

Volgian–Ryazanian ‘hot shales’ of the Bo Member (Farsund Formation) in the Danish Central Graben, North Sea: stratigraphy, facies and geochemistry

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Upper Jurassic – lowermost Cretaceous marine mudstones represent the most significant source of hydrocarbons in the Central and Northern North Sea. Of particular importance in the Danish sector of the Central Graben is a succession of radioactive ‘hot shales’ referred to the Bo Member, in the upper levels of the Farsund Formation (Kimmeridge Clay Formation equivalent). This mudstone-dominated succession is typically 15–30 m thick and has a total organic carbon (TOC) content of 3–8%, though locally exceeding 15%. Although truncated on some structural highs, the Bo Member is a persistent feature of the Danish Central Graben. Lateral variation in both thickness and organic richness is attributed to intrabasinal structural topography and to the location of sediment input centres.

Detailed study of the dinoflagellate cyst biostratigraphy of 10 wells indicates that the onset of enhanced organic carbon burial began in the middle–late Middle Volgian in this portion of the Central Graben. The Bo Member, representing the peak of organic carbon enrichment, is largely of Early Ryazanian age. Core data (Jeppe-1, E-1 wells) indicate that the organic-rich shales of the Bo Member are not wholly of hemipelagic origin, as commonly assumed, but may locally be dominated by fine-grained turbidites. Absence of bioturbation, well-preserved lamination and high TOC values suggest that bottom waters were predominantly anoxic although the presence of in-situ benthic bivalves at discrete horizons in the E-1 well suggests that suboxic conditions prevailed on occasion. The Bo Member is a good to very good source rock, showing very high pyrolysis yields (10–100 kg HC/ton rock) and Hydrogen Index (HI) values in the range 200–600. In particular, the Bo Member is characterised by an abundance of 28,30 bisnorhopane (H28), a compound that is indicative of anoxic environments.

These new data from the Danish sector of the Central Graben are compatible with the model of Tyson *et al.* (1979) in which the accumulation of organic-rich mudstones was controlled primarily by bottom-water anoxia beneath a stratified watermass. A number of factors probably contributed to the development of watermass stratification, both intrinsic such as the tectonic morphology of the graben system and extrinsic including climate and sea-level stand.

Keywords: Danish Central Graben, North Sea, Volgian–Ryazanian, organic-rich mudstones, marine source rock, depositional processes, geochemistry, anoxia

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One of the most important factors in the success of the North Sea basin as a major oil province is the presence of the thick succession of marine mudstones of Late

Jurassic to earliest Cretaceous age. This succession, which comes under various stratigraphic guises (Figs 1, 2), is generally accepted to have been the source of much

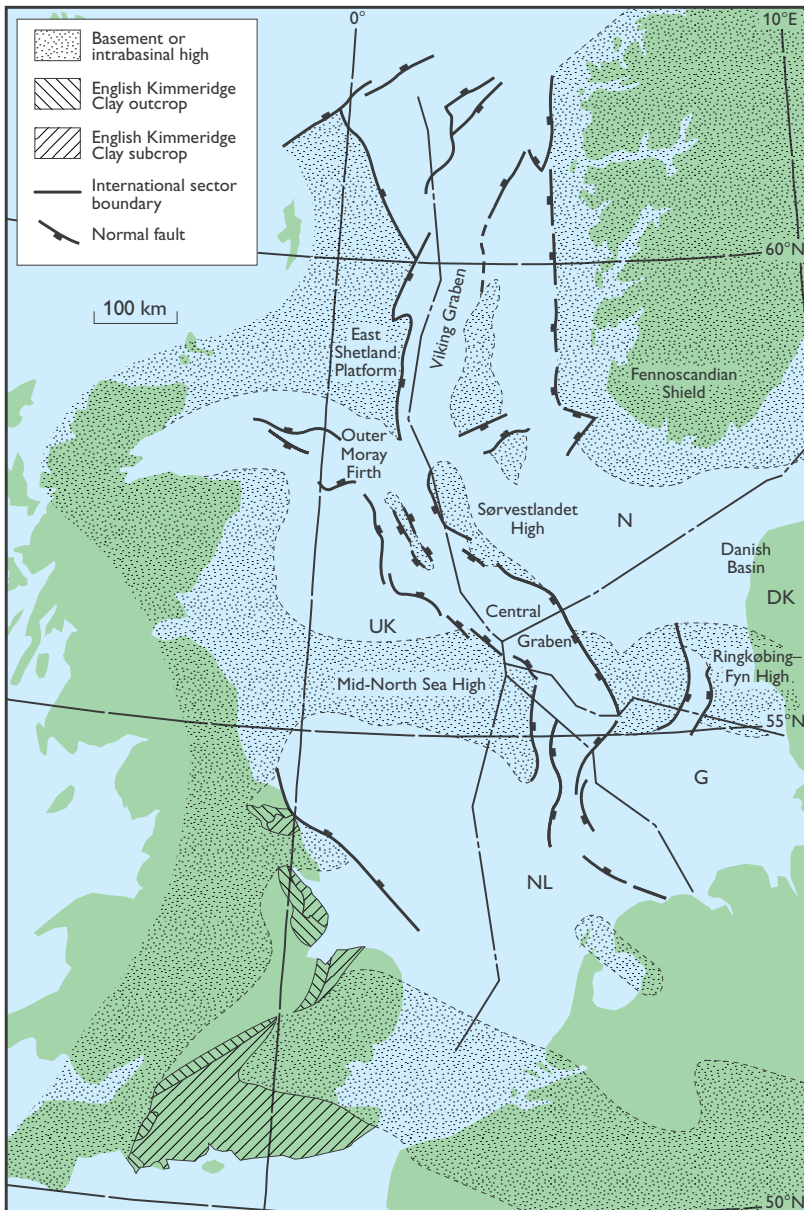


Fig. 1. Generalised Jurassic tectonic setting of the North Sea region (based on Doré *et al.* 1985). **DK**, Denmark; **G**, Germany; **N**, Norway; **NL**, the Netherlands; **UK**, United Kingdom.

of the hydrocarbons in the Central and Northern North Sea (Barnard & Cooper 1981; Cornford 1994, 1998; Kubala *et al.* 2003) and also forms the seal in many of the Jurassic oilfields in this area (e.g. South Brae, Turner *et al.* 1987). In well sections, this basinal mud-dominated succession is characterised by alternations of 'hot' and 'cold' mudstone intervals (Price *et al.* 1993). This refers to relative levels of radioactivity emitted by the mudstones, as recorded by the gamma-ray log. In general, such variation in the levels of gamma radioactivity in marine mudstones is assumed to reflect the organic richness of the sediment. Of particular note in the Central Graben of the North Sea are the 'hot shales' of Volgian to Ryazanian age which have been formalised

as the Clay Deep Member (Kimmeridge Clay Formation) in the Dutch sector and the Mandal Formation in the Norwegian sector (Fig. 2; Vollset & Doré 1984; van Adrichem Boogaert & Kouwe 1993). In the Danish sector, partially equivalent organic-rich mudstones have been described as the 'hot unit', an informal member of the Farsund Formation (Jensen *et al.* 1986); this unit is formally defined as the Bo Member of the Farsund Formation in this volume (Michelsen *et al.* 2003, this volume). In the UK sector of the Central North Sea, equivalent organic-rich mudstones are not given formal lithostratigraphic status within the Kimmeridge Clay Formation (see the 'hot shales' of Donovan *et al.* 1993). The aim of this paper is to integrate data from various

Age		Northern Dutch Central Graben	Danish Central Graben	Norwegian Central Graben	UK Central Graben	Outer Moray Firth	South Viking Graben (UK)
Cretaceous	Valanginian	Vlieland Claystone Fm	Cromer Knoll Group	Cromer Knoll Group	Cromer Knoll Group	Cromer Knoll Group	Cromer Knoll Group
	Ryazanian	Clay Deep Mb	Bo Member	Mandal Fm			Kimmeridge Clay Fm
Late Jurassic	Volgian	Scruff Greensand Fm Kimmeridge Clay Fm	Farsund Fm Poul Fm	Farsund Fm Ula Fm	Kimmeridge Clay Fm	Kimmeridge Clay Fm	Brae Fm
	Kimmeridgian		Heno Fm			Piper Fm	
	Oxfordian	Upper Graben Fm Middle Graben Fm	Lola Fm	Haugesund Fm	Fulmar Fm	Heather Fm	Heather Fm

Fig. 2. Generalised Upper Jurassic lithostratigraphy in the North Sea rift system from the Dutch sector of the Central Graben in the south to the Moray Firth and southern Viking Graben in the north (Vollset & Doré 1984; van Adrichem Boogaert & Kouwe 1993; Richards *et al.* 1993; Lott & Knox 1994; Michelsen *et al.* 2003, this volume). Sand-rich formations are indicated by a **stipple** ornament, lithostratigraphically defined ‘hot shales’ by **oblique** shading.

disciplines and thus to provide an up-to-date assessment of the nature, distribution and origin of the ‘hot shales’ of the Bo Member in the Danish Central Graben.

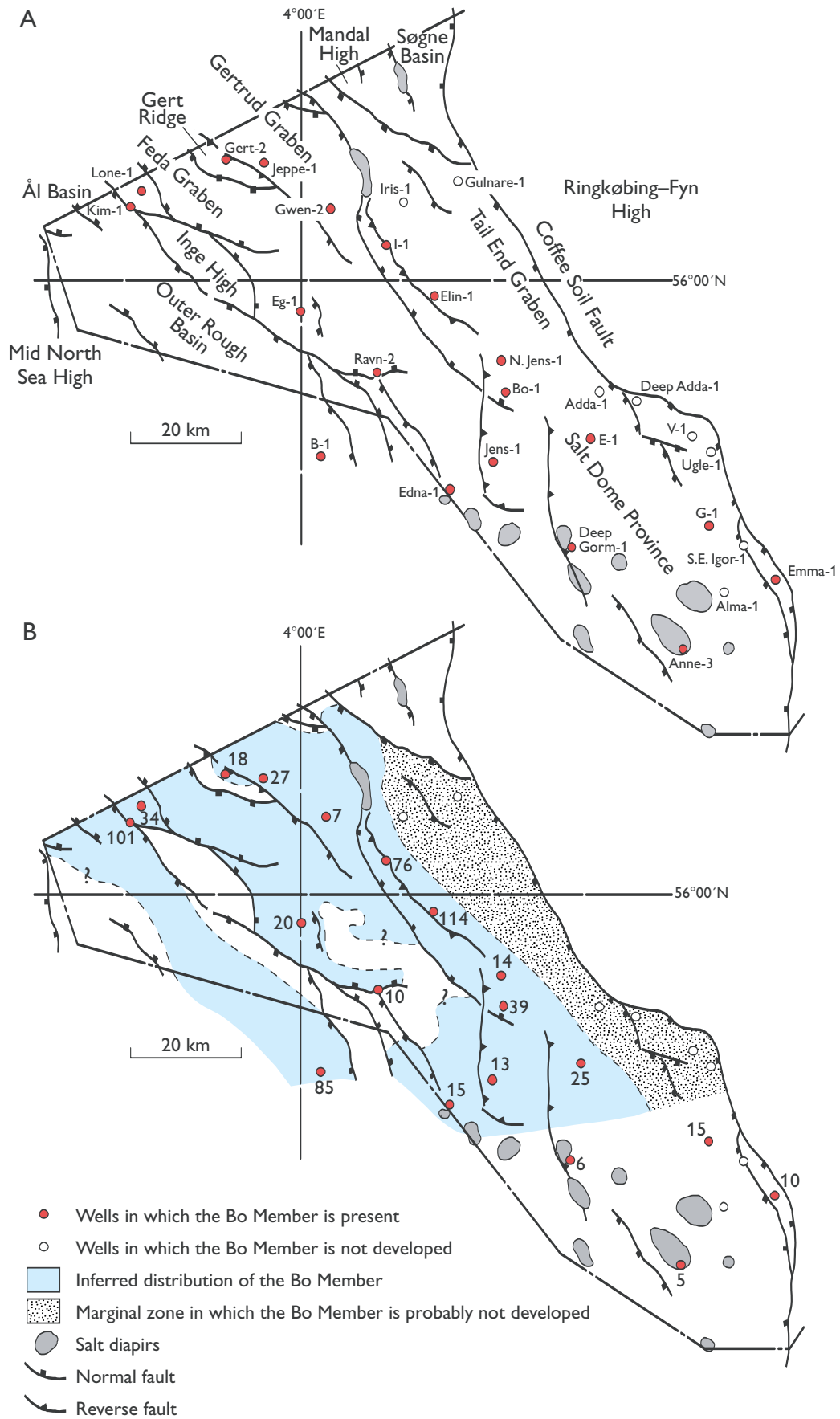
Geological setting

The Danish offshore area extends westwards to include a segment of the North Sea Central Graben, a complex Mesozoic rift system that trends roughly NNW–SSE (Fig. 1). Although probably following Palaeozoic lineaments (Glennie 1990), a discrete rift system is thought to have first developed in the earliest Triassic, trending north–south (Ziegler 1988, 1990; Sundsbø & Megson 1993). The dominant NW–SE structural trends in the Danish Central Graben developed during the Late Jurassic when the rift was at its most active, with the accumulation of up to 4 km of (compacted) sediment in the most rapidly subsiding sub-basins of the Danish Central Graben (Møller 1986; Sundsbø & Megson 1993; Japsen *et al.* 2003, this volume). The Early Cretaceous saw the transition from this phase of active extension and rapid fault-controlled subsidence in the Late Jurassic

to the regional subsidence pattern, centred on the axial graben system, that characterised the Late Cretaceous and Cenozoic.

Late Jurassic – earliest Cretaceous structural evolution

The Late Jurassic was characterised by the development and pronounced differential subsidence of successive half-grabens within the complex rift basin of the Central Graben. Subsidence and sedimentation was focussed on the eastern and southern area (Søgne Basin, Tail End Graben and Salt Dome Province; Fig. 3) during the latter part of the Middle Jurassic but extended north-westwards in the Late Jurassic due to both the overall sea-level rise and the development of secondary depocentres in successive half-grabens (Møller 1986; Damtoft *et al.* 1992; Andsbjerg & Dybkjær 2003, this volume; Møller & Rasmussen 2003, this volume). The Feda Graben in the Danish sector was probably initiated in the early Late Jurassic (?Oxfordian) and formed the dominant sediment depocentre in the area during the



Kimmeridgian (Söderström *et al.* 1991; Johannessen & Andsbjerg 1993; Rasmussen 1995). The Gertrud Graben developed as a discrete depocentre during the Late Kimmeridgian whereas onlap onto the Mid North Sea High to the west began in the Volgian (Møller 1986; Damtoft *et al.* 1992). In this western area, two discrete depocentres (the Ål and Outer Rough Basins) were initiated in the latest Jurassic and formed important sediment sinks in the Early Cretaceous (Japsen *et al.* 2003, this volume). During the latter part of the Late Jurassic, therefore, the Danish Central Graben was segmented into a number of NW–SE-trending depocentres, separated by elongate highs or broad plateaus (Johannessen & Andsbjerg 1993; Andsbjerg & Dybkjær 2003, this volume; Johannessen 2003, this volume).

This history of protracted extension during the Middle and Late Jurassic was interrupted in the latest Jurassic – earliest Cretaceous by a complex tectonic phase that was essentially extensional in character but involved block rotation associated with localised compression, reverse faulting and uplift (Rasmussen 1995; Møller & Rasmussen 2003, this volume). This end-Jurassic tectonic phase coincided broadly with deposition of the organic-rich Bo Member and thus is of direct relevance to this study.

Stratigraphy

History and stratigraphic status

The organic-rich ‘hot shales’ of the Bo Member occur within the uppermost levels of the Farsund Formation

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Fig. 3. **A:** Late Jurassic tectonic framework of the Danish Central Graben (modified from Damtoft *et al.* 1992) showing the location of released wells in which the Bo Member is recognised. The Bo Member is not recognised in the wells indicated by open circles in the east of the area, although the same stratigraphic interval is represented in these wells. Note that the Ål and Outer Rough Basins, although depicted on this map, were areas of active subsidence primarily in the Early Cretaceous (Japsen *et al.* 2003, this volume). **B:** Map showing the thickness (in metres) of the Bo Member in well sections and the lateral distribution of this member (**blue**) as deduced from well and seismic data. **Stipple** indicates the inferred area in which the Bo Member is not developed due to siliciclastic dilution of the organic matter proximal to the Ringkøbing–Fyn High. The distribution of the Bo Member is not inferred for the southern area (Salt Dome Province) due to sparse data points and the thin, irregular development of the ‘hot shales’ in this region.

