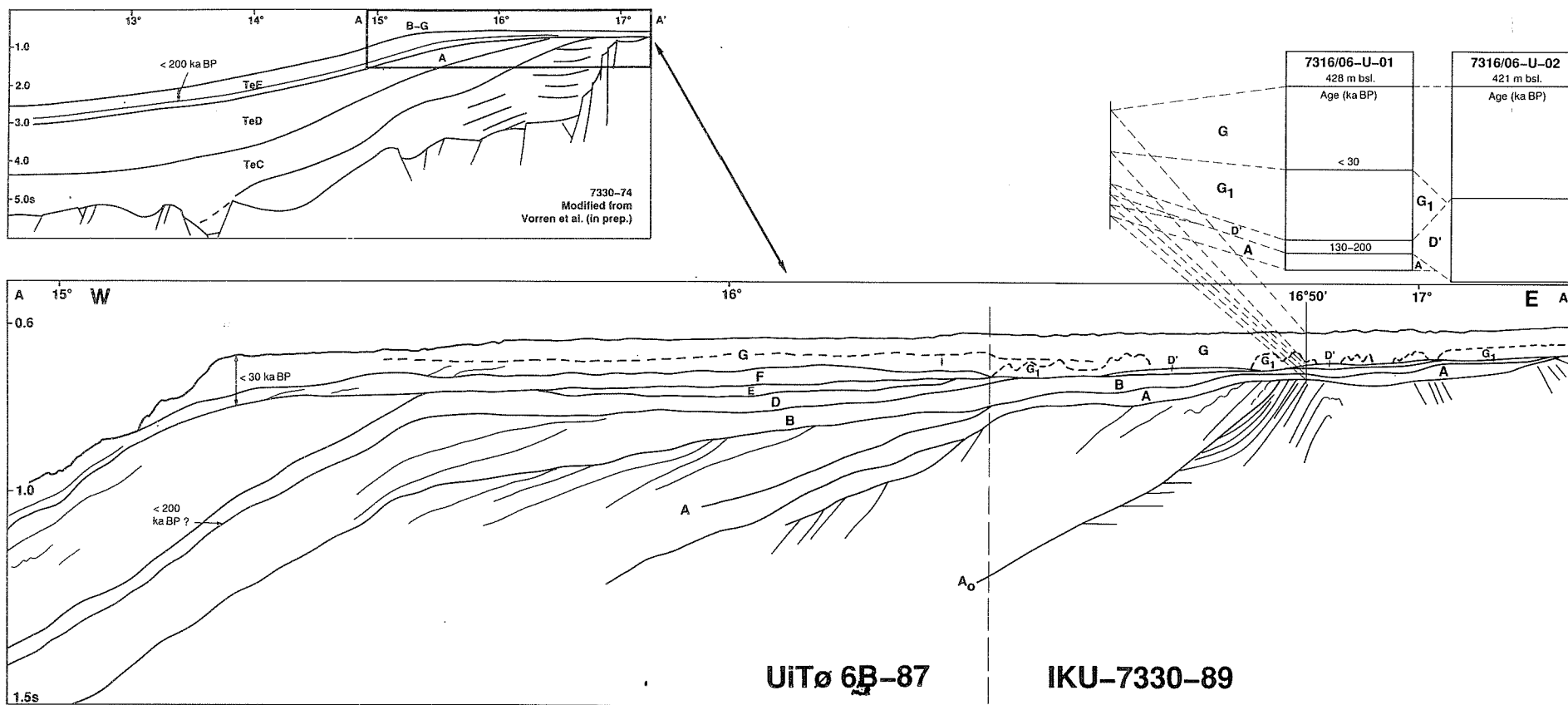


Figure 4.2.2

Interpretation of seismic lines at 73°30'N. Seismic units are correlated to the stratigraphy in boreholes 7316/06-U-01 and -02. The results are preliminary. The seismostratigraphy of Vorren et al. (in prep.) is included in the top of the figure, supplied with our age dating.

if we accept their assumptions of source and sink areas and assume our revised age estimates for the upper 70% of this succession, this gives an average erosion rate of 0.5 km/Ma during the last 0.2 Ma. Our results further show that 50 - 100 m of the sediments have been deposited on the upper slope since 30 000 BP. Most of this deposition occurred prior to 13 000 BP, thus implying a net erosion rate of 1.2 km/Ma during the last stadial (17 000 years).

The seismostratigraphy of Vorren et al. (in prep.) combined with the recent results of Eidvin and Riis (1989) seem to imply that unit A₀ is of Pliocene age, and that glacial sediments rest on the base unit A₀ unconformity at the Senja Ridge. Therefore, a glacial origin for this unconformity cannot be ruled out, although its level form might rather suggest that the erosion was controlled by sea level. (Note, however, that a glacial influence is indicated for some of the clasts found in core 7316/03-U-01, stratigraphically below this unconformity.)

North of about 73°25'N this unit has a very conspicuous internal seismic reflection pattern (Figure 4.2.3), interpreted as megaripples formed in an offshore environment. Their presence and internal structure are considered indicative of deposition in a shelf setting characterized by high sediment supply. This is consistent with the inferred high erosional rates in the Late Cenozoic. Younger erosional channels within unit A₀ cut down into these structures.

4.2.3 Cenozoic glacial erosion rates

In Appendix A.13 a number of reported rates of glacial erosion are summarized. Our inferred rates of glacial erosion (net erosion rate of 1.2 km/Ma during a glaciation and 0.5 km/Ma averaged over 200 ka) are comparable to erosional rates at Kongsvegen, Svalbard estimated by Elverhøi et al. (1980), three times the rates in the Ice stream B catchment area in West Antarctica inferred by Alley et al. (1987), and higher than found for many Norwegian glaciers eroding metamorphic rocks (Kjeldsen 1981). A number of studies, however, suggest values above 2 km/Ma, and two results from Alaska conclude with erosion rates as high as 19 km/Ma and 30 km/Ma.

Soft sediments with low erosion resistance were probably exposed during the early part of the glacial erosion in the Barents Sea, and the water may

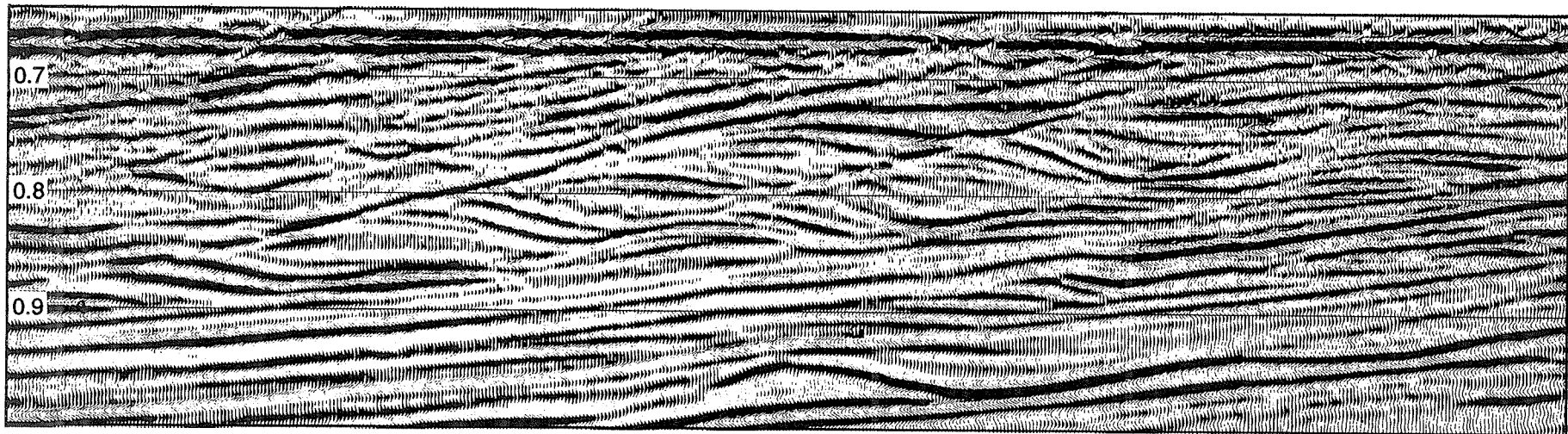
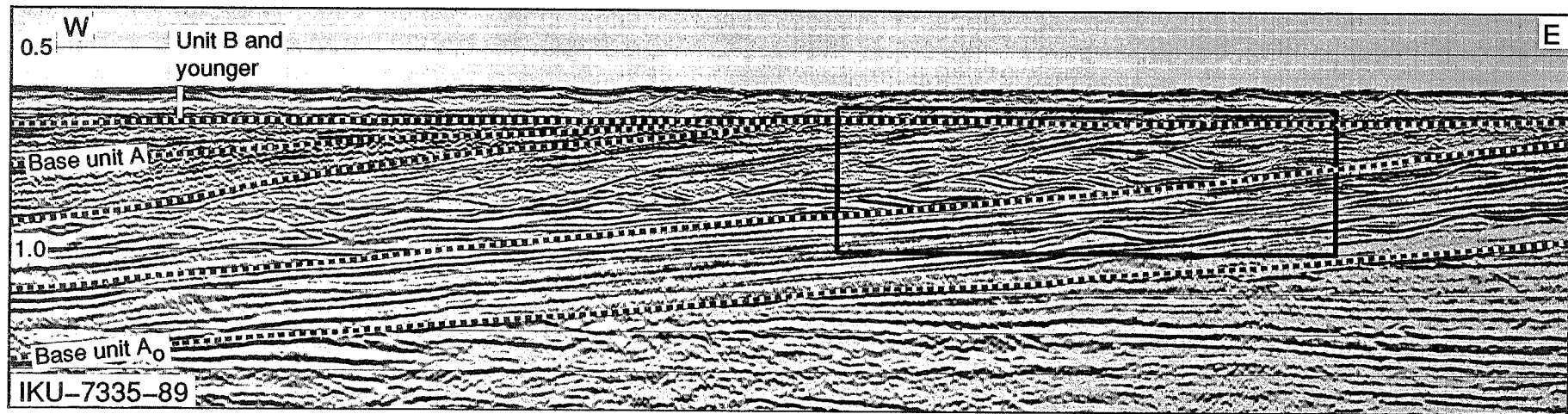


Figure 4.2.3 Internal reflection pattern of unit A₀ (corresponding to unit TeC of Vorren et al. (in prep.)). The climbing ripple pattern in the central part of the unit is interpreted as megaripples (?tidal sand ridges) deposited in an offshore environment characterized by high sediment supply.

have been significantly shallower. Glacial erosion at a rate of 1.2 km/Ma may therefore be considered reasonable, and probably a conservative estimate. In contrast to glacial erosion, subaerial erosion in the Arctic is at present two order of magnitudes lower. Holocene erosion rates in Linnédalen, Spitsbergen (mainly subaerially exposed) are found to be about 15 m/Ma (Svendsen et al. 1989).

Vorren et al. (in prep.) consider from volumetric considerations that 1.2 km has been eroded in the southern Barents Sea during the last 30 Ma. However, a combination of their volume estimates and source area assumptions, together with the recent results on the stratigraphy of the clastic fan in front of Bjørnøyrenna (Eidvin & Riis 1989), imply an average erosion in the southern Barents Sea of about 0.8 km since about 2.5 Ma. This is reasonably in line with rates of glacial erosion reviewed above, but seems too high to be explained by subaerial erosion (Svendsen et al. 1989). This implies that much of the Late Cenozoic erosion in the Barents Sea may have been glacial. Glacial erosion has clearly reached down to more than 500 m in the Barents Sea during the last glaciation (30 - 13 ka ago), and is, for instance, in Antarctica known to reach more than 1 km below sea level at present.

It may therefore be concluded that part of the Late Cenozoic erosion in the Barents Sea and the present bathymetry of the southern Barents Sea is not a result of uplift, erosion and subsequent subsidence, as classically assumed, but to a great extent due to glacial erosion. Uplift due to isostatic rebound must have occurred as a secondary effect of this erosion.

4.3 Evolution of the Bjørnøya West area

The Barents shelf has experienced a complex Cenozoic history with significant episodes of uplift, erosion and sediment deposition. Most of this history is reflected in the Bjørnøya West area because of its location in the main depositional area and its neighbourhood to the Stappen High which represents a key area for uplift and erosion during the Cenozoic.

We feel that the data and knowledge obtained during the present project therefore will contribute to an improved understanding of the Cenozoic history, and below we present a summary of this history, highlighted on the core data.

4.3.1 The pre-rift phase

By the close of the Cretaceous, the northern margin of the Barents platform was uplifted, and exposed Mesozoic sediments were truncated. Ziegler (1988) suggested that this uplift may have been associated with rifting of the Lomonosov Ridge, which separated from the Barents-Kara Sea Platform at this time. There was also local to subregional truncation and planation of faulted Cretaceous sediments along the Greenland-Barents Sea boundary. Apatite fission track analyses of our cores support this and suggest an erosion of the Stappen High in late Cretaceous (80-100 Ma).

Paleocene sedimentation over much of the western Barents Sea region occurred under conditions of broad subsidence on an epicontinental sea, and a thick Tertiary pre-rift sequence (Paleocene-Early Eocene) is interpreted for the region west of the Stappen High. A thin pre-rift sequence (Paleocene-?early Eocene) is reported for the Bjørnøya Basin (Nøttvedt et al. (1988). Extensional or transtensional subsidence commenced on Spitsbergen in the early Paleocene prior to the initiation of seafloor spreading, and infilling of the Spitsbergen Central Basin began at this time (Nøttvedt et al. 1988).

The 81.4 m thick section of bedrock recovered in core 7317/10-U-01 represents stacked upwardly coarsening successions of marine clastics ranging from clayey siltstones to sandy clayey siltstones with storm beds and some bioturbation. Deposition in a regressive shelf environment is proposed for these sediments. Microfossils are poorly preserved, but

moderately-preserved dinoflagellate cysts are common and include redeposited Aptian-Cenomanian species associated with an in situ Late Paleocene (Selandian) flora. This cyst assemblage is remarkably similar to that reported for the 1.55 m of dark olive grey burrowed claystones of core C85-A (7126/06-U-01) on the Finnmark Platform, south of the Nordkapp Basin. The sediments in both of the these cores are consistent with a period of epicontinental deposition above the basal Tertiary unconformity.

Stratigraphically the sediments of core 7317/10-U-01 underlie the regional basalt of proposed early Eocene age (see Faleide et al. 1988) (Figure 4.1.4). The sediments were probably, at least in part, sourced from the Stappen High, which then probably had experienced the initial part of its first significant uplift. The sediments from this core seem to correlate with Unit I of Spencer et al. (1984), who defined their units in the Senja Ridge/Tromsø Basin region further south.

4.3.2 Early Eocene-Early Oligocene transpressional shear phase

Differential shear propagated northward during early Eocene to early Oligocene time, which lead to break-up of the earlier broad basins. Approximately 500 km of lateral displacement occurred along the Senja Fracture Zone and 15-20 km of compressive shortening occurred in the West Spitsbergen fold belt (Nøttvedt et al. 1988). the smaller eocene basins were infilled rapidly by clinoformal sediments from areas raised by local and regional transpression.

The 131 m thick section of bedrock recovered in core 7316/06-U-02 represents stacked coarsening upward successions of compacted and bioturbated silty claystones with sandstone beds, glauconite, clinoptilolite and up to 21% opal-CT. Diverse benthic and planktonic siliceous and calcareous microfossils are preserved in these sediments as well as a rich dinoflagellate cyst assemblage. Redeposited Cretaceous microfossils are common, and an age ranging from 47 Ma to 54 Ma (Early Eocene-earliest Middle Eocene) is indicated by the Midway-type calcareous foraminiferal fauna. A quiet inner shelf depositional environment is inferred for the core.

Core 7317/02-U-01 (D85-V) was located on the western flank of a synclinal trend close to the Stappen High (Figure 4.1.7). Foraminifera and other

inorganic microfossils are essentially absent in the 10 m thick section of poorly consolidated clayey and silty, very fine-grained sand, but the dino-flagellate cyst flora of this core includes many species in common with that of core 7316/06-U-02. The two cores are regarded as approximately coeval, although core 7317/02-U-01 was previously interpreted to be Middle Eocene in age. The coarser sediments in core 7317/02-U-01 indicate a more proximal position to the paleo-shoreline than for core 7316/06-U-02.

The extensive deposition of Eocene sediments recognized in seismic data together with reworking of Cretaceous sediments into our Eocene cores support that the Stappen High was significantly uplifted and eroded during the period. Sediment supply was probably accelerated along the active Knølegga Fault zone.

4.3.3 Late Oligocene-Pliocene post-rift phase

The present study area is located in a transition zone with respect to the opening of the Norwegian - Greenland Sea, where the seafloor spreading occurred in earliest Eocene to earliest Oligocene (anomalies 25/24 to 13) with transcurrent movement along the Senja Fracture Zone to the south of our area. At anomaly 13 time (earliest Oligocene) the relative direction of plate movement changed to a more W-NW trend and caused extension in the study area (e.g. Faleide et al. 1988).

After the change in plate movement in earliest Oligocene, the area subsided, but the depth at the base of the evolving continental slope is not known. Much of the sediments in cores 7316/06-U-01 and 7316/03-U-01 are interpreted as debris flow deposits, turbidites or channel fill deposits, with an overall high sedimentation rate. They were probably deposited on the upper slope, but without knowing the depth of the base of the evolving continental slope, we cannot say whether they were part of an upper fan environment. In the description of depositional environment we have therefore preferred the term "upper slope".

The polymict conglomerate and coarse- to very fine-grained sandstones in core 7316/06-U-01 probably formed the floor of discharge channel on the upper slope, and the presence of high porosity, moderate to good sorting, subangular rock fragments and secondary glauconite are consistent with rapid redeposition of debris derived from a local provenance. Sedimentation

rates may have been too rapid to support bottom communities, as no inorganic microfossils have been found. Some intervals of lithologic Unit A are barren or contain only reworked Paleocene and Cretaceous dinoflagellate cysts. Other horizons contain a very low diversity flora of long-ranging Oligocene-Miocene species in assemblages dominated by reworked components. Biostratigraphic analysis is always difficult under these conditions, but a Late Oligocene/Early Miocene age is proposed with modest confidence. The high frequencies of older taxa in this assemblage attest to rapid redeposition, and the supply rate of erosional clastics must have been far in excess of the flux of planktonic debris from the overlying surface water. The paucity and low diversity of the in situ dinoflagellate cyst flora suggest that the oceanic conditions of this region had already deteriorated to a possibly pre-glacial state by the Late Oligocene/Early Miocene, far in advance of the Late Miocene decline evident in sediments from the Vøring Plateau.

The stratigraphic separation between cores 7316/06-U-01 and 7316/06-U-02 is estimated to be approximately 180 m, and the prominent reflector (Green of Chapter 4.1) within the interval is interpreted to represent truncation of at least 400 m. Dating of the two cores suggests a Middle Eocene to Late Oligocene/Early Miocene time interval between the cores. Core -02 contained reworked marine palynomorphs from Early Oligocene, indicating that such sediments were deposited in the area. Considering the short interval left up to the unconformity, we believe that they have been eroded. This could imply that the unconformity here represents a gap of missing sediments from Middle Eocene to Late Oligocene/Early Miocene, and that the depositional hiatus was from Early Oligocene to Late Oligocene/Early Miocene.

The 160.8 m of sediment recovered in core 7316/03-U-01 record a compound history that will be discussed in four parts: fan/upper slope sedimentation, volcanicity, siliceous ooze deposition, and glaciation. The bulk of the sediments in this core (subunits A, B, and F) represent rapid fan/upper slope deposition similar to that of core 7316/06-U-01. Overall, the section represents an upwardly coarsening succession of very fine- to medium-grained sandstones with interbedded finer grained units. The orange reflector (Chapter 4.1) is interpreted to represent the base of Subunit B, and this sequence may represent an interval of channel fill/debris flow sedimentation.

