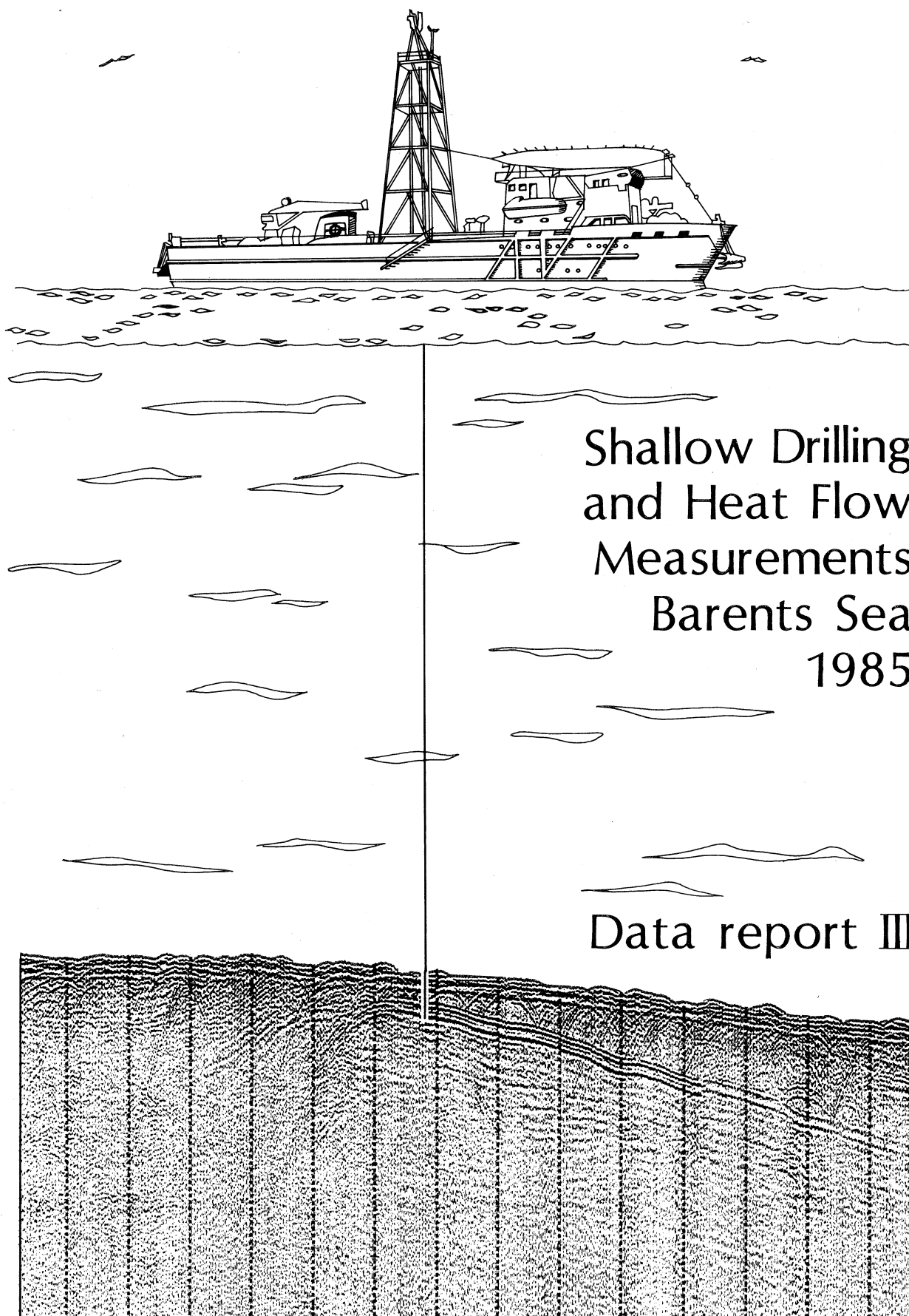




Shallow Drilling
and Heat Flow
Measurements
Barents Sea
1985

Data report III



INSTITUTT FOR KONTINENTALSOKKELUNDERSØKELSER OG PETROLEUMSTEKNOLOGI AS

Håkon Magnussons gt. 1 B • P.O. Box 1883, Jarlesletta • N-7001 Trondheim, Norway
Tel.: +47 7 92 06 11 • Telex: 55 434 iku n • Telefax: +47 7 92 09 24 (Aut.)

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SHALLOW DRILLING AND HEAT FLOW MEASUREMENTS
BARENTS SEA 1985. DATA REPORT III

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SUMMARY:

The project "Shallow seismic, shallow drilling and heat flow - Barents Sea 1985" was reported in two data reports (21.3405.00/01/85, 21.3405.00/02/85) in December 1985, one report for the shallow drilling and one for the heat flow measurements. In these two reports, further work on selected items was recommended. The present report gives the result, of this work presented in four chapters: revised palaeontology of cores D85-J and D85-N, diagenesis study of core D85-J, additional organic geochemistry analysis, and updated heat flow data.

Macrofossil studies and additional foraminiferal studies of cores D85-J and D85-N have given slightly revised and more accurate dating of these cores, and the relation between seismic sequence boundaries and lithological units has been changed because of this work. The diagenesis study was concentrated on the limestone unit in core D85-J and the reason for relatively high porosity but low permeability is explained. The organic geochemistry chapter presents further analytical results and some comments on kerogen type and relationships between cores. The heat flow chapter gives a comprehensive documentation for a revision for the heat flow values previously presented, and recommendation for development of the shallow probe method as a supplement to the down hole measurements.

KEY WORDS:	
Barents Sea	Diagenesis
Shallow drilling	Organic geochemistry
Heat flow measurements	Palaeontology

IKU Reports

21.3405.00/01/85: "Shallow Drilling Barents Sea 1985"

21.3405.00/02/85: "Heat Flow Measurements Barents Sea 1985"

21.3405.00/03/86: "Shallow drilling and heat flow measurements Barents Sea 1985. Data report 3."

21.3405.00/04/86: "Barents Sea mapping program 1984/85. Comparisons and correlations."

21.3405.00/05/86: "Shallow drilling Barents Sea 1985 Macrofossil dating of core D85-K."

ADDENDUM

The coreholes drilled in the Barents Sea in 1985 were given preliminary names by IKU, and these were used in all reports. The table shows these names along with the later assigned official NPD names and the final coordinates and water depths.

IKU name	NPD name	Latitude	Longitude	Water depth (m)
C85-A	7126/06-U-01	71° 38' 49.795" N	26° 52' 43.295" E	369
D85-H	7220/02-U-01	72° 45' 3.095" N	20° 34' 45.644" E	429
C85-E	7222/09-U-01	72° 24' 19.739" N	22° 47' 39.787" E	322.5
D85-V	7317/02-U-01	73° 49' 10.924" N	17° 22' 4.526" E	296
D85-I	7317/02-U-02	73° 47' 8.062" N	17° 23' 35.505" E	324
D85-J	7320/03-U-01	73° 51' 15.18" N	20° 55' 0.802" E	419
D85-K	7323/07-U-10	73° 16' 51.539" N	23° 1' 25.834" E	420
D85-W	7323/10-U-01	73° 13' 15.87" N	23° 18' 42.716" E	380
D85-N	7425/09-U-01	74° 29' 25.296" N	25° 46' 19.261" E	334
D85-O	7524/12-U-01	75° 12' 20.536" N	24° 43' 21.607" E	108

The coreholes drilled in the Barents Sea in 1984 will get official NPD names in 2008.

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1 PREFACE

This project is supported by the following companies:

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Fina Exploration Norway
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Their participation is greatly acknowledged.

The present report includes the results of the work carried out since the first two data reports of December 1985 (IKU reports 21.3405.00/01/85 and 21.3405.00/02/85).

The work has been carried out by:

Paleontological studies: Robert M. Goll, Simon R.A. Kelly, private consultant, Nils Aarhus.

Diagenesis study: Geir Elvebakk, Hans Rendall.
Lab. work: Tony Boassen, Øystein Rossing,
Unni M. Wang.

Organic geochemistry: Thomas L. Leith, Liv Schou.
Lab. work: Anne Due, Leif Husvik,
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Heat flow measurements: Joar Sættem, Martin Hamborg, Øistein Johansen.
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Calculation of effects from temperature
history models: Niels Balling, University of
Aarhus, Denmark

Drawing: Tone Aas, Rolf Angell-Petersen,
 Kirsti Bakken, May Britt Dybvik,
 Randi Soligard,

Typing: Anne Larsen, Lina Lauvsnes, Wenche Norum,
 Mette A. Sagmo, Ellen Solberg

Geotechnical investigations were carried by the Sintef Division of Geotechnical Engineering, comparisons of thermal conductivity by Geological Institute at Chalmers Technical University, Gothenburg, Sweden and analysis of porosity and permeability by Geco Petroleum Laboratory.

Robert M. Goll has integrated the new palaeontology data and added further foraminiferal work in Chapter 3. He also improved the English language for parts of the report. Tom Bugge co-ordinated the project and edited the present report together with Atle Mørk who also acted as a consultant during the laboratory work.

2 INTRODUCTION

This is the third data report in the project "Shallow seismic, shallow drilling and heat flow Barents Sea 1985". The first two reports were completed in December 1985; "Shallow Drilling Barents Sea 1985" (IKU report 21.3405.00/01/85) and "Heat Flow Measurements Barents Sea 1985" (IKU report 21.3405.00/02/85). This report gives the result of further analyses and studies and should be regarded as a supplement to the first two reports. All basic information on the field work, the cores and the data obtained are given in these reports.

On cores D85-J and D85-N macrofossil studies and additional foraminiferal studies have been carried out. Dr. S.R.A. Kelly, private consultant, UK, carried out the macrofossil work. Based on this and on his own foraminiferal studies, R.M. Goll revised the palaeontology chapters previously presented for cores D85-J and D85-N. Chapter 3 in this report are therefore updated and revised versions of the Chapters 3.4.1, 3.4.6, 3.6.1 and 3.6.6 in IKU report 21.3405.00/01/85.

In the first data report (21.3405.00/01/85), it was recommended that further work should be carried out on the complex diagenetic history of core D85-J. The result of this study is presented in Chapter 4. It is concluded that the limestone unit has high porosity, but very low permeability and the reasons for this are explained.

Additional organic geochemical analyses are discussed in Chapter 5 and are mainly concerned with a re-evaluation of the kerogen types previously described using Rock-eval pyrolysis, comparison of the extractable material in cores D85-K and D85-O, comparison of the catagenic hydrocarbon gases in cores D85-J and D85-V and a more detailed study of the extractable material in core D85-J.

Comprehensive investigations of the heat flow data showed that an incorrect calibration factor was used as input to computer calculation of thermal conductivity. Theoretical calculation of thermal conductivity indicates that the previously presented field data is too low by a factor of 1.81. The heat flow consultant, Omegalink, determined that the correction factor should be 1.799, and this value is used to correct all field data. New laboratory measurements of the core

material supports this result. All data from this work is presented in Chapter 6. An appendix to this chapter recommends further development of the shallow probe technique ($<10\text{cm}$) as a supplement to the down hole measurements.

The report is divided into separate chapters with separate authors.

CHAPTER 3



PALAEONTOLOGICAL STUDIES CORES D85-J and D85-N

Based on IKU report 21.3405.00/01/85

Compiled by Robert M. Goll

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3.1 Introduction

Cores D85-J and D85-N each penetrate separate regional seismic reflectors, both of which were previously believed to represent unconformities related to Late Kimmerian tectonism. At locality D85-N, the older reflector is too deep to have been penetrated by drilling. However, the younger reflector is represented in the recovery of core D85-N. At locality D85-J, the two reflectors have merged to form a single reflector, which is also represented in the recovery of core D85-J. In our previous report, we concluded that the reflector at D85-J must result from the lithologic contact between units B and C. New sedimentological and paleontological data indicate the presence of a major sequence boundary in core D85-J between units A and B, with Valanginian dolomite (unit B) overlying Kimmeridgian- Volgian siltstone (unit A). Our new interpretation is that the seismic reflector at locality D85-J, which represents two phases of eustatic disruption, coincides with this sequence boundary (the unit A/unit B contact, Chapter 3.2.2).

The precise location of the younger reflector in the recovery from core D85-N is unclear at the time of this writing and will be the subject of more intensive study in a subsequent report (Chapter 3.3.2).

Due to time constraints, macrofossil studies were not presented in the first data report. A variety of invertebrate fossils are abundantly preserved in portions of the recovery from these cores, however, and some, particularly Buchia, have important biostratigraphic value. Consequently Dr. Simon R.A. Kelly, a U.K. macrofossil specialist, was engaged to carry out a thorough examination of cores D85-J and D85-N. Additional studies of benthic foraminifera have been completed by R.M. Goll, IKU, who has also revised the summary and compiled the paleontology chapters on these two cores to include these new findings.

The revised chapters are presented below.

3.2 Core D85-J

3.2.1 Summary, core D85-J

On site D85-J, the overburden was only 2.12 m. About 34.1 m of pre-Quaternary rock was cored, which is a recovery of nearly 100%.

Two main units are represented in the cored interval, with the unit A/unit B boundary at 33.3 m. This is a seismic sequence boundary and appears as a well defined seismic reflection. The lowermost unit is referred to as A. The uppermost unit is subdivided into two sub-units, B and C, 33.3 - 30.0 m and 30.0 - 2.1 m below sea floor respectively.

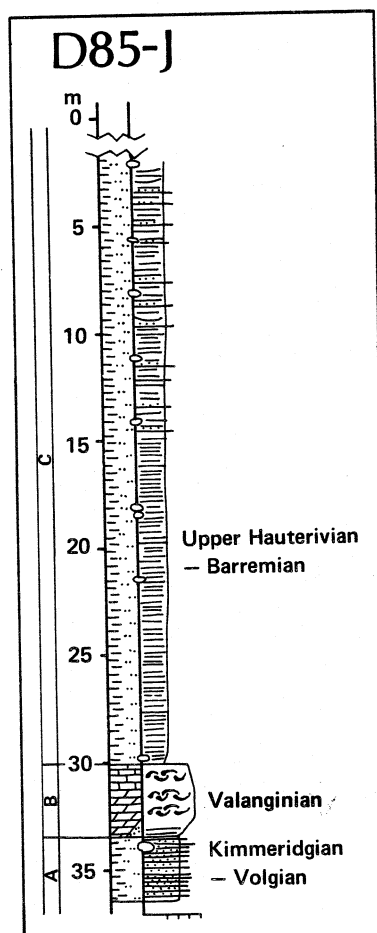


Fig. 3.2.1 Simplified lithological column of core D85-J.

Lithology. Unit A consists of a grey clayey to sandy siltstone. It has a sharp boundary to the overlying unit B (Fig. 3.2.1) which consists of

dolomitized carbonate with abundant fossil fragments. Unit C is lithologically similar to unit A. It consists of dark grey silty claystone to clayey, sandy, siltstone. The sedimentary environment is interpreted as a stormy shallow shelf. Both A and C are coarsening upwards units.

Dating. Palynomorphs, macrofossils and microfossils recovered from units A-C range in age from probably Kimmeridgian - Middle Volgian at the base to Early Hauterivian - Barremian at the top. Sparse dinoflagellate assemblages provide the only age control for unit A, which can be assigned with only moderate confidence to the Kimmeridgian - Middle Volgian. There is a modest discrepancy in the age interpretations of the bivalve and calcareous nannofossils assemblages of unit B, but Valanginian seems to be the most reliable age for this unit. Diverse dinoflagellate floras in unit C provide a confident assignment of Upper Hauterivian - Barremian for these silty mudstones.

Seismic reflection studies indicate a sequence boundary between units A and B. It was earlier interpreted to be between units B and C, but due to new palaeontological data and the diagenesis study, we are now able to redefine the exact location of this sequence boundary. We believe that it is the regional Late Kimmerian unconformity mapped elsewhere in the area.

Organic geochemistry. Unit A shows good to very good organic richness with an average TOC content of 2.7 wt%. The kerogen in this unit appears to be of type II/III on the basis of pyrolysis-gas chromatography data. The vitrinite reflectance values of 0.58% and 0.65% suggest a thermal maturity of very early "oil window".

Unit B is organic lean with an average TOC content of 0.2 wt%. One sample from this unit was frozen in liquid nitrogen on board the drill ship. Analysis of gas from this sample suggests a fairly immature gas.

The organic richness of unit C is fair to very good with an average TOC content of 3.2 wt%. Pyrolysis data for this unit indicate an upwards change in kerogen type from a type III kerogen at the base to a more hydrogen rich, type III/II kerogen.

3.2.2 Palaeontology, core D85-J

Age determinations for D85-J have been based on dinoflagellates, foraminifera, molluscan bivalves and calcareous nannofossils. The occurrences of these fossils are mutually exclusive in that palynomorphs were not recovered from the carbonate interval (unit B), which is the only part of the section where calcareous fossils were found.

36.20-33.50 m (unit A) Kimmeridgian - Middle Volgian

This interval cannot be confidently dated at this time. One poorly-preserved specimen of a dinoflagellate that became extinct in the basal Middle Volgian has been observed, and its occurrence suggests that unit A is no younger than this stage. Other dinoflagellates found in unit A have affinities to species recorded in Kimmeridgian sediments of Greenland, England and Europe. These lines of evidence are somewhat tenuous, however, and the reader is cautioned not to infer too much from the apparent hiatus between units A and B.

Palynology

A very poorly preserved specimen of Gonyaulacysta jurassica at 33.50 m, points to an earliest Middle Volgian (or older age) for the top of this dark siltstone interval (Fig. 3.2.2). Most samples contain Tubotuberella apatela, Paragonyaulacysta borealis and Sirmiodinium grossii. Similar assemblages may be found in Late Oxfordian - Ryazanian strata but are most characteristic for Kimmeridgian - Volgian beds. Additional more age significant taxa normally appear in Oxfordian and Ryazanian.

Some of the specimens of T. apatela bear a few very small spines at plate boundaries and may point to an age in the Jurassic part of its range. Gonyaulacysta sp. occurs fairly commonly both at 36.20 and 35.73 m. This form is closely related to those described and figured by Gitmez (1970) as Gonyaulacysta cf. helicoidea and Gonyaulacysta hyaloderma from the Kimmeridgian Liostrea delta Bed in Normandie. At 36.20 m we have also recorded several specimens of her basal Kimmeridgian Gonyaulacysta cf. mamillifera. Tasmanitids are common at 36.20 m and abundant at 35.73 m, whereas Cribroperidinium granulatum is

common at 33.50 m. This species may be referred to a form complex also including Cribroperidinium globatum and Cribroperidinium granuligerum. Gitmez & Sarjeant (1972) described C. globatum from the wheatleyensis - pectinatus Zones in Surrey, England. Piasecki (1980) recorded C. granuligerum from earliest Kimmeridgian and higher beds in East Greenland and used it as a stratigraphic marker. In western Europe others have found species of Cribroperidinium also in older strata.

33.30-29.95 m (unit B) Valanginian

Only the calcareous nannofossil and bivalve assemblages contain age-diagnostic constituents. The occurrence of two species of the bivalve Buchia provide a confident age of Valanginian for unit B. Late Hauterivian to Early Barremian is inferred for rare and poorly preserved calcareous nannofossils, but this younger age is somewhat less firmly established than the bivalve evidence.

Palynology

Samples at 33.00 and 31.85 m from this limestone interval were barren with respect to palynomorphs.

Macrofossils

Thirteen specimens of the bivalve Buchia from unit B are the only macrofossils identified in core D85-J. Of these 5 specimens could be identified only to the genus level. Five specimens of Buchia (B.) keyserlingi (Trautschold) were found in the interval 31.37-32.70 m and indicate a late Early Valanginian age. Late Valanginian to possibly Early Hauterivian is indicated for the interval 30.20-31.12 m by the presence of 3 specimens of Buchia (B.) sublaevis (Keyserling).

The above evidence suggests that unit B is a condensed section and includes sediments that differ in age by as much as 6 million years.

Foraminifera

One sample at 29.90 m yielded very rare arenaceous foraminifera and some Inoceramus prisms and "fish" teeth. One single arenaceous

foraminifer was met at 35.22 m. All other samples are barren. The foraminifera are characteristic in Early Tertiary, Cretaceous and Late Jurassic sediments with restricted arenaceous foraminiferal faunas.

Nannofossils

Five samples yielded very rare nannofossils, all damaged. Two forms occur repeatedly and their identification is reasonably well supported. Watznaueria biporta (29.98, 29.99, 30.04 m) suggests an age not earlier than Hauterivian. Circular discs of eight irregular segments occur in samples from 31.85 m and 32.33 m together with Braarudosphaera fragments at 31.85 m. Thus, it may be reasonable to assume some affinity to the Late Hauterivian - Early Barremian from the Sklinnabanken, where similar nannofossils occur abundantly in one restricted interval.

29.95 - 0.00 (unit C) Upper Hauterivian - Barremian

Palynology

Relatively diverse Hauterivian-Barremian dinocyst assemblages occur in the claystone from immediately above the limestone to the top of the core. Muderongia tetracantha occurs consistently in samples up to 23.45 m and more sporadically above. According to Davey (1974, 1979) it is restricted to Late Hauterivian and Early Barremian beds. Its co-occurrence with Muderongia simplex at 25.44 m may suggest an age no younger than C4 (middle Late Hauterivian) for the section up to this level, if the ranges from the Speeton Clay are applicable to this area. It may be noteworthy that Muderongia staurota is recorded only at 29.95 m, where it is common. Duxbury (1977) reported its first occurrence in the Upper Hauterivian slightly above the incoming of Chlamyphorella trabeculosa in upper Lower Hauterivian C9 beds. In core D85-J C. trabeculosa is first recorded at 28.94 m, and it has 3 other occurrences higher in the core.

Aptea anaphrissa occurs in most samples from 27.44 m to the top (common in the lower part). Duxbury (1977, 1980) and Davey (1979) found it in uppermost Lower and in Middle Barremian strata but in more northerly areas it may have a longer range.

Duxbury (1977) claimed that Nelchinopsis kostromiensis is restricted to Hauterivian beds. Its recovery here only at 29.95 and 28.94 m may seem to support this, but its occurrence at 49.72 m in core D85-N suggests that it may range into the Barremian.

Reworked material from Permo-Triassic, Lower-Middle Jurassic, Late Callovian-Oxfordian and Late Volgian-Ryazanian strata points to a complicated geological history in the area. Colour estimates on pollen (TAI 2- 2-/2) at 18.46 and 16.44 m are higher than in the shaly interval below the limestone.

G98/a/44/mas

3.3 Core D85-N

3.3.1 Summary, core D85-N

Core D85-N is a total of 64.38 m long, of which the upper 49.9 m is Quaternary overburden. The basal 14.5 m consists of Cretaceous bedrock that has been subdivided lithologically into four units, A-D, from oldest to youngest (Fig. 3.3.1). A prominent regional seismic reflector must be represented in the recovery from this core, but its exact location is unknown at the present time.

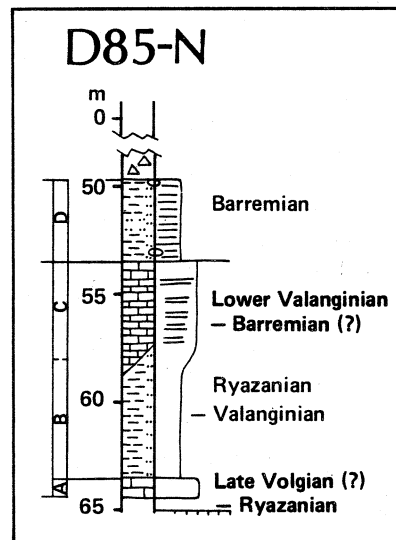


Fig. 3.3.1 Simplified lithological column of core D85-N.

Lithology. Unit A (64.38 - 63.68 m) consists of a greyish green limestone (wackestone) with abundant bivalve fragments. The boundary to unit B (at 63.68 m) is sharp. B consists of reddish brown silty fossil rich claystone. Its upper boundary is a gradational one. Unit C (58 to about 53.5 m) consists of limestone (wackestone). The lower 2.5 m is light grey to greyish green while the upper 2 m is dark grey to black.

3.3.2 Palaeontology, core D85-N

Age determinations for core D85-N have been based on dinoflagellates, benthic foraminifera, and the pelycepod Buchia. The latter two groups provide biostratigraphic control for units A-C, and dinoflagellates are the only age indicators for unit D.

64.38 - 63.68 m (unit A)

The age of this unit is probably Ryazanian, but it possibly may be as old as latest Volgian.

Palynology

Of the three samples from this unit that were examined for palynomorphs, one was barren (Fig. 3.3.2). The other two contained a very sparse flora that can be interpreted only as of general Early Cretaceous age.

Macrofossils

This bioclastic mudstone contains dense clusters of Buchia, many of which appear to be fairly complete. There is some evidence of current sorting, however. Two specimens of Buchia (Buchia) cf. unschensis (Pavlow) were found at 63.89 and 64.30 m (Fig. 3.3.3) and suggest a late Late Volgian to Early Ryazanian age. This interval is most likely Early Ryazanian, but a Late Jurassic age cannot be disregarded.

Foraminifera

Core D85-N was searched extensively for calcareous microfossils. Samples were prepared in the laboratory using three standard techniques: dilute peroxide, Glauber's salt, and dilute sodium hydroxide followed by freezing. Each technique was effective in disaggregating the sediments, but sparse foraminifera were the only microfossils encountered. Because of the variety of procedures followed, it is concluded that the paucity of microfossils cannot be attributed to artificial dissolution. On the contrary, it must be concluded that foraminifera are poorly preserved in these cores.

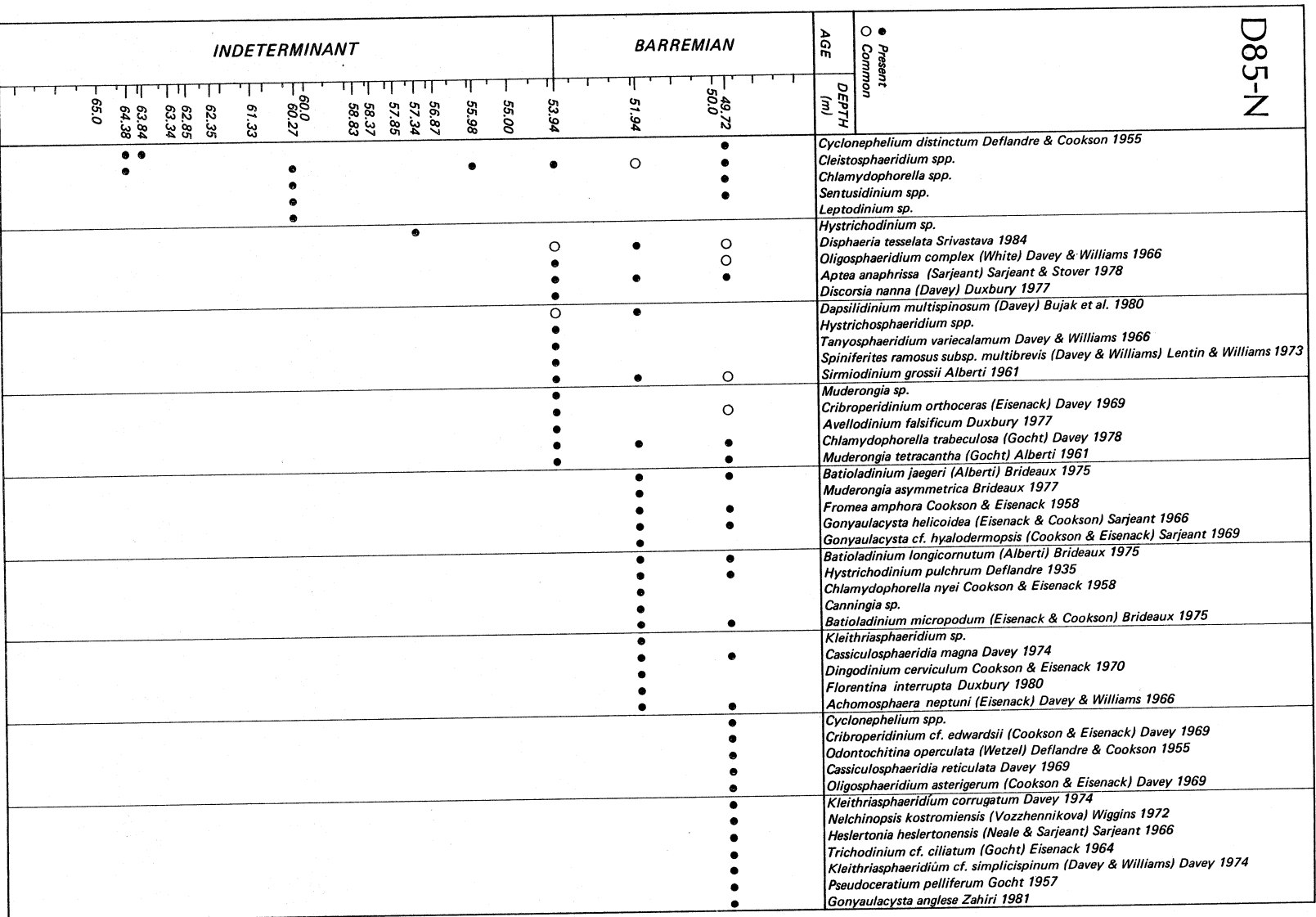


Fig. 3.3.2 Palynological range chart core D85-N.

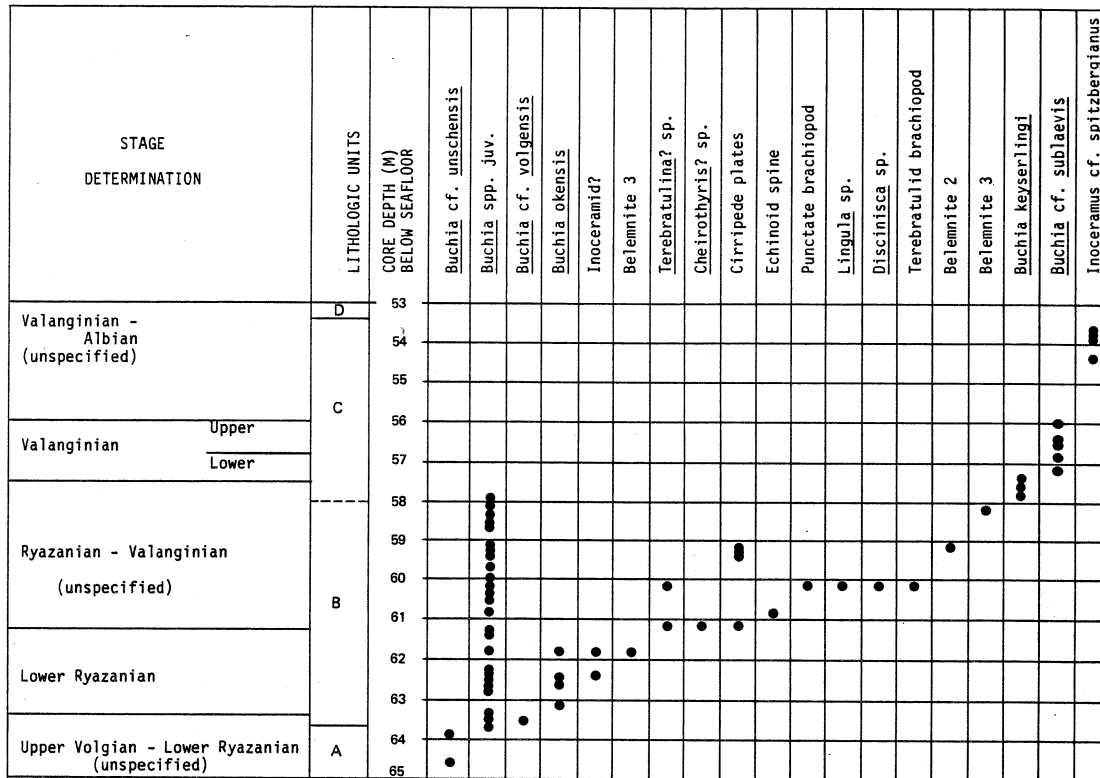


Fig. 3.3.3 Macrofossil occurrences in D85-J identified by S.R.A Kelly, 1986.

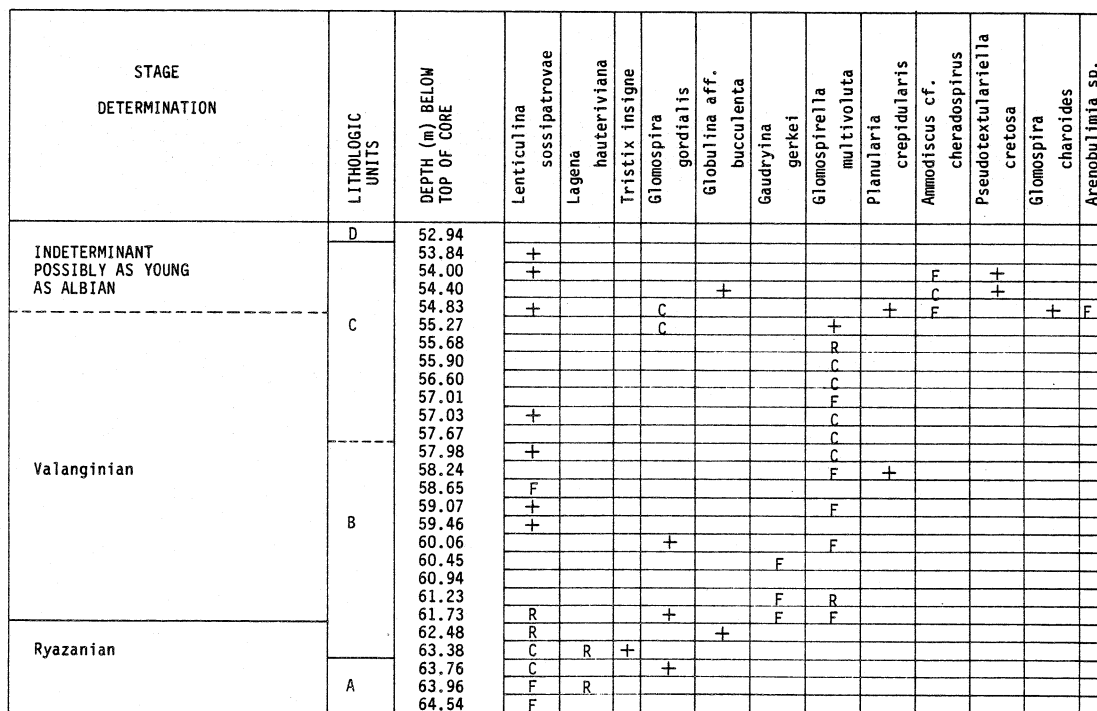


Fig. 3.3.4 Benthic foraminifera in D85-N.

Three samples from unit A at 63.76 and 64.54 m were examined for calcareous microfossils (Fig. 3.3.4). They contained only rare specimens of Lenticulina sossipatrovae Gerke and Ivanova, Lagena haueriviana Bartenstein and Brand and Glomospirella multivoluta Romanova as well as single specimen occurrences of Glomospira gordialis (Jones and Parker) and Tristix insigne Reuss.

L. sossipatrovae is diagnostic for the L. sossipatrovae Assemblage Zone of the Rurikfjellet Member, Janusfjellet Fm, Svalbard, which Nagy and Løfaldli (IKU report 04.6264/1/84) correlated to the Ryazanian of the Russian Platform. L. haueriviana ranges from Upper Ryazanian to Lower Barremian, according to Hart et al. (1981).

63.68 - ca. 58.0 m (unit B) Ryazanian - Valanginian

The bivalve and foraminiferal evidence for the age of unit B are in approximate agreement, although there is some discrepancy in the placement of the Ryazanian - Valanginian boundary. Foraminiferal biostratigraphy indicates that this boundary should be placed at ca 62 m, whereas bivalve distributions indicate no unequivocal Valanginian strata in unit B. Bivalve occurrences may be facies controlled, however, as the Buchia marker species has its first appearance above the limestone/silty mudstone contact.

Palynology

Eight samples in this unit were examined for palynomorphs, and all but one were barren (Figure 3.3.2). As was the case for unit A, the one remaining sample contained only very rare dinoflagellates without precise biostratigraphic significance.

Macrofossils

The red mudstones of unit B appear to have been moderately bioturbated, although the identified bivalves include only epifaunal or reclining filter feeders. There does not appear to be any evidence for current sorting, but all bivalves examined are disarticulated. The Buchia fauna is dominated numerically by juveniles having clearly preserved prodissoconchs.

Macrofossils found in unit B are tabulated in Fig. 3.3.3. The common occurrence of barnacle plates between 59.28 and 61.05 m is an interesting addition to the fauna. All the brachiopods and belemnites in unit B are isolated occurrences.

Age-diagnostic macrofossils are restricted to Buchia (Buchia) okensis (Pavlow), which occurs between 61.83 and 63.02 m, and Buchia (Buchia) cf. volgensis (Lahusen). B. okensis is most typical of the Early Ryazanian, where it usually occurs in association with the zonal index species, Hectoroceras kochi. Thus, the interval of its occurrence in unit B is firmly dated as late Early Ryazanian. B. volgensis, a more widespread Ryazanian indicator than B. okensis, becomes more abundant in the Late Ryazanian, but it also occurs in the Early Ryazanian.

Foraminifera

Twelve samples from unit B contain microfossils in the size range 62-500 μ m (Figure 3.3.4). Only sparse occurrences of benthic foraminifera were identified in these samples. Lenticulina sossipatrovae and Glomospirella multivoluta continue to be the dominant species, but their occurrences are sporadic in this sample set. Gaudryina gerkei is present in three of the samples near the base of unit B. Other species with very rare occurrences are: Glomospira gordialis, Lagena hauteriviana, Globulina aff. bucculenta Nagy and Planularia crepidularis. Unit B is correlative with the L. sossipatrovae and G. multivoluta Assemblage Zones of Nagy and Løfaldli. The first appearance of G. multivoluta is regarded as a marker for the Valanginian Stage, because this species is known only from the Valanginian of the Russian Platform.

Ca. 58.0 - 53.5 m unit C Lower Valanginian - Barremian, but possibly as young as Albian

Bivalves and foraminifera are the only age-diagnostic fossils recovered from the base of unit C, and one palynological sample near the top gives an age of Barremian. There is weak evidence, based on both macrofossils and foraminifera, that the upper part of unit C may be as young as Albian, but Barremian appears to be the most well-founded age.

Palynology

Six samples were examined for palynomorphs, and the lower 5 samples were either barren or contained only rare undiagnostic specimens, as was the case for samples from units A and B (Fig. 3.3.2).

Palynomorphs in sample from 53.94 m are not as diverse as samples from unit D, but they indicate a Barremian age as will be discussed in the chapter on the latter unit.

Macrofossils

The lithology ranges from a pale limestone at the base with many highly fragmented blocks of prismatic inoceramid shell to a dark, organic rich mudstone with some more complete inoceramid valves as well as fragments near the top. It is not clear whether the fragmentation has been caused by bioturbation or by current activity. None of the bivalves appeared to be in articulation and the large size (< about 8 cm) may suggest current reworking. The shell fragments would have impeded general small bivalve burrowing activity.

The base of unit C, which is transitional over an interval of 3 m, coincides approximately with the first occurrence of well-established Valanginian age indicators (Fig. 3.3.3). Buchia (Buchia) keyserlingi (Trautschold) occurs from 57.37-57.62 m. The species is a typical boreal Valanginian indicator, most abundant in the Lower Valanginian. The interval is firmly dated as mid to late Early Valanginian. Buchia (B.) cf. sublaevis (Keyserling) occurs from 55.95-57.10 m. The interval is dated as Upper Valanginian to possibly Early Hauterivian.

The mutually exclusive ranges of these two species over a depth range of ca 1.7 m in core D85-N is similar to their distribution in unit B of D85-J and indicates that these two carbonate units are coeval.

The top of the occurrence of B. cf. sublaevis marks the upper limit of firm macrofossil biostratigraphic control in unit C. Inoceramids occur throughout the core, but between 53.78 and 54.34 m, much larger shells of this group are common and are characterized by fragmented prismatic shell. This shell material is identified as Inoceramus aff. spitzbergianus Stolley. The specimens originally described by Stolley

were in association with crioceratid ammonites in the "Ditrupa Sandstein" (Carolinefjellet Formation) of Adventfjorden, Spitsbergen. The age was believed to be from Hauterivian to Aptian. The borehole specimens are only loosely identified with this species and the age cannot be more precisely given at this stage. Pchelina (1965, translation, p. 177) has recorded "I. spitzbergensis" from the Albian of Van Keulenfjord, Spitsbergen.

Foraminifera

The distribution of foraminifera in 11 samples from unit C are given in Fig. 3.3.3. The extension of the range of Lenticulina sossipatrovae to the top of unit C suggests that the entire unit is Valanginian, because this species is not reported for younger sediments in Siberia. Its occurrence is so sparse and sporadic above 57.03 m, however, that it is not regarded as age diagnostic higher in the section. The rather abrupt termination of the occurrence of Glomospirella multivoluta at 55.27 m is taken as the top of the Valanginian Stage, because of its consistent and common occurrence in D85-N. The distributions of these species in D85-N are similar to the L. sossipatrovae and G. multivoluta Assemblage Zones of the Rurikfjellet Member of the Agardhfjellet Formation, Svalbard.

Above 55.27 m in D85-N, however, similarities of the foraminiferal assemblage are not apparent. Glomospira charoides, which increases markedly in frequency in Rurikfjellet Member in the Agardhfjellet section, has been regarded only as a single trace occurrence in unit C.

53.50 - 49.72 m unit D Barremian

Palynology

Diverse dinocyst assemblages are found in two samples from this interval (Fig. 3.3.2). The presence of Sirmiodinium grossii, Pseudoceratium pelliferum, Heslertonella heslertonensis, Batioladinium longicornatum, Disphaeria tessellata, Aptea anaphrissa, Gonyaulacysta

anglese and Kleithriasphaeridium corrugatum together with Odontochitina operculata indicates a Barremian age. Muderongia tetracantha, Nelchinopsis kostromiensis and particularly B. longicornutum point to an age in the lower part of the Barremian. Some authors use N. kostromiensis as a marker for the Hauterivian, but our evidence (if not reworked) contradicts its usefulness as such in this area. Due to the recovery of Batioladinium jaegeri, Chlamydophorella nyei and Florentinia interrupta in addition to O. operculata, these samples may possibly be located somewhat higher in the sequence than the interval 29.95 - 0.0 m in core D85-J.

Macrofossils

none observed

Foraminifera

none observed

CHAPTER 4

DIAGENESIS STUDIES CORE D85-J

Remineralization and porosity relationships
in Lower Cretaceous carbonates in core D85-J
from the Barents Sea

by

Geir Elvebakk and Hans Rendall

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Summary

One of the seven cored sections (D85-J) recovered during IKU's 1985 drillship expedition to the Barents Sea includes a 3.3 m thick interval of partially dolomitized and ankeritized wackestone (unit B) of Valanginian age. Standard petrographic, cathodoluminescence and SEM microscopy as well as electron microprobe analyses have been conducted in order to determine the diagenetic history of these carbonates, which are separated from the underlying silty mudstones by a probable hiatus of at least 6-8 million years.

Ankerite and dolomite are the dominant carbonate minerals in the lower 2 m of unit B, while calcite predominates the upper part. Compactional squeezing of pore water from the underlying mudstones was the major source of the allogene ions.

The diagenetic sequence of events was: drusy calcite cement, porefilling dolomite cement, pseudomorphic replacements by dolomite, xenotopic ankerite, microcrystalline siderite cement and pyrite. Mouldic porosity was formed during the dolomitization and ankeritization reactions by selective dissolution of fossil fragments. Some vuggy porosity was formed during maturation of organic material.

During shallow burial, slightly diluted pore water derived from mudstone compaction entered unit B and caused precipitation of pore filling calcite and dolomite as well as dolomitization of microcrystalline calcite and fossil fragments. An increase in porosity occurred during this phase. Xenotopic ankerite crystallization occurred in the lower part of unit B, when burial depths reached the oil window. This second diagenetic phase was initiated by acidic pore water resulting from kerogen maturation, which also supplied the dissolved sulphur required for pyritizations. Iron probably was sourced by clay mineral diagenesis in the underlying mudstone.

Permeability is extremely low for the whole unit B, and although the middle part of the unit shows high porosity (moldic) the lack of communication between the pores gives this unit no value as a reservoir rock.

4.1 Introduction

In this chapter we present a detailed diagenetic study of core D85-J (from the project: "Shallow seismic, shallow drilling and heat flow - Barents Sea 1985"). The bedrock drilling was carried out in July-August 1985. The seismic and bedrock geology data of the project were presented in IKU report 21.3405.00/01/85, and the heat flow measurements were reported in report 21.3405.00/02/85.

Core D85-J consists of two silty claystone units (A and C) separated by 3.3 m of dolomite/wackestone (unit B, Fig. 4.2) that is the principal subject of this report. The silty claystone units are interpreted as storm shelf deposits, and unit B may represent a partly condensed section (see IKU report 21.3405.00/01/85).

Biostratigraphic datings of D85-J have been made on the basis of: palynomorphs, nannofossils and bivalves. The lower silty claystone (unit A) is of Kimmeridgian-Middle Volgian age, and the upper silty claystone (unit C) is regarded as Upper Hauterivian-Barremian. The best estimate of the age of the dolomite/wackestone unit B is Valanginian.

There are two main diagenetic problems associated with the dolomite/wackestone unit: the distribution of dolomite/calcite, and the fairly large amount of mouldic porosity. The main objectives of this report are, therefore, to investigate the dolomitization reactions and determine whether the limestone unit has been subjected to freshwater diagenesis.

Thin section analyses were carried out to solve these diagenetic problems. These thin sections were subjected to optical, cathodoluminescence, and scanning electron microscopy as well as electron microprobe analyses. In addition, some samples were studied by X-ray diffraction (XRD) analysis. The cathodoluminescence studies were only moderately successful, because the minerals only showed dull luminescent colours.

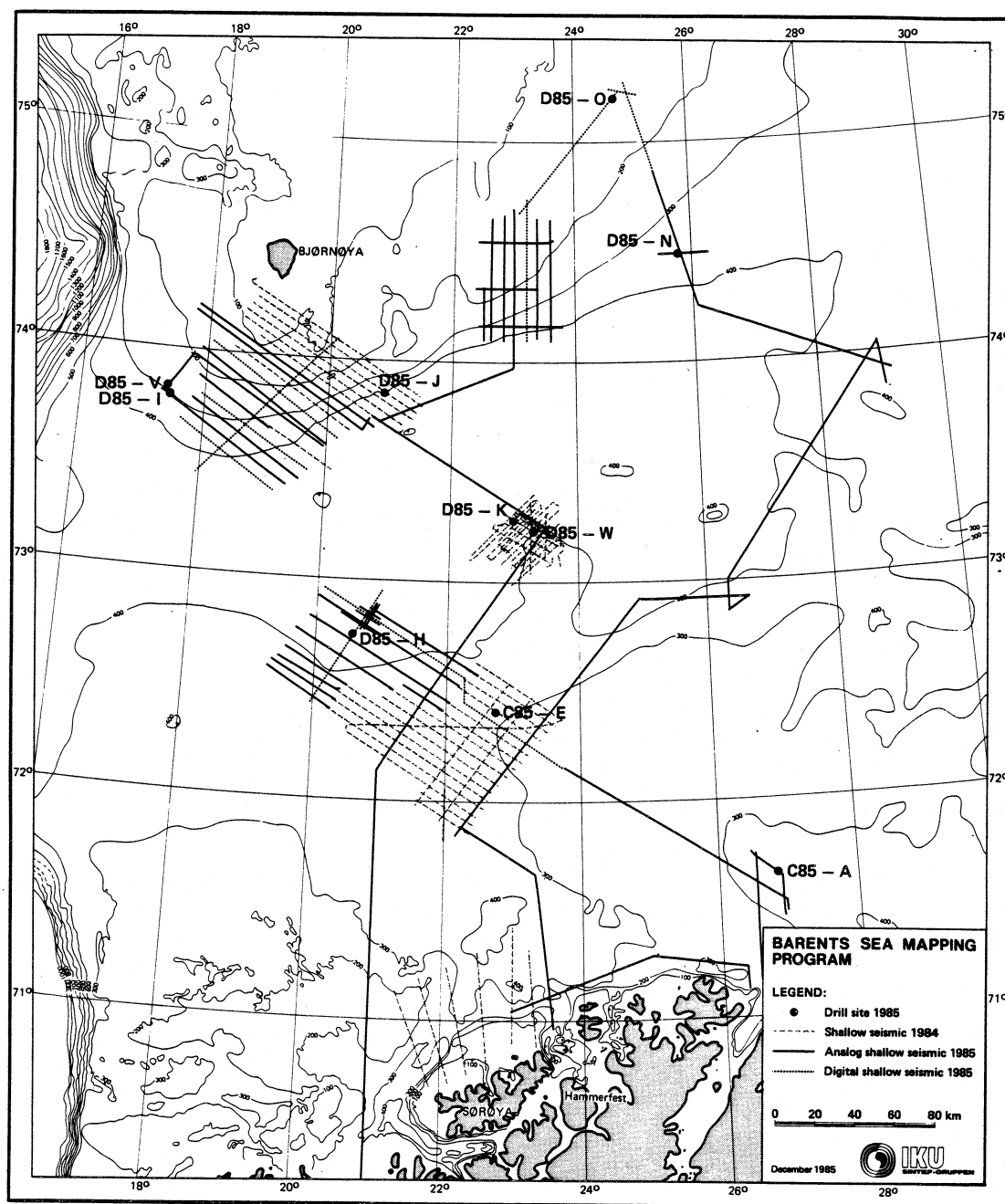


Fig. 4.1 General map of the survey area showing the location of core D85-J.

4.2 Methods

A short description of the various instruments and techniques that have been used in this study are outlined below.

Macroscopic descriptions of the samples, supplemented by a stereo microscope, were performed in order to relate the early diagenetic minerals to the host rocks.

Thin section microscopy: This investigation was undertaken on a Leitz Ortholux polarizing microscope.

Cathodoluminescence microscopy: A Nuclide Luminoscope was mounted on the above microscope for these investigations, which were performed under the following conditions:

- The vacuum chamber was filled with helium-gas with pressure in the range 150-250 millitorr.
- The beam current was in the range of 0.7-0.9 mA and the voltage in the range of 12-18 kV.

Scanning electron microscopy (SEM) and electron microprobe analysis:

This work was undertaken on a Jeol 733 Superprobe with a Tracor Northern energy dispersive (EDS) and four Jeol wave length dispersive (WDS) X-ray spectrometers as well as detectors for secondary and backscattered electrons and cathodoluminescence. The samples were carbon coated polished thin-sections.

Scanning electron microscope photographs presented in this report are back-scattered electron images (BEI). Consequently, the grey shade of a mineral phase is dependant on the mean atomic number of the mineral, i.e. the "heaviest" mineral will be light grey or white, while porosity will be dark grey or black.

The quantitative analyses were undertaken with a defocused electron beam (10 μm) to avoid migration of the lightest elements.

X-ray diffraction analysis (XRD): The XRD analyses were performed on bulk samples crushed in a mortar, suspended in water and sucked onto a Millipore filter. Qualitative and semi-quantitative mineralogy was then performed on a Phillips diffractometer calibrated as follows:

Preparation of bulk samples for XRD analysis involves crushing with a mortar suspension of the crushed rock in water, followed by entrapment of the coarse suspended fraction in a Millipore filter.

Radiation: monochromatic CuK_{α}
 Voltage: 40 kV
 Current: 30 mA
 Speed: 2° /minute, from 3° - 35° (2θ)

The semi-quantitative results are based on an area calculation (height x width, at half maximum height) multiplied by the weight factors given below, which have been experimentally determined at IKU by mixing minerals in different proportions.

Table 4.1 XRD weight factors used by IKU

Mineral	Weight factor	d-value ($\text{\AA}$)
Quartz (Qtz)	1.00	4.26
Alkalifeldspar (K-spar)	0.50	3.24
Plagioclase (Plag)	0.50	3.18
Illite (Ill)	1.40	10
Kaolinite (Kaol)	0.70	7
Chlorite	0.70	4.74, 7 and 14
Smect.	0.35	14
Mixed layer	0.55	10-14
Calcite (Calc)	0.25	3.04
Dolomite (Dol)	0.20	2.89
Siderite (Sid)	0.25	2.79
Apatite (Ap)	0.30	2.80
Pyrite (Py)	0.70	2.71

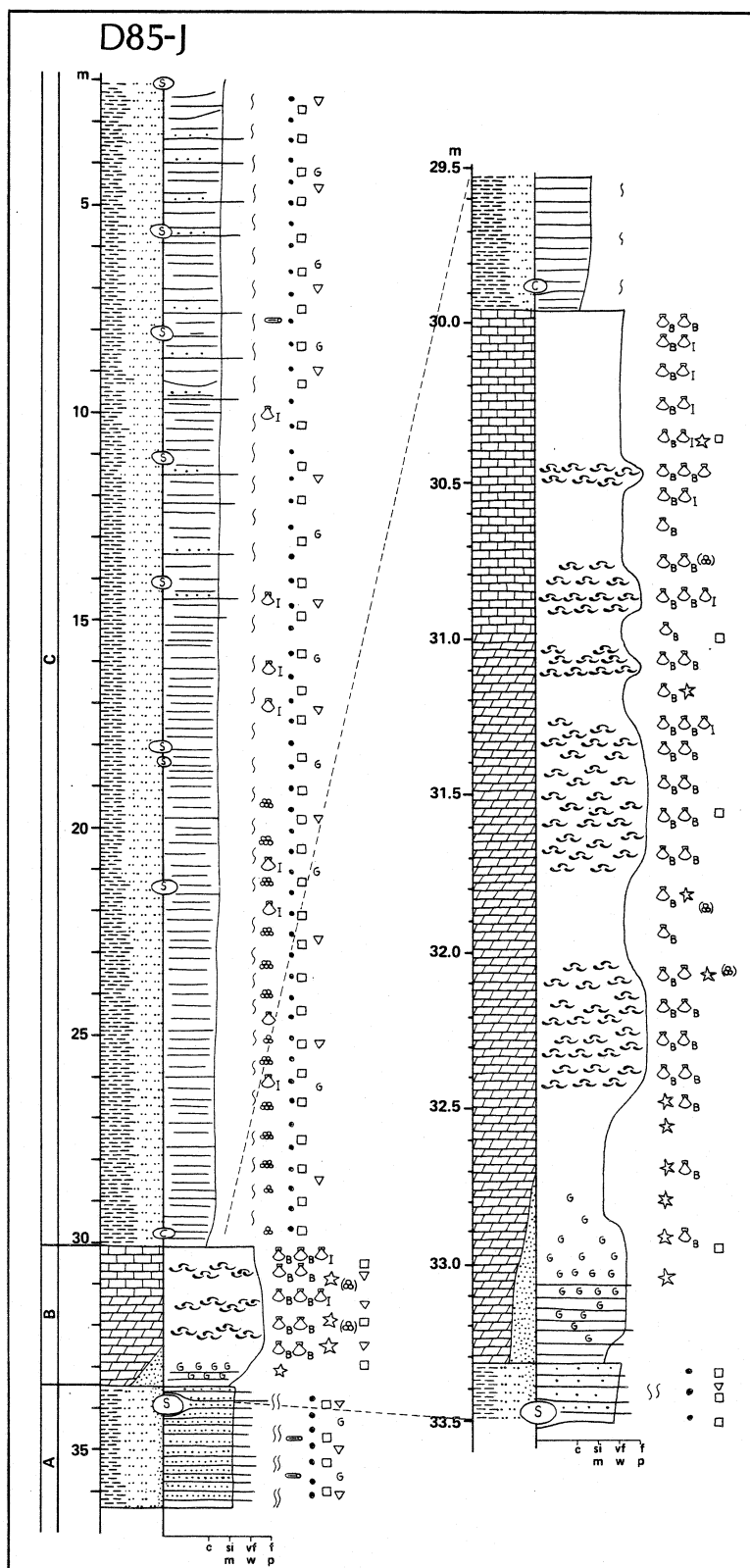


Fig. 4.2a Lithological sections of core D85-J. The right-hand column is an enlargement of the carbonate interval. Legend in Fig. 4.2b, next page.

Legend to the stratigraphical columns		
LITHOLOGY	SEDIMENTARY STRUCTURES	FOSSILS
 Claystone	 Horizontal lamination	 Buchia
 Siltstone	 Horizontal graded lamination	 Inoceramus
 Sandstone	 Coquina bed	 Echinoderma
 Limestone	 Bioturbation → intensity	 Foraminifera
 Dolomite		 Wood fragment
 Siderite		 Unidentified fossil
 Till	EARLY DIAGENETIC MINERALS	
 Concretion	 Glauconite	
	 Pyrite	
	 Chert	
	 Sideritized/Pyritized pellets	

Fig. 4.2b Legend to the stratigraphic column.

4.3 Description of unit B

Unit B consist of wackestone to packstone carbonate with varying amounts of fossils.

The lower half meter of the carbonate interval consists of a light grey horizontally laminated ankerite cemented rock with abundant clastic grains and glauconite (Fig. 4.3). Where the amount of clastic grains decreases, scattered dolomitized echinoderma plates and Buchia fragments occur (Fig. 4.3). The ankerite intervals show irregular colour changes from grey to brownish grey. Observation of thin sections from some of these boundaries indicated that they were caused by an abrupt change from microcrystalline dolomite to xenotopic ankerite texture.



Fig. 4.3 Units A, B and C, and lithological boundaries in core D85-J.
Part 1 shows the abrupt change from unit A to B. The lower part of unit B contains abundant clastic grains and glauconite.
Part 2 shows the interval 30.37-30.66 m of wackestone-packstone with mouldic porosity as small black spots.
Part 3 (29.73 to 30.20 m) shows the change from wackestones in B to black and dark grey silty claystone at the base of C.

Higher in the carbonate unit, the rocks consist of alternating layers of wackestone and packstone, with subordinate grainstone beds. The packstone beds consist predominantly of fragments of the bivalve Buchia. These beds alternate with wackestone beds that are characterized by Buchia fragments, echinoderma plates, Inoceramus fragments and a few foraminifera fragments as well as abundant moldic porosity. Higher up in the unit the rock consists predominantly of mudstones and wackestones. The upper 10 cm consists of sideritized wackestone which give the rock the spotted appearance (Fig. 4.3).

4.4 Mineralogy

The core can be divided lithologically into three units; A, B and C (Fig. 4.2). Units A and C are composed of silty claystones, and Unit B is a partially dolomitized and ankeritized limestone. The mineral quantification shown on Fig. 4.4, are based on X-ray diffraction analyses.

Unit A

The lower unit shows significantly higher quartz content than the upper unit (40-60%). The feldspar content is low, never exceeding 5%.

The clay mineral content is also lower than in the upper unit, and averages 20% consisting of mica/illite (7-15%), mixed layer (4-8%), chlorite (0-4%) and kaolinite (1-4%). Smectite was not recorded.

Of the carbonates only dolomite and siderite was recorded only in minor amounts (less than 7%).

The pyrite content averages 9%, but this value may be somewhat too high.

Unit B

Both dolomite/ankerite and calcite are present in the carbonate unit. Calcite occurs in the upper 1.5 m of this unit. Except for the uppermost sample dolomite and ankerite occur throughout the unit, and have their highest abundances in the lower interval. Thin section studies indicate that ankerite is the dominant carbonate mineral in the lower 1 m, and dolomite is the most dominant of the two minerals higher in the unit. Siderite occurs only in the lowest and uppermost samples of the unit.

The quartz content within this unit is generally below 20%, but values of 40-50% occur in the lower part of the section near the contact with unit A.

The clay mineral content in unit B is generally lower than 10% (but exceeds 20% in two samples), and the feldspar content is low throughout.

The pyrite content varies significantly throughout the unit and reaches very high values in the lowermost part. The high values may result from pyrite concretions. Thin section and macroscopic investigations indicate that the pyrite is about 10-20%.

Unit C

The quartz content (Fig. 4.4) of the upper unit is between 30 and 35% with occasional values slightly above 40%. There is no significant change in these concentrations with depth. The K-feldspar content is low (ca. 2-3%), with slightly higher values locally and the content of plagioclase is slightly higher (4-9%), with a single high value of 15%.

Clay minerals make up 30-50% of the total mineralogy and can be divided into the following categories mica/illite (10-20%), mixed layer minerals (5-12%), chlorite (1-6%), kaolinite (10-20%). Smectite, reaches 4% in rare cases.

Siderite normally makes up from 7 to 15%. Other carbonate minerals occur only in small amounts and rarely exceed 1%.

With a few exceptions, pyrite makes up about 2-3% and increases slightly towards the lower part of the unit.

4.5 Diagenesis

The rocks in unit B range in composition from mudstones to grainstones. During burial, the rocks went through diagenetic processes which changed the original fabrics and mineralogy. These diagenesis include mechanical compaction, pore filling by calcite and dolomite, replacement by dolomite and siderite, growth of xenotopic ankerite and formation of secondary porosity. The diagenetic stratigraphy is summarized in Table 4.2.

Table 4.2: The major diagenetic processes observed in carbonate unit B.
The numbers refer to the relative timing of the major events.

Mechanical compaction	1
Pore filling cements	
- drusy calcite	2
- pore filling dolomites	3
- blocky calcite	6
Replacements	
- microcrystalline dolomite cementation	4
- pseudomorphic replacement of fossils	4
- sideritization of micrite	7
- replacement by pyrite	5
Recrystallization	
- growth of xenotopic ankerite	5
Secondary porosity	
- mouldic porosity	4-5
- vuggy porosity	7

4.5.1 Mechanical compaction

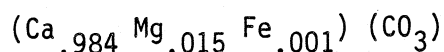
Mechanical compaction has been observed only in the packstone and grainstone beds, where some broken Buchia shell fragments occur. Later diagenetic dolomite and/or earliest diagenetic calcite cement grew at the fracture surfaces, indicating that mechanical compaction took place before the earliest recognized cementation events.

Mayers (1980) concluded that mechanical compaction with fracturing of fossils began at burial depths of approximately 10 m and was completed when a couple of hundred metres of overburden had accumulated. The scarcity of mechanical compaction features may indicate that the rocks were lithified at shallow depths.

4.5.2 Pore filling cements

Drusy calcite cement is seen in some pores in grainstone beds and in the shielded area beneath bivalve shells (umbrella effect) in wackestones, packstones and grainstones (Figs. 4.5 and 4.6). Along the inner surface of some pores, the cement grew as a continuous coating on the surface from which they nucleated. The cement also occurs as 6 to 20 μm long isolated crystals along the inner surface of some pore surfaces. As shown in Fig. 4.5, the cement grew prior to mouldic porosity formation and dolomitization. The cement displays a dull red luminescent colour under the cathodoluminescence microscope.

The cement has an average chemical composition of:



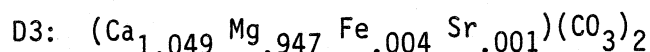
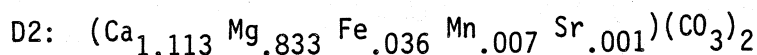
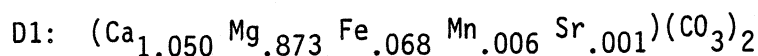
(for more details, see Appendix, Table 1).