(Fig. 2). Originally described informally by Jensen *et al.* (1986) as the ‘hot unit’, the stratigraphy of this unit was subsequently discussed briefly by Michelsen & Wong (1991) and the hydrocarbon source rock characteristics were reported by Østfeldt (1987) and Damtoft *et al.* (1987, 1992). A detailed study of these deposits was undertaken by Bojesen-Koefoed (1988). As noted earlier, Michelsen *et al.* (2003, this volume) formally define the ‘hot unit’ of Jensen *et al.* (1986) as the Bo Member in an accompanying paper.

Log character

The Bo Member is recognised primarily on the basis of log character since the gross lithological contrast between this member and the remainder of the Farsund Formation is often slight and rarely detectable in ditch cuttings. Jensen *et al.* (1986) described the unit from the Bo-1 well at the southern end of the Tail End Graben (Figs 3, 4). This description relied heavily on the gamma-ray log which thus forms the essential criterion for its recognition.

In Bo-1, the background gamma-ray values of the Farsund Formation mudstones are in the range 75–100 API (Fig. 4); the Bo Member shows values ranging from 120 to 160 API with marked upward shifts to higher and lower values defining the lower and upper boundaries of the member respectively. Although generally high, gamma-ray values may vary significantly within the Bo Member; in Bo-1 and many other wells, upward-decreasing gamma trends are evident, typically 3–5 m thick and separated by intervals showing more consistently high gamma-ray values. A marked feature of the Bo-1 section is the upward increase in gamma-ray values beneath the Bo Member over an interval of about 50 m (8724–8561 ft). This ‘warming-upwards’ interval (henceforth referred to as the W-U interval) is also a feature of many other wells (see below). Variation in the degree of development of this W-U interval in relation to the ‘hottest’ interval can result in ambiguity in locating the base of the Bo Member. In recognising the Bo Member throughout the Danish Central Graben, the Bo-1 gamma log pattern was used as the basic reference. It is important to note that the Bo Member of the Danish sector of the Central Graben is more restricted in its definition than the partially equivalent Mandal Formation in the Norwegian sector and the Clay Deep Member of the Dutch sector (Fig. 2; Dybkjær 1998). The bases of these units are defined at the point at which the overall gamma values begin to increase (i.e. at the base of

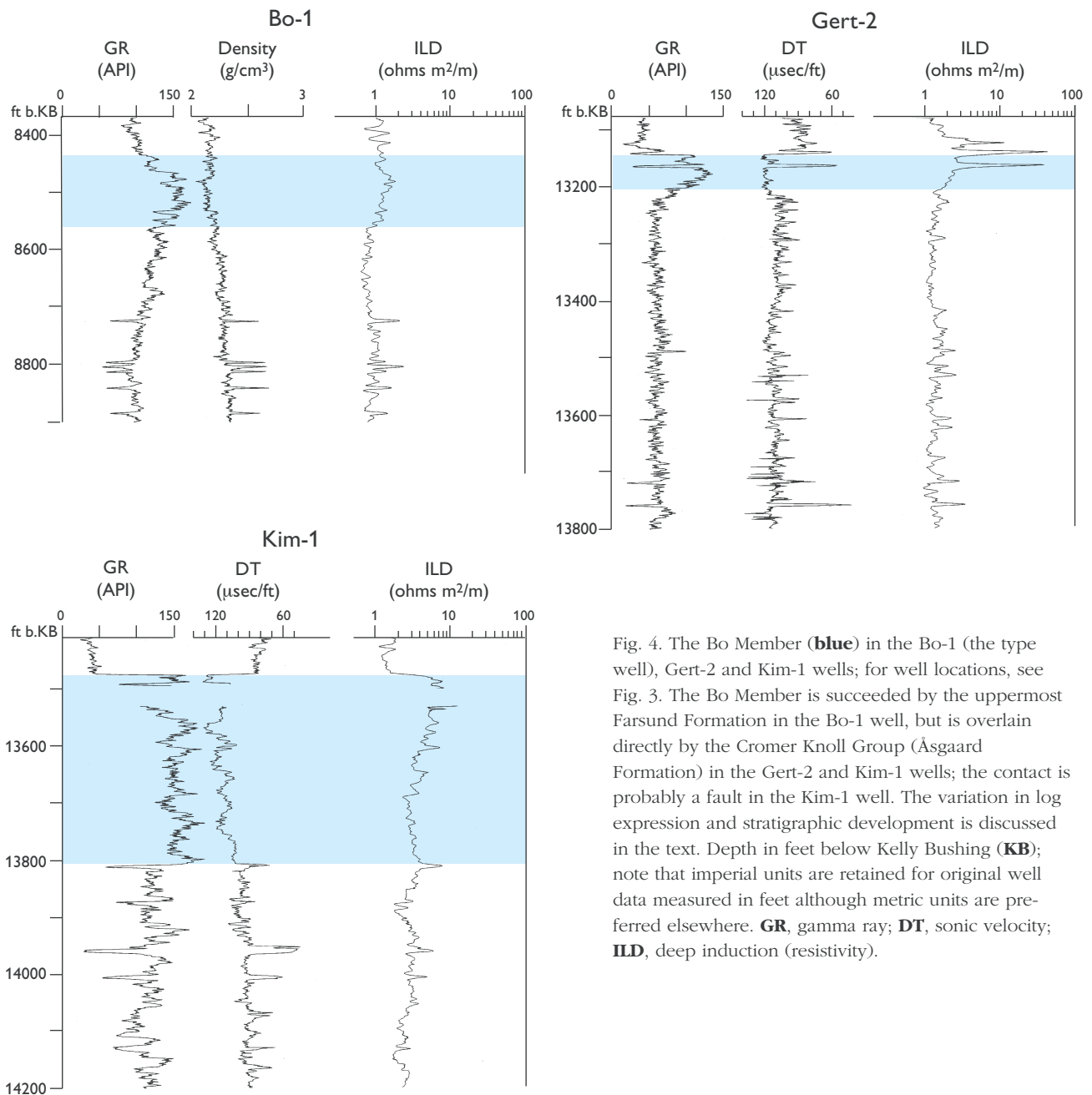


Fig. 4. The Bo Member (**blue**) in the Bo-1 (the type well), Gert-2 and Kim-1 wells; for well locations, see Fig. 3. The Bo Member is succeeded by the uppermost Farsund Formation in the Bo-1 well, but is overlain directly by the Cromer Knoll Group (Åsgard Formation) in the Gert-2 and Kim-1 wells; the contact is probably a fault in the Kim-1 well. The variation in log expression and stratigraphic development is discussed in the text. Depth in feet below Kelly Bushing (**KB**); note that imperial units are retained for original well data measured in feet although metric units are preferred elsewhere. **GR**, gamma ray; **DT**, sonic velocity; **ILD**, deep induction (resistivity).

the W-U interval) and they are succeeded by Cretaceous strata of the Cromer Knoll and Rijnland Groups (Völlset & Doré 1984; Michelsen & Wong 1991); the Mandal Formation and the Clay Deep Member clearly span a greater stratigraphic interval than the Bo Member as defined by Michelsen *et al.* (2003, this volume).

Spectral gamma-ray logs are not available for the Bo-1 well but the example illustrated from the Lone-1 well (Fig. 5) shows an overall gamma log pattern that is comparable with the Bo-1 section. It is clear from the spectral log that the increase in the overall gamma-ray values both beneath and within the Bo Member is the

result of an increased content of uranium (Jensen *et al.* 1986); the thorium and potassium values show little variation. As demonstrated by Swanson (1961), uranium in sedimentary rocks is typically bound to organic matter and the positive correlation between gamma radioactivity and total organic carbon (TOC) in the Bo Member is marked in most wells (see Fig. 14; Damtoft *et al.* 1992, fig. 5). Uranium may also be concentrated in biogenic phosphate material, however, and since such debris is common in the organic-rich mudstones of the Bo Member, this may provide an additional contribution to the total gamma response.

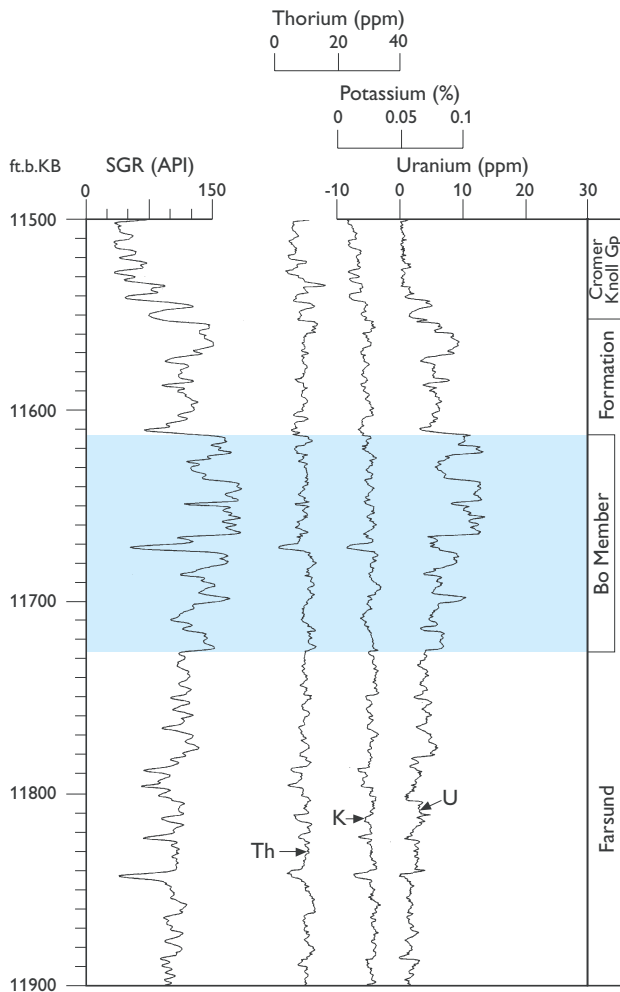


Fig. 5. Spectral gamma log through the Bo Member in the Lone-1 well. Note that the increased gamma radioactivity exhibited by this interval is attributable solely to the contribution from uranium. **SGR**, total gamma-ray log; **Th**, thorium; **K**, potassium; **U**, uranium.

Resistivity logs through this stratigraphic interval often show anomalously high readings which correspond, at least in part, to the Bo Member as defined on the gamma log. This gross correspondence is not surprising since resistivity logs are often used to recognise zones of organic richness (Meyer & Nederlof 1984; Passey *et al.* 1990). In Bo-1, the base of the Bo Member is marked by an abrupt increase in resistivity readings which remain relatively high throughout the unit (Fig. 4). It is noteworthy, however, that these high resistivities continue upwards to the top of the Farsund Formation, above the upper boundary of the Bo Member as defined by

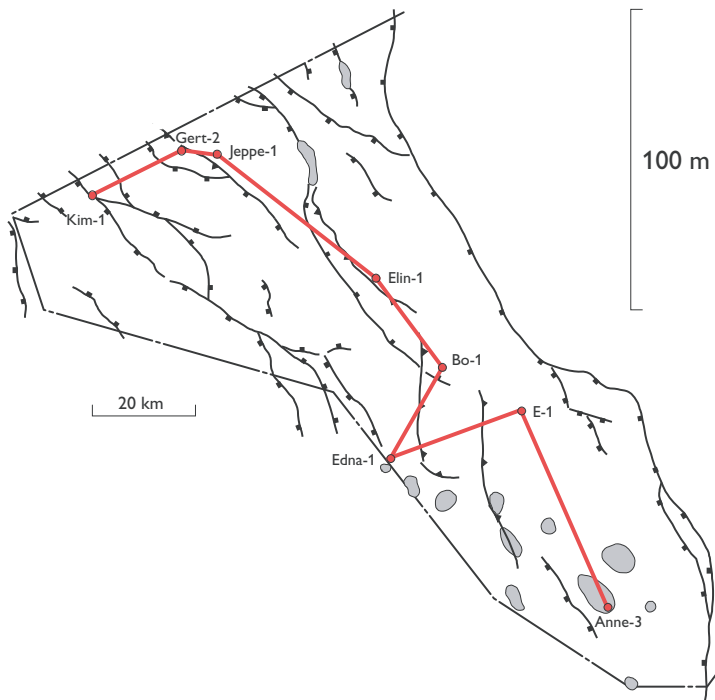
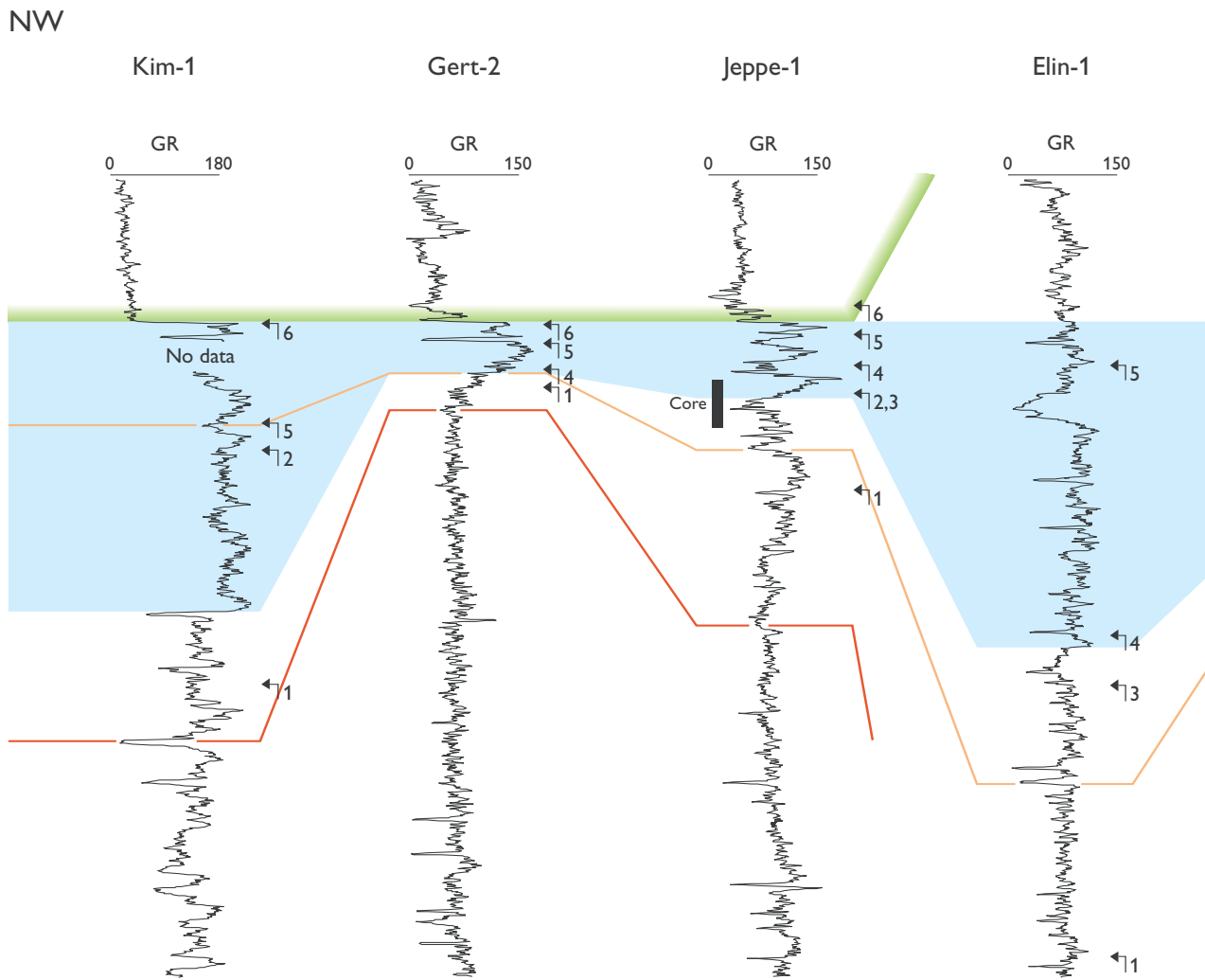
the gamma log. In a number of wells, the resistivity logs display a stepped increase, corresponding roughly to the W-U interval on the gamma log beneath the Bo Member. It should be acknowledged, however, that the marked log shifts on the gamma and resistivity logs often do not coincide in detail, despite the broad correspondence, and the boundaries defined on the gamma log are considered to take priority, following Jensen *et al.* (1986).

In Bo-1, the sonic and density–neutron logs do not vary substantially through this stratigraphic interval, indicating that gross lithological variation between the Bo Member and the host Farsund Formation is small; bulk densities and sonic velocities are slightly lower and neutron log values slightly higher in the ‘hottest’ zones of the Bo Member compared with the Farsund Formation in general.