Cretaceous and Paleocene microfossils and palynomorphs occur throughout this core, although the Middle Eocene fossils constitute the majority of the redeposited assemblages. Changes in provenance are indicated by older reworked dinoflagellate cyst floras in the subunits. Subunits C, D and E include only 12 of the 24 Cretaceous dinocysts and only 41 of the 85 Paleocene dinocysts occurring in the subunits A, B and F, and the former subunits include three Cretaceous dinocysts and 9 Paleocene dinocysts that are not present in subunits A, B or F. An age range of 5-11 Ma is indicated for the in situ microfossils and palynomorphs throughout the core, but the entire section could have been deposited within the interval 6-8 Ma (Late Miocene). As was noted in core 7316/06-U-01, the flux of redeposited clastics must have far exceeded the supply of contemporary planktonic debris, and low productivity, preglacial conditions persisted in this region, as evidenced by the paucity and low diversity of the fossil record of Late Miocene planktonic biota.

Approximately 20 pyroclastic beds are present in subunits B to F, ranging from thin tuffs to lapilli rich beds in subunits D and F. The olivine basalts present in this core are subalkaline to transitional alkaline, with high K, Na, and Si concentrations, and differ in composition from tholeiitic mid-ocean ridge and alkaline off-rift basalts. They exhibit strongest compositional similarities to the Late Tertiary basalts of North Spitsbergen. Although Faleide et al. (1988) found evidence of younger volcanic cones on the marginal high approximately 30 km to the west of core 7316/03-U-01, a more proximal source ca. 15 km to the southeast on the uplift east of the Central Fault Zone is suggested for these pyroclastics. Nevertheless, it is not unreasonable to assume that the late volcanism reported by Faleide et al. (1988) and the volcanism recorded in this core occurred during the same phase of igneous activity. If true, then the young volcanic cones on the marginal high are Late Miocene in age, rather than Oligocene as proposed by Faleide et al. (1988), and the sediments that drape them are Miocene or younger.

During the Late Miocene episode of igneous activity in the Bjørnøya West region, another event occurred that is recorded in core 7316/03-U-01. The lapilli beds in Subunit D are overlain and underlain by sediments (subunits C and E) that differ remarkably in character from the surrounding sediments. A large proportion of the silty clays and claystones of these subunits are composed of Middle Eocene siliceous ooze. The age and

composition of these sediments document their extraneous nature. They were probably redeposited by turbidity and debris flow processes from a nearby source on the adjacent shelf, and we suggest that minor reactivation along the Knølegga Fault Zone during the Late Miocene exposed a newly formed subcrop to rapid erosion.

Shallow burial depths for provenance of subunits C and F are documented by the fresh preservation of the Eocene siliceous microfossils, which retain the opal-A phase, although redeposition of older material is also indicated by the presence of Cretaceous and Paleocene dinocysts and cristobalite pseudomorphs of radiolaria. Erosion of this soft sediment subcrop probably proceeded rapidly, and the debris so derived must have dominated the fine-grained sedimentation of the nearby slope region upon which core 7316/03-U-01 is sited during subunits C and E time.

Among the sediments recovered in our Bjørnøya West shallow drillings, these redeposited sediments are our only record of ?proven Middle Eocene deposition on the Barents Shelf, but it is apparent that the Eocene paleoceanography of the shelf regions must have differed markedly from that of the adjacent slope. The Barents Shelf must have been the site of extremely high organic production during this time, and the siliceous oozes that accumulated under these conditions are closely similar to those of the Vøring Plateau. Other siliceous oozes of Late Cretaceous and Eocene age are reported for the Alpha Ridge of the Arctic Ocean (Sites 422 and 437 of Clark 1974 and 1981), and plate reconstructions indicate that this ridge and the Lomonosov Ridge were in juxtaposition with the Barents-Kara Platform during this time. Paleogene siliceous oozes have been reported at many localities in the Pechora Basin and on the Russian Platform (Locker and Martini 1987). Thus, it appears that the seas over this vast shelf region probably underwent intense coastal upwelling during ?Cretaceous to at least Middle Eocene time under conditions similar to the California margin during Middle Miocene time. Anoxic bottoms are common under selected zones of coastal upwelling regions, and the accumulation of organic matter in such zones led to formation of hydrocarbons in the Monterey Formation of California. Such a play concept should not be ignored during future exploration in the Barents Sea.

A glacial origin has been suggested for faint striations and abrasion pits on some of the pebbles dispersed throughout the core. Although the type and

degree of glaciation is presently unknown, cold ocean conditions for the Barents Sea region beginning as early as the Late Oligocene have been postulated on the basis of the biostratigraphy of cores 7316/06-U-01 and 7316/03-U-01, and formation of glaciers in the highlands is not unreasonable. If substantiated, these findings suggest that glacial transport may have contributed to and progressively accelerated erosion of the Barents shelf during the interval of intense build-up of the Tertiary wedge (see also Chapter 4.3).

4.4 Burial depth

Much effort has been used in trying to resolve the magnitude and timing of the many uplift and erosional events during the Late Cretaceous and Cenozoic of the Barents Shelf.

Our shallow drillings of the Tertiary sediments from the western Barents Shelf provides an opportunity in trying to gain data from both timing and the extent of the different erosional episodes. We have applied mineralogy/diagenesis, organic geochemistry, seismic data, apatite fission track analysis and oedometer tests in our study to estimate the maximum burial depth/temperature of the sediments.

Mineralogy and diagenesis

Petrographic analyses of the cored rocks in the Bjørnøya West area have been carried out by X-ray diffraction analysis and thin section studies using both optical microscope and SEM/electron microprobe. In addition, bulk chemical analysis of volcanic glass has been performed. This discussion will focus on temperature sensitive minerals/rocks and diagenesis.

The petrographic analysis of the sediments has demonstrated the existence of opal-CT, clinoptilolite, smectite and illite-smectite mixed layer clay, mixed layer clay, and alkali feldspar cements. The optical and chemical studies of volcanics in core 7316/03-U-01 have also shown that basaltic glass has been excellently preserved from diagenesis.

The presence of opal-CT has been documented by X-ray diffraction analysis in cores 7316/06-U-02 and 7316/03-U-01. Opal cement has also been found in association with volcanic material in core 7316/03-U-01. In addition, Paleogene siliceous microfossils (opal-A type) has been documented in the same core; these microfossils may be derived from sediments previously subjected only to shallow burial. The Cretaceous - Paleogene, cristobalite replaced Radiolaria in core 7316/03-U-01 probably resulted from an older phase of redeposition that occurred during accumulation of the parent sediments from which subunits C and E are derived. The alteration of opal-A to opal-CT in general requires temperatures of 40-50°C. Bjørlykke (1989) indicated temperatures of 60-70°C for transformation of opal-CT to quartz.

Clinoptilolite is present in core 7316/03-U-01 as pore lining cement associated with basaltic glass and opal-CT. (The relative timing of the diagenetic reactions has not been established). In cores 7316/06-U-01 and 7316/06-U-02 the mineral is only documented by X-ray diffraction analysis. Clinoptilolite may form as a late reaction product during diagenesis by alteration of volcanic glass, but the most common occurrence is in deep marine sediments. Its formation is a time-dependent process, and may proceed via metastably formed phillipsite (Singer & Müller 1983).

Alteration (palagonitization) of basaltic glass proceeds rapidly at temperatures above 50°C (Fisher & Schmincke 1984). During low temperature diagenesis both pore water chemistry and flow rate are controlling factors on diagenetic hydration of the glass, while temperature and burial depths and time are far less important (Davies et al. 1979).

Smectite has been identified in core 7316/03-U-01, while mixed layer clay minerals have been identified in all cores. Smectite will be replaced by disordered illite-smectite mixed layer clays if a source of potassium is present. The smectite diagenesis is sensitive to the temperature and smectite will start developing to illite-smectite mixed layer clays at temperatures of 65-70°C.

To conclude, the mineralogical analyses indicate that the cores could only have been subjected to maximum temperatures of 40-70°C.

Apatite fission track results

Apatite fission track analyses have been performed on two samples from core 7316/03-U-01 and one sample from each of the cores 7316/06-U-01, 7316/06-U-02 and 7317/10-U-01 at the Geotrack International laboratory (see enclosed report, Appendix A.8). It was concluded that none of the samples show evidence of having been subjected to significantly paleotemperatures at any time since deposition. The mean or pooled fission track ages (Table 1 in Appendix A.8) record provenance history rather than the latest burial history. The data suggest that the temperatures were never sufficiently high to induce fission track shortening after deposition. Due to the limitations of the apatite fission track method at low temperatures (10-50 °C range), only burial temperatures exceeding 50°C can be given by this method. In the present case it can be concluded that the maximum burial temperatures did not exceed this limit of 50°C.

Thermal maturation from organic geochemical parameters

The extensive reworking of the organic matter indicated by the palynology of these cores created inherent uncertainties in using data obtained from the kerogen to make deductions regarding the post-depositional thermal history of the cored successions. It should also be remembered that the observed organic geochemical maturation indices could reflect the thermal maturation of the source area if burial of the redeposited material has not been great enough.

Rock-Eval Tmax values show the greatest variation within core 7316/03-U-01, but are within the uncertainties of the method in the other three cores, i.e. $\pm 5^{\circ}\text{C}$. Tmax values are similar in the three northern cores, 7316/03-U-01, 7316/06-U-01 and 7316/06-U-02, i.e. 420°C . In contrast, the Tmax values from the southernmost core, 7317/10-U-01, are significantly higher with an average value of 430°C . Vitrinite reflectance data also suggest that the most mature kerogen occurs in core 7317/10-U-01, although the difference is not as pronounced as that observed in the Tmax data. Instead, mean vitrinite reflectance values for cores 7316/03-U-01, 7316/06-U-02 and 7317/10-U-01 shows a southwards increase from ca. 0.36% to ca. 0.40%. The poor quantity and quality of the kerogen in core 7316/06-U-01 prevented usable vitrinite reflectance data being obtained from this core.

These trends are consistent with the increasing age of the cores from north to south. This in turn may be taken to suggest that burial of the reworked kerogen has exceeded that of the source area and that the maturation data reflect the burial history of the sampled successions.

Gas chromatography and GC-MS data obtained from the extractable organic matter (EOM) in these cores could be interpreted as indicating a low level of thermal maturity. However, the lack of a possible source rock within the cored succession makes it hazardous to use these data regarding the burial history of the rocks.

The available organic maturation data from these rocks clearly indicate that the kerogen present is thermally immature. Comparison with other areas on the Barents shelf would suggest that burial of the cored succession has been in order of 1 to 1.5 kilometres, the southernmost core having been more deeply buried than the three northern cores. It is also possible to say, based on the reasoning outlined above, that the source area of the cored rocks had not experienced much burial prior to erosion.

Oedometer test

Compressional tests (oedometer tests) have been carried out in order to quantify the maximum previous load on the sediment.

The SINTEF Division of Geotechnical Engineering has performed a total of 5 oedometer test on sedimentary rocks from cores 7317/10-U-01, 7316/06-U-02 and 7316/03-U-01. The main conclusions are:

- Two samples (207 m and 147 m) of siltstones and silty shales from core 7317/10-U-01 show apparent preconsolidation stresses of 35-40 MPa. Disregarding possible cementation effects this corresponds to 3.5-4 km of burial. The deformation behavior indicates, however, that cementation is important and may have contributed considerably to the high values.
- Two samples (138 m and 181 m) of siltstones and silty shales from core 7316/06-U-02 show an apparent preconsolidation stress level of 14-20 MPa, indicating a maximum previous sediment overburden of 1.4-2 km.
- The sandstone sample from well 7316/03-U-01 shows an apparent preconsolidation stress slightly in excess of the present effective stress.

There is reason to believe that cementation between grains and possible secondary pore filling influence the measured values. Secondary electron images from the analysed samples and from parallel reference sample have therefore been studied. From the siltstones and silty shales of core 7317/10-U-01 and 7316/06-U-02 cementating effects could not be detected by SEM. From the sandstone in core 7316/03-U-01, needleshaped clinoptilolite crystal were sporadically encountered as pore lining cements (Figure 3.3.9). However, the small amounts of clinoptilolite cement probably played a minor role in increasing the measured preconsolidation stress level.

As expected, these relationships are quite complicated and should be studied further. Before this has been done, we can only conclude that the measured values represent maximum values for burial depth. Although the measured values for core 7317/10-U-01 are unrealistically high, the measured trend of the five analyses seems, however, to be reasonable, with a minimum of burial of core 7316/03-U-01, intermediate values for core 7316/06-U-02 and highest values for core 7317/10-U-01.

Seismic evidence of uplift and erosion

The seismic data in the study area show evidence of several phases of uplift and erosion. The first, possible Eocene phase, can be connected to the beginning of the rifting and opening of the Norwegian-Greenland Sea. This phase separated the Stappen High from the areas further west by several large boundary faults, as can be seen on for instance the NPD line BV-09-86, sp 1100, approximately 10 km to the northwest of well 7317/10-U-01. Signs of heavy erosion of the fault plane indicate that this fault and the area to the east were exposed in Eocene time, while a basin with approximately 2 km of sediments developed to the west. This also confirms that the well 7317/10-U-01 is located in a different sedimentary environment than the other wells, and illustrates the difficulties of tying this well by seismic data to the west.

The relative erosion in the Oligocene represented by the hiatus between cores 7316/06-U-02 and -01 ("Green reflector time") can be measured to about 400 m over a distance of 25 km (along line 163730, from 73°22'N to 73°35'N). It is not possible from seismic data to identify any area that was particularly uplifted by tectonic movements, and this erosion may as well have been caused by sea level changes with a general tilt in a NS direction. Below the Miocene/Pliocene unconformities, as described in Section 4.1, including the light purple "Base of Pliocene fan unconformity", erosion of sediments on a high to the west of the 1989 drill sites has occurred. On Figure 4.1.2 we can see that the reflections become more or less concordant approaching the Central Fault Zone (Figure 4.1.7), thus indicating very little erosion at drill sites 7316/06-U-01 and -02 associated with these unconformities. However, the seismic data give evidence that the Central Fault Zone has been active during Miocene, and the area between these faults and the Knølegga Fault Zone has been uplifted. The two wells will then be situated in an intermediate position within the Central Fault Zone.

The conclusion must then be that the main erosion experienced in the area of wells 7316/06-U-01 and -02 must have occurred in post-Mid Pliocene time (later than 2.5 Ma). The magnitude of this erosion can be estimated to approximately 1 km, depending on how much of the Pliocene fan that was deposited at the two sites.

Conclusions

Because of the difference in structural setting and stratigraphic position it would be reasonable if variations in maximum burial depth were found in the four cores. Figure 4.4.1 summarizes the results of the different methods used in evaluation of maximum burial.

Some of the analyses are based on temperature reactions, and conversion into burial depth should therefore rely on present and paleo temperature gradients for each site. Keeping in mind the Miocene volcanic activity in the area and the probability for local variations in heat flow, there are obvious sources for errors in conversion from temperature to depth. However, the most important limitations in this case are the scarcity of temperature sensitive mineral reactions in the most relevant temperature interval ($<40-70^{\circ}\text{C}$), and the strong kinetic controls on those that are present. We have applied a geothermal gradient of $37^{\circ}\text{C}/\text{km}$ and a surface temperature of 2°C measured earlier in the area (Sættem 1985).

Organic geochemistry and oedometer tests and partly the mineralogy/diagenesis results indicate a deeper maximum burial for core 7317/10-U-01 than for the other three. There is also a slight indication that the two youngest cores (7316/06-U-01 and 7316/03-U-01) have experienced less burial than core 7316/06-U-02. These trends are, however, vague. For the exact depth of burial, the results seem to concentrate around 1 - 1.5 km, but because most analyses cannot give a minimum value, we would suggest that the most probable value is closer to 1 km but would not exclude even less for the areas of the three northernmost drill sites.

From discussions elsewhere in this report we may conclude that a major uplift of the study area occurred in Eocene, possibly starting in the Late Cretaceous. A second erosional event occurred in the Oligocene. This is reflected by the unconformity between cores 7316/06-U-01 and -02 ("Green reflector"), and from our data in the Bjørnøya West area it seems to represent a general relative lowering of sealevel rather than a tectonic event.

A Late Miocene uplift was associated with crustal deformation, volcanism and redeposition of Eocene, Early Oligocene and possibly also early Miocene deposits. The extent of erosion is hard to quantify and could only be of local importance.

Our data indicate that the post-Mid Pliocene (later than 2.5 Ma) erosion was the most important Cenozoic erosional episode in the study area and probably also for large parts of the Barents Sea. Seismic data indicate an erosion of the order of 1 km in the area of cores 7316/06-U-02 and 7316/06-U-01.

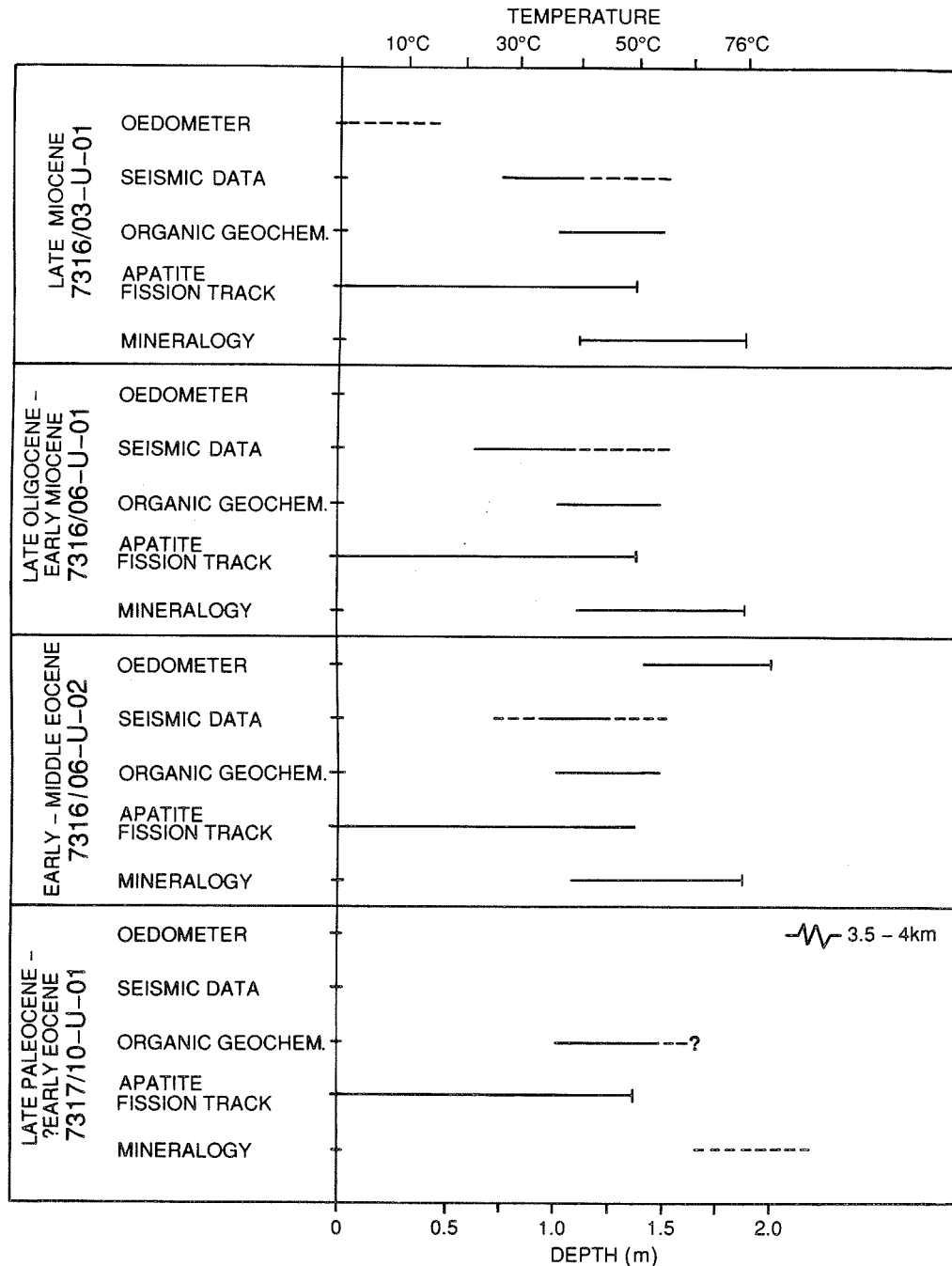


Figure 4.4.1 Depth of burial estimated from different analysis of the four cores. The seismic data and oedometer tests give depth estimates, while the mineralogy/diagenesis, the apatite fission track analysis and the organic geochemical data are temperature indicators. Based on a heat flow measurement in the area, temperature has been converted to depth using a temperature gradient of 37°C/km and a surface temperature of 2°C.