Longman (1970) and Bathurst (1983) described CaCO_3 cement precipitated from marine porewater as either Mg-calcite or aragonite. These cements show characteristic fine-grained needle shaped crystals (aragonite) or small blades and/or rhombs (Mg-calcite). Such fine-grained textures are assumed to be caused by magnesium poisoning of the cement growth (Berner 1975). Removal of magnesium from marine pore water by dilution and/or dolomitization may give rise to drusy calcite growth. Chemical

analysis of the calcite in unit B indicates a fairly Mg-rich and Fe and Mn poor pore water. Fe and Mn are important constituents in freshwater calcite cements (Richter and Füchtbauer 1978), and the chemical composition of our calcite indicates a non-fresh pore water. The dolomitization with Mg-removal is thus supported by the low Fe- and Mn-content of the drusy calcite cement.

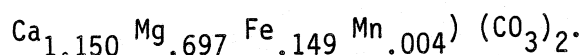
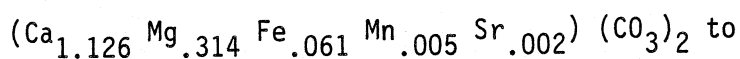
Pore filling dolomite may either consist of coarse-grained subhedral to euhedral crystals or drusy fabrics (Figs. 4.5, 4.6 and 4.7). Both these fabrics are restricted to packstone and grainstone beds or to larger pores under bivalve shells in wackestone beds. Subhedral to anhedral dolomite is most common. Back-scattered electron images of the drusy dolomite reveal three different growth zones (D1, D2 and D3 from oldest to youngest) characterized by different shades of gray (Fig. 4.9). The contacts between the zones are sharp with no signs of corrosion.

The average chemical composition of these different zones are



(For more details, see Appendix, Table 3).

Pore filling subhedral to euhedral dolomite cement in a thin section from 31.43 m shows a range of composition between (Table 3D):



These are richer in Ca and Fe, and poorer in Mg than the dolomites of D1, D2 and D3.



Fig. 4.5 A partially dolomitized packstone in plane polarized light. The brown and black coloured areas are partially dissolved *Buchia* fragments. The dark and fine-grained areas are dominantly microcrystalline calcite (micrite), the white areas are drusy calcite and dolomite cements. Dissolution of *Buchia* fragments took place later than precipitation of the calcite and dolomite cements. Sample 30.93 m.

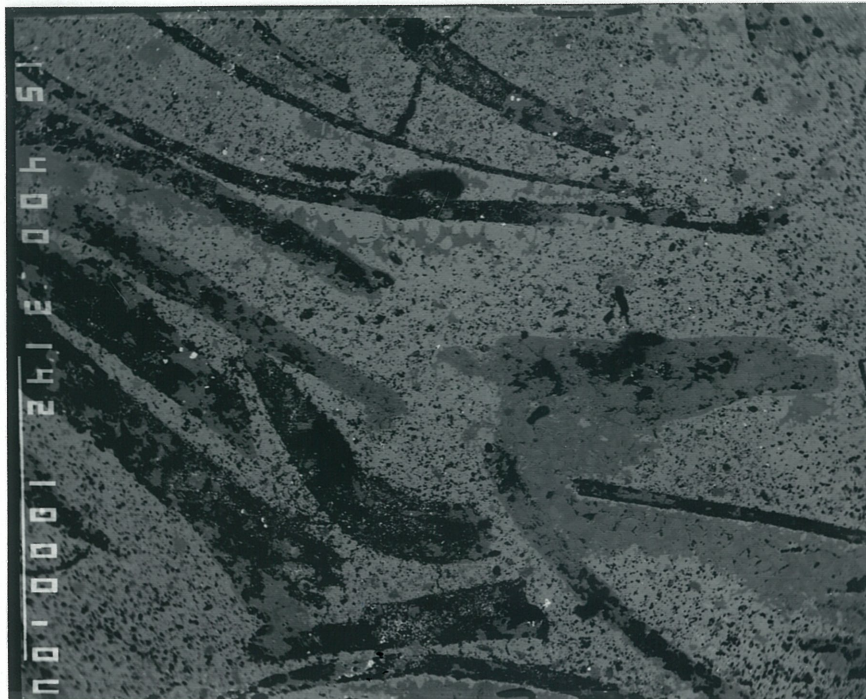


Fig. 4.6 A back-scatter electron image (BEI) of approximately the same area as Fig. 4.5. Black is porosity, darkest grey areas are dolomite, and the lighter grey areas are calcite. Notice that the drusy calcite cement in the middle part of the photomicrograph precipitated prior to the dolomite cements. Sample 30.93 m.

Blocky calcite cement has been observed only in a thin section from 31.42 m. This cement partial fills porosity from earlier dissolved fossils. Remnants of the dolomitized fossils can be seen along the rim of some pores, which indicates that the cement is younger than the dolomite which replaced the fossil.

The calcite cement has the following average chemical composition



(for more analytical details - see Appendix, Table 2).

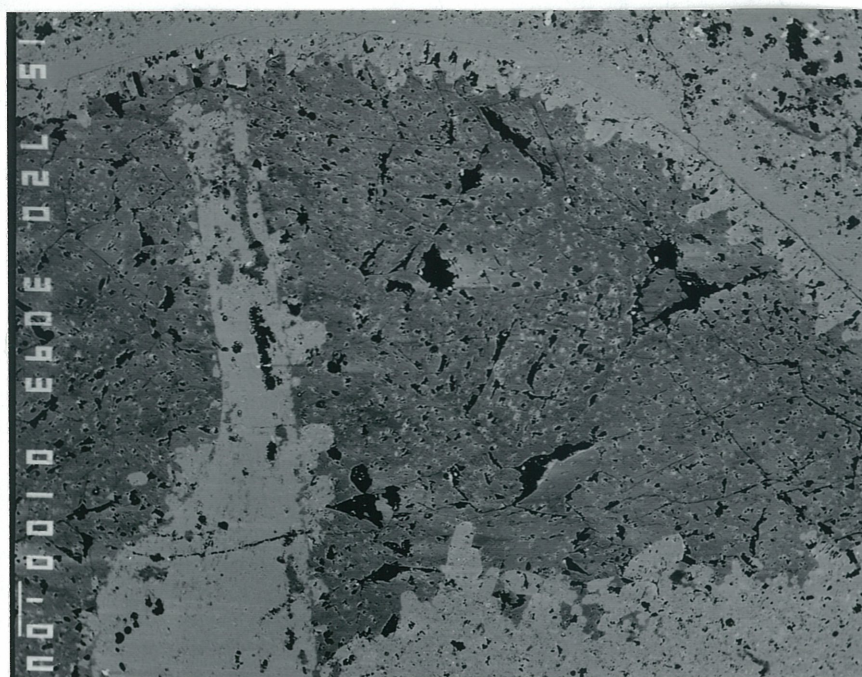


Fig. 4.7 A back-scatter electron image (BEI) of a pore lining calcite (lightest grey) and subhedral dolomite (darkest grey) cement filling a large pore under a Buchia shell. Notice that the shell, the microcrystalline calcite and the pore lining calcite show no evidence of corrosion or replacement by the pore filling dolomite. Sample 31.37 m.

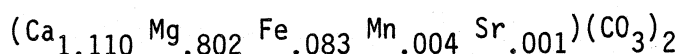
4.5.3 Replacement cements

There are three different replacement fabrics recognized within our thin sections: microcrystalline dolomite, microcrystalline siderite and pseudomorphic replacement of fossils by dolomite.

The microcrystalline dolomite is dark in plane-polarized light, because of its very fine-grained texture (1-5 μm) and the numerous inclusions of clay minerals (indicated by XRD) and organic material. Back-scattered electron images (Fig. 4.11) indicate that porosity of the microcrystalline dolomite is fairly high (about 10-20%).

Dolomite replacement of fossils is very common. In replaced fossils the primary shell fabric is commonly well preserved. Echinoderma plates, Buchia fragments and foraminiferas are almost always replaced while Inoceramus prisms are not. The selectivity of replacement may be caused by primary mineralogy, fossil fabric or grain size (Sibley 1980).

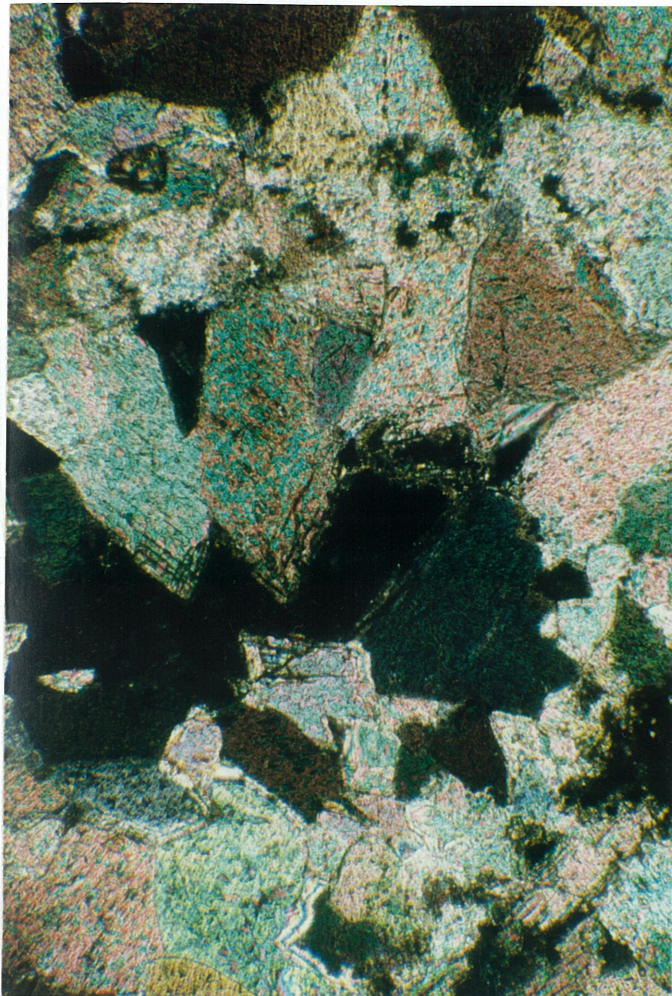
Chemical analyses of microcrystalline dolomite and dolomite replacement of fossils show the following average composition:



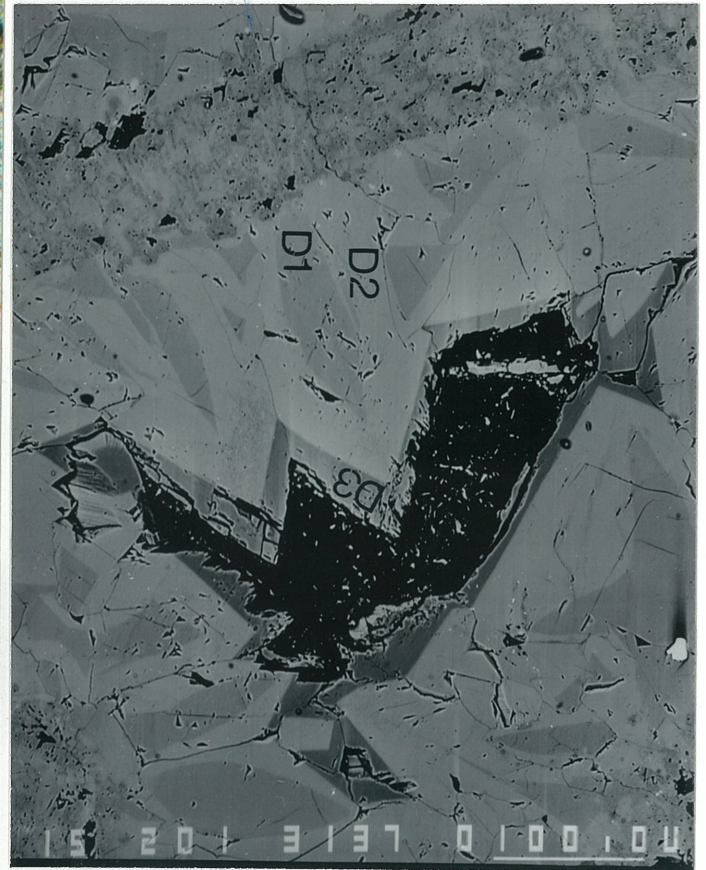
(For more analytical details, see Appendix, Table 5).

The back scattered electron image (Fig. 4.9) also shows that two different phases of these replacement textures exist. This may explain the fairly wide range in chemical composition within this group, and may be comparable with the D1 and D2 phases of the porefilling dolomite.

Microcrystalline siderite occurs in the upper 10 cm (Figs. 4.3 and 4.4) and has a relative abundance of 18%. It also occurs in trace amounts in the lowermost few cms of the unit. Back-scattered electron images show that siderite occurs as 1-5 μm long grains in close association with microcrystalline calcite, which indicates that the siderite may have formed by pseudomorphic replacement of some calcite crystals.



100 μm



100 μm

Fig. 4.8 A cross polarized image of a dolomitized fossil (finer grained zone crosscutting the image in the upper half of the picture) and drusy dolomite cement that partly fills a pore (black). Sample 31.37 m.

Fig. 4.9 A back-scatter image of the same area as Fig. 4.8. Notice the three different growth zones in the drusy dolomite: D1, D2 and D3. D1 and D2 also are recognized in the dolomitized fossil. Notice also the corrosion of D3 along the pore (black).

4.5.4 Pyrite

Pyrite occurs as subhedral, often cube-shaped crystals associated with porosity. Sometimes small concretions up to 2x1 cm occur. The pyrite content varies considerable from 0 to 49% (Fig. 4.4). Occasional inclusions of partially dissolved bivalve fragments composed of ankerite occur within the pyrite (Fig. 4.12), indicating that the pyrite is a very late diagenetic cement. The pyrite is very often associated with vuggy porosity.

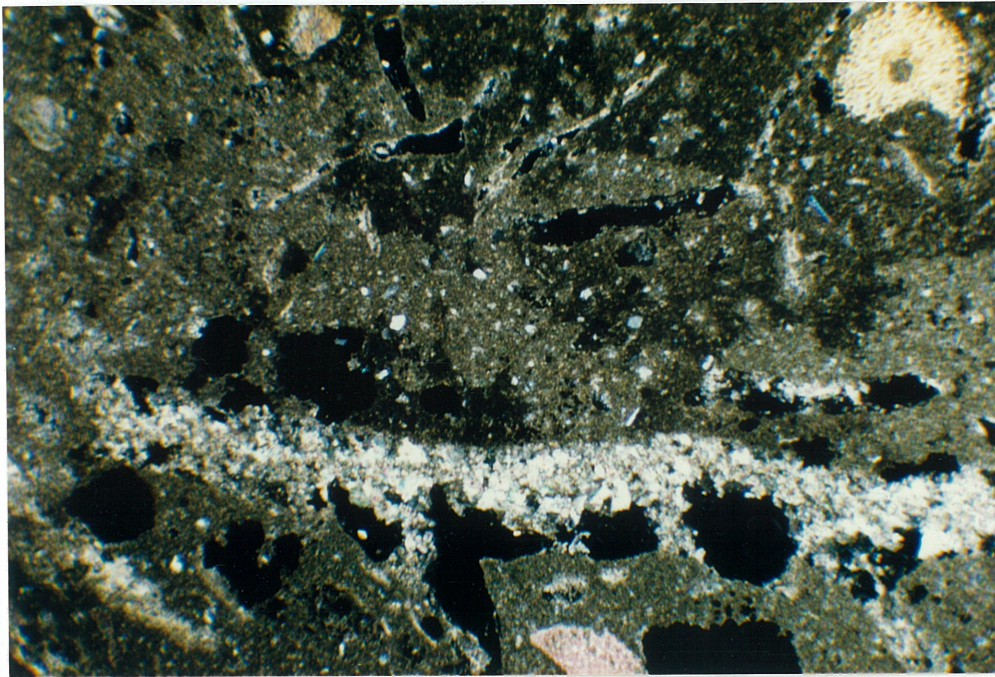


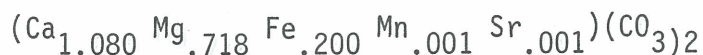
Fig. 4.10 A cross-polarized image of a partially recrystallized dolomitized wackestone. Buchia-fragments (lightest grey) have a much coarser textures than the recrystallized microcrystalline dolomite. Black areas are secondary porosity formed after recrystallization. Sample 31.37 m.

4.5.5 Recrystallization cement

The only recrystallization cement recognized within the rocks is xenotopic ankerite, which occurs as crystals 10-30 μm long. These crystals have no inclusions. Where this fabric is in contact with microcrystalline dolomite, the boundary is abrupt and irregular (Figs. 4.8 and 4.9), thus indicating that the ankerite grew by recrystallization of microcrystalline dolomite. Ankerite is also seen as recrystallization fabrics of fossils, and no signs of the original fossil fabric are then recognized.

This cement is dominant in the lower part of the core. The xenotopic ankerite is also more abundant along fossil-rich zones with abundant secondary porosity. Back-scattered electron images (Fig. 4.11) of the microcrystalline dolomite reveal higher porosity than the xenotopic ankerite. This implies that new material was added to the xenotopic ankerite during recrystallization.

The chemical analyses of xenotopic ankerite have the following average composition:



(For more analytical details, see Appendix, Table 4).

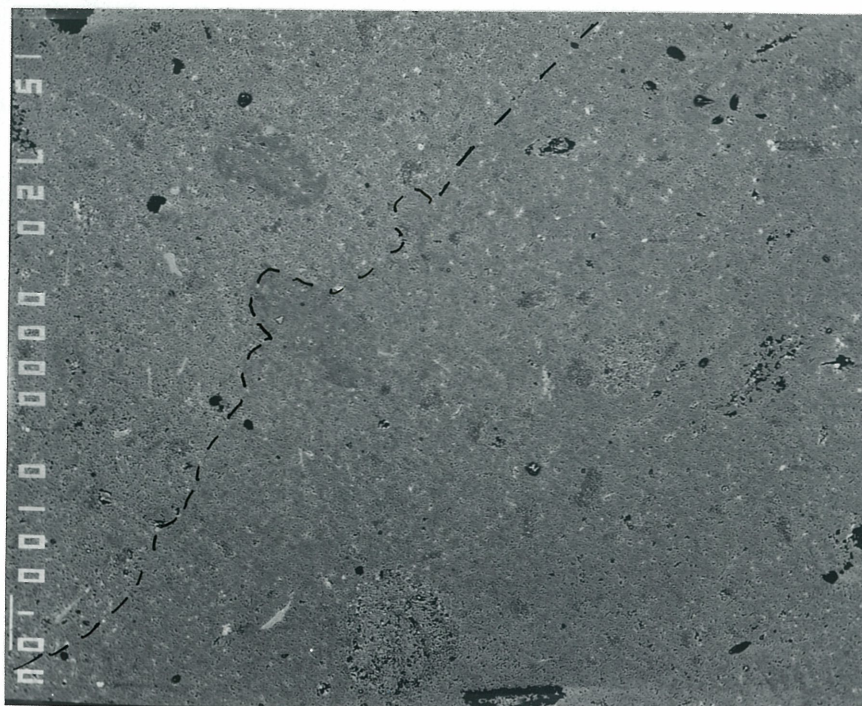


Fig. 4.11 A back-scatter image (BEI) of the boundary (stippled) between microcrystalline dolomite (left part) and xenotopic ankerite. Porosity decreases along the boundary between these two carbonate minerals. Sample 31.40 m.

4.5.6 Secondary porosity

Two phases of secondary porosity formation are recognized in the investigated rocks. The earliest phase involves partial-to-complete dissolution of fossils. Observed fossil moulds are predominantly Buchia, but pseudomorphs of echinoderma plates have also been observed. It is hard to give a relative date for this phase with respect to the other diagenetic events. We suggest that it may have occurred contemporaneously with the dolomitization/ankeritization reactions.

A later phase of secondary porosity formation was associated with pyrite cementation and post-dates the dolomite replacement of fossils (Fig. 4.12). This later episode, which is commonly associated with pyrite cement is difficult to distinguish from the earlier phase, Fig. 4.12.

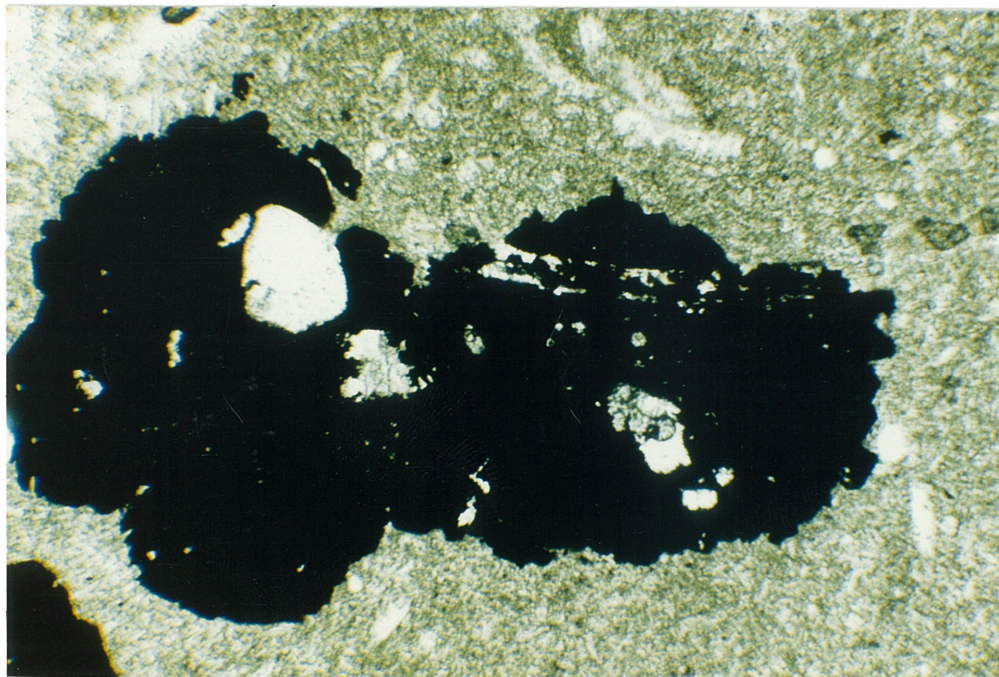


Fig. 4.12 A cross-polarized image of xenotopic ankerite and pyrite cement. Notice the inclusions of xenotopic ankerite and some remains of bivalve shell with mouldic porosity in the pyrite cement.

4.5.7 Porosity and permeability measurements

Permeability values measured from unit B are extremely low (Table 4.3), but high porosity values were recorded for an interval in the middle of this unit. This anomalously high porosity is interpreted to be moldic in nature and has not, therefore, improved permeability. Lack of intercommunication between pores in the remainder of unit B is regarded to have resulted from the cementation and remineralization processes described in this report. Thus, unit B as a whole has little reservoir rock potential.

Table 4.3 Porosity and permeability measurements from unit B (analysed by GECO Petroleum Laboratory).

Depth m	Permeability (mD)		Porosity (He) %
	KA	KEL	
30.00	0.04	0.02	4.6
30.70	0.04	0.02	1.4
31.40	NMP		19.4
31.82	NMP		28.3
32.25	0.041	0.030	4.4
32.90	0.04	0.02	2.1
33.30	0.04	0.02	5.0

4.6 Discussion

The ankerite, dolomite and limestones have predominantly wackestone structure with some packstone and grainstone beds (Fig. 4.2). The transition between dolomite and limestone (Figs. 4.3 and 4.4) is gradual and complex, indicating that the major diagenetic process was dolomitization of original limestone. Additionally the transition between dolomite and ankerite is gradual and complex, and the ankerite formed by recrystallization of dolomite.

The sequence of major diagenetic events is summarized in Table 4.2.

Textural relationships indicate that ankerite formed later than dolomite. We suggest that dolomitization took place at fairly shallow burial depths (before organic maturation and smectite diagenesis), and ankeritization took place during smectite diagenesis (about 70°C). We also suggest that formation of secondary porosity took place contemporaneously with both dolomitization and ankeritization.

4.6.1 Dolomitization and ankeritization by compactional squeezing of pore water from underlying mudstones

There are several different ways in which original calcium carbonate rocks may be dolomitized. At or close to the sediment surface evaporation and precipitation of gypsum may raise the Mg/Ca ratio from about 5 (normal sea water) to more than 7 (the value required for dolomite precipitation). Other models stress the importance of dilution of marine pore water by meteoric water as a necessary precondition for dolomitization (Badiozimani 1973, Folk & Land 1975, Muir et al. 1980, Kaldi & Gidman 1982 and Ward & Hally 1985).

Folk & Siedlecka (1974) postulated extensive dolomitization in "schizohaline" environments (evaporation during dry seasons followed by freshwater flushing during more humid seasons) in Late Paleozoic rocks from Bjørnøya.

There is no evidence of evaporite minerals or chemically clean (Mg:Ca=1) dolomite, which Lumsden & Chimahusky (1980) found in close association with evaporite minerals, supporting a non-evaporite model

for the dolomitization in the investigated rocks. Dating of units A, B and C indicate that a significant gap in time exists between unit A and B, while a continuous sedimentation took place during unit B and C. All the units are interpreted as shelf deposits. The relationship between unit B and C therefore suggests that fresh water flushing of the carbonate interval was unlikely.

Neomorphic replacement of Mg-calcite by calcite liberates Mg^{2+} which may cause some dolomitization. However, the original concentration of Mg-calcite in the rocks probably was too low to cause the observed distribution of dolomite. However, neoformation may be one source of Mg for the dolomitization reactions.

The lower half of the carbonate unit consists nearly exclusively of terrestrial and glauconite grains, dolomite and ankerite, while the upper one third (except the upper 10 cms) consists predominantly of calcite. Thus, dolomite-supersaturated water flowed upwardly through unit B and probably occurred during compaction, when such an upwardly directed water flow would predominate.

The original pore waters of units A, B and C are assumed to have been normal saline sea water, because of the marine depositional environments. During compactional squeezing of pore water, one would expect that the expelled water has a marine composition as well. However, Singer & Müller (1983), demonstrated that the salinity of the expelled pore water from marine deposited mud decreased during compaction. Mg^{2+} and Ca^{2+} were among the studied ions. The reason for the observed decrease in salinity was assumed to be the ionic filtration effect of clay minerals. Divalent cations and potassium are most efficiently filtered by clay minerals. Decreasing salinity of the expelled pore water during progressive compaction, may be explained by changes in mud fabric as compaction progresses. During deposition and very shallow burial, the clay minerals were oriented randomly within the sediment in a card-house manner, and intergranular pores were fairly large. Absorption of the ions, particularly cations, on negatively-charged clay mineral surfaces was less effective. During increasing compaction the clay minerals started to undergo parallel orientation, which reduced porosity and permeability. Hydrated

cations were more easily filtered from the expelled porewater, giving rise to lower saline pore water. As this fluid entered the carbonate rock, it progressively diluted the salinity of the ambient pore water as compaction progressed.

It is evident in unit B that most of the dolomitization reactions took place when pore water supersaturated with respect to dolomite entered the limestone. The dolomitization reaction front moved up into the calcium carbonate rock as new ions were added from below.

The same general trend seems to be valid for xenotopic ankerite. In the lower metre of the core, xenotopic ankerite is the carbonate mineral. Ankerite occurs more sporadically higher in unit B. It therefore seems likely that the underlying silty claystone were the major source of iron needed for the formation of the xenotopic ankerite. Clay minerals within this unit may also supply some iron and magnesium.

Calculation of the thermodynamic stability of different carbonate minerals in mixtures of marine and fresh water by Badiozamani (1973) showed that the solutions were undersaturated with respect to calcite, and supersaturated with respect to dolomite in a large mixing region. Badiozamani concluded that the dolomitization reaction was caused by thermodynamic drive, which caused calcite to dissolve and dolomite to precipitate. Morrow (1982) concluded that dolomitization caused by dilution of marine porewater undersaturated with respect to calcite is not important, because dolomitization takes place in solutions that are saturated or supersaturated with both phases (eg. evaporitic environments). Dolomitization takes place simply because of kinetic effects of the attendant decrease in rate of precipitation, according to Morrow (1982). Three major kinetic effects seems important:

- (1) rapid crystalization of dolomite from supersaturated solutions,
- (2) difference in the hydration energy between Mg^{2+} , Ca^{2+} and CO_3^{2-} (Mg^{2+} approximately 20% greater than Ca^{2+} and much greater than the carbonate ion), and
- (3) the low activity of carbonate ions relative to Mg^{2+} and Ca^{2+} in most solutions.

Factor (1) and (2), are the main reasons why dolomites studied by Morrow (1982) are calcium-rich. Rapidly growing crystals were more likely to incorporate ions that had adhered in incorrect lattice positions. Because of its lower hydration energy, Ca^{2+} substitutes more easily for Mg^{2+} than vice versa. A solution rich in dissolved carbonate promotes dolomitization, because it has a higher probability of containing energies high enough to shoulder away the hydration shield of magnesium. Additionally, higher temperature reduces kinetic inhibition effects. Length of time and the supply of Mg^{2+} are other important factors that control dolomitization reactions, according to Morrow (1982).

4.6.2 Chemistry of the carbonate minerals

Our diagenetic models indicate that pore water chemistry changed gradually during burial. Consequently, carbonate minerals precipitated at different times and positions in unit B have quite different chemical compositions. Large chemical changes probably took place when the rocks were buried to oil window depths and organic material matured. Organic molecules, CO_2 and sulphur compounds may have lowered the pH of the porewater at this time. Smectites probably were transformed to other clay minerals at these high temperatures, and significant amounts of Mg^{2+} , Fe^{2+} and silica may have been liberated into the carbonate rock.

The dolomites and ankerite in the investigated samples have a mole % Ca that varies between 52.5 to 57.4, and the ankerite contains 54.0 mole % Ca. These results show that the dolomites and ankerite are calcium rich. Lumsden & Chimahusky (1980) concluded that dolomite composition is not related to insoluble residue, porosity, rock type, percentage of crinoidal fragments, recrystallization, or percentage of the dolomite fraction in dolomitized limestones, however, they observed that crystalline dolomites have a lower Ca^{2+} content (mode 50-51% CaCO_3) than dolomitized micrite (modes 51-52% evaporitic, 54-55% non-evaporitic CaCO_3). According to our data, the mole percent CaCO_3 of pore filling dolomite varies between 52.4 and 57.4%, while the microcrystalline dolomite and dolomitized fossils contained an average of 55.5% CaCO_3 . Thus, it should be noted that replacement dolomites

show a variable chemical composition (varies from 53.5 to 58.1 mole %). The range of compositional variation between pore filling and replacement dolomites in unit B is far greater than those of the rocks investigated by Lumsden & Chimahusky (1980). The great compositional inhomogeneity of unit B replacement dolomites may be explained by the low permeability of the replaced material. Chemical inhomogeneities of the original material may give rise to significant local chemical variations in the replacement dolomites. Comparison of the chemical composition of pore filling and replacement dolomites show that variation in porewater compositions probably occurred between the precipitation of the D1 and D2, dolomites. This is indicated by the zoning seen in back-scattered electron picture of the replacement dolomites (Fig. 4.9).

Ward & Halley (1984) described selective leaching of the most calcium-rich dolomite in strata containing both limpid dolomite ($\text{Ca}_{1.14} \text{Mg}_{.86} (\text{CO}_3)_2$) and zoned dolomite composed of alternating zones with compositions $\text{Ca}_{1.14-1.18} \text{Mg}_{.86-.82} (\text{CO}_3)_2$ and $\text{Ca}_{1.24} \text{Mg}_{.76} (\text{CO}_3)_2$. The limpid dolomite composition and the low-calcium dolomites show chemical compositions comparable with the investigated dolomites from D85-J. Ward & Halley (1984) interpreted their dolomites and calcites as having precipitated from pore water produced by freshwater after shallow burial or contemporaneous with sedimentation. This model does not provide a satisfactory explanation for the carbonates of unit B.

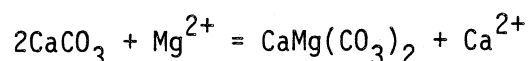
Both Lumsden & Chimahusky (1980) and Morrow (1982) noticed that temperature may influence the chemical composition of dolomites. High temperature should result in chemically clean dolomites ($\text{Mg}:\text{Ca}=1$), because the importance of the hydration energies of Mg^{2+} and Ca^{2+} decreases with increasing temperature. D1 is more calcium-poor than D2, which indicates that a significant change in the pore water chemistry took place before crystallization of D2 rather than an increase in temperature. On the other hand, D3 has a much more chemical clean composition than the earlier phases, and an increase in temperature may explain this unique chemical composition. The trace element content (Fe, Mn and Sr) in D₃ is very low, indicating that it was derived from approximately the same pore water as the blocky calcite cement.

4.6.3 Dolomitization and ankeritization reactions

Dolomitization probably started at very shallow burial. The scarcity of mechanical compaction breakage of shell fragments may indicate that the rock was partly lithified prior to major mechanical compaction.

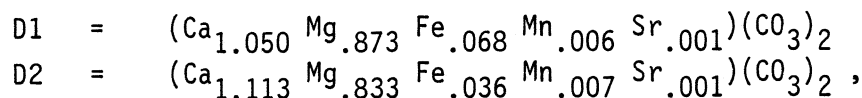
Except Inoceramus prisms (calcite), all CaCO_3 minerals such as aragonite, Mg-calcite (micrite and echinoderma plates, forams) and calcite (Buchia-shells) are dolomitized. Thus, neither mineralogy nor grain-size controlled the dolomitization reaction. Structure of the original material was the primary causal factor. The general dolomitization model implies that the dolomitization/ankeritization took place over a long time interval. Therefore, it is likely that Mg-calcite and aragonite were recrystallized to calcite, mostly prior to the dolomitization and ankeritization reactions.

Our study shows that the dolomitization reactions were selective, at least in the dolomitization front. Euhedral dolomite cement precipitated in the pores, but drusy calcite cement as well as a Buchia fragment and the microcrystalline calcite cement were unaltered (Figs. 4.7 and 4.8). Because there is no evidence of corrosion of the calcite material, solution in the dolomitization front are assumed to have been also saturated with calcite. Folk & Land (1975) pointed out that sparry calcite can form only in supersaturated solutions with low Mg/Ca ratio. Drusy calcite cement may then precipitate from pore water depleted in Mg^{2+} by the dolomitization reaction, thus liberating calcium to the pore water. The drusy calcite cementation may therefore have taken place contemporaneous with dolomitization in the lower part of the carbonate units. The calcite is fairly "clean", with low Mg^{2+} and negligible concentrations of other trace elements. The simplest chemical equation that describes the dolomitization reaction is:



In this equation, Mg^{2+} is consumed during the reaction and Ca^{2+} is liberated. The liberated Ca^{2+} may result in a calcite-supersaturated pore water just above the dolomitization front.

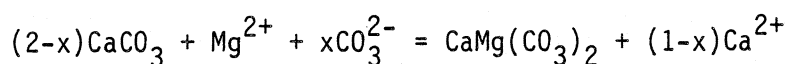
The chemical composition of the two most important pore-filling dolomites are as follows:



The oldest phase (D1) has lower Ca and higher Mg and Fe than D2. The lower Ca^{2+} content of the D1 may be the result of Ca^{2+} removal by contemporaneous precipitation of drusy calcite just above the D1 precipitation areas. This interpretation is consistent with the co-precipitation of drusy calcite and D1. Dolomitized fossils and microcrystalline dolomite display broad ranges in chemical composition although the precipitation may have taken place from both the pore water that precipitated D1 and to a larger extent D2. Back-scattered electron images have two different shades of gray, supporting the above hypothesis. This may indicate that all the dolomitization reactions and pore filling cementation took place almost contemporaneously close to the dolomitization front.

4.6.4 Dolomitization and ankeritization reactions and formation of secondary porosity

An important aspect of dolomitization of CaCO_3 is the formation of about 13% and 6% secondary porosity when dolomitization calcite and aragonite respectively, when equal moles of calcium carbonate are replaced. Visual comparison of back-scattered electron images from microcrystalline calcite and dolomite indicates that no significant new porosity was created during dolomitization of microcrystalline calcite cements. Morrow (1982) presented a dolomitization equation that compensates for volume conservation during dolomitization:



where $x = 0.11$ or 0.25 for the dolomitization of aragonite or calcite respectively. Thus, magnesium and carbonate ions must be supplied during the dolomitization reaction, if mineral volume is to be conserved.

The presence of mouldic porosities indicate that the supplied ions were derived most likely by dissolution of fossils, particularly Buchia fragments and to a lesser extent echinoderm plates. The large amount of microcrystalline dolomite compared to fossil fragments may have caused significant dissolution of the fossil fragments, and hence created mouldic porosity.

Later, xenotopic ankerite started to recrystallize the microcrystalline dolomite and some of the fossil fragments. Iron must have been supplied during this reaction. Because the porosity of the xenotopic ankerite is somewhat lower than that of the microcrystalline dolomite, some carbonate ions must have been supplied. Calcium and magnesium also may have been supplied as well. The preferential distribution of xenotopic ankerite around fossil moulds may indicate that the ions were supplied from pore water flow in the higher permeability zones in fossil moulds. Alternatively, the additional carbonate, magnesium and calcium ions may have been sourced by dissolution of fossil fragments.

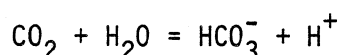
The source of the iron required for ankeritization of dolomite may have been the underlying silty mudstone. Vitrinite reflectance analyses indicate that these rocks were subjected to oil window maturation. The temperature may have been sufficiently high for smectite alteration. Boles & Franks (1979) found that large quantities of Mg^{2+} , Fe^{2+} , SiO_4^{4-} and water were liberated during smectite diagenesis. In Unit A, the illite-mixed clay and chlorite may have formed by the decomposition of smectite, which is consistent with the above reaction. Some Fe^{2+} and to a lesser extent Mg^{2+} may then have entered the carbonates, causing xenotopic ankerite and microcrystalline siderite formation. These ions may have been carried by the expelled pore water. Sideritization in the upper 10 cm of the carbonate rock may have taken place by diffusion. Alternatively, water may have preferentially moved along the lithological boundary and cause this sideritization. However, diffusion was probably also active during the latter mechanism.

4.6.5 Permeability and porosity relationships

The permeability of all samples in unit B is very low (Table 4.3), which must be a result of the different cementation and remineralization processes that reduced fluid migration to extremely low values. In general, porosity is also low, however some high values (19 and 28%) are recorded for the middle of the unit. These values are due to moldic porosity that is not characterized by interconnecting channels. Consequently, unit B as a whole is regarded to have no potential as a source rock for hydrocarbons.

4.6.6 Pyritization

When burial depths reach the oil window, important chemical reactions in organic matter takes place, during which CO_2 and sulphur compounds are liberated. The reaction between CO_2 thus produced and water can be described in the following equation:



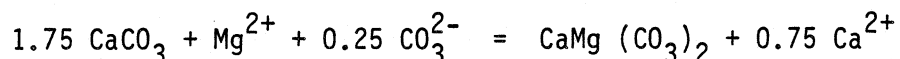
Thus, pH is lowered because of H^+ liberation, which is a very important factor in the stability of carbonate minerals. Only a slight lowering of pH may enhance the carbonate dissolution. The last carbonate dissolution episode that took place in unit B rocks was accompanied by pyrite precipitation. Chemical reactions associated with organic material maturation may have resulted in lowered pH and also supplied the sulphur needed for pyrite precipitation. As discussed above, iron may have been liberated during smectite diagenesis. With negative Eh and pH lower than 6.5, pyrite will precipitate, and carbonate minerals such as siderite may dissolve, even if sulphur activity is as low as 10^{-6} and CO_3^{2-} activity is equal to 1.

4.7 Conclusions

The mineralogical, textural and chemical study has been carried out on a 3.3 m thick partial dolomitized and ankeritized wackestone, to study its diagenesis. A summary of the diagenetic sequence of events is shown in Table 4.2.

Ankerite and dolomite are the dominant carbonate minerals in the lower 2 m of unit B, while calcite becomes dominant in the upper part (Fig. 4.4). We suggest that dolomitization took place at shallow to moderate depth of burial, while ankeritization and pyritization took place at deeper burial. The distribution of the carbonate minerals indicates that compactional squeezing of pore water from the underlying mudstones was the major source of the allogene ions.

Dolomitization reactions of carbonate have been described (Morrow 1982) as follows:



During shallow burial, slightly diluted pore water derived from compaction of the underlying mudstones entered the carbonate interval and caused precipitation of pore filling calcite and dolomite as well as dolomitization of microcrystalline calcite and fossil fragments. The latter process reduced the solid material volume. Because no additional porosity was formed in the microcrystalline dolomite during dolomitization of the microcrystalline calcite, CO_3^{2-} was supplied by selective dissolution of fossil fragments. During deeper burial, xenotopic ankerite recrystallized the microcrystalline dolomite in the lower part of the unit. The iron required for pyritization probably was sourced by clay mineral diagenesis in the underlying mudstone.

During kerogen maturation, acidic pore water containing dissolved sulphur entered the carbonate interval, caused pyrite precipitation, and probably also created some vuggy porosity. The cementation and remineralization processes described in this report have resulted in carbonates with very low permeability. Although high porosity values occur in the middle part of the unit, the moldic nature of these pores and the low permeability exclude unit B as a possible reservoir rocks for hydrocarbons.

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4.9 Appendix

Table 1 CHEMICAL ANALYSES (wt% oxide and mole fraction of cations exclusive C)

	CaO	Ca	MgO	Mg	FeO	Fe	MnO	Mn	SrO	Sr	Ca/Mg
30.93 m	54.24	.983	.65	.016	.05	.001	.00	.000	.00	.000	61
	57.07	.979	.77	.019	.05	.001	.01	.000	.01	.000	52
	54.22	.980	.76	.019	.04	.001	.00	.000	.07	.001	52
	53.92	.980	.78	.020	.02	.001	.00	.000	.00	.000	49
	53.83	.987	.50	.013	.04	.001	.00	.000	.00	.000	76
	54.32	.989	.33	.008	.11	.001	.03	.001	.07	.001	124
	54.79	.982	.66	.017	.04	.001	.04	.001	.00	.000	58
	53.62	.982	.64	.017	.04	.001	.02	.001	.00	.000	58
	53.64	.982	.67	.017	.00	.000	.01	.000	.00	.000	58
31.42 m	54.51	.986	.44	.011	.05	.001	.02	.000	.08	.001	90
	53.41	.991	.28	.008	.03	.001	.03	.001	.06	.001	124
	54.02	.991	.34	.009	.01	.000	.00	.000	.00	.000	110

Drucy calcite

(Ca_{.984} Mg_{.015} Fe_{.001}) (CO₃)

Table 2:

31.42 m	47.00	.995	.05	.002	.11	.002	.01	.000	.02	.001	498
	59.37	.993	.09	.002	.19	.003	.12	.002	.02	.001	497
	55.03	.993	.13	.003	.13	.002	.07	.001	.08	.001	331
	54.24	.993	.06	.001	.17	.002	.12	.002	.09	.001	993
	54.83	.992	.20	.005	.07	.001	.09	.001	.00	.000	198
	54.11	.991	.28	.007	.08	.001	.00	.000	.11	.002	142
	54.35	.991	.28	.007	.11	.001	.03	.000	.01	.000	142
	54.20	.994	.09	.002	.15	.002	.02	.001	.04	.000	497
	55.10	.990	.12	.003	.16	.002	.15	.002	.14	.001	330

Latest calcite

(Ca_{.992} Mg_{.004} Fe_{.002} Mn_{.001} Sr_{.001}) (CO₃)

CHEMICAL ANALYSES

Table 3a

	CaO	Ca	MgO	Mg	FeO	Fe	MnO	Mn	SrO	Sr	Ca/Mg
31.37 m	30.18	1.056	17.76	.870	2.51	.068	.22	.006	.00	.000	1.214
	29.68	1.037	17.74	.875	2.73	.075	.18	.005	.05	.001	1.185
	30.11	1.058	17.08	.841	3.51	.096	.12	.003	.07	.001	1.258
	30.89	1.068	18.06	.875	1.82	.049	.26	.007	.06	.001	1.221
	29.82	1.054	17.38	.859	2.96	.081	.18	.005	.03	.001	1.227
	29.22	1.038	17.82	.886	2.50	.070	.17	.005	.04	.001	1.172
	29.91	1.048	17.86	.876	2.55	.069	.21	.006	.07	.001	1.196
	30.47	1.058	18.10	.879	2.03	.055	.23	.007	.04	.001	1.204
	29.46	1.038	18.07	.891	2.33	.064	.24	.007	.00	.000	1.165
	30.28	1.045	18.56	.879	1.85	.050	.28	.008	.04	.001	1.165

Dolomite, darker BEI

Table 3B

31.37 m	34.94	1.113	18.82	.839	.99	.024	.25	.006	.08	.002	1.327
	33.38	1.115	18.20	.852	1.01	.027	.24	.006	.00	.000	1.309
	32.50	1.125	17.42	.844	.90	.025	.21	.006	.00	.000	1.333
	32.92	1.126	17.21	.824	1.52	.041	.29	.008	.12	.002	1.337
	32.71	1.068	17.45	.798	1.55	.040	.29	.007	.00	.000	1.338
	32.81	1.127	17.13	.825	1.50	.040	.24	.007	.04	.001	1.366
	32.59	1.125	17.16	.829	1.39	.037	.25	.007	.09	.002	1.357
	32.48	1.123	16.99	.822	1.84	.049	.22	.006	.02	.000	1.366
	32.62	1.112	17.20	.828	1.52	.041	.23	.007	.09	.002	1.355
	31.98	1.109	17.36	.843	1.43	.039	.29	.008	.04	.001	1.316
	31.86	1.103	17.61	.854	1.36	.037	.26	.007	.00	.000	1.292

Dolomite lightest BEI

(Ca_{1.113} Mg_{.833} Fe_{.036} Mn_{.007} Sr_{.001})

CHEMICAL ANALYSES

Table 3c

	CaO	Ca	MgO	Mg	FeO	Fe	MnO	Mn	SrO	Sr	Ca/Mg
31.37 m	31.41	1.060	19.81	.937	.11	.003	.00	.000	.00	.000	1.131
	31.63	1.051	20.35	.946	.13	.004	.01	.000	.00	.000	1.111
	31.46	1.054	20.09	.942	.16	.004	.02	.001	.00	.000	1.119
	31.39	1.042	20.56	.955	.11	.003	.01	.000	.06	.001	1.091
	30.88	1.038	20.29	.955	.17	.005	.00	.000	.11	.002	1.087
	31.48	1.050	20.21	.945	.17	.005	.00	.000	.03	.001	1.111

Dolomite, darkest BEI

Table 3d

31.37 m	32.50	1.138	16.27	.798	2.15	.059	.18	.005	.00	.000	1.462
	32.26	1.126	16.67	.814	1.99	.054	.19	.005	.03	.001	1.383
	32.16	1.126	16.45	.807	2.24	.061	.19	.005	.07	.002	1.395

Dolomite post drucy calcite

 $(\text{Ca}_{1.130} \text{Mg}_{.806} \text{Fe}_{.058} \text{Mn}_{.005} \text{Sr}_{.001}) (\text{CO}_3)_2$

Table 3e

31.37 m	32.63	1.148	15.28	.752	3.47	.095	.15	.004	.04	.001	1.527
	32.56	1.146	15.46	.761	3.27	.090	.12	.003	.00	.000	1.506
	32.97	1.150	14.28	.697	5.49	.149	.13	.004	.00	.000	1.650

Dolomite

 $(\text{Ca}_{1.148} \text{Mg}_{.737} \text{Fe}_{.111} \text{Mn}_{.004}) (\text{CO}_3)_2$

Table 4

33.05 m	30.59	1.079	14.55	.719	7.25	.199	.07	.002	.06	.001	1.501
	30.77	1.093	14.18	.705	7.16	.198	.02	.001	.14	.003	1.550
	28.90	1.078	13.82	.721	6.87	.199	.06	.002	.00	.000	1.495
	30.67	1.104	13.86	.698	7.02	.197	.02	.001	.00	.000	1.582
	30.01	1.071	14.57	.727	7.26	.201	.03	.001	.00	.000	1.473
	29.03	1.031	15.20	.752	7.54	.209	.05	.002	.04	.001	1.362
	29.40	1.088	13.71	.710	6.97	.201	.04	.001	.00	.000	1.532
	30.73	1.093	14.22	.709	7.12	.197	.04	.001	.00	.000	1.542

Xenotopic ankerite in clastic rich lower interval

 $(\text{Ca}_{1.080} \text{Mg}_{.718} \text{Fe}_{.200} \text{Mn}_{.001} \text{Sr}_{.001}) (\text{CO}_3)_2$

CHEMICAL ANALYSES

Table 5

	CaO	Ca	MgO	Mg	FeO	Fe	MnO	Mn	SrO	Sr	Ca/Mg
31.42 m	31.76	1.132	15.36	.766	3.54	.098	.13	.004	.00	.000	1.478
	33.52	1.161	15.93	.772	2.36	.064	.08	.002	.02	.001	1.504
	31.51	1.111	15.92	.786	3.59	.098	.07	.002	.08	.002	1.413
	31.23	1.100	16.70	.823	2.68	.073	.09	.003	.00	.000	1.327
	30.93	1.113	15.83	.798	3.07	.087	.11	.003	.00	.000	1.395
	31.49	1.092	17.45	.848	1.98	.054	.16	.005	.04	.001	1.288
	28.71	1.086	14.93	.791	4.03	.123	.16	.004	.00	.000	1.373
	29.48	1.085	15.32	.790	4.16	.119	.18	.005	.01	.000	1.373
	29.01	1.070	15.51	.800	4.34	.125	.16	.004	.06	.001	1.338

Dolomite pre late calcite in fossils and microcrystalline dolomite
 $(\text{Ca}_{1.110} \text{Mg}_{.802} \text{Fe}_{.083} \text{Mn}_{.004} \text{Sr}_{.001}) (\text{CO}_3)_2$

31.37 m	32.70	1.137	16.74	.815	1.51	.041	.20	.006	.03	.001	1.395
	31.95	1.121	17.04	.836	1.30	.035	.25	.007	.06	.001	1.341

Dolomitized fossils

Table 6

XRD - ANALYSE (wt%)

	Qtz	Feldsp.	Clay-min	Calc	Dol	Sid	Py
2.47-20.95	34	8	42	tr.	1	10	5 (average 30 samples)
30.00	22	3	9	35	-	18	11
30.27	5	1	11	58	24	-	-
30.50	14	1	24	35	26	-	1
30.90	4	-	8	59	27	-	2
31.35	13	3	4	-	80	-	-
31.42	6	-	-	46	35	-	13
31.67	15	-	-	1	58	-	26
31.90	24	4	22	1	43	1	4
32.33	7	2	11	-	80	-	-
32.40	16	-	-	1	67	-	16
32.70	12	3	11	1	73	-	1
33.05	48	-	7	1	23	-	22
33.29	39	2	9	-	-	1	49
33.33-36.20	55	8	19	-	4	4	9 (average 4 samples)

CHAPTER 5

ORGANIC GEOCHEMISTRY

by

Thomas L. Leith and Liv Schou

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5.1 Summary

Additional analysis of selected samples has lead to some slightly changed conclusions regarding the organic geochemistry of the samples from the 1985 shallow drilling cruise.

The most important conclusion, with regard to the evaluation of kerogen type in the rocks, is that the Rock-Eval data for the samples appears to have been largely misleading due to the effects of oxidation. Use of the more detailed Py-GC technique has indicated that the kerogen types in cores D85-J, D85-K and D85-N are more hydrogen-rich than was originally indicated, consisting mostly of type II/III kerogen. A terrestrial organic matter input to some of these rocks is still likely, but this material will tend to consist of more waxy components than has been suggested previously.

The additional analyses of the Triassic rocks in cores D85-K and D85-0 suggest different organic matter inputs to these rocks, the D85-0 rocks containing a more hydrogen-poor, terrestrial kerogen.

The data for the sediment gases from D85-V and D85-J were generally inconclusive due to sample problems, but suggest that the gases are of catagenic origin and may be associated with crude oil.

Detailed analysis of the extractable organic matter (EOM) in D85-J, with particular emphasis on the non-indigenous material, suggests that the unit C EOM is largely indigenous, with possible minor contributions from the non-indigenous EOM in the underlying unit B. The non-indigenous material from units A and B appears to have a broadly similar source type. The differences previously observed may be due to maturity and/or migration differences.

5.2 Detailed kerogen analysis of selected samples

In the 1985 data report, kerogen typing was largely based on Rock-Eval bulk pyrolysis data. However, this data showed evidence of oxidation effects and was considered to be of dubious reliability. In order to obtain a clearer view of kerogen variations in the cores, it was suggested that the samples be further analysed by programmed pyrolysis-gas chromatography (Py-GC). To this end, samples were selected from cores D85-J, D85-K, D85-N and D85-O.

D85-J: Three samples from unit C (dated as Upper Hauterivian/-Barremian), see Fig. 3.2.1, and two samples from unit A (Kimmeridgian/-Middle Volgian) were analysed. The pyrograms are shown in Fig. 5.1.

The kerogen in the claystones of unit C appears to be similar in the three samples analysed. A well-developed homology of aliphatic components is present in the pyrograms, although the samples also contain prominent aromatic peaks including benzene, toluene, xylene and phenols. The aromatic components suggest the presence of terrestrial higher plant material, possibly including waxy cuticle and spore coatings. (m+p)-xylene/n-octene ratios for these samples vary from 1.29 to 1.53 (Table 5.1), values which are typical of waxy terrestrial tissues (Larter & Douglas 1980).

The two samples of unit A are appreciably more aliphatic in nature than the samples of unit C, this being reflected by the significantly lower (m+p)-xylene/n-octene ratios of 0.65 and 0.54 for the two samples. These values are considered to be indicative of type II/III kerogens. It is worth noting that, whereas the hydrogen index data for the 36.20 m sample would suggest a type IV kerogen, the Py-GC data suggest a type II/III kerogen which is even more aliphatic than the 34.22 m sample. This illustrates the inaccuracies which may occur in Rock-Eval data due to oxidation.

Core D85-K: Five samples from this core were analysed using Py-GC. Rock-Eval hydrogen indices for these samples vary from 61 mg/g TOC to 201 mg/g TOC, the majority of the samples having hydrogen indices of between 60 and 70 mg/g TOC. This suggests that a hydrogen-poor type III kerogen is the main kerogen type present.

However, the pyrograms (Fig. 5.2) show that the kerogen is generally quite rich in aliphatic components, as reflected in (m+p)-xylene/-n-octene ratios of between 0.65 and 0.78 (Table 5.1), with a high value of 1.33 at 97.51 m. These data would point to a type II/III or hydrogen-rich type III kerogen more typical of the hydrogen index of 201 mg/g TOC at 123.70 m. The occurrence of prist-1-ene in most of the samples may suggest at least a contribution of type II kerogen. The presence some quite prominent aromatic peaks could suggest a higher plant input, although this would be quite waxy given the aliphatic nature of the pyrograms.

D85-N: Two samples of unit D (see Fig. 3.3.1) and one sample of unit C were evaluated for kerogen type using Py-GC. Hydrogen index data for these samples at 49.72 m, 57.44 m and 53.94 m vary from 72 mg/g TOC to 41 mg/g TOC suggesting a very hydrogen-poor type III or type IV kerogen.

The Py-GC data from the samples, shown in Fig. 5.3 and Table 5.1, show quite aliphatic-rich kerogens with moderate contents of aromatic components. This is reflected by (m+p)-xylene/n-octene ratios of 0.78 to 0.96, which are more suggestive of a type II/III or hydrogen-rich type III kerogen than the very hydrogen-poor kerogen types suggested by the Rock-Eval data.

Core D85-0: One sample at 56.38 m was analysed using Py-GC. This sample has a hydrogen index of 59 mg/g TOC, suggesting a hydrogen-poor type III kerogen. The pyrogram in Fig. 5.4 would also tend to suggest a type III kerogen, based on the mostly aromatic nature of the sample. The aliphatic components are generally quite poorly developed, and this is reflected in the (m+p)-xylene/n-octene ratio of 1.72, which is typical of type III kerogen.

5.3 Comparison of Triassic sediments from cores D85-K and D85-0

A more detailed characterization of the extractable organic matter (EOM) in cores D85-K and D85-0 was carried out in order to examine and compare the EOM occurring in these two cores (Dated as Middle and Late Triassic).

Pyrolysis-gas chromatography (Py-GC) data from these two cores indicate of significant differences in the kerogen. The kerogen from the D85-K core appears to be quite hydrogen-rich, possibly a type II/III kerogen. The D85-0 kerogen, based on analysis of one sample, appears to consist of a terrestrial type III kerogen of significantly less hydrogen richness than the D85-K core. This difference is illustrated by the generally lower (m+p)-xylene/n-octene ratio of the core K samples, relative to the core 0 sample (Table 5.1). The only exception to this is the sample at 97.51 m in core K, which has an intermediate (m+p)-xylene/n-octene ratio of 1.33.

Examination of the extracted material (EOM) from the D85-K and D85-0 cores (Tables 5.2-5.5), indicates that the extraction yields from the D85-K core are generally higher than that from D85-0. The highest EOM yields are obtained at 97.51 m (83.3 mg/g TOC) in D85-K. The actual quantity of EOM obtained from the D85-0 sample was so small (1.7 mg), that problems were encountered in weighing the extract fractions. The data concerning the extract fractions from this sample are therefore unreliable.

Comparison of the saturated hydrocarbon gas chromatograms (Fig. 5.5) of the six samples reveal some similarities between the samples at 97.51 m and 106.60 m in D85-K and the sample from D85-0. The D85-K saturate distributions tend to be less mature than the D85-0 saturate distribution, based on carbon preference indices of between 1.8 and 2.2 in core D85-K, and a lower value of 1.5 in the D85-0 sample (Table 5.6). The D85-K samples tend to be more bimodal in appearance, with a relatively greater content of lower molecular weight hydrocarbons (nC_{16} - nC_{20}). This would agree with the more aliphatic kerogen type observed in the D85-K samples. The pristane/phytane ratios of both the D85-K and D85-0 samples are similar, varying from 1.0 to 1.4 and suggesting that the RedOx conditions of deposition were similar in both core sections.

point to different organic matter types in D85-K and D85-0, although deposited under similar RedOx conditions.

All of the samples studied have aromatic hydrocarbon distributions of very immature appearance. The aromatic hydrocarbon gas chromatograms (Fig. 5.6) are dominated by large "humps" of generally high molecular weight compounds, although the phenanthrene and methyl-phenanthrene peaks are distinguishable in most of the samples. Methyl-phenanthrene index (MPI) values (Table 7) are roughly similar in both cores, although the samples in core D85-K tend to have slightly higher values than the D85-0 sample. The values, varying from 0.54 to 0.73 (MPI-1) suggest a maturity typical of the early 'oil window' (Ro ca. 0.5%). The higher values in D85-K tend to suggest that these samples are the more thermally mature, in contrast to the previously discussed CPI data.

The GC-MS data of the saturated fractions (Fig. 5.7a,b Tables 8 and 9) support the low maturity seen in the previously discussed parameters. Due to the low EOM concentration in the D85-0 sample only weak mass chromatograms were obtained from this sample, making any meaningful comparison between the two cores impossible. Maturity ratios obtained from the five samples from core D85-K indicate hydrocarbons of early to middle oil window maturity. Only small internal variations were seen, and no specific source characteristic features were observed.

Generally low extraction weights prevented $\delta^{13}\text{C}$ data from being obtained for most of the samples (Table 10). It was only possible to obtain data for the saturate fraction of the sample at 123.70 m in D85-K (-30.6°oo). Given the lack of data it is difficult to compare the samples.

The data from the D85-K and D85-0 samples indicate some differences between the non-indigenous EOM in D85-0 and the EOM in D85-K. The immature appearance of the D85-0 EOM would tend to suggest that the source of this material is relatively close to the sampled unit. The analysed unit in D85-K appears to contain a mixed type II/III kerogen which has probably contributed to the EOM present in these rocks.

5.4 Comparison of Quaternary gco (D85-V) and Mesozoic gas (D85-J)

The compositional data on these gas samples have been previously described in IKU report no. 21.3405.00/01/85. Based on these data, the Quaternary gas in core D85-V was indicated as being of catagenic origin, but insufficient data were available to permit any meaningful comparison with the catagenic gas recovered from early Cretaceous carbonates in core D85-J.

In order to examine and compare the origin of these samples, $\delta^{13}\text{C}$ isotope data were obtained on the $\text{C}_1\text{-C}_4$ hydrocarbons, where possible (Table 5.11). Due to lack of sample, data could not be obtained from the 14.64 m gas in D85-V, and only partial data were obtained from the other two samples.

The $\delta^{13}\text{C}$ isotope data for the D85-J gas are typical of gases associated with crude oil (Fig. 5.8), according to Schoell (1983). It is difficult to be certain about the 20.20 m gas in D85-V given the lack of any $\delta^{13}\text{C}$ data on the ethane component. However, the methane $\delta^{13}\text{C}$ value would be consistent with a crude oil-associated gas.

Fig. 5.9 shows a plot of $\delta^{13}\text{C}$ methane against $\% \text{C}_2+$ which suggests that the D85-J gas is of greater thermal maturity than the D85-V gas. This is supported by $i\text{C}_4/n\text{C}_4$ values of 0.79 (D85-J: 31.15 m) and 1.34 (D85-V: 20.20 m), the latter value indicating the lower thermal maturity.

These data suggest that the gas samples analysed are catagenic gases associated with crude oils, but that the D85-J gas appears to be substantially more mature than the D85-V gas. Connan & Cassou (1980), suggest that gases with $i\text{C}_4/n\text{C}_4$ ratios of greater than 0.8 are associated with immature condensates, while gases with $i\text{C}_4/n\text{C}_4$ ratios of less than 0.8 are associated with waxy crude oils/mature condensates where the source material is of terrestrial origin, i.e. waxy plant and spore cuticles and resins.

5.5 Further analysis of the extractable organic matter in core D85-J

The presence of non-indigenous hydrocarbons in the carbonate rocks of unit B was noted in the earlier IKU report 21.3405.00/01/85. The presence of possible non-indigenous hydrocarbons in unit A of uppermost Jurassic age was also noted. It was indicated that these hydrocarbons were derived from different sources, although these sources could not be identified. In an attempt to clarify the situation, additional samples from units A, B and basal C were analysed using gas chromatography and combined gas chromatography - mass spectrometry (GC-MS).

The extraction data tables include data from the earlier IKU report indicated above and also data for the additional samples analysed for GC-MS (Tables 5.2-5.5). The GC-MS data are given in Tables 5.12 and 5.13.

The additional gas chromatographic data obtained on sample 29.44 m (Fig. 5.10) shows a similar saturated hydrocarbon distribution to those seen in the underlying carbonate. However, the aromatic hydrocarbons (Fig. 5.11), steranes and triterpanes (Fig. 5.12) are more typical of unit C. It is possible that some seepage of the more mobile hydrocarbons from the underlying carbonate ((Unit B) has altered the n-alkane pattern of the 29.44 m sample, although this cannot be proved with any certainty.

The GC-MS data for the hydrocarbons in these additional samples clearly distinguish two main sample groups. The three shallowest samples at 22.94 m, 27.94 m and 29.44 m from unit C have very similar biomarker distributions which suggest a relatively low thermal maturity, probably equivalent to the early oil window. The sterane mass chromatograms (m/z 217) show very prominent C₂₉ sterane peaks which would suggest an input of terrestrial organic matter and which would support an indigenous origin for the hydrocarbons in these samples.