Distribution and regional development

The Bo Member is recognised widely in the Danish Central Graben in wells where this portion of the upper Farsund Formation is preserved (Fig. 3). The uppermost Farsund Formation is often truncated on structural highs (e.g. the inverted Søgne Basin, Fig. 3) such that the original extent and variation in development of the Bo Member in the various subbasins is difficult to evaluate. Furthermore, although forming part of the uppermost seismic sequence of the Farsund Formation mapped by Møller (1986), the Bo Member alone cannot be differentiated on seismic data. Knowledge of the lateral extent and variation in development of the member is thus fragmentary.

It is possible, however, to make certain broad observations concerning both the distribution and lateral variation in the stratigraphic development of the Bo Member. Firstly, it is absent in a number of wells flanking the eastern margin of the Danish Central Graben, from Gulnare-1 in the north to Alma-1 in the south (Fig. 3). In these wells, biostratigraphic data indicate the presence of the stratigraphic interval occupied by the Bo Member elsewhere yet the anomalously high gamma values characteristic of this member are not observed. It is likely that the lack of development of organic-rich mudstones in these marginal wells is largely the result of increased siliciclastic input and consequent dilution of the organic matter. It is noteworthy that discrete sand-rich intervals are characteristic of the uppermost Jurassic – lowermost Cretaceous in a number of these wells and indeed are recognised at formation

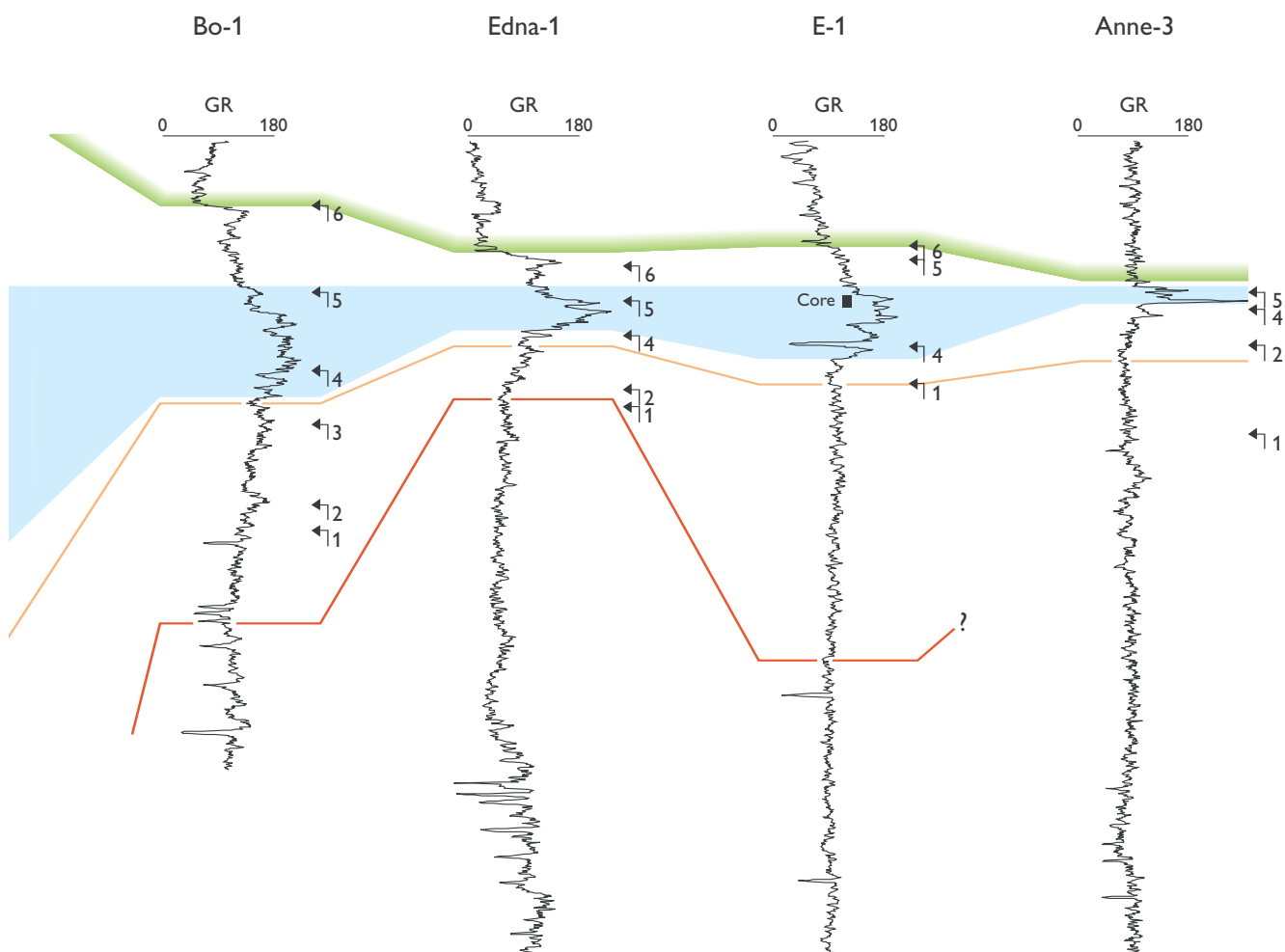


Last occurrence datum, dinoflagellate cyst species:

- 6 *Dichadogonyaulax culmula*/*Dingodinium?* *spinosum*
- 5 *Rotosphaeropsis thula*
- 4 *Amphorula expirata*
- 3 *Egmontodinium polyplacophorum*
- 2 *Dichadogonyaulax panna*/*Glossodinium dimorphum*
- 1 *Senoniasphaera jurassica*

- Base Cromer Knoll Group
- Bo Member
- Base Ryaz-1 sequence boundary
- Base Volg-4 sequence boundary

Fig. 6. Variation in the development of the Bo Member in the Danish Central Graben, illustrated by gamma logs (datum: top Bo Member). Sequence boundaries 'Base Volg-4' and 'Base Ryaz-1' are from Andsbjerg & Dybkjær (2003, this volume).



level (Poul and Vyl Formations; Michelsen *et al.* 2003, this volume) in the Deep Adda-1, V-1 and Ugle-1 wells. Furthermore, the Bo Member is poorly defined, and only tentatively recognised, in several wells that lie adjacent to this marginal belt. In the Elin-1 well in the Tail End Graben (Fig. 3), for example, biostratigraphic data indicate that the Upper Volgian to Ryazanian succession is highly expanded relative to other wells (Fig. 6). The Bo Member in this well is thick (114 m compared with 39 m in Bo-1) but shows only slightly higher gamma values than the background Farsund Formation.

The thickness of the Bo Member varies greatly in the Danish Central Graben, from less than 10 m in the southern Salt Dome Province to over 100 m in the western part of the Danish Central Graben (Fig. 3). Although this overall trend appears significant, thickness variation over much of the Danish Central Graben does not show any systematic regional trend. It was probably con-

trolled by local factors such as structural position within individual subbasins and local variations in the rate of sediment supply. The complexity is exemplified in the central portion of the Danish Central Graben where the Bo Member is 39 m and 25 m thick, respectively, in the Bo-1 and E-1 wells but under 15 m in nearby wells such as North Jens-1 and Jens-1 (Fig. 3). The influence of structural position (and hence proximity to depocentres) is illustrated by comparing the Gert-2 well on the Gert Ridge (18 m) with the Jeppe-1 well within the Gertrud Graben (27 m) and the I-1 well in the major depocentre of the Tail End Graben (76 m). It should be emphasised again, however, that the Bo Member, although potentially thick, is only weakly developed in the major depocentre of the Tail End Graben.

The log character of the Bo Member also varies through the Danish Central Graben (Figs 4, 6), often in association with variation in thickness. On structural

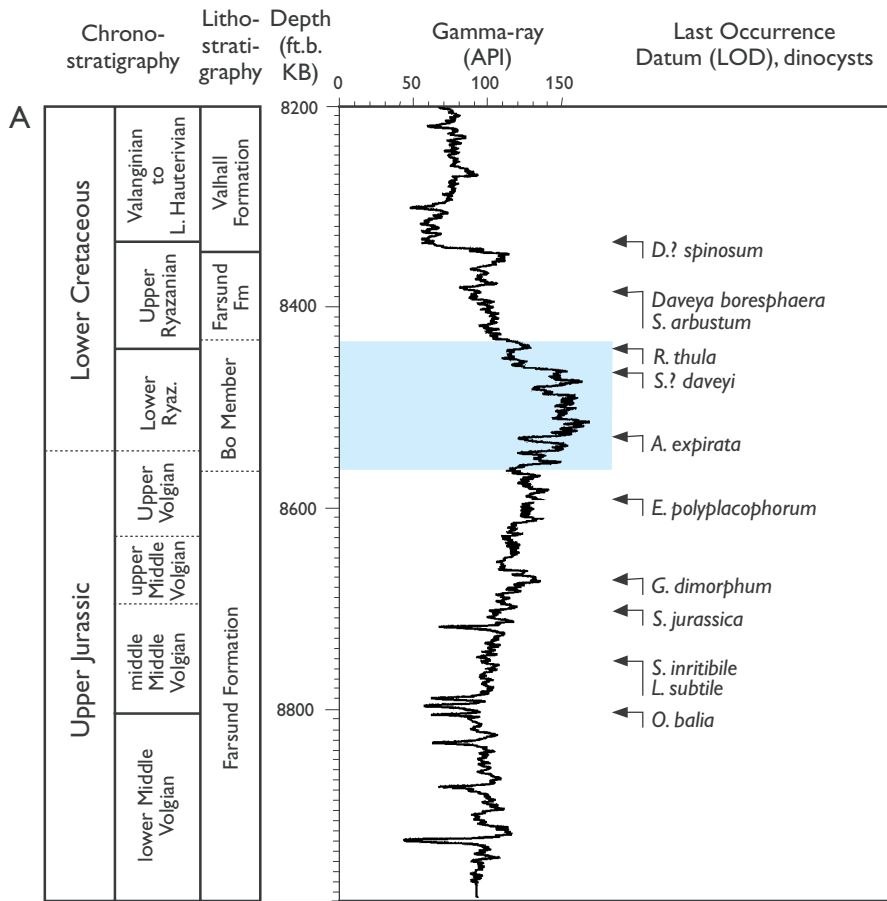
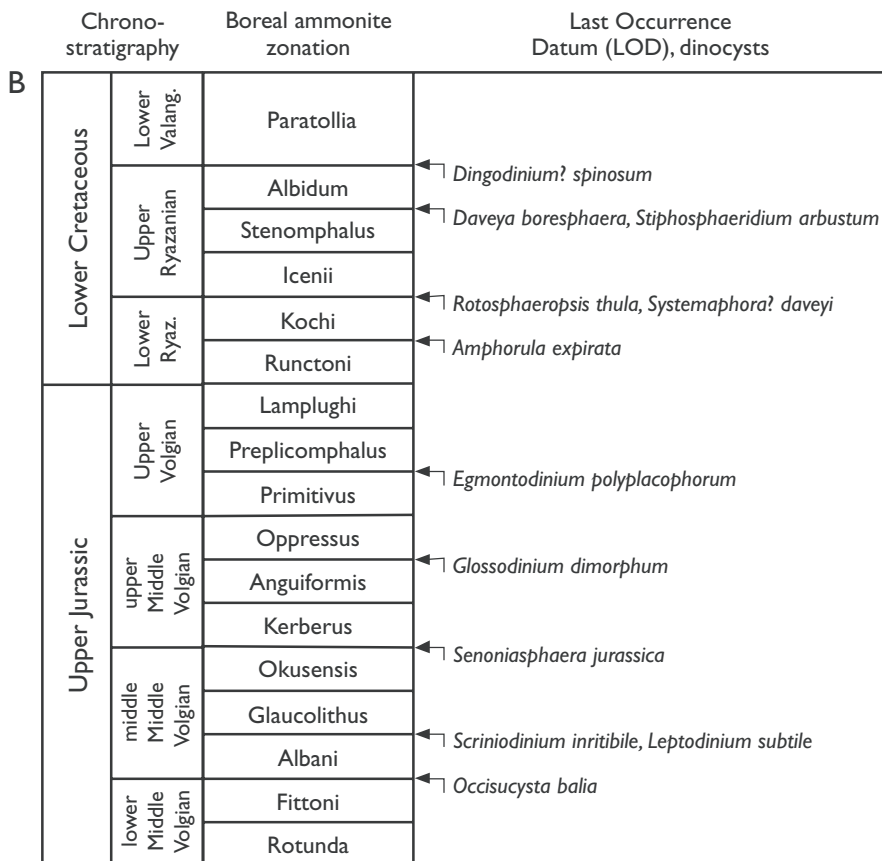


Fig. 7. **A:** Biostratigraphy of the Bo-1 well, the type well of the Bo Member, based on analysis of ditch cuttings. The stippled chronostratigraphic boundaries cannot be accurately positioned in the succession on the basis of the palynology (Fig. 7B). **B:** Correlation of dinocyst 'tops' to the standard Boreal ammonite zonation compiled from Costa & Davey (1992) and Riding & Thomas (1992).



highs, such as the Gert Ridge (Gert-2, Fig. 4), both the Bo Member and the underlying W-U interval may be thin in comparison to the type Bo-1 well, although still well-defined on petrophysical logs. In a number of wells (e.g. Jeppe-1, Fig. 6), an important erosional surface has been recognised beneath the Bo Member, truncating the W-U interval and in places succeeded by deep-water sandstones. This surface has been interpreted as a sequence boundary of regional significance by Andsbjerg & Dybkjær (2003, this volume; see also below).

Biostratigraphy

The Farsund Formation in the Danish Central Graben has a maximum age range of Kimmeridgian to Ryazanian. The Bo Member (Michelsen *et al.* 2003, this volume) occurs in the uppermost Farsund Formation, in the interval broadly dated as latest Middle Volgian to Early Ryazanian (Poulsen 1991). Previous work on both ammonites and dinoflagellate cysts (hereafter referred to as dinocysts) from the cored portion of the Bo Member in the E-1 well (Fig. 6) indicated an Early Ryazanian age (Kochi Chronozone) for at least the uppermost Bo Member (Birkelund *et al.* 1983). In the present study, the biostratigraphy of the Bo Member and the immediately underlying and overlying Farsund Formation was investigated by means of palynological analysis of core where available (E-1, Jeppe-1), sidewall cores and ditch cuttings from a total of 10 wells. The results of this study are summarised in Figures 6–8; detailed biostratigraphic data from Bo-1 in comparison with the type well of the Mandal Formation (7/12-3A) have been presented by Dybkjær (1998).

A number of key dinocysts (Fig. 8C–H) can be used to bracket the Bo Member over much of the Danish Central Graben; these dinocysts are well-known as stratigraphically useful species in the North Sea region (Davey 1979, 1982; Riding 1984; Riding & Thomas 1992). As illustrated for Bo-1 (Fig. 7), the top of the Bo Member typically occurs immediately above the Last Occurrence Datum (LOD) of *Rotosphaeropsis thula* whereas the base falls between the LOD of *Amphorula expirata* and the LOD of *Egmontodinium polyplacophorum*. The interval showing the highest gamma values is bracketed by the LOD's of *R. thula* and *A. expirata*, corresponding to the Kochi Chronozone of the Early Ryazanian (Fig. 7). The results from the Bo-1 well, supported in general by the regional data (Fig. 6), indicate a maximum age range for the Bo Member of Late Volgian

(Preplicomphalus Chronozone) to middle Late Ryazanian (Stenomphalus Chronozone). The point at which the gamma values begin to increase, i.e. the base of the W-U interval underlying the Bo Member (equivalent to the base of the Mandal Formation in the Norwegian sector), corresponds roughly to the LOD of *Senoniasphaera jurassica* in Bo-1 and many other wells (Figs 6, 7), giving a middle–late Middle Volgian age for the onset of increased preservation of organic carbon in the succession.

Over much of the Danish Central Graben, the biostratigraphy of the Bo Member is closely comparable to that in Bo-1 (Fig. 6), despite the variation in thickness and local problems of definition (see discussion above). In most wells, the top of the Bo Member lies between the LOD of *Dichadogonyaulax culmula/Dingodinium? spinosum* and that of *R. thula* as observed in the type well (Figs 6, 7), although deviations from this pattern were recorded (compare Gert-2 with E-1). Over most of the transect in Figure 6, from Gert-2 in the north to Anne-3 in the south, the base of the Bo Member coincides roughly with the LOD of *A. expirata*. Key Volgian dinocysts are absent from the underlying Farsund Formation in the Edna-1 and E-1 wells; in the latter case this may be attributable to an erosional hiatus near the base of the Bo Member (Fig. 6; Andsbjerg & Dybkjær 2003, this volume).