5. REFERENCES

- AKSU, A.E., 1985: Planktonic foraminifera and oxygen isotope stratigraphy of CESAR cores 102 and 103; preliminary results. In Jackson, H.R., Mudie, P.J. and Blasco, S.M. (Eds.) Initial geologic report on CESAR-the Canadian expedition to study the Alpha Ridge, Arctic Ocean. Geological Survey of Canada. Paper 84-22, 115-124.
- AKSU, A.E. and MUDIE, P.J., 1985: Magnetic stratigraphy and palynology demonstrate at least 4 million years of Arctic Ocean sedimentation. *Nature*, volume 318, 280-283.
- ALLEY, R.B., BLANKENSHIP, D.D., BENTLEY, C.R. and ROONEY, S.T., 1987: Till beneath ice stream B 3. Till deformation: evidence and implications. *Journal of Geophysical Research*, vol. 92, no. B9, 8921-8929.
- AMUNDSEN, H.E.F., GRIFFIN, W.L. and O'REILLY, S.Y., 1987: The lower crust and upper mantle beneath northwestern Spitsbergen: evidence from xenoliths and geophysics. *Tectonophysics* 139, 169-185.
- ARTHUR, M., SHRIVASTAVA, S., AKSU, A., BALDAUF, J., BOHRMAN, G., BUSCH, W., CEDERBERG, T., CLEMENT, B., CREMER, M., DADEY, K., De VERNAL, A., FIRTH, J., HALL, F., HEAD, M., HISCOLT, R., JARRARD, R., KAMINSKY, M., LAZARUS, D., MONJANEL, A.-L., NIELSEN, O., STEIN, R., THIEBAULT, F., ZACHOS, J. and ZIMMERMANN, H., 1986: Ocean Drilling Program. High-latitude paleoceanography. *Nature* volume 320, 17-18.
- BETTENSTAEDT, F., 1962: Diatomeae. In: Bartenstein, H., et al. *Leit Fossilien der mikropaläontologie*, Gebrüder Borntraeger (Berlin). 357-358.
- BJORØY, M., BUE, B. and ELVSBORG, A., 1983: Organic geochemical analysis of the first two wells in the Troms I area (Barents Sea). In: M. Bjarøy et al (eds.), *Advances in Organic Geochemistry 1981*, pp 16-27.
- BJØRKLUND, K.R., 1976: Radiolaria from the Norwegian Sea, Leg 38 of the Deep Sea Drilling Project. In: Talwani, M., Udintsev, G., et al. *Initial Reports of the Deep Sea Drilling Project*, vol. 38. Washington (US Gov. Printing Office) pp. 1101-1168.
- BJØRLYKKE, K., 1989: *Sedimentology and Petroleum Geology*. Springer Verlag. pp 363.
- BOUMA, A.H., BERRYHILL, H.L., KNEBEL, H.J. and BRENNER, R.L., 1982: Continental shelf. In: Scholle, P.A. and Spearing, D., (eds): *Sandstone depositional environments*. American Association of Petroleum geology, Memoir 31, 281-328.
- BUGGE, T., 1983: Submarine slides on the Norwegian continental margin, with special emphasis on the Storegga area. *IKU Publ.* 110, 152 pp.
- BUJAK, J.P., DOWNIE, C., EATON, G.L. and WILLIAMS, G.L., 1980: Taxonomy of some Eocene dinoflagellate cyst species from southern England. In: Bujak, J.P., Downie, C., Eaton, G.L., Williams, G.L. (eds.), *Dinoflagellate cysts and acritarchs from the Eocene of southern England: Palaeont. Assoc., Spec. Papers Palaeont.*, 24: 26-36, 22 pls.
- CHERNOVA, L.P., 1981: Influence of mass balance and run-off on relief-forming activity of mountain glaciers. *Annals of Glaciology*, 2, 69-70.

CLARC, D.L., 1974: Late Mesozoic and early Cenozoic sediment cores from the Arctic Ocean. *Geology*, 2: 41-44.

CLARC, D.L., 1981: Geology and geophysics of the American Basin. In: Nairn, A.E.M., Churkin, M., Stehli, F.G., (eds.), *The ocean basins and margins*, 5. The Arctic Ocean, Plenum Press. 1-20.

CLARC, D.L., ANDREE, M., BROECKER, W.S., MIX, A.C., BONANI, G., HOFMAN, H.J., MORENZONI, E., NESSI, M., SUTER, M. and WÖLFELI, W., 1986: Arctic Ocean chronology confirmed by accelerator c-14 dating. *Geophysical research letter*.

CORBEL, J., 1962: *Neiges et glaciers*. Paris.

COSTA, L.I. and DOWNIE, C., 1979: Cenozoic dinocyst stratigraphy of sites 403 to 406 (Rockall Plateau), IPOD, Leg 48. In: Montadert, L. et al., (eds), *Init. Repts. D.S.D.P.* 48: 513-529, 3pls.

CURRY, R.R., 1966: Glaciations about 3 000 000 years ago in Sierra Nevada. *Science*, volume 154, 770-771.

DAM, A. ten 1944: Die stratigrafische Gliederung des niederländischen Paläozäns und Eozäns nach Foraminiferen. *Mededelingen Geologische Stichting C-V-3*, pp. 1-142.

DAVIES, D.K., ALMON, W.R., BONIS, S.B. and HUNTER, B.E., 1979: Deposition and diagenesis of Tertiary-Holocene volcanoclastics, Guatemala. *SEPM, Special Publication No. 26*, 281-306.

DOPPERT, J.W. CHR. and NEELE, N.G., 1983: Biostratigraphy of marine Paleogene deposits in the Netherlands and adjacent areas. *Mededelingen Rijk Geologische Dienst*. 37, 2, pp. 3-79.

DREWRY, D., 1986: *Glacial Geologic Processes*, 276 pp. Edward Arnold, London.

DZINORIDZE, R.N., JOUSÉ, A.P., KOROLEVA-GOLIKOVA, G.S., KOZLOVA, G.E., NAGAEVA, G.S., PETRUSHEVSKAYA, M.G. and STRELNIKOVA, N.I., 1978: Diatom and Radiolarian Cenozoic stratigraphy, Norwegian Basin. *DSDP Leg 38*. In: *Initial Report of the Deep Sea Drilling Project, Supplement to volumes XXXVIII and XXXIX and XLI*, pp. 289-428.

EATON, G.L., 1976: Dinoflagellate cysts from the Bracklesham beds (Eocene) of Isle of Wight, Southern England: *Bull. Br. Mus. nat. Hist. (Geol.)*, 26(6): 227-332, 21 pls.

EDWARDS, L.E., 1984: Miocene Dinocysts from Deep Sea Drilling Project Leg 81, Rockall Plateau, Eastern North Atlantic Ocean. In: Roberts, D.G. et al., (eds), *Init. Repts. D.S.D.P.*, 81: 581-594, 5 pls.

EIDVIN, T. and RIIS, F., 1989: Nye dateringer av de tre vestligste borehullene i Barentshavet. Resultater og konsekvenser for den Tertiære hevingen. *NPD -Contribution No. 27*, 44 pp, 9 figures, 5 plates.

EINARSSON, T., HOPKINS, D.M., and DOELL, R.R., 1967: The stratigraphy of Tjörnes, Northern Iceland, and the history of the Bering Land Bridge. In Hopkins, D.M. (Ed.), *The Barents Land Bridge*. Stanford University Press, 312-325.

ELDHOLM, O., FALEIDE, J.I. and MYHRE, A.M., 1987: Continent- Ocean transition at the western Barents Sea/Svalbard continental margin. *Geology* 15, 1118-1122.

ELVERHØI, A., LIESTØL, O. and NAGY, J., 1980: Glacial erosion, sedimentation and microfauna in the inner part of Kongsfjorden, Spitsbergen. *Norsk Polarinstitutt Skrifter* 172, 33-61.

ERIKSSON, J., 1980: Tjörnes, North Iceland: A bibliographical review of the Geological Research History. *Jökull*, 30, Jöklaránnsóknafelag Island, 1-9.

FALEIDE, J.I., MYHRE, A.M. and ELDHOLM, O., 1988: Early Tertiary volcanism at the western Barents Sea margin. In: Morton, A.C. and Parson, L.M. (eds): Early Tertiary volcanism and the Opening of the NE Atlantic. *Geol. Soc. Spec. Publication* 39, 135-146.

FAURE, G., 1986: Principles of Isotope Geology. Sec. edition. John Wiley and Sons, New York Chichester Brisbane Toronto Singapore, pp. 589.

FISHER, R.V., SCHMINCKE, H.-U., 1984: Pyroclastic rocks. Springer Verlag, Berlin Heidelberg New York Tokyo pp. 472.

GABRIELSEN, R.H., FÆRSETH, R.B., JENSEN, L.N., KALHEIM, J.E. and RIIS, F., in press: Structural elements of the Norwegian continental shelf - Part I: The Barents Sea region. To be printed in Norwegian Petroleum Directorate Bulletin.

GEROCK, S., NOVAK, W., 1984: Proposal of zonation for the Late Tithonian-Late Eocene, based upon arenaceous foraminifera from the outer Carpathians, Poland, Benthos '83; 2nd International Symposium Benthic Foraminifera, Pan and Bordeaux, 225-239.

GOLL, B., 1989: A Synthesis of Norwegian Sea biostratigraphies: ODP LEG 104 on the Vøring Plateau. In: Eldholm, O., Thiede, J. and Taylor, E. (eds.): Proceedings of the Ocean Drilling Program, Scientific Results, 104, 777-826.

GOLL, R.M. and BJØRKLUND, K.R., 1989: A new radiolarian biostratigraphy for the Neogene of the Norwegian Sea. ODP leg 104. In: Eldholm, O., et al. Proc. ODP, Sci. Results 104: College Station, Texas. (Ocean Drilling Program), 697-738.

HAGEN, H., 1988: Modelling of chemical processes in magma chambers: the evolution of a Tertiary olivine tholeiite sequence from Okstindane, NW Spitsbergen. Unpublished Cand. Scient. Thesis, University of Oslo, pp. 90.

HARLAND, R., 1982: Recent dinoflagellate cyst assemblages from the southern Barents Sea. *Palynology*, 6: 9-18.

HAQ, B.U., HARDENBOL, J. and VAIL, P.R., 1987: Chronology of fluctuating sea level since the Triassic. *Science* 35, 1156-1167.

HAY, R.L., 1981: Geology of zeolites in sedimentary rocks. In: Mumton, F.A. (Ed): Mineralogy and geology of natural zeolites. Review in geology, volum 4, 53-64.

HENRICH, R., 1988: Glacial/interglacial cycles in the Norwegian Sea: Sedimentology, paleoceanography, and evolution of Late Pliocene to Quaternary northern hemisphere climate. In Eldholm, O., Thiede, J. and Taylor, E. et al., 1988: Proceedings ODP, Scientific results, 104: College Station, TX (Ocean Drilling Program).

HENRICH, R., WOLF, T., BOHRMANN, G. and TIEDE, J., 1989: Cenozoic paleoclimatic and paleoceanographic changes in the northern hemisphere revealed by variability of coarse fraction composition in sediments from the Vøring plateau - ODP Leg 104 drill sites. In Eldholm, O., Thiede, J. and Taylor, E. et al., 1989: Proceedings ODP, Scientific results, 104: College Station, TX (Ocean Drilling Program), 189-232.

HERMAN, Y. and HOPKINS, D.M., 1980: Arctic oceanic climate in late Cenozoic time. *Science*, volume 209, 557-562.

HULSBOS, R.E., KROON, D., JANSEN, M.S.M. and VAN HINTE, J.E., 1986: Early Eocene Benthic Foraminifers from the Vøring Plateau, Norwegian Sea (DSDP 338), and their paleoenvironmental implications. Manuscript Geomarine Centre, Vrije Universiteit, Amsterdam.

IMSLAND, P., 1978: The Geology of the volcanic island Jan Mayen, Arctic Ocean, *Nordic Volcanol. Inst. Res. Rep.* 7813, pp 74.

IMSLAND, P., 1986: The volcanic eruption on Jan Mayen, January 1985: Interaction between a volcanic island and a fracture zone. *J. Volcanol. Geotherm. Res.* 28, 45-53.

IRVINE, T.N. and BARAGAR, W.R.A., 1971: A guide to chemical classification of the common volcanic rocks. *Canadian J. Earth Sci.* 8, 523-548.

JAKOBSSON, S.P., 1972: On the consolidation and palagonitization of the tephra of the Surtsey volcanic island, Iceland. *Surtsey Res. Progr. Paper* 6, 121-128.

JANSEN, E., BLEIL, U., HENRICH, R., KRINGSTAD, L. and SLETTEMARK, B., 1987: Climatic changes in the Norwegian Sea during the last 2.8 Ma. *Polar Research* 5 n.s.329-332.

JANSEN, E., BLEIL, U., HENRICH, R., KRINGSTAD, L. and SLETTEMARK, B., 1988: Paleoenvironmental changes in the Norwegian Sea and the northeast Atlantic during the last 2.8 m.y.: Deep Sea Drilling Project/Ocean Drilling Program sites 610, 642, 643 and 644. *Paleoceanography*, volume 3, No. 5, 563-581.

JANSEN, E., SJØHOLM, J., BLEIL, U. and ERICHSEN, J.A., in prep.: Neogene and Pleistocene glaciations in the northern hemisphere and late Miocene-Pliocene global ice volume fluctuations: Evidence from the Norwegian Sea.

KASTNER, M., KEENE, J.B. and GIESKES, J.B., 1977: Diagenesis of siliceous oozes-I. Chemical controls on the rate of Opal-A to opal-CT transformation - an experimental study. *Geochimica Cosmochimica Acta*, 41, 1041-1059.

KEEN, M., 1978: The Tertiary-Palaeozoene. In: Bate, R.H., Robinson, E., (ed) *A stratigraphical index of British Ostracoda*, *Geological Journal Spec. Issue* No. 8, p. 385-450.

KING, C., 1983: Cainozoic micropalaeontological biostratigraphy of the North Sea. *Rep. Inst. Geol. Sci.*, No. 82/7, 40 pp.

KIRSCH, H.J., 1983: Mineralogy and petrology of burial diagenesis (burial metamorphism) and incipient metamorphism in clastic rocks. In: Larsen, G. and Chilingar, G.V. (eds.): Diagenesis in sediments and sedimentary rocks, 2. Elsevier, 289-494.

KJELDSSEN, O. (ed.), 1981: Materialtransportundersøkelser i Norske Breelver 1980. Norges vassdrags og elektrisitetsvesen; Vassdragsdirektoratet Hydrologisk avdeling Rapport 1981 N14, 41 pp.

KUL'KOVA, I.A., 1988: Eocene plant microfossils of west Siberia. In: S.B. Shatsky (ed.): Plant microfossils and stratigraphy of Mesozoic and Cenozoic of Siberia, 25-36. Novosibirsk "Nomka", Siberian Branch, 212 pp.

LARSEN, G. and CHILINGAR, G.V., 1983: Diagenesis in sediments and sedimentary rocks, 2. Elsevier.

LIESTØL, O., 1967: Storbreen glacier in Jotunheimen, Norway. Norsk Polarinstitutt Skrifter 141, 63 pp.

LOCKER, S. and MARTINI, E., 1987: Silicoflagellaten aus einigen russischen Paläogen-Vorkommen. Senckenbergiana lethaea 68: 21-67.

MANGERUD, J. and SVENDSEN, J.I., 1990: Hvor lenge lå isen over Svalbard under siste istid? (Abstract). Geonytt, Årgang 17, No. 1, p. 77.

MANUM, S.B., 1976: Dinocysts in Tertiary Norwegian-Greenland sea sediments (Deep Sea Drilling Project Leg 38), with observations on palynomorphs and palynodebris in relation to environment. In: Talwani, M., et al. (eds), Init. Repts. D.S.D.P., 38: 897-919, 6 pls.

MANUM, S.B., BOULTER, M.C., GUNNARSDOTTIR, H., RANGNES, K. and SCHOLZE, A., 1989: Eocene to Miocene palynology of the Norwegian Sea (ODP leg 104). In: Eldholm, O., Thiede, J., Taylor, E. et al. (eds.). Proceedings of the Ocean Drilling Program, Scientific Results, vol. 104, 611-662.

MARTINI, E. and MÜLLER, C., 1976: Eocene to Pleistocene silicoflagellates from the Norwegian-Greenland Sea (DSBP leg 38). In: Talwani, M., Udintsev, G., et al., Initial Reports of the Deep Sea Drilling Project, vol. 38. Washington (US Gov. Printing Office) pp. 857-896.

MATSUOKA, K. and BUJAK, J.P., 1988: Cenozoic dinoflagellate cysts from the Navarin Basin, Norton Sound and St. George Basin, Bering Sea. Bulletin of the Faculty of Liberal Arts, Nagasaki University, Natural Sciences, vol. 29, pp. 1-147.

MATTHEWS, R.K., 1984: Oxygen isotope record of ice volume history: 100 Million Years of glacio-eustatic sea level fluctuation. In: Schlee, J.S., 1984: Interregional unconformities and hydrocarbon accumulation. AAPG Memoir 36, pp. 97-107.

Mc DOUGALL, I. and WENSIK, H., 1966: Paleomagnetism and geochronology of the Plio-Pleistocene lavas in Iceland. Earth planet Scientific Letters, 1, 232-236.

MIYASHIRO, A., 1978: Nature of alkalic volcanic rock series. Contrib Mineral Petrol 66, 91-104.

MURÉ, E., 1987: Frigg. In: A.M. Spencer et al. (eds), Geology of Norwegian Oil and Gas Fields, pp 203-213.

MURRAY, J.W., CURRY, D., HAYNES, J.R. and KING, C., 1981: Paleogene. In: Jenkins, D.G., Murray, J.W., (eds) Stratigraphic atlas of fossil foraminifera, pp. 228-267.

NEUMANN, E.-R. and SCHILLING, J.-G., 1984: Petrology of basalts from the Mohns-Knipovich Ridge; the Norwegian-Greenland Sea. Contrib Mineral Petrol 85, 209-223.

NØTTVEDT, A., BERGLUND, L.T., RASMUSSEN, E. and STEEL, R.J., 1988: Some aspects of Tertiary tectonics and sedimentation along the western Barents Shelf. In: Morton, A.C. and Parson, L.M. (eds.), Early Tertiary Volcanism and the Opening of the NE Atlantic, Geol. Soc. Spec. Publ. 39, pp. 421-425.

ODIN, G.S. and FULLAGAR, P.D., 1988: Geological significance of the glaucony facies. In: Odin, G.S. (ed.): Green marine clays. Developement in sedimentology, 45, 295-332.

OERLEMANS, J., 1984: Numerical experiments on large-scale glacial erosion. Zeitschrift for Gletscherkunde und Glasiologie. Band 20, 107-126.

OSKARSSON, N., SIGVALDASON, G.E. and STEINTHORSSON, 1982: A Dynamic Model of Rift Zone Petrogenesis and the Regional Petrology of Iceland. J. Petrol. 23, 28-74.

PETTIJOHN, F.T., POTTER, P.E. and SIEVER, R., 1973: Sand and sandstone. Springer-Verlag, New York Heidelberg Berlin, pp 618.

POORE, R.Z., 1981: Temporal and spatial distribution of ice rafted mineral grains in Pliocene sediments of the North Atlantic: Implications for Late Cenozoic climatic history. SEPM Special Publication No. 32, 505-515.

POSAMENTIER, H.W. and VAIL, P.R., 1988: Eustatic controls on clastic deposition II - Sequence and systems tract models. Society of Economic Paleontologists and mineralogists, Special Publication No. 42, 125-154.