The remaining samples from units B (31.35 m and 32.33 m), and A (33.50 m and 34.22 m) have biomarker distributions indicating a higher level of thermal maturation than was seen in the unit C samples, with most of the maturity-dependant ratios having values at or close to thermal stability. This would suggest a maturity equivalent to the main phase of oil generation. In all of these samples, the C₂₉ sterane peaks are

the C₂₇ sterane peaks are more prominent, which may suggest a more marine source kerogen type, or it may simply reflect increased maturity.

The deepest sample at 34.22 m differs from the other three unit A/B samples in a higher abundance of tricyclic terpanes, seen as the two doublets immediately before the C₂₇ hopane peaks. This difference could reflect a slightly higher thermal maturity in the 34.22 m sample, or it could reflect a slight difference in source input.

5.6 References

CONNAN, J. and CASSOU, A.M. 1980: Properties of gases and petroleum liquids derived from terrestrial kerogen at various maturation levels. Geochim. Cosmochim. Acta. 44, 1-23.

LARTER, S.R. and DOUGLAS, A.G. 1980: A pyrolysis-gas chromatographic method for kerogen typing. In: Douglas and Maxwell (eds.), Adv. in Org. Geochem. 1979, p. 579.

SCHOELL, M. 1983: Genetic characterization of natural gas. Amer. Assoc. Petrol. Geol. Bull. 67, 2225-2238.

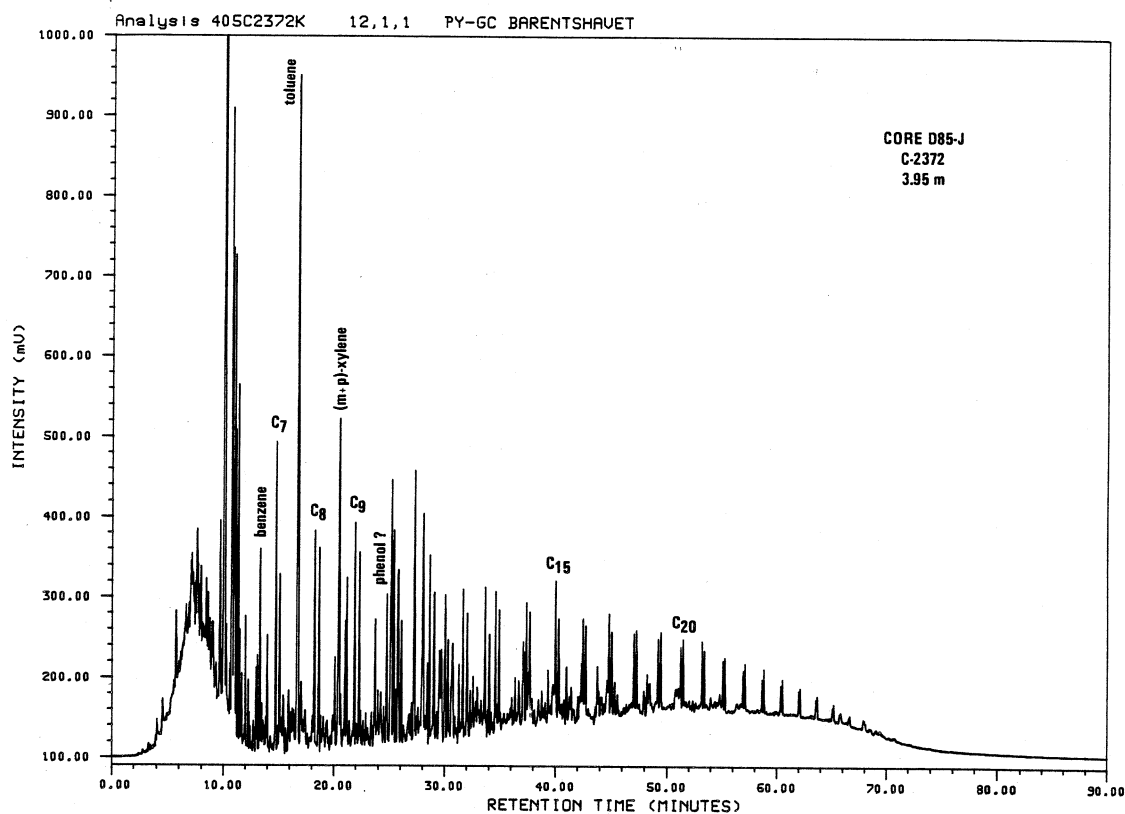


Fig.5.1a Pyrogram of core D85-J, 3.95 m.

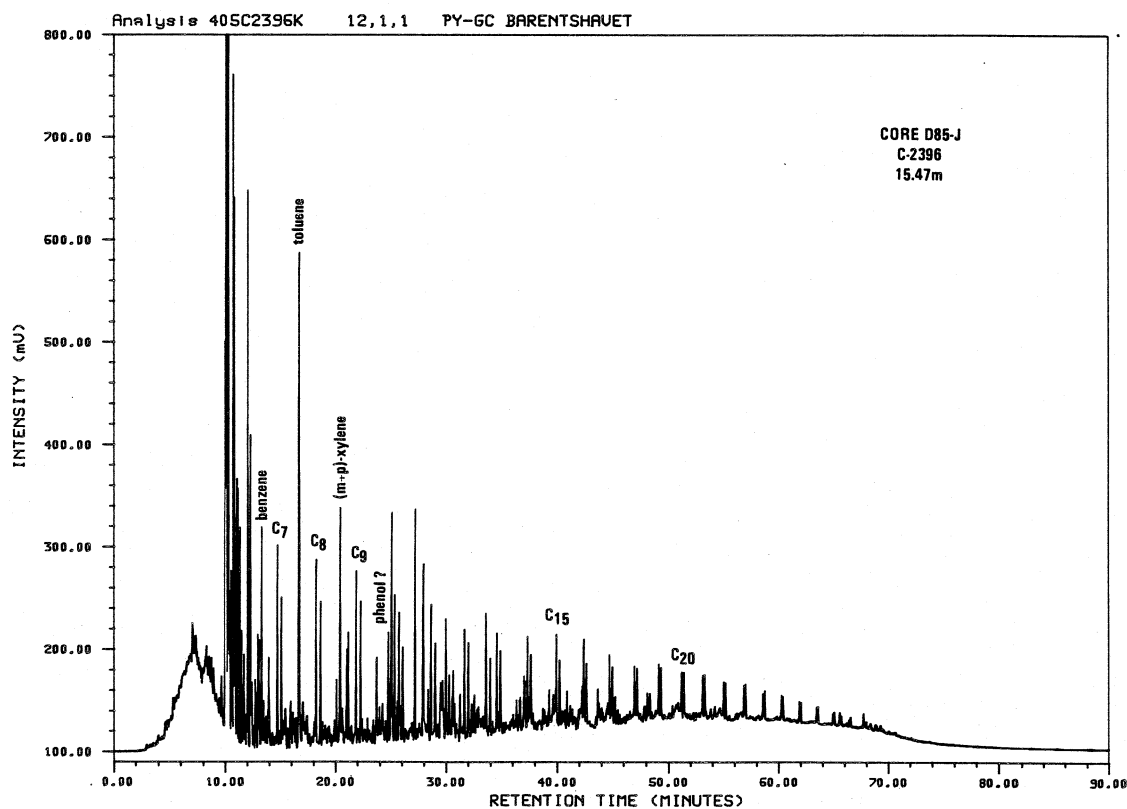


Fig.5.1b Pyrogram of core D85-J, 15.47 m.

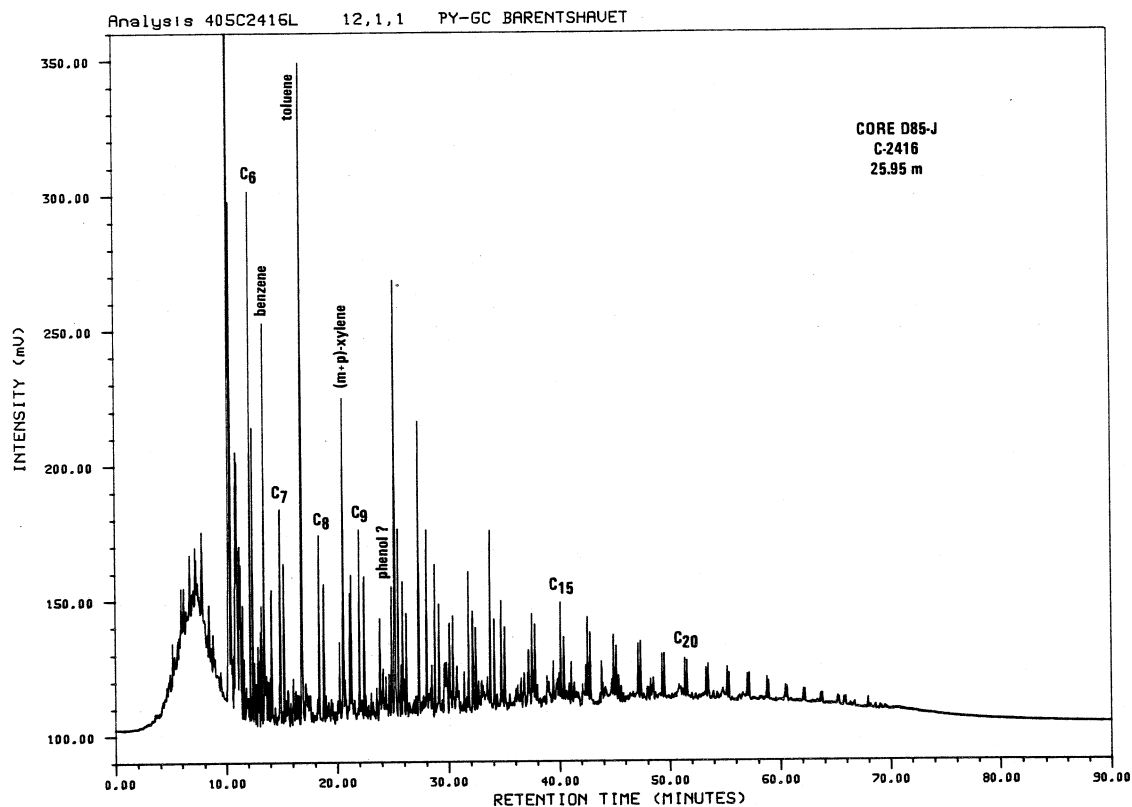


Fig.5.1c Pyrogram of core D85-J, 25.95 m.

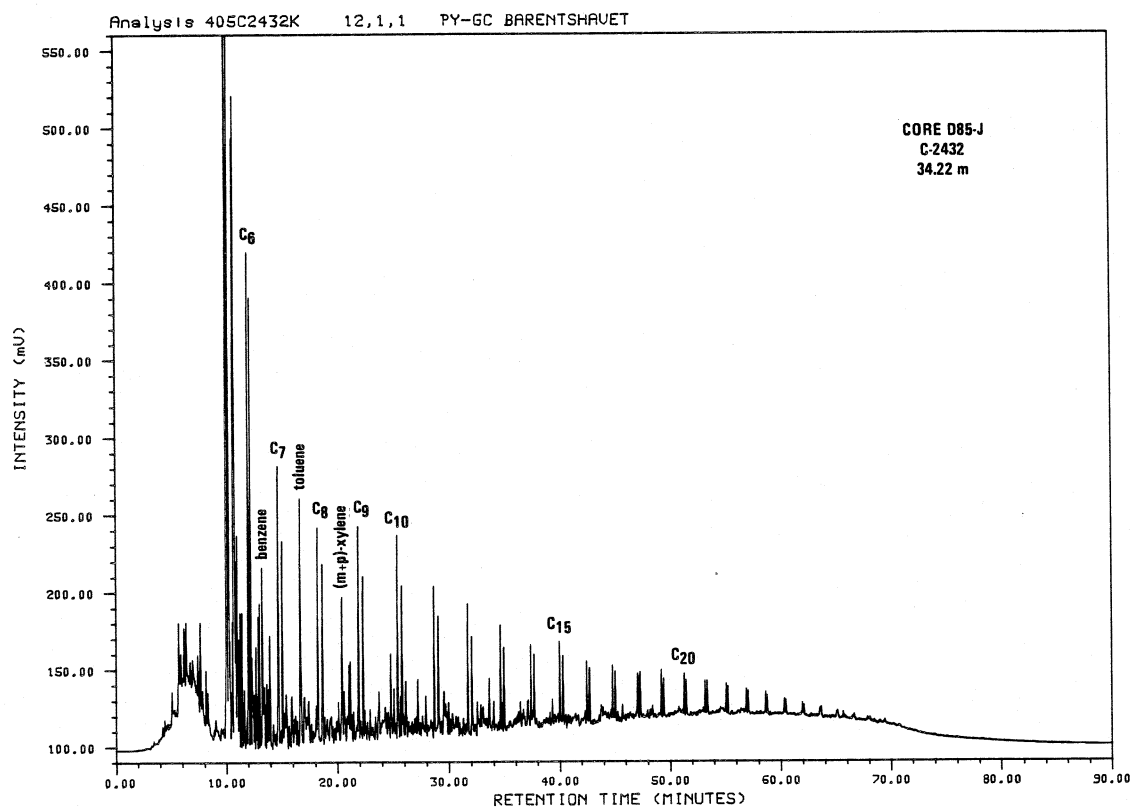


Fig.5.1d Pyrogram of core D85-J, 34.22 m.

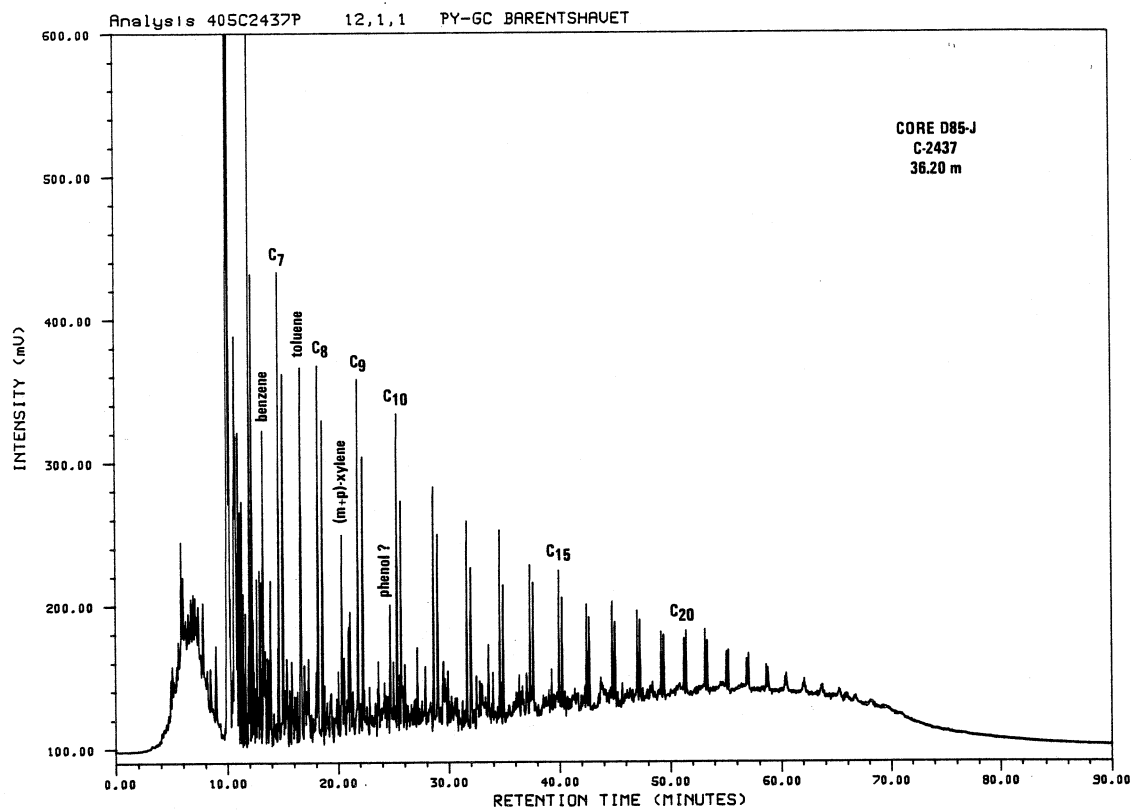


Fig.5.1e Pyrogram of core D85-J, 36.20 m.

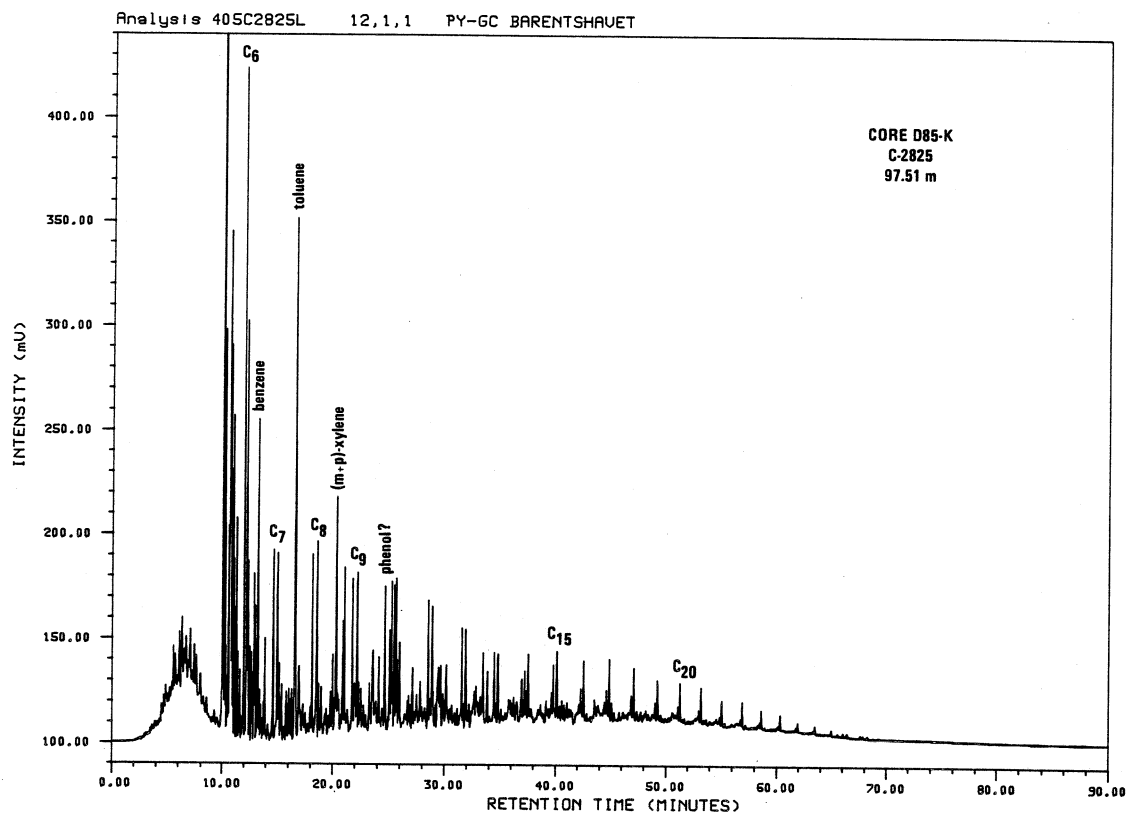


Fig.5.2a Pyrogram of core D85-K, 97.51 m.

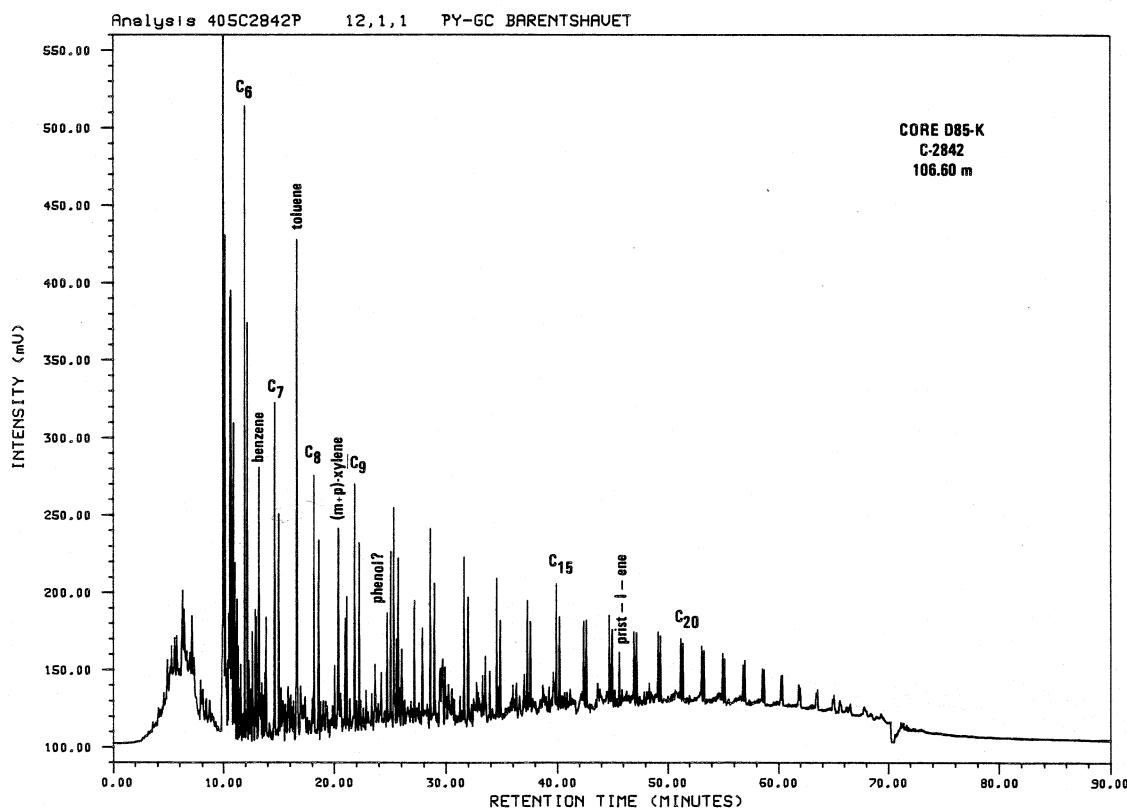


Fig.5.2b Pyrogram of core D85-K, 106.60 m.

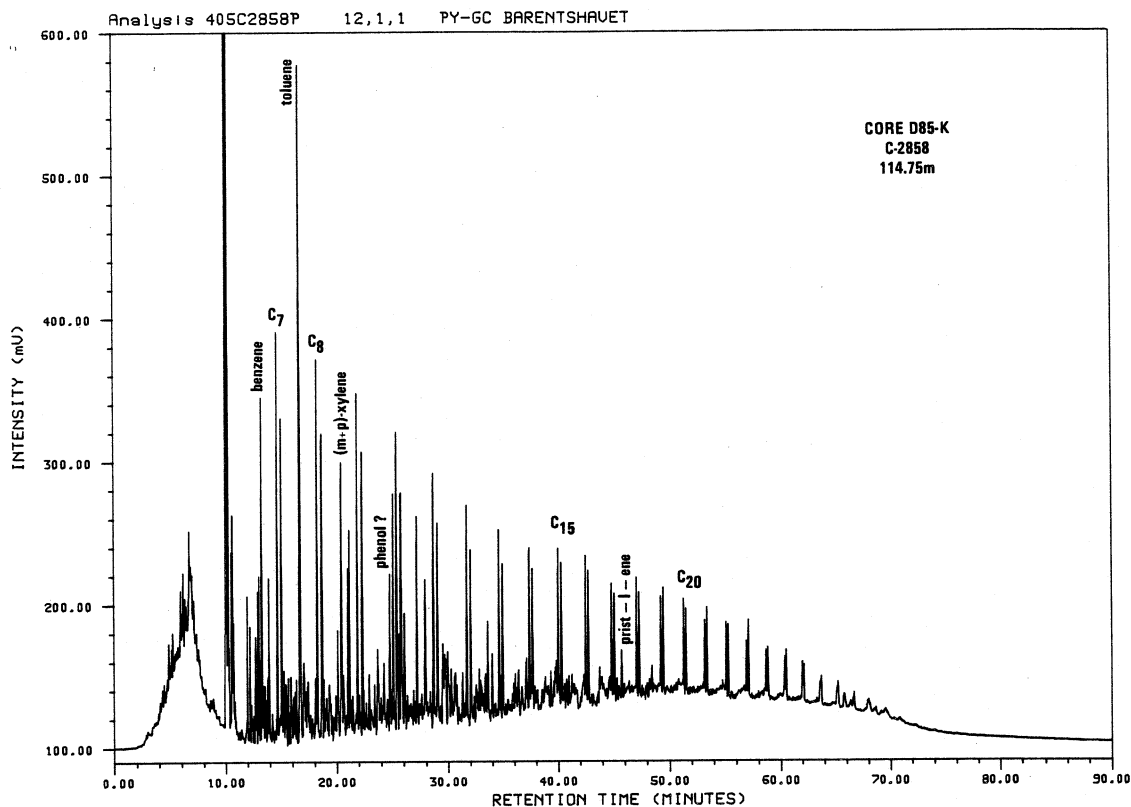


Fig. 5.2 c Pyrogram of core D85-K, 114.75 m.

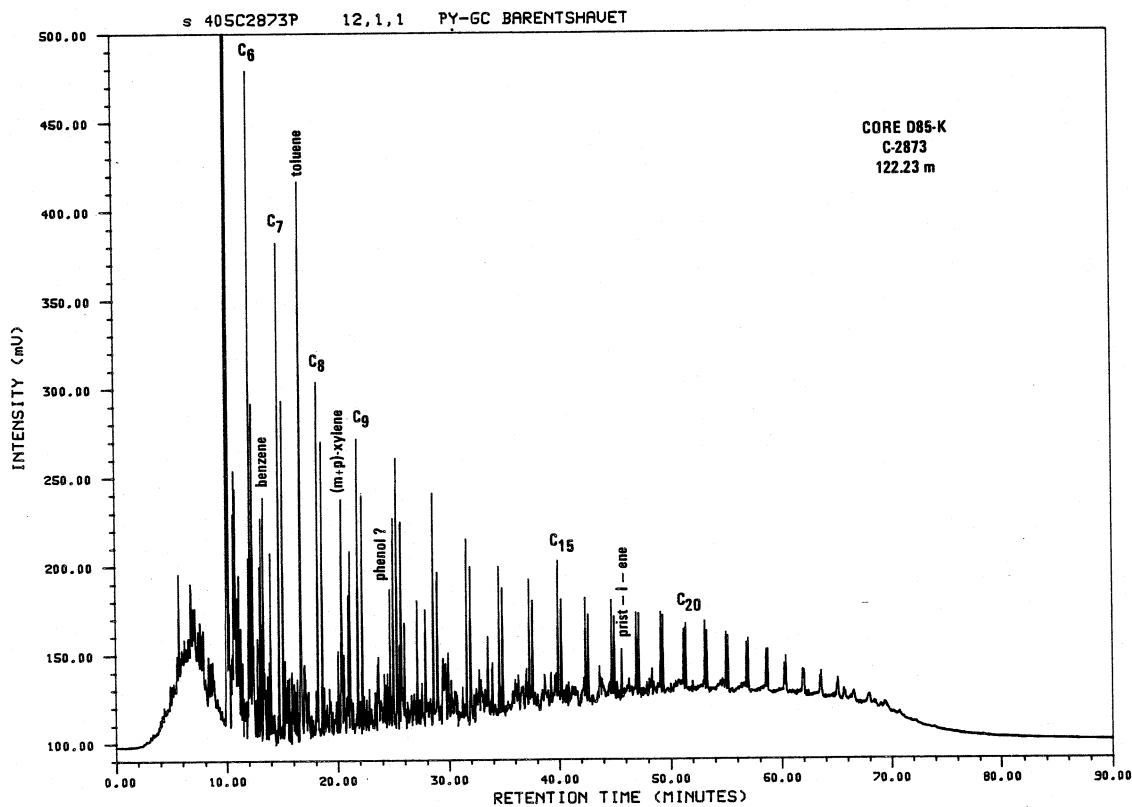


Fig. 5.2 d Pyrogram of core D85-K, 122.23 m.

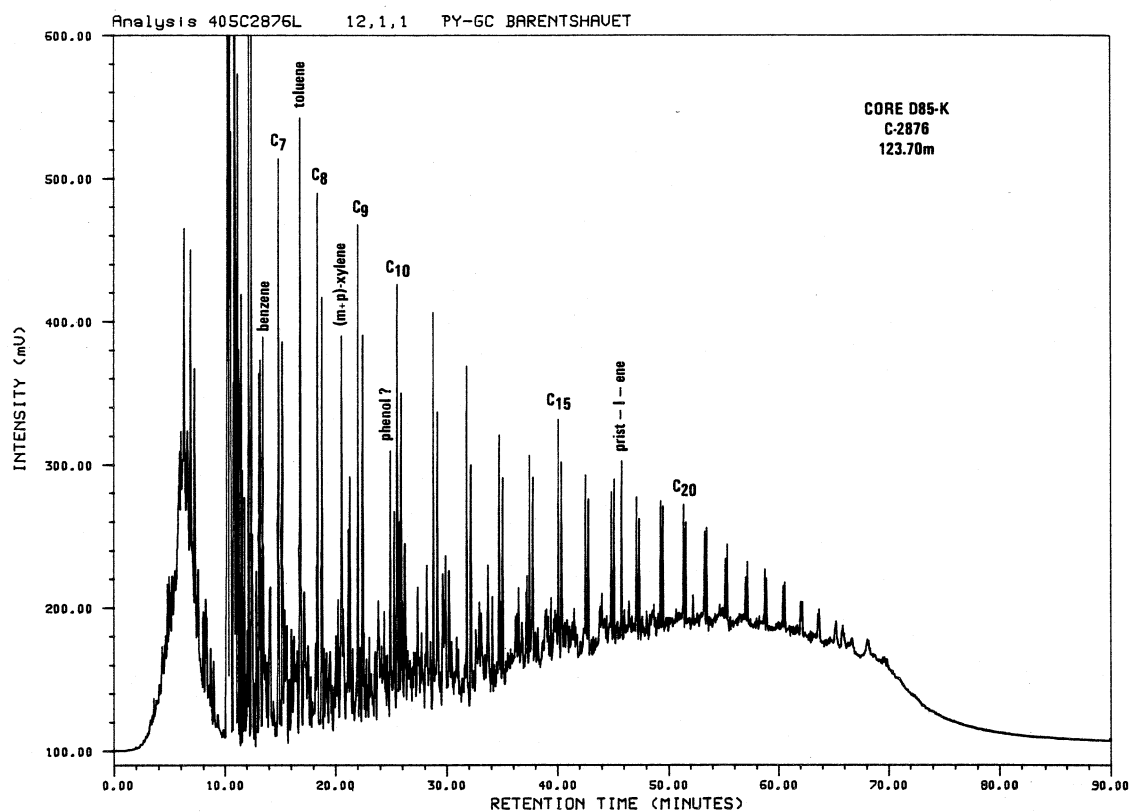


Fig.5.2e Pyrogram of core D85-K, 123.70 m.

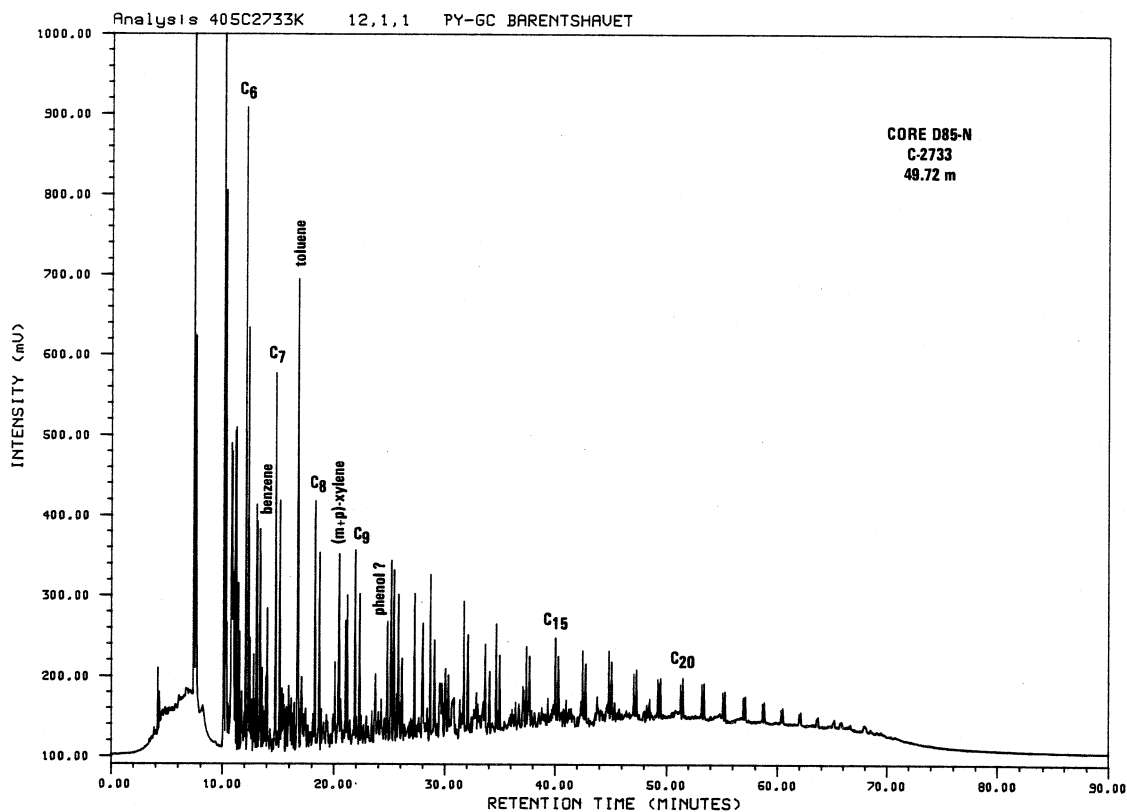


Fig.5.3a Pyrogram of core D85-N, 49.72 m.

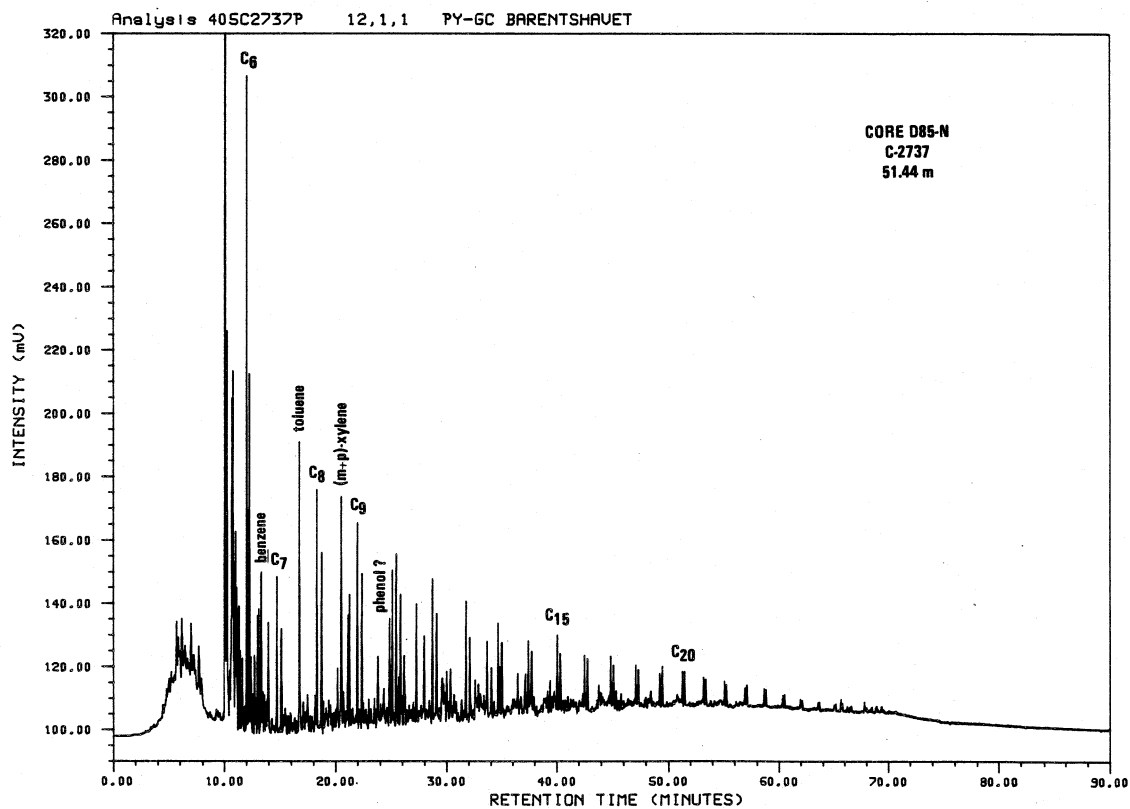


Fig.5.3b Pyrogram of core D85-N, 51.44 m.

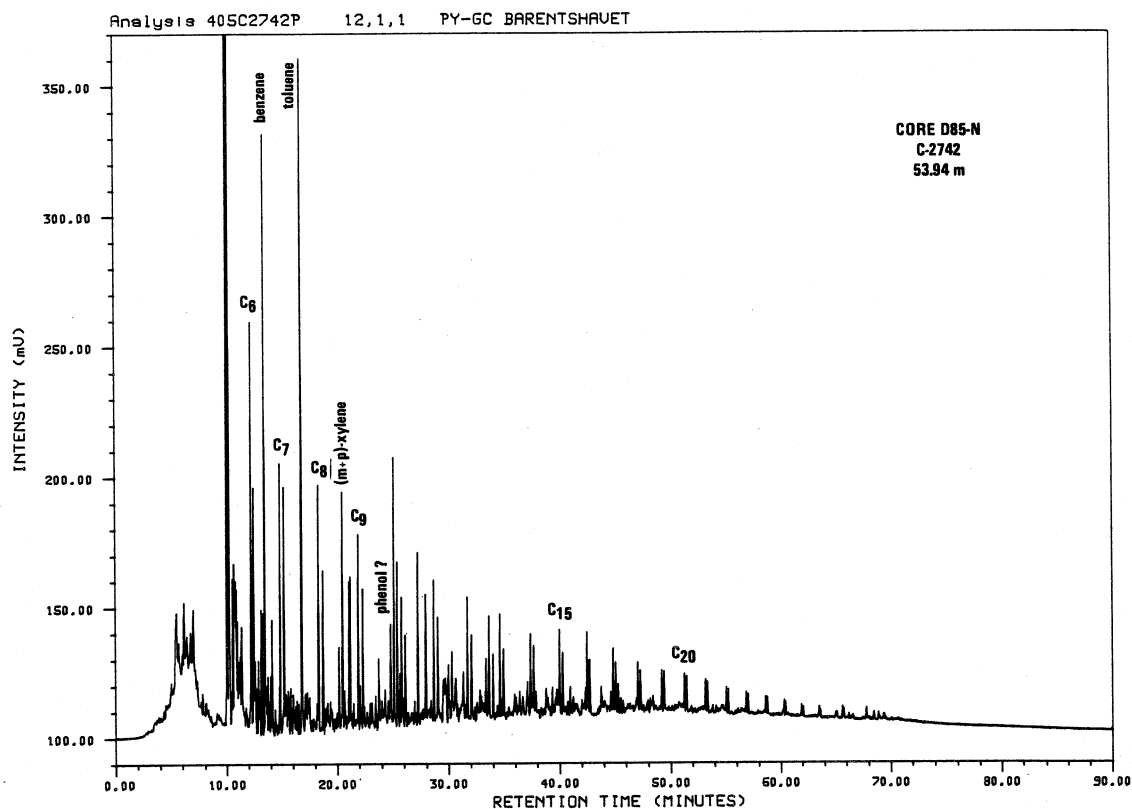


Fig. 5.3 c Pyrogram of core D85-N, 53.94 m.

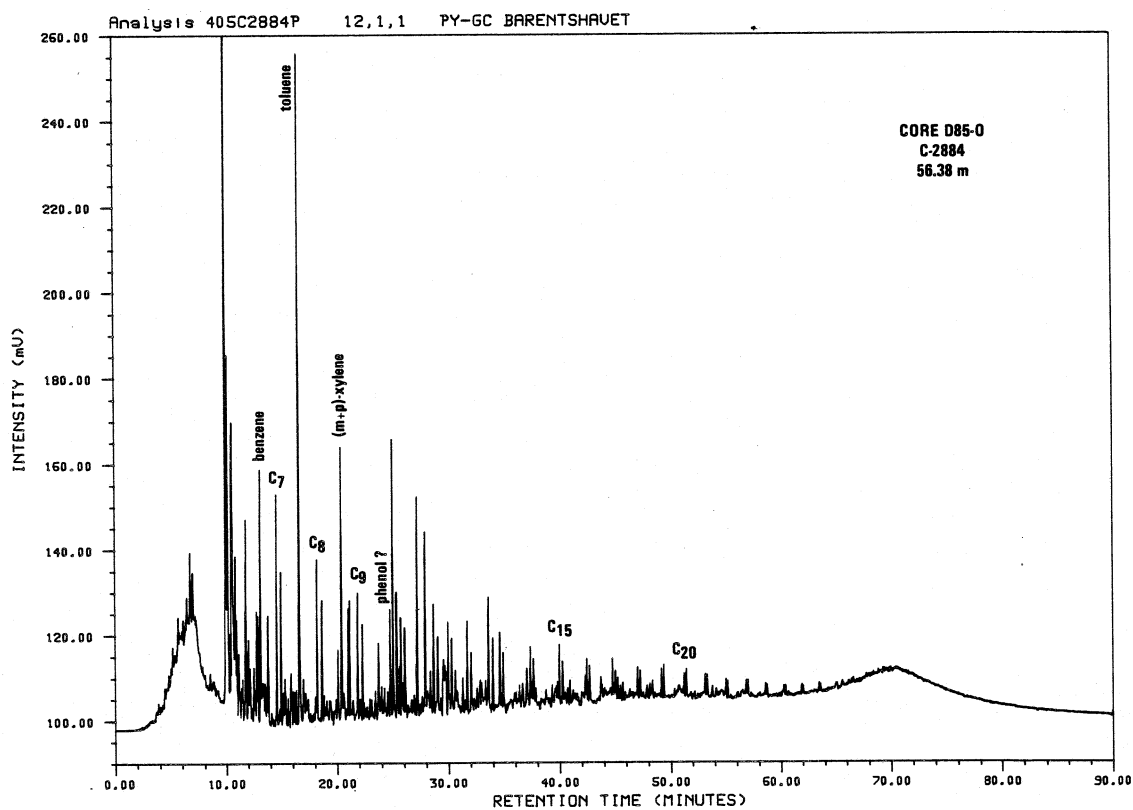


Fig. 5.4 Pyrogram of core D85-0, 56.38 m.

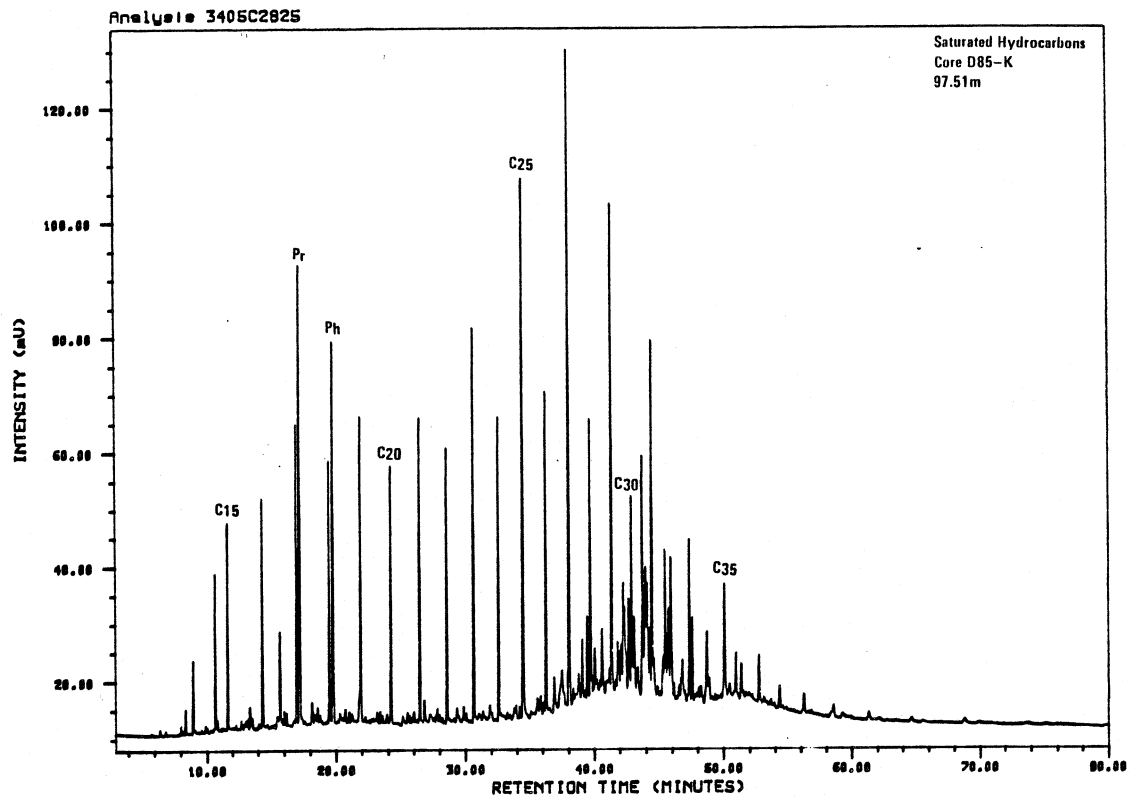


Fig.5.5a Gas chromatogram of saturated hydrocarbons from core D85-K,
97.51 m.

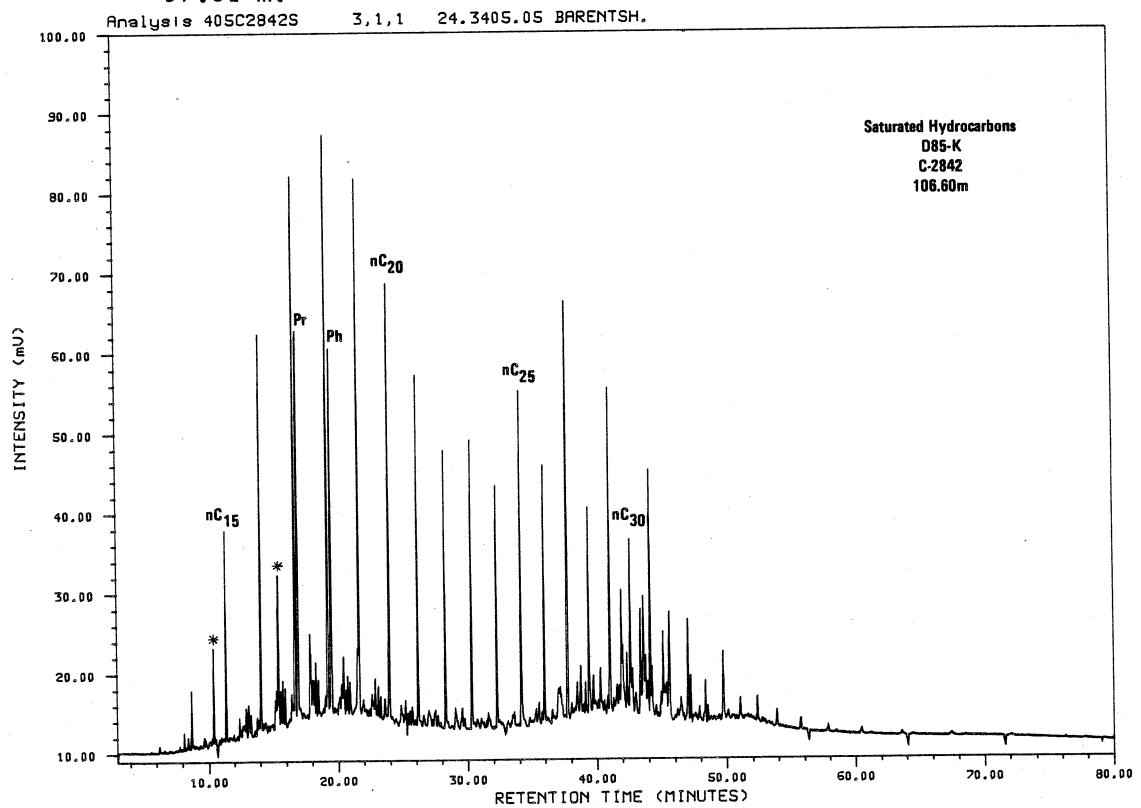


Fig.5.5b Gas chromatogram of saturated hydrocarbons from core D85-K,
106.60 m.

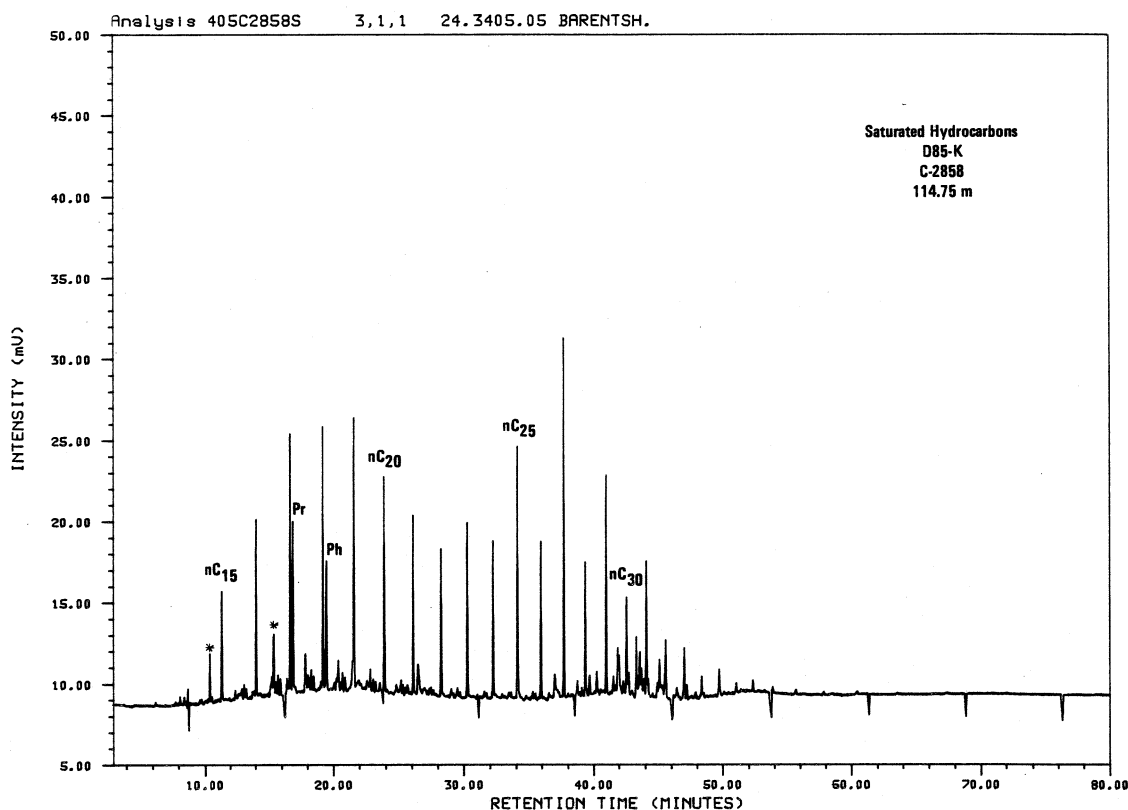


Fig.5.5c Gas chromatogram of saturated hydrocarbons from core D85-K, 114.75 m.

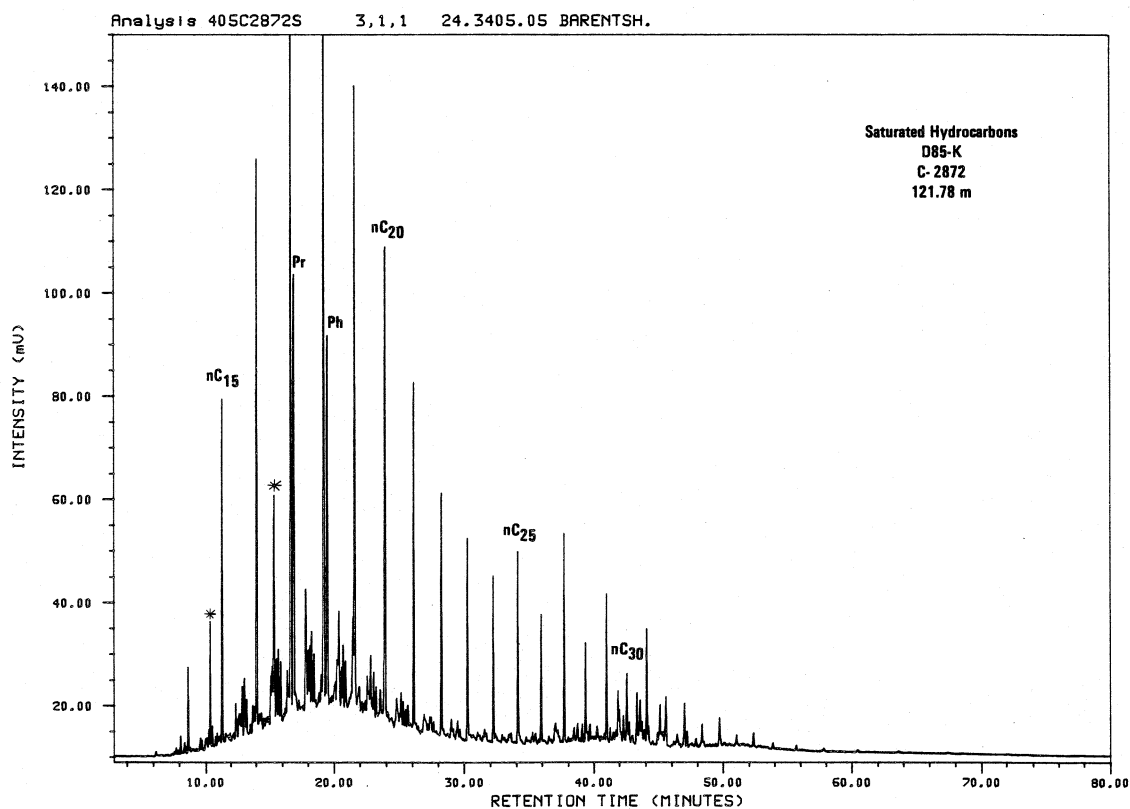


Fig.5.5d Gas chromatogram of saturated hydrocarbons from core D85-K, 121.78 m.

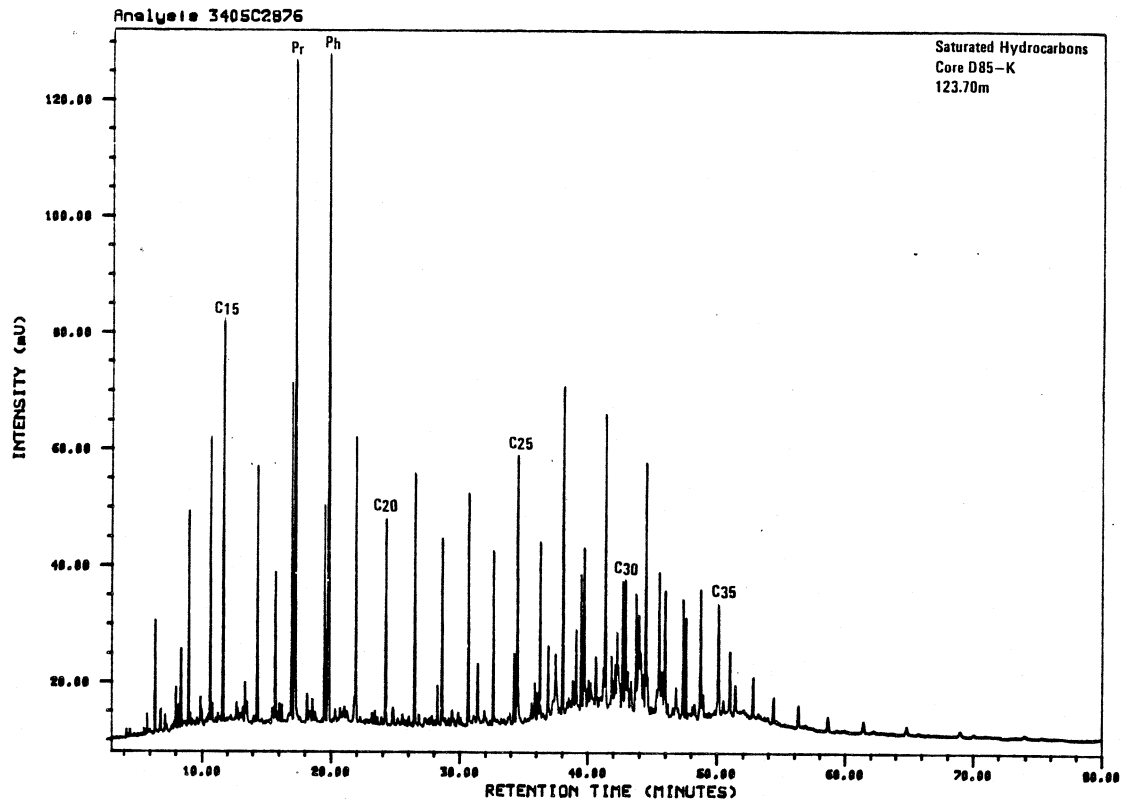


Fig.5.5e Gas chromatogram of saturated hydrocarbons from core D85-K,
123.70 m.

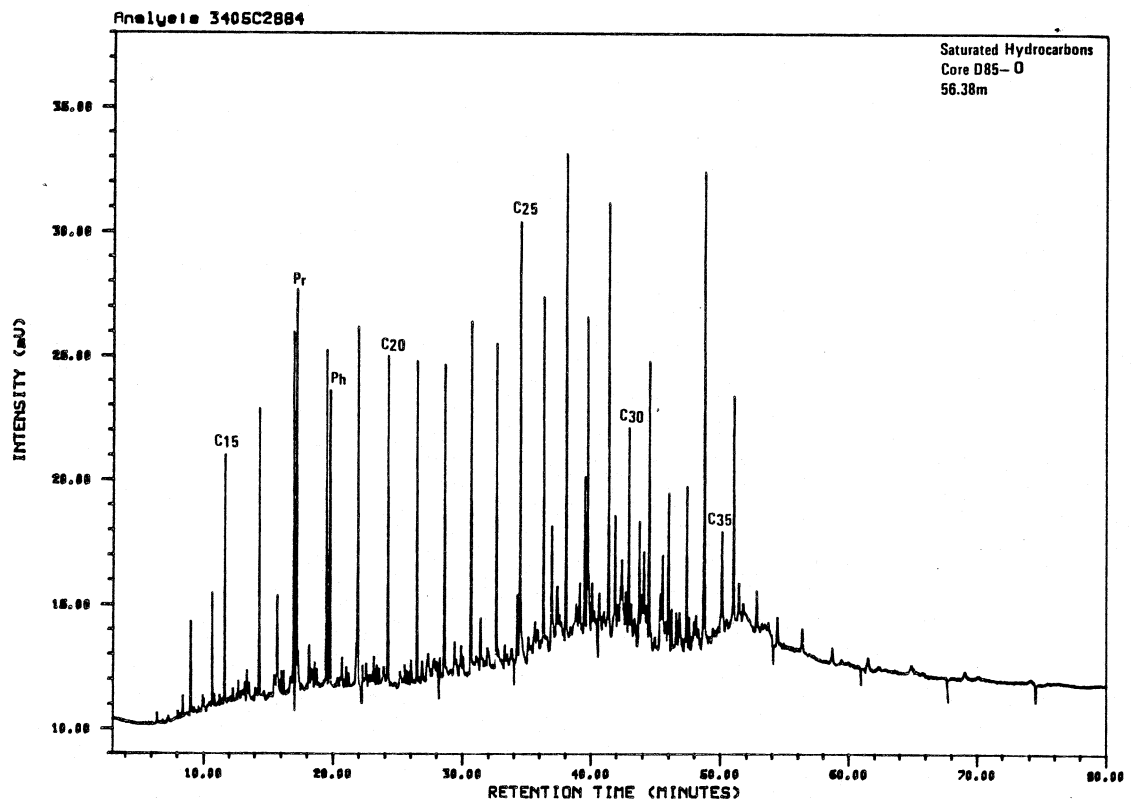


Fig.5.5f Gas chromatogram of saturated hydrocarbons from core D85-0,
56.38 m.

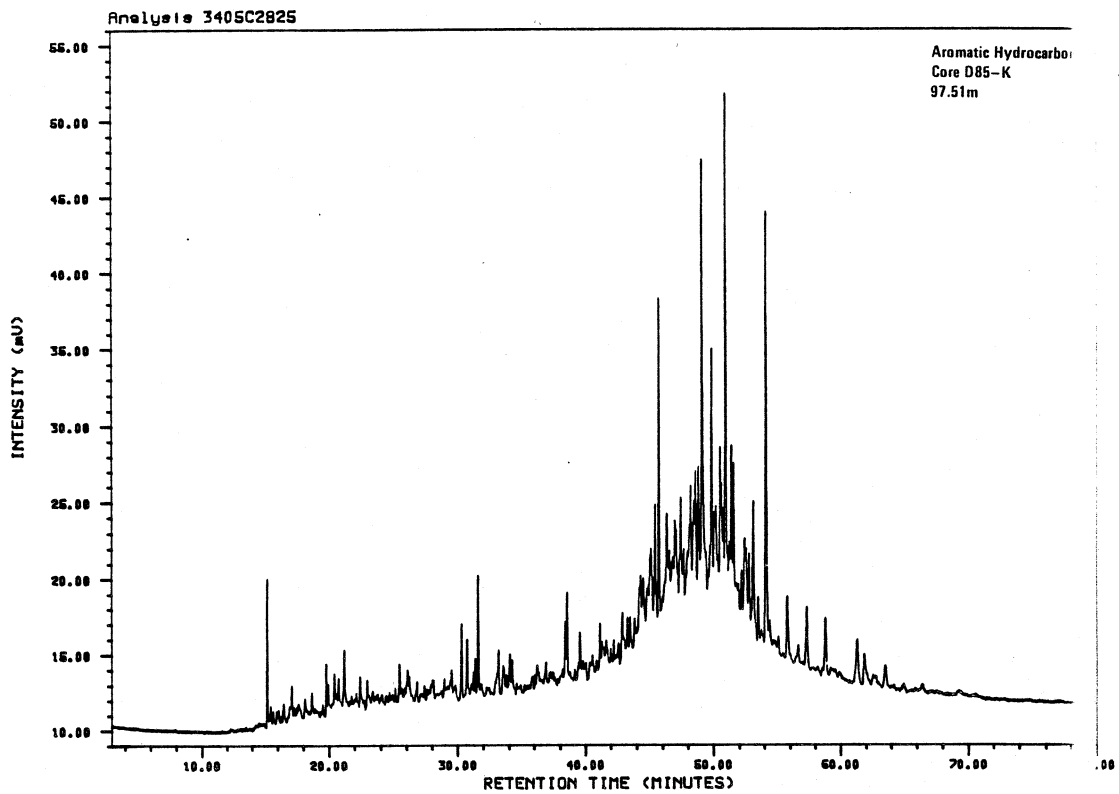


Fig.5.6a Gas chromatogram of aromatic hydrocarbons from core D85-K,
97.51 m.