In the Jeppe-1 well, the ranges of *Dichadogonyaulax? pannea/Glossodinium dimorphum* and *E. polyplacophorum* extend into the Bo Member and share a common LOD (Fig. 6). It is thought likely that this anomaly can be attributed to reworking of these species, extending their ranges upwards. As described in detail below, the strata underlying the Bo Member in Jeppe-1 are characterised by abundant evidence of slumping and sediment gravity flow and the potential for redeposition of older sediments must be considered high. In the Gert-2 well on the Gert Ridge, in contrast, these dinocysts are not recorded and uppermost Middle to Upper Volgian strata may be absent in this well; a sequence boundary is recognised at this level in Gert-2 by Andsbjerg & Dybkjær (2003, this volume; Fig. 6). Thus, the anomalous biostratigraphic results from this northern portion of the Danish Central Graben are perhaps best explained by local erosion of structurally positive regions during the latest Volgian with concomitant redeposition in adjacent lows; this subject is discussed further below.

As noted earlier, the Bo Member is well-developed and anomalously thick in the Kim-1 and B-1 wells in the westernmost part of the Danish Central Graben (the incipient Ål and Outer Rough Basins). In addition, the

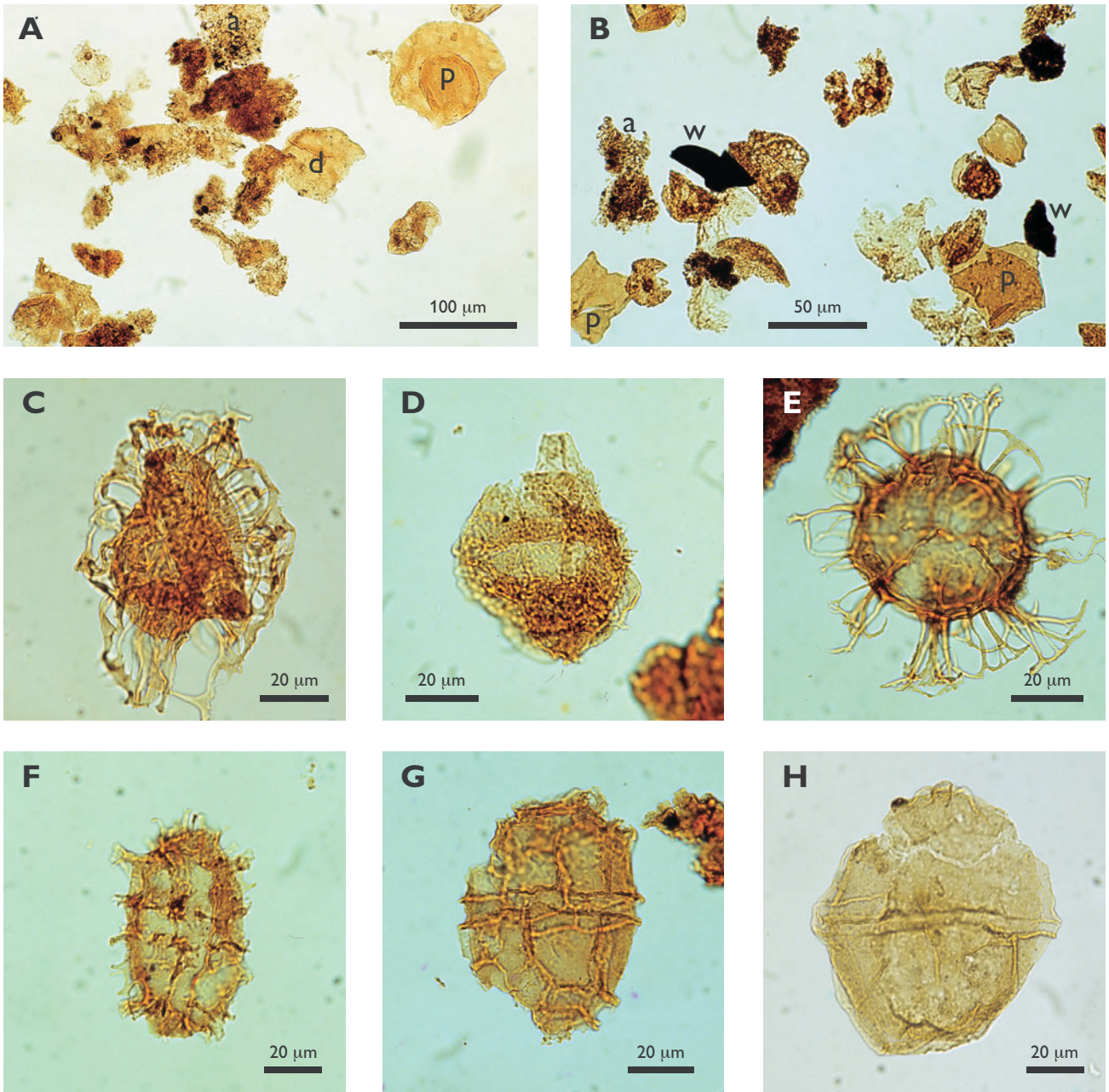


Fig. 8. Palynofacies assemblages (**A**, **B**) and stratigraphically important dinocysts (**C–F**) from the Bo Member and the upper Farsund Formation in general. The figured specimens are stored at the Geological Survey of Denmark and Greenland, Copenhagen, under the catalogue numbers provided. **A**: A typical palynofacies assemblage from the organic-rich mudstones (facies 1, 2) of the Bo Member showing a dominance of AOM (**a**), dinocysts (**d**) and prasinophyte algae (**p**). Jeppe-1, core 1, 4402.71 m (drill depth). **B**: Palynofacies assemblage from a muddy sandstone bed (facies 3b; Fig. 13, sample 2). Note the heterogeneous nature of the assemblage, comprising wood particles (**w**), AOM (**a**) and palynomorph fragments, including prasinophyte algae (**p**). Jeppe-1, core 1, 4418.73 m (drill depth). **C**: *Rotosphaeropsis thula*. Bo-1, cuttings sample 8670–8680 ft, Geological Survey of Denmark and Greenland (GEUS) Catalogue No. 2000-KD-001. **D**: *Dingodinium? spinosum*. Bo-1, cuttings sample 8370–8380 ft, GEUS Catalogue No. 2000-KD-002. **E**: *Systematophora? daveyi*. Bo-1, cuttings sample 8790–8800 ft, GEUS Catalogue No. 2000-KD-003. **F**: *Amphorula expirata*. Bo-1, cuttings sample 8640–8650 ft, GEUS Catalogue No. 2000-KD-004. **G**: *Egmontodinium polyplacophorum*. Bo-1, cuttings sample 8580–8590 ft, GEUS Catalogue No. 2000-KD-005. **H**: *Senoniasphaera jurassica*. Bo-1, cuttings sample 8760–8770 ft, GEUS Catalogue No. 2000-KD-006.

W-U interval that is characteristic of the strata underlying the Bo Member over much of the Danish Central Graben is absent or very thin in this western area. The biostratigraphic data are poor but the results from Kim-1 suggest that the lower portion of the Bo Member in this well may be time-equivalent to at least part of the W-U interval in the remainder of the Danish Central Graben (Fig. 6).

Sequence stratigraphic framework

The sequence stratigraphy of the Jurassic in the Central Graben presented by Andsbjerg & Dybkjær (2003, this volume) is adopted here. Based primarily on new biostratigraphic data integrated with detailed log analysis, these workers subdivided the Upper Jurassic mudstone-dominated Farsund Formation into 11 sequences. In their study, sequence boundaries are recognised in well sections on the basis of parasequence stacking patterns (derived from log analysis), abrupt facies shifts and biostratigraphic evidence for hiatuses. In most wells, the uppermost sequence boundary (base Ryaz-1) lies beneath or at the base of the Bo Member (Fig. 6). In wells exhibiting an expanded section (e.g. Bo-1, Fig. 6) this surface lies within the upper levels of the W-U interval and there is no evidence of a significant stratigraphic gap. In such apparently conformable sections, the sequence boundary is of Late Volgian or Early Ryazanian age. In other wells, particularly on Late Jurassic structural highs (e.g. the Gert Ridge), this sequence boundary lies close to, or is coincident with, the base of the Bo Member and is marked by a significant stratigraphic gap (e.g. Gert-2, Fig. 6). Andsbjerg & Dybkjær (2003, this volume) did not present a sequence stratigraphic interpretation of the succession overlying this sequence boundary up to the base of the Cromer Knoll Group, acknowledging the probable complexity of the Bo Member itself and the unresolved debate concerning the sequence stratigraphic identity of the 'base Cretaceous unconformity' (Rawson & Riley 1982; Donovan *et al.* 1993).

Donovan *et al.* (1993) presented the results of an integrated sequence stratigraphic study of the Middle–Upper Jurassic in a portion of the UK Central Graben, some 200 km north-west of the Danish sector. They recognised eight depositional sequences, the uppermost of which includes Volgian–Ryazanian 'hot shales' that are equivalent in part to the Bo Member of the Danish Central Graben. The lower boundary of this sequence (the 'Purple' sequence boundary of Donovan

et al. 1993) occurs beneath the 'hot shales'; in some wells, this is a major truncation surface at which the Volgian and uppermost Kimmeridgian are absent. At its correlative conformity, this surface is suggested to be of medial Middle Volgian age (base Portlandian, *sensu anglico*). Donovan *et al.* (1993) suggested further that the maximum gamma peak within the 'hot shales' represents a maximum flooding surface; the succeeding sequence boundary was placed at the base of the Cromer Knoll Group although Donovan *et al.* (1993) acknowledged that evidence for truncation at this surface is scarce. There is a striking similarity between the log character of the Bo Member in the Danish sector and the equivalent succession in the UK sector of the Central Graben, particularly with respect to the relationship of these 'hot shales' to the regional sequence stratigraphic framework. However, the ages referred to the sequence boundaries underlying the 'hot shales' do not match and there is clearly a problem in integrating these two sequence stratigraphic frameworks.

Facies, processes and depositional environment

The uppermost Jurassic – lowermost Cretaceous 'hot shales' in the North Sea Central Graben occur within a thick Upper Jurassic – Lower Cretaceous mudstone-dominated marine succession (Farsund Formation and Cromer Knoll Group; Fig. 2), an unattractive stratigraphic position with respect to exploration for hydrocarbon reservoirs. As a result, this interval is rarely cored (Pegrum & Spencer 1990) and lithological, biostratigraphic and geochemical studies are generally based on ditch cuttings and sidewall cores. The Bo Member of the Danish Central Graben has, however, been partially cored in two wells, E-1 and Jeppe-1 (Figs 3, 6). The following discussion of facies, processes and depositional environment is based on these cores. It is well-established that the Farsund Formation was deposited in a fully marine environment (Michelsen *et al.* 1987); the Bo Member yields ammonites, inoceramid bivalves and dinocysts confirming its overall marine character (Birkelund *et al.* 1983 and this study).

Jeppe-1 well

Core 1 in Jeppe-1 spans the lower boundary of the Bo Member (Fig. 6), exhibiting 11.97 m of the underlying strata and 6.21 m of the Bo Member itself (Figs 9–11).

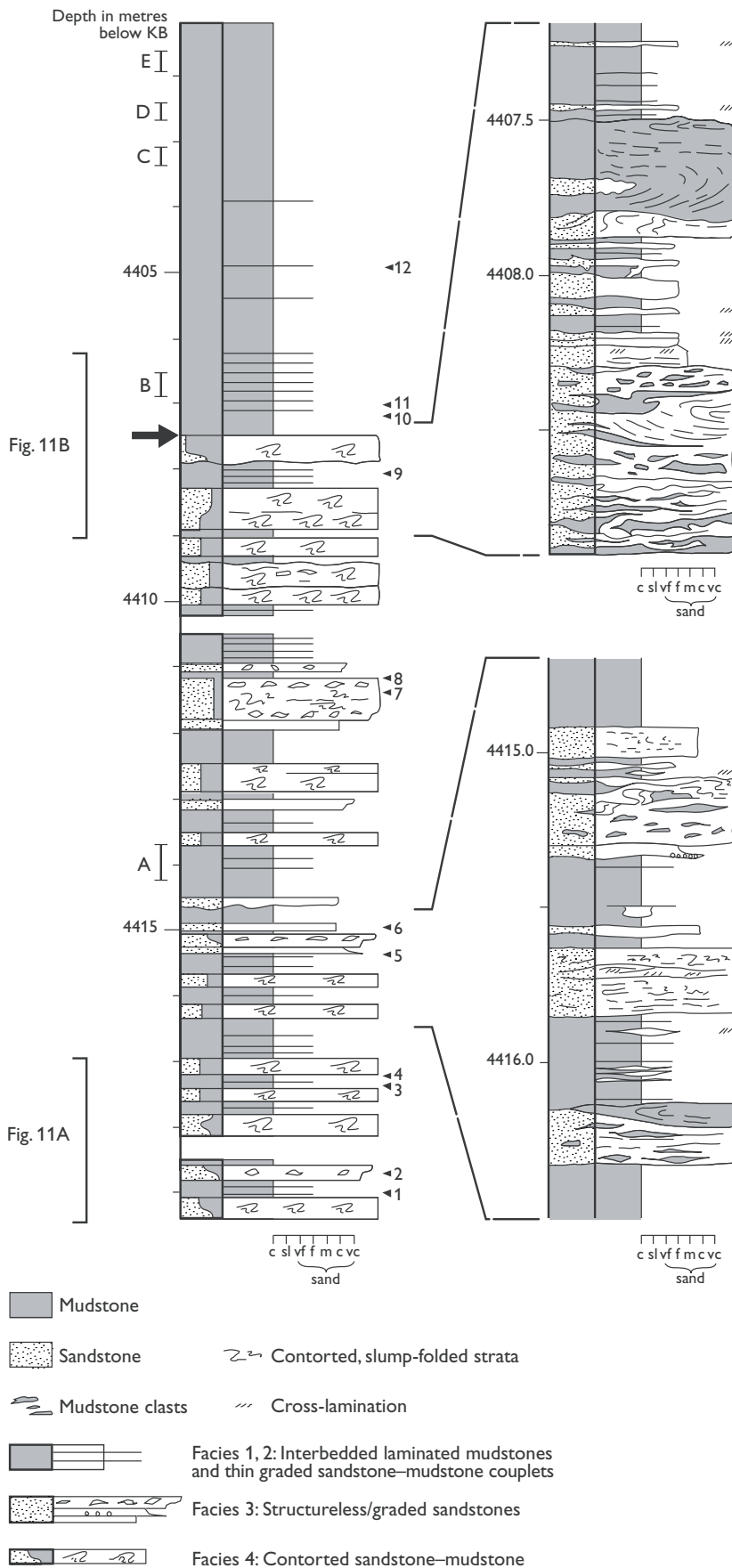


Fig. 9. Log of core 1 from the Jeppe-1 well, spanning the base of the Bo Member (large arrow). Note the clear subdivision into a heterogeneous, sand-rich lower portion (up to base Bo Member) and an upper mud-dominated portion showing a fining-upwards trend. Detailed logs illustrate the lithofacies described in the text; inferred sandstone injection structures occur at 4408 m and at 4415 m in these detailed sections. Small arrows 1–12 indicate locations of palynofacies analyses exhibited in Figure 13. The results of detailed lamina-by-lamina logs (A–E) through the fine-grained fraction are presented on Table 1 and Figure 10. The intervals illustrated in Figure 11 are indicated.

The core is well-preserved, in contrast to the E-1 core, and thus forms the dominant data source for detailed facies analysis, including palynofacies analysis.

Facies and depositional processes

Facies 1. Black claystones

This facies occurs intimately interbedded with thin-bedded sandstone–mudstone couplets (facies 2) where it ranges from laminae less than a millimetre thick to beds several centimetres thick. In representative detailed lamina-by-lamina sections (Fig. 10), the black claystone laminae are typically 1–3 mm thick, increasing both in thickness and overall proportion of the section upwards from the sub-Bo Member strata into the Bo Member itself (Fig. 12; Table 1). Laminae and beds are typically parallel-sided with sharp, flat boundaries, except where scoured and/or loaded at the contact with an overlying sandstone bed (facies 2).

The claystones are black or very dark grey, locally with a faint brownish cast, and show a weak, yet pervasive planar structure defined by discontinuous organic wisps, an overall platy fabric and, in places, by concentrations of calcareous microfossils (largely *Rhaxella perforata*, sponge reproductive cysts). Bioturbation is absent. A characteristic feature of this facies is the occurrence of phosphatic fragments, typically concentrated along specific horizons; on bedding plane surfaces these are often identifiable as fish scales. Larger vertebrate remains occur in this facies within the Bo Member at 4402.45 m (Figs 9, 10); these have been preliminarily identified as vertebrae, ribs and other bones of a marine reptile, most likely of plesiosaur or ichthyosaur type (S.E. Bendix-Almgreen, personal communication 1994).