POWELL, A.J., 1988: A modified dinoflagellate cyst biozonation for latest Palaeocene and earliest Eocene sediments from the central North Sea. Rev. Palaeobot. Palynol., 56: 327-344.

PRESTVIK, T., 1978: Cenozoic plateau lavas of Spitsbergen - a geochemical study. Norsk Polarinst. Årbok 1977, 9-143.

REID, H.F., 1892: Studies of Muir Glacier, Alaska. Natnl. Geogr. Mag. 4. 19 - 84.

REKSTAD, J.B., 1911-12: Die Ausfüllung eines Sees vor dem Ergabrae, dem grössten Ausläufer des Svartisen, als Mass der Gletchererosion. Zeit. Gletscherkunde 6, 212-214.

RUDDIMAN, W.F. and RAYMO, M.E., 1988: Northern Hemisphere climate regimes during the past 3 Ma: possible tectonic connections. Phil. Trans. R. Soc. London. B 318, 411-430.

SCHRADER, H.J. and FENNER, J., 1976: Norwegian Sea Cenozoic diatom biostratigraphy and taxonomy. In: Talwani, M. and Udintsev G., et al., Initial Reports of the Deep Sea Drilling Project, vol. 38. Washington (US Gov. Printing Office) pp. 921-1100.

SINGER, A. and MÜLLER, G., 1983: Diagenesis in argillaceous Sediments. In: Larsen, G. and Chilingar, G.V. (eds.): Diagenesis in sediments and sedimentary rocks, 2. Elsevier, 115-212.

SKJELKVÅLE, B.-L., 1987: En vulkanologisk og geokjemisk undersøkelse av tre kvartære vulkansentere i Bockfjord-Woodfjord området, NV-Spitsbergen. Cand Scient Thesis, University of Oslo, pp 194.

SKJELKVÅLE, B.-L., AMUNDSEN, H.E.F., O'REILLY, S.Y., GRIFFIN, W.L. and GJELSVIK, T., 1987: A primitive alkali basaltic stratovolcano and associated eruptive centres, Northwestern Spitsbergen: Volcanology and tectonic significance. J. Volcanol. Geotherm. Res. 37, 1-19.

STOVER, L.E., 1977: Oligocene and Early Miocene Dinoflagellates from Atlantic Corehole 5/5b, Blake Plateau. In: Contributions of stratigraphic palynology (with emphasis on North America). Volume 1 - Cenozoic Palynology: Amer. Assoc. Stratigr. Palynol. Contr. Ser. 5A: 66-89, 3 pls.

SVENDSEN, J.O, MANGERUD, J. and MILLER, G.H., 1989: Denudation rates in the arctic estimated from lake sediments on Spitsbergen, Svalbard. In press in Palaeography, Palaeoclimatology, Palaeoecology.

SÆTTEM, J., 1985: Heat Flow Measurements Barents Sea 1985. IKU Report 21.3405.00/02/85, 63 pp. Restricted.

TAYLOR, P.M. and MORTON, A.C., 1989: Sr, Nd, and Pb isotope geochemistry of the Upper and Lower volcanic series at site 642. In: Eldholm, O., Thiede, J. and Taylor, E. et al. (eds.): Proceedings of the Ocean Drilling Program, Scientific Results, Vol 104, 429-435.

THORARINSON, S., 1939: Hoffellsjökull, its movements and drainage. Geografiska Annaler, 21, 189-215.

TOXWENIUS, B.B., 1986: Biostratigraphical zonation and correlation of five Late Cretaceous-Tertiary wells, offshore central West Greenland. Grønlands Geologiske Undersøgelse, Rapport nr. 130, 36-43.

UPTON, B.G.J., 1988: History of Tertiary igneous activity in the N Atlantic boarderlands. In: Morton, A.C. and Parson, L.M. (eds.): Early Tertiary Volcanism and the Opening of the NE Atlantic, Geol. Soc. Spec. Publ. 39, 429-453.

VORREN, T.O., RICHARDSEN, G., KNUTSEN, S.M. and HENRIKSEN, E., in prep.: The western Barents Sea during the Cenozoic. Manuscript to: "Geologic history of the Polar Oceans: Arctic versus Antarctic." Eds. Bleil, U. and Tiede, J., NATO ASI series, Klüver Acad. Publ.

VÅGENES, E., FALEIDE, J.I. and GUDLAUGGSON, S.T., 1990: Tertiær tektonikk og sedimentasjon i sørvestre Barentshav (vest for 18 Ø). (Abstract). Geonytt 17, Nr 1, p. 121.

WALKER, R.G., 1984: Shelf and shallow marine sands. In: Walker, R.G. (ed): Facies Models. Geoscience of Canada, Reprint Series 1, 2. edition, 141-189.

WALL, D., DALE, B., LOHMANN, G.P. and SMITH, W.K., 1977: The environmental and climatic distribution of dinoflagellate cysts in modern marine sediments from regions in the North and South Atlantic Oceans and adjacent seas. Mar. Micropaleontol., 2: 121-200.

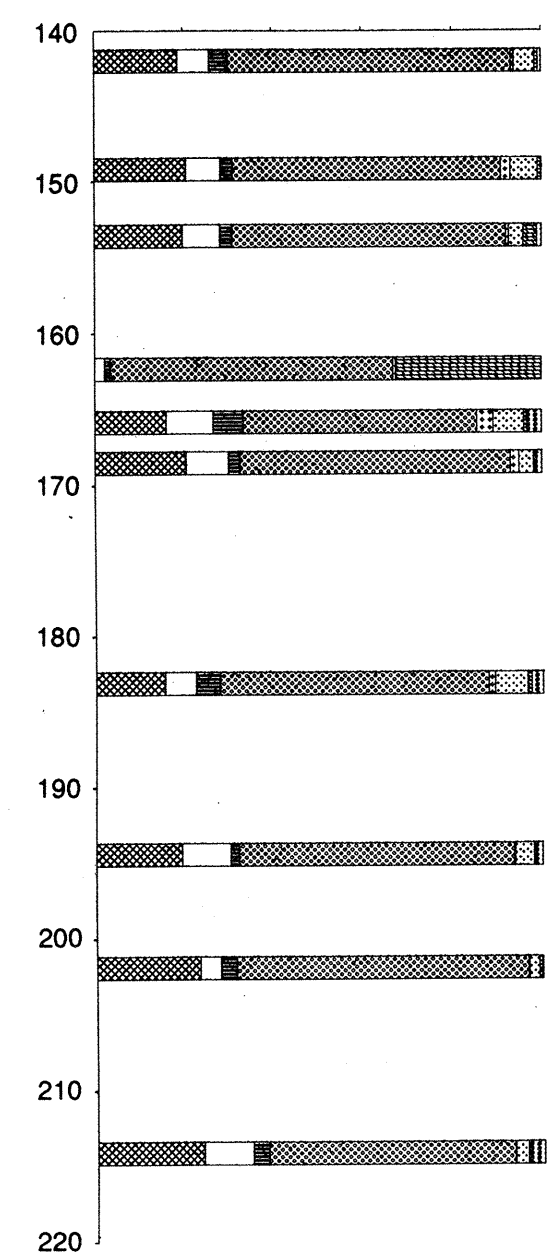
WILLIAMS, G.L. and BUJAK, J.P., 1985: Mesozoic and Cenozoic dinoflagellates. In: Bolli, H.M., Saunders, J.B., Perch-Nielsen, K. (eds.), Plankton stratigraphy: Cambridge Earth Sciences Series: 847-964, 41 pls., Cambridge.

WORSLEY, D., JOHANSEN, R. and KRISTENSEN, S.E., 1988: The Mesozoic and Cenozoic succession of Tromsøflaket. In: Dalland, A., Worsley, D. and Ofstad, K., (eds.): A lithostratigraphic scheme for the Mesozoic and Cenozoic succession offshore mid- and northern Norway. NPD-BULLETIN NO 4, 42-63.

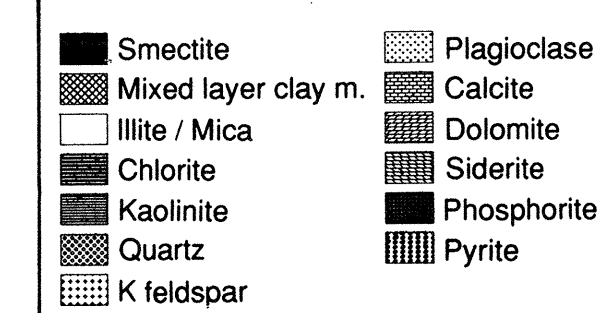
ZIEGLER, P.A., 1988: Evolution of the Arctic-North Atlantic and Western Tethys. AAPG Memoir 43, p. 197, 4 maps.

SEMI-QUANTITATIVE MINERALOGY (XRD)

(relative percent)

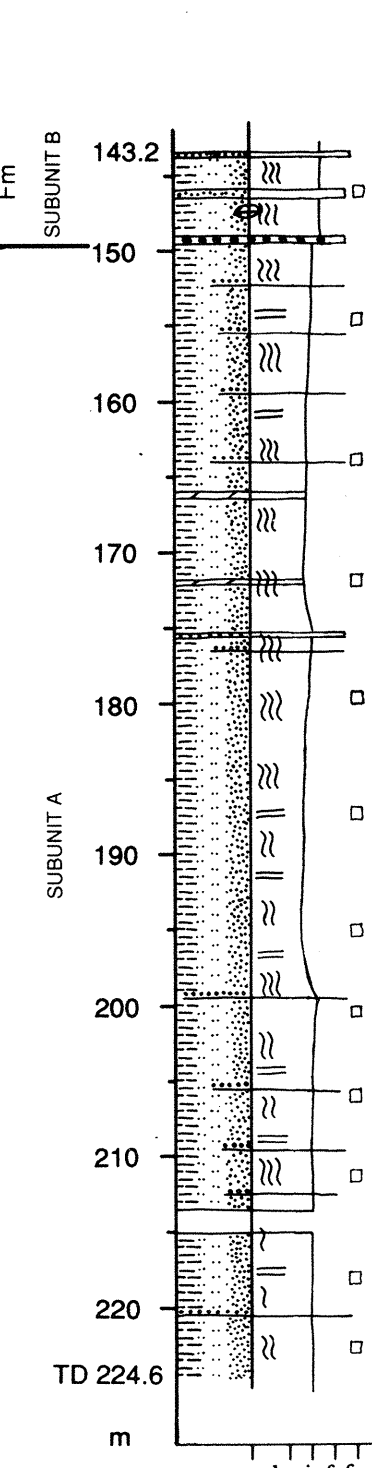


LEGEND TO XRD-PLOT



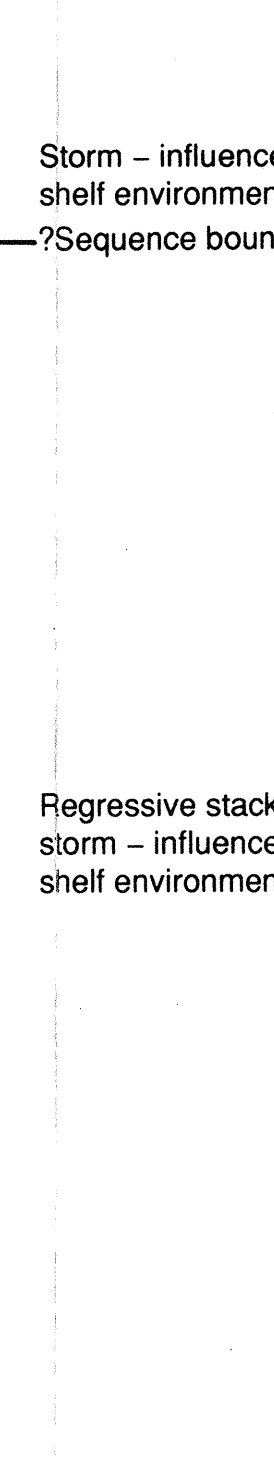
SEDIMENTOLOGICAL CORE LOG

CORE LOG



DEPOSITIONAL ENVIRONMENT

DEPOSITIONAL ENVIRONMENT



FOSSIL TAXA

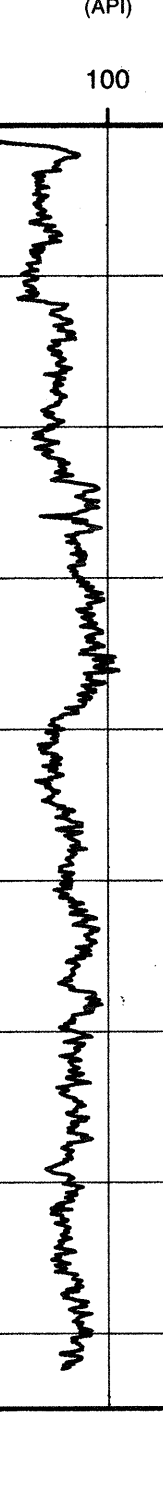
(First/last occurrence of significant taxa)



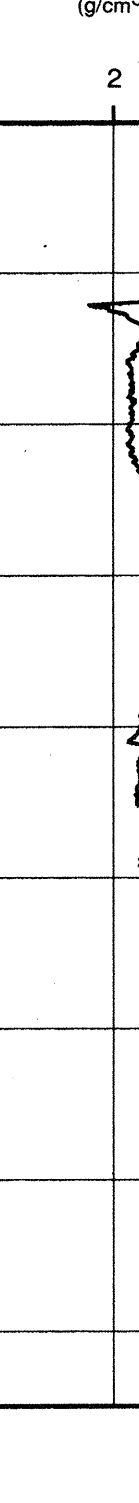
PETROPHYSICAL LOGS

GAMMA

(API)

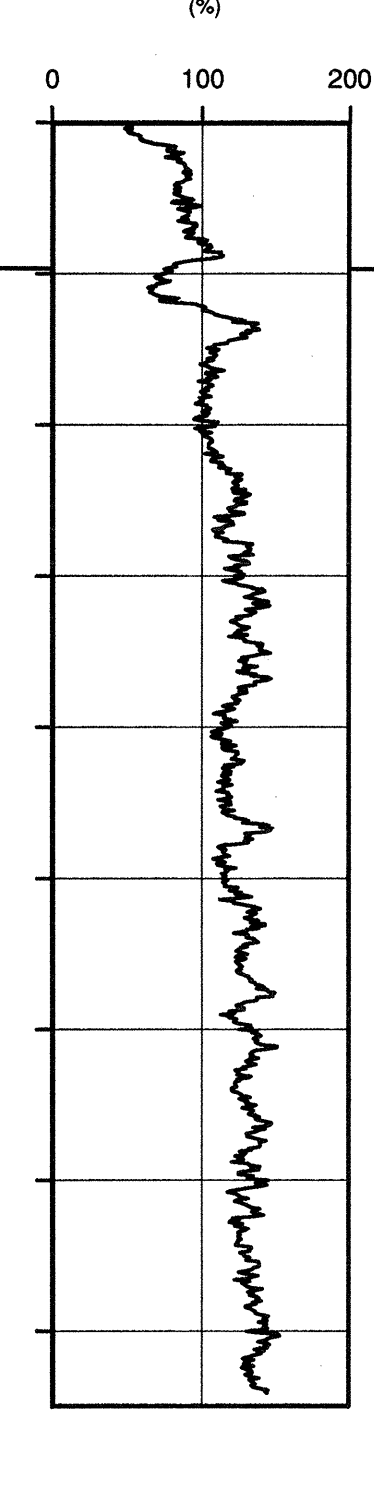


DENSITY

(g/cm³)

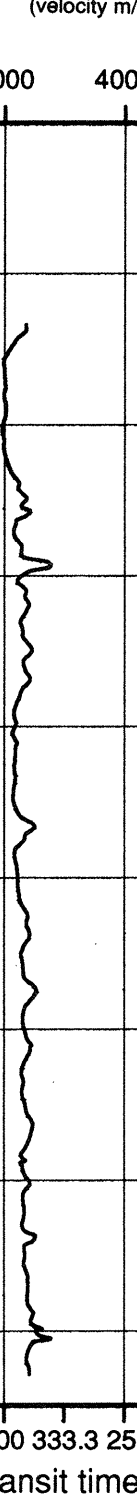
NEUTRON POROSITY

(%)



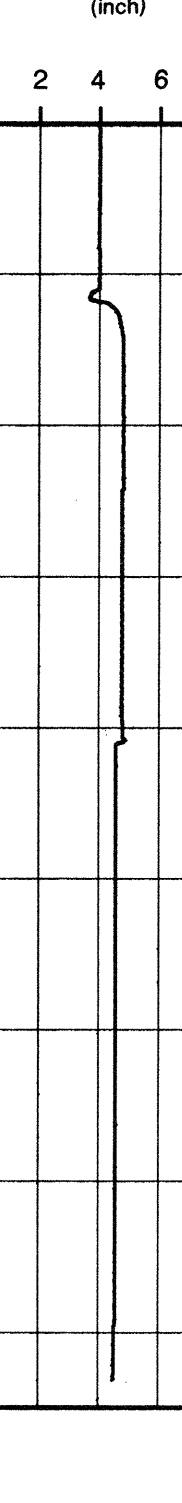
SONIC

(velocity m/s)



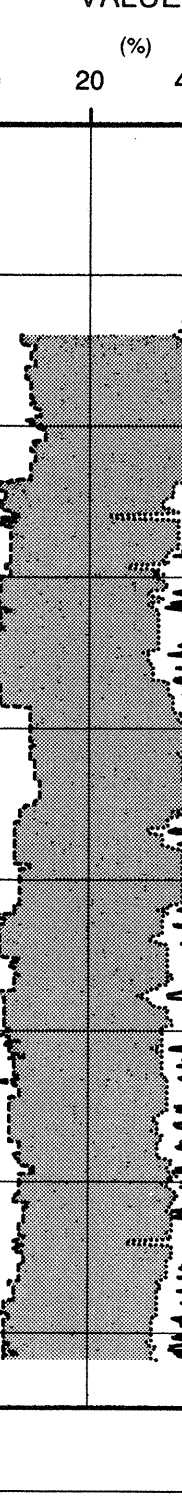
CALIPER

(inch)



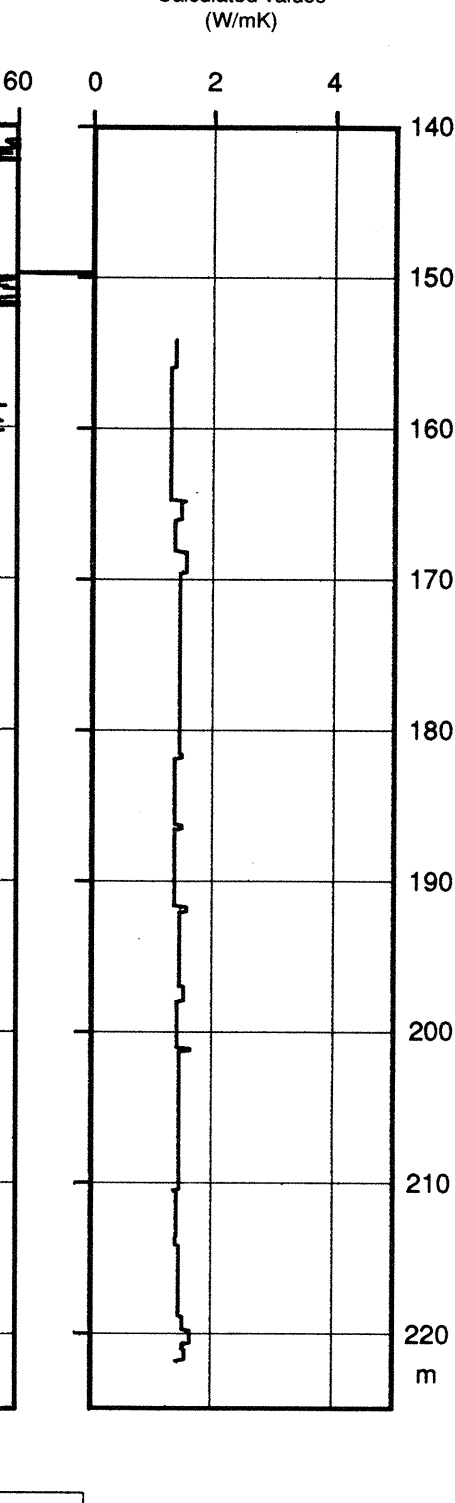
POROSITY VALUES

(%)