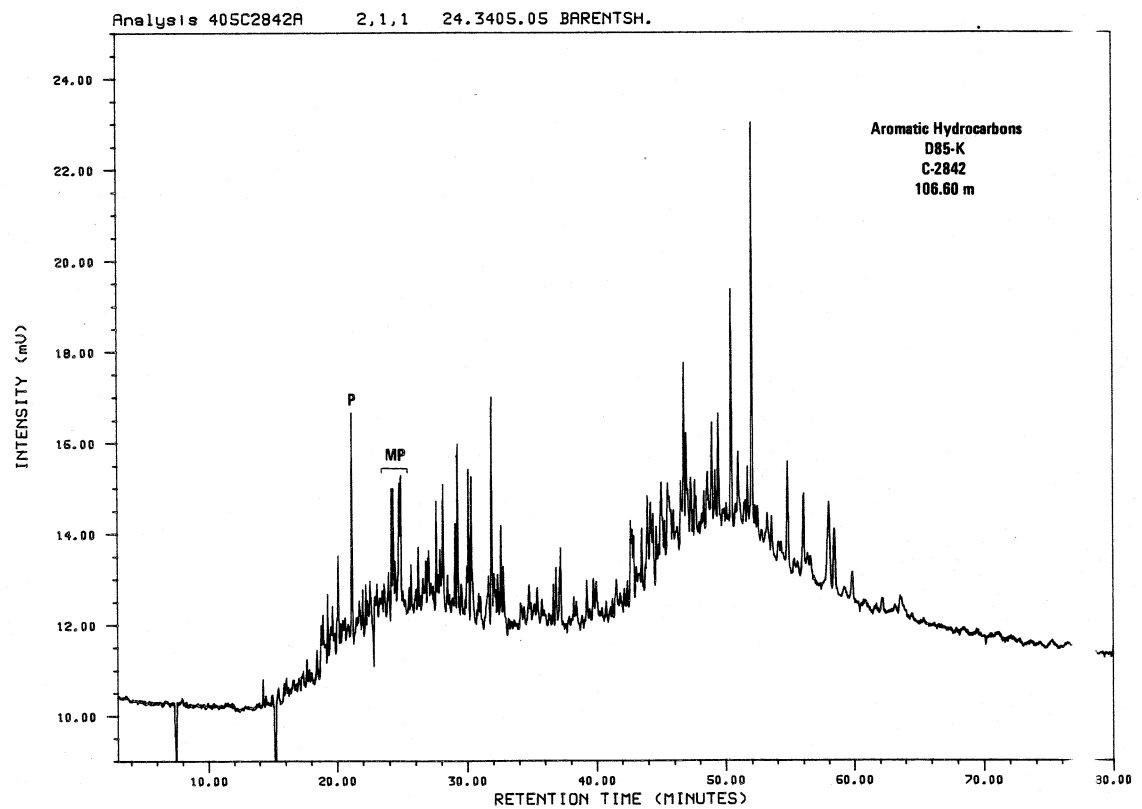


Fig.5.6b Gas chromatogram of aromatic hydrocarbons from core D85-K,
106.60 m.

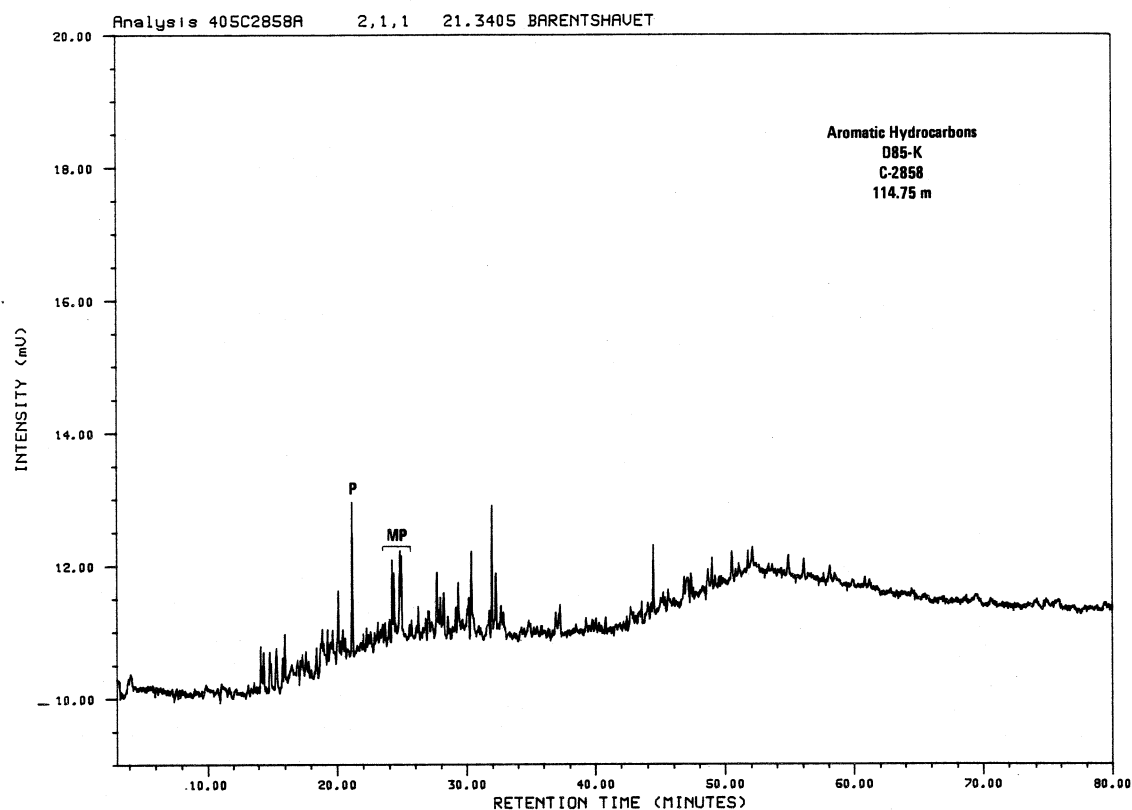


Fig.5.6c Gas chromatogram of aromatic hydrocarbons from core D85-K,
114.75 m.

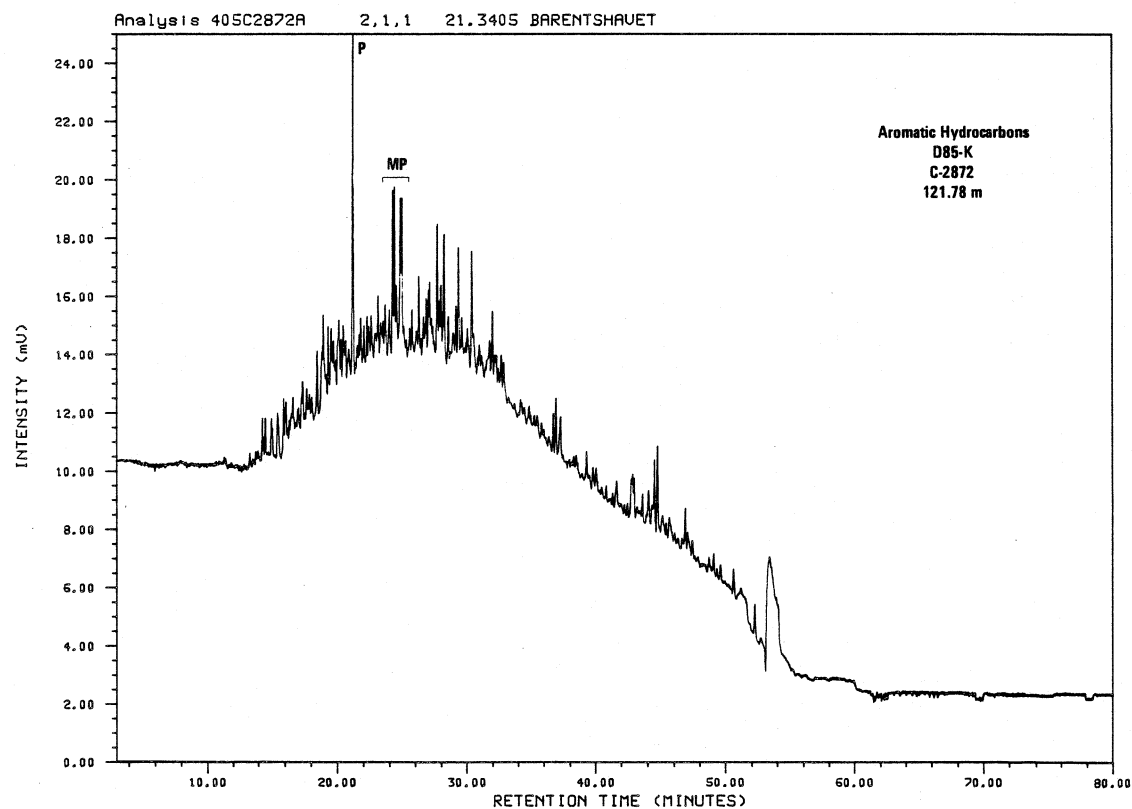


Fig.5.6d Gas chromatogram of aromatic hydrocarbons from core D85-K,
121.78 m.

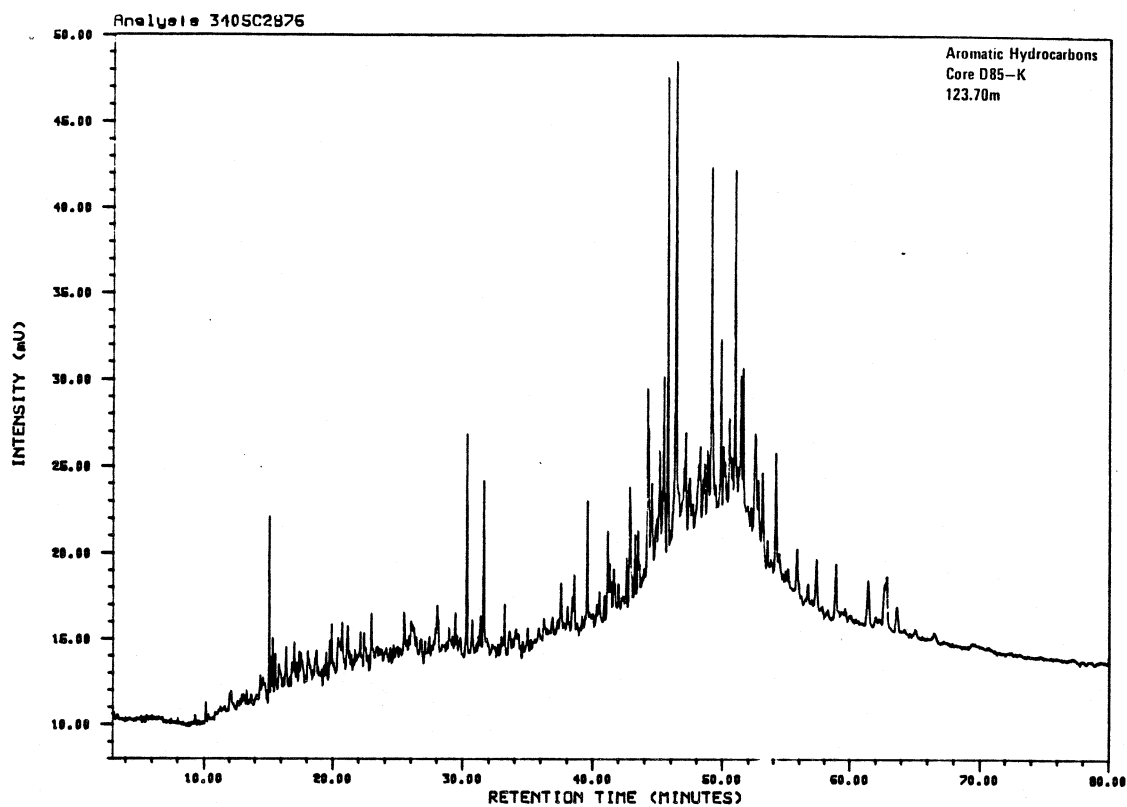


Fig.5.6e Gas chromatogram of aromatic hydrocarbons from core D85-K, 123.70 m.

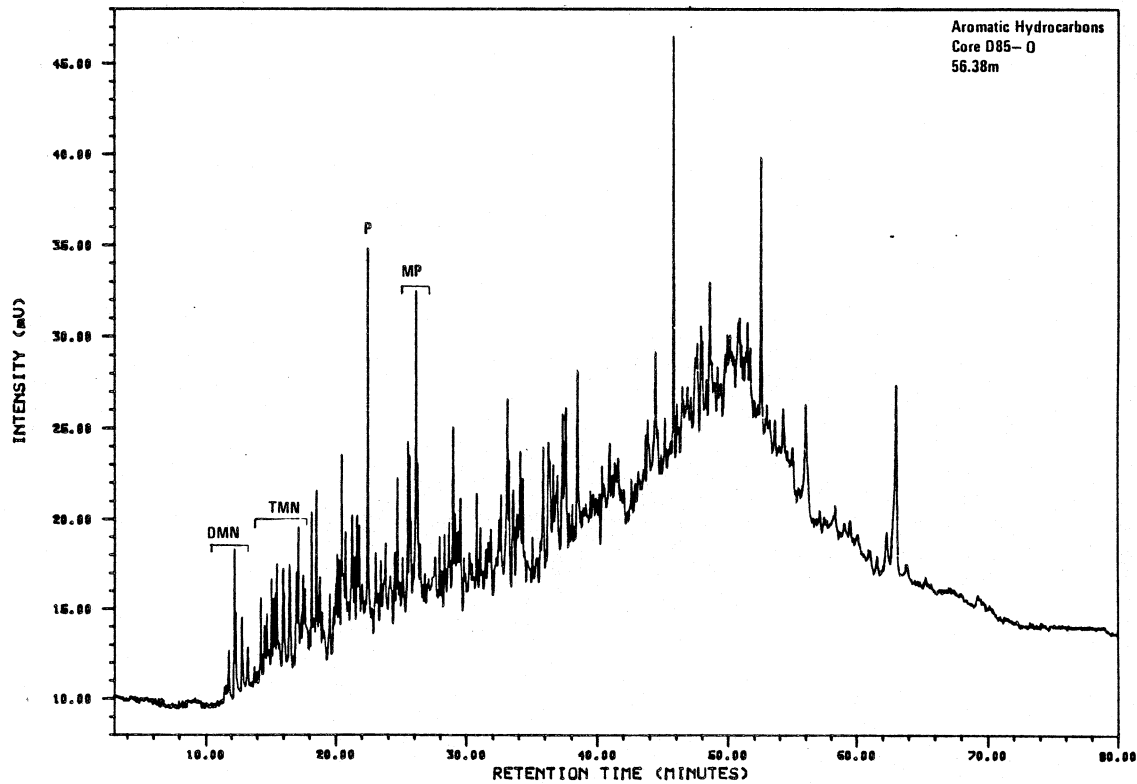


Fig.5.6f Gas chromatogram of aromatic hydrocarbons from core D85-0, 56.38 m.

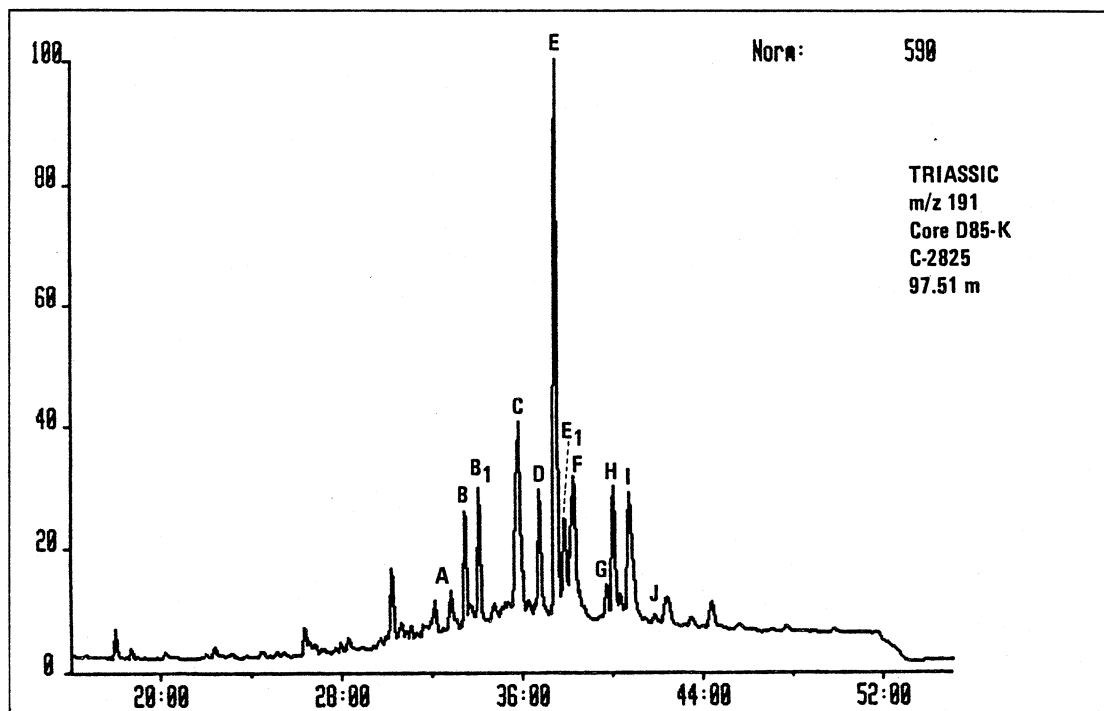


Fig. 5.7a (1) Mass chromatogram of terpanes (m/z 191), core D85-K, 97.51 m.

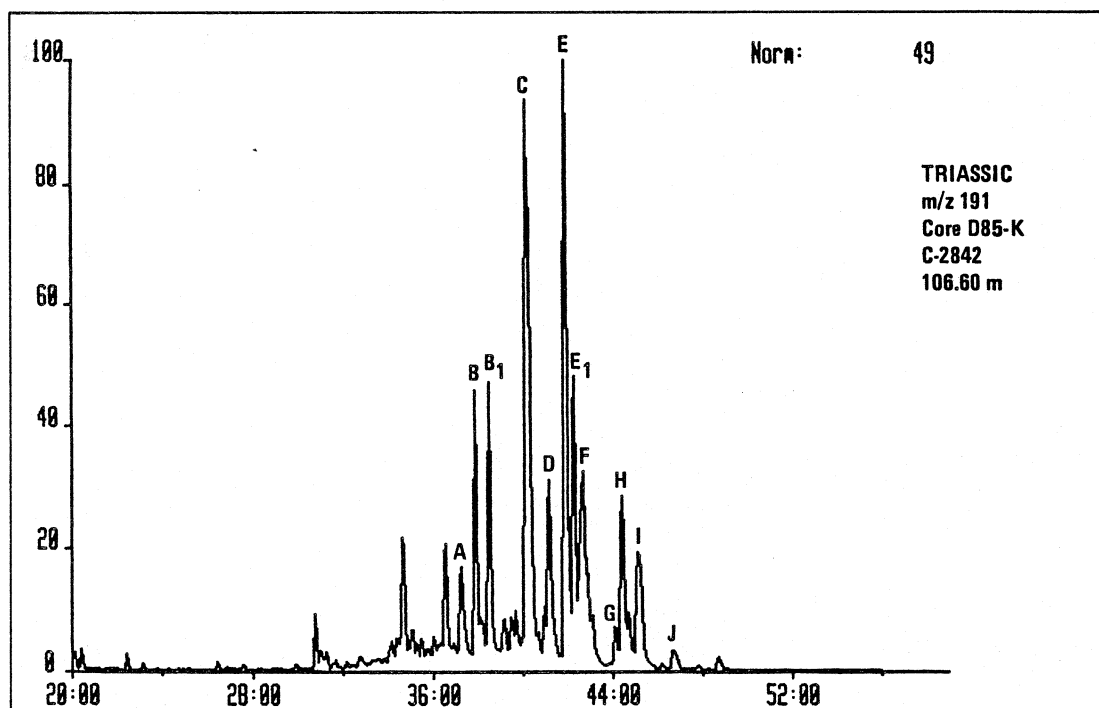


Fig. 5.7a (2) Mass chromatogram of terpanes (m/z 191), core D85-K, 106.60 m.

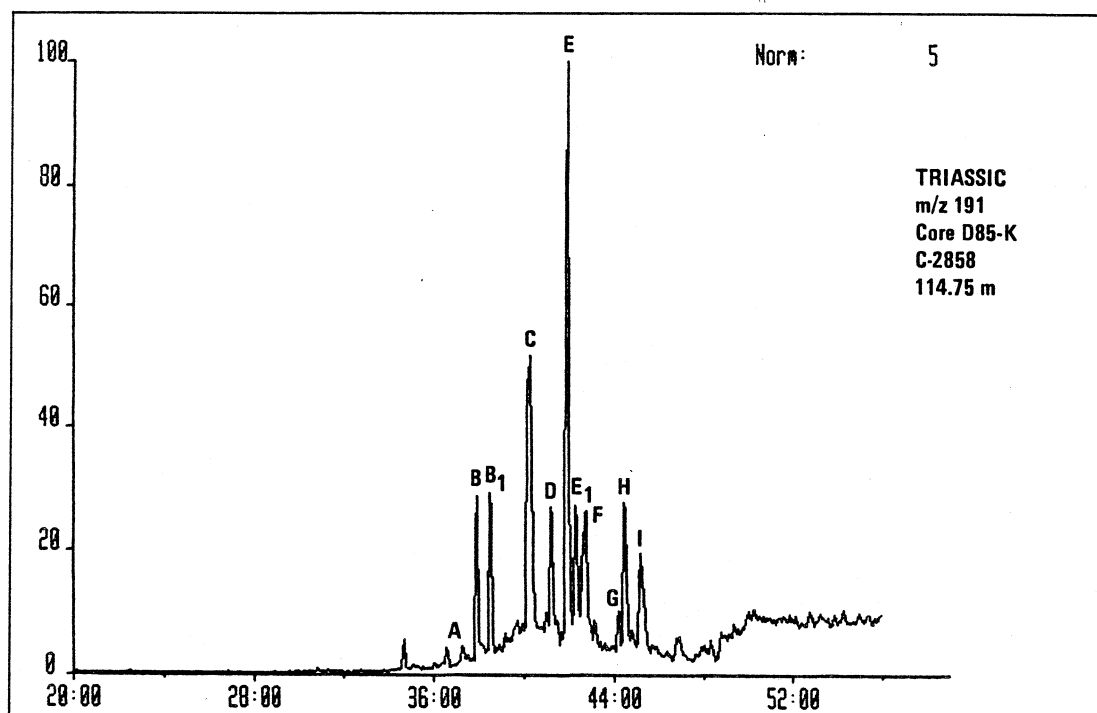


Fig. 5.7a (3) Mass chromatogram of terpanes (m/z 191), core D85-K, 114.75 m.

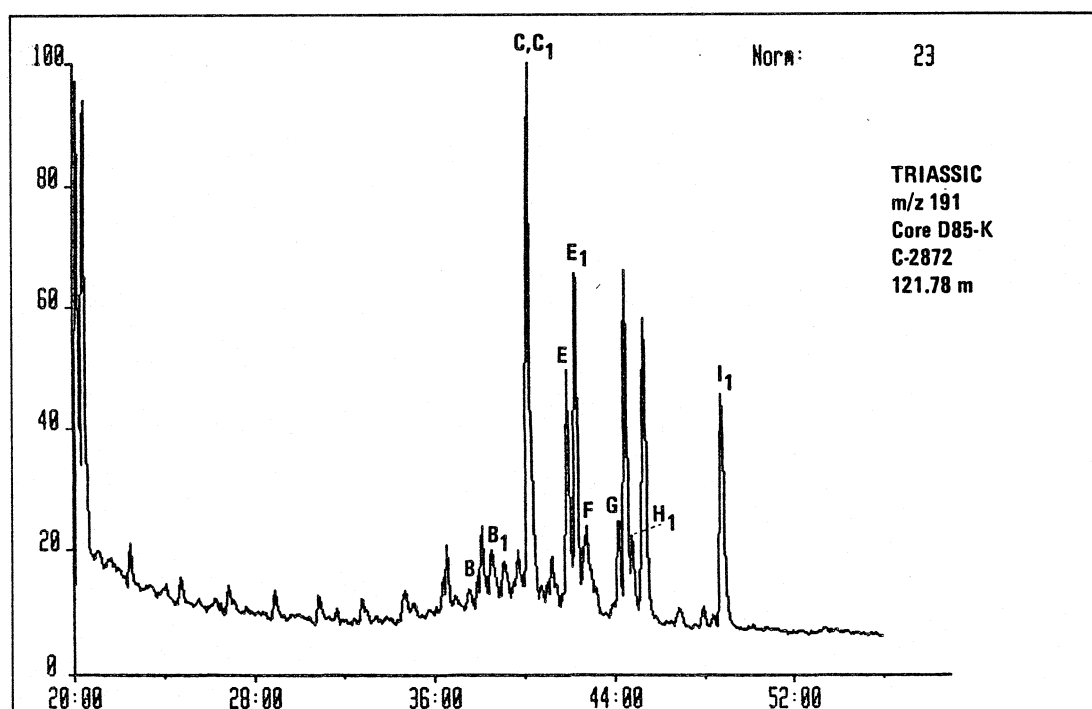


Fig. 5.7a (4) Mass chromatogram of terpanes (m/z 191), core D85-K, 121.78 m.

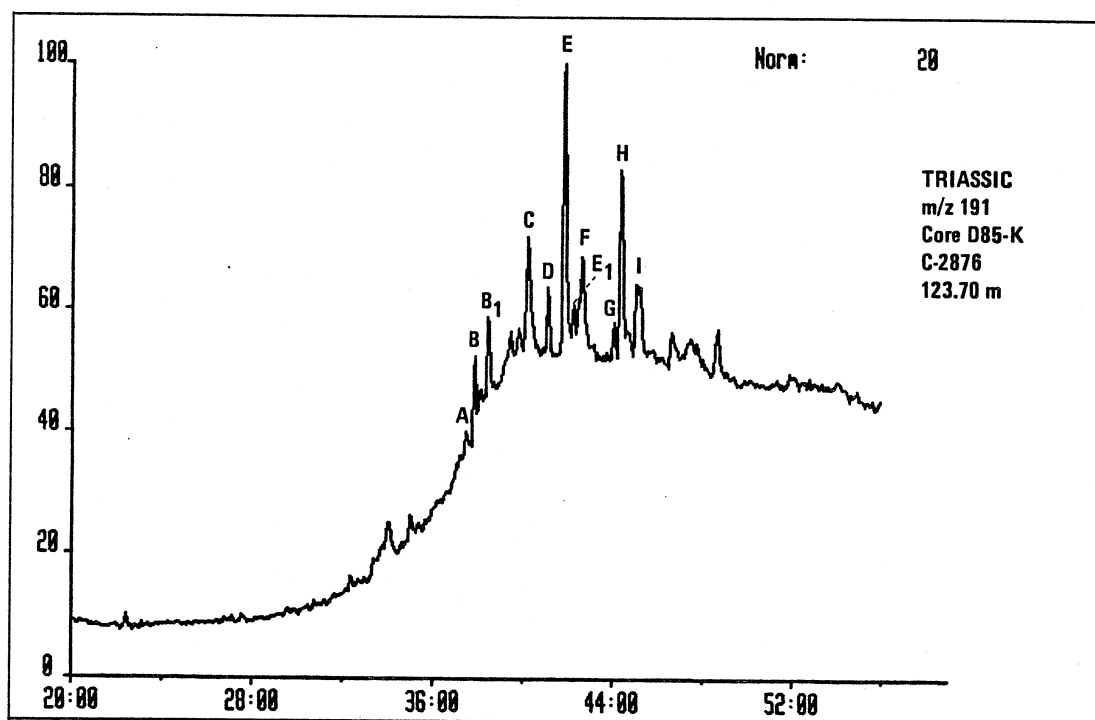


Fig. 5.7a (5) Mass chromatogram of terpanes (m/z 191), core D85-K, 123.70 m.

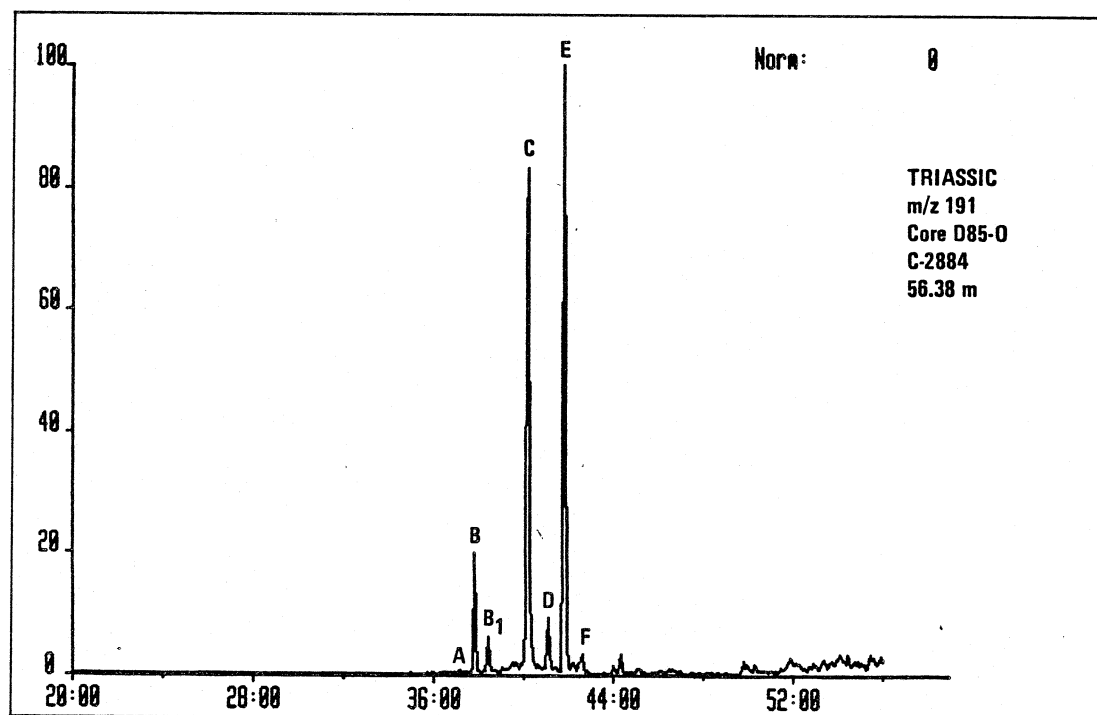


Fig. 5.7a (6) Mass chromatogram of terpanes (m/z 191), core D85-O, 56.38 m.

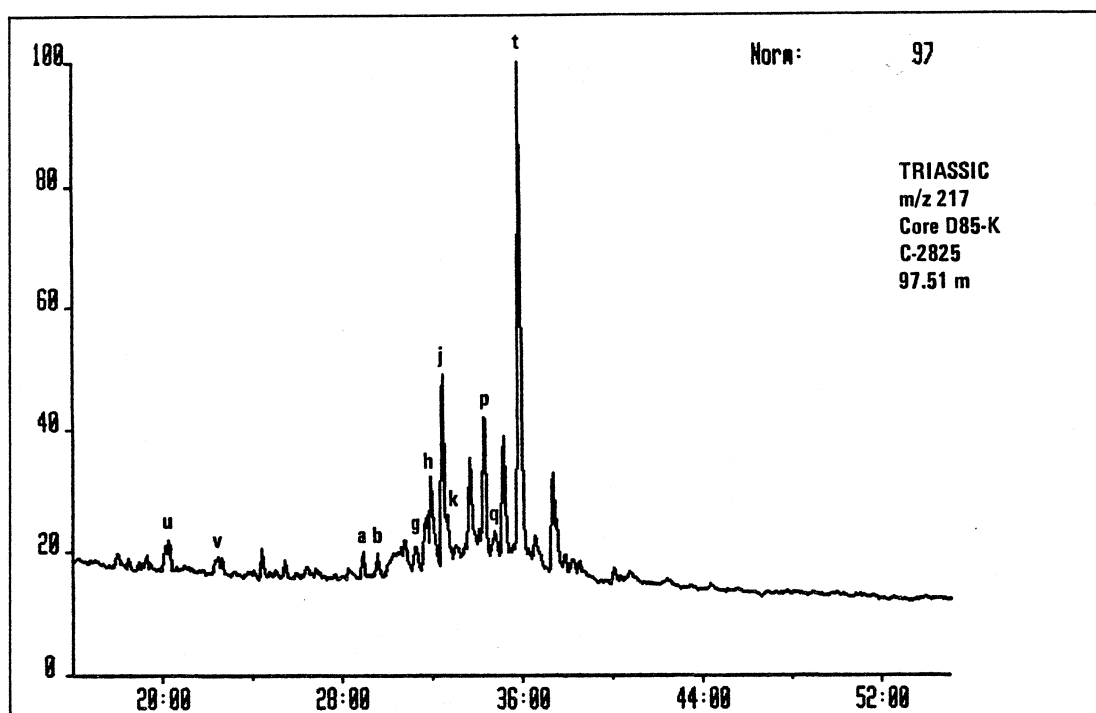


Fig. 5.7b (1) Mass chromatogram of steranes, core D85-K,
97.51 m.

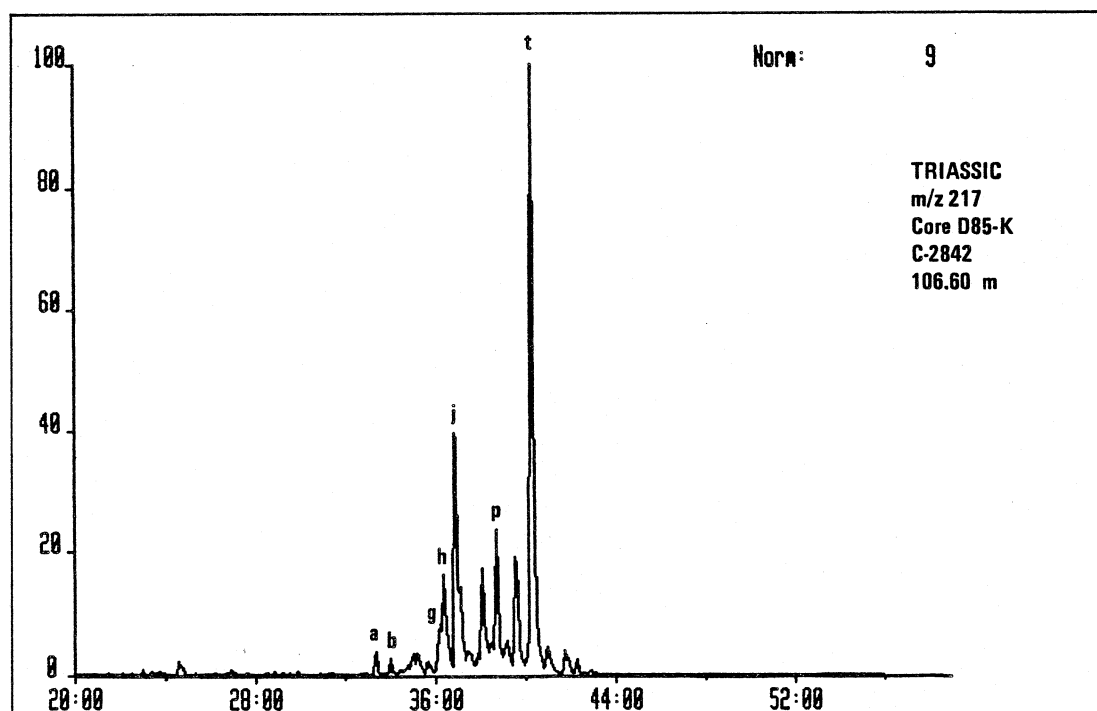


Fig. 5.7b (2) Mass chromatogram of steranes, core D85-K,
106.60 m.

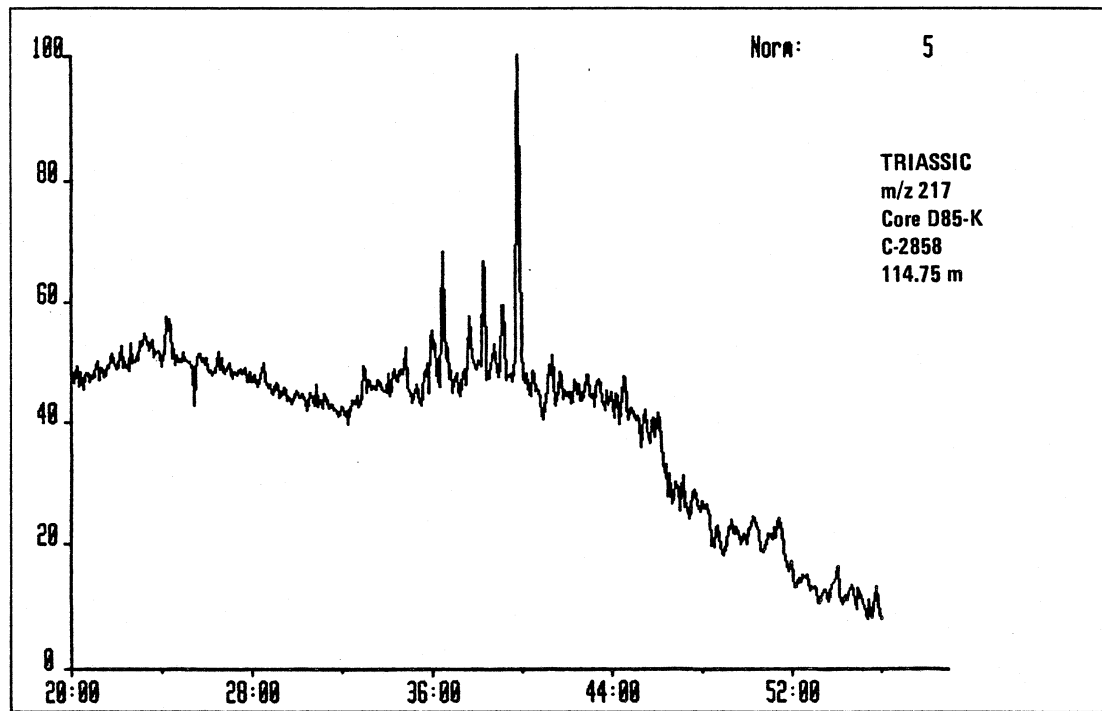


Fig. 5.7b (3) Mass chromatogram of steranes, core D85-K,
114.75 m.

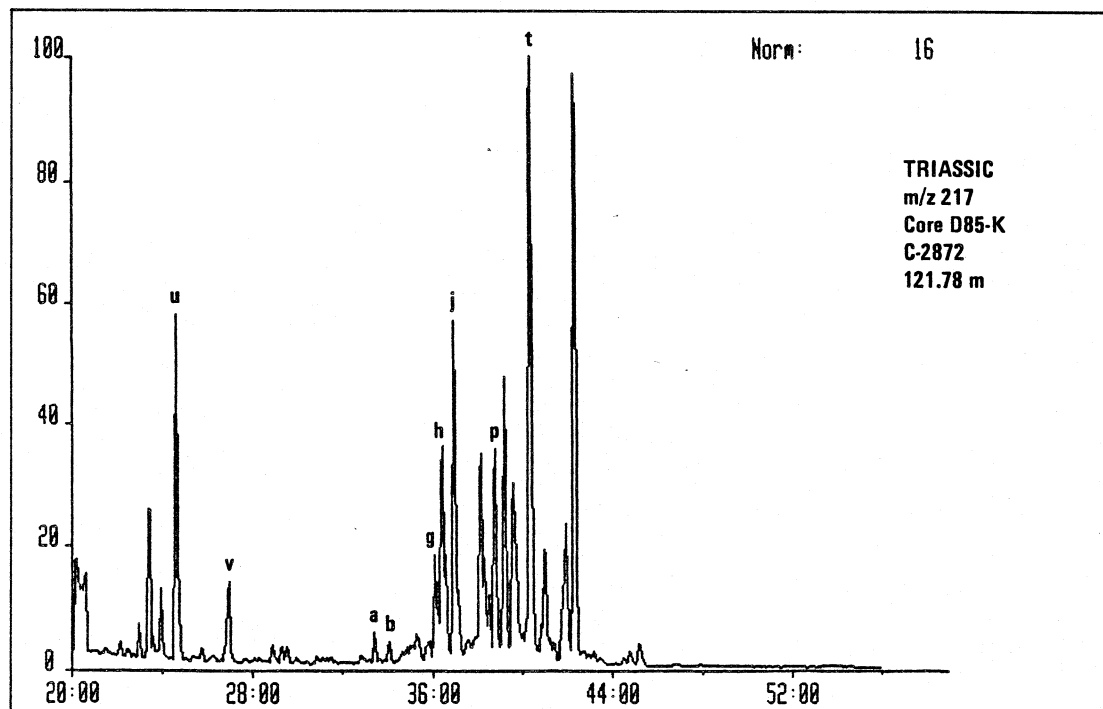


Fig. 5.7b (4) Mass chromatogram of steranes, core D85-K,
121.78 m.

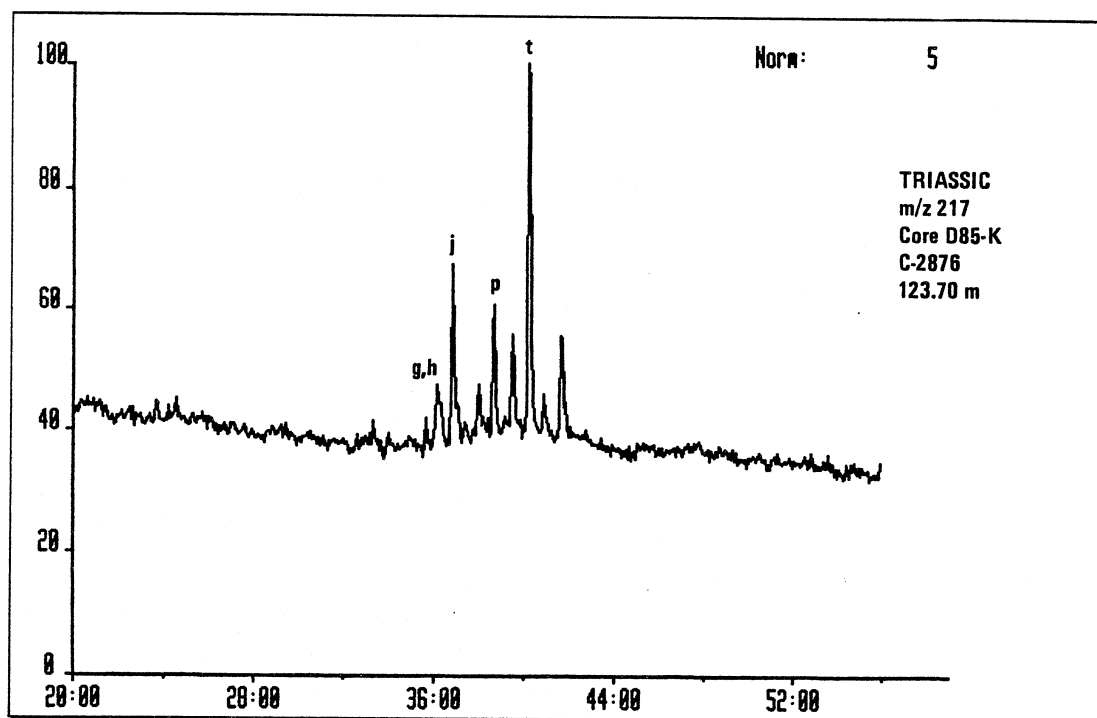
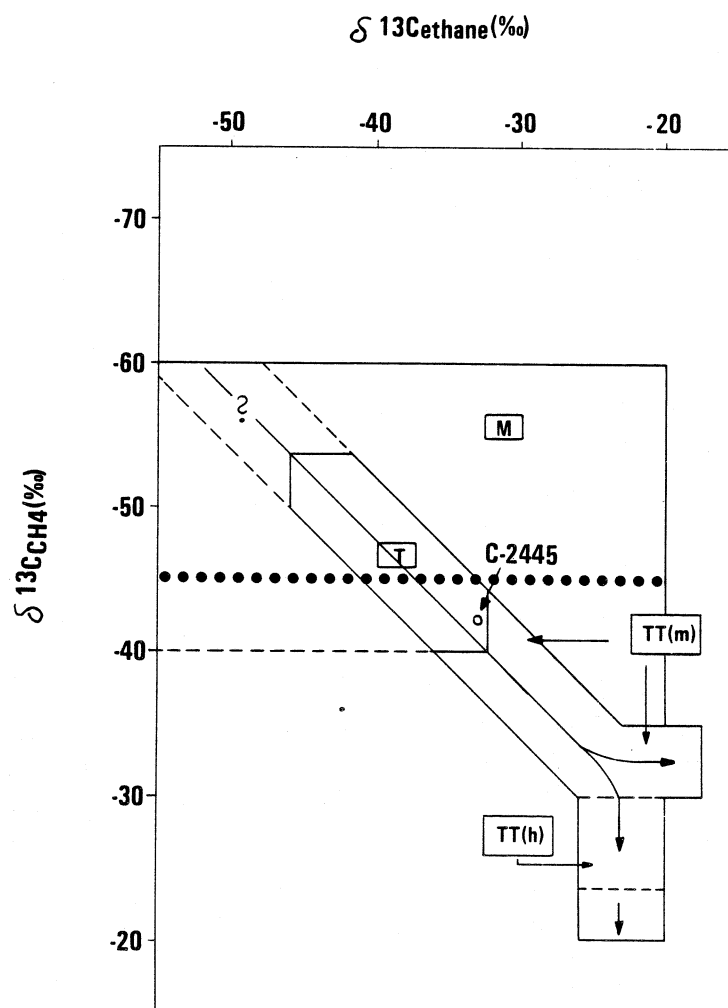


Fig. 5.7b (5) Mass chromatogram of steranes, core D85-K,
123.70 m.



C-2445 plots in the zone of gases associated with crude oils (T). C-2451 plots somewhere along the dotted line. (After Schoell, 1983).

Fig. 5.8 Plot of $\delta^{13}\text{C}$ for methane and ethane.

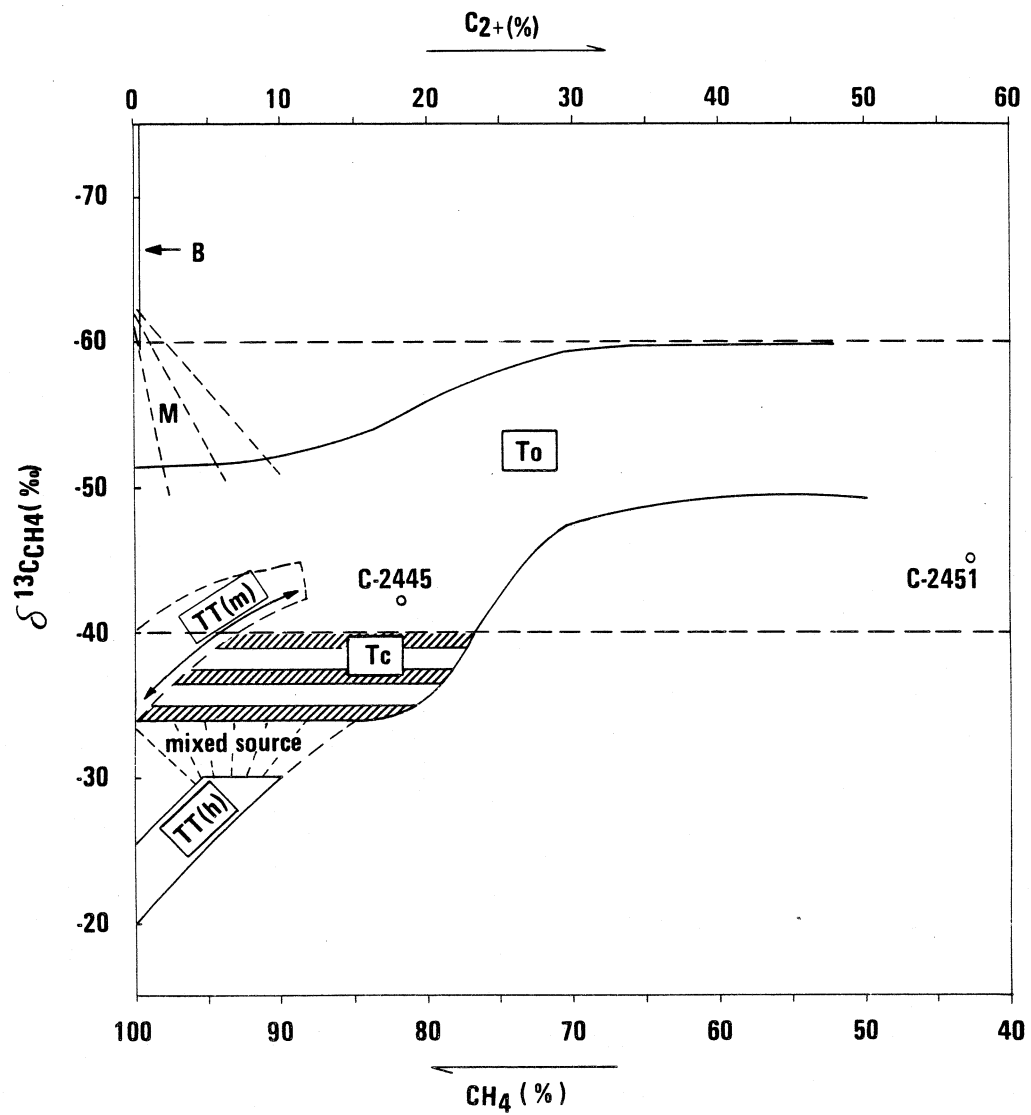


Fig. 5.9 Plot of $\delta^{13}C_{CH_4}$ vs% C_{2+} components.

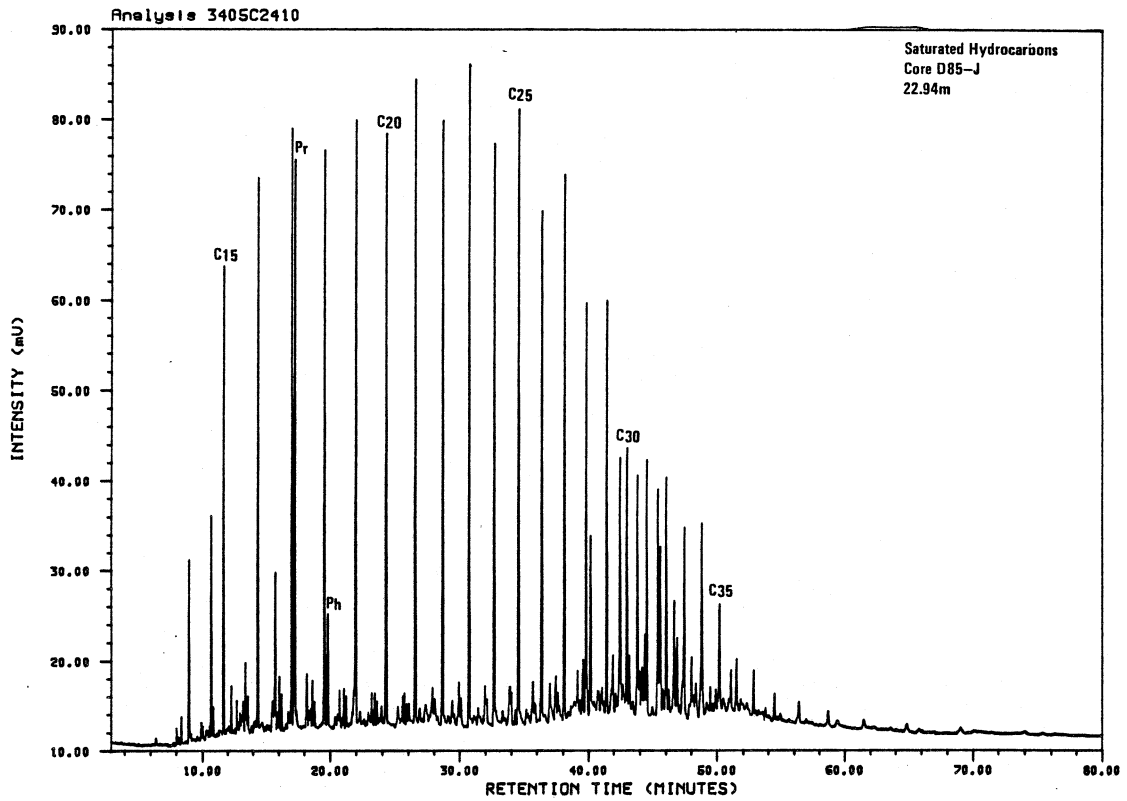


Fig.5.10a Gas chromatogram of saturated hydrocarbons, D85-J, 22.94 m.

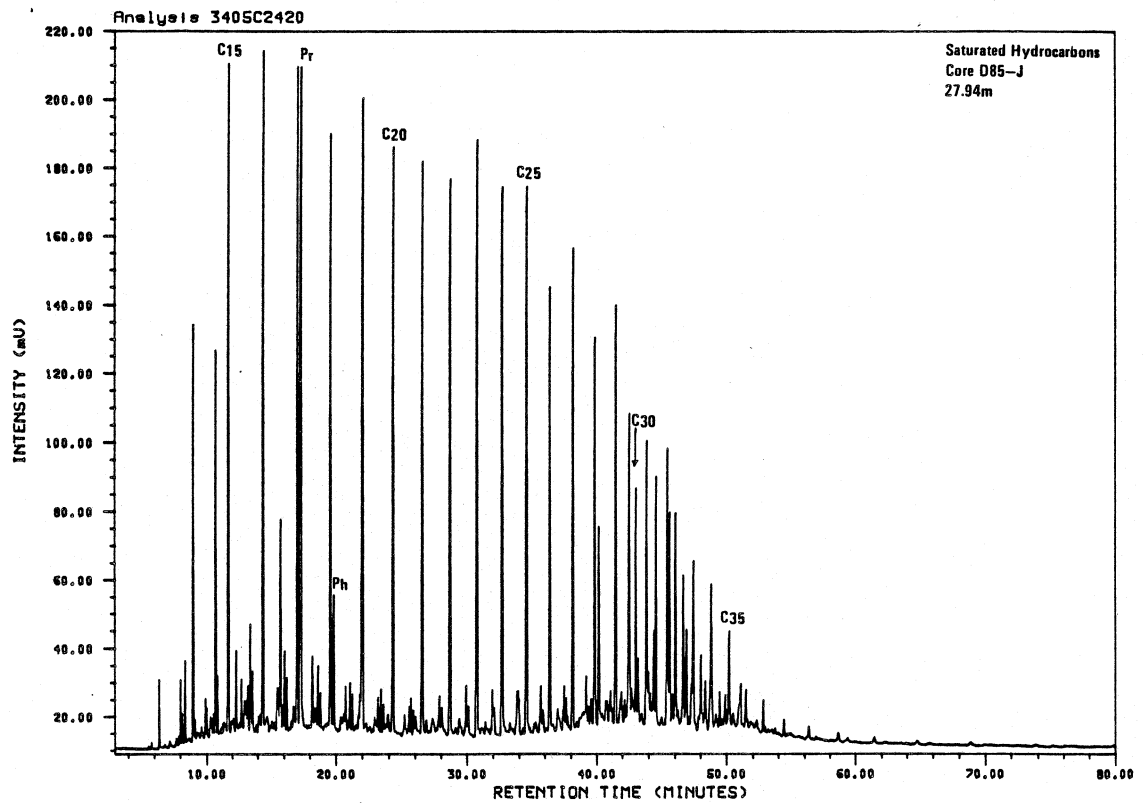


Fig.5.10b Gas chromatogram of saturated hydrocarbons, D85-J, 27.94 m.

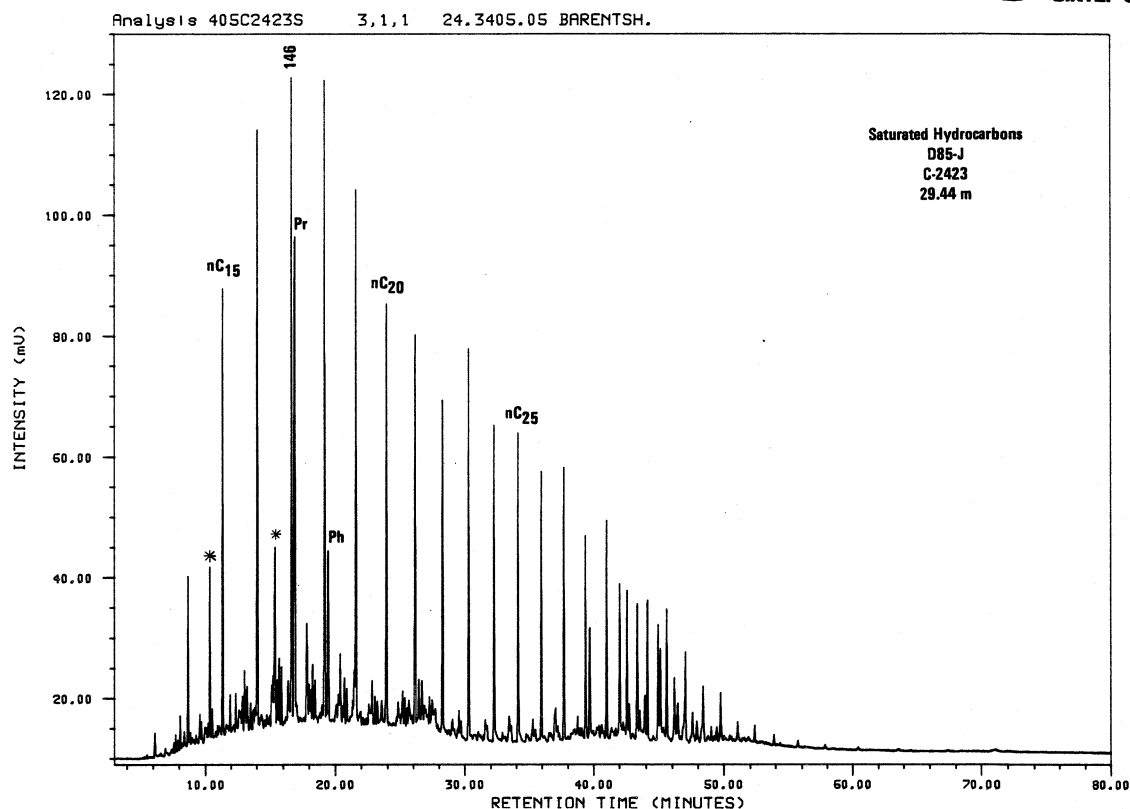


Fig.5.10c Gas chromatogram of saturated hydrocarbons from core D85-J, 29.44 m.

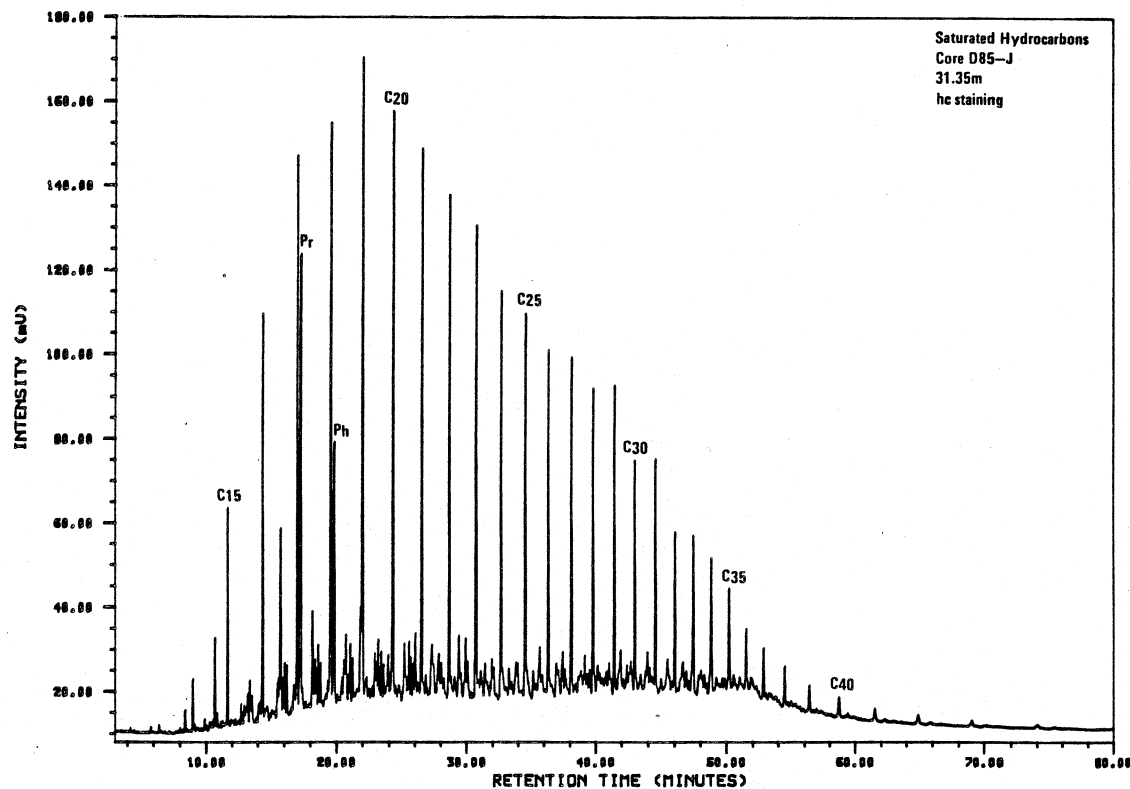


Fig.5.10d Gas chromatogram of saturated hydrocarbons from core D85-J, 31.35 m.

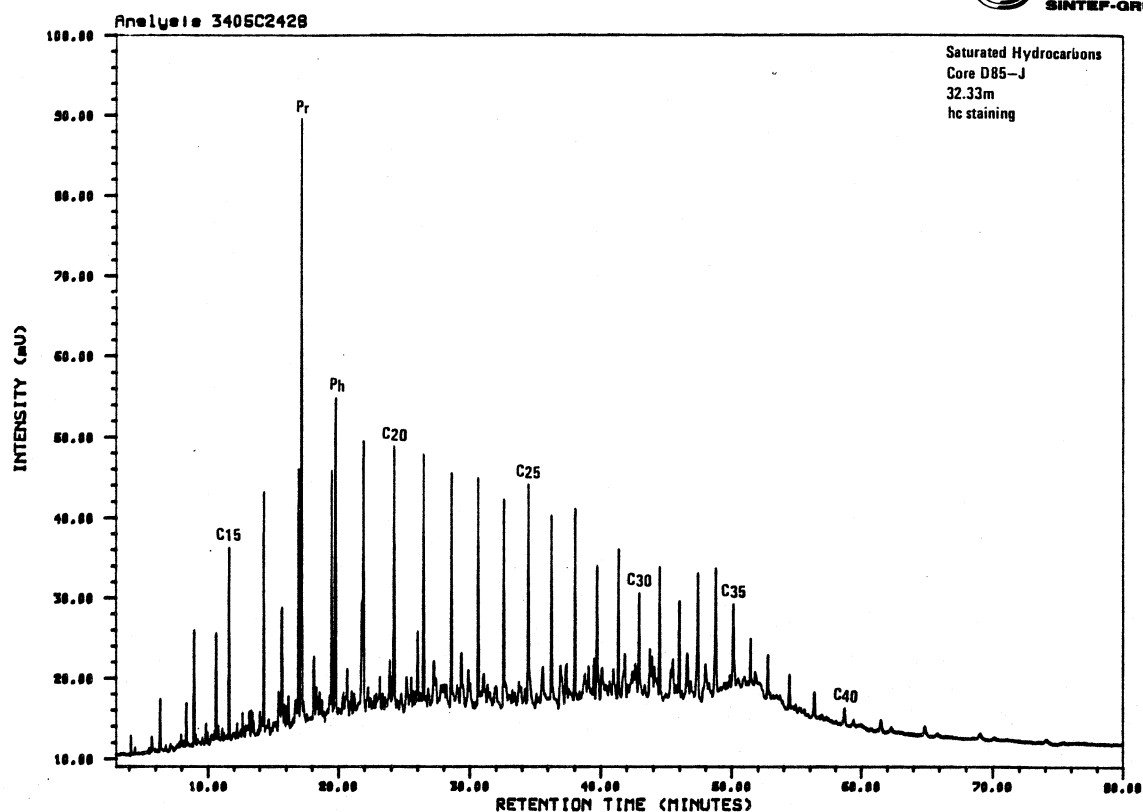


Fig.5.10e Gas chromatogram of saturated hydrocarbons from core D85-J,
32.33 m.