Due to the finely interstratified nature of facies 1 and 2, analysis of palynofacies and source rock potential was largely undertaken on mudstone plug samples that included both facies. As shown on Figure 13 (see also Fig. 8A), the composition of the organic matter in these composite samples is very uniform. Amorphous organic matter (AOM) and dinocysts together form over 75% of the organic matter, dominating over the terrestrial component (wood fragments, spores and pollen). The total organic carbon (TOC) content of the mudstones (facies 1 and 2) in the Bo Member itself has a range of 5.2–7.1 wt% with an average of 6.0 wt% (Table 2). In the succession beneath the Bo Member, the mudstones show TOC values in the range 3.8–7.1 wt% (average of 5.0 wt%). Supplementary analyses were undertaken on two

samples of facies 1 and one sample of facies 2 from the Bo Member. These few analyses (Table 2) suggest that the facies 1 mudstones show slightly higher TOC values (7.3, 8.4 wt%) than the facies 2 mudstones (6.8 wt%), the slight difference being attributable largely to the higher silt content of the latter facies.

Interpretation. This facies is characterised by its fine-grained nature, the high organic carbon content, the dominance of AOM and dinocysts over terrestrial components, the lack of bioturbation and the weak yet pervasive planar fabric. All these features indicate deposition in a low-energy, oxygen-deficient marine environment, distant from terrestrial influence. The facies is interpreted to represent hemipelagic/pelagic fines, deposited by settling through the water column. The preservation of very thin (millimetric) sedimentation units in this succession, as a result of the suboxic–anoxic bottom conditions (terminology from Tyson & Pearson 1991) and consequent absence of bioturbation, permits the identification and differentiation of hemipelagic deposits from fine-grained sediment gravity flow deposits (see facies 2); in many deep-water fine-grained deposits, such differentiation is impossible (e.g. Ineson 1989). This facies is equivalent to the ‘fissile-laminated non-bioturbated mudrock’ facies of Stow & Atkin (1987), in their study of Upper Jurassic mudrocks from the UK sector of the North Sea, and to facies E2.2 of the deep-water facies scheme of Pickering *et al.* (1986).

Facies 2. Sandstone–mudstone couplets

This facies is an important component of the sub-Bo Member succession and forms over 70% of the cored portion of the Bo Member itself (Table 1). Where fully developed (Figs 9–11), it comprises two well-differentiated components: a lower parallel- to cross-laminated, fine- to very fine-grained sandstone succeeded by a structureless or coarse-tail graded, weakly laminated silty mudstone. Such couplets may be up to 5 cm thick, but are typically 0.5–2 cm thick; the sandstone portion is typically in the range 1–5 mm whereas the mudstone portion is typically 5–10 mm thick. This facies, together with facies 1, is a common component of the slump sheets assigned to facies 4. Syndepositional extensional microfaults and boudinage are observed locally. The organic composition of facies 2 is discussed under facies 1 (see above).

The basal sand–silt layer has a sharp base, often erosional and loaded. Some of these sandstones are normally graded but most are ungraded and show parallel-

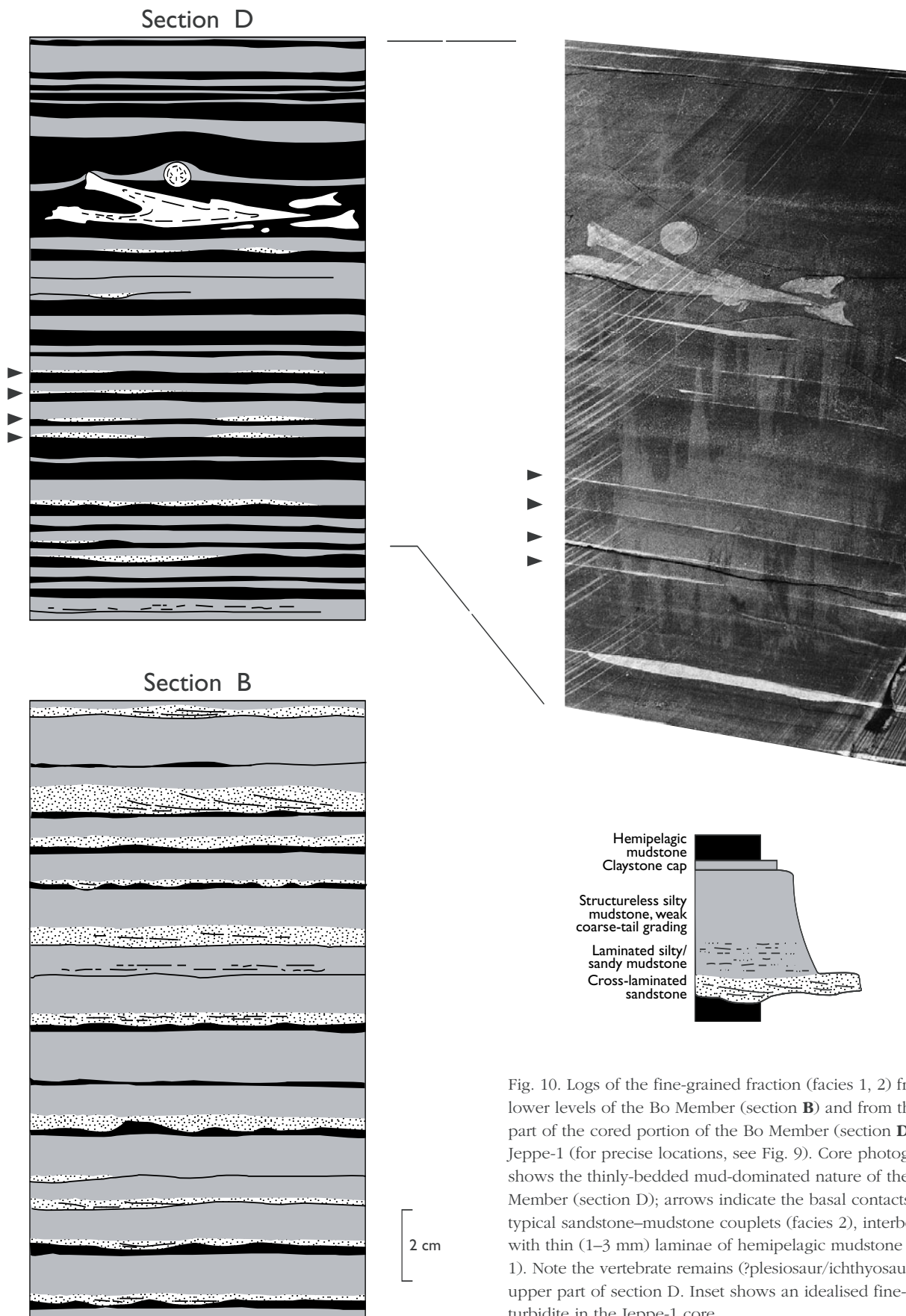


Fig. 10. Logs of the fine-grained fraction (facies 1, 2) from the lower levels of the Bo Member (section **B**) and from the upper part of the cored portion of the Bo Member (section **D**) in Jeppe-1 (for precise locations, see Fig. 9). Core photograph shows the thinly-bedded mud-dominated nature of the Bo Member (section D); arrows indicate the basal contacts of four typical sandstone–mudstone couplets (facies 2), interbedded with thin (1–3 mm) laminae of hemipelagic mudstone (facies 1). Note the vertebrate remains (?plesiosaur/ichthyosaur) in the upper part of section D. Inset shows an idealised fine-grained turbidite in the Jeppe-1 core.

Table 1. Jeppe-1 well, core 1: Sedimentological data from the Bo Member (sections B–E), and underlying strata (section A)

Section	Thickness (cm)	Facies 1 Hemipelagic mud		Facies 2 Sand–mud couplets (fine-grained turbidites)		
		Average thickness (mm)	Proportion of section (%)	Average thickness (mm)	Proportion of mud component in section (%)	Proportion of sand component in section (%)
E	27.3	2.4 (n = 33)	28.9	3.8 (n = 50)	60.4	10.7
D	26.5	2.7 (n = 31)	31.9	4.6 (n = 39)	64.6	3.5
C	25.4	3.1 (n = 16)	16.8	7.9 (n = 26)	76.5	6.7
B	35.9	1.5 (n = 18)	7.6	12.3 (n = 27)	73.4	19.0
A	53.0	1.1 (n = 12)	2.4	16.2 (n = 32)	81.9	15.7

or cross-lamination. Such cross-lamination is low-angle, sometimes with muddy toesets, and the sandstones are often lenticular, resembling the ‘fading ripples’ of Stow & Shanmugam (1980). Such well-developed sand–mud couplets (sand component > 2 mm thick) are subordinate in the facies (under 20% of facies 2 beds in the detailed sections) and many beds (about 30% in the detailed sections; Fig. 10; Table 1) possess only a discontinuous sand or silt lamina, less than 2 mm thick. In the Bo Member itself, nearly 70% of facies 2 beds lack the sand portion (i.e. ‘base cut-off’).

The mudstone portion of these couplets typically abruptly overlies the sand component, if present, although graded transitions were observed in a few beds. The silty mudstones are commonly structureless but may show weakly defined coarse-tail grading, particularly in the lower levels in association with diffuse lamination. On polished slabs, the grading is picked out by an upward decrease in the proportion of dispersed very fine sand and coarse silt grains, or *Rhaxella* cysts at some levels in the core.

An interesting feature of this facies is the presence of a very thin (0.1–0.2 mm) but persistent cap of mid-brown claystone (Fig. 10). In detailed lamina-by-lamina sections, over 75% of beds with non-erosional upper contacts possess this ultra-thin claystone cap. The base of this layer is sharp although rapid grading from dark grey silty mudstone to paler-coloured claystone is observed; the upper contact with succeeding hemipelagic deposits (facies 1) is sharp and planar.

Interpretation. Although the two components of these sandstone–mudstone couplets are often well-differen-

tiated, their close association and the occurrence, in places, of a graded transition from sand to mud indicates that they are the result of a common process. The scoured basal contacts and the presence of grading indicates deposition from a waning turbulent bottom current. In the absence of evidence of wave or storm activity, and in view of the close similarity to descrip-

Table 2. Jeppe-1 well: Geochemical data from core 1 spanning the base of the Bo Member (dashed line)

Depth (m)	Facies	TOC (wt%)	T_{max} (°C)	S1 (mg/g)	S2 (mg/g)	HI	Palyno-facies sample no.
4401.36	1, 2	5.5	445	4.5	28.0	508	
4401.37	1	7.3	446	3.6	33.0	454	
4402.67	1	8.4	445	4.5	39.5	473	
4402.69	2	6.8	444	4.0	31.2	458	
4402.71	1, 2	7.1	444	5.2	37.7	533	
4404.18	1, 2	5.2	443	4.1	24.7	477	
4404.97	1, 2	6.6	441	5.5	35.7	543	12
4406.98	1, 2	5.9	443	3.6	28.7	488	11
4407.19	1, 2	5.6	444	3.0	27.9	497	10
4408.09	1, 2	4.7	441	2.5	20.8	443	9
4409.35	1, 2	3.8	441	2.6	14.5	378	
4410.86	1, 2	5.3	443	2.6	22.4	425	
4411.12	3a	0.5	434	2.0	0.8	169	8
4411.36	3a	n.d.	n.d.	n.d.	n.d.	n.d.	7
4412.94	1, 2	7.1	433	5.4	36.9	518	
4414.72	1, 2	5.3	445	3.1	22.6	431	
4414.97	3b	0.2	n.d.	n.d.	n.d.	n.d.	6
4415.49	1, 2	4.7	445	2.6	19.3	412	5
4417.33	1, 2	4.5	442	2.9	18.3	405	4
4417.47	4	2.3	443	2.3	6.3	276	3
4418.73	3b	1.2	442	1.7	3.5	281	2
4419.36	4	4.7	445	2.7	22.0	466	1

n.d., not detected

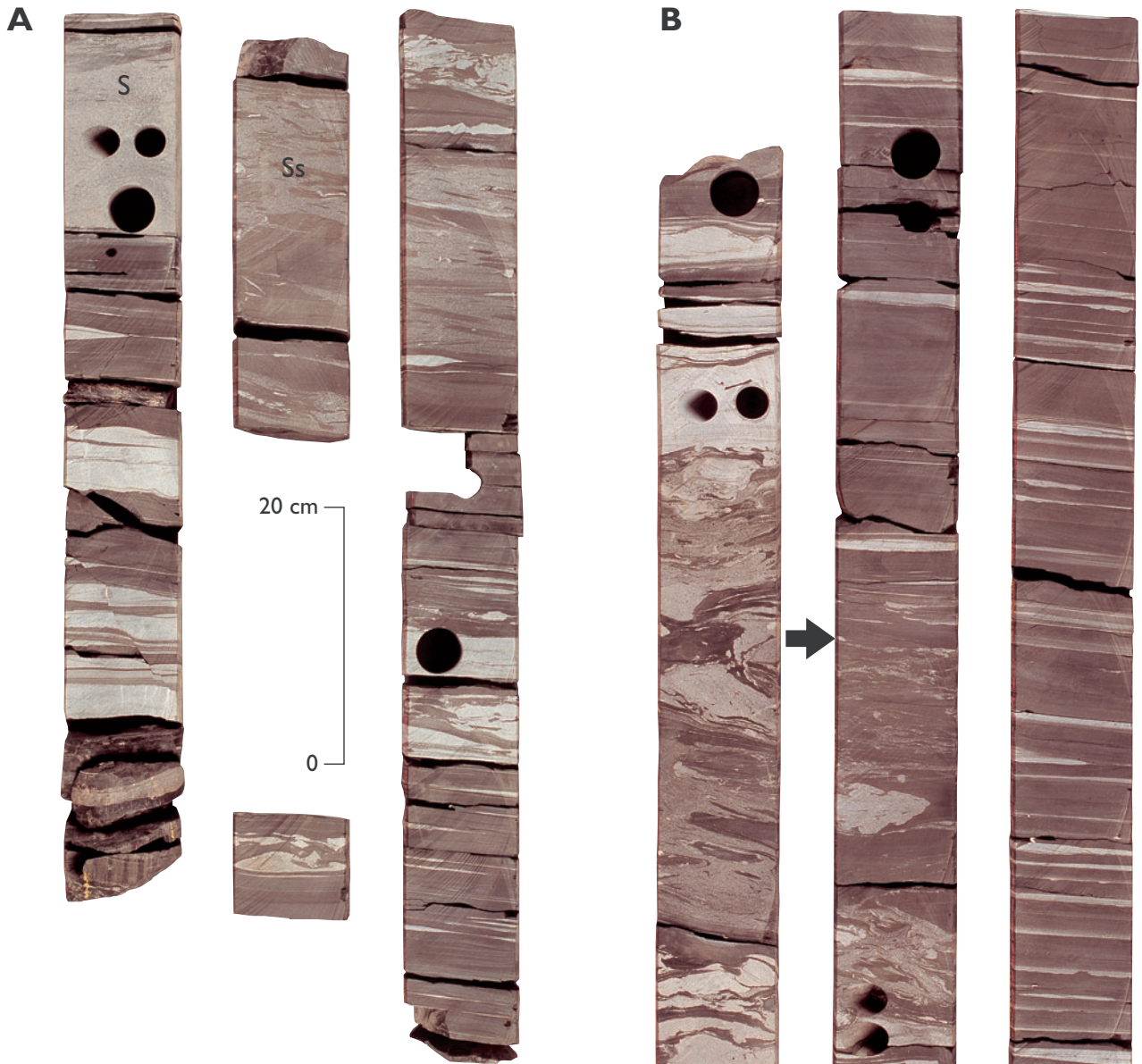


Fig. 11. Selected intervals of core 1, Jeppe-1 (Fig. 9); the section youngs from bottom left to top right. Depths (drill) in metres below well reference point. **A:** The lowermost beds (4419.4–4416.9 m) typify the cored interval beneath the Bo Member. Massive sandstones (**S**; facies 3) and slump sheets (**Ss**; facies 4) are interbedded with fine-grained turbidites and hemipelagic mudstones (facies 1, 2). **B:** This core section (4408.9–4406.1 m) spans the lower boundary of the Bo Member (arrow; see also Fig. 9) and illustrates the thin-bedded, mud-dominated nature of this member.

tions of fine-grained sediment gravity flow deposits in the literature (Piper 1972; Stow 1979; Stow & Shanmugan 1980; Pickering *et al.* 1986), these beds are attributed to deposition from sediment-starved, dilute turbidity

currents. The basal laminated and ripple cross-laminated sandstone portion, where present, is indicative of tractional processes whereas the succeeding structureless or graded mudstone portion was deposited pri-

marily from suspension. The silty lamination observed locally within the mudstones may be the result of shear sorting within the bottom boundary layer (Stow & Bowen 1978).