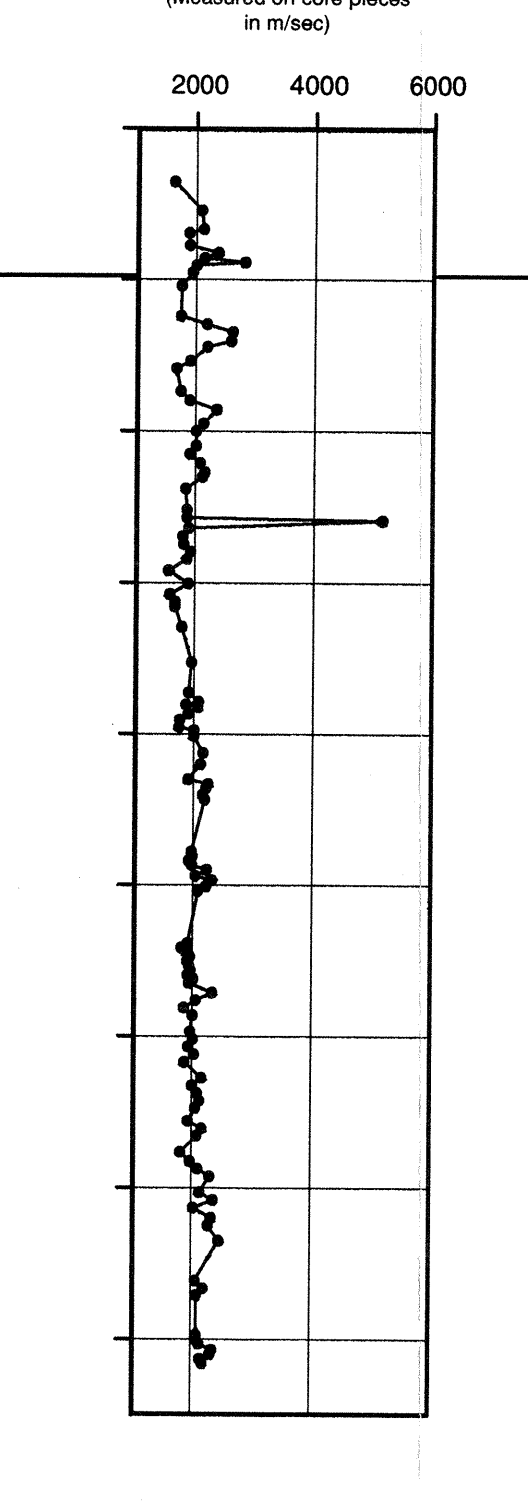
THERMAL CONDUCTIVITY

Calculated values (W/mK)



HORIZONTAL SOUND VELOCITY

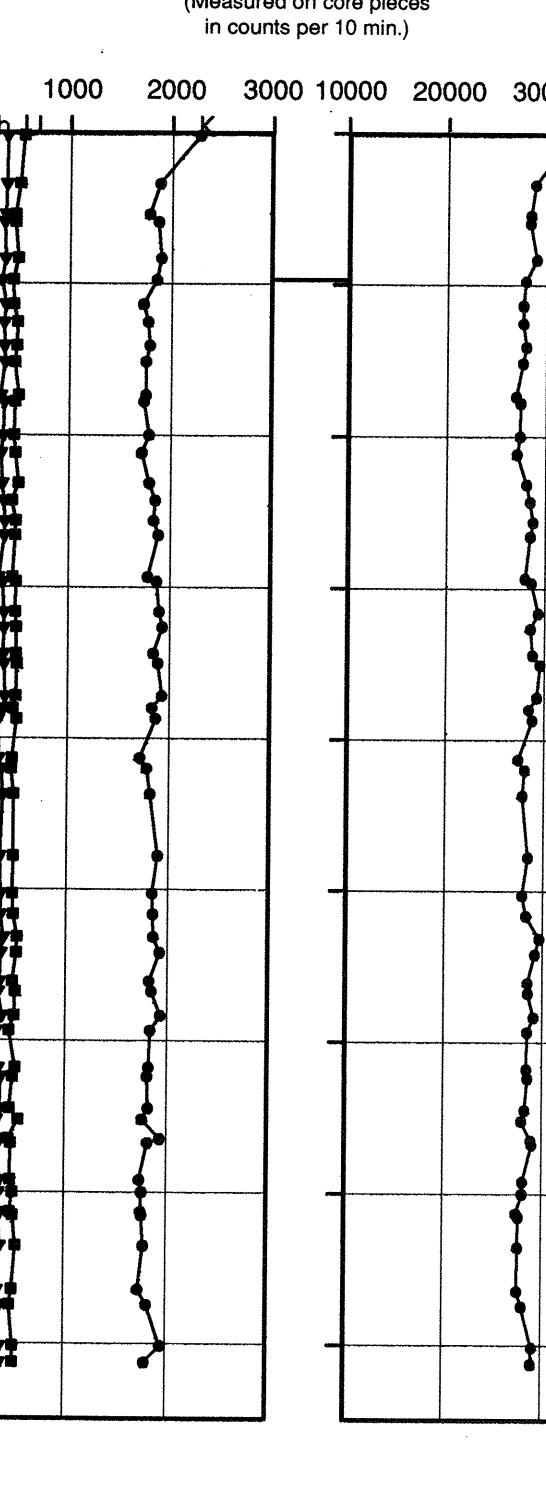
(Measured on core pieces in m/sec)



CORE MEASUREMENT

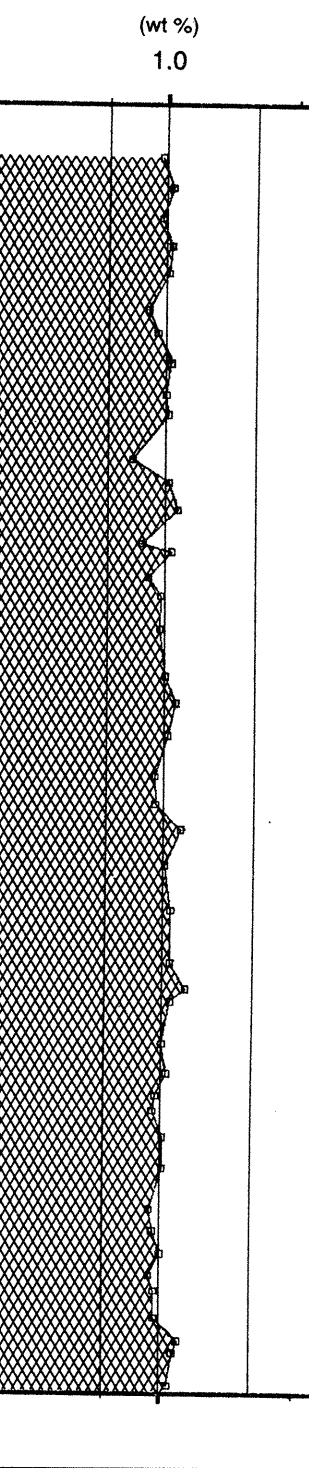
NATURAL GAMMA ACTIVITY

(Measured on core pieces in counts per 10 min.)

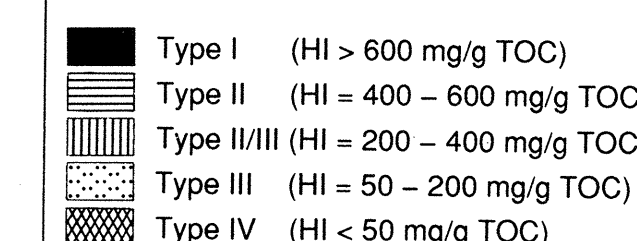


TOTAL ORGANIC CONTENT (TOC)

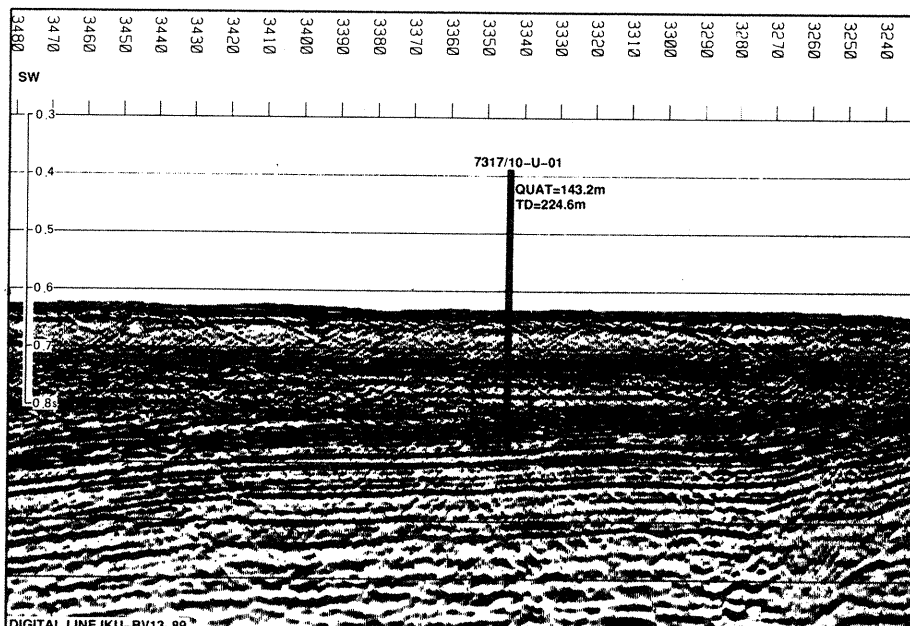
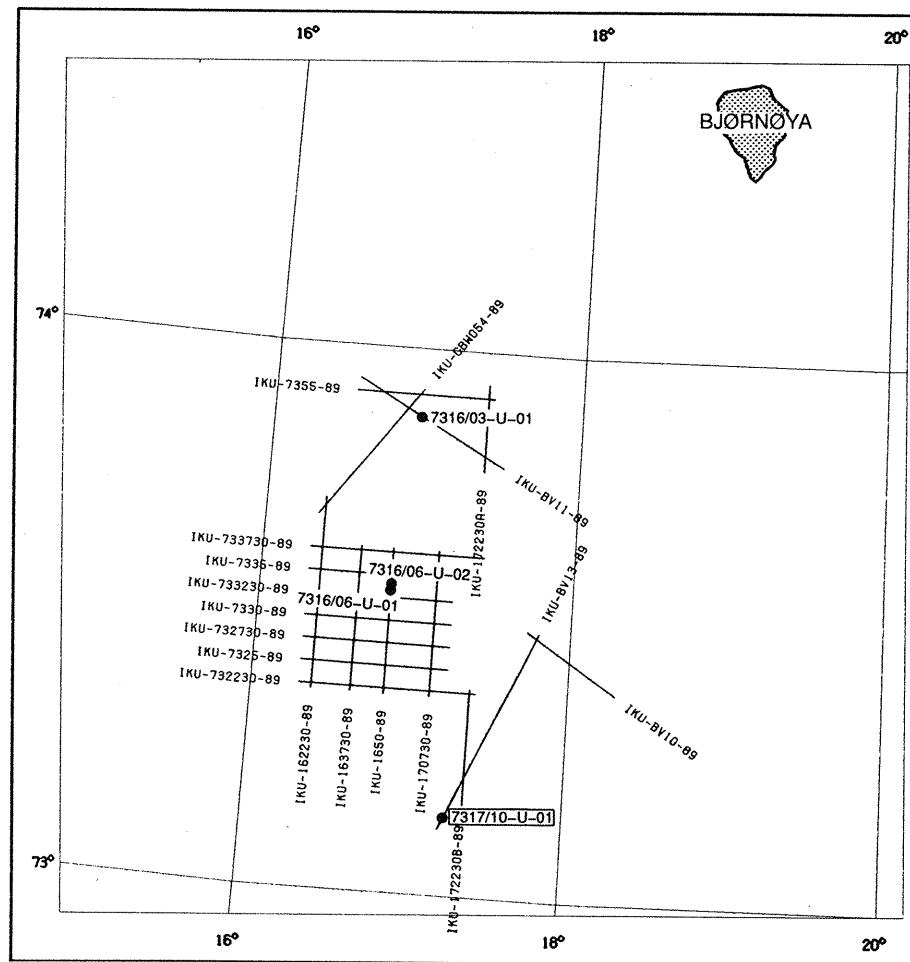
(wt %)



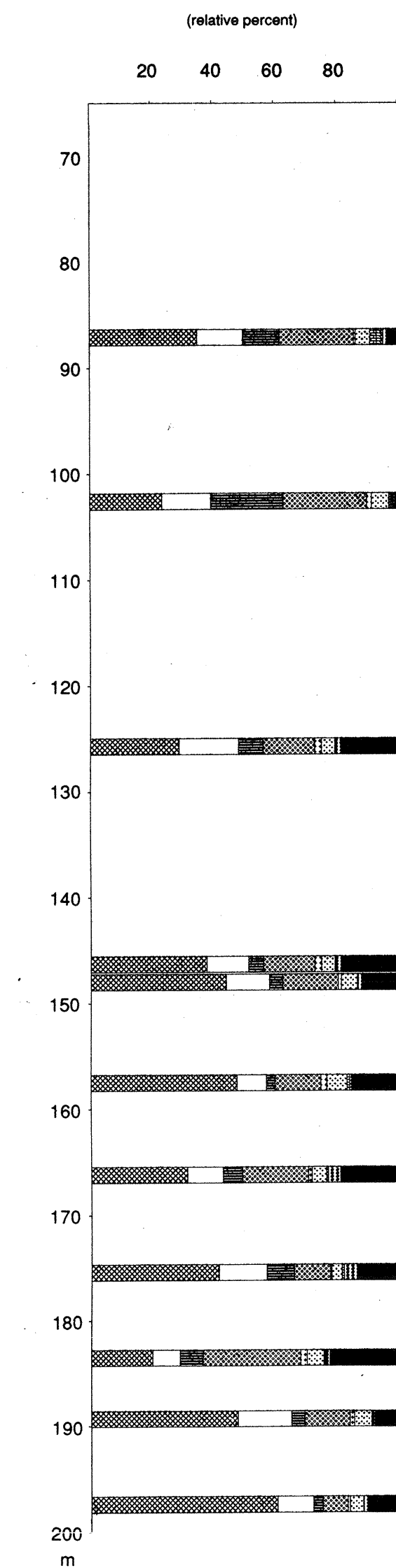
LEGEND TO KEROGEN TYPE

IKU
SINTEF GROUP
7317/10-U-01APPENDIX A.1.1
SUMMARY SHEET
21.3465.00/02/90

Scale 1 : 500

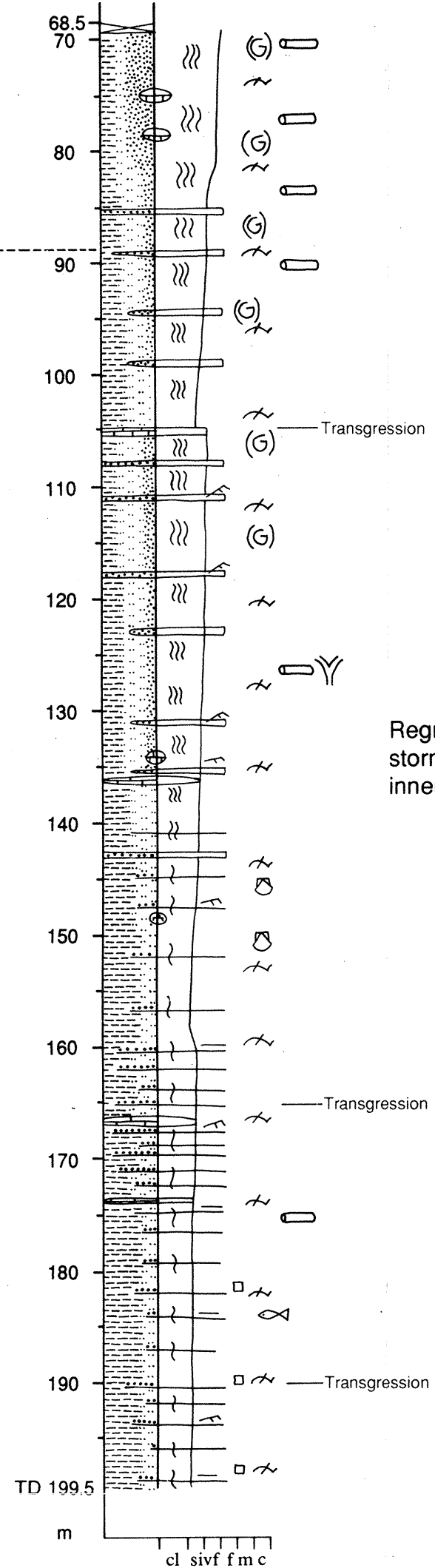
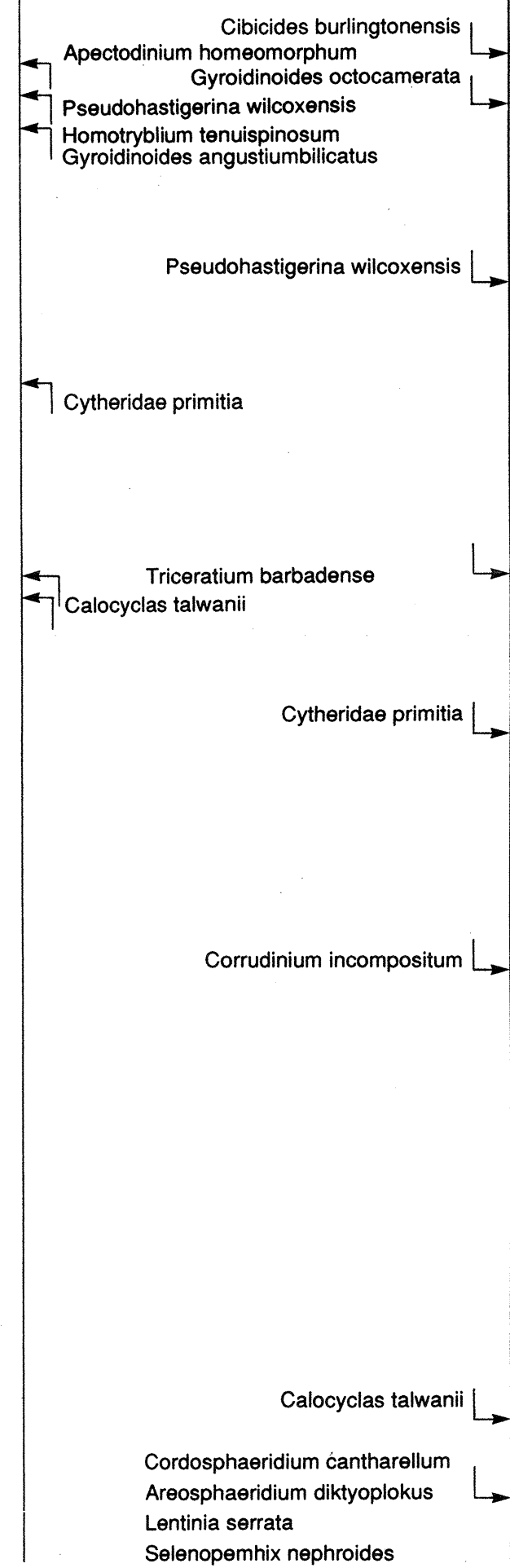


SEMI-QUANTITATIVE MINERALOGY (XRD)