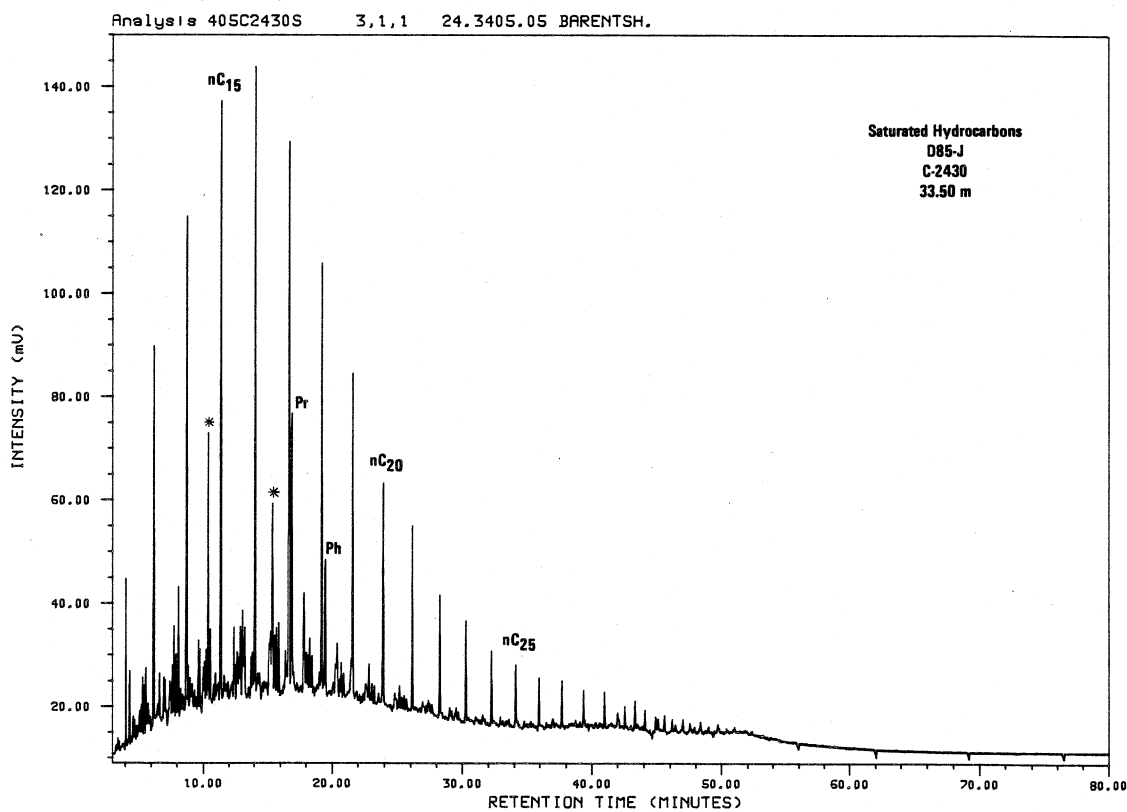


Fig.5.10f Gas chromatogram of saturated hydrocarbons from core D85-J,
33.50 m.

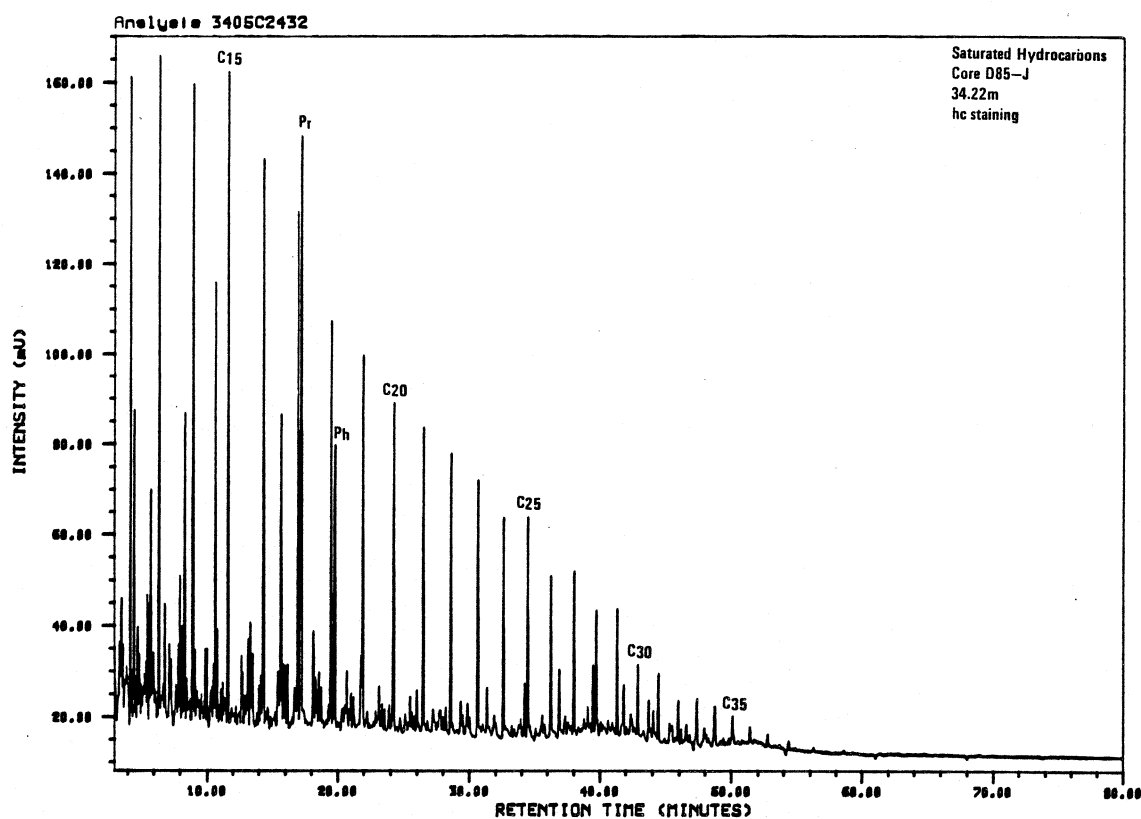


Fig. 5.10 g Gas chromatogram of saturated hydrocarbons from core D85-J, 34.22 m.

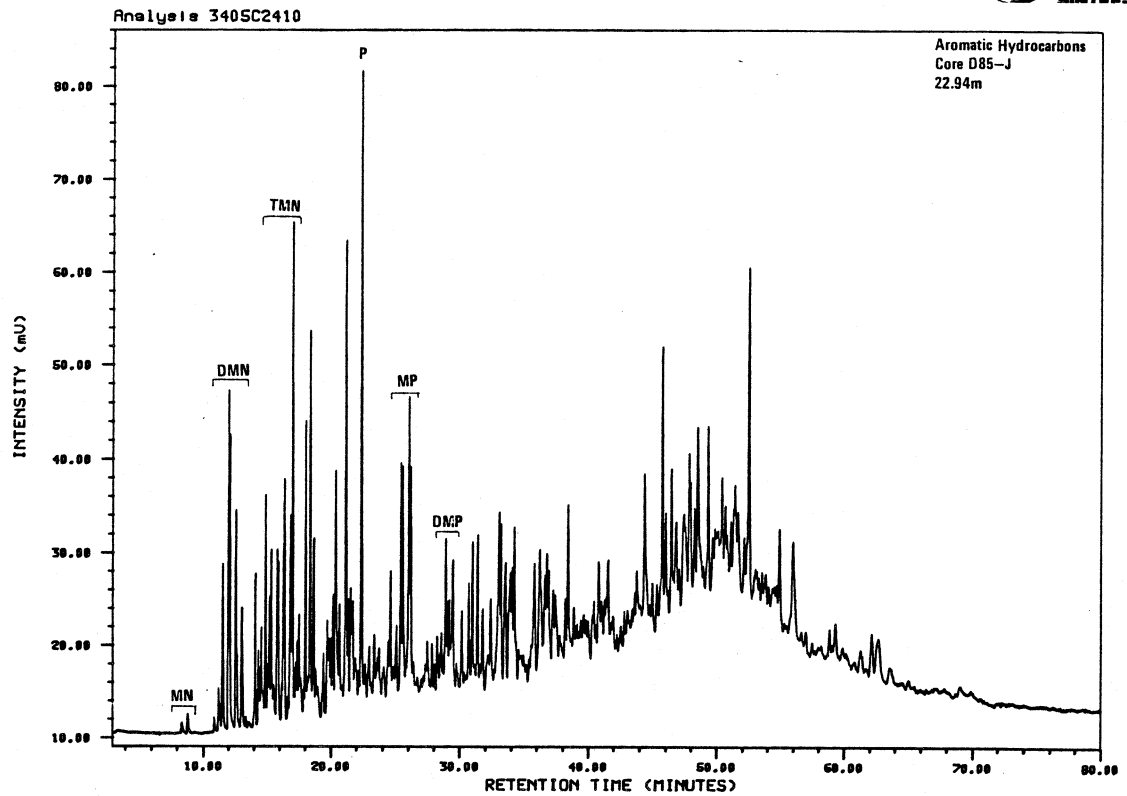


Fig. 5.11 a Gas chromatogram of aromatic hydrocarbons, D85-J,
22.94 m.

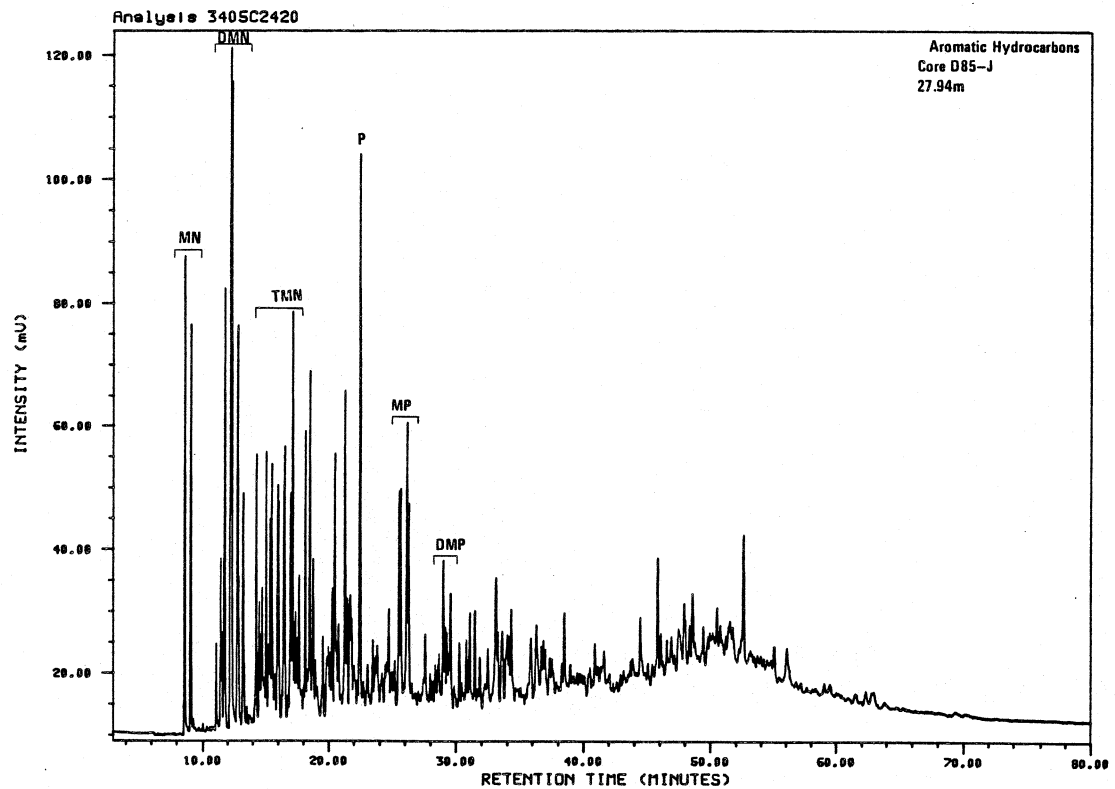


Fig. 5.11 b Gas chromatogram of aromatic hydrocarbons, D85-J,
27.94 m.

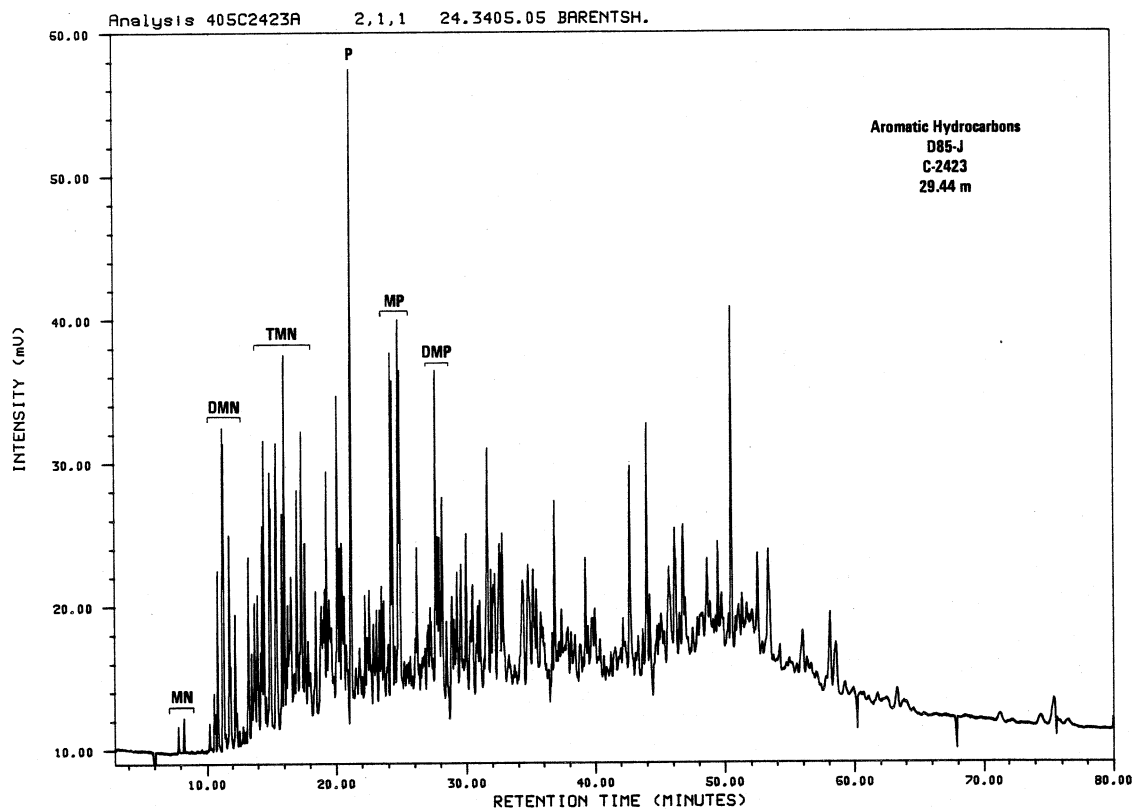


Fig. 5.11 c Gas chromatogram of aromatic hydrocarbons from core D85-J, 29.44 m.

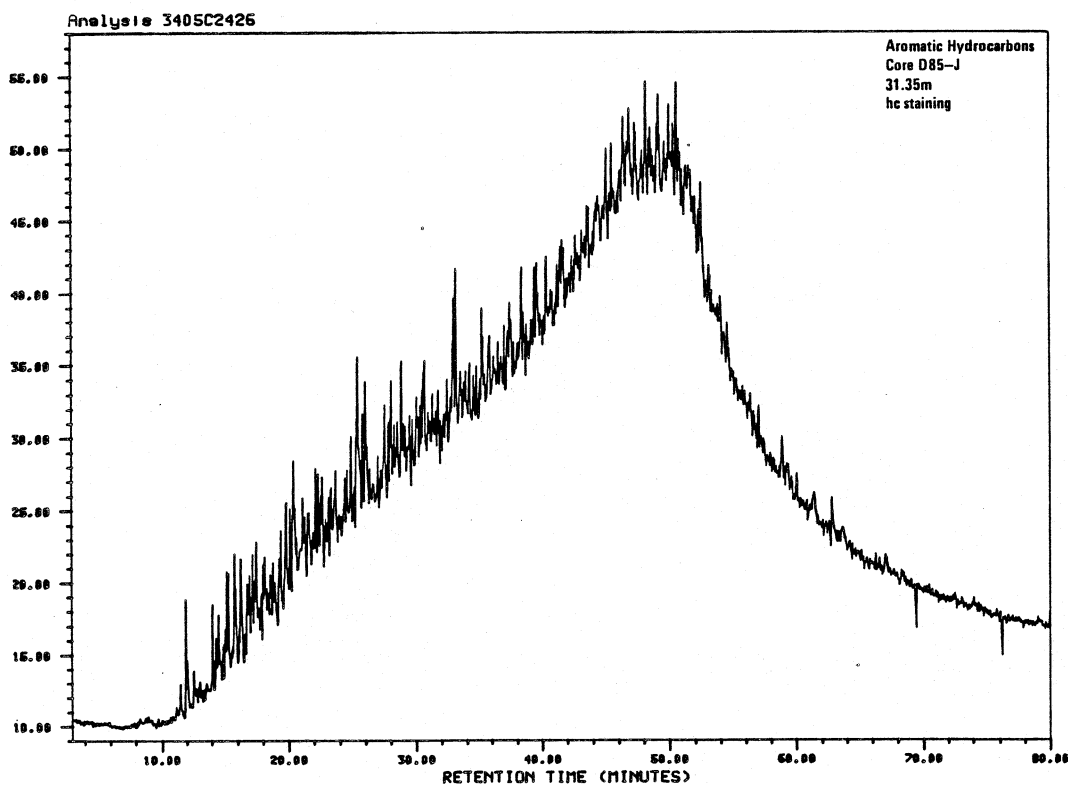


Fig. 5.11 d Gas chromatogram of aromatic hydrocarbons, D85-J, 31.35 m.

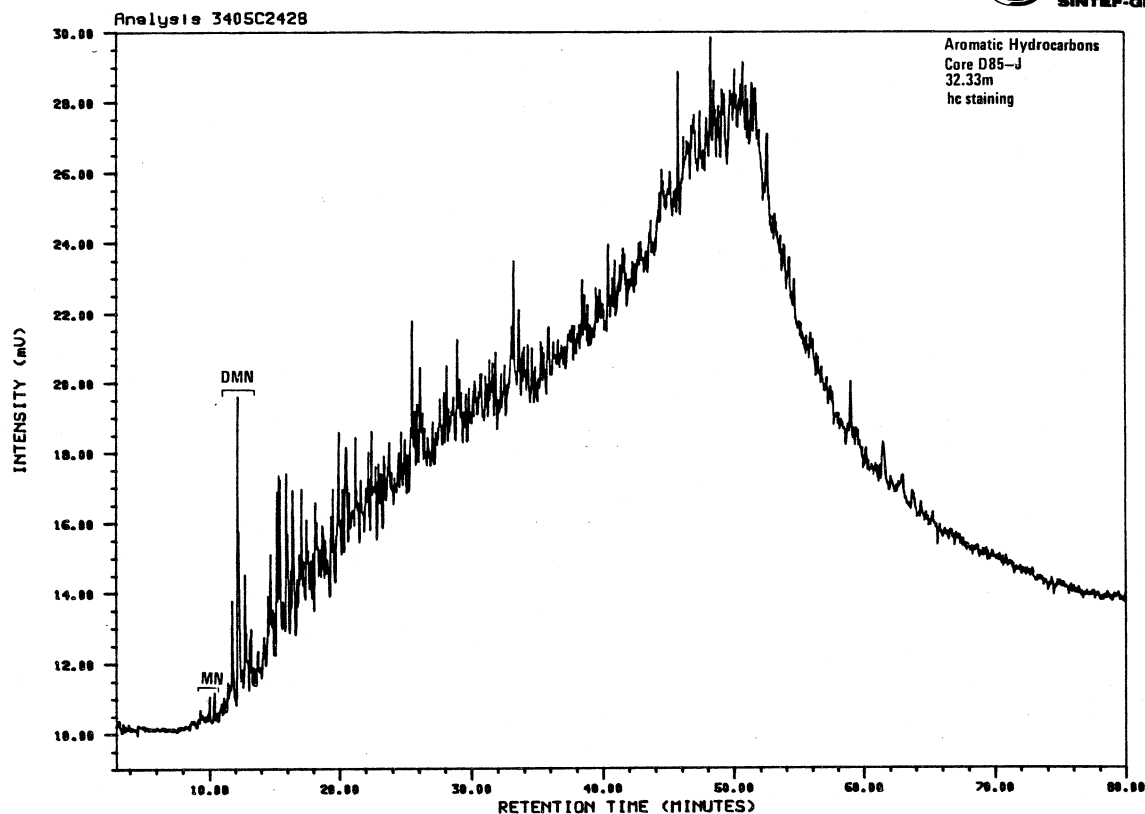


Fig. 5.11 e Gas chromatogram of aromatic hydrocarbons, D85-J, 32.33 m.

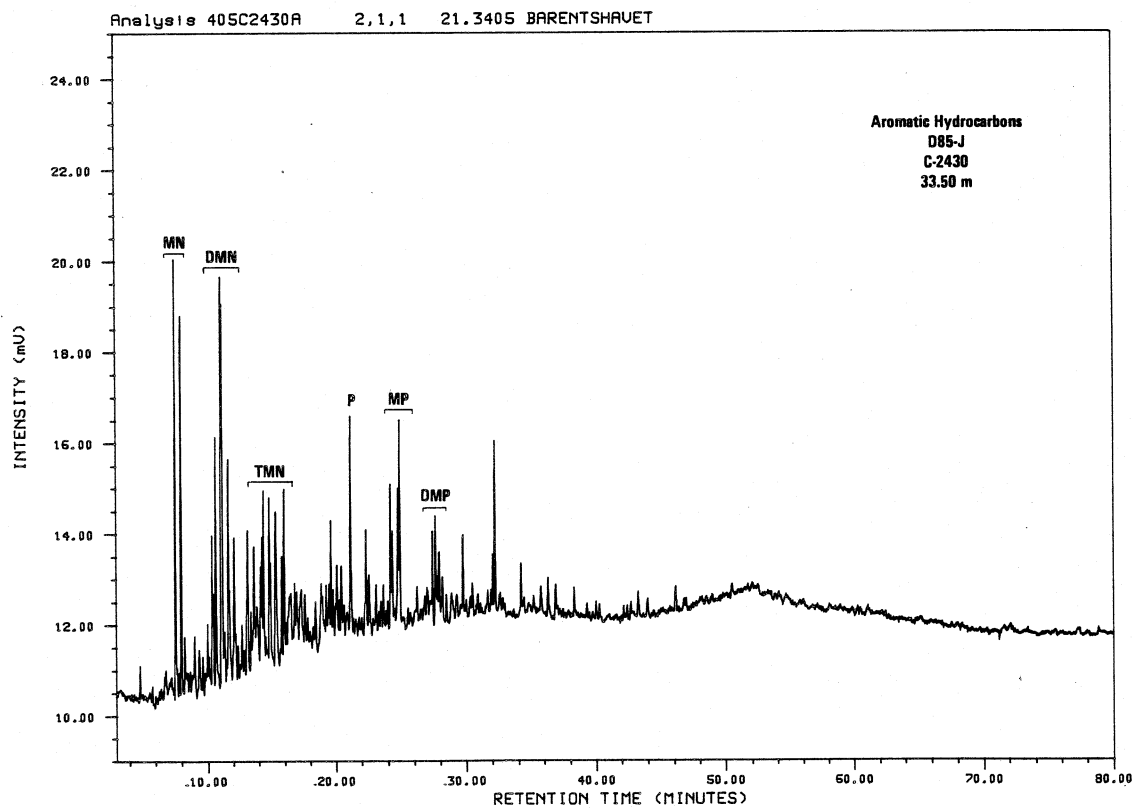


Fig. 5.11 f Gas chromatogram of aromatic hydrocarbons from core D85-J, 33.50 m.

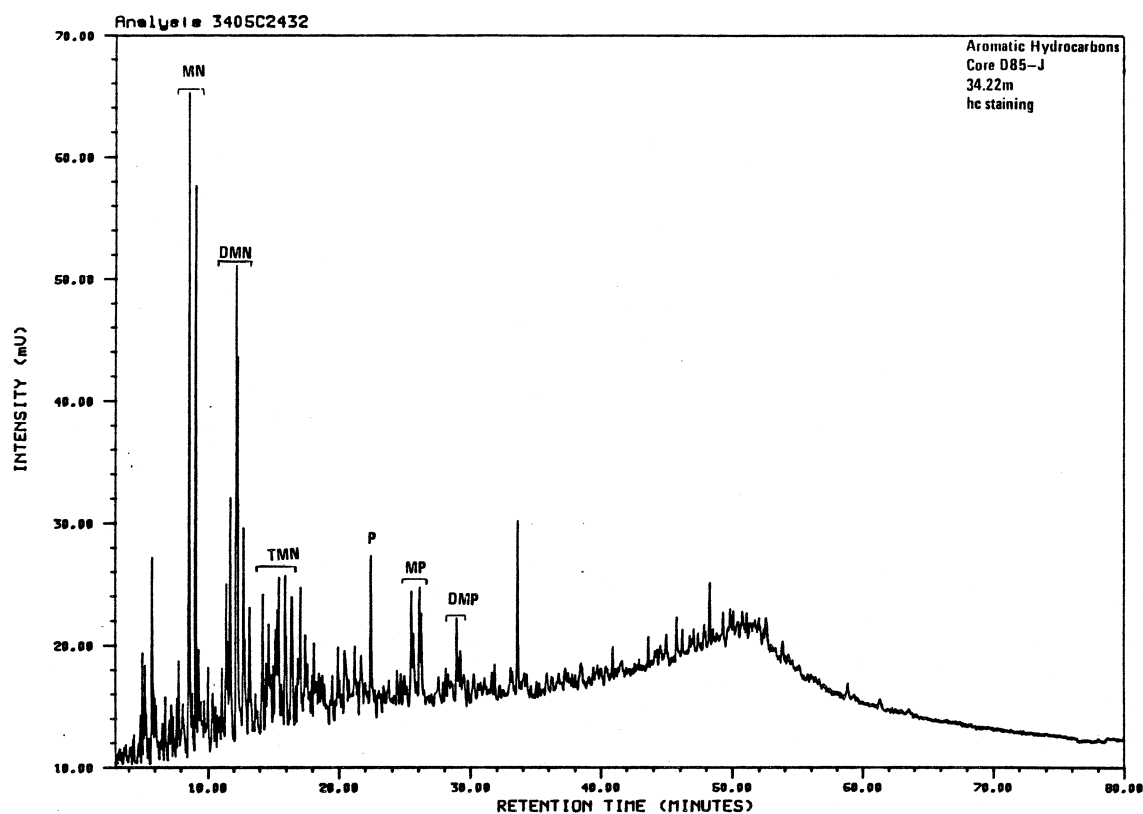


Fig. 5.11 g Gas chromatogram of aromatic hydrocarbons, D85-J,
34.22 m.

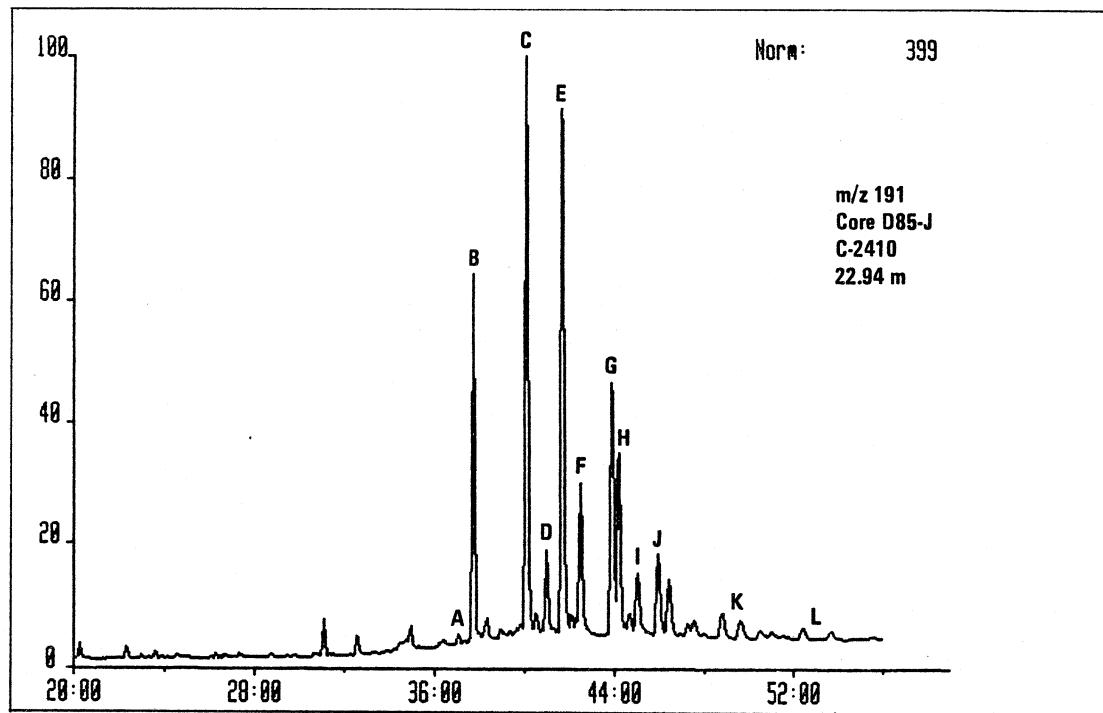


Fig. 5.12a (1) Mass chromatogram of terpanes (m/z 191), core D85-J, 22.94 m.

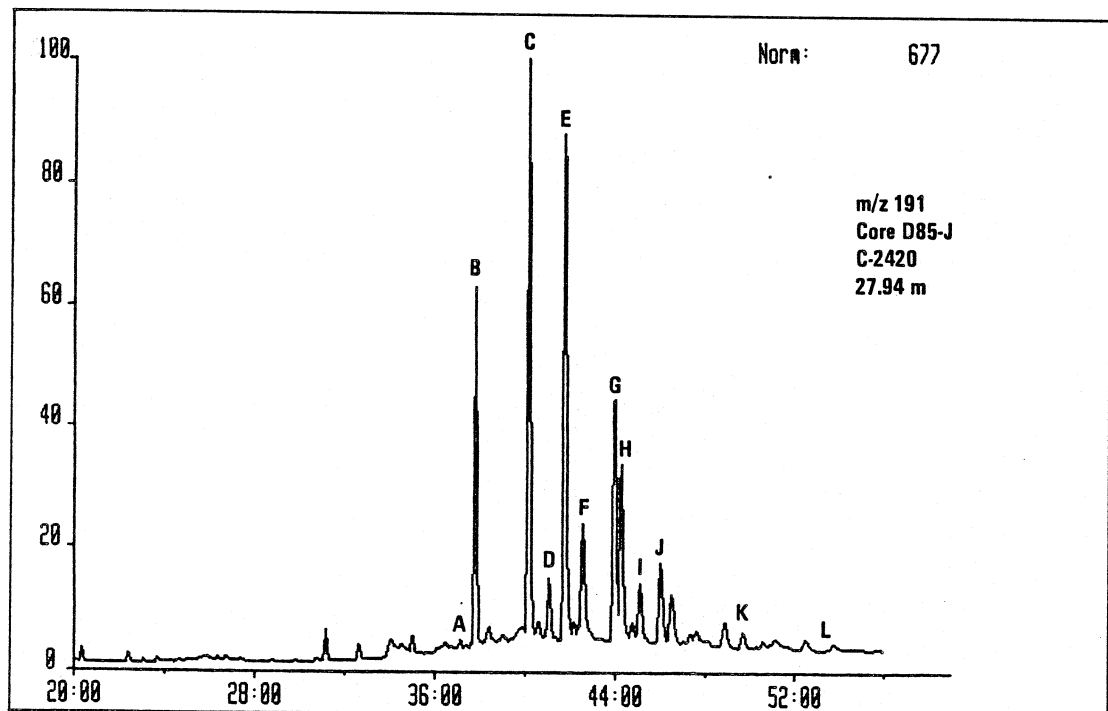


Fig. 5.12a (2) Mass chromatogram of terpanes (m/z 191), core D85-J, 27.94 m.

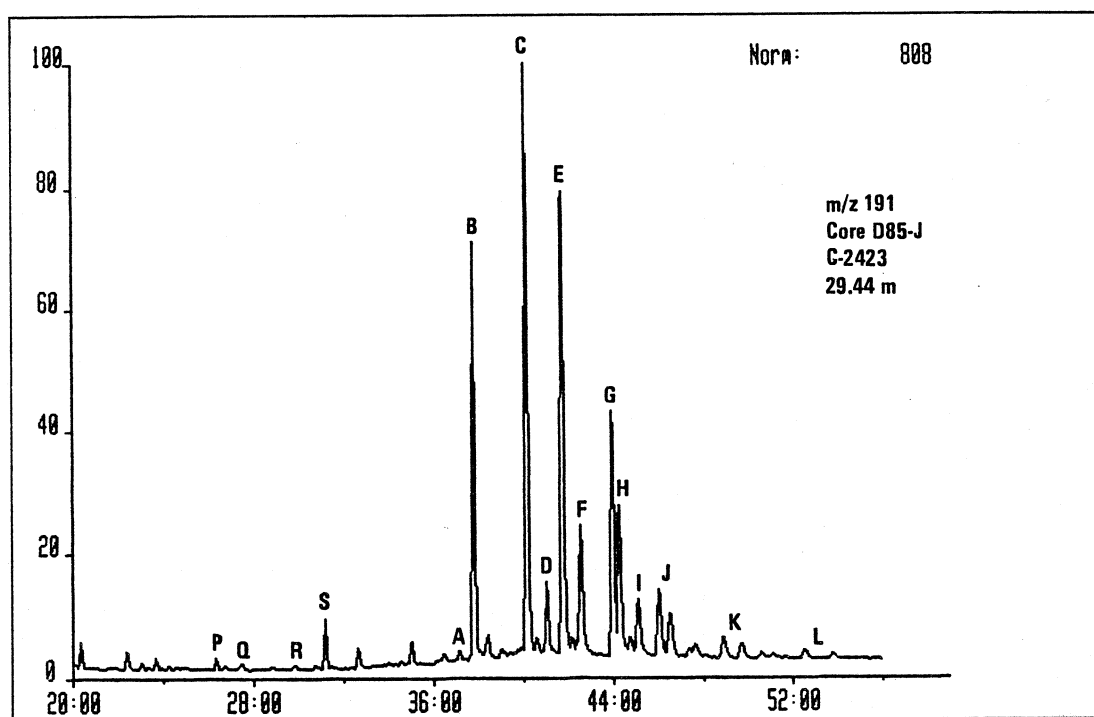


Fig. 5.12a (3) Mass chromatogram of terpanes (m/z 191), core D85-J, 29.44 m.

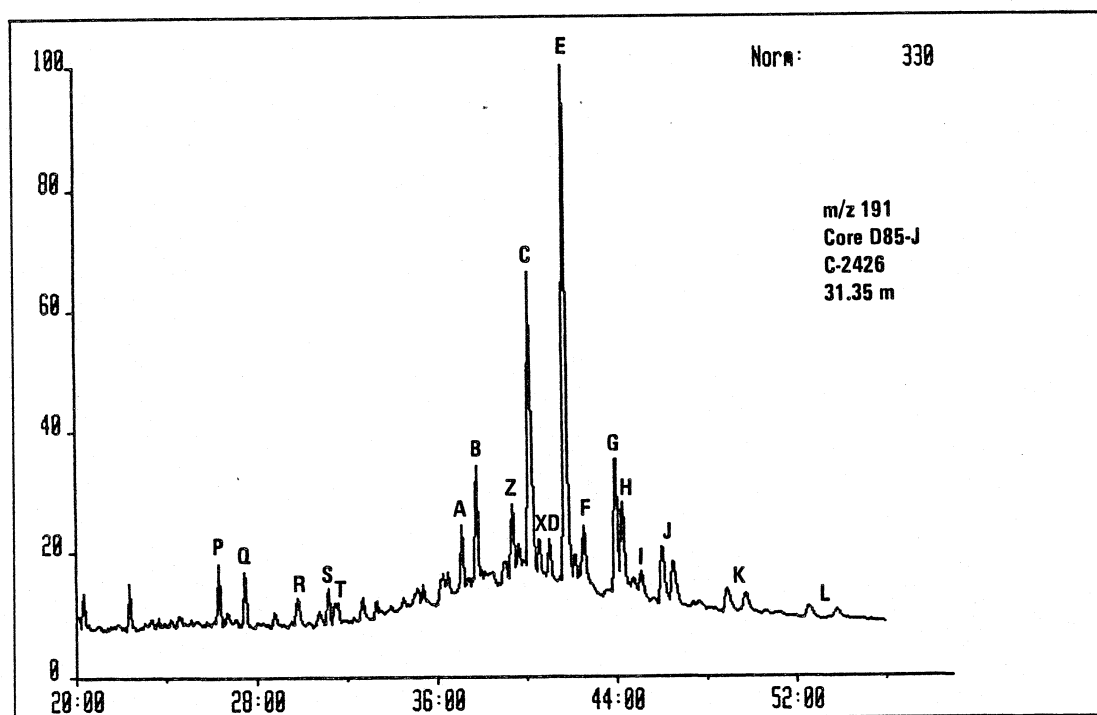


Fig. 5.12a (4) Mass chromatogram of terpanes (m/z 191), core D85-J, 31.35 m.

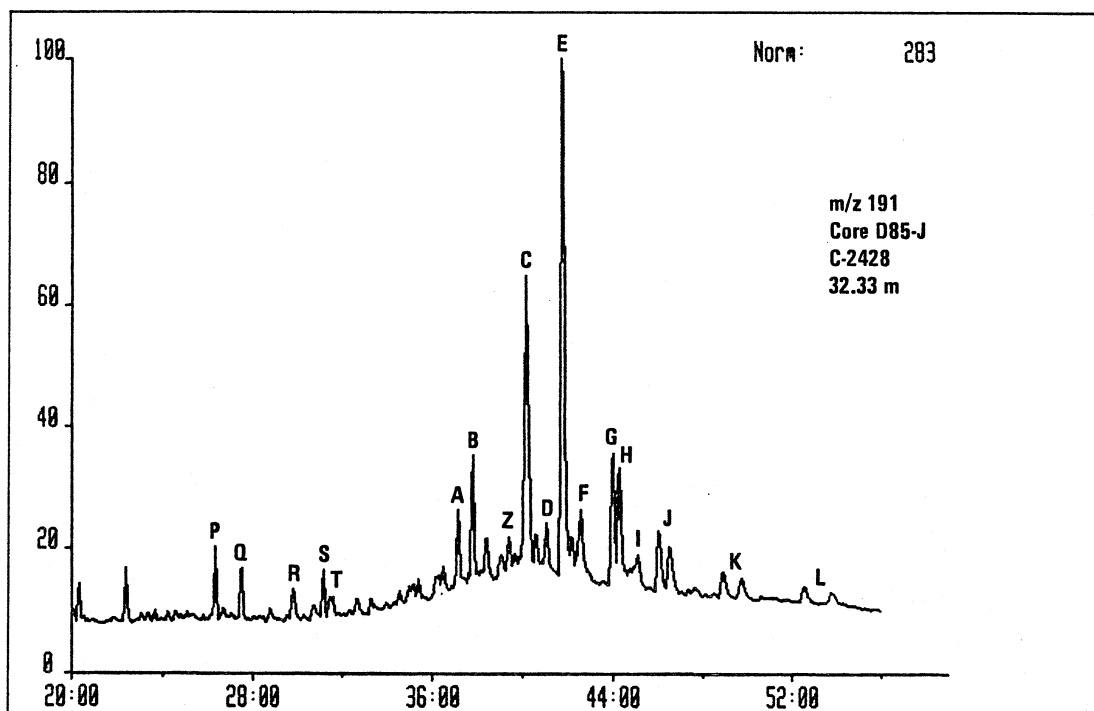


Fig. 5.12a (5) Mass chromatogram of terpanes (m/z 191), core D85-J, 32.33 m.

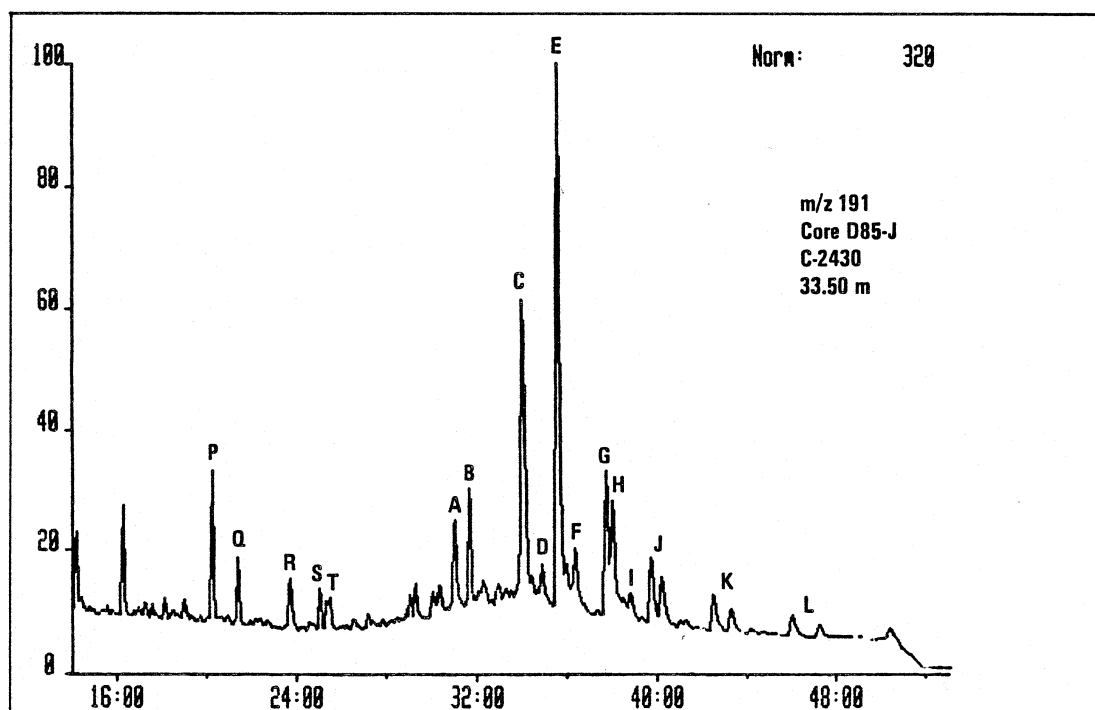


Fig. 5.12a (6) Mass chromatogram of terpanes (m/z 191), core D85-J, 33.50 m.

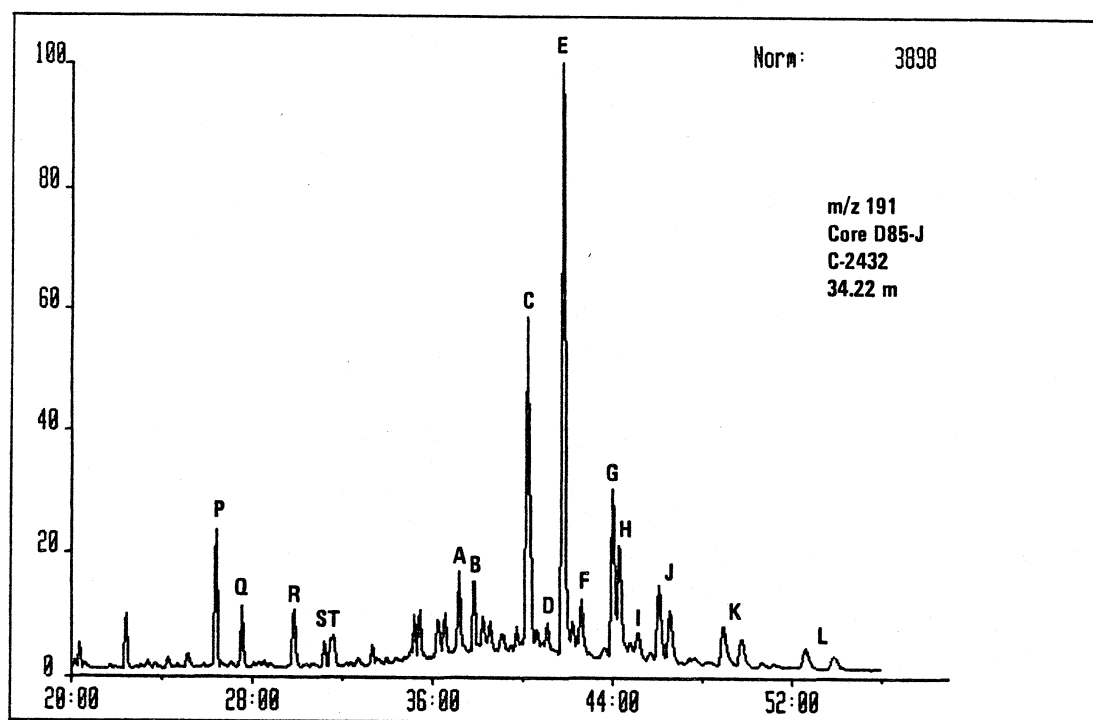


Fig. 5.12a (7) Mass chromatogram of terpanes (m/z 191), core D85-J, 34.22 m.

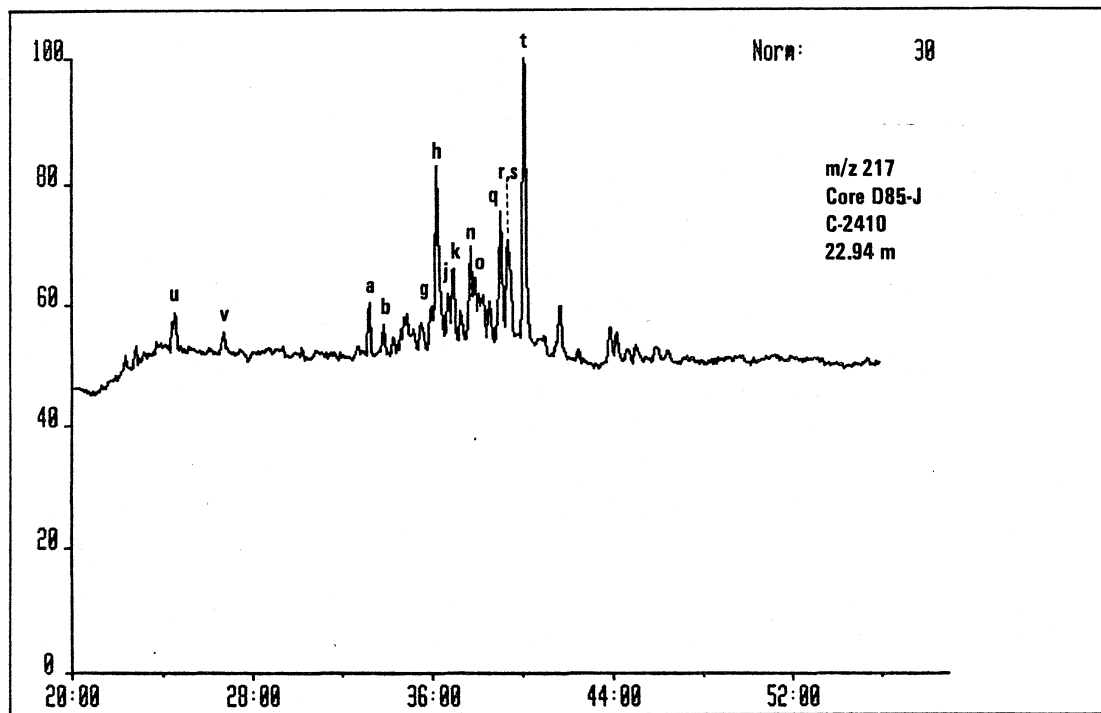


Fig. 5.12b (1) Mass chromatogram of steranes, core D85-J,
22.94 m.

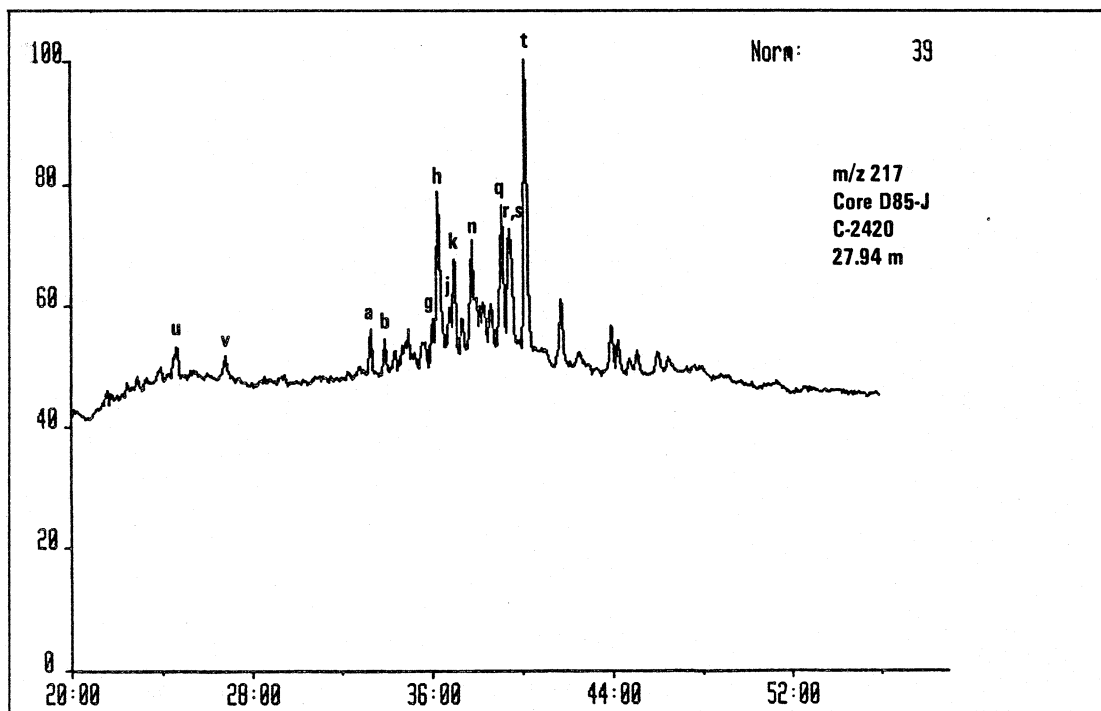


Fig. 5.12b (2) Mass chromatogram of steranes, core D85-J,
27.94 m.

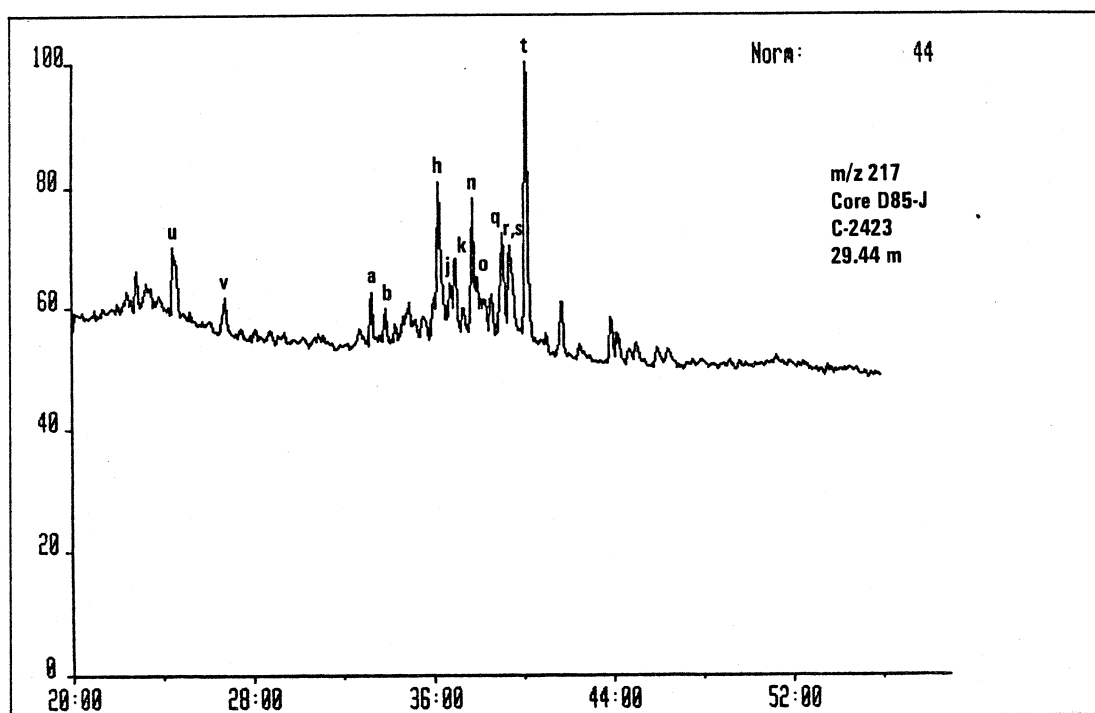


Fig. 5.12b (3) Mass chromatogram of steranes, core D85-J, 29.44 m.

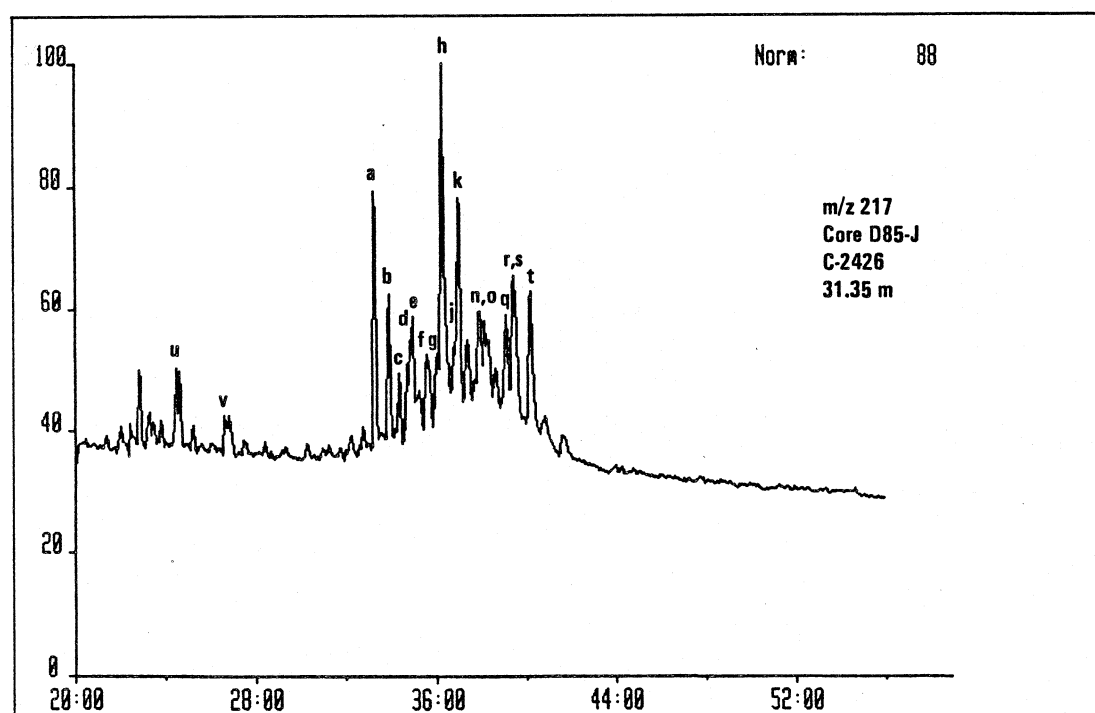


Fig. 5.12b (4) Mass chromatogram of steranes, core D85-J, 31.35 m.

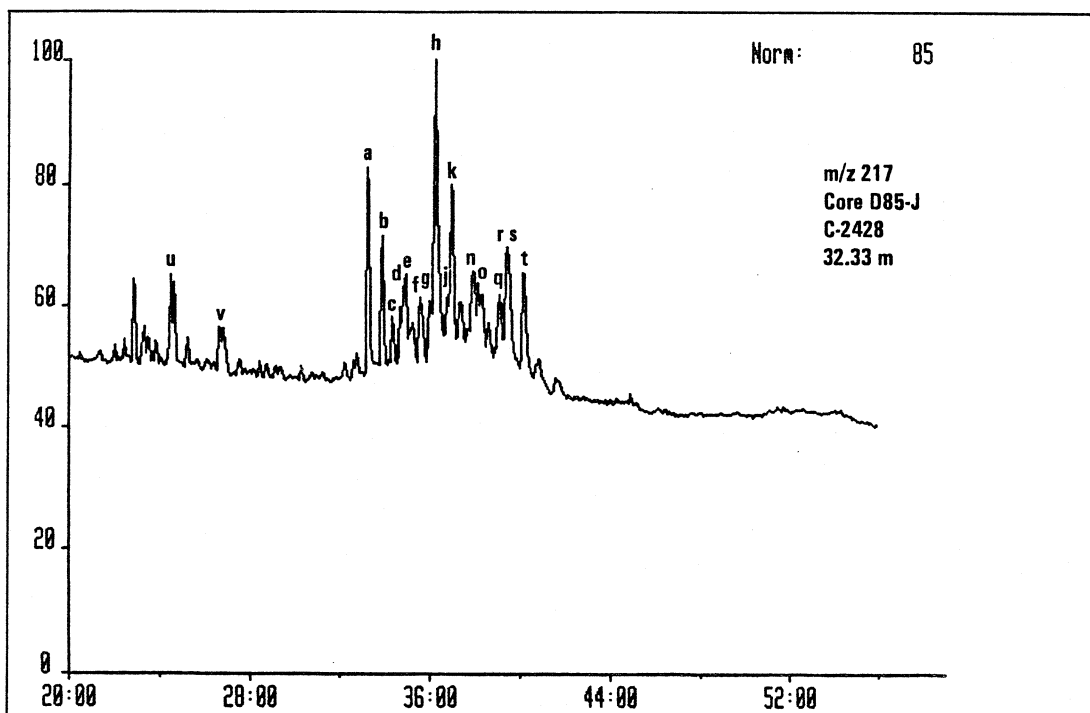


Fig. 5.12b (5) Mass chromatogram of steranes, core D85-J,
32.33 m.

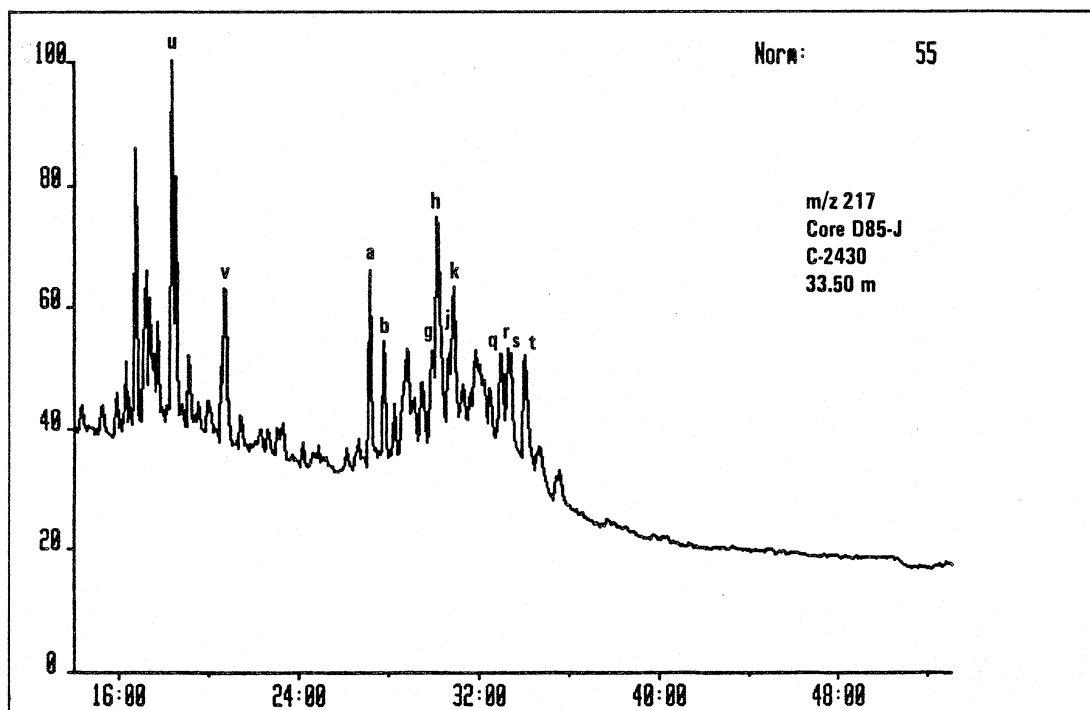


Fig. 5.12b (6) Mass chromatogram of steranes, core D85-J,
33.50 m

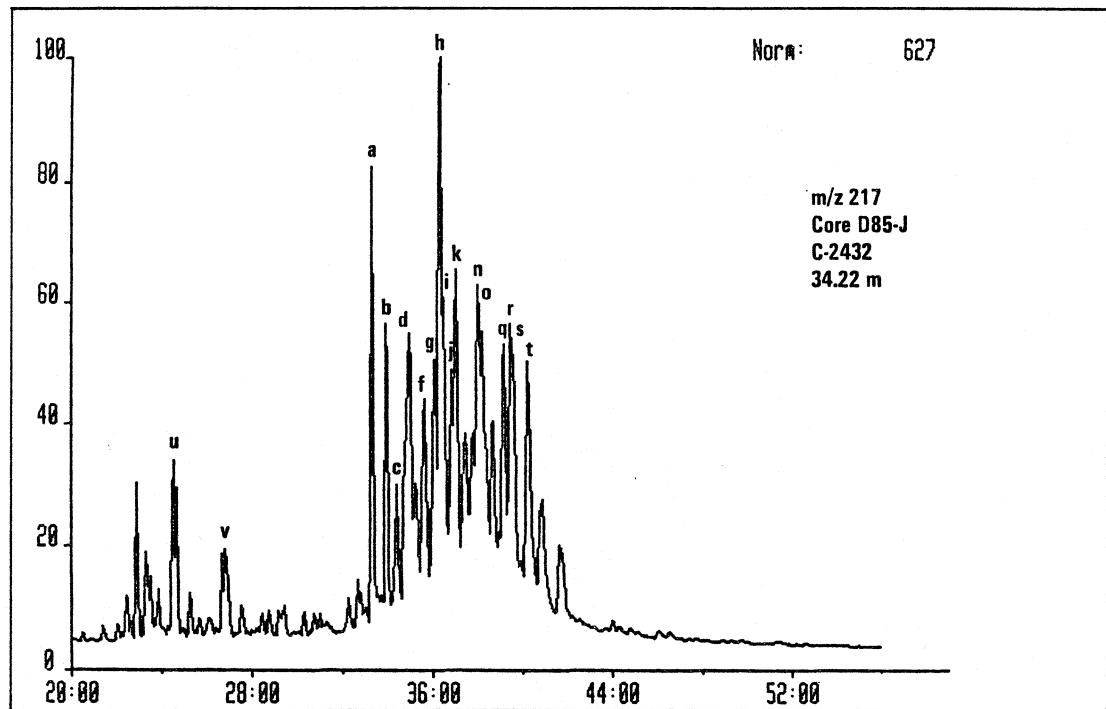


Fig. 5.12b (7) Mass chromatogram of steranes, core D85-J,
34.22 m.