These couplets are comparable to Facies C2.3/D2.1 of Pickering *et al.* (1986) and closely resemble the 'mudstone facies' described from the Upper Jurassic Brae oil-field of the UK sector of the North Sea (Stow *et al.* 1982). These beds can be described as $T_{c(d)e}$ turbidites on the classical Bouma (1962) scheme; within the mudstone portion, the divisions E1, E2, E3 of Piper (1972) and T3/T4, T6, T7 of Stow & Shanmugan (1980) are represented. Interestingly, the discrete, ultra-thin (0.1–0.2 mm) claystone cap observed in this study is not evident in these facies schemes for fine-grained turbidites. This cap is clearly differentiated from the graded or structureless mudstone division and from overlying hemipelagic deposits; it probably represents the fall-out of the finest sediment fraction following the passage of the muddy turbulent cloud. Preservation of this thin lamina requires the total absence of burrowing infauna, perhaps explaining its lack of recognition in other fine-grained successions. As noted under facies 1, the absence of bioturbation in association with the high TOC content and the nature of the palynofacies indicates deposition in an oxygen-deficient marine environment.

Facies 3. Sandstones

This facies is absent from the cored portion of the Bo Member but forms about 17% of the underlying Farsund Formation (Fig. 9). The sandstones are typically medium- to fine-grained, although locally coarse- or very coarse-grained, and commonly contain up to 20% abraded shell fragments (bivalves, echinoderms, sponge spicules, bryozoans). The sandstones are mineralogically sub-mature, containing up to 30% feldspar grains, mica flakes and rock fragments. The lithics include basement lithologies, such as meta-quartzite, acid plutonics and foliated mica-quartz aggregates, in addition to volcanic and dolomite fragments probably derived from the Permian section. Coalified wood fragments, locally up to 1 cm across, are present in some beds. Glaucony is a characteristic although minor component. The TOC content of this facies varies considerably, dependent on the proportion of mud either as matrix or as discrete clasts (Table 2). Two subfacies are recognised, based on the presence or absence of mud matrix.

Subfacies 3a. These mud-free, calcite-cemented sandstones are rare, being represented by just three beds

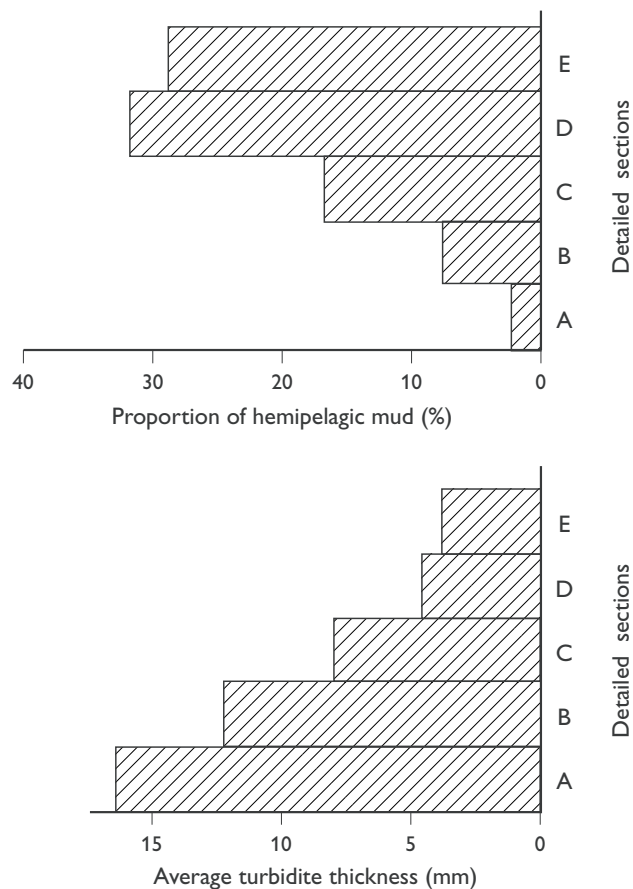


Fig. 12. Histograms showing the stratigraphic variation in the average thickness of the fine-grained turbidites (facies 2) and in the relative importance of the hemipelagic deposits (facies 1), based on the detailed sections A–E (Fig. 9).

ranging in thickness from 5 cm to 20 cm. They are of medium or medium-coarse sand grade and are structureless or show diffuse parallel to low-angle (deformed?) lamination. Bed boundaries are typically sharp and planar but one bed has an erosional, scoured base and a normally graded, pebbly basal few centimetres (Fig. 9). This facies occurs in association with muddy sandstones (facies 3b) and slumped sediment (facies 4).

Interpretation. The sharp, locally erosive basal contacts, the presence of normal grading, the massive to parallel-stratified structure and the lack of mud matrix suggests deposition from energetic, waning currents that were capable of sorting sand from mud-grade sediment. These beds are interpreted as the deposits of sandy turbidity currents and can be broadly classified as Bouma T_{ab} turbidites. The implications of their close association with facies 3b and 4 is discussed below. The

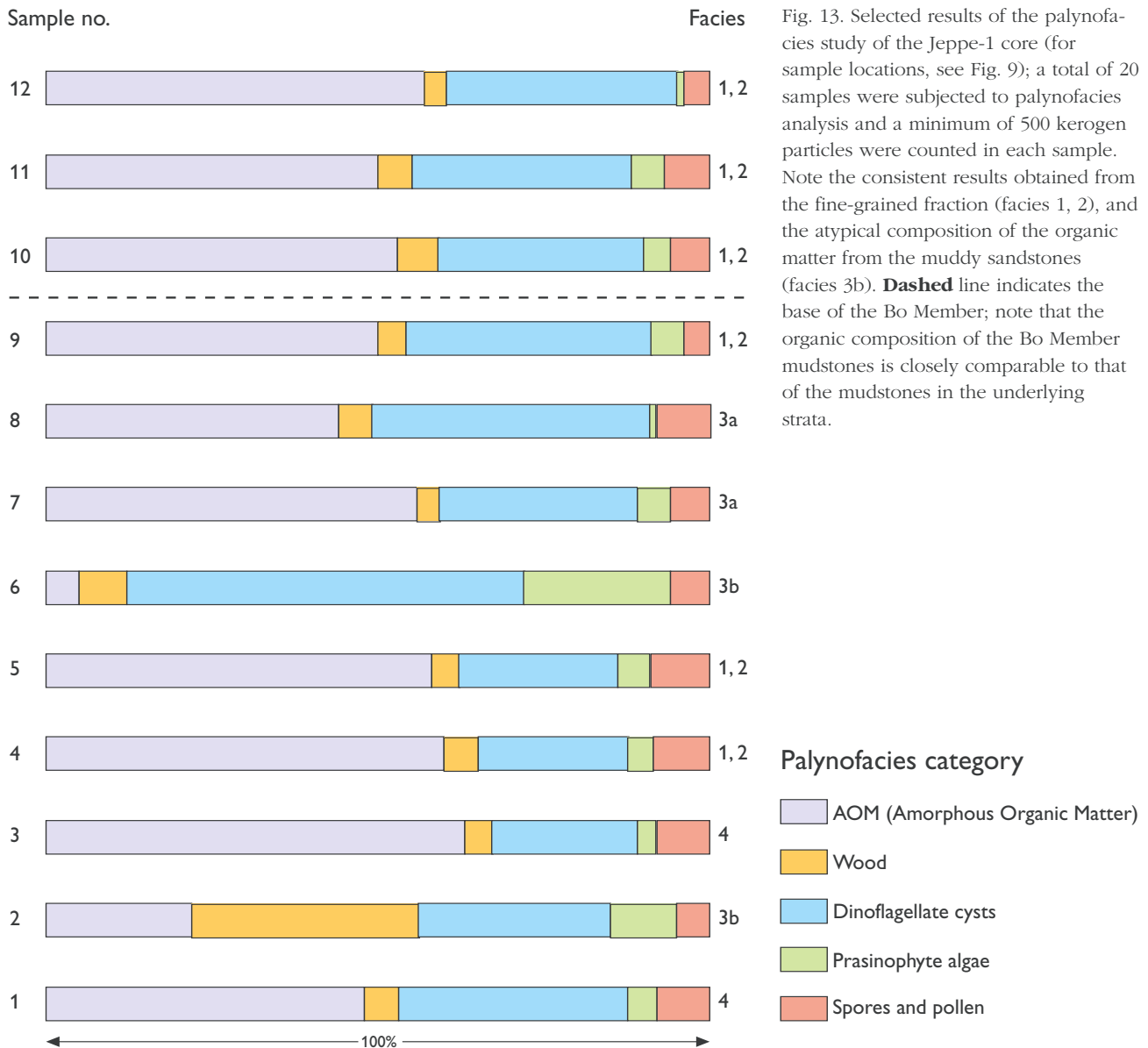


Fig. 13. Selected results of the palynofacies study of the Jeppe-1 core (for sample locations, see Fig. 9); a total of 20 samples were subjected to palynofacies analysis and a minimum of 500 kerogen particles were counted in each sample. Note the consistent results obtained from the fine-grained fraction (facies 1, 2), and the atypical composition of the organic matter from the muddy sandstones (facies 3b). **Dashed** line indicates the base of the Bo Member; note that the organic composition of the Bo Member mudstones is closely comparable to that of the mudstones in the underlying strata.

organic content of this subfacies shows a typically marine signature, AOM and dinocysts together dominating the assemblage (Fig. 13).

Subfacies 3b. Mud-rich, medium- to coarse-grained sandstones dominate facies 3 and occur in close association with slumped sediments (facies 4). Indeed, there is a complete gradation between muddy sandstones containing discrete intraformational sandstone or mudstone clasts and intervals of contorted sandstone and mudstone (cf. *in situ* facies 1/2) in which semi-coherent, slump-folded sediment rafts are separated by zones of muddy sandstone (cf. facies 3b). Sandstone beds referred to this subfacies are 5–25 cm in thickness (typ-

ically about 10 cm thick) and generally show sharp, flat bed boundaries; loading and water escape structures are observed in places. They are typically structureless but a few beds show faint parallel stratification. Elongate intrabasinal clasts ranging in length from a few centimetres to the width of the core (10 cm) are oriented parallel to bedding and in several cases are concentrated in the upper levels of the bed.

Two muddy sandstone beds assigned to this facies were subjected to palynofacies analysis (Fig. 13, samples 2, 6); the results contrast with the fine-grained fraction (see discussion under facies 1) but also differ from each other. Both samples show a decrease in the relative importance of AOM compared with the mudstone

facies, but the lower sample (Fig. 8B) shows a relative increase in the wood component whereas the upper sample shows an increase in the proportion of dinocysts and prasinophyte algae.

Interpretation. The poor sorting, mud matrix, structureless ungraded character, flat non-erosional boundaries and gradation to slumped strata indicate deposition from viscous sediment gravity flows of debris flow type. This facies is equivalent to Facies C1.1 of Pickering *et al.* (1986).

A few thin beds (max. 10 cm thick) assigned broadly to this facies show some features that are not wholly compatible with the interpretation given above (Fig. 9). They show sharp, subparallel boundaries that are locally slightly oblique to the general bedding and display elongate, tapering offshoots up to a few centimetres long and a centimetre across, both on the upper and lower bed boundaries. Although largely structureless, weak lamination may be present in the middle zone of the bed. These sandstones are interpreted to be of intrusive origin (i.e. sandstone sills) rather than representing primary sediment gravity flow deposits, although confirmation is impossible in core.

Facies 4. Contorted sandstone–mudstone

Units of contorted intraformational sediment form an important and striking part of the sub-Bo Member succession in the Jeppe-1 core (Figs 9, 11). They are 10–50 cm thick and typically show flat, non-erosional bases and flat or slightly irregular tops. A complete gradation is represented from sheets composed entirely of slump-folded but essentially coherent thinly interbedded sandstones and mudstones (cf. facies 1, 2) to sheets composed of lenses (phacoids *sensu* Voigt 1962) of internally deformed bedded sandstone and mudstone floating in a muddy sandstone matrix. With increasing disintegration of intraformational slabs, this facies grades into the muddy sandstones assigned to facies 3b. The organic composition and TOC content of this facies are very variable, most likely due to the variable proportion of mud in these heterogeneous deposits.

Interpretation. This facies records remobilisation and partial disaggregation of thin-bedded sandstone and mudstone and is thought to have originated largely by surficial downslope transport by processes ranging between sliding or slumping and viscous debris flow. It is thus equivalent to Facies F2.1 transitional to C1.1 of Pickering *et al.* 1986. It is possible, however, that sub-

surface injection processes may have created some of the fabrics illustrated by this facies (cf. Anderton 1997).

Stratigraphic distribution of facies

As shown on Figure 9, the cored section in the Jeppe-1 well is readily subdivided into a lower heterogeneous, relatively sand-rich interval capped by a fining-upwards, mud-dominated unit, representing the basal beds of the Bo Member. The base of the sand-rich interval beneath the Bo Member occurs at a log depth of 4431 m, at an inferred sequence boundary (Fig. 6; Andsbjerg & Dybkjær 2003, this volume).

In the cored sub-Bo Member interval, packets of thin-bedded sandstone–mudstone turbidites (facies 2) with intervening hemipelagic laminae and beds (facies 1) make up around half (47%) of the succession, interbedded with sandstone turbidites and debris flows (facies 3; 17%) and sandstone–mudstone slump sheets (36%). In a representative detailed section (section A) through the fine-grained fraction (Figs 9, 12; Table 1), hemipelagic mud (facies 1) forms less than 5% of the fine-grained fraction at this level, typically occurring as thin laminae (about 1 mm thick) sandwiched between sand–mud turbidites (facies 2; average thickness 16 mm).

The cored section of the Bo Member is composed solely of facies 1 and 2. The abundance and thickness of the sand component in facies 2 decreases upwards from the sub-Bo Member succession into the Bo Member itself, in parallel with an upward decrease in turbidite thickness and a relative increase in the proportion of hemipelagic mud (Figs 10, 12; Table 1). In the uppermost detailed sections (Fig. 9, sections D, E), hemipelagic mudstone forms about 30% of the succession in laminae up to 10 mm thick.

Depositional setting

The cored section from the Jeppe-1 well records deposition in a low-energy marine environment characterised by background sedimentation of muds and subordinate thin sands from dilute, muddy turbidity currents and by suspension settling through the water column. The absence of bioturbation and the preservation of high levels of organic carbon indicate very low levels of free oxygen (suboxic–anoxic) within the bottom waters.

In contrast, coarse poorly-sorted shelly sandstone turbidites, debris flow deposits and slump sheets record a more dynamic depositional environment. Firstly, the

shelly sands testify to a source of immature sediment, including basement rock fragments; the presence of a varied assemblage of coarse shell fragments and glaucony suggest relatively high-energy, shallow marine conditions in the source area. Secondly, the slump sheets of turbiditic sandstone and mudstone and associated debris flow deposits provide evidence of intrabasinal slopes in the vicinity of the Jeppe-1 well. Indeed, the sandstone injection structures, the evidence of minor slope creep and the close similarity between the *in situ* sediments (facies 1, 2) and the slump sheet components indicate that this succession accumulated very close to the base-of-slope.

The Jeppe-1 well is located near the western margin of the Gertrud Graben, a half-graben that was most active in Volgian times when it was bounded by the Gert Ridge to the west and the Mandal High to the east (Fig. 3). As noted earlier, block rotation in association with subsidence of the Feda and Gertrud Grabens resulted in local compression and uplift of the Gert Ridge during the latest Jurassic (Rasmussen 1995), yielding a potential sediment source just west of the Jeppe-1 location. In the Gert Field at the northern end of the Gert Ridge, Permian, Carboniferous and metamorphic basement were encountered beneath the Jurassic succession (Rasmussen 1995). In the immediate area of the field, such potential sediment sources were draped by Upper Jurassic sediment during deposition of the Bo Member, but it is possible that these strata were locally exposed in uplifted fault slices at the southern end of the Gert Ridge.

It is likely, however, that the Gert Ridge was most important as a source of intraformational sediment in the latest Jurassic and earliest Cretaceous. Both the sedimentological and palynological data from Jeppe-1 testify to significant reworking of Middle to Upper Volgian sediments in the form of slumps, slides and debris flows. The well data from the Gert Field atop the Gert Ridge also suggest that the ridge experienced significant erosion in the latest Jurassic. The uppermost Farsund Formation, including the Bo Member, is absent from the Gert-1 and Gert-3 wells and a significant stratigraphic gap spanning the late Middle and Late Volgian is recognised immediately beneath the Bo Member in Gert-2 (Fig. 6 and previous discussion).