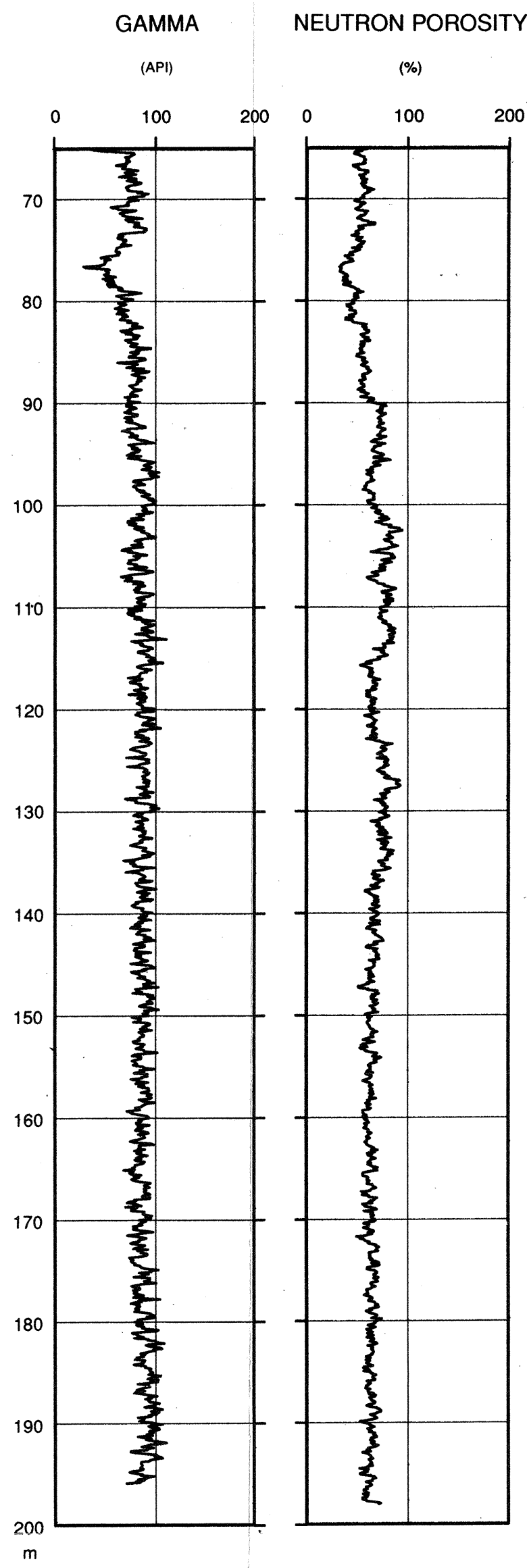
SEDIMENTOLOGICAL
CORE LOG

MIDDLE EOCENE

LOWER EOCENE

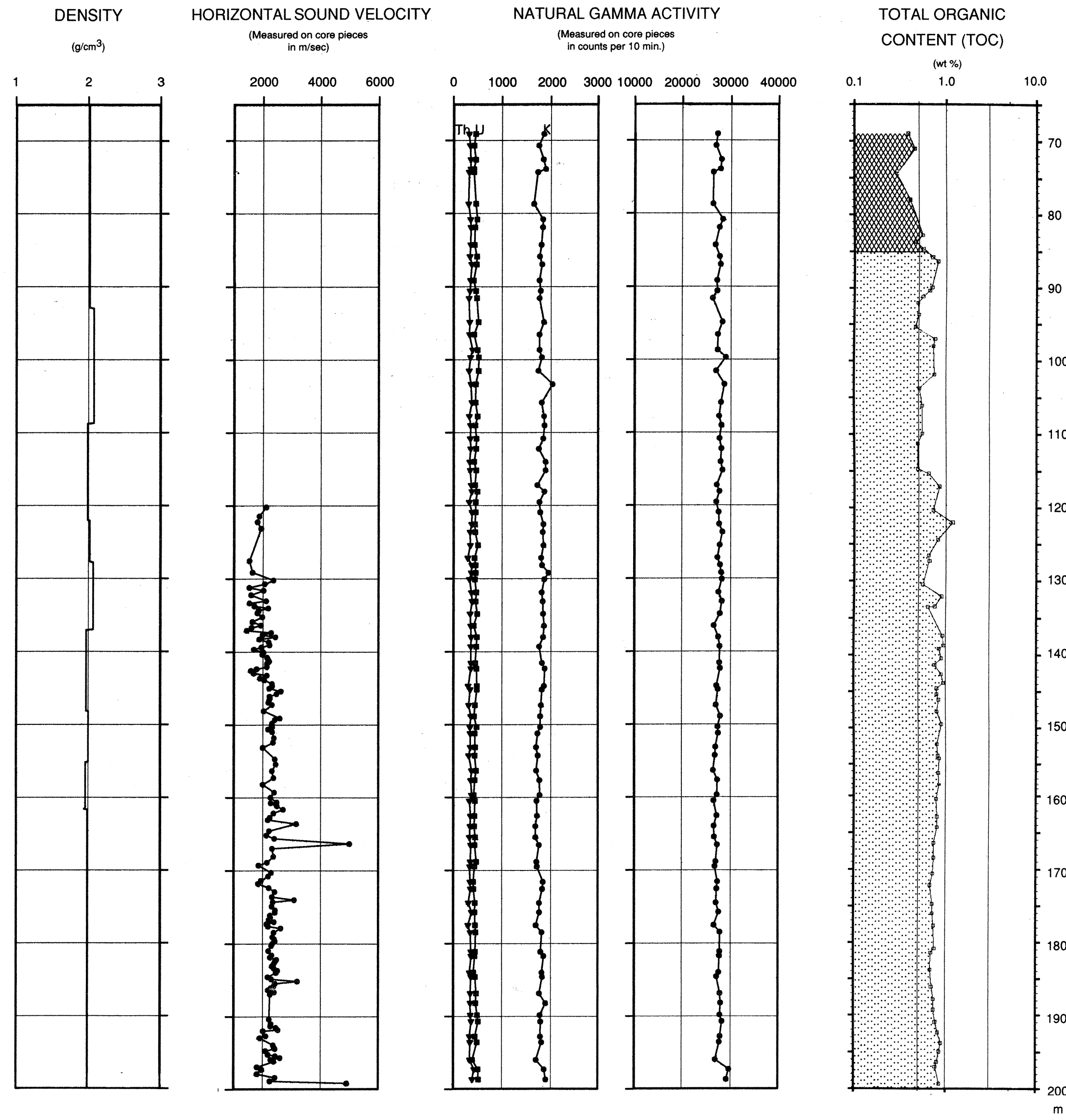
DEPOSITIONAL
ENVIRONMENTRegressive stacked
storm - influenced
inner shelf successionsFOSSIL TAXA
(First/last occurrence
of significant taxa)

PETROPHYSICAL LOGS

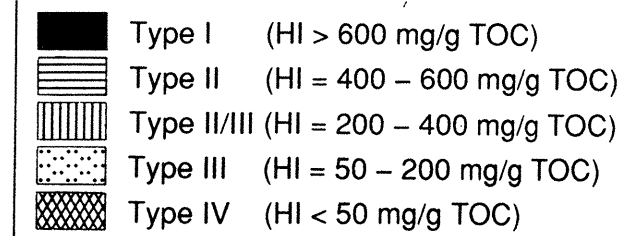


Logged inside HR-string

CORE MEASUREMENT



LEGEND TO KEROGEN TYPE

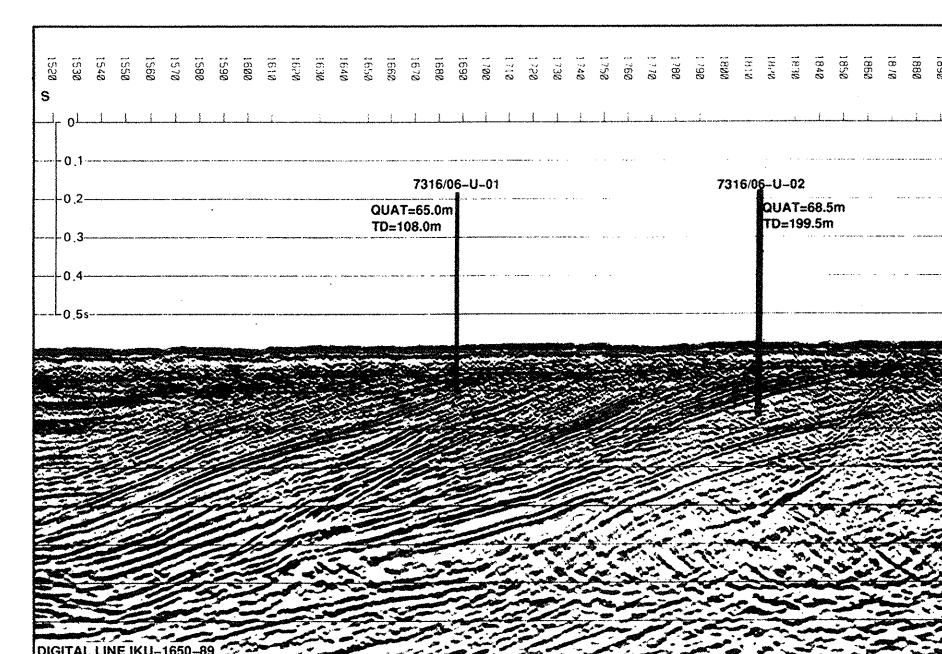
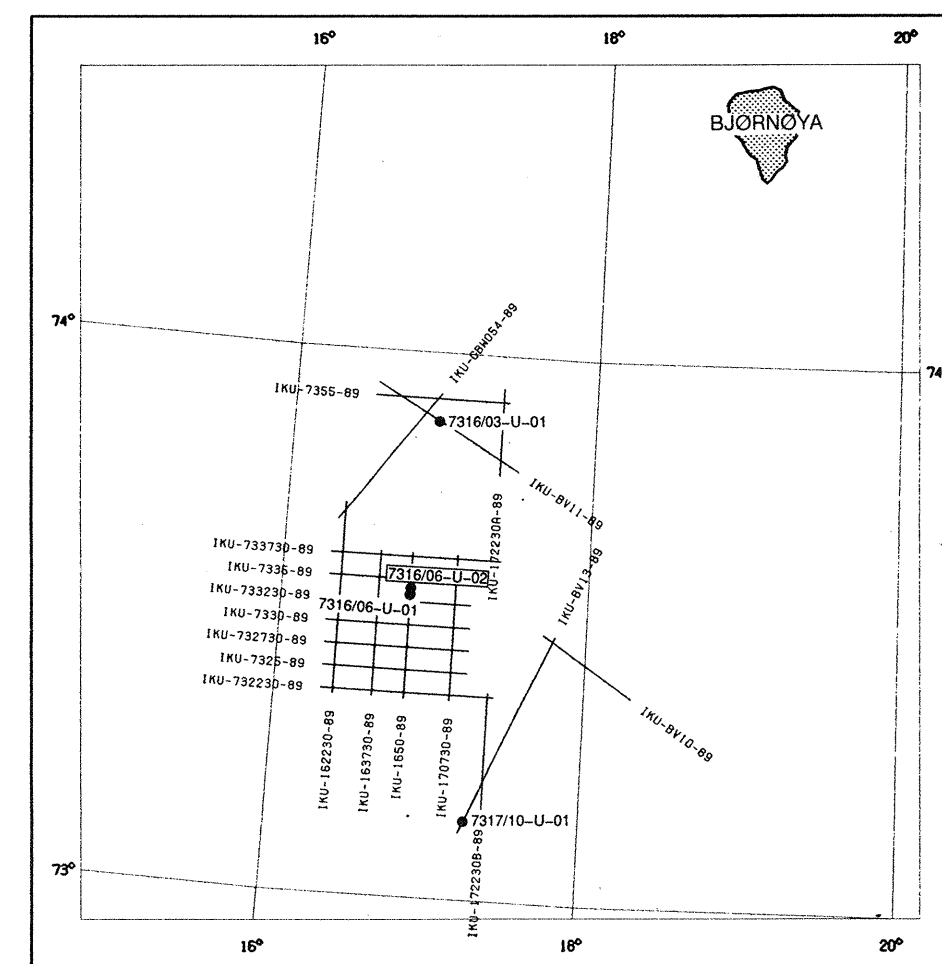


7316/06-U-02

APPENDIX A.1.2
SUMMARY SHEET

21.3465.00/02/90

Scale 1 : 500



SEMI-QUANTITATIVE MINERALOGY (XRD)

(relative percent)

SEDIMENTOLOGICAL CORE LOG

DEPOSITIONAL ENVIRONMENT

FOSSIL TAXA

(First/last occurrence of significant taxa)

PETROPHYSICAL LOGS

CORE MEASUREMENT

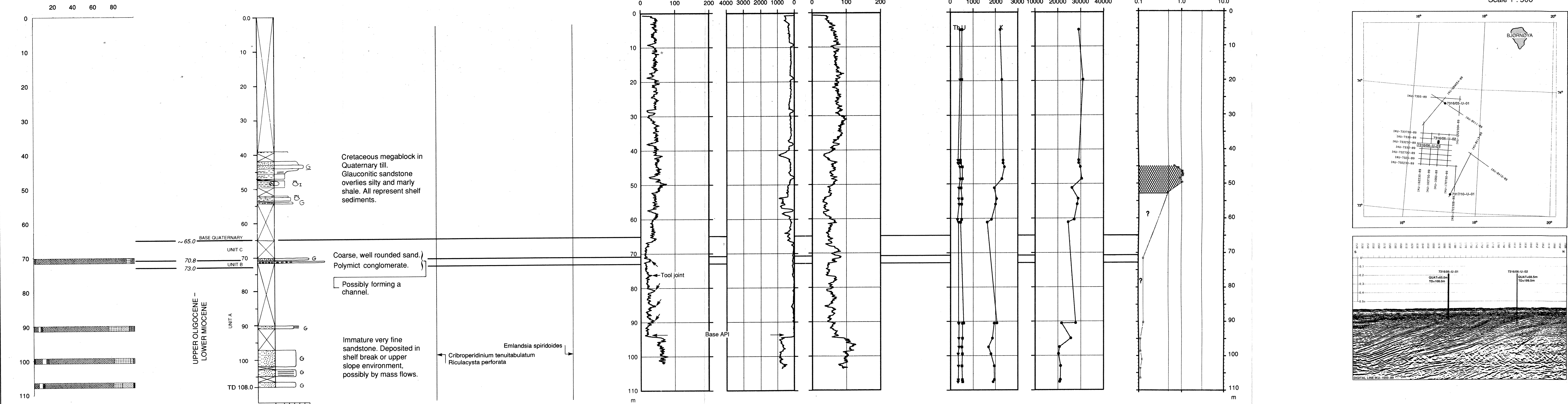


7316/06-U-01

APPENDIX A.1.3
SUMMARY SHEET

21.3465.00/02/90

Scale 1 : 500

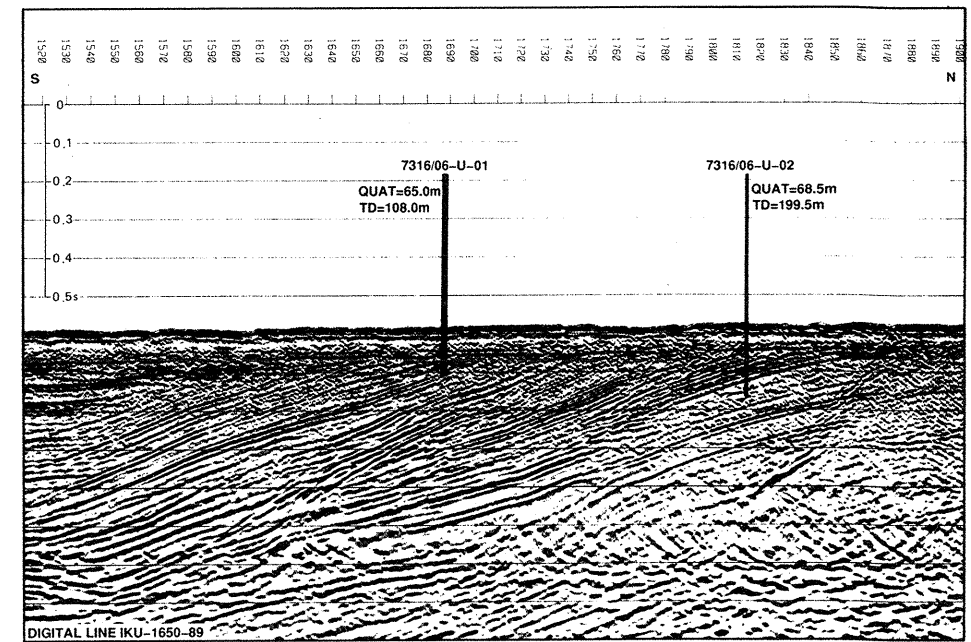
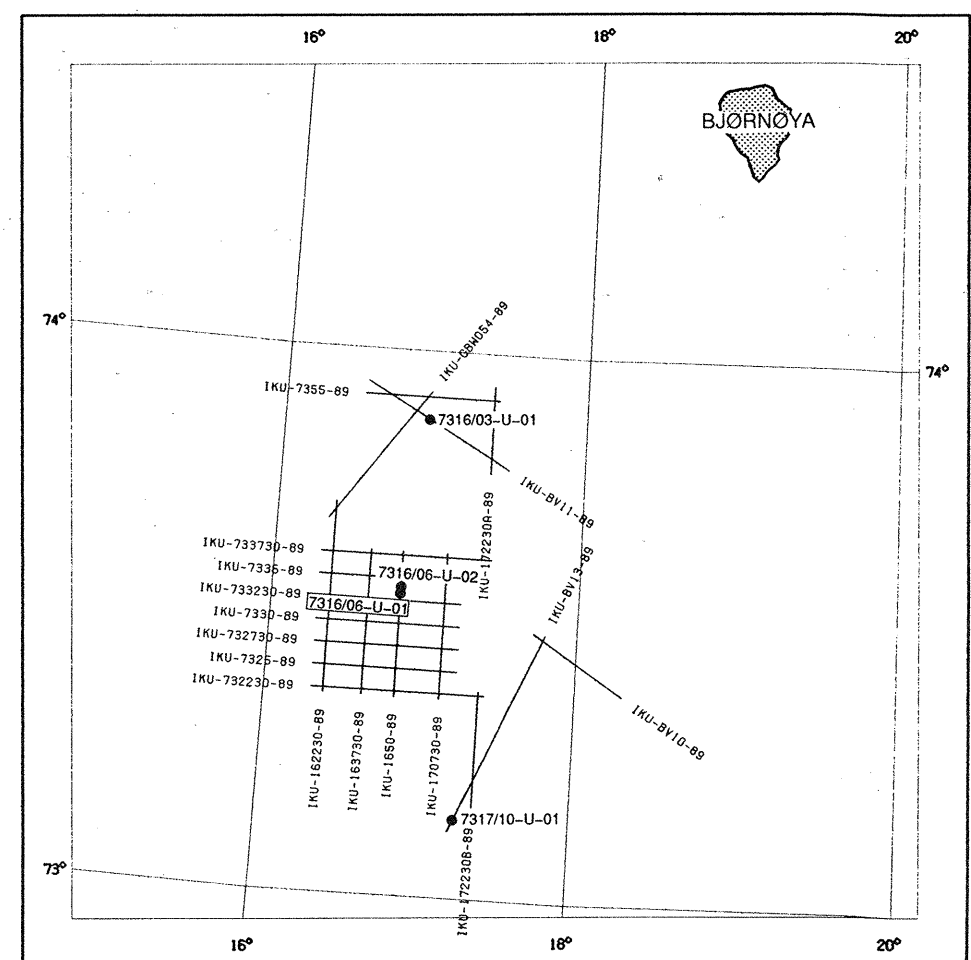