Table 5.1 (m+p)-xylene/n-octene ratios calculated from pyrolysis-gas chromatograms.

IKU no.	Core	Depth (m)	(m+p)-xylene n-octene
C-2372	J	3.95	1.53
C-2396	J	15.47	1.29
C-2416	J	25.95	1.74
C-2432	J	34.22	0.65
C-2437	J	36.20	0.54
C-2825	K	97.51	1.33
C-2842	K	106.60	0.78
C-2858	K	114.75	0.70
C-2873	K	122.23	0.65
C-2876	K	123.70	0.72
C-2733	N	49.72	0.78
C-2737	N	51.44	0.95
C-2742	N	53.94	0.96
C-2884	O	56.38	1.72

Table 5.2 Concentration of EOM and chromatographic fractions.

IKU-No	DEPTH (m)	Rock Extr. (g)	EOM (mg)	Sat. (mg)	Aro. (mg)	HC (mg)	Non HC (mg)	TOC (%)
D85-J								
C 2410	22.94	9.2	9.1	1.0	2.6	3.6	5.5	2.20
C 2420	27.94	16.4	18.5	0.6	2.6	3.2	15.3	3.04
C 2423	29.44	8.8	8.5	0.8	1.1	1.9	6.6	2.22
C 2426	31.35	17.1	19.7	13.1	4.4	17.5	2.2	0.21
C 2428	32.33	18.7	41.7	10.4	4.3	14.8	26.9	0.36
C 2430	33.50	16.6	30.0	5.3	10.3	15.6	14.4	2.01
C 2432	34.22	12.2	32.0	9.5	8.9	18.4	13.6	2.43
D85-K								
C 2825	97.51	27.6	27.1	1.0	1.7	2.6	24.5	1.18
C 2842	106.60	4.3	1.8	1.0	0.6	1.6	0.2	1.14
C 2858	114.75	6.2	5.4	0.8	0.6	1.4	4.0	1.22
C 2872	121.78	4.9	5.6	1.1	0.6	1.7	3.9	1.38
C 2876	123.70	8.1	5.0	2.6	1.4	4.1	0.9	2.23
D85-O								
C 2884	56.38	12.3	1.7	1.6	1.1	2.6	-0.9	0.65

Table 5.3 Weight of EOM and chromatographic fractions (weight ppm of rock).

I	I	I	I	I	I	I	I	I	I
I	IKU-No	DEPTH	EOM	Sat.	Aro.	HC	Non	I	I
I		(m)					HC	I	I
I								I	I
I	D85-J							I	I
I	C 2410	22.94	991	105	288	392	599	I	I
I	C 2420	27.94	1125	36	161	197	928	I	I
I	C 2423	29.44	971	96	123	219	752	I	I
I	C 2426	31.35	1151	764	259	1023	127	I	I
I	C 2428	32.33	2230	558	231	789	1441	I	I
I	C 2430	33.50	1807	318	622	940	867	I	I
I	C 2432	34.22	2623	777	728	1505	1118	I	I
I	D85-K							I	I
I	C 2825	97.51	983	35	61	96	887	I	I
I	C 2842	106.60	419	223	140	363	56	I	I
I	C 2858	114.75	870	135	97	232	638	I	I
I	C 2872	121.78	1131	218	121	339	792	I	I
I	C 2876	123.70	619	327	178	505	114	I	I
I	D85-O							I	I
I	C 2884	56.38	139	127	88	215	-77	I	I

Table 5.4 Concentration of EOM and chromatographic fractions (mg/g TOC).

I	I	I	I	I	I	I	I	I	I
I	IKU-No	DEPTH	EOM	Sat.	Aro.	HC	Non	I	I
I		(m)					HC	I	I
I								I	I
I	D85-J							I	I
I	C 2410	22.94	45.1	4.8	13.1	17.8	27.2	I	I
I	C 2420	27.94	37.0	1.2	5.3	6.5	30.5	I	I
I	C 2423	29.44	43.8	4.3	5.6	9.9	33.9	I	I
I	C 2426	31.35	548.0	363.8	123.5	487.3	60.6	I	I
I	C 2428	32.33	619.4	155.1	64.2	219.3	400.2	I	I
I	C 2430	33.50	89.9	15.8	30.9	46.8	43.2	I	I
I	C 2432	34.22	107.9	32.0	30.0	61.9	46.0	I	I
I	D85-K							I	I
I	C 2825	97.51	83.3	2.9	5.2	8.1	75.2	I	I
I	C 2842	106.60	36.7	19.6	12.2	31.8	4.9	I	I
I	C 2858	114.75	71.3	11.1	7.9	19.0	52.3	I	I
I	C 2872	121.78	82.0	15.8	8.8	24.6	57.4	I	I
I	C 2876	123.70	27.7	14.7	8.0	22.6	5.1	I	I
I	D85-O							I	I
I	C 2884	56.38	21.3	19.6	13.6	33.1	-11.8	I	I

Table 5.5 Composition of material extracted from the rock.

I	IKU-No	DEPTH	Sat	Aro	HC	SAT	Non HC	HC	I
I	:	:	---	---	---	---	---	---	I
I	:	:	EOM	EOM	EOM	Aro	EOM	Non HC	I
I	:	(m)	%	%	%	x 100	%	x 100	I
I	D85-J	:	:	:	:	:	:	:	I
I	C 2410	22.94	10.5	29.0	39.6	36.4	60.4	65.5	I
I	:	:	:	:	:	:	:	:	I
I	C 2420	27.94	3.2	14.3	17.5	22.7	82.5	21.2	I
I	:	:	:	:	:	:	:	:	I
I	C 2423	29.44	9.9	12.7	22.6	77.8	77.4	29.2	I
I	:	:	:	:	:	:	:	:	I
I	C 2426	31.35	66.4	22.5	88.9	294.6	11.1	803.7	I
I	:	:	:	:	:	:	:	:	I
I	C 2428	32.33	25.0	10.4	35.4	241.7	64.6	54.8	I
I	:	:	:	:	:	:	:	:	I
I	C 2430	33.50	17.6	34.4	52.0	51.2	48.0	108.3	I
I	:	:	:	:	:	:	:	:	I
I	C 2432	34.22	29.6	27.8	57.4	106.8	42.6	134.6	I
I	D85-K	:	:	:	:	:	:	:	I
I	C 2825	97.51	3.5	6.2	9.7	57.1	90.3	10.8	I
I	:	:	:	:	:	:	:	:	I
I	C 2842	106.60	53.3	33.3	86.7	160.0	13.3	650.0	I
I	:	:	:	:	:	:	:	:	I
I	C 2858	114.75	15.6	11.1	26.7	140.0	73.3	36.4	I
I	:	:	:	:	:	:	:	:	I
I	C 2872	121.78	19.3	10.7	30.0	180.0	70.0	42.9	I
I	:	:	:	:	:	:	:	:	I
I	C 2876	123.70	52.8	28.8	81.6	183.3	18.4	443.5	I
I	D85-O	:	:	:	:	:	:	:	I
I	C 2884	56.38	91.8	63.5	155.3	144.4	-55.3		I
I	:	:	:	:	:	:	:	:	I

Table 5.6 Tabulation of data from the gas chromatograms.

I	DEPTH	:	PRISTANE	PRISTANE			I
I	IKU No.	:	-----	-----	CPI-1	CPI-2	I
I	(m)	:	n-C17	PHYTANE			I
I	=====						
I	D85-J	:					I
I	C 2410	22.94	0.9	5.0	1.2	1.2	I
I		:					I
I	C 2420	27.94	1.0	4.9	1.2	1.2	I
I		:					I
I	C 2423	29.44	0.8	2.9	1.2	1.1	I
I		:					I
I	C 2426	31.35	0.9	1.7	1.0	1.1	I
I		:					I
I	C 2428	32.33	2.4	1.9	1.2	1.2	I
I		:					I
I	C 2430	33.50	0.5	2.0	1.1	1.1	I
I		:					I
I	C 2432	34.22	1.1	2.1	1.2	1.2	I
I	D85-K	:					I
I	C 2825	97.51	1.5	1.2	2.2	2.0	I
I		:					I
I	C 2842	106.60	0.7	1.1	1.8	1.7	I
I		:					I
I	C 2858	114.75	0.7	1.1	2.4	1.9	I
I		:					I
I	C 2872	121.78	0.6	1.2	1.8	1.7	I
I		:					I
I	C 2876	123.70	2.1	1.0	1.9	1.9	I
I	D85-O	:					I
I	C 2884	56.38	1.1	1.4	1.5	1.5	I
I		:					I

Table 5.7 Methyl-phenanthrene index (MPI) ratios calculated from aromatic hydrocarbon gas chromatograms.

IKU no.	CORE	Depth (m)	MPI-1	MPI-2
C-2410	J	22.94	0.58	0.57
C-2420	J	27.94	0.60	0.61
C-2423	J	29.44	0.73	0.70
C-2426	J	31.35	not calculable	- -
C-2428	J	32.33	not calculable	- -
C-2430	J	33.50	0.41	0.48
C-2432	J	34.22	0.73	0.54
C-2825	K	97.51	not calculable	- -
C-2842	K	106.60	0.64	0.64
C-2858	K	114.75	0.65	0.58
C-2872	K	121.78	0.73	0.74
C-2876	K	123.70	not calculable	- -
C-2884	O	56.38	0.54	0.51
$\text{MPI-1} = \frac{1.5(2\text{-MP}+3\text{-MP})}{\text{P}+1\text{-MP}+9\text{-MP}} \qquad \text{MPI-2} = \frac{3(2\text{-MP})}{\text{P}+1\text{-MP}+9\text{-MP}}$				

Table 5.8 Molecular ratios from sterane and terpane mass chromatograms.
Maturity ratios.
Triassic sediments in core D85-K and D85-0.

IKU code	Depth (m)	$\alpha\beta/\alpha\beta+\beta\alpha$ ¹⁾	%22S ²⁾	% $\beta\beta$ ³⁾	%20S ⁴⁾
Core D85-K					
C-2825	97.51	0.82	20.0	31.7	4.7
C-2842	106.60	0.73	16.7	24.8	2.9
C-2858	114.75	0.84	20.7	31.0	8.6
C-2872	121.78	0.79	21.4	25.6	31.0 (?)
C-2876	123.70	0.78	12.8	20.3	3.2
Core D85-0					
C-2884	56.38	0.97	-	-	-

1) E/E+F in m/z 191

* only from C₃₁ homologs

2) Average % distribution between first and second eluting isomers of extended hopanes (G-M in m/z 191)

3) q/q+t in m/z 217

4) $2(r+s)/(q+t+2(r+s))$ in m/z 217

Table 5.9 Molecular ratios from terpane and sterane mass chromatograms.
Maturity and source characteristic ratios.
Triassic sediments in Core D85-K and D85-0.

IKU no.	Depth (m)	Q/E ¹⁾	T _m /T _s ²⁾	X/E ³⁾	Z/E ⁴⁾	a/a+j ⁵⁾
Core D85-K						
C-2825	97.51	-	2.17	-	-	0.13
C-2842	106.60	-	3.07	-	-	0.10
C-2858	114.75	-	8.33	-	-	-
C-2872	121.78	-	-	-	-	0.08
C-2876	123.70	-	3.33	-	-	-
Core D85-0						
C-2884	56.38	-	~20	-	-	-

- 1) Relative abundance of tricyclic terpanes (Q/E in m/z 191)
- 2) B/A in m/z 191
- 3) Relative abundance of unknown (X/E in m/z 191)
- 4) Relative abundance of bisnorhopane (Z/E in m/z 191)
- 5) Relative abundance of C₂₇ rearranged steranes (a/a+j in m/z 217)

Table 5.10 $\delta^{13}\text{C}$ carbon isotope data from rock extracts.

IKU no.	Core	Depth (m)	SAT (o/oo)	AROM (o/oo)
C-2420	J	27.94	-	-25.6
C-2423	J	29.44	-	-26.2
C-2426	J	31.35	-29.1	-27.8
C-2428	J	32.33	-29.1	-28.4
C-2430	J	33.50	-31.1	-30.4
C-2432	J	34.22	-32.2	-31.0
C-2876	K	123.70	-30.6	-

Table 5.11 $\delta^{13}\text{C}$ carbon isotope data from gas samples.

IKU no.	Core	Depth (m)	C ₁ (o/oo)	C ₂ (o/oo)	C ₃ (o/oo)	iC ₄ (o/oo)	nC ₄ (o/oo)
C-2445	J	31.15	-42.1	-33.0	-31.3	-33.0	-
C-2450	V	14.64	-	insufficient sample -			-
C-2451	V	20.20	-45.4	- insufficient sample			-

Table 5.12. Molecular ratios from sterane and terpane mass chromatograms.
 Maturity ratios.
 Non-indigenous hydrocarbons - Core J.

IKU code	Depth (m)	$\alpha\beta/\alpha\beta+\beta\alpha$ ¹⁾	%22S ²⁾	%8B ³⁾	%20S ⁴⁾
C-2410	22.94	0.79	59.0	47.2	29.9
C-2420	27.94	0.82	59.2	50.4	32.4
C-2423	29.44	0.79	62.0	45.6	25.8
C-2426	31.35	0.91	59.7	71.5	37.8
C-2428	32.33	0.89	55.1	71.9	36.0
C-2430	33.50	0.93	58.7	69.1	32.0
C-2432	34.22	0.92	59.8	68.5	47.1

1) E/E+F in m/z 191

* only from C₃₁ homologs

2) Average % distribution between first and second eluting isomers of extended hopanes (G-M in m/z 191)

3) q/q+t in m/z 217

4) $2(r+s)/(q+t+2(r+s))$ in m/z 217

Table 5.13. Molecular ratios from terpane and sterane mass chromatograms.

Maturity and source characteristic ratios.

Non-indigenous hydrocarbons - Core J.

IKU no.	Depth (m)	Q/E ¹⁾	T _m /T _s ²⁾	X/E ³⁾	Z/E ⁴⁾	a/a+j ⁵⁾
C-2410	22.94	<0.01	31.0	0.03	<0.01	0.56
C-2420	27.94	<0.01	30.0	0.03	<0.01	0.64
C-2423	29.44	0.01	35.0	0.04	<0.01	0.60
C-2426	31.35	0.10	1.9	0.06	0.13(?)	0.83
C-2428	32.33	0.10	1.7	0.06	0.06(?)	0.83
C-2430	33.50	0.18	1.2	0.03	-	0.80
C-2432	34.22	0.10	0.8	0.02	0.02(?)	0.73

- 1) Relative abundance of tricyclic terpanes (Q/E in m/z 191)
- 2) B/A in m/z 191
- 3) Relative abundance of unknown (X/E in m/z 191)
- 4) Relative abundance of bisnorhopane (Z/E in m/z 191)
- 5) Relative abundance of C₂₇ rearranged steranes (a/a+j in m/z 217)

CHAPTER 6

HEAT FLOW MEASUREMENTS

by

Joar Sættem, Martin Hamborg and Øistein Johansen

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6.1 Summary

Chapter 6 summarizes all work carried out as a continuation of the study reported in IKU report 21.3405.00/02/85.

Because of the surprisingly low heat flow values that were calculated quite some effort was put into investigations to look for any errors in the heat flow calculation. The first to be checked was the thermal conductivity, and from theoretical calculations based on input of mineral composition, water content and gas voids (in some cores) it was found that the originally used values for thermal conductivity should have been multiplied by a factor of about 1.81. Further analyses of the needle probes used for the field measurements showed that the correct factor is 1.799, which therefore should be applied to all values given in the first report.

Other effects which were investigated were possible changes in thermal conductivity due to pressure and temperature changes from in situ conditions to the laboratory, gas influence, anisotropy and possible influences on the temperature gradient due to Quaternary glaciations.

The thermal conductivity was measured in the laboratory on board and it is concluded that neither the temperature nor the pressure change itself influenced significantly. A secondary but significant effect of pressure release has occurred at sites D85-N, D85-O and D85-V, because samples from the depths in question from these sites were found to contain gas dissolved in the pore water. This resulted in gas voids being formed in the laboratory which influenced the thermal conductivity measurements. The necessary correction for this is found to be an increase in the average measured thermal conductivity from these sites by 16%, 8% and 4% respectively.

28 representative samples have been subjected to measurement of both vertical and horizontal thermal conductivity and these values were found to differ in average by 13%. As the on board measurements were

carried out slightly obliquely to the horizontal direction, we suggest a reduction of 10% in these values for the calculation of vertical heat flow.

Correction of the measured temperatures and temperature gradients due to transient effect from late Quaternary climatic changes have been quantified for two alternative glacial history models. This effect is found to be different for the different sites. Maximum effect is found to occur on site C85-A where the disturbance on the present day gradient is between

$$-3.6 \cdot 10^{-3} \text{ }^{\circ}\text{C m}^{-1} \text{ and } -6.2 \cdot 10^{-3} \text{ }^{\circ}\text{C m}^{-1}$$

depending on temperature history models.

From the results of the different topics studied we have arrived at the following suggested final heat flow values:

Site	Field results		Suggested adjusted results	
	(HFU)	(mW/m ²)	(HFU)	(mW/m ²)
C85-A	1.78	74.5	1.77	74.1
C85-E	1.30	54.5	1.32	55.2
D85-K	1.33	55.7	1.29	54.0
D85-N	1.57	65.7	1.72	72.0
D85-O	1.54	64.5	1.52	63.6
D85-V	1.62	76.8	1.61	67.4

The results are adjusted for anisotropy of thermal conductivity, gas voids in some cores and late Quaternary climatic effects.

6.2 Summary of field results

As a part of the IKU Barents Sea Mapping Program, shallow drilling and heat flow measurements were carried out on 7, respectively 9 sites in the southwestern Barents Sea during July-August 1985.

The basic field data results are presented in IKU report 21.3405.00/02.85 (Sættem et al. 1985). A brief summary of this report and the field results are given below. Data from the six sites where accurate temperature gradients were obtained are given in Table 6.2.1.

The heat flow data has been obtained by measuring temperature and thermal conductivity in Quaternary sediments down to about 70 m below the sea bed. The survey was carried out from the Norwegian drillship M/S "Bucentaur". Downhole temperature measurements were performed by use of a high precision temperature probe developed by Fugro/Omegalink. The probe was mounted on a penetrometer device inside the API 5" drill string and pushed into undisturbed Quaternary sediments.

At levels corresponding to each temperature measurement, Quaternary sediment samples were collected and subjected to thermal conductivity measurement by use of a needle probe. Based on this information mean heat flow values have been calculated at six of the investigated sites.

In addition to these measurements a 3 m long spear with thermistors mounted in fixed positions was used in order to obtain temperatures close to the sea bed. However, this data was not used to calculate heat flow.

Both the needle probes and the thermistors as well as the respective amplifiers have been subjected to thorough pre- and post-calibration. Final system resolution and accuracy for the temperature measurements are estimated to be within 0.008 and 0.02 deg C respectively.

The values for thermal conductivity of the Quaternary sediments as presented by Sættem et al. (1985) have all to be corrected by a constant factor of 1.799. This is due to an input error in a computer program. A consequence of this is that also the reported heat flow values must be corrected correspondingly. This is discussed further in Chapter 6.3 and Appendix 1.

Table 6.2.1 Final measured gradients for the 1985 IKU sites. R-SQUARE is the standard coefficient of determination for regression (least square fit for all temperature-depth points). Values near 1.0 indicate good fit; values near zero indicate poor fit. N = number of temperature-depth points. 1) Top bedrock was not penetrated.

SITE	GEOGRAPHICAL COORDINATES	WATER DEPTH (m)	QUATERNARY THICKNESS (m)	GRAD (mdegC/m)	N	R-SQUARE
C85-A	71°38'49.795" 26°52'43.295"	369.0	37.80	42.8	3	0.9996
C85-E	72°24'19.739" 22°47'39.787"	323.5	app. 102.00 1)	33.1	3	0.9925
D85-K	73°16'51.539" 23°01'25.834"	419.5	97.05	34.4	16	0.9976
D85-N	74°29'25.296" 25°46'19.261"	333.5	48.89	38.2	3	0.9998
D85-O	75°12'20.536" 24°43'21.607"	107.5	55.70	31.0	3	0.9996
D85-V	73°49'10.924" 17°22'04.526"	295.5	45.69	36.8	5	0.9953

6.3 Thermal conductivity

6.3.1 Summary

The relatively low thermal conductivity values reported by Sættem et al. (1985) are found to be affected by a constant error (too low by a factor of 1.799) due to a computer programming error (Zielinski 1986 a). Due to this all the thermal conductivity values on Quaternary samples from Sættem et al. (1985) are multiplied by this factor.

Analyses of physical properties of the Quaternary cores are summarized in Chapter 6.3.2. In addition to standard soil mechanical testing mineral composition analyses (XRD) carried out at a representative selection of samples.

Results from the above tests are used as input for theoretical calculation of thermal conductivity based on known thermal conductivity of water and the various mineral compounds. Linear regression through a plot of these theoretical values versus the corresponding values from Sættem et al. (1985) gives a line with slope 1.81, which is practically equal to the factor 1.799 reported by Zielinski (1986 a).

The need for correction due to differences in temperature and pressure from in situ to laboratory conditions has also been investigated. The temperature is found to be insignificant because of the low water content of the samples.

For the samples from the sites D85-N, D85-0 and D85-V it has been found that due to the pressure release from in situ to laboratory conditions, gas dissolved in the pore water was released and formed gas voids. These gas voids have resulted in reduced thermal conductivity being measured in the laboratory. The necessary average correction expressed as a correction factor to the measured values for these three sites were 1.16, 1.08 and 1.04 respectively.

Results from comparison tests of thermal conductivity performed at independent institutions is presented in Chapter 6.3.4. These results fit very well with both the correction factor reported by Zielinski (1986 a) and the theoretically calculated thermal conductivity values.

Differences in vertical and horizontal thermal conductivity have also been investigated (Table 6.3.8) and the mean factor of anisotropy in thermal conductivity is calculated to 1.11. As the on board measurements mainly are believed to represent the horizontal thermal conductivity, we suggest that these values are reduced by 10%.

6.3.2 Analyses of physical properties

Several physical properties are analysed on selected Quaternary samples. A purpose of these analyses is to enable correlations of these properties with thermal conductivity. In the literature such correlations are given both on a theoretical basis (conductivity of multi compound media) as well as on an empirical basis (a.o. Bullard & Day, 1961; Lachenbruch & Marshall, 1961; Johansen 1975 and Matsuda & von Herzen 1985). Thus such correlations of the Barents Sea data both provide quality control and increase the confidence of the final thermal conductivity values.

Results from these analyses are given in Table 6.3.1 - Table 6.3.4b. Some of the results below are previously reported by Fjordingstad et al. (1985) (IKU report 21.3405.00/01/85, Chapter 4). However, the tables below include corrected thermal conductivity values and other corrections and additions.

The analysis techniques follow standard procedures. For a brief description reference is made to the Appendix A.3 in IKU report 21.3405.00/01/85.

Table 6.3.1 summarizes the following parameters:

- thermal conductivity, K , ($\text{mcal } ^\circ\text{C}^{-1} \text{ cm}^{-1} \text{ s}^{-1}$)
- water content, w , (% of dry weight)
- bulk density, γ , (kN/m^3)
- undrained shear strength, S_u , (kN/m^2)
- sound velocity, V_s , (m/s)

Table 6.3.1 Results from laboratory analyses of the Quaternary core material. This table include some additions and corrections relative to the version in IKU report 21.3405.00/01/85.

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985								
SITE	DEPTH (m)	T.CON. mcal/ °Ccms	WATER %	DENS. kN/m ³	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM. TRIAX.
C85-A	0.03	2.95						
	0.10				8.9			
	0.20				9.0			
	0.30				7.4			
	0.36	2.72						
	0.40				7.4			
	0.50				11.7			
	0.60				8.1			
	0.70		67.2	16.3	9.8			
	0.80				5.5			
	0.87	2.57						
	0.90				11.7			
	0.94	2.56						
	1.00				11.7			
	1.10				9.8			
	1.20	2.47			8.9			
	1.30				8.9			
	1.40		88.5	15.5	9.8			
	1.50				7.0			
	1.60	2.41			8.1			
	1.70				13.1			
	1.85	2.45						
	1.90				8.1			
	2.00				8.9			
	2.10				11.7			
	2.20	2.66			9.8			
	2.30				11.7			
	2.40		70.6	16.0	7.0			
	2.55	2.63						
	2.95	2.48						
	4.65	3.17						
	8.24		27.7	19.8	34.3	x	1650	x
	8.38	3.83	29.9	20.0				
	8.53		27.7	19.8	38.7		1670	
	14.02	4.21	22.0	20.8	>250.0	x	1710	x
	14.17	4.37	22.3	20.6				
	18.25				15.2			
	18.50	3.69						
	18.65	4.16	20.4	21.0	>250.0	x	1785	x
	23.15	4.23	20.6	21.3	>250.0		1760	x
	23.36	4.17	20.4	21.0	>250.0	x	1785	
	33.93	4.01						
	34.02		20.7	20.7	>250.0		1820	
	34.09	4.21						
	34.15	4.08						
	34.18		20.6	20.9	>250.0	x	1885	x
	34.32	4.52	24.6	21.7				
	37.40		22.2	19.6	>250.0	x	1880	
	37.65		26.1	20.0	109.9		1740	x

Table 6.3.1 cont.

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985									
SITE	DEPTH (m)	T.CON. mcal/ °Ccms	WATER %	DENS. kN/m ³	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM.	TRIAX
C85-E	0.04	3.62	36.9	19.3		x			
	0.10				16.6				
	0.20				9.0				
	0.30				6.5				
	0.35	3.83							
	0.40				11.7				
	0.50				18.6				
	0.60				16.2				
	0.70	3.08			11.7				
	0.80				18.6				
	0.90				11.7				
	1.00	3.69							
	1.03	3.63							
	1.10				20.0				
	1.20				15.2				
	1.30				18.6				
	1.40	3.02			11.7				
	1.50				15.2				
	1.60				15.2				
	1.70				18.6				
	1.80	3.00	40.8	18.2	11.7				
	1.90				11.7		1590		
	2.00				15.2				
	2.10	3.22			21.1	x			
	10.53	3.92			61.0				
	10.60	4.17							
	10.64	3.78							
	10.73		23.5	20.5					
	10.75		21.6	20.8	115.0		1800	x	
	21.28		4.12						
	21.34	4.25							
	21.40	3.90							
	21.47	3.94							
	21.55	3.06							
	21.75		21.5	20.9	180.0		1750		
	22.02		23.8	20.5			1830	x	x
	31.03	4.07	22.5					x	x
	31.08	4.25	20.8	20.7			1910		
	31.13	4.48							
	37.20		24.3	20.6	49.1	x	1680		
	40.13	3.94			>250.0				
	40.21	3.87	20.0	20.9	>250.0		1805		
	40.30	4.03	21.6	21.1	>250.0		1825	x	x
	40.37	3.71							
	50.05		22.5	20.3	85				
	50.10				85		1755		
	50.12	4.03			85				
	50.15	3.96			85				
	50.22	3.89			85				
	78.55		22.1	20.4					
	78.60		19.1	20.9	>250.0		1860	x	x
	88.98		18.2	21.4	>250.0	x	1815		
	89.21		17.7	21.5	>250.0		1875		
	89.49		18.1	21.2	>250.0		1905		
	89.66		17.6	21.2	>250.0	x	1905		
	97.10		18.6	21.1	>250.0	x	1930		
	97.40					x			
	98.25					x			
	98.50		19.9	20.9	>250.0	x	1810		
	99.25				>250.0	x			
	100.25		21.8	20.6	>250.0	x	1820		

Table 6.3.1 cont.

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985									
SITE	DEPTH (m)	T.CON. mcals/ °Ccms	WATER %	DENS. kN/m ³	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM.	TRIAx
D85-H	0.03	2.81							
	0.10				5.0	x			
	0.20				3.7				
	0.24	2.79							
	0.30				11.7				
	0.40				18.6				
	0.44	3.81							
	0.47		35.1	19.0					
	0.50	3.56			18.6		1675		
	0.60				18.6				
	0.70				26.0				
	0.72	3.00							
	0.80				30.4				
	0.90				30.4				
	0.97	2.90							
	1.00				16.6				
	1.10				34.3				
	1.20				26.0				
	1.22	3.17							
	1.30				27.0				
	1.40		34.0	19.5	27.0				
	1.44	3.42							
	1.50				24.0				
	11.25	4.68	15.8		>250.0	x	1790		
	13.56	4.62							
	13.65		13.8	21.8	>250.0	x	1990		
	13.78		12.9	22.1				x	

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985									
SITE	DEPTH (m)	T.CON. mcals/ °Ccms	WATER %	DENS. kN/m	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM.	TRIAx
D85-I	0.03	3.69							
	0.20				4.8				
	0.22	2.90							
	0.30				21.1				
	0.40				6.0				
	0.47	3.38							
	0.60				9.8				
	0.70				7.0				
	0.72	3.56							
	0.80				8.9				
	0.90				62.3				
	0.97	4.48							
	5.20				159.0				
	5.44	3.83							
	5.54	4.35	19.2	21.1	159.0	x	1970		

Table 6.3.1 cont.

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985									
SITE	DEPTH (m)	T.CON. mcal/ °Ccms	WATER %	DENS. kN/m ³	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM.	TRIAX
D85-K	0.03	2.72							
	0.10		60.2	16.8	8.1	x			
	0.20				6.5				
	0.30				7.0				
	0.33	2.90							
	0.40				7.0				
	0.50				15.2				
	0.60				16.6				
	0.70				15.2				
	0.71	2.93							
	0.80				15.2				
	0.90				8.9				
	1.00				7.0				
	1.09	3.15							
	1.10				10.8				
	1.17	3.02							
	1.20				10.8				
	1.30				15.2				
	1.40				18.6				
	1.50		39.0	18.6	16.6				
	1.56	3.00							
	1.60				18.6				
	1.70				21.1				
	1.80				24.0				
	1.86	3.04							
	1.90				18.6		1585		
	2.00				15.2				
	2.10				15.2				
	2.15		38.4	18.8		x			
	2.33	3.27							
	9.76		29.5	19.5	44.1	x	1620		
	9.93		29.4	19.5	49.1				
	9.97	3.47							
	10.00	3.56							
	10.08	3.71							
	10.11		29.7	19.6	32.4		1620		
	10.45		30.2	19.5	43.2		1675		
	10.46		31.5	19.5	43.2				
	10.60				52.5		1640	x	x
	10.71	3.58	30.3						
	14.61		28.9	19.6	49.1		1630		
	14.64	3.60							
	14.71	3.26							
	14.77	3.62							
	14.80		29.5	19.4	39.2		1640		
	14.83	3.56			39.2				
	14.89	3.26							
	14.92	3.62							
	14.95		31.0	19.2	32.4		1595		
	19.50		24.7	20.8	74.1			x	x
	19.60		23.4	20.5	87.8		1725		
	19.63	4.07							
	19.71	3.62	22.2	20.4			1700	x	x
	19.78	4.08							
	22.12				196.0				
	22.34		21.8	20.8	159.0		1725		

Table 6.3.1 cont.

D85-K	22.37	3.63							
	22.46	3.72			196.4				
	22.53	4.23							
	22.56		21.8	20.7	159.0	x	1725		
	25.27				159.0				
	25.52	3.94							
	25.59	3.81							
	25.63	4.03							
	25.66		22.2	20.7	196.4	x			
	29.94		22.3	20.9	115.8				
	30.19				102.1			x	x
	30.21	3.72							
	30.27	3.85							
	30.31	4.08	21.0				1765		
	34.45		21.0						
	34.49		20.3	21.0	>250.0			x	x
	34.51	4.03					1780		
	34.57	4.10							
	34.63	4.41			>250.0		1825		
	43.69				>250.0				
	43.85		20.3	21.0	>250.0		1795		
	43.98	4.12	20.1	20.8				x	x
	44.03	4.23							
	44.07	4.37							
	54.45		18.0	21.2	>250.0	x	1830		
	54.66	3.90					1830		
	54.72	3.72	21.2	20.9			1770	x	x
	54.79	4.01					1770		
	64.52				>250.0				
	64.67		17.9	21.3	>250.0	x		x	x
	64.74	3.89	18.6	21.4			1845		
	64.79	3.92							
	74.13		13.4	22.3	>250.0				
	84.27	3.99	16.3	21.0				x	x
	94.75		18.8	20.8	>250.0				
	94.87	3.89				x			
	94.92	3.53							
	97.20				150.0				
	97.36		21.8	20.4					

Table 6.3.1 cont.

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985									
SITE	DEPTH (m)	T.CON. mcal/ °Ccms	WATER %	DENS. kN/m ³	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM.	TRIAx
D85-N	0.03	3.04				x			
	0.10				11.7				
	0.20				7.4				
	0.23	3.40							
	0.30				21.1				
	0.40		27.7	19.9	21.1	x			
	0.43	3.42							
	0.50				21.1				
	0.60				38.7				
	0.62	3.76							
	9.94		14.0	21.6	>250.0		1900		
	12.00		16.4	20.6	>250.0	x	1855		
	12.08	3.78		21.7					
	12.16	4.21					1965	x	x
	21.81				245.0	x	1845		
	21.83	3.90							
	21.89	3.83							
	21.92		16.3	21.2	44.1		1170		
	27.24		15.7	21.7	>250.0		1800	x	
	27.26	3.99							
	27.31	3.51							
	27.37		16.7	20.5	>250.0	x			
	30.12		15.1	21.5	>250.0		1670	x	
	30.23	4.50							
	30.26	4.79							
	30.29		14.7	21.2	>250.0	x			
	35.10		15.5	20.2	>250.0	x	1535		
	35.16		14.5	21.5	>250.0			x	
	36.11	4.50							
	36.14	4.52							
	39.60		15.7	21.4	>250.0	x		x	
	39.66	4.01	15.9	21.5	>250.0		1395		
	39.70	4.39							
	39.78	4.03			38.7				
	41.44				>250.0		1000	x	x
	41.55	4.17	15.5	21.0					
	41.60		16.5	20.3	>250.0	x			
	41.75	4.25					920		
	41.78		15.0	20.7	>250.0				
	41.95						800		
	42.07	3.63							
	42.10		15.2	20.4	>250.0	x			
	45.61		14.7	21.1	>250.0	x	945		
	45.65						1000		
	46.00						870		
	46.20						980		
	46.58		14.4	21.2	>250.0	x	1070		
	48.47		16.7	21.2	>250.0	x	2020		
	49.40		16.7		245.5	x	810		

Table 6.3.1 cont.

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985								
SITE	DEPTH (m)	T.CON. mcal/ °Ccms	WATER %	DENS. kN/m ³	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM. TRIAX
D85-O	0.03	2.30						
	0.10				16.6			
	0.20				8.9			
	0.30	2.30	78.7	15.7	10.8			
	0.40				7.4			
	0.50				21.1			
	0.60	2.41			16.6			
	0.70				8.1			
	0.80				11.7			
	0.90	2.45						
	1.00				10.8			
	1.10				4.8			
	1.20		20.9	21.0	38.7			
	1.25	4.39						
	5.27		16.4	21.0	200.0	x	2030	x
	5.28	3.94						
	5.34	4.17	15.8	21.2				
	5.38	4.34						
	9.80		14.7	21.5	>250.0	x	1930	
	9.85	5.04						
	9.94	5.40						
	22.08		15.6	21.0	>250.0	x	2025	x
	22.14	3.99						
	22.80	3.87						
	25.38		23.9	19.9	>250.0	x	1730	
	25.41	3.18						
	25.48	3.27						
	30.16		12.9	22.2	>250.0	x	1995	x
	30.23	4.75	13.2	21.8				
	30.28	5.11						
	35.05		17.3	20.7	159.0	x	2185	
	35.12		11.7	22.2	>250.0	x	2005	x
	35.16	4.77	11.1	22.4				
	35.23	5.11						
	42.01		12.4	22.0				x
	42.11	5.09						
	42.14		12.2	20.4	>250.0	x	1905	

SAMPLE ANALYSIS SHALLOW DRILLING BARENTS SEA 1985								
SITE	DEPTH (m)	T.CON. mcal/ °Ccms	WATER %	DENS. kN/m ³	SH.STR. kN/m ²	GR.SIZE	VELOC. m/s	OEDOM. TRIAX
D85-V	0.03	3.49						
	0.03	3.62						
	0.14	4.21						
	0.24	4.73						
	0.27	4.19						
	0.42	5.20						
	14.67	4.07	23.2	20.5		x		
	14.77	3.83						x
	14.82		22.3	20.6	>250.0	x	1720	
	20.06		21.2	20.6	>250.0		1810	x
	20.10	4.32						
	20.20	4.28	20.3	20.8				
	24.48	3.94						
	24.58	4.03						
	24.61		20.6	20.7	>250.0	x	1715	
	24.75		19.4	20.8	>250.0	x	1800	
	24.84	3.44						
	24.97	3.89						
	25.12	3.74			>250.0			
	25.34	3.67						
	25.46	4.32						
	25.49		21.1	20.6	>250.0	x	1790	
	29.63	4.59	15.7	21.1			1950	x
	29.76	5.09	16.2	21.7				
	29.79					x		
	39.95		16.4	21.4				x
	40.06		18.2	21.0		x	1780	
	40.12	4.53						
	40.19	4.41						
	44.40		20.6	20.5		x	1810	
	44.50	4.62						
	44.56		19.1	19.6		x		
	44.59	4.88						
	45.82	4.91						
	45.94	4.50						
	45.97		18.0	19.5		x	970	
	54.84		28.0	19.0		x	1680	

Advanced soil mechanical property testing (triaxial and oedometer testing) is carried out at SINTEF's Division of Geotechnical Engineering (Senneset 1986) and the results from these tests are summarized in Table 6.3.2. The following abbreviations are used.

w	-	water content (% water of dry weight)
w_L	-	liquid limit (% water of dry weight)
w_p	-	plasticity limit (% water of dry weight)
a	-	Coulombian attraction (kN/m^2)
c_v	-	coefficient of consolidation (m^2/yr)
m	-	deformation modulus number
S_u	-	undrained shear strength (kN/m^2)
$\tan\phi$	-	tangens of angle of failure
ϵ_f	-	deformation at failure
p_o'	-	effective overburden pressure (kN/m^2)
p_c'	-	effective preconsolidation pressure (kN/m^2)

Table 6.3.2 Results from soil mechanical testing of Quaternary core material.

Boring No.	Sample depth m	W %	Index properties				Overburden pressure $\sim p_o$ kN/m ²	Oedometer tests			Triaxial tests			Remarks Sample quality
			w_L %	w_p %	Unit weight kN/m ³	S_u kN/m ²		$-p_c$ kN/m ²	Modulus number m	Coeff. of consolid. c_v m ² /gr.	Attraction a kN/m ²	$\tan \phi$	Def. at failure ϵ_f %	
C 85-A	8.24- 8.38	29.9	36.2	18.9	20.0	43	81	120	16	(0.9)	20	0.43	4	Partly disturbed
	14.02-14.17	22.3	44.6	22.3	20.6	135	145	350	22	(0.8)	20	0.55	10	Disturbed
	23.15-23.29	20.6	38.6	21.4	21.3	210	241	1550	9.5	2.9	20	0.56	4	Good
	34.18-34.32	24.6	46.5	23.0	21.7	240	370	1700	15	(0.4)	50	0.42	4	Good
	37.60-37.85	26.1	41.6	21.4	20.0	140	403	-	21.5	-	20	0.41	8	Disturbed
C 85-E	10.67-10.73	23.5	33.2	20.2	20.5	-	112	210	18	1.1	-	-	-	Partly disturbed
	21.98-22.02	23.8	38.7	21.7	20.5	120	223	230	17	1.3	50	0.42	5	Disturbed
	31.0 -31.14	20.8	39.9	23.1	20.7	140	332	330	25	1.1	20	0.51	8	Disturbed
	40.25-40.46	21.6	39.9	23.1	21.1	250	423	700	20	0.6	20	0.52	8	Partly disturbed
	78.42-78.55	22.1	42.3	22.0	20.4	410	817	-	24	1.1	50	0.49	8	Disturbed
D 85-H	13.71-13.78	12.9	33.8	19.1	22.1	-	163	800	35	2.3	-	-	-	Good
D 85-K	10.60-10.70	30.3	36.4	22.9	-	40	-	-	-	-	20	0.42	6	Partly disturbed
	19.40-19.81	22.2	42.1	20.4	20.4	75	201	400	15	0.5	20	0.45	8	Partly disturbed
	30.23-30.33	-	-	-	-	120	-	-	-	-	20	0.42	4	Good
	34.32-34.45	21.0	45.9	23.1	-	170	350	630	17	0.5	20	0.56	6	Disturbed
	43.95-44.1	20.1	44.9	21.2	20.8	220	450	-	14	2.7	50	0.50	6	Disturbed
	54.67-54.82	21.2	46.1	17.7	20.9	180	565	800	15	1.7	50	0.48	6	Partly disturbed
	64.67-64.82	18.6	39.4	21.0	21.4	390	674	1050	20	4.0	50	0.63	8	Partly disturbed
	84.14-84.3	16.3	45.7	23.6	21.0	480	910	2700	25	5.0	50	0.49	9	Partly disturbed
D 85-N	12.09-12.19	-	32.8	18.4	21.7	135	140	-	-	-	50	0.48	6	Disturbed
	27.24-27.28	15.7	30.2	18.4	21.7	-	319	320	32	1.8	-	-	-	Good
	30.12-30.20	15.1	29.4	19.3	21.5	-	347	1300	32	4.3	-	-	-	Good
	35.10-35.16	14.5	31.9	18.5	21.5	-	404	1200	24.5	(3.0)	-	-	-	Disturbed
	39.60-39.65	15.9	50.2	19.4	21.5	-	456	-	23.5	-	-	-	-	Disturbed
	41.44-41.58	15.5	30.3	18.4	21.0	300	477	-	24.5	-	20	0.48	6	Disturbed
D 85-O	5.27- 5.41	15.8	28.1	17.3	21.2	65	60	150	29	-	20	0.55	7	Partly disturbed
	22.08-22.68	-	30.5	18.4	-	160	-	-	-	-	20	0.49	6	Good
	30.16-30.31	13.2	27.9	17.3	21.8	320	367	570	28	2.7	50	0.52	6	Good
	35.12-35.26	11.1	23.9	15.9	22.4	420	430	-	26.5	-	50	0.52	6	Disturbed
	42.01-42.06	12.4	24.0	15.7	22.0	-	450	-	34.5	-	-	-	-	Disturbed
D 85-V	14.64-14.78	23.2	50.5	27.3	20.5	135	154	900	7	0.5	20	0.51	8	Partly disturbed
	20.06-20.21	20.3	50.4	25.8	20.8	175	213	1000	12.5	0.9	50	0.45	5	Good
	29.63-29.77	16.2	43.1	22.7	21.7	(480)	341	1400	13	-	20	0.73	4	Good
	39.92-40.06	16.4	40.7	21.2	21.4	340	447	500	27.5	1.0	50	0.52	6	Disturbed

Table 6.3.3 Total organic carbon (TOC) and total carbon (TC) in per cent of sediment.

SITE	DEPTH (M)	TOC%	TC%
C85-E	10.67	1.12	1.73
C85-E	21.75	1.16	1.79
C85-E	31.03	1.40	1.98
D85-K	10.10	1.13	1.66
D85-K	14.61	1.18	1.60
D85-K	54.45	1.21	1.52
D85-K	64.67	1.15	1.65
D85-K	94.87	0.94	1.25
D85-N	21.89	1.95	2.56
D85-N	27.31	1.45	1.68
D85-N	30.26	1.28	1.60
D85-N	35.11	1.29	1.77
D85-N	39.66	1.47	2.00
D85-N	41.75	1.37	1.75
D85-N	42.07	1.26	1.76
D85-O	5.20	1.90	1.61
D85-O	22.14	2.52	2.94
D85-O	35.12	1.43	1.94
D85-O	42.14	1.54	2.17
D85-V	24.60	1.24	1.58
D85-V	24.75	1.27	1.76
D85-V	25.49	1.24	1.62
D85-V	44.40	0.92	1.17
D85-V	44.56	0.46	0.67

Table 6.3.4a Semiquantitative mineral composition interpreted from X-ray diffractograms, bulk samples (wt% minerals).

Sites depth Minerals	C85-E 40.21 m	C85-E 31.08 m	C85-E 10.73 m	D85-K 34.49 m	D85-K 43.95 m	D85-K 14.61 m	D85-K 14.80 m	D85-K 22.34 m	D85-K 54.82 m	D85-K 25.66 m	D85-N 41.78 m	D85-N 27.31 m	D85-N 21.92 m	D85-N 35.16 m		
Quartz * Alk.fld. Plagioclase Musc./Ill.10A Mix.clay 10-14A Chlorite Smectite Kaolinite Calcite Dolomite Siderite Pyrite	38 6 17 14 5 5 1 11 1 tr - 2	31 4 17 14 1 14 - 17 - tr - 2	23 5 16 19 5 16 - 14 - 2	31 8 22 17 6 7 - 6 tr 1 - 2	31 3 20 14 9 15 - 6 - 2	26 3 17 22 10 10 1 8 - - 3	36 3 23 14 5 8 - 6 - 2	32 4 23 14 9 7 - 9 2	34 3 17 10 3 7 - 22 tr - 2 2	31 4 19 14 7 5 tr 17 2 - - 2	42 3 12 20 8 3 1 10 - 10 - - 2	34 3 9 18 5 7 - 17 - - 3 5	47 4 12 11 5 7 - 10 - - 1 3	36 2 14 20 5 10 - 11 - - 2 2		
	Quartz Alk. fld. Plagioclase Musc. Ill. 10A Mix.clay 10-14 A Chlorite Smectite Kaolinite Calcite Dolomite Siderite Pyrite	38 1 5 30 8 4 tr 9 - - 4	22 2 15 32 10 8 - 8 tr - 2	37 2 6 27 11 4 1 8 - 8 - 5	47 2 8 13 6 7 - 6 - 3 9	51 2 6 12 13 5 tr 6 tr 1 4 6	51 1 3 9 16 4 tr 7 - - 3 6	46 tr 5 12 13 - - - 14 1 - - 3 7	46 1 5 13 9 10 - 9 - - 3 4	45 6 10 17 7 3 - 11 - - - 1	19 2 5 25 24 7 9 6 - - tr 1	30 5 19 13 4 9 1 14 - - 1 2	28 4 18 15 11 5 - 17 - - 1 1	57 5 8 16 14 6 8 - 2 - - tr	29 3 15 16 7 7 1 18 - - 2 2	D85-V 14.77 m <

Alk.fld. = Alkalifeldspar

Musc./ill = Muscovite/Illite (mainly illite in these samples).

Dolomite and siderite are questionable, the reflections interfere with mica and feldspars.

Table 6.3.4 b Semiquantitative mineral composition interpreted from X-ray diffractograms, fractioned samples (wt % minerals).

Site C85-E Depth: 10.10 m Fraction $\leq 4 \mu\text{m}$ Mineral	Site C85-E Depth: 21.75 Fraction $\leq 4 \mu\text{m}$ 4-63 μm $>63 \mu\text{m}$ bulk (48.47%) (34.89%) (16.64%) Mineral					Site C85-K Depth: 14.65 m Fraction $\leq 4 \mu\text{m}$ Mineral
Quartz 2.2	Quartz 4.1	7.1	66.7	15.6		Quartz 2.7
K-feldspar 1.4	K-feldspar 1.3	2.7	5.8	2.5		K-feldspar 1.3
Plagioklase 2.8	Plagioklase 3.2	9.8	27.6	9.6		Plagioklase 2.4
Mica 30.8	Mica 32.6	22.2	25.8	27.8		Mica 32.3
Chlorite 41.8	Chlorite 44.9	35.6	-	34.2		Chlorite 43.0
Kaolinite -	Kaolinite 1.9	4.9	-	2.6		Kaolinite -
Smectite 6.4	Smectite -	-	-	-		Smectite 8.2
Mixed 10-14 Å 14.2	Mixed 10-14 Å 11.7	17.8	-	11.9		Mixed 10-14 Å 10.0
Calcite -	Calcite -	-	-	-		Calcite -
Dolomite -	Dolomite -	-	-	-		Dolomite -
Siderite -	Siderite -	-	-	-		Siderite -
Pyrite 0.5	Pyrite 0.4	-	-	tr		Pyrite tr
Amphibole -	Amphibole -	-	tr	tr		Amphibole -

Site C85-K Depth: 54.45 Fraction $\leq 4 \mu\text{m}$ Mineral	Site C85-K Depth: 64.67 Fraction $\leq 4 \mu\text{m}$ 4-63 μm $>63 \mu\text{m}$ bulk (49.5%) (33.7%) (16.8%) Mineral					Site C85-K Depth: 94.87 Fraction $\leq 4 \mu\text{m}$ Mineral
Quartz 3.4	Quartz 3.5	6.7	74.5	16.5		Quartz 4.7
K-feldspar 1.5	K-feldspar 1.8	3.0	4.3	2.6		K-feldspar 1.0
Plagioklase 3.0	Plagioklase 2.8	6.0	21.3	7.0		Plagioklase 2.0
Mica 33.7	Mica 32.8	25.8	-	25.0		Mica 15.1
Chlorite 43.2	Chlorite 38.0	31.9	-	29.6		Chlorite 39.1
Kaolinite -	Kaolinite 1.0	9.7	-	3.8		Kaolinite 0
Smectite 2.5	Smectite 5.6	-	-	2.8		Smectite 22.3
Mixed 10-14 Å 12.3	Mixed 10-14 Å 13.9	16.8	-	12.6		Mixed 10-14 Å 15.9
Calcite -	Calcite -	-	-	-		Calcite -
Dolomite -	Dolomite -	-	-	-		Dolomite -
Siderite -	Siderite -	-	-	-		Siderite -
Pyrite 0.3	Pyrite 0.5	-	-	tr		Pyrite -
Amphibole -	Amphibole -	-	-	-		Amphibole -

Table 6.3.4 b cont.

Site C85-N Depth: 21.89 m Fraction $\leq 4 \mu\text{m}$ Mineral	Site C85-N Depth: 27.31 m Fraction $\leq 4 \mu\text{m}$ Mineral	Site C85-N Depth: 30.26 Fraction $\leq 4 \mu\text{m}$ Mineral	Site C85-N Depth: 35.11 m Fraction $\leq 4 \mu\text{m}$ Mineral
Quartz 4.4 K-feldspar 2.0 Plagioklase - Mica 39.8 Chlorite 31.5 Kaolinite 10.9 Smectite - Mixed 10-14 A 10.7 Calcite - Dolomite - Siderite - Pyrite 0.7 Amphibole -	Quartz 3.5 K-feldspar 1.2 Plagioklase 2.0 Mica 38.0 Chlorite 38.1 Kaolinite 5.3 Smectite 0 Mixed 10-14 A 11.2 Calcite - Dolomite - Siderite - Pyrite 0.6 Amphibole -	Quartz 3.2 K-feldspar 0.9 Plagioklase 1.9 Mica 48.1 Chlorite 36.7 Kaolinite 2.9 Smectite - Mixed 10-14 A 6.0 Calcite - Dolomite - Siderite - Pyrite 0.2 Amphibole 0	Quartz 3.4 K-feldspar 1.1 Plagioklase 2.7 Mica 43.3 Chlorite 34.0 Kaolinite 4.8 Smectite 0 Mixed 10-14 A 10.8 Calcite - Dolomite - Siderite - Pyrite - Amphibole -

Site C85-N Depth: 39.66 Fraction $\leq 4 \mu\text{m}$ Mineral	Site C85-N Depth: 41.75 Fraction $\leq 4 \mu\text{m}$ Mineral	Site C85-N Depth: 42.07 Fraction $\leq 4 \mu\text{m}$ (35%) 4-63 μm (32%) >63 μm (33%) bulk
Quartz 3.9 K-feldspar 1.3 Plagioklase 3.0 Mica 30.8 Chlorite 43.2 Kaolinite 4.0 Smectite 0 Mixed 10-14 A 13.3 Calcite - Dolomite - Siderite - Pyrite 0.4 Amphibole 0	Quartz 3.1 K-feldspar 1.0 Plagioklase 2.8 Mica 50.0 Chlorite 30.8 Kaolinite 6.7 Smectite 0 Mixed 10-14 A 5.8 Calcite - Dolomite - Siderite - Pyrite - Amphibole -	Quartz 4.1 K-feldspar 0.8 Plagioklase 2.2 Mica 42.6 Chlorite 42.0 Kaolinite - Smectite - Mixed 10-14 A 8.2 Calcite - Dolomite - Siderite - Pyrite tr Amphibole 0
		8.7 1.0 5.6 33.3 36.4 - - 14.9 - - - - - 1.5 -
		29.6 2.0 7.9 25.6 26.4 - - 7.6 - - - - - -

Site C85-0 Depth: 5.28 Fraction <u>4 um</u> 4-63 um >63 um bulk Mineral	Site C85-0 Depth: 22.14 m Fraction <u>4 um</u> Mineral	Site C85-0 Depth: 35.12 m Fraction <u>4 um</u> Mineral
Quartz 2.6 14.1 76.2 29.3 K-feldspar 0.4 0.5 2.4 2.85 Plagioklase 0.8 3.0 8.3 3.9 Mica 55.6 35.2 1.2 31.9 Chlorite 26.2 17.1 7.1 16.1 Kaolinite - 18.1 - 7.1 Smectite - - - - Mixed 10-14 A 13.4 11.6 - 8.6 Calcite - - - - Dolomite - - - - Siderite 0.3 - 2.4 tr Pyrite 0.6 0.5 2.4 1.1 Amphibole - - - -	Quartz 4.1 K-feldspar 0.3 Plagioklase 1.5 Mica 39.4 Chlorite 30.6 Kaolinite 3.2 Smectite 0 Mixed 10-14 A 20.2 Calcite - Dolomite - Siderite - Pyrite 0.7 Amphibole -	Quartz 3.6 K-feldspar 0.8 Plagioklase 1.4 Mica 38.7 Chlorite 15.2 Kaolinite 13.4 Smectite 0.3 Mixed 10-14 A 26.0 Calcite - Dolomite - Siderite 0.2 Pyrite 0.3 Amphibole -

Table 6.3.4 b cont.

Site C85-V Depth: 25.49 Fraction $\leq 4 \mu\text{m}$ Mineral		Site C85-V Depth: 44.40 m Fraction $\leq 4 \mu\text{m}$ Mineral	
Quartz	3.7	Quartz	2.6
K-feldspar	0.6	K-feldspar	1.1
Plagioklase	2.2	Plagioklase	2.7
Mica	30.3	Mica	37.3
Chlorite	35.8	Chlorite	32.7
Kaolinite	-	Kaolinite	1.4
Smectite	11.6	Smectite	19.7
Mixed 10-14 Å	15.7	Mixed 10-14 Å	1.9
Calcite	-	Calcite	0.1
Dolomite	-	Dolomite	-
Siderite	-	Siderite	-
Pyrite	0.1	Pyrite	0.5
Amphibole	-	Amphibole	-

6.3.3 Theoretical calculation of thermal conductivity from sediment composition

Objective and main findings

The purpose of the theoretical calculation of thermal conductivity was initially to investigate for explanations of the rather poor correlation between thermal conductivity and water content shown by Sættem et al. (1985) (IKU report 21.3405.00/02/85, Fig. 4.2.5.1) and to perform corrections for the in situ versus laboratory conditions for the thermal conductivity measurements.

Various possible explanations for the above mentioned scatter were investigated and the main conclusions are.

1. Correction on measured thermal conductivity due to different temperature in situ and in the laboratory can be neglected.
2. The measured variation in thermal conductivity can be correlated with a variation in quartz content, and generally the range in the scatter of the plot of thermal conductivity versus water content can be explained by the actual range in quartz content (10% - 50% of the mineral fractions).
3. Samples from sites D85-N, D85-O and D85-V were expanded due to gas release in the pores at the time when thermal conductivity was measured on board. To correct for this the measured values must be increased (a maximum effect of 16% for the data from site D85-N).
4. The average ratio between the thermal conductivities reported by Sættem et al. (1985) and the corresponding theoretical thermal conductivities is 1.81.

Methods and input data

In marine sediments, the major factor determining the thermal conductivity is generally assumed to be the water content. Since the sediments will be fully saturated, this parameter simply reflects the volumetric composition of the material (solid fraction - water saturated void fraction). Another factor of significance in such binary mixtures is the thermal conductivity of the solid fraction, which is determined by the mineral composition of the sediment.

A method for computation of the thermal conductivity of water saturated sediments which has gained wide acceptance is based on the geometric mean equation (Johansen 1975; Farouki 1981; Matsuda & von Herzen 1986), i.e.:

$$K = K_w^n K_s^{(1-n)} \quad (1)$$

where K_w is the thermal conductivity of water,
 K_s , mineral thermal conductivity,
 n , void fraction (water)
 $(1-n)$, solid fraction (minerals)

Equation 1 may also be used inversely to determine thermal conductivity of the solid fraction, presumed that the saturated conductivity and the volumetric composition are known.

The thermal conductivity of the solid fraction may also be estimated by a corresponding equation, presumed that information is available on the mineral composition and the thermal conductivity of the mineral constituents:

$$K_s = K_1^{n_1} K_2^{n_2} \dots K_i^{n_i} \quad (2)$$

where K_i , n_i is the thermal conductivity and volume fractions of the mineral constituents.

By an inspection of the mineralogical composition determined from the samples, the mineralogical composition is found to be acceptably represented by three constituents, i.e. quartz, clorite and an unspecified rest. Thermal conductivities for these are given in Table 6.3.5.

When the measurements are made on board a ship, the core samples may have undergone changes implying loss of water due to drainage, or expansion of gas causing gas filled void in the samples. The temperature of the samples will also deviate from the in situ conditions. In order to adjust the measurements to in situ condition, corrections for these factors must be applied.

As rather insignificant change in the thermal conductivity with temperature has been found for the Barents Sea sediments in the actual temperature range, as the opposite temperature effect on the thermal conductivity of water and minerals largely cancel each other for those low water content sediments. The temperature deviation may therefore be neglected. For the remaining factors, the following procedures are applied:

Presumed that the gas filled voids are limited by volume to a few percent, corrections for lack of saturation may be made by the use of the Hashin-Shtrikman upper bound (HS-bound), which represents a mixture where a low conductivity material (gas) is homogeneously dispersed in a high conductivity (solid) matrix (Johansen 1975).

When the conductivity ratio is large (i.e. $K_s/K_f \gg 1$), the upper HS-bound may be approximated by:

$$K = f(v) K_{sat} \quad (3)$$

where $f(v) = 2(1 - v)/(2 + v)$

K_{sat} , conductivity of saturated sediments,
 v , void volume fraction (gas filled).

The adjustment of a measured conductivity to saturation thus implies division of the measured value by the factor $f(v)$, where v is derived from the density data of the sample. This presumes knowledge of the following parameters:

- Wet density: ρ (measured)
- Water content by dry weight: w (measured)
- Density of solids: ρ_s (estimated to 2700 kg/m^3)

From these parameters, the following may be derived:

$$\text{- Dry density: } \rho_d = \rho / (1+w) \quad (4a)$$

$$\text{- Volume fraction of solids: } n_s = \rho_d / \rho_s \quad (4b)$$

$$\text{- Volume fraction of water: } n_w = \rho_d w / \rho_w \quad (4c)$$

$$\text{- Void volume fraction (gas): } v = 1 - (n_s + n_w) \quad (4d)$$

The density data refers to the sample under the laboratory conditions. By assuming that the measured water content represents all water present in the undisturbed sample, an adjusted dry density value may be obtained, i.e.:

$$n_w + n_s = 1, \quad (5a)$$

$$\text{which implies: } w \rho'_d / \rho_w + \rho'_d / \rho_s = 1,$$

i.e.:

$$\text{- Adjusted dry density: } S'_d = 1 / (w / \rho_w + 1 / \rho_s) \quad (5b)$$

$$\text{- Adjusted void fraction: } n = 1 - \rho'_d / \rho_s \quad (5c)$$

$$\text{- Volume fraction of water: } n'_w = 1 - \rho'_d / \rho_s$$

The thermal conductivities for the different compounds all refer to laboratory conditions, i.e. 20°C and atmospheric pressure. The following values are applied.

Table 6.3.5 Thermal conductivity of compounds in atmospheric pressure and 20°C.

Compound	Thermal conductivity	
	$\text{Wm}^{-1} \text{K}^{-1}$	$\text{mcal cm}^{-1} \text{ } ^\circ\text{C}^{-1} \text{ s}^{-1}$
Water	(0.599)	(1.431)
Chlorite	3.5	8.36
Quartz	7.7	18.36
Mineral rest.	2.0	4.78

Input values to equations 1-5c about composition of each sample are summarized in Table 6.3.6. All the values are related to on board thermal conductivity measurements at certain depths. In reality the measurements may have been carried out within a depth range of a few cm to a few dm away from the nominal depth. The correlation is, however, done after a careful representativity consideration. (For actual depth of each analysis, see previous chapter).

Table 6.3.6 Summary of some physical properties and input values for the calculations in eq. 1-5 above.

Site	Depth (m)	Therm cond. (mcal/ cm °Cs)	Water content (wt%)	Wet density (kN/m ³)	Temperature (°C)		Mineral content (wt. %) Qtz. Chl.	Sound velocity (m/s)
					in situ	lab.		
C85-A	0.03	1.64						
	0.36	1.51						
	0.87	1.43						
	0.94	1.42						
	1.20	1.37						
	1.60	1.34						
	1.85	1.36						
	2.20	1.48						
	2.55	1.46						
	2.95	1.38						
	5.90	1.76						
	8.53	2.38	27.7	19.8	4.1	19.4		1670
	14.02	2.34	22.0	20.8	4.3	20.6		1710
	18.65	2.31	20.4	21.0	4.6	20.6		1785
	23.15	2.35	20.6	20.9	4.8	20.6		1760
	23.36	2.32	22.2	19.6	4.8	20.4		1785
	34.15	2.27	21.0		5.2	19.0		1844
C85-E	0.04	2.01	36.9					
	0.35	2.13						
	0.70	1.71						
	1.00	2.05						
	1.03	2.02						
	1.40	1.68						
	1.80	1.67	40.8	18.2				1590

Table 6.3.6 cont.