The data suggest, therefore, that the cored succession in Jeppe-1 was derived from two sources. The thin muddy turbidites that form the bulk of the succession were probably derived from regional sediment sources; transport paths are difficult to infer, especially given the complex nature of the Danish Central Graben in the Late Jurassic, but were most likely axial in the various elon-

gate subbasins. In this context, it should be noted that Rasmussen *et al.* (1999) have proposed, on the basis of seismic data, the existence of a channelised sandy fan system in the axis of the Gertrud Graben that was broadly coeval with the cored section beneath the Bo Member in Jeppe-1. These workers suggested that the thin-bedded turbidites observed in the Jeppe-1 core represent 'fan-fringe' or levee/overbank deposits related to the axial channel system. Rasmussen *et al.* (1999) postulated that the siliciclastic source for this fan system was the Sørvestlandet High (Fig. 1), although the Mandal High and the inverted Søgne Basin area are also possible candidates (Fig. 3). It is notable, however, that the mudstone component of the thin turbidites is palynologically and geochemically closely comparable to the hemipelagic mudstones. This suggests that the clay fraction was mainly of intrabasinal origin, cannibalised by erosive, turbulent flows entering the Central Graben.

In contrast, the coarse shelly sands and the slumps, slides and debris flow deposits of intraformational sediment may have been of local origin, perhaps shed from the Gert Ridge immediately to the east of the Gertrud Graben. Dispersal of this intrabasinal sediment from the flanks of the Gert Ridge may have occurred during storms; the close association of shelly sands with slump sheets or debris flow deposits suggests a common triggering mechanism. Although acknowledging the possibility of local derivation, Rasmussen *et al.* (1999) suggested that the shelly sands and slump–debris flow deposits may alternatively be related to the axial fan system, perhaps recording periodic levee collapse or breach.

A number of the facies displayed by the Jeppe-1 well are closely comparable to those described from the Brae and Miller oilfields in the Viking Graben of the North Sea (Stow *et al.* 1982; Turner *et al.* 1987; McClure & Brown 1992). In particular, the sandstone–mudstone couplets (facies 2) are closely comparable to the 'tiger stripe' facies described by Stow *et al.* (1982). Such facies are commonly termed 'interchannel' or 'levee' deposits when observed in close association with coarse-grained channelised turbidites (Mutti 1977; Walker 1985). The inferred existence of a channelised fan system in the axis of the Gertrud Graben (Rasmussen *et al.* 1999) is interesting in this respect. The large-scale fining-upwards trends observed in the core and in the overlying uncored portion of the Bo Member (Fig. 6), record pulses of erosion and sediment dispersal both from regional and intrabasinal sediment sources. As discussed above, the mud-rich turbidites in the cored section become thinner, and in general finer-grained, upwards whereas the

interbedded hemipelagic muds form an increasing proportion of the succession (Fig. 12; Table 1). Although the biostratigraphic resolution does not permit direct measurement of sedimentation rates at this scale, this pattern is suggestive of a waning supply of turbiditic mud relative to the background hemipelagic rain, perhaps related to rising sea level.

E-1 well

The cored section in the E-1 well is from the uppermost portion of the Bo Member, where the gamma-ray values are consistently high (Fig. 6). Although originally totalling some 2 m of core (9783–9792 ft, 77% recovery), this core dates from 1968 and has been intensively sampled. Representative slabs remain, totalling about 1 m of core. This core and the boundary between the Farsund Formation and the overlying Cromer Knoll Group in this well were subjected to a detailed biostratigraphic study by Birkelund *et al.* (1983). The core fragments that remain are composed solely of parallel-laminated black or very dark grey claystone, comparable to facies 1 of the Jeppe-1 core.

Facies and depositional processes

Facies 1. Black claystones

The lamination in this dark organic-rich claystone is defined by slight colour variation, concentrations of silt-sized calcite grains (*?Rbaxella* sp.) and phosphatic fish fragments. Birkelund *et al.* (1983; fig. 3) illustrated the well-laminated nature of this facies by means of an X-radiograph, and suggested that two orders of lamination are present. The sub-millimetric parallel lamination is interrupted at regular intervals (5–20 mm) by discrete paler homogeneous claystone laminae (0.1–0.3 mm thick); these laminae define the second order lamination of Birkelund *et al.* (1983). Ammonites and inoceramid bivalves are present in this core; the latter occur at discrete horizons and in places appear to extend across the full width of the core. Birkelund *et al.* (1983) figured complete inoceramid valves (assigned to *Inoceramus* aff. *vereshagini*) and cross-sections indicate the presence of compacted yet entire valves; no evidence of transport is observed.

The organic composition of the Bo Member in the E-1 core is very uniform (Table 3); TOC values range from 6.7 to 8.5 wt% (average of 7.3 wt%), comparable

Table 3. E-1 well: Geochemical data from the Bo Member in core 8

Depth (ft)	Facies	TOC (wt%)	T_{\max} (°C)	S1 (mg/g)	S2 (mg/g)	HI
9784	1	7.0	431	2.7	40.6	582
9786	1	6.8	424	4.0	40.6	601
9787	1	7.0	426	3.5	42.9	614
9788	1	7.8	426	4.4	45.5	587
9789	1	6.7	433	3.6	40.2	599
9790	1	8.5	430	4.8	56.3	660

to the individual analyses of the hemipelagic facies (facies 1) in Jeppe-1. The content of organic matter is also very homogeneous, being strongly dominated by amorphous organic matter (AOM), dinocysts and prasinophyte algae.

Interpretation. In common with facies 1 of the Jeppe-1 core, this well-laminated organic-rich mudstone facies is attributed to hemipelagic settling of fines through the water column. The discrete light-coloured claystone laminae are reminiscent of those that cap the mudstone turbidites (facies 2) in the Jeppe-1 core and may have a similar origin i.e. they may represent the fine-grained tail of individual turbidite events. Sandstone–mudstone couplets of the type seen in Jeppe-1 (facies 2) are, however, not recognised in E-1. The well-preserved parallel lamination with no sign of bioturbation (even in radiographs), and the high content of organic matter (TOC), particularly AOM, suggest that anoxic conditions prevailed for much of the time in the bottom waters. The occurrence of in-situ inoceramid bivalves at certain horizons, however, indicates at least periodic suboxic conditions on the sea floor (Birkelund *et al.* 1983). Bioturbation is not evident in association with these faunas suggesting that conditions on the sea floor were close to the boundary between anoxic and dysoxic (i.e. suboxic in the terminology of Tyson & Pearson 1991). Comparable apparently anomalous sediment–faunal associations have been described and similarly interpreted by Savrda & Bottjer (1987) from the Miocene of California and by Doyle & Whitham (1991) from the Upper Jurassic – Lower Cretaceous of the Antarctic Peninsula. Savrda & Bottjer (1987) suggested that the development of such associations (their exaerobic biofacies) is favoured by the development of bacterial mats on the sediment surface (see also Tyson & Pearson 1991).

The association of benthic inoceramid faunas, in particular, with laminated ‘black shales’ has been noted by many workers (see discussion by MacLeod & Hoppe 1992). On the basis of facies criteria (Kauffman &

Sageman 1990) and isotope data (MacLeod & Hoppe 1992), it has been suggested that inoceramids may have benefited from bacterial chemosymbiosis, thus extending their potential environmental range. Although the isotopic evidence has been disputed (Grossman 1993), there is well-documented evidence that inoceramid bivalves were tolerant of conditions that excluded most other forms (MacLeod & Hoppe 1992), whether this was the result of a highly efficient metabolism, chemosymbiosis or a combination of these factors. In any event, inoceramids are not observed in the Jeppe-1 core and only occur at specific levels in the E-1 core, indicating that even the most tolerant benthic invertebrates were largely excluded from the floor of the Danish Central Graben during the deposition of the Bo Member.

Depositional setting

The E-1 core consists solely of laminated organic-rich hemipelagic mudstones that accumulated under suboxic–anoxic bottom conditions. In contrast to the Jeppe-1 core, turbidite processes were apparently unimportant in this setting. As shown on Figure 6, however, the E-1 and Jeppe-1 cored sections are not time-equivalent so that it is not clear whether this contrast in sedimentation style reflects a geographical or a temporal shift in the dominant depositional processes.

Organic geochemistry and source rock potential

A number of studies have documented the good to very good source potential of the Farsund Formation in the Danish sector (Damtoft *et al.* 1987, 1992) and a positive correlation between source characteristics and the produced hydrocarbons in Danish fields has been achieved in a number of cases (Østfeldt 1987; unpublished GEUS data). In this section, the source characteristics and geochemistry of the Farsund Formation are described with particular emphasis on the Bo Member.

Farsund Formation

The total organic content (TOC) of the Farsund Formation is very variable, being dependent on lithology, stratigraphic position, geographical setting and

level of thermal maturity. It ranges from less than 1 wt%, typically in sandstone or dolomite/limestone stringers and in mudstones in the lower levels of the formation, to more than 15 wt% in the mudstones of the Bo Member. Pyrolysis yields (Rock-Eval S2) vary from less than 1 kg HC/ton rock to more than 90 kg HC/ton rock. Corresponding values of the hydrogen index range from less than 100 to approximately 600. However, in general, the Farsund Formation can be considered a good or even very good hydrocarbon source rock.

The source rock potential varies with depth through the formation, as well as geographically within the Central Graben. In general terms, the lower Farsund Formation is poorer in organic carbon and the proportion of terrigenous organic matter is significant, leading to a mixed gas-/oil-prone kerogen type (Damtoft *et al.* 1987, 1992). Upwards, the terrigenous component of the organic matter decreases in abundance, leading to highly oil-prone kerogen of predominantly marine/bacterial origin. In the south-eastern part of the Central Graben, and along the eastern border fault (Coffee Soil Fault), the source rock potential deteriorates, probably due to dilution with mineral matter (siliclastic detritus) and incorporation of larger proportions of inert terrigenous organic matter.

In the upper Farsund Formation, *n*-alkane distributions are unimodal, with low to moderate proportions of 'Unresolved Complex Mixture' (UCM), centred in the range C_{15–19}. With a few exceptions, the *n*-alkane distributions are smooth, with little or no preference for odd or even numbered compounds. The abundance of linear isoprenoids is generally low to moderate. Pristane/phytane ratios are mainly in the range 0.9–1.6. Occasionally, an increased contribution of terrigenous organic matter to the kerogen is manifest in slightly increased abundance of waxy components (*n*C₂₂₊), slight predominance of odd-numbered *n*-alkanes in the range *n*C_{23–31}, and a bimodal distribution of UCM. In the lower Farsund Formation these features are pronounced, testifying to a general increase in the proportion of the kerogen component derived from terrigenous organic matter.

The terpane distributions show very variable, but mostly modest amounts of tricyclic triterpanes, which may form a homologous series ranging from C₂₀ to C₃₀ (see below). The pentacyclic terpanes of the hopane series are dominated by hopane and norhopane. In low maturity samples, moretanes and ββ-hopanes, in particular 17β(H)-trisnorhopane, may be rather abundant, whereas 28,30-bisnorhopane, where present, forms only a minor proportion. The proportions of Ts and Tm,

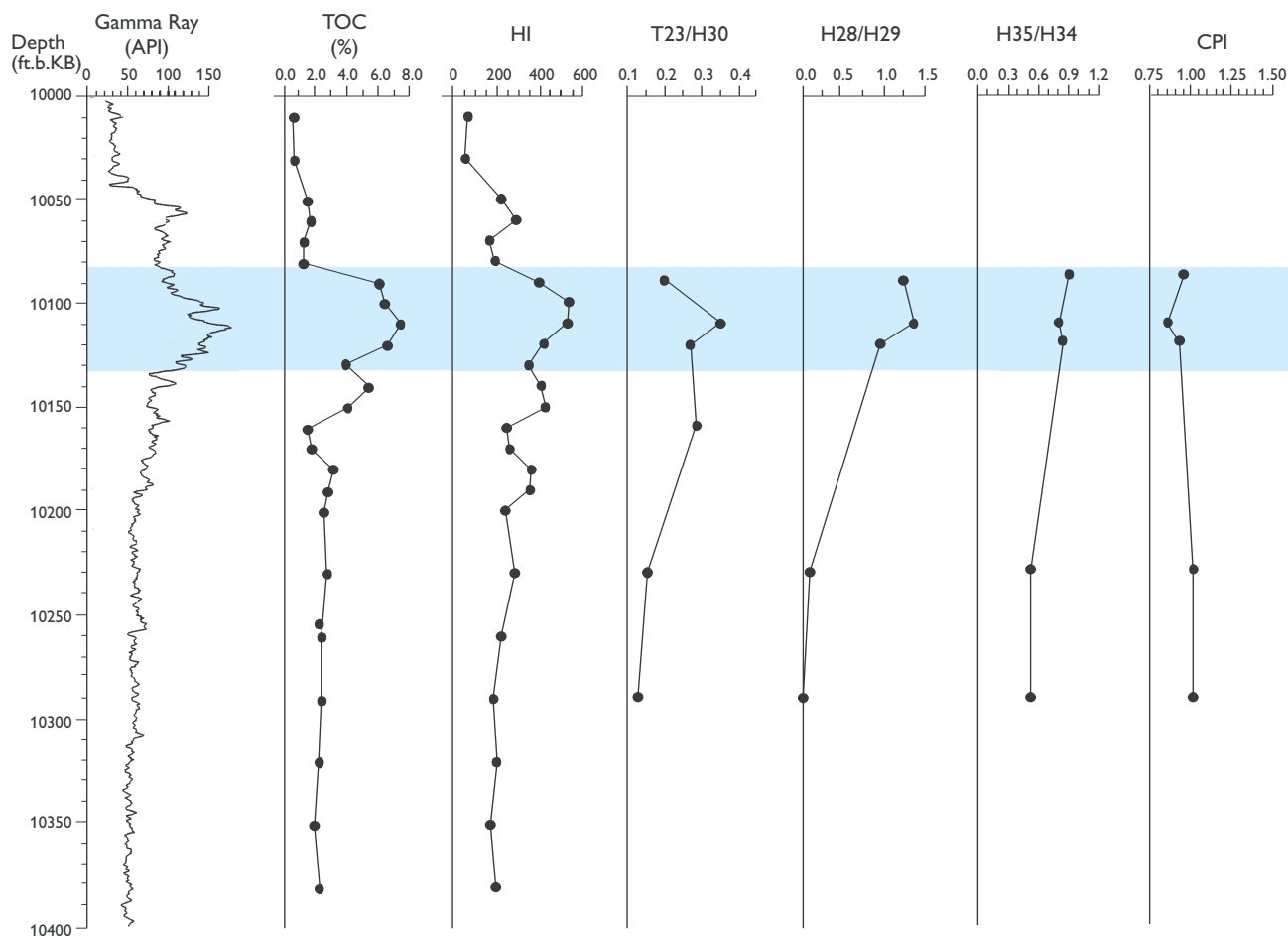


Fig. 14. Geochemical profile of the upper Farsund Formation, including the Bo Member (blue), in the Edna-1 well. **TOC**, total organic carbon; **HI**, hydrogen index; **T23/H30**, ratio of C₂₃-tricyclic terpene to C₃₀ hopane; **H28/H29**, ratio of 28,30-bisnorhopane to norhopane; **H35/H34**, ratio of pentakishomohopane to tetrakishomohopane; **CPI**, carbon preference index.

and 29Ts vary with the level of thermal maturity. C₃₀ diahopane (compound 'X' of Philp & Gilbert 1986; Moldowan *et al.* 1991), is generally present in very low proportions, but tends to become increasingly prominent with maturation. Extended hopanes are abundant, displaying a regular decrease in abundance with increasing carbon number from C₃₁ to C₃₅.

The distribution of regular steranes is very homogeneous, featuring a slight predominance of C₂₇ steranes over C₂₈ and C₂₉ steranes, which are roughly equal in abundance. Diasteranes are rather abundant, generally increasing with level of thermal maturity. C₃₀ steranes are present in all samples.