LEGEND TO XRD-PLOT

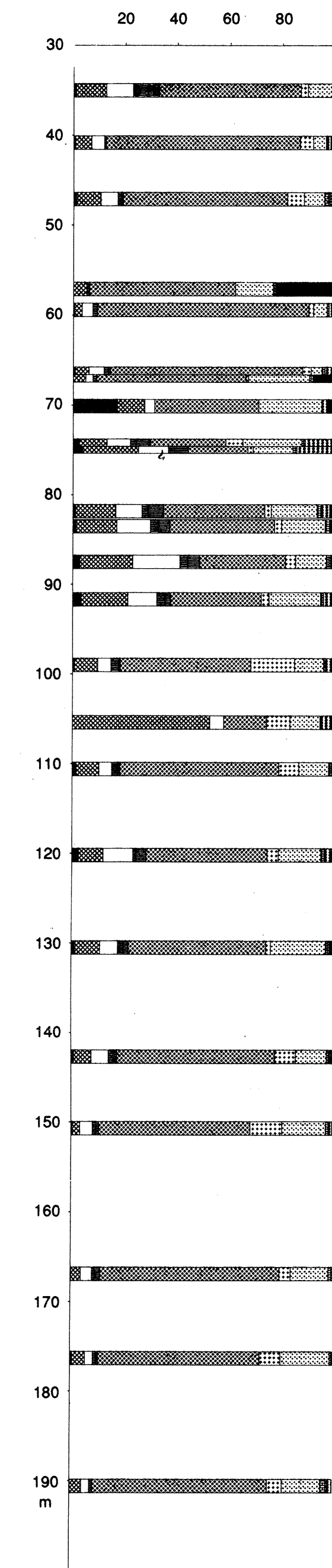
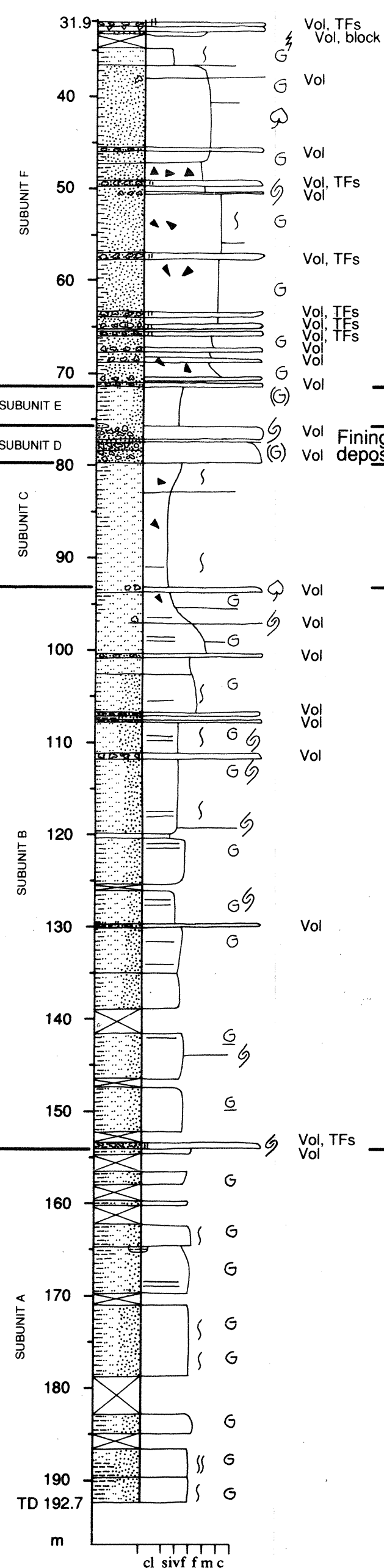
Smectite	Calcite
Mixed layer clay m.	Dolomite
Illite / Mica	Siderite
Chlorite	Phosphorite
Kaolinite	Pyrite
Quartz	Opal
K feldspar	Zeolite
Plagioclase	

LEGEND TO KEROGEN TYPE

Type I (HI > 600 mg/g TOC)
Type II (HI = 400 - 600 mg/g TOC)
Type II/III (HI = 200 - 400 mg/g TOC)
Type III (HI = 50 - 200 mg/g TOC)
Type IV (HI < 50 mg/g TOC)



Logged inside HR-string

SEMI-QUANTITATIVE MINERALOGY (XRD)
(relative percent)SEDIMENTOLOGICAL
CORE LOGDEPOSITIONAL
ENVIRONMENT

Fine to medium grained, glauconitic sandstone with volcanic debris beds. Deposited by debris flow and mass flow mechanisms in upper slope environment by resedimentation of Eocene shelf sediments.

Silty shale deposited by turbidity currents or possibly by hemipelagic deposition.

Fining upward beds with volcanic debris deposited by debris flows in slope setting

Silty shale deposited by turbidity currents or possibly by hemipelagic sedimentation.

Very fine to fine-grained glauconitic immature sandstone deposited by debris flows and turbidity currents in slope settings.

Very fine-grained glauconitic immature sandstone deposited in slope channels and by turbidity currents in slope setting.

FOSSIL TAXA
(First/last occurrence
of significant taxa)

Hystriochostroglyon sp. Edwards 1984

Tesserastrum thielei

Coscinodiscus vigilans

Opephora gemmata

Achomosphaera andalousiensis

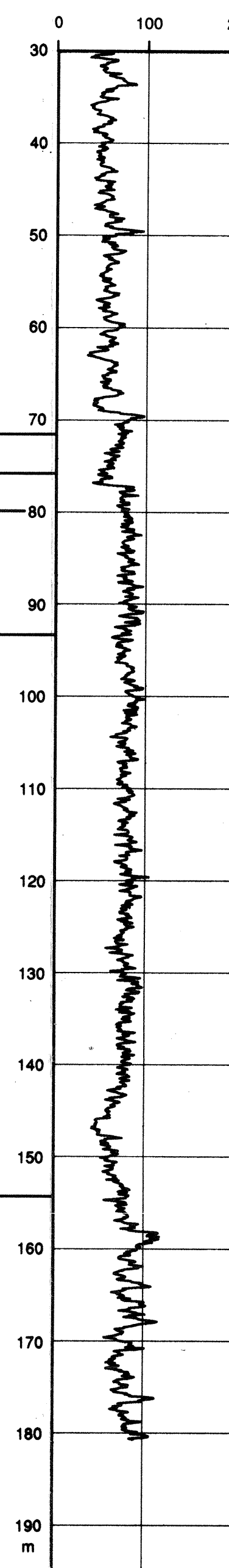
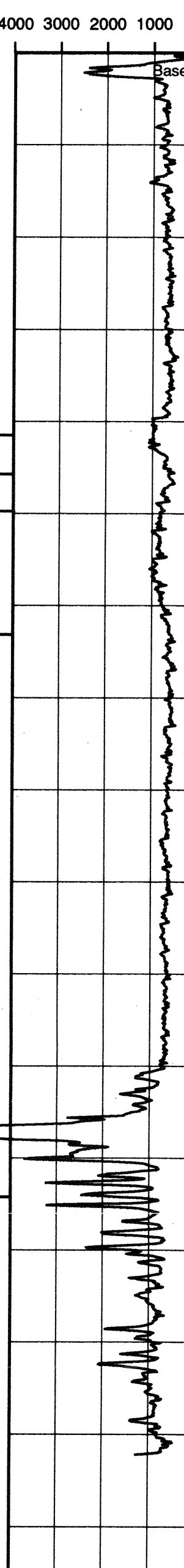
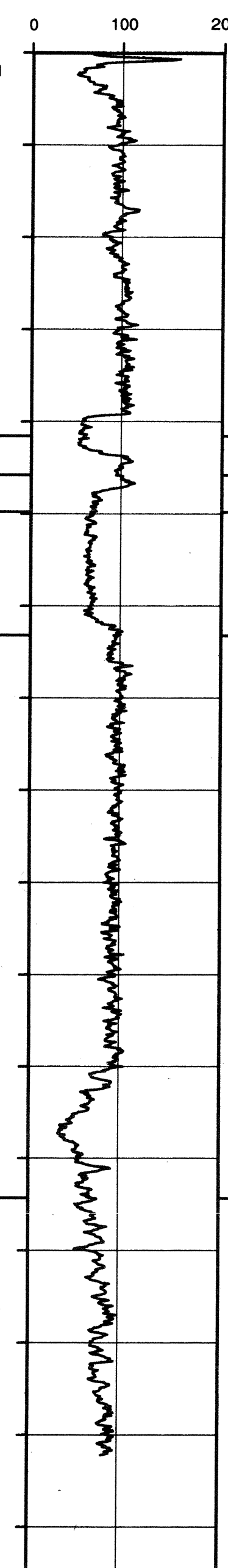
Invertocysta tabulata

Cribroperidinium tenuitubulatum

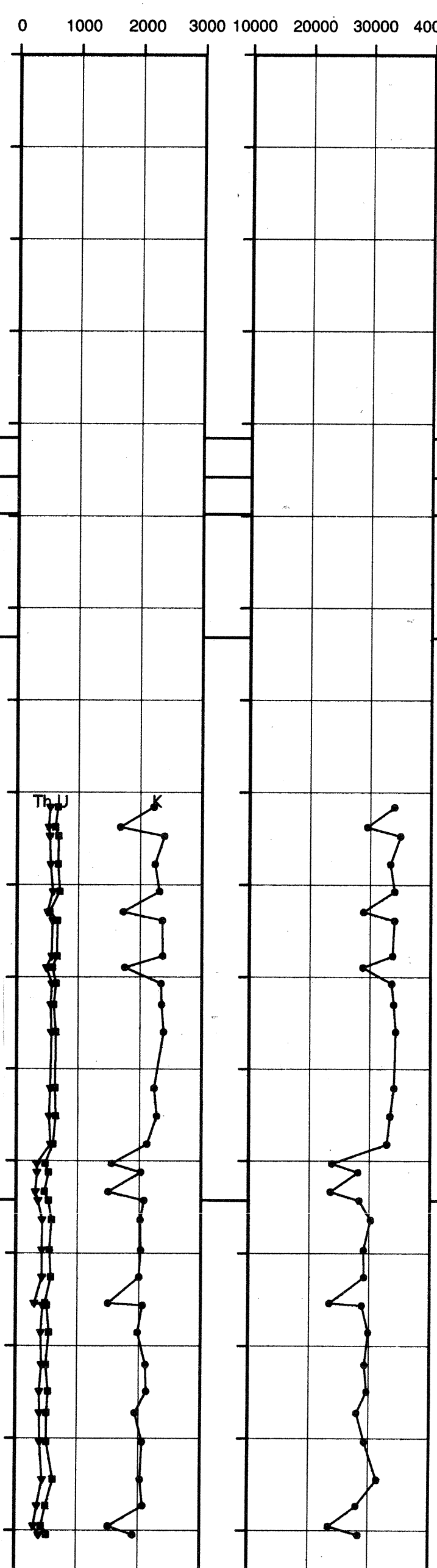
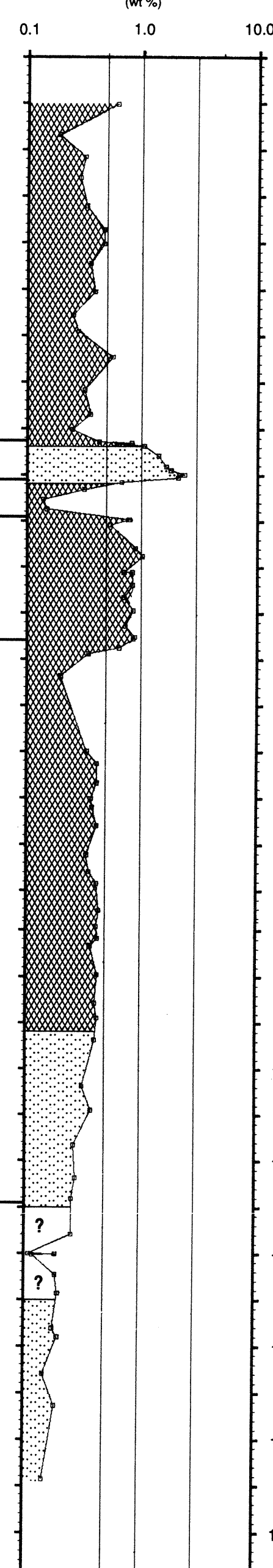
Labyrinthidium cf. truncatum

Impagidinium patulum

PETROPHYSICAL LOGS

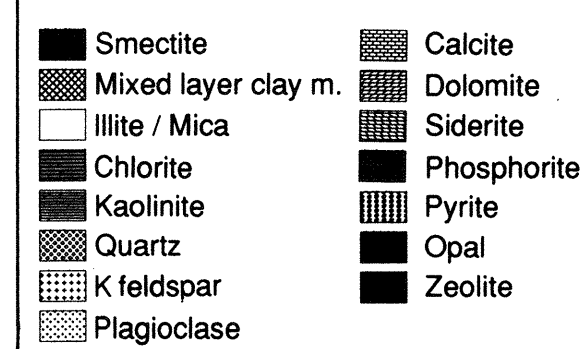
GAMMA
(API)DENSITY
(LSDU)NEUTRON POROSITY
(%)

CORE MEASUREMENT

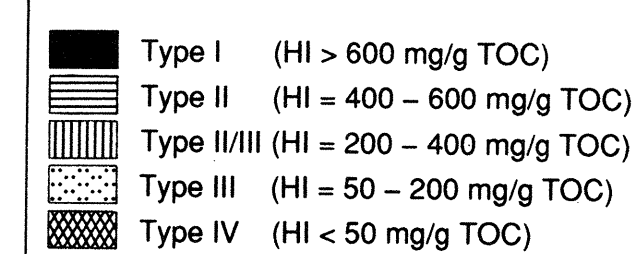
NATURAL GAMMA ACTIVITY
(Measured on core pieces
in counts per 10 min.)TOTAL ORGANIC
CONTENT (TOC)
(wt %)

Logged inside HR-string

LEGEND TO XRD-PLOT



LEGEND TO KEROGEN TYPE

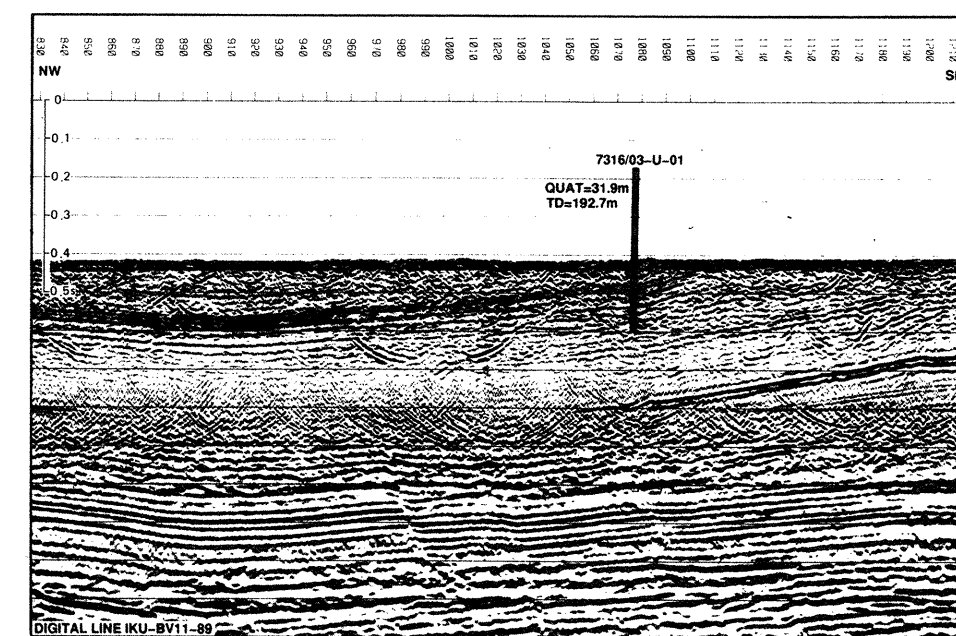
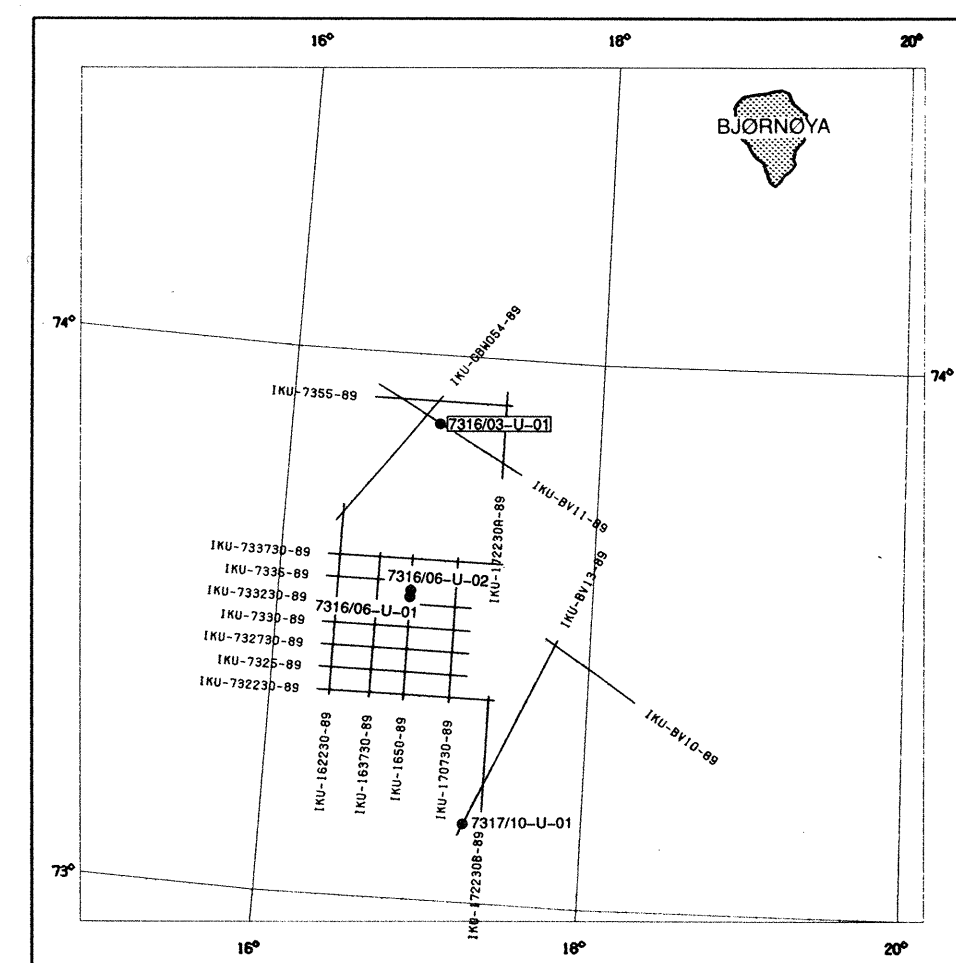


IKU
SINTEF GROUP
7316/03-U-01

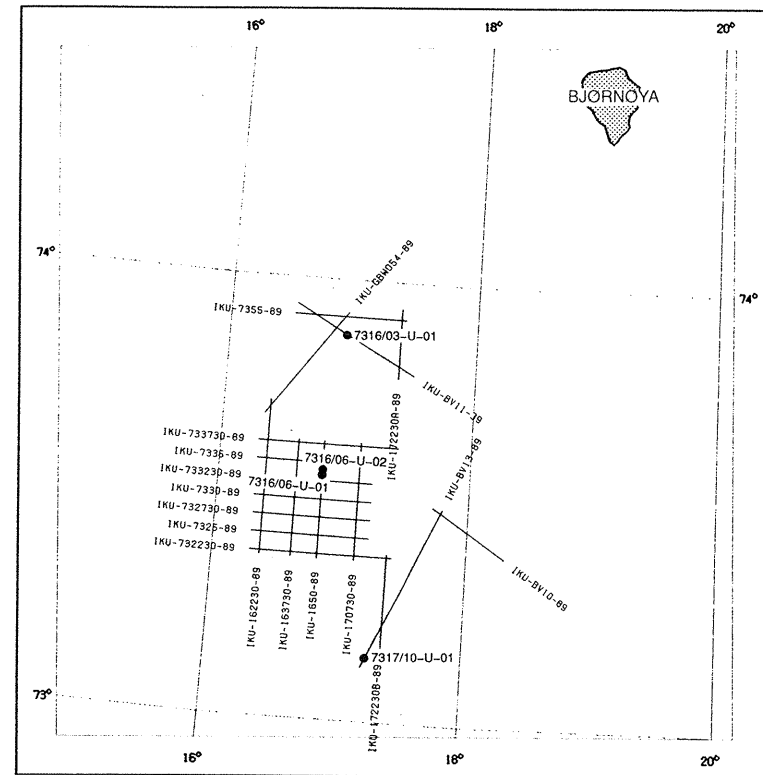
APPENDIX A.1.4
SUMMARY SHEET

21.3465.00/02/90

Scale 1 : 500



Appendix A.2.