Site	Depth (m)	Therm cond. (mcal/ cm °Cs)	Water content (wt%)	Wet density (kN/m ³)	Temp. (°C)		Mineral content (wt. %) Qtz. Chl.	Sound velocity (m/s)
					in situ	Lab.		
C85-E cont.	2.10	1.79						
	10.53	2.18			4.0	24.7		1800
	10.60	2.32			4.0	25.2		1800
	10.64	2.10	21.6	20.8	4.0	23.9	23.0 16.0	1800
	21.28	2.29			4.4	23.2		
	21.34	2.36			4.4	23.2		
	21.40	2.17			4.4	23.1		
	21.47	2.19			4.4	23.0		
	21.55	1.70	21.5	20.9	4.4	23.4	15.6 34.2	1750
	31.03	2.26	22.5		4.7	22.5		
	31.08	2.36			4.7	22.6	31.0 14.0	1910
	31.13	2.49			4.7	22.8		
	40.13	2.19			5.0	20.9		
	40.21	2.15	20.0	20.9	5.0	20.9	38.0 5.0	1805
	40.30	2.24			5.0	21.6		1825
	40.37	2.06	22.5	20.3	5.0	21.6		
	50.12	2.24			5.3	24.1		1755
	50.15	2.20			5.3	23.8		1755
	50.22	2.16			5.3	23.6		1755
D85-H	0.03	1.56			3.2			
	0.24	1.55			2.5			
	0.44	2.12	35.1	19.0	2.0			
	0.50	1.98	35.1	19.0	2.0			1675
	0.72	1.67			1.9			
	0.97	1.61			1.8			
	1.22	1.76			1.8			
	1.44	1.90	34.0	19.5	1.7			

Table 6.3.6 cont.

Site	Depth (m)	Therm cond. (mcal/ cm °Cs)	Water content (wt%)	Wet density (kN/m ³)	Temp. (°C)		Mineral content (wt. %)		Sound velocity (m/s)
					in situ	Lab.	Qtz	Chl	
C85-H cont.	11.25	2.60	15.8		2.5	21.4			1790
	13.56	2.57	13.8	21.8	3.0	21.8			1990
C85-I	0.03	2.05							
	0.22	1.61							
	0.47	1.88							
	0.72	1.98							
	0.97	2.49							
	5.44	2.13			2.2				
	5.54	2.42	19.2	21.1	2.2				1970
D85-K	0.03	1.51	60.2	16.8					
	0.33	1.61							
	0.71	1.63							
	1.09	1.75							
	1.17	1.68							
	1.56	1.67	39.0	18.6					
	1.86	1.69							
	2.33	1.82	38.4	18.8					
	9.97	1.93	29.4	19.5	2.1				
	10.00	1.98							
	10.08	2.06	29.7	19.6	2.1	19.8			1620
	10.71	1.99	31.5	19.5	2.1	19.4			1640
	14.64	2.00	28.9	19.6	2.3	20.8	26.0	10.0	1630
	14.71	1.81			2.3	20.9			
	14.77	2.01	29.5	19.4	2.3	21.0	36.0	8.0	1640

Table 6.3.6 cont.

Site	Depth (m)	Therm cond. (mcal/ cm °Cs)	Water content (wt%)	Wet density (kN/m ³)	Temperature (°C)		Mineral content (wt. %)		Sound velocity (m/s)
					in situ	Lab.	Qtz.	Chl.	
D85-K cont.	14.83	1.98	29.5	19.4	2.3	21.1	36.0	8.0	1640
	14.89	1.81			2.3	21.2			
	14.92	2.01	31.0	19.2	2.3	21.2			1595
	19.63	2.26	23.4	20.5	2.5	21.6			1700
	19.71	2.01			2.5	21.7			1700
	19.78	2.27			2.5	21.5			1700
	22.37	2.02	21.8	20.8	2.5	21.5	32.0	7.0	1725
	22.46	2.07	21.8	20.8	2.5	21.8			1725
	22.35	2.02			2.5	21.5			1725
	25.52	2.19			2.7	21.9			
	25.59	2.12	22.2	20.7	2.7	21.6			2135
	25.63	2.24	22.2	20.7	2.7	21.6	31.0	5.0	2135
	30.21	2.07			2.8	21.8			
	30.27	2.14			2.8	21.6			
	30.31	2.27			2.8	21.6			1765
	34.51	2.24	20.3	21.0	3.0	20.6			1780
	34.57	2.28			3.0	20.4			
	34.63	2.45			3.0	20.4			1825
	43.98	2.29	20.3	21.0	3.3	20.9	31.0	15.0	1795
	44.03	2.35			3.3	21.0			
	44.07	2.43			3.3	21.0			
	54.66	2.17	18.0	21.2	3.7	22.0			1830
	54.72	2.07			3.7	22.0			1770
	54.79	2.23			3.7	21.9	34.0	7.0	1770
	64.74	2.16	17.9	21.3	3.9	22.1	16.5	29.6	1845
	64.79	2.18			3.9	22.2			1845
	84.27	2.22			4.6	22.2			
	94.87	2.16	18.8	20.8	4.9	22.6			
	94.92	1.96			4.9	22.2			

Table 6.3.6 cont.

Site	Depth (m)	Therm cond. (mcal/ cm °Cs)	Water content (wt%)	Wet density (kN/m ³)	Temperature (°C)		Mineral content (wt. %)		Sound velocity (m/s)
					in situ	Lab.	Qtz.	Chl.	
D85-N	0.03	1.69							
	0.23	1.89							
	0.43	1.90	27.7	19.9					
	0.62	2.09							
	12.08	2.10	16.4	20.6	1.9	23.6			1855
	12.16	2.34	16.4	20.6	1.9	23.5			1845
	21.83	2.17			2.3	23.3			
	21.89	2.13	16.3	21.1	2.3	23.2	47.0	7.0	
	27.26	2.22			2.4	23.4			
	27.31	1.95	16.7	20.5	2.4	23.8	34.0	7.0	1800
	30.23	2.50			2.5	22.9			
	30.26	2.66	14.7	21.2	2.5	22.8			
	36.11	2.50	15.5	20.2	2.7	20.7			
	36.14	2.51			2.7	21.1			
	39.66	2.23	15.7	21.4	2.9	24.1			
	39.70	2.44			2.9	24.0			
	39.78	2.24			2.9	24.3			
	41.55	2.32	16.5	20.3	3.0	23.5			
	41.75	2.36	15.0	20.7	3.0	23.6	42.0	3.0	
	42.07	2.02	15.2	20.4	3.1	23.8	29.6	26.3	
D85-0	0.03	1.28							
	0.30	1.28	78.7	15.7					
	0.60	1.34							
	0.90	1.36							
	1.25	2.44	20.9	21.0					
	5.28	2.19	16.4	21.0			51.0	5.0	2030
	5.34	2.32							
	5.38	2.41							

Table 6.3.6 cont.

Site	Depth (m)	Therm cond. (mcal/ cm °Cs)	Water content (wt%)	Wet density (kN/m ³)	Temperature (°C)		Mineral content (wt. %)		Sound velocity (m/s)
					in situ	Lab.	Qtz.	Chl.	
D85-0 cont.	9.85	2.80	14.7	21.5	-0.2	21.8	47.0	7.0	1930
	9.94	3.00			-0.2	21.9	46.0	13.0	
	22.14	2.22	15.6	21.0	0.2	22.4	37.0	4.0	
	22.80	2.15			0.2	21.6			
	25.41	1.77	23.9	19.9	0.3	22.2	11.1	32.5	1730
	25.48	1.82			0.3	21.8	22.0	4.0	1730
	30.23	2.64	12.9	22.2	0.5	21.7	38.0	4.0	1995
	30.28	2.84			0.5	21.7			1995
	35.16	2.65	11.7	22.2	0.6	21.6			2005
	35.23	2.84			0.6	21.7	46.0	0.0	
	42.11	2.83	12.2	20.4	0.8	23.2	51.0	4.0	1905
D85-V	0.03	1.94							1720 1810 1715 1800 1790
	0.03	2.01							
	0.14	2.34							
	0.24	2.63							
	0.27	2.33							
	0.42	2.89							
	14.67	2.26			2.3	20.8			
	14.77	2.13	22.3	20.6	2.3	20.9	18.0	26.0	
	20.10	2.40	21.2	20.6	2.5	20.8			
	20.20	2.38			2.5	20.6			
	24.48	2.19			2.7	19.9			
	24.58	2.24	20.6	20.7	2.7	20.1	11.1	32.5	
	24.84	1.91	19.4	20.8	2.7	19.9	29.0	7.0	
	24.97	2.16			2.7	20.2	30.0	9.0	
	25.12	2.08			2.7	20.1			
	25.34	2.04			2.7	20.4			
	25.46	2.40	21.1	20.6	2.7	20.2	28.0	5.0	

Table 6.3.6 cont.

Site	Depth (m)	Therm cond. (mcal/ cm °Cs)	Water content (wt%)	Wet density (kN/m ³)	Temperature (°C)		Mineral content (wt. %)		Sound velocity (m/s)
					in situ	Lab.	Qtz.	Chl.	
D85-V cont.	29.63	2.55	15.7	21.1	2.9	20.8	45.0	3.0	1950
	29.76	2.83			2.9	20.5			
	40.12	2.52	18.2	21.0	3.3	20.2	19.0	7.0	1780
	40.19	2.45			3.3	20.0			
	44.50	2.57	20.6	20.5	3.5	18.9			1810
	44.59	2.71	19.1	19.6	3.5	18.8	57.0	8.0	
	45.82	2.73			3.5	20.5			
	45.94	2.50	18.0	19.5	3.5	20.5			

Results

Fig. 6.3.1 shows a plot of measured thermal conductivity (uncorrected values) versus theoretical thermal conductivity calculated by equations 1-5 above.

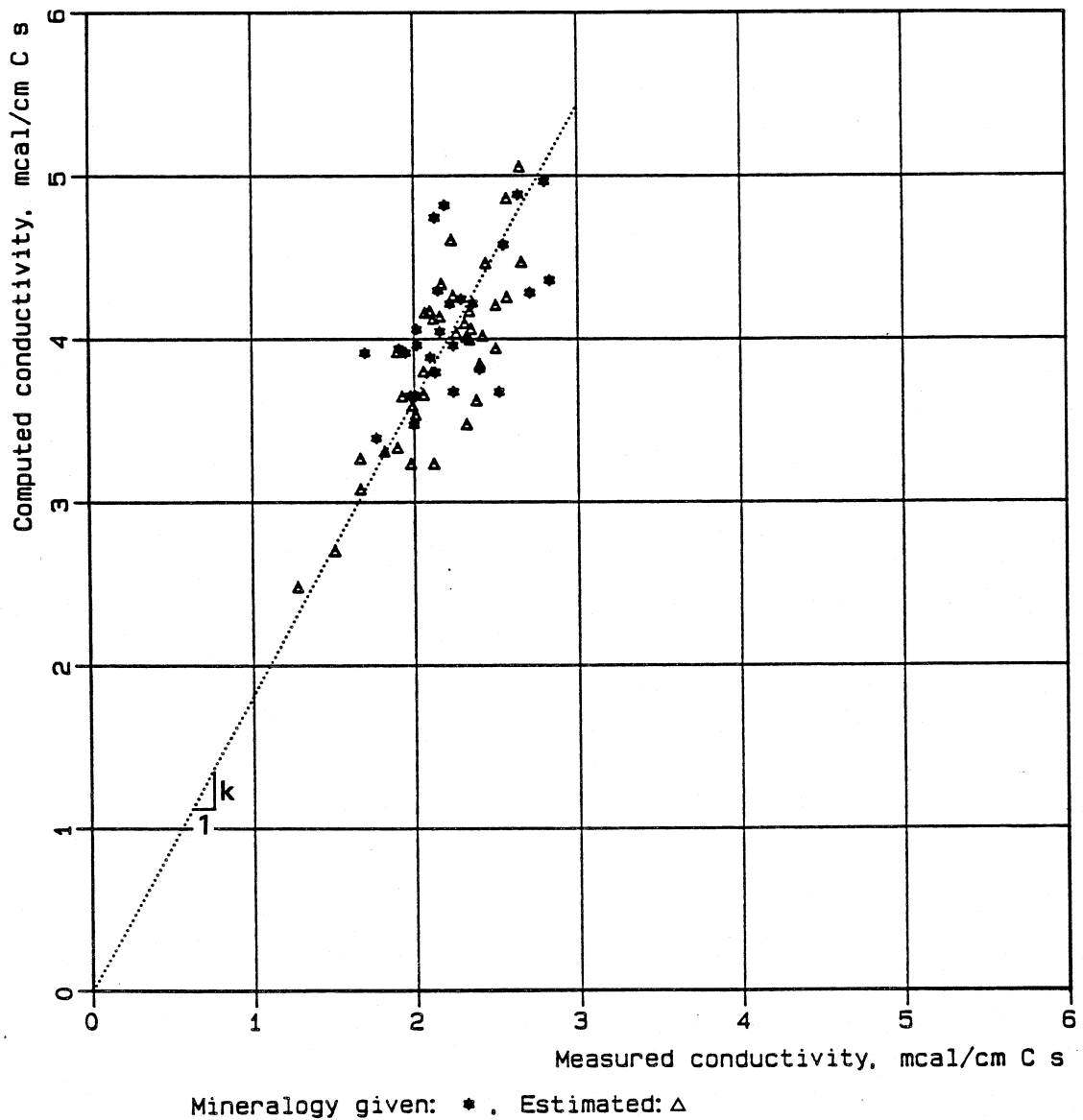


Fig. 6.3.1 Uncorrected measured thermal conductivity from on board measurements versus theoretical computed thermal conductivity. $k = 1.81$.

A straight least square regression line of the plot, forced through the origin of the diagram, is shown as a dotted line. The slope of this line gives a correlation factor of 1.81.

The correlation factor of 1.81 is in very good agreement with the findings presented in Appendix I of this chapter which indicate that all the measured thermal conductivity values on Quaternary samples given in IKU report 21.3405.00/02/85 (Sættem et al. 1985) should be multiplied by a factor of 1.799.

Sættem et al. (1985) suspected lithological variations to contribute to the relatively large scatter in thermal conductivity versus water content. In Fig. 6.3.2 below we have plotted measured thermal conductivity, corrected for saturation (and multiplied by a factor of 1.8), versus water content. Different symbols are used for different quartz content. Theoretical curves for thermal conductivity versus water content for a range of quartz content is superimposed. (Thermal conductivity of rest mineral composition is fixed to $5.98 \text{ mcal cm}^{-1} \text{ } ^\circ\text{C}^{-1} \text{ s}^{-1}$). Though there is a significant scatter, the trend clearly indicates a correlation between quartz content and thermal conductivity. The scatter in the plot of measured thermal conductivity versus water content (corrected to saturation) is furthermore shown to be largely explainable by a variation in quartz content in the mineral fraction between 10% and 50%. This is actually the range in quartz content for the bulk part of the analyzed samples.

Thermal conductivity adjusted to in situ conditions

In situ values of thermal conductivity are required in order to calculate heat flow. The significant deviations in conditions from in situ to the laboratory are temperature and pressure. The temperature effect is found to be negligible. In Sættem et al. (1985) also the effect of pressure release on the thermal conductivity of the individual components is found to be negligible.

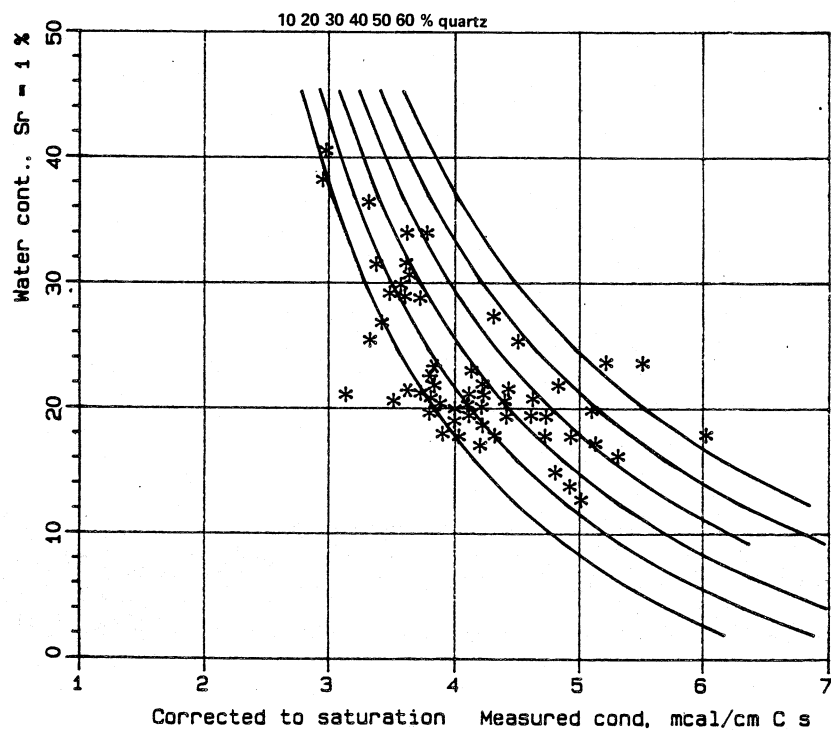
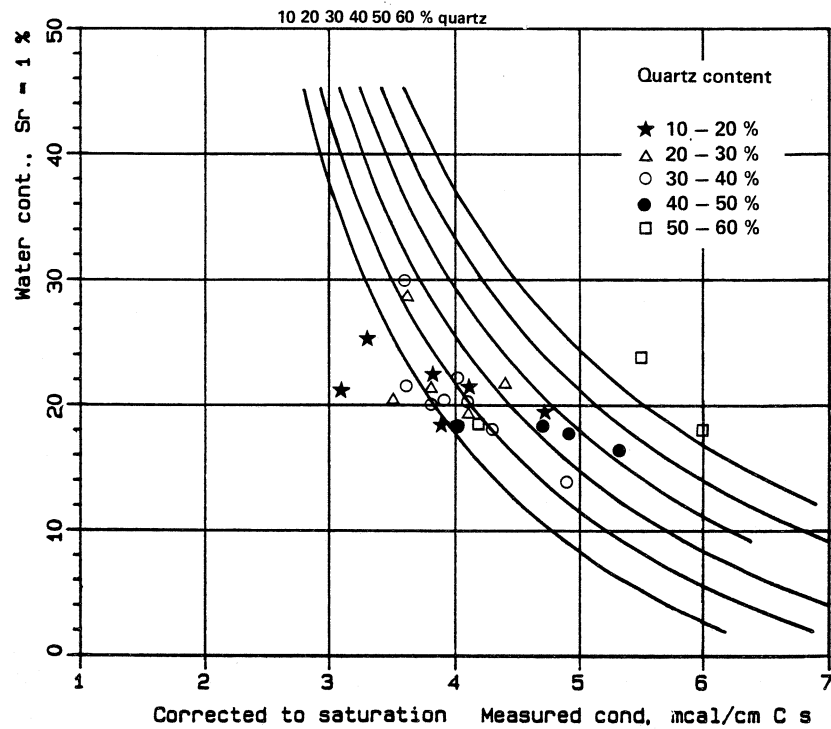


Fig. 6.3.2 Measured thermal conductivity corrected to saturation as a function of water content (wt %). A) Samples with known quartz content. B) General plot. Theoretical curves for a range of quartz content is superimposed.

However, due to the pressure release from downhole to the laboratory gas dissolved in the pore water may form isolated gas voids in the sample. The result is that the sample in the laboratory comprises mineral grains, water and gas voids, while it in situ comprises only grains and water. To arrive at in situ thermal conductivity it is therefore necessary to correct for the effect of these gas voids. The procedure for doing this is based on equations 1 to 5 above in this chapter and can be summarized as follows.

- The measured thermal conductivity is adjusted to saturation (eq. 3) based on the gas void fraction determined from the density data (eq. 4d).
- The conductivity of the mineral fraction is derived from this saturated conductivity by inverse application of the geometric mean equation (eq. 1).
- The in situ conductivity is determined by the geometric mean equation from this solid conductivity and the volumetric composition corresponding to the adjusted dry density (eq. 5b).

The results of this exercise is shown in Table 6.3.7. The shallowest data which are of no relevance for the down hole heat flow calculations is not included.

Fig. 6.3.3 shows both the final values from the laboratory analyses and the same values adjusted for in situ conditions. The conclusion that can be drawn from this figure is that adjustments due to gas expansion only influence significantly upon the thermal conductivity for the sites D85-N, D85-O and D85-V.

Table 6.3.7

Adjustment of thermal conductivity to in situ conditions (correction of effects due to gas voids forming in the samples under laboratory conditions.)
D-DENS (kg/m^3) = measured dry density, SAT (%) = water saturation in pore space, C-MEAS ($\text{mcal/cm}^{\circ}\text{Cs}$) = measured thermal conductivity, C-SAT ($\text{mcal/cm}^{\circ}\text{Cs}$) = thermal conductivity corrected for lack of water saturation, C-DENS (kg/m^3) = dry density corrected for gas voids in bulk sample under laboratory conditions, C-INSITU ($\text{mcal/cm}^{\circ}\text{Cs}$) = in situ thermal conductivity.

SITE	DEPTH	D-DENS	SAT	C-MEAS	C-SAT	C-DENS	C-INSITU
C85-A	8.53	1550.5	100.9	4.28	4.28	1550.5	4.28
	14.02	1704.9	101.8	4.21	4.21	1704.9	4.21
	18.65	1744.2	100.5	4.16	4.16	1744.2	4.16
	23.15	1733.0	99.7	4.23	4.24	1735.0	4.24
	23.36	1603.9	87.7	4.18	4.50	1688.1	4.78
C85-E	10.64	1710.5	100.8	3.78	3.78	1710.5	3.78
	21.55	1720.2	101.9	3.06	3.06	1720.2	3.06
	40.21	1741.7	98.1	3.87	3.91	1753.2	3.93
	40.37	1657.1	96.5	3.71	3.78	1679.6	3.83
	13.56	1915.6	91.0	4.63	4.81	1967.1	4.97
D85-H	10.08	1511.2	101.9	3.71	3.71	1511.2	3.71
	10.71	1482.9	103.6	3.58	3.58	1482.9	3.58
	14.64	1520.6	100.6	3.60	3.60	1520.6	3.60
	14.77	1498.1	99.3	3.62	3.64	1502.9	3.65
	14.83	1498.1	99.3	3.56	3.58	1502.9	3.59
D85-K	14.92	1465.6	99.4	3.62	3.63	1469.8	3.64
	19.63	1661.3	101.0	4.07	4.07	1661.3	4.07
	22.37	1707.7	101.3	3.64	3.64	1707.7	3.64
	22.46	1707.7	101.3	3.73	3.73	1707.7	3.73
	25.59	1693.9	100.9	3.82	3.82	1693.9	3.82
	25.63	1693.9	100.9	4.03	4.03	1693.9	4.03
	34.51	1745.6	100.3	4.03	4.03	1745.6	4.03
	43.98	1745.6	100.3	4.12	4.12	1745.6	4.12
	54.66	1796.6	96.7	3.91	3.97	1817.0	4.02
	64.74	1806.6	97.7	3.89	3.93	1820.3	3.96
	94.87	1750.8	93.6	3.89	4.02	1790.9	4.12
	12.08	1769.8	84.2	3.78	4.10	1871.4	4.36
	12.16	1769.8	84.2	4.21	4.57	1871.4	4.89
	21.89	1814.3	90.1	3.83	4.02	1874.9	4.17
	27.31	1756.6	84.0	3.51	3.82	1860.9	4.05
	30.26	1848.3	86.1	4.79	5.11	1932.9	5.42
	36.11	1748.9	77.0	4.50	5.09	1903.4	5.70
	39.66	1849.6	92.2	4.01	4.16	1896.2	4.28
	41.55	1742.5	81.1	4.18	4.62	1867.9	5.03
	41.75	1800.0	81.0	4.25	4.67	1921.7	5.06
	42.07	1770.8	78.2	3.64	4.07	1914.4	4.43
	9.85	1874.5	90.1	5.04	5.27	1932.9	5.49
	22.14	1816.6	86.6	4.00	4.27	1899.8	4.49
	25.41	1606.1	94.8	3.19	3.29	1641.0	3.35
	30.23	1966.3	93.4	4.75	4.88	2002.5	4.99
	35.16	1987.5	88.1	4.77	5.00	2051.8	5.21
	42.11	1818.2	67.9	5.09	5.98	2031.0	7.07
	14.77	1684.4	99.9	3.83	3.84	1685.3	3.84
	20.10	1699.7	97.3	4.32	4.39	1717.1	4.44
	24.58	1716.4	97.1	4.03	4.10	1735.0	4.14
	24.84	1742.0	95.3	3.44	3.53	1771.9	3.58
	25.46	1701.1	97.0	4.32	4.39	1720.1	4.45
	29.63	1823.7	88.2	4.59	4.86	1896.2	5.10
	40.12	1776.6	94.6	4.54	4.66	1810.4	4.77
	44.50	1699.8	94.5	4.63	4.77	1735.0	4.89
	44.59	1645.7	80.5	4.88	5.48	1781.4	6.12
	45.94	1652.5	76.7	4.50	5.17	1817.0	5.87

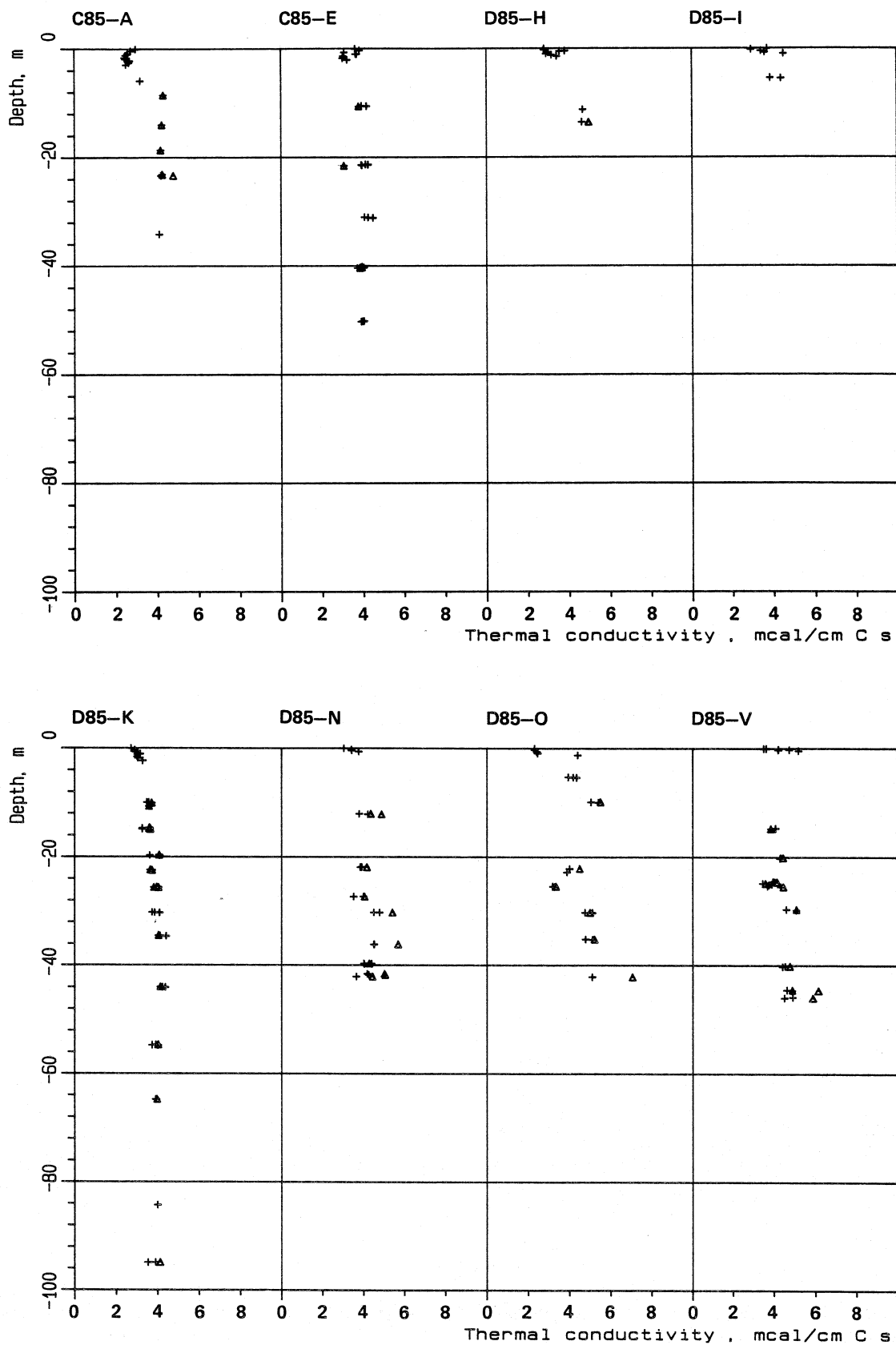


Fig. 6.3.3 Thermal conductivity versus depth. + - laboratory conditions, Δ adjusted to in situ conditions.

6.3.4 Comparative measurements of thermal conductivity

Measurements and results

A selection of Quaternary samples have been subjected to thermal conductivity measurements at two independent laboratories, the Chalmers Technical University in Gothenburg, Sweden and the Geophysical Laboratory at the University of Aarhus, Denmark.

All the measurements are performed by use of a standard needle probe with the following specifications:

Chalmers:

probe length 40.0 mm
probe diameter 0.9 mm

University of Aarhus:

probe length 50 mm and 80 mm
probe diameter 1.5 mm

The results of these measurements are presented in Table 6.3.8 together with the thermal conductivities given by Sættem et al. (1985) from corresponding sites and levels.

All conductivity values are given in $\text{mcal } ^\circ\text{C}^{-1} \text{ cm}^{-1} \text{ s}^{-1}$ ($1 \text{ mcal } ^\circ\text{C}^{-1} \text{ cm}^{-1} \text{ s}^{-1} = 0.4186 \text{ Wm}^{-1} \text{ K}^{-1}$).

Discussion

The results from the test sequence in Table 6.3.8 are found to support the correction factor reported by Zielinski (1986a). The best correlation to the reported factor of 1.799 is found in the data from Aarhus which suggests a correction factor of 1.81 (st.dev. 0.19). The results from Chalmers Technical University give a slightly lower value for this factor (1.70 with st.dev. 0.11). It is, however, suggested that these values may be slightly too low due to undersaturation of the samples (Wilén pers. commun. 1986).

The following abbreviations are used:

Sættem et al. (1985):

K_T - thermal conductivity from measurements on board the M/S Bucentaur

$\alpha \cdot K_T$ - K_T multiplied by a correction factor , $\alpha = 1.799$

University of Aarhus:

K_h^1 - horizontal thermal conductivity

K_v^1 - vertical thermal conductivity

$A = \frac{K_h^1}{K_v^1}$ - coefficient of anisotropy

$\alpha^1 = \frac{K_h^1}{K_T}$ - correlation coefficient between results from the University of Aarhus and results from on board measurements.

Chalmers Technical University:

K^2 - thermal conductivity

$\alpha^2 = \frac{K^2}{K_T}$ - correlation coefficient between results from the Chalmers Technical University and results from on board measurements.

Table 6.3.8 Comparison of different thermal conductivity measurements on Quaternary samples from the Barents Sea.

Site	Depth(m)	Settem et al. 1985		Geophysical Laboratory, University of Aarhus				Department of geology Chalmers Technical University	
		K_T (mcal °C ⁻¹ cm ⁻¹ s ⁻¹)	$\alpha \cdot K_T$ (mcal °C ⁻¹ cm ⁻¹ s ⁻¹)	K_h^1 (mcal °C ⁻¹ cm ⁻¹ s ⁻¹)	K_v^1 (mcal °C ⁻¹ cm ⁻¹ s ⁻¹)	$A = \frac{K_h^1}{K_v^1}$	$\alpha^1 = \frac{K_h^1}{K_T}$	K^2 (mcal °C ⁻¹ cm ⁻¹ s ⁻¹)	$\alpha^2 = \frac{K^2}{K_T}$
C85-A	8.42-8.53	2.13	3.83	3.96	3.27	1.21	1.86	3.6	1.69
	13.94-14.02	2.30	4.14	4.01			1.74		
	18.25-18.50	2.20	3.96	3.88	2.67	1.45	1.76	3.4	1.66
	23.05-23.15	2.33	4.19	4.16	3.35	1.24	1.79		
C85-E	1.74-1.94	1.67	3.00	3.16	3.08	1.02	1.89	3.1	1.86
	21.75-21.93	1.70	3.05	4.04			2.38		
	40.10-40.24	2.17	3.90	4.02	3.53	1.14	1.85	3.6	1.66
D85-H	0.36-0.46	1.95	3.51	3.49	2.68	1.30	1.79		
	0.46-0.56	2.05	3.69	3.33	3.20	1.04	1.62	3.2	1.56
	1.34-1.44	1.85	3.33	3.51	3.20	1.09			
D85-I	5.20-5.51	2.13	3.83	4.51	4.08	1.11	2.12	(3.4)	1.60
D85-K	0.00-0.10	1.55	2.79	2.82	2.63	1.08	1.82		
	1.46-1.56	1.67	3.00	3.11	3.20	0.97	1.86	3.0	1.78
	2.26-2.36	1.80	3.24	3.27	2.80	1.17	1.82		
	9.94-10.06	1.93	3.47	3.68	3.03	1.22	1.91	3.3	1.71
	10.46-10.60	2.00	3.60	3.44			1.72		
	14.61-14.74	1.91	3.44	3.52	3.13	1.12	1.84	3.1	1.62
	14.80-14.90	1.98	3.56	3.44			1.74		
	19.50-19.59	2.26	4.07	4.10			1.81	3.6	1.82
	22.34-22.44	2.02	3.03	3.68	3.89	0.97	1.82		
	25.27-25.53	2.19	3.94	3.94			1.80		
	54.45-54.67	2.17	3.90	4.48	3.39	1.33	2.06	4.0	1.84
	94.79-94.89	2.16	3.89	2.25			1.50		
D85-N	0.43-0.53	1.90	3.42	3.80	3.65	1.04	2.00	3.4	1.79
	12.00-12.08	2.10	3.78	3.56			1.70		
	21.81-21.92	2.15	3.87	3.32	3.44	1.02	1.54	3.2	1.52
	27.12-27.23	2.20	3.96	4.20	3.75	1.12	1.91		
	30.12-30.20	2.50	4.50	3.66	3.34	1.09	1.46		
	39.49-39.57	2.30	4.14	3.89	4.49	0.89	1.69		
	41.78-41.94	2.36	4.25	4.15	3.77	1.12	1.76	4.1	1.74
D85-O	0.30-0.42	1.30	2.34	2.56			1.97		
	1.05-1.15	-	-	4.09	3.94	1.04	1.86	3.7	-
	5.07-5.22	2.20	3.96	5.04	1.44	1.47	2.29		
	9.89-9.97	3.00	5.40	5.17	4.27	1.21	1.72	4.3	1.43
	25.22-25.33	1.80	3.23	3.15			1.75		
D85-V	19.97-20.05	-	-	3.76	3.77	0.99	1.57	3.4	-
	24.45-24.52	2.25	4.05	3.49			1.55		
	24.74-25.00	2.04	3.67	3.65	3.05	1.21	1.66	2.8	1.37
	25.26-25.45	2.04	3.67	3.75			1.84		
	29.40-29.60	2.55	4.59	4.79	4.56	1.05	1.88	4.8	1.88
	40.12-40.25	2.45	4.41	4.16			1.70		

Anisotropy of thermal conductivity

Numerous papers and reports describe anisotropy in the thermal conductivity of sedimentary rocks. The possible presence of this effect in Quaternary sediments has, however, generally been neglected (Balling pers. commun. 1986).

At the University of Aarhus a number of the tested samples have been measured with respect to thermal conductivity both in the horizontal and the vertical direction. Though the statistical base is somewhat scarce the results clearly indicate that anisotropy in thermal conductivity is present in the stiff, consolidated Quaternary samples from the Barents Sea. This anisotropy, A, defined as horizontal-, divided by vertical thermal conductivity is presented in Table 6.3.8. The mean value of A is 1.13 (st.dev. 0.14).

All the thermal conductivity values reported from the on board measurements are measured by the needle probe method by either pushing the needle axially into the core samples or with an angle that may deviate in the order of $50-60^{\circ}$ from the axis of the cores. All the cores are obtained with an vertical axis. As the needle probe measures thermal conductivity perpendicular to the needle axis it can be seen from a simple geometrical sketch that the results from the on board measurements are dominated by the horizontal thermal conductivity component.

Our conclusion is therefore that the thermal conductivity measurements as they are given in Table 6.3.1 may be approximately 10% higher than the vertical thermal conductivity of the tested samples, and we suggest that for calculation of vertical heat flow one should correct the measured values by a factor of 0.91% ($0.91=1/1.1$).

6.4 Correction for perturbations in measured temperatures due to the late Quaternary history

6.4.1 Summary

Possible perturbations in the measured data due to topographic relief and the glacial history are investigated.

The effects from topographic relief and lateral variations in Quaternary sediment thickness close to the heat flow sites are found to be negligible.

The climatic effect is treated individually for each site. Based on published information and data gathered within the Barents Sea Mapping Program we have deduced two alternative glacial history models causing minimum and maximum transient effects on the measured temperatures. These models are used as input to computer calculations of maximum and minimum perturbations.

The maximum and minimum transient effects on the measured gradients are summarized in Table 6.4.3.

As was suggested by Sættem et al. 1985, this effect varies from site to site. Maximum effect is found to occur on site C85-A where the disturbance on the present day gradient is between

$$-3.6 \cdot 10^{-3} \text{ }^{\circ}\text{Cm}^{-1} \text{ and } -6.2 \cdot 10^{-3} \text{ }^{\circ}\text{Cm}^{-1}$$

depending on the temperature history models. The lowest effect is found on site D85-0 where the maximum glaciation effect implies a disturbance of

$$-1.1 \cdot 10^{-3} \text{ }^{\circ}\text{Cm}^{-1}.$$

6.4.2 Objective and limitations

Supracrustal geological processes and climatic changes will cause variations in the boundary conditions for heat flow through the earth's crust.

The transient effect on the present vertical heat flow from previous climatic changes will vary with depth as a function of the time span since the events occurred and of the actual temperature amplitudes involved.

Therefore, in order to be able to compare the heat flow data collected in the present project (15-70 m below the sea bed) with similar data from exploration wells (1000-4500 m below the sea bed) it is necessary to apply a correction model which implies none or negligible corrections at an average well depth of 2000-3000 m.

Werner (1985) relates a time span of 200 000 years to a depth range of 10 km. Nielsen & Balling (1985) show an example where the effects from a temperature change of 10°C 10 000 years ago are totally damped at 2000 m. With the temperature amplitudes and the combinations of different temperature events through time that are used in our models, the deduced corrections to be applied at 2000-3000 m can be neglected.

Conclusively, the shallow heat flow data (15-70 m) should be comparable to exploration well data if they are corrected for late Quaternary climatic changes according to the models below.

In our study we have chosen a time span of 245 000 years for the climatic correction models. This span is chosen because 245 000 y. BP is regarded as the best estimate of the onset of the warm Holsteinian interglacial before the Saalian glacial period (Balling, pers. commun. 1986).

The temperature history models presented represent a refinement of the models outlined by Sættem et al. (1985) and in the following chapters a quantification of the corresponding correction to the measured heat flow values is included.

Correction for isostatic crustal movements is not considered in the present study. A discussion of the arguments for this is given by Sættem et al. (1985).

The effect from erosion and deposition within the same time span as for the climatic history should also be considered in order to have a more complete reduction of shallow data for comparison with well data. However, such events are less established than the temperature history at our sites. Furthermore, as several episodes of erosion and deposition may have occurred, the summarised effect of these processes may largely be cancelled. The effect from the topographic relief and lateral variations in Quaternary sediment thickness is found to be negligible for all the heat flow sites. Thus only corrections for climatic variations are included in the present study.

6.4.3 Status of knowledge about the glacial history of the area

Introduction

This Chapter provides a rather extensive discussion of the status regarding knowledge about the glacial history of the Barents Sea. Based on this discussion we have suggested two alternative temperature history models from the area implying a maximum (model I) and a minimum (model II) effect on heat flow measured in shallow boreholes (Chapter 6.4.4 and chapter 6.4.5).

The Quaternary is characterised by large changes in climate with glaciations and interglaciations.

Investigations in the north Atlantic (Shackleton et al. 1982) indicate that extensive ice sheets existed in the northern hemisphere about 2.4 mill y BP. From then on there have been alterations between glacial and interglacial periods. The exact time for the first glaciation may, however, be disputed.

Published knowledge

The knowledge of how the last ice age (the Weichselian) developed and receded in the Barents Sea are scarce compared with the western and southern part of the European ice sheet. This relation is even more pronounced going back in time to the last interglacial (the Eemian) and the preceding glacial (the Saalian) and interglacial (the Holsteinian).

For the present purpose the need for a detailed glacial history decreases with increasing age, and we have done the following simplifications:

For the interstadials Holsteinian and Eemian we have assumed that the climate was as today (Holsteinian in Europe was actually slightly cooler and Eemian slightly warmer). The duration of Eemian is in this context fixed between 115 000-130 000 B.P. (Sejrup in press) and we assume that the Holsteinian interglacial terminated around 185 000 BP (Denton & Hughes 1983).

The second last glacial period in our model, the Saalian had a more extensive ice cover than the Weichselian (Nilson 1983) and there are no stratigraphical work done in Scandinavia which are of any help in reconstructing the conditions in the Barents Sea. We have therefore assumed that the Barents Sea was covered by an ice sheet in large parts of this period. As a crude simplification we have not taken into account any interstadials in this period.

Our tentative relation between period nomenclature and age for the present purpose can thus be summarized as follows:

Weichselian ice age	10 000 - 115 000 BP
Eemian interglacial	115 000 - 130 000 BP
Saalian ice age	130 000 - 185 000 BP
Holsteinian interglacial	185 000 - 245 000 BP

Complicating factors for the glacial history of the area may be:

- The ice cover on Svalbard and adjacent areas was possibly not in phase with the southern and western margin of the European ice sheet. (Boulton 1979; Boulton et al. 1982; Landvik pers. commun. 1986).
- The ice sheet in the Barents Sea will be depending on ice feeding from as much as three possible different glacier domes. (Denton et al. 1981): A) the ice sheet draining northwards through Finnmark (sev. authors). B) Southerly drainage from a postulated ice dome east of Svalbard (Salvigsen 1981, Salvigsen & Nydal 1981,

Häggbom 1982, Boulton et al. 1982, Elverhøi & Solheim 1983, Solheim & Kristoffersen 1984). C) Feeding from a postulated culmination to the east (the Kara ice dome) (Schytt et al. 1968; Grosswald 1980; Denton & Hughes 1981).

The present published knowledge of the glacial history of the Barents Sea, which is based mainly on recent marine geological data and shore line investigations from the Svalbard archipelago, can briefly be summarised as done by Vorren & Kristoffersen (1986):

1. Grounded glaciers reached the shelf edge in front of Bjørnøyrenna at least four times during the late Cenozoic (Solheim & Kristoffersen 1984).
2. The northern Barents Sea was covered by an ice sheet in Late Weichselian (Salvigsen 1981; Salvigsen & Nydal 1981).
3. Spitsbergenbanken was covered by an ice sheet, possibly as late as the Late Weichselian (Elverhøi & Solheim 1983; Solheim & Kristoffersen 1984).

Vorren & Kristoffersen (1986) describes undated ridges in the southwestern Barents Sea which they argue to represent the maximum of a Weichselian ice sheet.

In other words there does not exist an unambiguous model of glaciation for the last 200 000 years BP in the study area. To bypass this problem we are largely bound to use models of glaciations from other places. This is generally the same approach as presented by Sættem et al. (1985). However, in the present work we pay more specific attention to results from Svalbard and the Norwegian mainland.

Western Norway. In this area there are many papers summing up the Weichselian stratigraphy (a.o. Mangerud 1981; Andersen et al. 1981). For the coastal areas Sejrup (in press) has summarised and correlated the existing works. His juxtaposition is the northernmost well-funded model of Weichselian stratigraphy and we use this model as a support in the Barents Sea. His model consists of four interstadials and five stadials in the timespan between Eem and Allerød. He regards this as a minimum number of interstadials.

The duration of the interstadials are in principle uncertain and in our models we use Sejrup's inferred time estimate of the different events (Fig. 6.4.2 and Fig. 6.4.3).

Troms. In the coastal area the ice sheet has transgressed sorted sediments dated to 27 000 BP (Andreassen et al. 1984). Vorren et al. (1983) have shown that the ice sheet occupied the outer part of the continental shelf after about 36 000 BP, and Rokoengen et al. (1979) indicated that the ice sheet may have reached the shelf edge as late as about 13 300 BP.

Finnmark. In the southwestern part of Finnmarksvidda several ice free periods are recognised, but none of them are dated (Olsen & Hamborg 1983, 1984).

Svalbard. The investigations by Salvigsen (1981) and Salvigsen & Nydal (1981) show that Svalbard was completely covered by an ice-sheet at least once, probably in Early Weichselian. The Late Weichselian glaciation was limited and local in the western part, while the eastern part of the archipelago was covered by an ice sheet.

In the scope of this presentation Boulton et al. 1982 are largely in agreement with Salvigsen & Nydal (1981). Boulton et al. (1982) is modified and presented in Figs. 6.4.2 and 6.4.3.

Barents Sea. Solheim & Kristoffersen (1984) regard the pre-Late Weichselian glacial chronology as completely open due to lack of stratigraphical tie points. However, presenting arguments for a maximum age of a few hundred thousand years for a moraine ridge complex in the western part of Bjørnøyarena, they suggest that the last ice sheet that extended westwards into the deeper parts of Bjørnøyarena was of Saalian or Early Weichselian age.

This is in agreement with Salvigsen & Nydal (1981) and Landvik (pers. commun.) who find that the most extensive ice cover on Svalbard occurred in early Weichselian, while Troitsky et al. (1979) and Miller (1982) suggest a Saalian age. For the Late Weichselian Elverhøi & Solheim (1983) find it most probable that the ice sheet was limited to the bank areas (present water depth 300-400 m) and thus forming a calving bay in Bjørnøyarena.

Still they are open for the possibility that also a Late Weichselian ice sheet extended westwards to form parts of the ridge complex in the western part of Bjørnøyareenna.

Sparker data collected within the IKU Barents Sea Mapping Program shows evidences of a wedge shaped (in NW-SE cross section) deposit north of Nordkapp. The northwestern limitation of this deposit subparallels the present 300 m water depth contour on the map in Fig. 6.4.1 and could tentatively be correlated with Elverhøi & Solheim's (1983) ice margin at 300-400 m water depth on the northern flank of Bjørnøyareenna, thus supporting the Bjørnøyareenna calving bay theory.

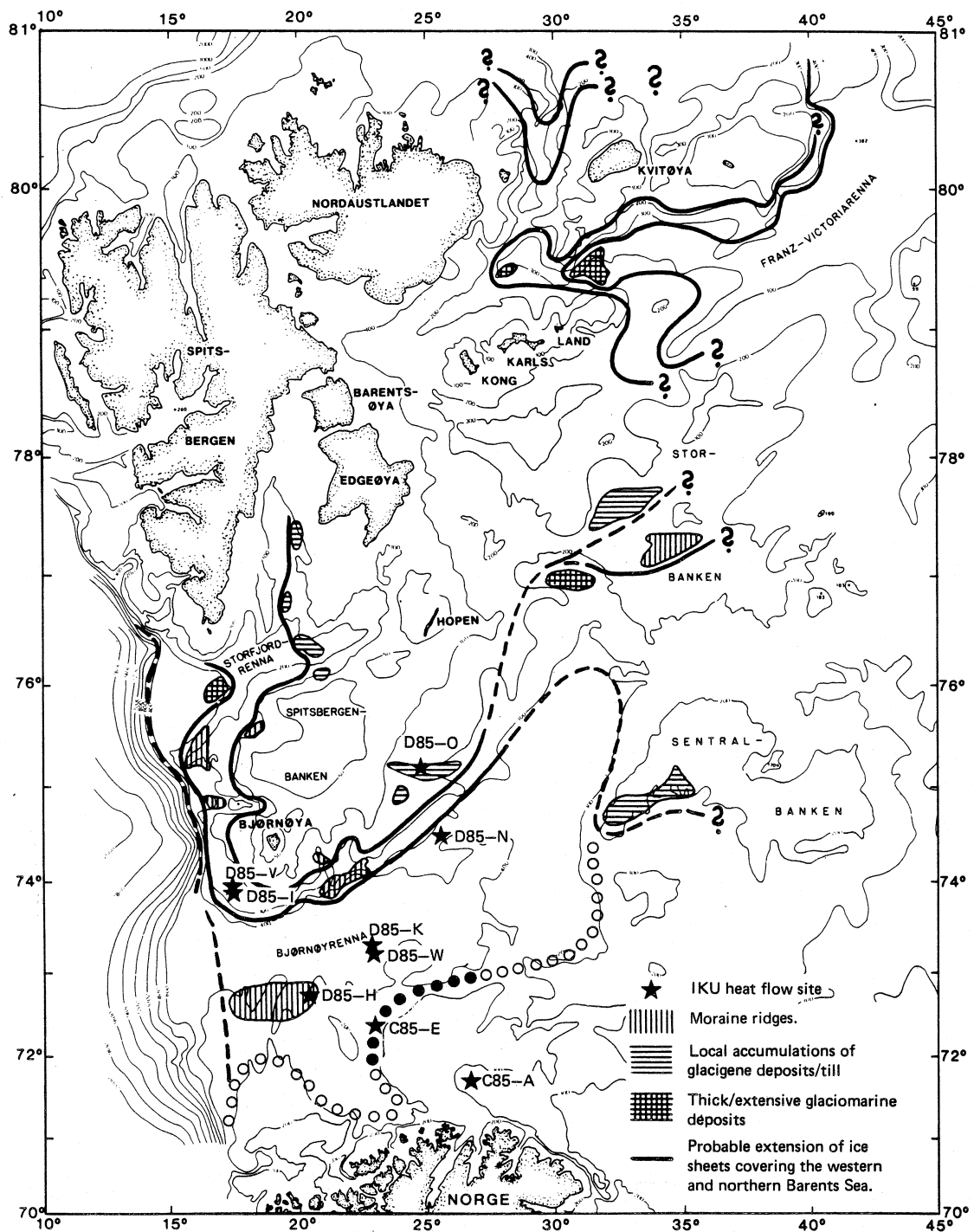


Fig. 6.4.1 Bathymetric map showing features related to former ice margins in the Barents Sea. Modified after Elverhøi and Solheim (1983). The heavy, dotted line shows the approximate western termination of a wedge shaped sediment thickening interpreted from IKU sparker data. Open circles indicate a tentative ice sheet margin related to this feature.

Construction of glaciation models

As can be seen from the previous discussion the geological data makes several models of glaciation possible.

In this work we are interested in the "coldest" and "warmest" probable temperature history.

We have therefore reconstructed what we regard as the probable maximum and minimum glacial influence in the area. These reconstructions are respectively combined with ice thicknesses which give the "warmest" and "coldest" condition at the base of the glacier.

These models are based on the following strong simplifications for the positions of the ice margin:

- 1) The ice-sheet was non existent or had retreated to the mainland.
- 2) The ice-margin was situated at 300-400 m present water depth forming a calving bay in Bjørnøyareenna.
- 3) The ice-sheet was calving at the continental margin.

I Maximum glaciation

Glaciers extending to the shelf edge occurred during Saale and Early Weichselian. In addition there were two periods when ice sheets grounded on the shallow banks in the Middle and Late Weichselian.

II Minimum glaciation

Saalian glaciation with an ice-sheet extending to the shelf edge. Early, Middle and Late Weichselian ice sheets grounded on the shallow banks and did not extend to the shelf edge.

Temperatures underneath the Barents Sea ice sheets

There are several opinions with regard to gradients and thicknesses of a Barents Sea ice sheet. In large there are two main temperature conditions to be considered, namely frozen (T_c) and wet-based (T_m) glaciers (cold and temperate glaciers).

For wet based glaciers there are grossly two main alternative approaches when calculating surface profiles of an ice sheet: A low profile ice sheet is a consequence of the model by Boulton & Jones (1979) which assumes that 90% of the movement of a glacier is due to deformation in the underlying sediments. The profile of an ice sheet will thus mainly depend on the shear strength of the underlying confined mixture of water and sediments.

The other model for temperate glaciers is based in internal flow of the ice (Paterson 1981) and ice sheet profiles given by Hughes (1981, Fig. 5.8). The same profiles are used in the reconstruction by Hughes et al. (1981) of a Late Weichselian ice sheet in the northern hemisphere.

Our model II, representing a "warm" ice, is based on a low profile, wet-based ice sheet. We have assumed a shear strength of 20kPa (0.2 bar) in the sediments just underneath the ice sheet (Boulton & Jones 1979, Fig. 5).

In cold based ice sheets the deformation in the sole of the glacier is supposed to decrease strongly. Most of the movement will take part within the ice body. This is expected to give steeper gradients of the ice sheet surface and a greater maximum ice sheet thickness. According to the reconstruction of the maximum Late Weichselian ice sheet by Hughes et al. (1981), the temperature below the maximum Barents Sea glacier would be below the pressure melting point, T_m , for all our heat flow sites except from C85-A and C85-E. Though Orheim (pers. commun. 1985) tends to assume that most of the large Barents Sea ice sheet (sheets) had temperatures equal to T_m .

In our model I, for a "cold" ice we have chosen the steepest probable ice profile by Hughes (1981, 1 in Fig. 5.8). We have furthermore chosen $T_c = T_m$ as an average temperature condition for the lifetime of a cold based ice sheet.

In a closed system underneath an ice sheet the pressure melting point temperature, T_m , will follow the phase equilibrium function for ice and fresh water. As a function of pressure this equilibrium temperature decreases with about $7 \cdot 10^{-3}$ K/atm or approximately 0.7 K/1000 m ice thickness (Sugden & John 1976).

From our alternative assumptions about ice margin positions and ice sheet profiles we have arrived at alternative ice sheet thicknesses for each heat flow site. The corresponding temperatures are presented in Table 6.4.1 and Table 6.4.2.

Table 6.4.1 Temperature ($T_c = T_m$), underneath an ice sheet at the IKU heat flow sites assuming the "cold" glaciation model (Model I)

- 1) The ice sheet terminates at the continental margin.
- 2) The ice sheet is restricted to the shallow banks.

Site	Average temperature under a cold based ice. Model I ($^{\circ}\text{C}$)	
	1) ($^{\circ}\text{C}$)	2) ($^{\circ}\text{C}$)
C85-A	-1.8	-1.3
C85-E	-1.6	-0.8
D85-H	-1.4	
D85-W	-1.5	
D85-K	-1.5	
D85-I	-0.9	-0.8
D85-V	-0.9	-0.8
D85-N	-1.8	-0.8
D85-O	-1.8	-1.2

Table 6.4.2 Pressure melting point temperature, T_m , underneath an ice sheet for the IKU heat flow sites assuming the warm glaciation model (Model II).

- 1) The ice-sheet terminates at the continental edge.
- 2) The ice-sheet is restricted to the shallow banks.

Site	Pressure melting point temperature. Model II.	
	1) (°C)	2) (°C)
C85-A	-1.0	-1.3
C85-E	-0.8	-0.8
C85-H	-0.7	
D85-W	-0.8	
D85-K	-0.8	
D85-I	-0.5	-0.8
D85-V	-0.5	-0.8
D85-N	-1.0	-0.8
D85-O	-1.0	-1.2

Sea water temperatures

For a simplified temperature model it may be assumed that the annual mean sea bottom temperature was as today during the whole Holocene period and in all periods between the glaciations. This temperature has been obtained by extrapolation of temperature gradients from the boreholes. The effect of variations in thermal conductivity with depth is accounted for.

The shelf area in front of a marine ice sheet will normally have abundant icebergs or be covered by an ice sheet or drifting ice. Under such

conditions it can be assumed that the sea water even down to 400 m depth was between the freezing point for sea water (about -2.1°C) and the freezing point for fresh water (about 0°C). The actual temperature would vary with the salinity of the water and with depth.

Hald (pers. commun. 1985) tends to believe that most likely the sea water temperatures for a setting as described above would be between 0°C and -1°C . The exceptions would be if salt brines were developed as a result of increased salinity due to freezing of ice in the upper part of the water column - a process that could result in temperatures near -2.1°C close to the sea bed.

A temperature of -0.5°C may probably be chosen as an average value under such conditions, and if this approximation also is valid for site D85-0 it implies that the sea water temperature at this site has been roughly the same since the deglaciation as it is today.

6.4.4 Temperature history models

Based on the discussions in Chapter 6.4.3 we have arrived at two temperature history models which we believe give realistic boundaries for maximum (Model I) and minimum (Model II) perturbation on heat flow data measured in the upper 10-100 m below the sea bed (Fig. 6.4.2 and 6.4.3).

The corresponding correction to apply on the measured heat flow for each individual site is calculated in Chapter 6.4.5.

**MAXIMUM GLACIATION MODEL
EXTENSIVELY COLD BASED ICE SHEET**

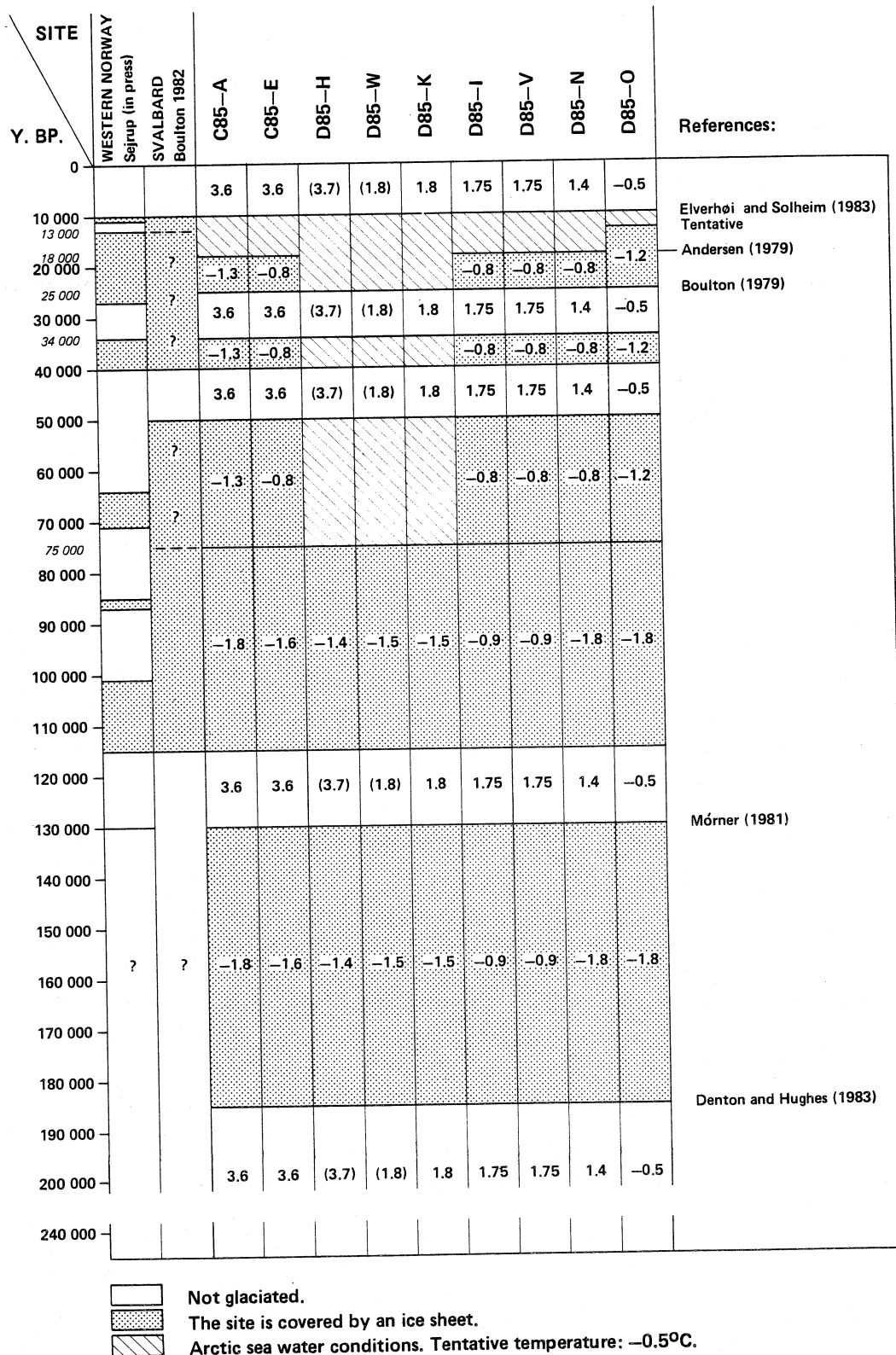


Fig. 6.4.2 Temperature history model I. Extensive cold based ice sheets combined with a maximum glaciation model.