Key to kerogen type:

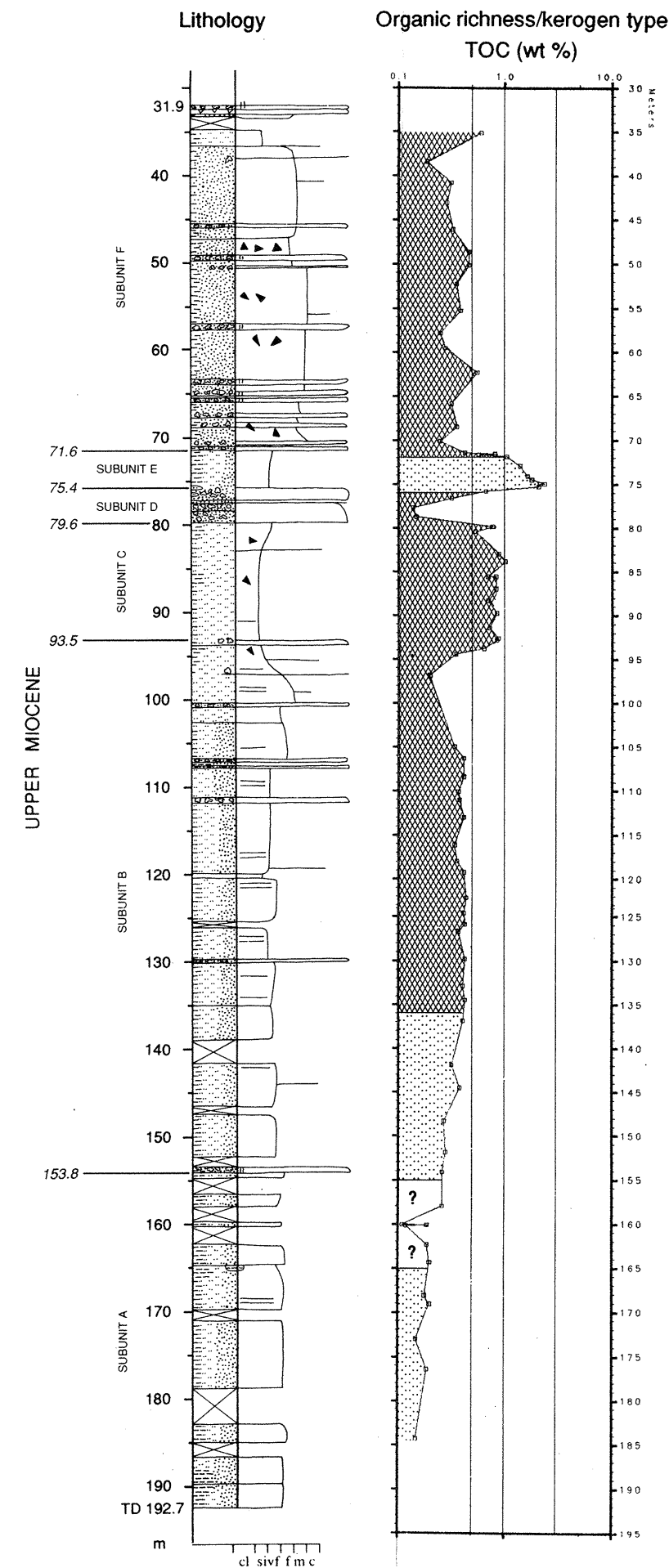
- Type I (HI > 600 mg/g TOC)
- Type II (HI = 400 – 600 mg/g TOC)
- Type II/III (HI = 200 – 400 mg/g TOC)
- Type III (HI = 50 – 200 mg/g TOC)
- Type IV (HI < 50 mg/g TOC)

21.3465.00/02/90
BJØRNØYA WEST AREA
 Organic richness/Rock-Eval kerogen type of
 the cored sections
 Compiled by T.L.Leith
 February 1990



7316/03-U-01

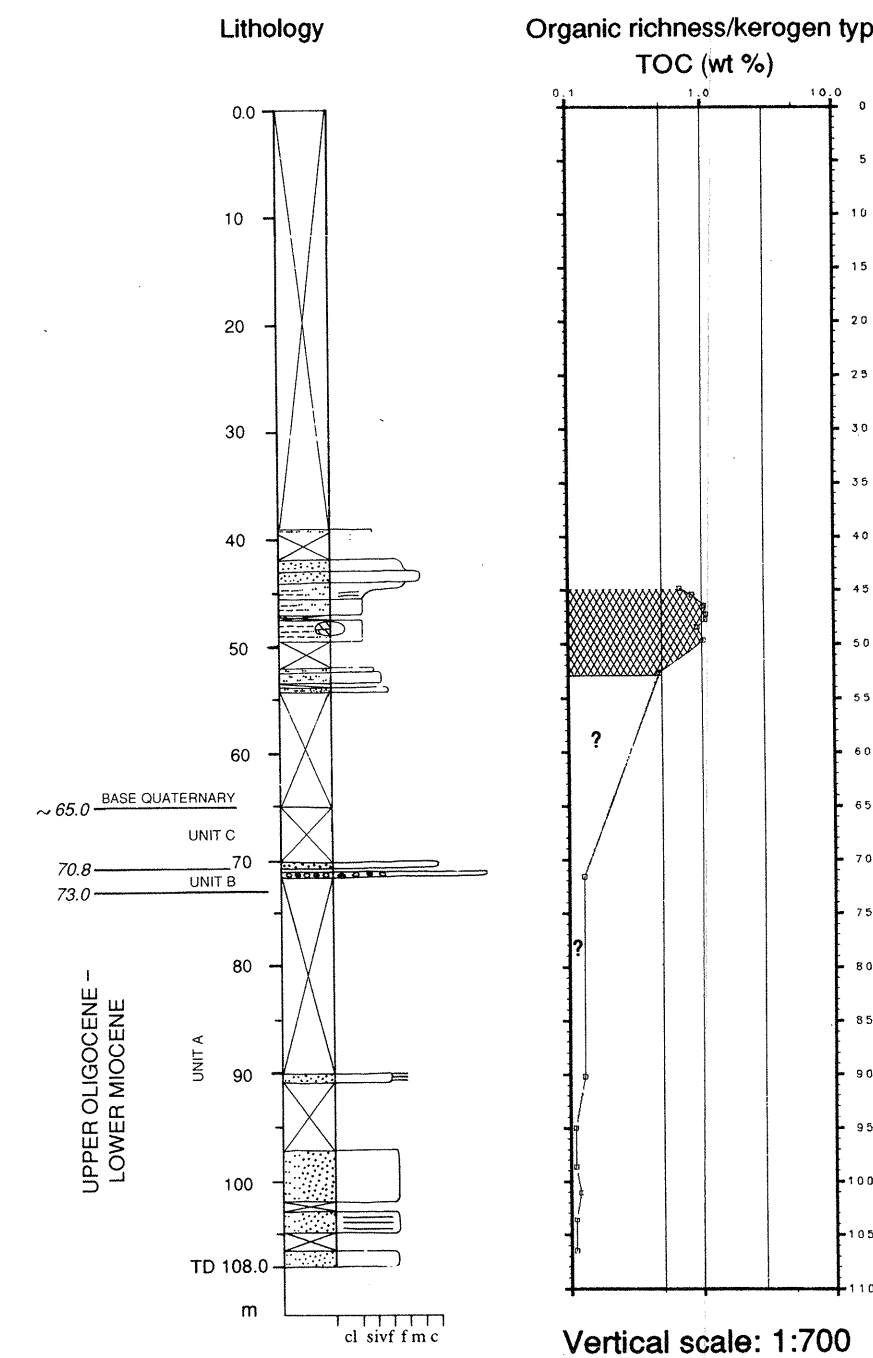
Age: Late Miocene



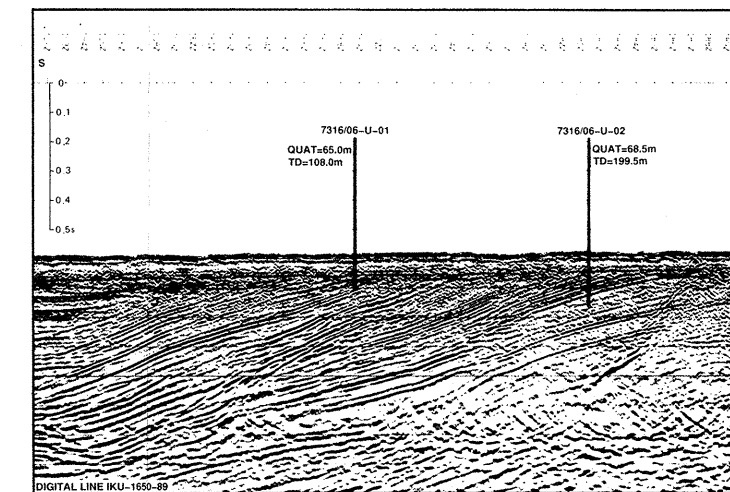
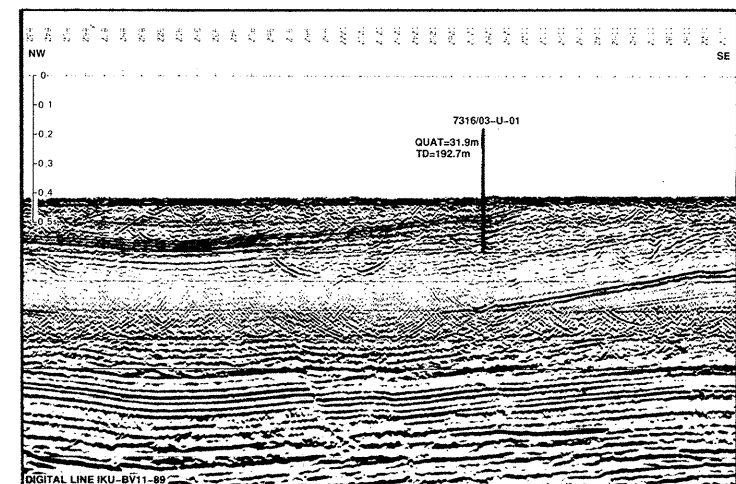
Vertical scale: 1:700

7316/06-U-01

Age: Late Oligocene – Early Miocene

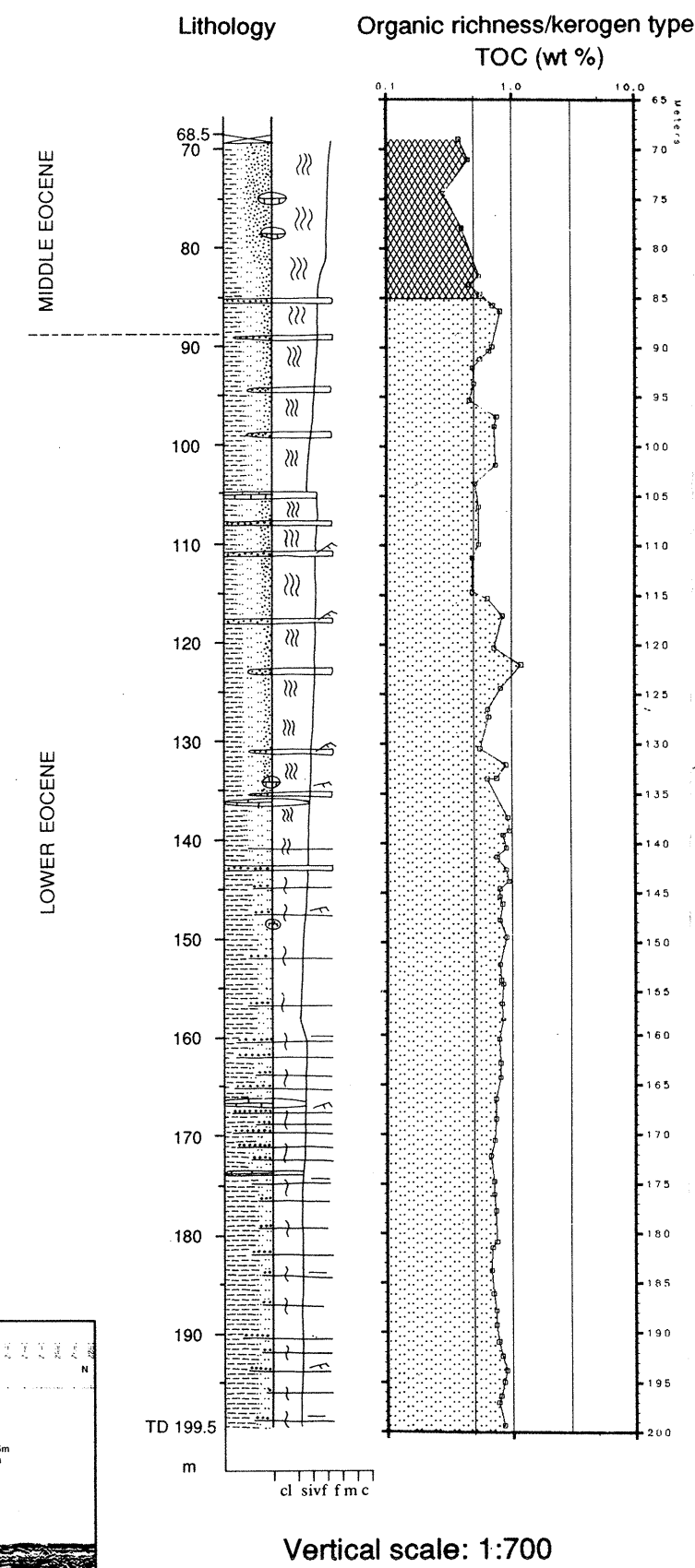


Vertical scale: 1:700



7316/06-U-02

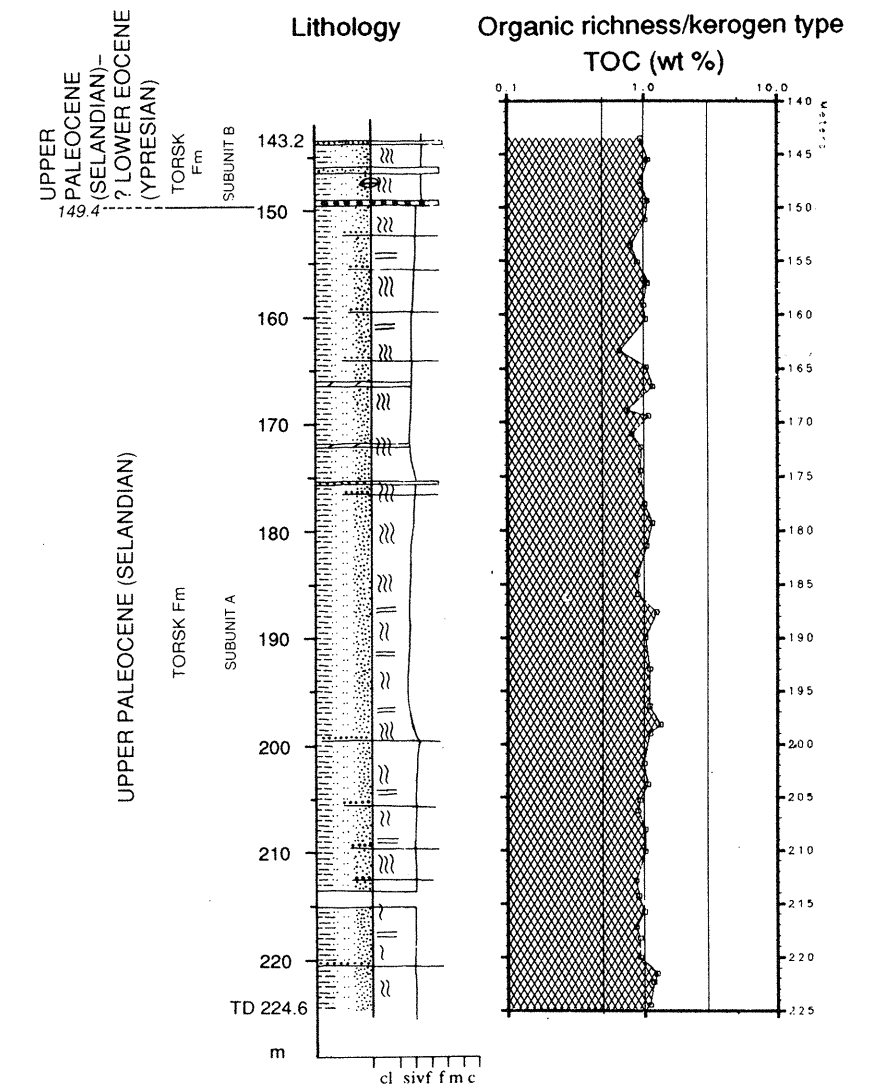
Age: Early – Middle Eocene



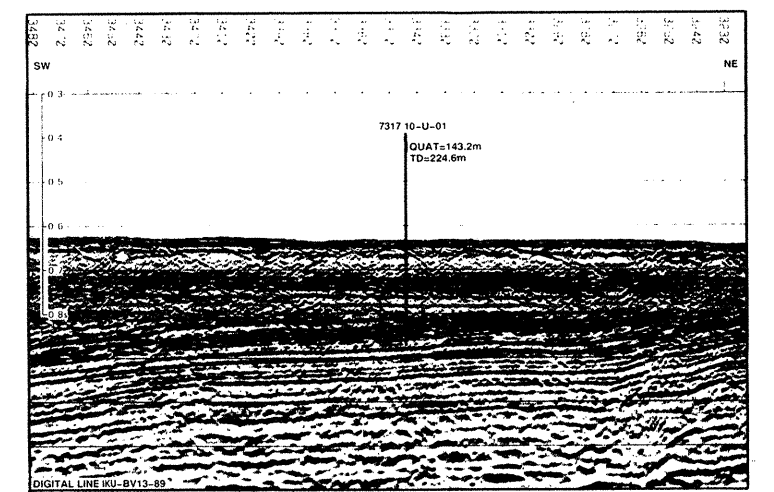
Vertical scale: 1:700

7317/10-U-01

Age: Late Paleocene – Early Eocene

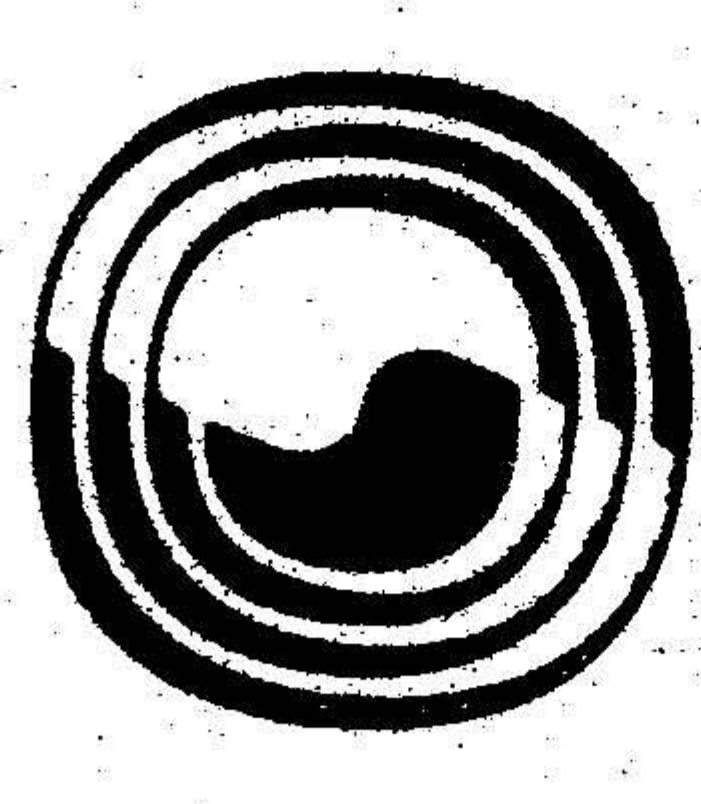


Vertical scale: 1:700



Shallow Drilling Bjørnøya West 1989

Main Report



SINTEF GROUP