**MINIMUM GLACIATION MODEL
EXTENSIVELY WET BASED ICE SHEET**

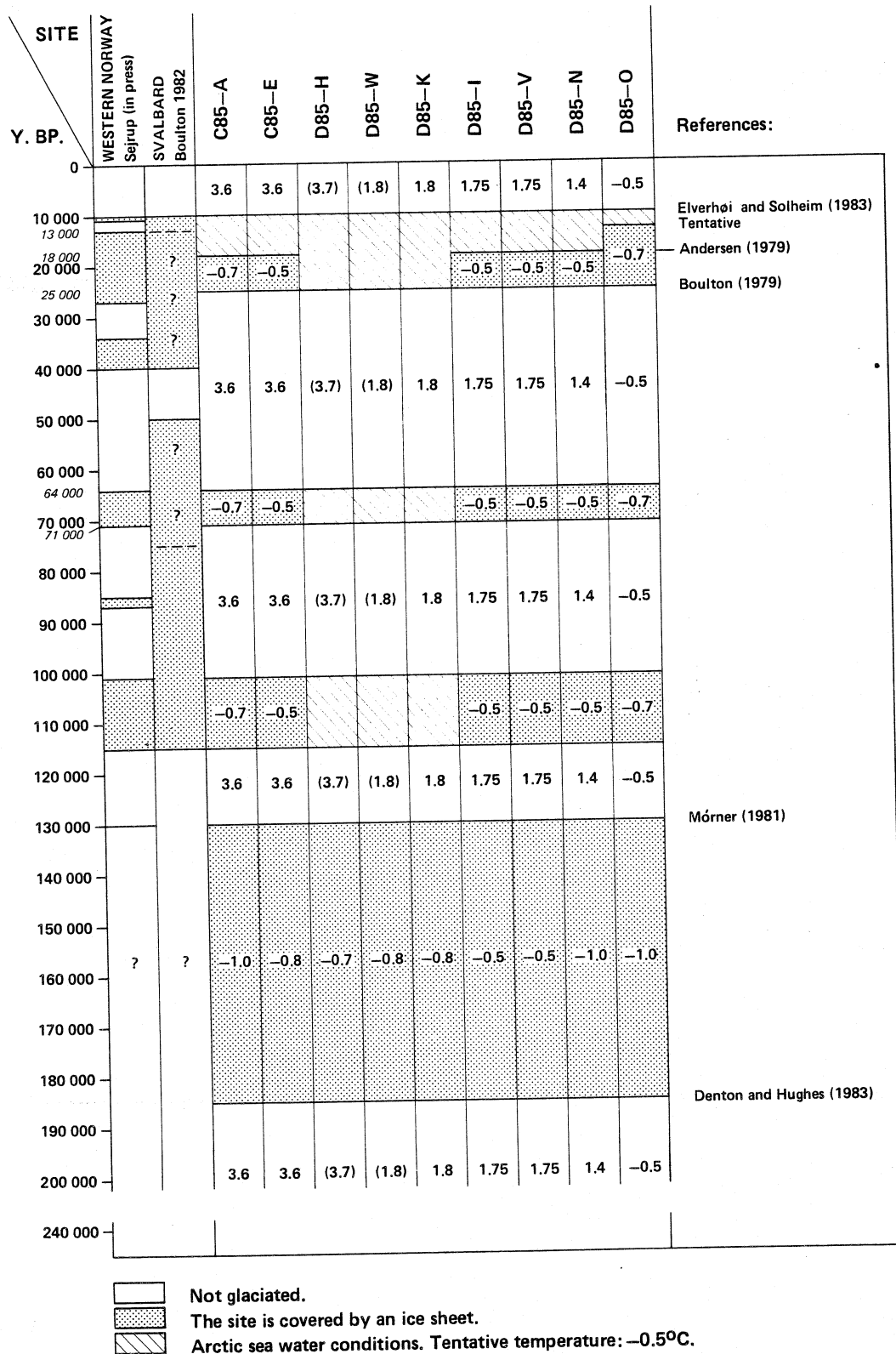


Fig. 6.4.3 Temperature history model II. Extensive wet based ice sheets combined with minimum glaciation models.

6.4.5 Calculation of climatic perturbations

Method and assumptions

The calculation of the theoretical transient effect upon the measured temperatures is calculated at the Geophysical Laboratory, Department of Geology, at the University of Aarhus in Denmark. This work follows a scheme earlier published by this institution (Balling et al. 1981) and the mathematical part will not be further discussed here.

A stepped temperature versus time-function is applied. Input to this function is taken from the temperature history in Fig. 6.4.2 and Fig. 6.4.3.

The present day surface temperature is chosen as reference level. This temperature is deduced from an extrapolation of the depth/temperature profile obtained in the boreholes, taking into account the vertical variations in thermal conductivity.

The calculations are based on a theoretical half space with a constant thermal diffusivity. The thermal diffusivity is proportional to the thermal conductivity and inversely proportional to the specific heat per volume. For the Barents Sea samples, the average value for thermal diffusivity will be about $0.4 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ to $0.6 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$. Calculations are done using both these alternatives.

Table 6.4.3 summarizes the calculated disturbances of the temperature gradients for each site. To arrive at maximum and minimum correction alternatives, temperature history model I is combined by a thermal diffusivity of $0.4 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$, whereas temperature history model II is combined with a thermal diffusivity of $0.6 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$. This table thus provides extreme boundaries for the details presented in Table 6.4.4.

Table 6.4.3 Maximum and minimum disturbance based on temperature history models I and II.

Site	Depth (m)	Temperature gradient disturbance ($^{\circ}\text{Cm}^{-1}$)	
		Temperature history model I. Thermal diffusivity $0.4 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ ($10^{-3} \text{ }^{\circ}\text{C m}^{-1}$)	Temperature history model II. Thermal diffusivity $0.6 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ ($10^{-3} \text{ }^{\circ}\text{C m}^{-1}$)
C85-A	0-50	-6.2	-3.6
C85-E	"	-5.9	-3.5
D85-H	"	-5.9	-3.6
D85-I	"	-3.2	-1.9
D85-K	"	-3.5	-1.9
D85-N	"	-3.2	-1.6
D85-O	"	-1.1	-0.1
D85-V	"	-3.2	-1.9
D85-W	"	-3.5	-1.9

Table 6.4.4 a Calculated disturbance of measured temperatures and temperature gradients for site C85-A.

A: Thermal diffusivity of underground: $0.6 \times 10^{-6} \text{ m}^2/\text{s}$.

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0051	0.000	-.0036
5.0	-.025	-.0051	-.018	-.0036
10.0	-.051	-.0051	-.036	-.0036
15.0	-.076	-.0051	-.054	-.0036
20.0	-.102	-.0051	-.071	-.0036
25.0	-.127	-.0051	-.089	-.0036
30.0	-.152	-.0051	-.107	-.0036
35.0	-.178	-.0051	-.125	-.0036
40.0	-.203	-.0051	-.143	-.0036
45.0	-.229	-.0051	-.161	-.0036
50.0	-.254	-.0051	-.178	-.0036
55.0	-.279	-.0051	-.196	-.0036
60.0	-.305	-.0051	-.214	-.0036
65.0	-.330	-.0051	-.232	-.0036
70.0	-.355	-.0051	-.250	-.0035
75.0	-.380	-.0050	-.267	-.0035
80.0	-.406	-.0050	-.285	-.0035
85.0	-.431	-.0050	-.303	-.0035
90.0	-.456	-.0050	-.320	-.0035
95.0	-.481	-.0050	-.338	-.0035
100.0	-.506	-.0050	-.356	-.0035

B: Thermal diffusivity of the underground: $0.4 \times 10^{-6} \text{ m}^2/\text{s}$.

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0062	0.000	-.0044
5.0	-.031	-.0062	-.022	-.0044
10.0	-.062	-.0062	-.044	-.0044
15.0	-.093	-.0062	-.066	-.0044
20.0	-.124	-.0062	-.088	-.0044
25.0	-.156	-.0062	-.109	-.0044
30.0	-.187	-.0062	-.131	-.0044
35.0	-.218	-.0062	-.153	-.0044
40.0	-.249	-.0062	-.175	-.0044
45.0	-.280	-.0062	-.197	-.0044
50.0	-.311	-.0062	-.218	-.0044
55.0	-.342	-.0062	-.240	-.0043
60.0	-.373	-.0062	-.262	-.0043
65.0	-.404	-.0062	-.284	-.0043
70.0	-.434	-.0062	-.305	-.0043
75.0	-.465	-.0062	-.327	-.0043
80.0	-.496	-.0062	-.348	-.0043
85.0	-.527	-.0061	-.370	-.0043
90.0	-.557	-.0061	-.391	-.0043
95.0	-.588	-.0061	-.413	-.0043
100.0	-.619	-.0061	-.434	-.0043

Table 6.4.4 b Calculated disturbance of measured temperatures and temperature gradients for site C85-E.

Assumed thermal diffusivity of underground:

$$0.6 \times 10^{-6} \text{ m}^2/\text{s}.$$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance ($^{\circ}\text{C}$)	Gradient disturbance ($^{\circ}\text{C}/\text{m}$)	Temperature disturbance ($^{\circ}\text{C}$)	Gradient disturbance ($^{\circ}\text{C}/\text{m}$)
0.0	0.000	-.0048	0.000	-.0035
5.0	-.024	-.0048	-.017	-.0035
10.0	-.048	-.0048	-.035	-.0035
15.0	-.073	-.0048	-.052	-.0035
20.0	-.097	-.0048	-.069	-.0035
25.0	-.121	-.0048	-.087	-.0035
30.0	-.145	-.0048	-.104	-.0035
35.0	-.170	-.0048	-.121	-.0035
40.0	-.194	-.0048	-.139	-.0035
45.0	-.218	-.0048	-.156	-.0035
50.0	-.242	-.0048	-.173	-.0035
55.0	-.266	-.0048	-.191	-.0035
60.0	-.290	-.0048	-.208	-.0035
65.0	-.315	-.0048	-.225	-.0034
70.0	-.339	-.0048	-.242	-.0034
75.0	-.363	-.0048	-.260	-.0034
80.0	-.387	-.0048	-.277	-.0034
85.0	-.411	-.0048	-.294	-.0034
90.0	-.435	-.0048	-.311	-.0034
95.0	-.459	-.0048	-.328	-.0034
100.0	-.483	-.0048	-.345	-.0034

B: Thermal diffusivity of the underground: $0.4 \times 10^{-6} \text{ m}^2/\text{s}.$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance ($^{\circ}\text{C}$)	Gradient disturbance ($^{\circ}\text{C}/\text{m}$)	Temperature disturbance ($^{\circ}\text{C}$)	Gradient disturbance ($^{\circ}\text{C}/\text{m}$)
0.0	0.000	-.0059	0.000	-.0043
5.0	-.030	-.0059	-.021	-.0043
10.0	-.059	-.0059	-.043	-.0042
15.0	-.089	-.0059	-.064	-.0042
20.0	-.119	-.0059	-.085	-.0042
25.0	-.148	-.0059	-.106	-.0042
30.0	-.178	-.0059	-.127	-.0042
35.0	-.208	-.0059	-.149	-.0042
40.0	-.237	-.0059	-.170	-.0042
45.0	-.267	-.0059	-.191	-.0042
50.0	-.296	-.0059	-.212	-.0042
55.0	-.326	-.0059	-.233	-.0042
60.0	-.355	-.0059	-.254	-.0042
65.0	-.385	-.0059	-.275	-.0042
70.0	-.414	-.0059	-.296	-.0042
75.0	-.444	-.0059	-.317	-.0042
80.0	-.473	-.0059	-.338	-.0042
85.0	-.502	-.0059	-.359	-.0042
90.0	-.532	-.0058	-.380	-.0042
95.0	-.561	-.0058	-.401	-.0042
100.0	-.590	-.0058	-.422	-.0042

Table 6.4.4 c Calculated disturbance of measured temperatures and temperature gradients for site D85-H.

Assumed thermal diffusivity of underground:

$$0.6 \times 10^{-6} \text{ m}^2/\text{s}.$$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0048	0.000	-.0036
5.0	-.024	-.0048	-.018	-.0036
10.0	-.048	-.0048	-.036	-.0036
15.0	-.072	-.0048	-.053	-.0036
20.0	-.096	-.0048	-.071	-.0036
25.0	-.119	-.0048	-.089	-.0036
30.0	-.143	-.0048	-.107	-.0036
35.0	-.167	-.0048	-.124	-.0035
40.0	-.191	-.0048	-.142	-.0035
45.0	-.215	-.0048	-.160	-.0035
50.0	-.239	-.0048	-.178	-.0035
55.0	-.262	-.0048	-.195	-.0035
60.0	-.286	-.0048	-.213	-.0035
65.0	-.310	-.0048	-.231	-.0035
70.0	-.334	-.0047	-.248	-.0035
75.0	-.357	-.0047	-.266	-.0035
80.0	-.381	-.0047	-.283	-.0035
85.0	-.405	-.0047	-.301	-.0035
90.0	-.428	-.0047	-.319	-.0035
95.0	-.452	-.0047	-.336	-.0035
100.0	-.476	-.0047	-.354	-.0035

B: Thermal diffusivity of the underground: $0.4 \times 10^{-6} \text{ m}^2/\text{s}.$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0059	0.000	-.0044
5.0	-.029	-.0059	-.022	-.0044
10.0	-.059	-.0059	-.044	-.0044
15.0	-.088	-.0058	-.065	-.0044
20.0	-.117	-.0058	-.087	-.0044
25.0	-.146	-.0058	-.109	-.0043
30.0	-.175	-.0058	-.131	-.0043
35.0	-.205	-.0058	-.152	-.0043
40.0	-.234	-.0058	-.174	-.0043
45.0	-.263	-.0058	-.196	-.0043
50.0	-.292	-.0058	-.217	-.0043
55.0	-.321	-.0058	-.239	-.0043
60.0	-.350	-.0058	-.261	-.0043
65.0	-.379	-.0058	-.282	-.0043
70.0	-.408	-.0058	-.304	-.0043
75.0	-.437	-.0058	-.325	-.0043
80.0	-.466	-.0058	-.347	-.0043
85.0	-.495	-.0058	-.368	-.0043
90.0	-.524	-.0058	-.389	-.0043
95.0	-.553	-.0057	-.411	-.0043
100.0	-.581	-.0057	-.432	-.0043

Table 6.4.4 d Calculated correction of measured temperatures and temperature gradients for site D85-I and site D85-V.

Assumed thermal diffusivity of underground:

$$0.6 \times 10^{-6} \text{ m}^2/\text{s}.$$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0026	0.000	-.0019
5.0	-.013	-.0026	-.010	-.0019
10.0	-.026	-.0026	-.019	-.0019
15.0	-.039	-.0026	-.029	-.0019
20.0	-.052	-.0026	-.039	-.0019
25.0	-.065	-.0026	-.049	-.0019
30.0	-.079	-.0026	-.058	-.0019
35.0	-.092	-.0026	-.068	-.0019
40.0	-.105	-.0026	-.078	-.0019
45.0	-.118	-.0026	-.088	-.0019
50.0	-.131	-.0026	-.097	-.0019
55.0	-.144	-.0026	-.107	-.0019
60.0	-.157	-.0026	-.117	-.0019
65.0	-.170	-.0026	-.126	-.0019
70.0	-.183	-.0026	-.136	-.0019
75.0	-.196	-.0026	-.146	-.0019
80.0	-.209	-.0026	-.155	-.0019
85.0	-.222	-.0026	-.165	-.0019
90.0	-.235	-.0026	-.174	-.0019
95.0	-.248	-.0026	-.184	-.0019
100.0	-.261	-.0026	-.194	-.0019

B: Thermal diffusivity of the underground: $0.4 \times 10^{-6} \text{ m}^2/\text{s}.$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0032	0.000	-.0023
5.0	-.016	-.0032	-.012	-.0023
10.0	-.032	-.0032	-.023	-.0023
15.0	-.048	-.0032	-.035	-.0023
20.0	-.064	-.0032	-.047	-.0023
25.0	-.080	-.0032	-.058	-.0023
30.0	-.096	-.0032	-.070	-.0023
35.0	-.112	-.0032	-.082	-.0023
40.0	-.128	-.0032	-.093	-.0023
45.0	-.144	-.0032	-.105	-.0023
50.0	-.160	-.0032	-.116	-.0023
55.0	-.176	-.0032	-.128	-.0023
60.0	-.192	-.0032	-.140	-.0023
65.0	-.208	-.0032	-.151	-.0023
70.0	-.224	-.0032	-.163	-.0023
75.0	-.240	-.0032	-.174	-.0023
80.0	-.255	-.0032	-.186	-.0023
85.0	-.271	-.0032	-.197	-.0023
90.0	-.287	-.0032	-.209	-.0023
95.0	-.303	-.0031	-.220	-.0023
100.0	-.319	-.0031	-.231	-.0023

Table 6.4.4 e Calculated disturbance of measured temperatures and temperature gradients for site D85-K and site D85-W.
Assumed thermal diffusivity of underground:

$$0.6 \times 10^{-6} \text{ m}^2/\text{s}.$$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0028	0.000	-.0019
5.0	-.014	-.0028	-.010	-.0019
10.0	-.028	-.0028	-.019	-.0019
15.0	-.042	-.0028	-.029	-.0019
20.0	-.056	-.0028	-.038	-.0019
25.0	-.071	-.0028	-.048	-.0019
30.0	-.085	-.0028	-.057	-.0019
35.0	-.099	-.0028	-.067	-.0019
40.0	-.113	-.0028	-.076	-.0019
45.0	-.127	-.0028	-.086	-.0019
50.0	-.141	-.0028	-.095	-.0019
55.0	-.155	-.0028	-.105	-.0019
60.0	-.169	-.0028	-.114	-.0019
65.0	-.183	-.0028	-.124	-.0019
70.0	-.197	-.0028	-.133	-.0019
75.0	-.211	-.0028	-.142	-.0019
80.0	-.225	-.0028	-.152	-.0019
85.0	-.239	-.0028	-.161	-.0019
90.0	-.253	-.0028	-.171	-.0019
95.0	-.267	-.0028	-.180	-.0019
100.0	-.281	-.0028	-.189	-.0019

B: Thermal diffusivity of the underground: $0.4 \times 10^{-6} \text{ m}^2/\text{s}.$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0035	0.000	-.0024
5.0	-.017	-.0035	-.012	-.0024
10.0	-.035	-.0035	-.024	-.0024
15.0	-.052	-.0035	-.036	-.0024
20.0	-.069	-.0035	-.048	-.0024
25.0	-.086	-.0035	-.060	-.0024
30.0	-.104	-.0035	-.071	-.0024
35.0	-.121	-.0035	-.083	-.0024
40.0	-.138	-.0034	-.095	-.0024
45.0	-.156	-.0034	-.107	-.0024
50.0	-.173	-.0034	-.119	-.0024
55.0	-.190	-.0034	-.131	-.0024
60.0	-.207	-.0034	-.143	-.0024
65.0	-.224	-.0034	-.154	-.0024
70.0	-.241	-.0034	-.166	-.0024
75.0	-.259	-.0034	-.178	-.0024
80.0	-.276	-.0034	-.190	-.0023
85.0	-.293	-.0034	-.202	-.0023
90.0	-.310	-.0034	-.213	-.0023
95.0	-.327	-.0034	-.225	-.0023
100.0	-.344	-.0034	-.237	-.0023

Table 6.4.4 f Calculated correction of measured temperatures and temperature gradients for site D85-N.

Assumed thermal diffusivity of underground:

$$0.6 \times 10^{-6} \text{ m}^2/\text{s}.$$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0026	0.000	-.0016
5.0	-.013	-.0026	-.008	-.0016
10.0	-.026	-.0026	-.016	-.0016
15.0	-.039	-.0026	-.024	-.0016
20.0	-.052	-.0026	-.032	-.0016
25.0	-.066	-.0026	-.040	-.0016
30.0	-.079	-.0026	-.048	-.0016
35.0	-.092	-.0026	-.056	-.0016
40.0	-.105	-.0026	-.064	-.0016
45.0	-.118	-.0026	-.072	-.0016
50.0	-.131	-.0026	-.080	-.0016
55.0	-.144	-.0026	-.088	-.0016
60.0	-.157	-.0026	-.096	-.0016
65.0	-.170	-.0026	-.104	-.0016
70.0	-.183	-.0026	-.112	-.0016
75.0	-.196	-.0026	-.120	-.0016
80.0	-.209	-.0026	-.128	-.0016
85.0	-.222	-.0026	-.136	-.0016
90.0	-.235	-.0026	-.144	-.0016
95.0	-.248	-.0026	-.152	-.0016
100.0	-.261	-.0026	-.160	-.0016

B: Thermal diffusivity of the underground: $0.4 \times 10^{-6} \text{ m}^2/\text{s}.$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0032	0.000	-.0020
5.0	-.016	-.0032	-.010	-.0020
10.0	-.032	-.0032	-.020	-.0020
15.0	-.048	-.0032	-.030	-.0020
20.0	-.064	-.0032	-.039	-.0020
25.0	-.080	-.0032	-.049	-.0020
30.0	-.096	-.0032	-.059	-.0020
35.0	-.112	-.0032	-.069	-.0020
40.0	-.128	-.0032	-.079	-.0020
45.0	-.144	-.0032	-.089	-.0020
50.0	-.160	-.0032	-.098	-.0020
55.0	-.176	-.0032	-.108	-.0020
60.0	-.192	-.0032	-.118	-.0020
65.0	-.208	-.0032	-.128	-.0020
70.0	-.224	-.0032	-.137	-.0019
75.0	-.240	-.0032	-.147	-.0019
80.0	-.256	-.0032	-.157	-.0019
85.0	-.272	-.0032	-.166	-.0019
90.0	-.288	-.0032	-.176	-.0019
95.0	-.303	-.0032	-.186	-.0019
100.0	-.319	-.0032	-.195	-.0019

Table 6.4.4 g Calculated correction of measured temperatures and temperature gradients for site D85-0.

Assumed thermal diffusivity of underground:

$$0.6 \times 10^{-6} \text{ m}^2/\text{s}.$$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0009	0.000	-.0001
5.0	-.004	-.0009	-.001	-.0001
10.0	-.009	-.0009	-.001	-.0001
15.0	-.013	-.0009	-.002	-.0001
20.0	-.017	-.0009	-.003	-.0001
25.0	-.022	-.0009	-.003	-.0001
30.0	-.026	-.0009	-.004	-.0001
35.0	-.030	-.0009	-.005	-.0001
40.0	-.035	-.0009	-.005	-.0001
45.0	-.039	-.0009	-.006	-.0001
50.0	-.043	-.0009	-.007	-.0001
55.0	-.048	-.0009	-.008	-.0001
60.0	-.052	-.0009	-.008	-.0001
65.0	-.056	-.0009	-.009	-.0001
70.0	-.061	-.0009	-.010	-.0001
75.0	-.065	-.0009	-.010	-.0001
80.0	-.069	-.0009	-.011	-.0001
85.0	-.074	-.0009	-.012	-.0001
90.0	-.078	-.0009	-.012	-.0001
95.0	-.082	-.0009	-.013	-.0001
100.0	-.087	-.0009	-.014	-.0001

B: Thermal diffusivity of the underground: $0.4 \times 10^{-6} \text{ m}^2/\text{s}.$

Depth below sea bed (m)	Temperature history model I		Temperature history model II	
	Temperature disturbance (°C)	Gradient disturbance (°C/m)	Temperature disturbance (°C)	Gradient disturbance (°C/m)
0.0	0.000	-.0011	0.000	-.0002
5.0	-.005	-.0011	-.001	-.0002
10.0	-.011	-.0011	-.002	-.0002
15.0	-.016	-.0011	-.003	-.0002
20.0	-.021	-.0011	-.003	-.0002
25.0	-.027	-.0011	-.004	-.0002
30.0	-.032	-.0011	-.005	-.0002
35.0	-.037	-.0011	-.006	-.0002
40.0	-.043	-.0011	-.007	-.0002
45.0	-.048	-.0011	-.008	-.0002
50.0	-.053	-.0011	-.008	-.0002
55.0	-.059	-.0011	-.009	-.0002
60.0	-.064	-.0011	-.010	-.0002
65.0	-.069	-.0011	-.011	-.0002
70.0	-.074	-.0011	-.012	-.0002
75.0	-.080	-.0011	-.013	-.0002
80.0	-.085	-.0011	-.013	-.0002
85.0	-.090	-.0011	-.014	-.0002
90.0	-.096	-.0011	-.015	-.0002
95.0	-.101	-.0011	-.016	-.0002
100.0	-.106	-.0011	-.017	-.0002

6.4.6 Topographic effects

Topographic corrections to the thermal gradient is necessary in areas with marked local variations in surface altitude. This is necessary when the surface towards which the heat flows (representing a boundary condition) is at different levels and has different temperatures in short enough distances away from the site to affect the downward temperature increase (i.e. the topography causes the heat flow to deviate from the vertical).

The topographic effect on temperature gradients can be estimated by an equation derived by Jeager (1965) (in Kappelmayer & Haenel 1974):

$$G = G_u - \frac{x \cdot \Delta T_o}{T_i (x^2 + z^2)}$$

where

- G is the apparent gradient influenced by topography
- G_u is corrected thermal gradient in the underground
- x is the distance away from the site to a change in annual surface temperature ΔT_o
- z is the depth below the surface of the site.

On land the use of this equation is relatively simple as the mean annual temperature for a given region generally can be expressed as a linear function of altitude. In the sea water masses, however there may exist different sheets of water with different temperatures. Mean annual temperature difference within a given water depth range at a spesific geographical position is therefore influenced by 1) the depth to the bundary between different water masses, and 2) the mean annual temperatures within the actual water masses.

In our case we have considered the topographic relief on all our sites (as revealed by our shallow seismic data), and for the "worst case" quantified what magnitude of ΔT_o that is necessary to make the topographic effect significant.

The "worst case" in this respect was found to be site D85-0 which is situated on a ridge of Quaternary sediments. At this site the sea bed is about 20 m lower about 1.7 km away from the borehole. It can be shown by the equation above that even with oceanographic conditions causing a 4 °C difference in mean annual temperature between the sea floor at site D85-0 and at 20 m deeper water at the slope of the sediment ridge, the effect on the measured gradient would be less than 2%. Because of the very unlikely oceanographic conditions required, we find such a temperature difference to be unrealistically high and therefore choose to neglect this effect.

A similar discussion can be carried out to show that effects from actual lateral variations in Quaternary thickness can be neglected.

The topographic effect from relict plough marks from drifting icebergs (Lien 1983) is also found negligible. For this case the topographic conditions may be less favourable. ΔT_0 , however, is quite insignificant, and this will imply a negligible correction.

6.5 Shallow probe data

6.5.1 Introduction

During the 1985 IKU shallow drilling and heat flow cruise a gravity spear with fixed thermistors was used to measure temperature to a maximum depth of approximately 3 m below the sea floor. Corresponding thermal conductivity values were obtained by use of the needle probe method on sediment samples collected by a gravity corer (Sættem et al. 1985).

The aim of these shallow probe data was to evaluate the potential for measuring heat flow by use of a relatively simple device lowered to the sea floor.

An evaluation of the results from these measurements is carried out by Omegalink International Ltd. (Zielinski 1986b), and Omegalink's report to IKU is included as Appendix II in this volume.

A brief summary of the results from this work is presented in Chapt 6.5.2.

6.5.2 Methods

The shallow probe data were investigated with a two-fold purpose:

- 1) To evaluate the feasibility and limitations of using the shallow probe equipment to measure heat flow
- 2) To (if possible) calculate heat flow from shallow probe data

Zielinski & Bruchhausen (1983) describes a method for calculating steady state temperature gradient from measurements in the upper few metres of the underground. This method relies on input about mean annual surface temperature T_s , annual temperature amplitude, A , and the time t_0 when $T = T_s$.

For the Barents Sea data A and t_0 are unknown. T_s can be established with high accuracy from extrapolation from the borehole temperature data, taking into account the variation in thermal conductivity with depth.

Zielinski et al. (1986b) have proposed a method whereby values for A and t_0 are obtained as best fit values based on an algorithm similar to those described by Zielinski & Bruchhausen (1983), with the following known input parameters.

- the steady state gradient
- temperatures at approximately 1, 2 and 3 metres
- mean annual surface temperature

6.5.3 Results

An attempt has been made to extrapolate the calculated values for A and t_0 to the sites where only shallow probe data exists and at these

sites calculate the steady state thermal gradient by following the scheme of Zielinski & Bruchhausen (1983).

The conclusion is that this is not possible with only three temperature depth points from the upper 3 m below the sea bed.

However, further investigation and sensitivity tests of the method have given rather optimistic results regarding the use of a longer (up to 10 m) shallow probe device with several thermistors to measure heat flow. Our conclusion can be summarized as follows:

1. Temperature/depth data from shallow drilling boreholes (10-50 m) can be used to calibrate measurements from the upper 10 m below the sea bed (which are influenced from the seasonal temperature variations).
2. The results from this calibration can be extrapolated to a certain distance from the shallow drilling site. When doing this proper consideration of oceanographic conditions will be required.
3. The resolution of a heat flow value resulting from 8 temperature/depth points between 0 m and 10 m below the sea bed with a resolution of 0.001°C of each temperature measurement, may be better than 20% of the actual steady state heat flow value.

Though the resolution of the data obtained by the method indicated above is lower than is the case in the state of the art measurements from shallow drillings, we feel that the method has a high potential which should be investigated further. This method may especially be useful in studies of fluid flow in the underground. (a.o. Zielinski & Bruchhausen 1983; Zielinski et al. 1985).

For such studies high accuracy and resolution of the measured heat flow is not of the same importance as when the objective of the measurements is to extrapolate temperatures at depth. However, fluid flow studies would require collection of heat flow data with a spatial density far greater than what is realistic for the shallow drilling method (provided that sufficient density of shallow drilling data or actual annual temperature curves for the bottom water for calibration purposes

exist) and this method would economically enable heat flow profiles to be gathered across structures like basin rims, major fault zones, traps etc.

6.6 Final heat flow values for the 1985 IKU Barents Sea sites

6.6.1 Final measurement results (field results)

Table 6.6.1 summarizes the final measured gradients and heat flow values from the IKU heat flow sites in the Barents Sea. In this table the error in previously reported thermal conductivity values (IKU report 21.3405.00/02/85) is accounted for. The values given are not subjected to any influence from in situ versus laboratory conditions, climatic corrections etc. and thus represent final field results.

Table 6.6.1 Final measured gradients and heat flow values (field results). R-SQUARE is the standard coefficient of determination for regression (least square fit for all temperature-depth points). Values near 1.0 indicate good fit; values near zero indicate poor fit. 1 HFU=1 mcal cm⁻²s⁻¹.

SITE	GRAD(mdegC/m)	N	R-SQUARE	K _{av} (mcal °C ⁻¹ cm ⁻¹ s ⁻¹)	MHF(HFU)
C85-A	42.8	3	0.9996	4.16	1.78
C85-E	33.1	3	0.9925	3.94	1.30
D85-K	34.4	16	0.9976	3.87	1.33
D85-N	38.2	3	0.9998	4.10	1.57
D85-O	31.0	3	0.9996	4.97	1.54
D85-V	36.8	5	0.9953	4.41	1.62

6.6.2 Correction of thermal conductivity values

Pressure release effects

The correction to apply for difference in temperature and pressure in laboratory versus in situ conditions have been found negligible.

From data about water content, bulk density and mineralogy it is found that the samples from sites D85-0, D85-N and D85-V have expanded since they were retrieved from the borehole. This is probably due to dissolved gas in the pore water, being released and forming gas voids in atmospheric pressure. This is also supported by results from organic geochemical analyses (Fjerdingsstad et al. 1985). A further support to this theory is the large scatter in thermal conductivity versus depth for these three stations. The correction to apply for this is calculated in Chapter 6.3.4.

The representative correction for the actual depth interval can be expressed as a correction factor " k_g " that can be applied directly to K_{av} in the Table 6.6.1 above.

From Table 6.3.7, " k_g " is defined as

$$k_g = \frac{C-IN\ SITU}{C-MEAS}$$

The resulting correction factors " k_g " found representative for the depth intervals corresponding to the temperature measurements are:

Site	k_g
D85-N	1.16
D85-0	1.08
D85-V	1.04

Thermal conductivity anisotropy

The thermal conductivity measurements carried out at the University of Aarhus (Chapter 6.3.4) give clear indications of an anisotropy in the thermal conductivity of the Quaternary samples from the Barents Sea.

The factor of anisotropy, A , is defined as:

$$A = \frac{k_h}{k_v}$$

A is calculated from 28 pairs of horizontal versus vertical thermal conductivity, and though the statistics is limited, the trend is clear and the following mean value can be calculated.

$$A_{\text{mean}} = 1.13$$

(Standard deviation 0.14)

K_{av} in Table 6.6.1 above is believed to be largely representative for the horizontal thermal conductivity and a reduction of K_{av} should be included in the vertical thermal conductivity. We therefore suggest a correction factor k_{av} of 0.91 to be applied ($0.91 = \frac{1}{1.1}$).

6.6.3 Correction due to topographic effects and glacial history

Effects on measured temperature gradients from topographic relief on the sea floor are found to be negligible.

The variations in paleotemperatures due to the glacial history is evaluated for the last 245 000 years. This effect is found to vary depending upon glaciation models and site locations. Maximum and minimum effects are calculated based on two glaciation models. The result of this is summarized in Table 6.4.3 and in the Table 6.6.2 below this effect is expressed as a correction factor to the actual measured temperature gradient for each individual site.

Table 6.6.2 Correction factors to apply on measured temperature gradients due to the maximum and minimum glaciation models.

SITE	Correction factor	
	Model I	Model II
C85-A	1.14	1.08
C85-E	1.18	1.11
D85-K	1.10	1.06
D85-N	1.08	1.04
D85-O	1.04	1.00
D85-V	1.09	1.05

6.6.4 Discussion

Table 6.6.3 summarizes the various correction factors that can be applied to the measurement results. In Fig. 6.6.1 the most likely combination of these corrections is applied, and the resulting adjusted heat flow values are plotted together with the final field measurement results.

This figure also summarizes some published examples of heat flow from different provinces. Within the scope of the present report we will, however, not perform any comparison to those values.

We feel that the correction for anisotropy in the thermal conductivity is relatively clearly expressed in our data and therefore include this correction in our recommended adjusted heat flow values.

Table 6.6.3 Heat flow sites in the Barents Sea 1985.
Final heat flow values and possible corrections
(expressed as correction factors).

SITE	Field results		Correction of thermal conductivity		Correction for glacial history	
	(HFU)	(mW/m ²)	k _g	$k_A = \frac{1}{A_{\text{mean}}}$	Model I	Model II
C85-A	1.78	74.5	1.00	0.91	1.14	1.08
C85-E	1.30	54.5	1.00	0.91	1.18	1.11
D85-K	1.33	55.7	1.00	0.91	1.10	1.06
D85-N	1.57	65.7	1.16	0.91	1.08	1.04
D85-O	1.54	64.5	1.08	0.91	1.04	1.00
D85-V	1.62	76.8	1.04	0.91	1.09	1.05

Correction due to erroneous measurements due to gas expansion in the samples under laboratory conditions is documented with somewhat less certainty. An important input to these calculations is the solid density of the mineral fraction. This is estimated from the mineralogy and not measured. Still we feel that the results from these calculations supports earlier observations about gas content and large scatter in plots of thermal conductivity versus depth (Fjordingstad et al. 1985 and Sættem et al. 1985) as all results point out that this effect may be present especially in the samples from sites D85-N, D85-O and D85-V. We therefore suggest that also this effect is accounted for.

The correction for the effects from the glacial history can be done according to either of the two glaciation models presented in Chapter 6.4. These two models are believed to represent the maximum and minimum likely transient effect on the present shallow temperature gradients. From scientific evidences it is not possible to prefer one of the models. However, tentatively we would suggest to use Model II for this correction.

From the above discussion and Table 6.6.3, we may therefore summarize the field results and final adjusted results for the heat flow sites as presented in Table 6.6.4.

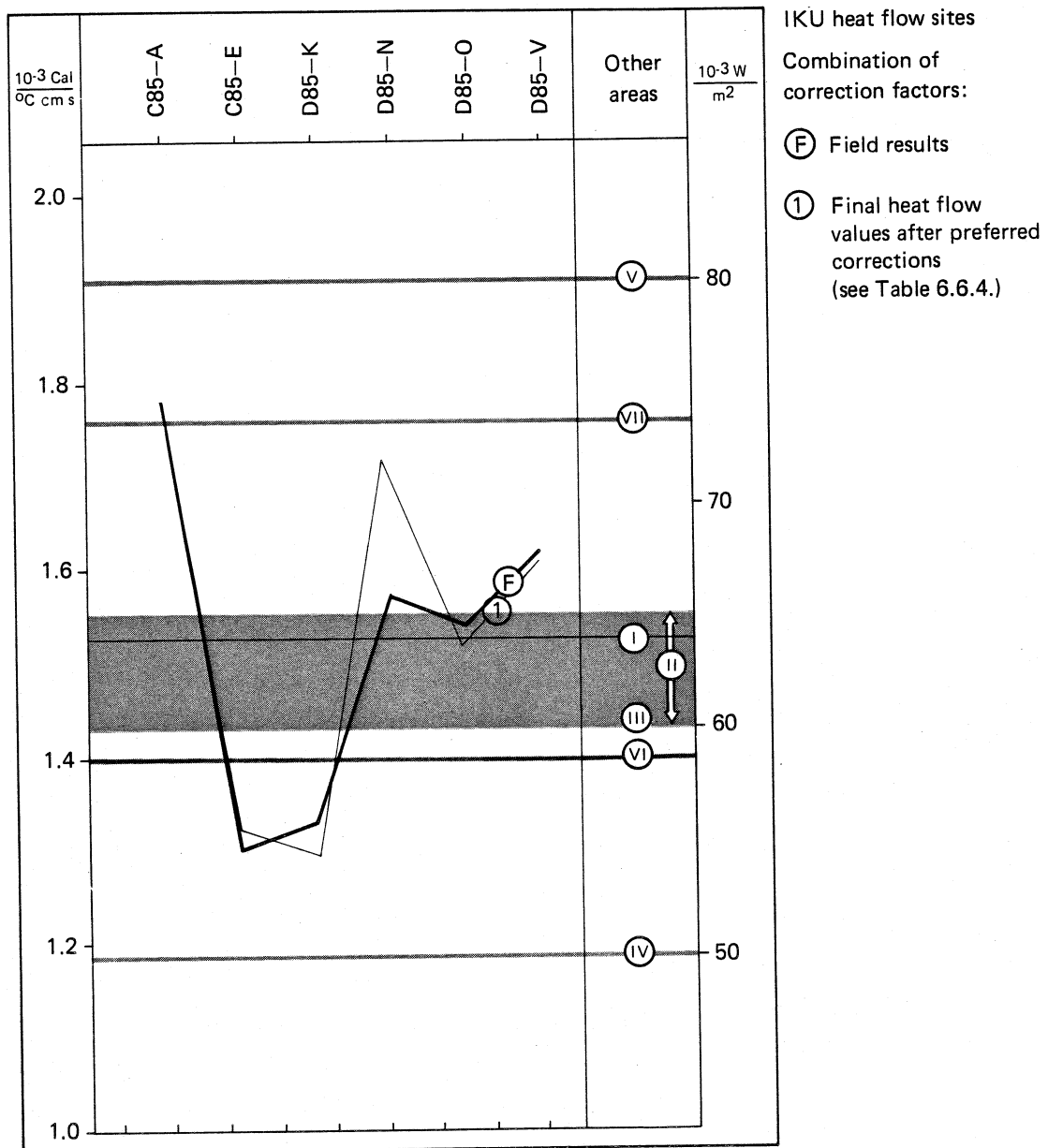
These two alternatives are also plotted in Fig. 6.6.1.

Table 6.6.4 Heat flow data from the IKU 1985 sites in the Barents Sea.

A: Field results.

B: Field results subjected to suggested adjustments:
Anisotropy, gas, climatic effects (see discussion in text).

Site	A: Field results (HFU) mW/m^2		B: Suggested adjusted results (HFU) (mW/m^2)	
C85-A	1.78	74.5	1.77	74.1
C85-E	1.30	54.5	1.32	55.2
D85-K	1.33	55.7	1.29	54.0
D85-N	1.57	65.7	1.72	72.0
D85-O	1.54	64.5	1.52	63.6
D85-V	1.62	76.8	1.61	67.4



- (I) Europe (Čermak 1979)
- (II) North Sea (52° – 56° N) (Andrews – Speed *et al.* 1984)
- (III) Mainland UK (Richardson and Oxburg 1979 in Andrews – Speed *et al.* 1984)
- (IV) Central Graben and Anglo Dutch Basin (Andrews – Speed *et al.* 1984)
- (V) Mid North Sea High and East Midland Shelf (Andrews – Speed *et al.* 1984)
- (VI) Continental crust (Lee 1963 in Gretner 1981)
- (VII) Mesozoic- Cenozoic orogenic areas (Lee 1969 in Zielinski 1986 a)

Fig. 6.6.1 Heat flow values from the Barents Sea. Published heat flow values from other areas are shown for comparison.

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6.8 Appendix I

ZIELINSKI, G.W., 1986a: Correction to thermal conductivity values reported for 1985 field program. Omegalink International Ltd. Confidential report, 8 pp + appendices.

CORRECTION TO THERMAL CONDUCTIVITY
VALUES REPORTED FOR 1985 FIELD PROGRAM

-OMEGALINK INTERNATIONAL LTD.-

The relation upon which the needle-probe thermal conductivity measurements are based can be written (Von Herzen and Maxwell, 1959):

$$T = \frac{q}{4K\pi} (\ln 4at/b) \quad (1)$$

where

a = thermal diffusivity of sediment sample

b = probe radius squared

t = time

q = heat input per unit length per unit time

K = thermal conductivity of sediment sample

T = temperature rise of probe

Kappelmeyer and Haenel (1974) essentially rearrange this equation to:

$$K = \frac{0.0438bP}{lT} \quad (2)$$

where in this case,

P = power input to the probe (Watts)

T = temperature rise of probe over one logarithmic unit of time (such that $l nt = 2.303$)

b = constant = 1 for our needle probes

l = length of heater wire.

Essentially constants and conversion factors (e.g. Joules to Watts) have been consolidated and q is expressed in terms of P/l where P can be monitored (in terms of volts, amps, ohms) during conductivity measurement.

For typical needle-probes (e.g. Fig. 1, Von Herzen and

Maxwell, 1959) including those used during our 1985 field surveys, $l = 14$ cm, and for a typical heat-wire resistance value of 50 ohms, one can compute from Fig. 2 in Von Herzen and Maxwell (1959), for the given conductivity value, the value for $P=2$ watts which is also given in their text. All seems to be in agreement except for the fact that for P/l to be consistent with q from the infinite line source theory, upon which equation (1) is based, l (equation 2) must equal half the heater wire length. The heater wire runs the length of the probe and back or two times the length of our "infinite line source" (the probe), whereas, q is the heat input per unit length along the line source per unit time.

With respect to Fig. 2 in Von Herzen and Maxwell (1959), Langseth (1965) used that same figure, citing heater power as being typically between 0.5 and 1.0 watt. This value is consistent (according to equation (2) for no change in K) with $l=7$ cm or half the heater wire length. Hence the apparent dilemma regarding equation 2 from Kappelmeyer and Haenel (1974) and Von Herzen and Maxwell's (1959) Figure 2 appears resolved.

In the last-minute haste amidst final preparations for the 1985 field work, l was programmed into the new conductivity apparatus as the total (round trip) length of the heater wire instead of the one-way length (half the total length). Had this been the only error, the resulting measured values would have been half of the actual values (equation 2). However, a second, much less dramatic discrepancy between the length of the heater wire, the reported resistance per unit length at 25 deg. C. and the resistance at 25 deg. C. reported by the probes' manufacturer gave rise to a small error in the opposite direction. This was confirmed and resolved in a recent series of interchanges with the manufacturer.

As a result of the above, all previously reported thermal values for the 1985 field season should be multiplied by the correction factor f , where

$$f = \frac{\text{corrected heater wire constant}}{\text{old heater wire constant}} = 1.799$$

In order to test this result on the apparatus employed in 1985, a crude thermal conductivity standard was constructed by mixing high-organic sphagnum peat moss with distilled water. The resulting highly saturated test sample (Table 1) would be expected to return thermal conductivity values close to those of distilled water. A few hours after initial preparation of the standard it was observed that the previously homogeneous mixture began to separate and the peat began to float on a thin layer of mostly water. The range(s)

in physical property values (Table 1) reflect this occurrence. The bulk density is seen to be less than the density of water alone, probably due to minute amounts of gases (air) trapped within the plant cells forming most of the solid component.

The results of ten thermal conductivity tests using the two needle probes in question (48 and 51) are summarized in Table 2. Except for tests 4 and 5 which were performed in the exact same location, the needle-probes were moved to different points in the sample. (Probe #48 became intermittent during the tests so that additional measurements were not possible.) The resulting values reflect the variability within the sample and generally fall around the value reported for the thermal conductivity of water at 20 deg. C., 1.43 mcal/deg.C cm sec (e.g. Clark, 1966). Hillel (1980) reports an average value of 1.2 mcal/deg C cm sec for peat with a volumetric wetness of 0.8 (similar to our test sample) and Redfield (1964) reports a value of 1.46 mcal/deg. C. cm sec for salt marsh peat at 10 deg. C., 78% wt. saturation and density 1.15 gm/cc. Our results in Table 2 fall within this range; whereas, with the old heater wire constant values between 0.72 and 0.81 mcal/deg.C. cm sec would have resulted.

As a final rough check on the corrected thermal conductivity values we have plotted thermal conductivity vs. water content in Fig. 1. Also plotted are two empirical curves, Bullard and Day (1961) and Lachenbruch and Marshall (1966), which are based on measurements performed on deep ocean sediments. Departures of actual data from such curves (e.g. Kasameyer et al, 1972) are most often the result of the variability in the thermal conductivity of the solid fraction, which is not adequately accounted for by the empirical relations. For purposes of comparison, we have also included a similar plot from Horai (1982).

On Fig. 1 note that the scales have been changed from the 1985 IKU report. Water content is now expressed as % wet weight and thermal conductivity has been multiplied by the correction factor f , and has been reduced an additional 5%. The latter is intended to compensate for the difference in temperature from shipboard to seabed conditions. The data in Fig. 1 are seen to be in good agreement with the empirical curves and appear to exhibit less scatter than is seen in the Horai (1982) data (Fig. 2).

As a result of the foregoing analyses, we have made the appropriate corrections to the 1985 field data report (Appendix 1). The revised thermal conductivity values, from every indication, are the product of state of the art measurements and are representative of existing in situ conditions.

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TABLE 1 - PHYSICAL PROPERTIES OF TEST SAMPLE

"High-organic" Sphagnum Peat Moss and Distilled Water

Bulk Density	0.93 - 0.95 gm/cc
% water (wet wt.)	76 - 83
Approx. volume	600 cc

TABLE 2 - THERMAL CONDUCTIVITY TEST RESULTS

Test #	Probe #	K @ 20 deg.C (mcal/deg.C.cm sec)
01	48	1.30
02	48	1.33
03	51	1.46
04	51	1.33
05 *	51	1.33
06	51	1.40
07	51	1.33
08	51	1.46
09	48	1.46
10	48	1.30
mean/std. dev.		1.37/0.07

* exact same location in sample as test #04

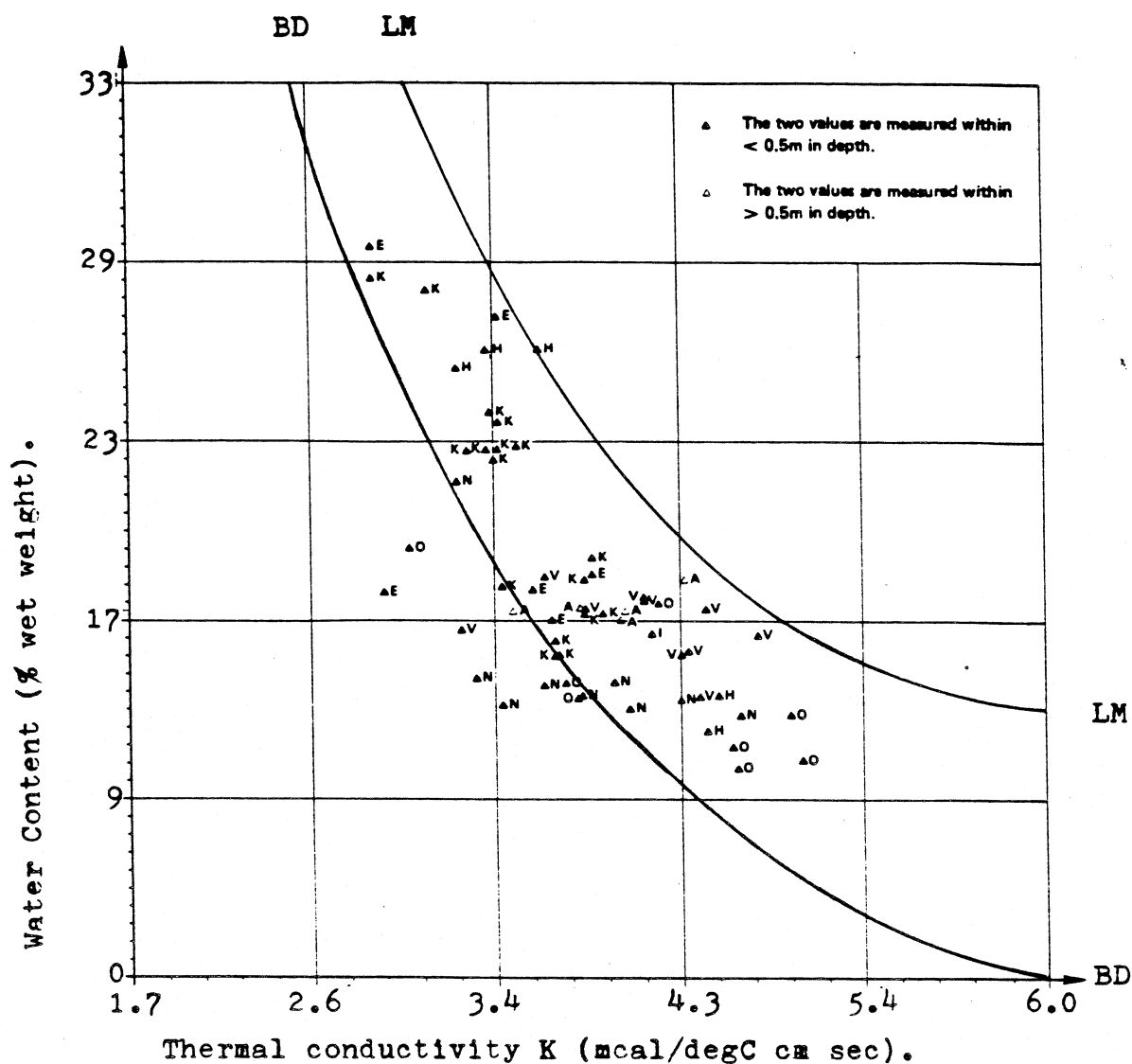


Fig. 1 Thermal conductivity versus water content for Quaternary samples obtained during the 1985 IKU drilling cruise.

BD: Bullard & Day
LM: Lachenbruch & Marshall

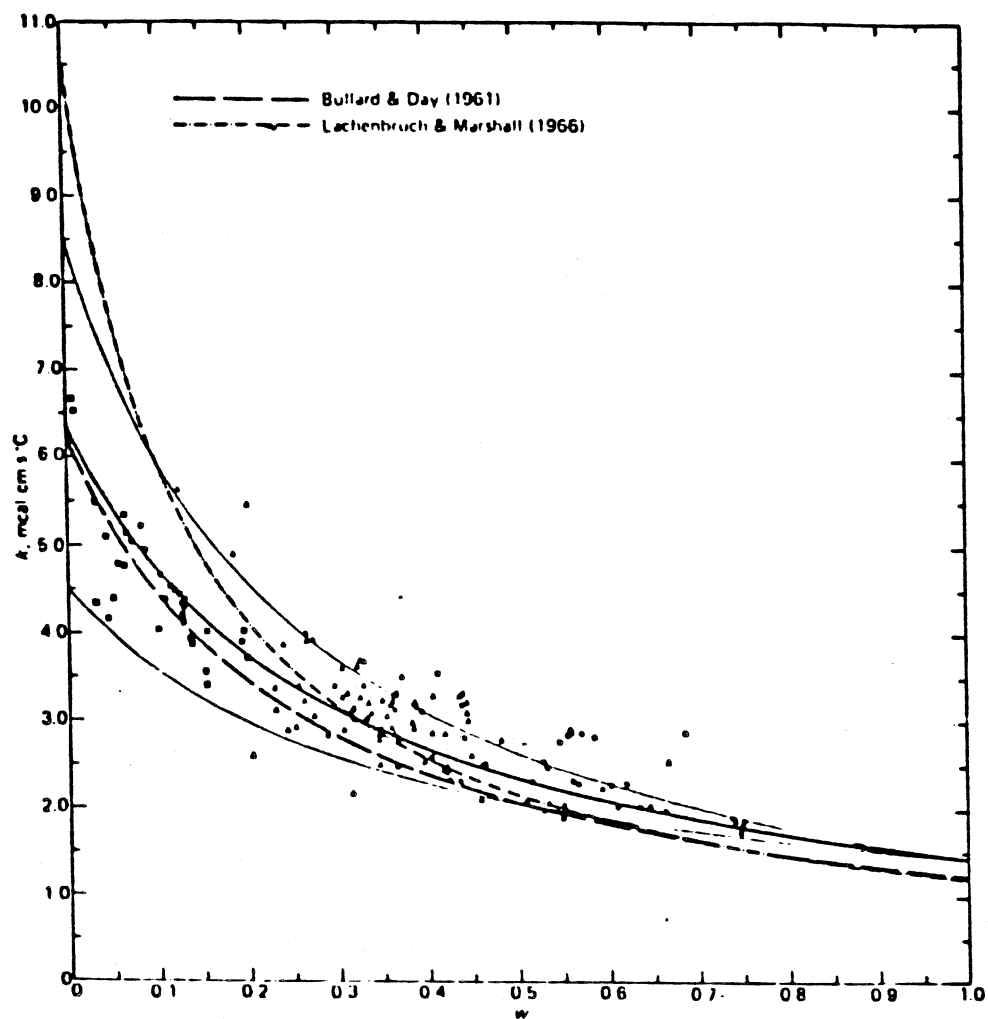


Figure 13. Thermal conductivity k versus water content w plot for Leg 60 ocean crust samples. Data size: $n = 145$. Curves: —, best fit to the data ($k_i = 6.35 \text{ mcal/cm s } ^\circ\text{C}$); ---, range of variations corresponding to $k_i = 8.50$ and $4.50 \text{ mcal/cm s } ^\circ\text{C}$.

Fig. 2 Thermal conductivity versus water content from Horai (1982).

6.9 Appendix II

ZIELINKSKI, G.W., 1986b: Feasibility of shallow probe heat flow surveys in the Barents Sea. Omegalink International Ltd. Confidential report, 9 pp.

FEASIBILITY OF SHALLOW-PROBE HEAT FLOW SURVEYS IN THE BARENTS SEA

Gary W. Zielinski, Omegalink International Ltd.

ABSTRACT

Analyses performed on shallow probe and downhole heat flow data from the 1985 Barents Sea survey define the seasonal variation in the shallow (less than 10 meters) thermal regime. Based on these data an operationally feasible scheme for calibrating shallow probe measurements using downhole data has been developed. Analyses of test data generated from the 1985 measurements suggest that high density shallow probe/shallow penetration heat flow surveys can be used to map salt related thermal anomalies, the ocean/continent heat flow transition, and fluid flow related anomalies such as might be due to active hydrocarbon seepage or migration in the Barents Sea. The latter application suggests an operational and interpretational advantage in performing combined heat flow/surface geochemical surveys in the Barents Sea and elsewhere.

INTRODUCTION

As part of the 1985 IKU Barents Sea heat flow program, reconnaissance shallow-probe heat flow measurements were performed at nine site locations, Saettem et al (1985). Corresponding thermal conductivity measurements were performed on gravity-core samples covering similar depth intervals (0 - 3 meters). The intent of these measurements was to combine the shallow thermal data with the deeper downhole data to assess the feasibility and limitations of future shallow-probe surveys in the Barents Sea as a cost effective means of increasing data density.

The main problem associated with shallow penetration (less than 10 meters) heat flow measurements in shallow water areas is due to periodic fluctuation in bottom water temperature. These fluctuations propagate into the sediment and generally constitute a serious disturbance to the background heat flow. This disturbing effect is generally absent in abyssal plain areas where most marine heat flow measurements are confined. The main task with respect to the Barents Sea data is (1) to assess the magnitude and variability of this disturbance over the 1985 survey area and, based on these results, (2) to estimate the minimum resolution possible for a fully configured (as opposed to

reconnaissance) shallow-probe heat flow survey.

METHODOLOGY

In order to accomplish this, we first focus our attention on sites A, E, V, O, N and K containing both shallow-probe and multiple depth point downhole thermal measurements. Based on a varying degree of contrast between the thermal conductivities of shallow and deep samples, we have adopted a simple two-layer model of shallow lower value thermal conductivity (K_s) sediments overlying deeper values (K_d) to some depth (D), which is less than or equal to 10 meters. Generally speaking, K_s corresponds to the depth interval of the shallow-probe measurements and K_d to the downhole measurements, both being harmonic means of the measured values. Via this crude model, values of K_d/K_s and D were combined with the values of the linear temperature gradients (GRAD) and their surface intercept temperatures (T_i) from the downhole data analysis. Thereby, the shallow gradients (SGRAD) and the mean annual surface temperatures (T_s) for each site were computed (TABLE 1A).

Based on visual inspection of the temperature profiles for each site (Saettem et al, 1985) we have adopted a simple model for the observed temperatures in the upper 10 meters of sediment:

$$T = T_s (\text{GRAD}) Z + A \exp(-ZX) \sin(wt' - ZX)$$

where $X = (w/2\mu)^{1/2}$, $t' = t - t_0$ and $w = 2\pi/\text{year}$ (1/sec).

(1)

Accordingly, it is assumed that a seasonal sinusoidal variation in bottom water temperature of amplitude A (half the seasonal range) and mean annual bottom water temperature T_s is superimposed upon the steady-state thermal gradient GRAD (in this case SGRAD). μ is thermal diffusivity and t_0 is the time when $T = T_s$ (i.e., bottom water temperature is at its mean annual value), t is the time of measurement and Z is depth. A linear least squares algorithm similar to those described by Zielinski and Bruchhausen (1983) and Zielinski et al (1985) was used to compute best-fit values of A and t_0 for the sites listed in Table 1 to equation 1 using SGRAD, T_s and the shallow-probe temperature/depth data. (Saettem et al, 1985) Essentially, by this procedure the downhole data are used to calibrate the shallow-probe data.

RESULTS

The high degree of agreement of the data to equation 1 is evidenced by the low values of RMS error (RMSE) in Table 1, which are generally well below the millidegree C level. All but site O yielded minimum error solutions by this procedure. A second algorithm was developed to perform the inverse process, i.e., to compute SGRAD from the shallow probe data, this time specifying only A and t-to. The results of this exercise (SGRAD') are also given in Table 1B and the close correspondence between SGRAD and SGRAD' for each site attests to self-consistency of the methods used. The direct implication is that if we remain close enough to a given site location such that A and t-to remain relatively constant, we can extract the steady-state geothermal gradient from shallow-probe data alone. Multiplying this value by the mean thermal conductivity measured from equivalent depth gravity core values would theoretically result in a reliable heat flow determination, based on the results in Table 1.

In actuality, significant variations in the values of A and t-to occur over the survey area, some of which are likely to be the result of uncertainty in the penetration of the shallow probe and too few temperature/depth points. These shortcomings would be largely eliminated using a fully-configured heat flow probe designed to measure heat flow, not just the potential for doing so (our 1985 objective).

From the revised measured thermal conductivity values we have computed mean thermal diffusivities (μ) for each site using Bullard (1963). These values were used to provide additional constraint on the regression algorithm based on equation 1. The results of this exercise are given in Table 2. Although there is a deterioration in fit as seen by larger RMS error, we have evidence that this is largely due to the uncertainties just discussed with respect to the shallow probe and that depth correction could eliminate this error. Note that there is a reduction in the variation of A and t-to, compared with the results in Table 1, for the survey area.

The algorithm used to generate A and t-to in Table 1 also yielded values for μ (for all but site O) independent of the measured thermal conductivity values. Interestingly, the mean value for thermal diffusivity (μ) based on equation 1 and the temperature/depth data is 0.0050 square cm/second.

This is in good agreement with the value of 0.0047 sq.cm/sec. (Table 2) using measured thermal conductivity and Bullard (1963). The standard deviation is nearly twice as large as the Table 2 value, again suggesting better overall agreement of the Table 2 parameters with observation, despite the higher RMS error.

We have used the mean values of A and t-to, together with $T_s=1$ deg. C., $\mu=0.005$ sq. cm/second and $GRAD=4E-4$ (4mK/m) to compute a typical Barents Sea temperature profile for the upper 10 meters of sediment at the time of the 1985 field survey (Figure 1).

Quantitative modeling of the temperature data indicates that the bottom water temperatures at the time of the 1985 survey were not as close to the seasonal maxima as assumed and, in fact, they were relatively close to the mean values ($t-t_o=0$). The temperature profiles shown by the dashed curves in Saettem et al (1985), therefore, somewhat exaggerate the positive temperature excursion from the shallowest measured depth points to the sediment/water interface (the interval without data).

We have used the model data illustrated in Fig. 1 to test further the sensitivity of the algorithms mentioned previously to the various input parameters/constraints. The initial result, that the variability in the parameters A and t-to over the survey rendered the likelihood of determining the steady state gradient (GRAD) from only three temperature points at one, two and three meters respectively, came as no surprise. This was, unfortunately, verified by attempts to derive GRAD for sites H and W. Similar attempts on site I, which has a fourth temperature value at six meters, met with somewhat better results. For site I, several variations in the control parameters consistently yielded gradient values between 60 and 70 mK/m (RMSE = 0.02) which is 50% higher than the gradient at site V. Only about 20% of this difference can be accounted for by the thermal conductivity being higher at site V. We draw attention to this result, but regard its reliability as questionable.

As was discussed earlier, much of the uncertainty in defining A and t-to over the survey area would be eliminated with use of a fully-configured heat flow probe by first performing shallow probe measurements at the downhole sites to determine the parameters exactly as described per Table 1. This would be required, in any event, since it cannot be assumed that A and t-to for each site remain fixed from year to year. Since these parameters are linked to the hydrography of the area and are related to depth and latitude, a study of their distribution should improve interpolation of their values between control points. Accepting, however, their variations as indicated in Table 2,

we can use the test data (Fig. 1) to estimate conservatively the minimum resolution attributable to a relatively modestly configured heat flow probe.

We accomplished this by generating test temperatures at 0, 0.5, 1, 2, 3 and 4 meters (the probable shallow probe depth limit) and rounding the temperature values to only two decimal places (± 0.01 deg. C. resolution). Assuming that it is possible to define μ to within 5% by direct measurement, we stepped through values of t -to for a full range of 26 days (one standard deviation). All other parameters were allowed to vary freely. A minimum RMS error resolution resulted (RMSE = 0.001) in a 50% departure from the known value for GRAD. This translates to a heat flow resolution of about 0.8 HFU.

DISCUSSION

This resolution is adequate for such tasks as mapping salt related thermal anomalies which may cause local heat flow to vary by as much as a factor of five. Salt deposits/diapirs exist in several key areas of the Barents Sea. The transition to oceanic heat flow values west of our 1985 survey area represents perhaps a factor of two increase in heat flow over a relatively short distance, with potentially great implication to hydrocarbon generation. Furthermore, heat flow anomalies associated with vertical fluid migration are known to vary by a factor of ten (Bredenhoeft and Papadopoulos, 1965; Zielinski and Bruchhausen, 1983; Zielinski et al, 1985). Hence, for example, a surface geochemical anomaly believed to be associated with an active hydrocarbon seep should also exhibit a large heat flow anomaly. For such purposes, the large data density possible with shallow probe surveys and cost far exceed the loss of resolution compared to downhole measurements. Combined heat flow/surface geochemical surveys can be run from a single vessel, with little delay to either individual program, with an advantage that the heat flow data can be processed aboard ship practically in real time.

Test data results based on the 1985 survey data further indicate a dramatic increase in potential heat flow resolution for "state of the art" measurements (± 0.001 deg. C.) and for an increased number of temperature/depth points, both of which are possible. Alternate penetration devices such as vibrocorers (Fig. 2) and seabed penetrometers (Figs. 3 and 4) may be fitted with heat flow instrumentation and adapted for penetration in the 4 - 10 meter range. The result of the increased penetration is that nearly a full wavelength of the seasonal transient is defined (see Fig.1). Test data rounded to three decimal places (0.001 deg. C. resolution), from 0 - 10 meters at 1 meter intervals (8 temperature/depth points) suggest a resolution of better

than 20% or 0.3 HFU.

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Data report

III



IKU

SINTEF-GRUPPEN