Bo Member

At 'bulk level', the mudstones of the Bo Member are characterised by high or even very high organic carbon contents, generally in the range 4–8 wt% TOC, occa-

sionally exceeding 15 wt% TOC. Pyrolysis yields are very high, 10–100 kg HC/ton rock (Rock-Eval S2), with corresponding values of the Hydrogen Index occasionally exceeding 500, somewhat dependent on the level of thermal maturity (Figs 14, 15; Tables 2, 3). Interestingly, organic matter enrichment may extend above the upper boundary of the Bo Member as defined by the gamma log (Ravn-2, G-1). As discussed earlier, the high gamma radiation exhibited by the Bo Member mudstones is attributed primarily to uranium bound to organic matter. The occurrence of organic-rich shales with relatively low gamma values above the Bo Member may, therefore, be either the result of exhaustion of available uranium in the geochemical system or a decrease in the ability of the sediments to incorporate uranium. Factors that may conceivably influence the incorporation of uranium into sediments are organic matter type, redox conditions and sedimentation rate. The data show no marked changes in organic matter type and, in any event, complexation and subsequent reduction of U⁶⁺

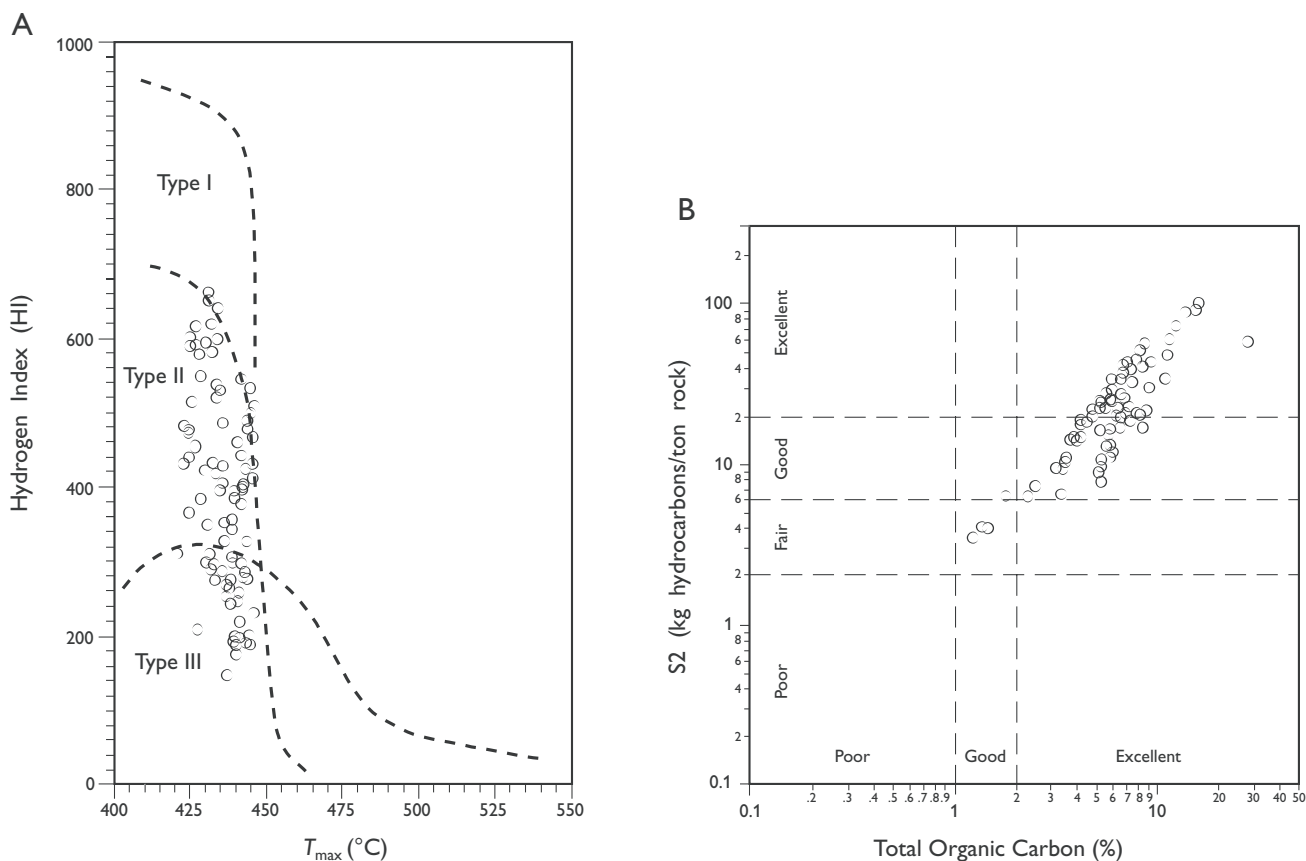


Fig. 15. Rock-Eval/TOC data for samples of the Bo Member in various sub-basins of the Danish Central Graben. **A:** T_{max} vs. Hydrogen Index (HI); **dashed** lines indicate maturity evolution paths for kerogen types I, II and III. Note the elongate distribution of data points, mainly reflecting variation in thermal maturity of the Bo Member in the different sub-basins. **B:** Total organic carbon (TOC) vs. S2 pyrolysis yield; note the overall organic richness and high pyrolysis yields.

takes place in deposits containing both type II and type III kerogen, apparently with no significant differences in enrichment factors (Disnar & Sureau 1990). Cuttings samples from this interval indicate the persistence of black laminated mudstones, suggesting that there were no significant changes in redox conditions in the sedimentary environment. The third possible factor, sedimentation rate, cannot be evaluated meaningfully on the basis of the available data.

In general terms, the geochemical characteristics of the Bo Member conform to those of the remainder of the Farsund Formation as outlined above. However, a number of specific characteristics serve to geochemically differentiate the Bo Member from the remainder of the Farsund Formation (Figs 14–16). Firstly, *n*-alkane distributions may display a slightly increased abundance of *n*C₁₅ and/or *n*C₁₇, and in the C_{20–28} range, a slight predominance of even carbon numbered components is sometimes noted, for example in the Edna-1 well. In the m/z 191 ion fragmentogram, the proportion of tricyclic triterpanes relative to pentacyclic triterpanes,

shown by the ratio of C₂₃ tricyclic triterpane (T23) to C₃₀ hopane (H30), may be increased (Fig. 14). The abundance of 28,30-bisnorhopane (H28) is often very high (Fig. 16), and in samples of low to moderate thermal maturity, this compound may even dominate the m/z 191 fragmentogram. Extended hopanes may be abundant. The C₃₅ and, in some cases, the C₃₃ homologues are slightly enriched, leading to C₃₅/C₃₄ homologue ratios close to unity or above.

Geochemical interpretation of the Bo Member

A predominance of *n*C₁₅ and *n*C₁₇ is generally assumed to indicate algal organic matter (Gelpi *et al.* 1970; Tissot & Welte 1984). An abundance of tricyclic triterpanes has been linked to the occurrence of *Tasmanites*-type alginite (Azevedo *et al.* 1992; Revill *et al.* 1994), which is present in large proportions in the Bo Member (Bojesen-Koefoed 1988). The presence of 28,30-bisnorhopane is

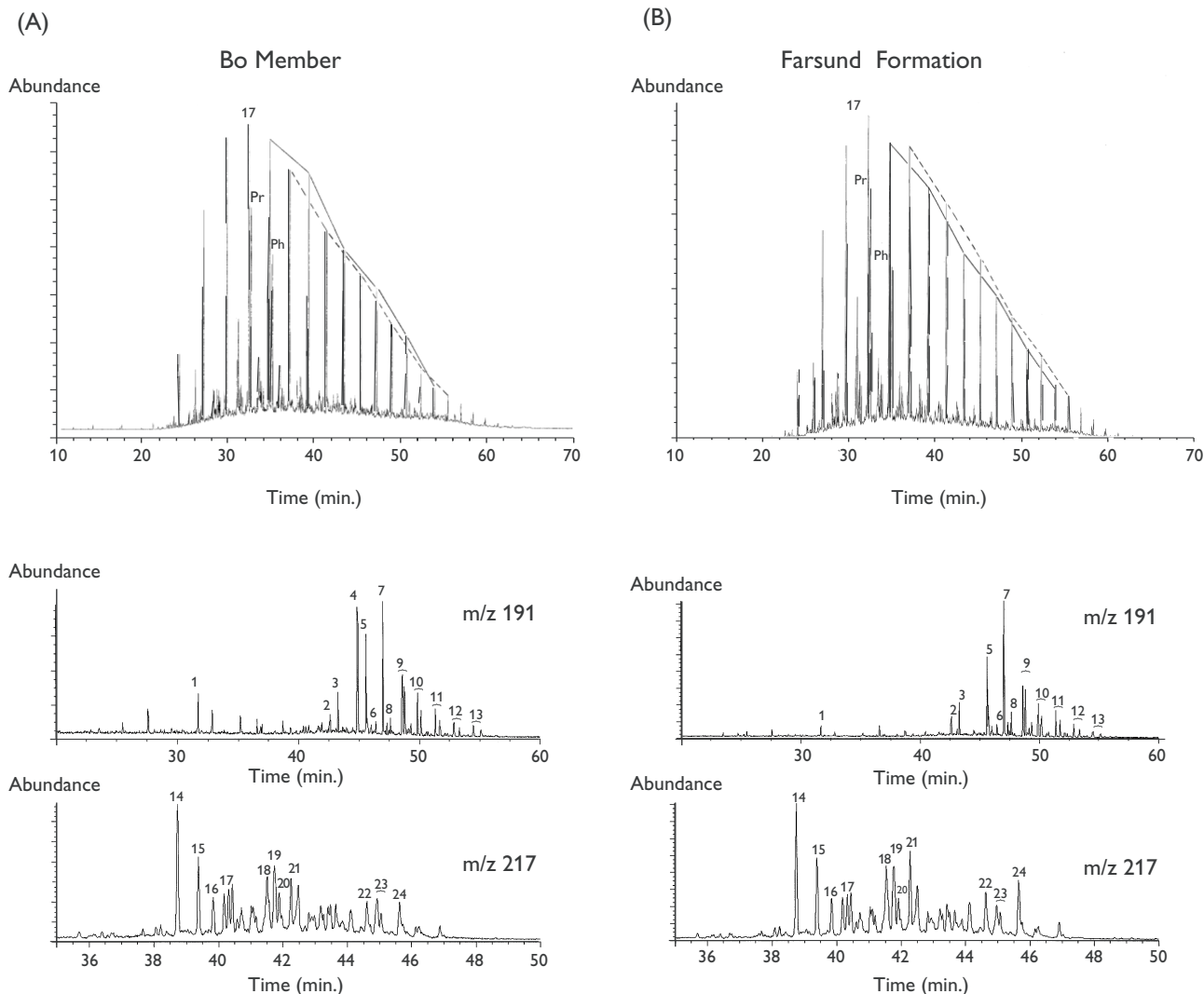


Fig. 16. Comparison of the geochemical characteristics of the Bo Member (A) and 'background' Farsund Formation mudstones (B). The uppermost pair of traces are gas chromatograms; the **solid** line connects *n*-alkanes with even numbers of carbon atoms, the **dashed** line connects *n*-alkanes with odd numbers of carbon atoms. Note the dominance of even-numbered *n*-alkanes in the Bo Member. The two pairs of traces beneath illustrate biomarker data, ion fragmentograms *m/z* 191 and *m/z* 217. Note the relative abundance of C₂₃ tricyclic triterpanes (peak number **1**) and 28,30-bisnorhopane (**4**) in the Bo Member mudstones. Additional peaks: **2**, Ts; **3**, Tm; **5**, norhopane; **6**, C₂₉-moretane; **7**, hopane; **8**, C₃₀-moretane; **9–13**, homohopanes; **14–17**, C₂₇-diasteranes; **18–21**, C₂₇ regular steranes; **22–24**, C₂₉ regular steranes.

generally assumed to indicate highly anoxic environments, and its occurrence has also been linked to bacterial activity (Katz & Elrod 1983; Williams 1984; Moldowan *et al.* 1985; Peters & Moldowan 1993). A predominance of even-numbered *n*-alkanes in the C_{20–28} range is often observed in mildly hypersaline depositional environments (Welte & Waples 1973; Nishimura & Baker 1986; Grimalt & Albaiges 1987; Bojesen-Koefoed *et al.* 1997) but has also been recorded from a marine setting with little terrigenous input (Kennicutt & Brooks 1990). A high proportion of homohopanes is generally favoured by strongly reducing environments (Peters &

Moldowan 1993), and enrichment in C₃₃ and C₃₅ homologues is often observed in carbonate and hypersaline environments (Mello *et al.* 1988). Hence, the biomarker distribution is indicative of anoxia, a minimal input of organic matter derived from higher land plants and perhaps of conditions of mild hypersalinity.

As noted earlier, the vast majority of commercial as well as non-commercial petroleum accumulations in the Danish North Sea can, with a high degree of certainty, be genetically related to the Farsund Formation (and its stratigraphic equivalents). A common feature of almost all crude oils from the Danish North Sea is

the presence of varying proportions of 28,30-bisnorhopane. This compound is thermally labile and its presence will to some extent be governed by maturity; indeed, the variation observed roughly parallels the maturity of the oils (unpublished GEUS data). Since 28,30-bisnorhopane is only found in appreciable proportions in the sediments of the Bo Member, it may be assumed that this member has contributed to most of the hydrocarbon occurrences in the Danish North Sea. This further supports the idea that the Bo Member is a persistent feature in the Danish Central Graben, and probably retains its identity in areas outside present well control, such as in the central portions of the Feda and Gertrud Grabens and along the western flank of the Central Graben. Furthermore, the occurrence of a large number of 'immature' oils, particularly in the southern part of the Danish Central Graben, all carrying notable proportions of 28,30-bisnorhopane, suggests that the kerogen of the Bo Member and the upper Farsund Formation in general is able to generate and expel petroleum at low levels of thermal maturity.

Discussion

Upper Jurassic organic-rich shales: current models

The importance of Upper Jurassic source rocks in Northwest Europe has resulted in numerous studies focussing on the mode of accumulation and preservation of organic matter in this intra-cratonic setting (e.g. Tyson *et al.* 1979; Oschmann 1988; Miller 1990; Wignall 1991a). There is general agreement that a stratified water column with anoxic or dysoxic bottom waters, at least periodically, is indicated by the lithofacies, biofacies, palynofacies and geochemistry. Rather less agreement has been reached, however, concerning the dominant mechanism(s) controlling such stratification and anoxia.

Tyson *et al.* (1979) suggested that stagnation and stratification in the Kimmeridgian sea was influenced both by the regional palaeogeography and the overall high sea level; the complexity of the former inhibited circulation and open ocean transfer while the increased water depths over the shelf favoured stratification of bottom waters beneath wave-base. In some respects, this resembles the 'silled basin model' of Demaison & Moore (1980), using the Quaternary of the Black Sea as a broad analogy; the onset of stratification on the Late Jurassic shelf was envisaged as a response to thermocline devel-

opment, however, rather than to salinity stratification. The role of sea-level variation in controlling the regional extent of bottom-water anoxia has been emphasised (Wignall 1991a; Wignall & Hallam 1991) while the link between cyclical variations in source rock development and climatic fluctuations has been discussed by a number of workers (e.g. Oschmann 1988).

Alternative models have been presented to explain the regional development of a stratified water mass in Northwest Europe during the Late Jurassic – earliest Cretaceous. Oschmann (1988) suggested that the 'silled basin' model was inappropriate, given the regional extent of the organic-rich facies and the well-established connections to the Tethyan and Boreal seas. He argued for the seasonal development of extensive anoxia due to the southward migration of cold, oxygen-poor Boreal waters during the summer in response to a northerly, wind-driven surface current. Turnover of this stratified system probably occurred during the winter in all but the deepest submarine grabens. Miller (1990) proposed a simple two-layer oceanographic model that is grossly the reverse of Oschmann (1988): cold oxygenated Boreal waters formed the surface layer flowing southward above warm saline bottom waters that originated in hypothetical shallow evaporative bays and flowed sluggishly northwards collecting in local depocentres and major rift axes. This model requires a sensitive balance between temperature and salinity of the two water masses since slight cooling of the Boreal waters or decrease in the elevated salinities of the warmer southern waters would result in complete overturn. Such a scenario, initiated by climatic or oceanographic shifts, was envisaged by Miller (1990) for the regional destabilisation of the stratified system in the latest Ryazanian (the basin 'flushing' of Rawson & Riley 1982).

Volgian–Ryazanian 'hot shales'

In discussion of the models proposed for the development of the 'Kimmeridge Clay Formation' sea, it should be noted that such models are based largely on onshore data, particularly from the type area (see discussion by Miller 1990). The facies are thus not directly comparable to those of the Central Graben, and indeed the depositional models are only strictly applicable to the period represented by the onshore section (i.e. Kimmeridgian – Middle Volgian). Furthermore, such models are designed to explain the long-term controls on the deposition and preservation of organic-rich shales that characterise the 'Kimmeridge Clay Formation' in all

its stratigraphic guises throughout Northwest Europe. The 'hot shales' of the uppermost Jurassic – lowermost Cretaceous in the Central Graben of the North Sea record a distinctive event within the background of organic-rich shale sedimentation, perhaps precipitated by the enhancement of one controlling factor or a coincidental combination of factors. The data presented here from the Danish Central Graben are assessed below in the light of the current models summarised above.

Stratigraphic continuity

In attempting to understand the origin of these organic-rich shales, it is clearly important to establish the regional extent and degree of synchronicity of these deposits. Black, organic-rich shales are a characteristic feature of the Kimmeridgian–Ryazanian of Northwest Europe and indeed farther afield (Ager 1975; Doré *et al.* 1985; Klemme 1994). Clearly, this represents a time during which burial and preservation of organic carbon was favoured on a global scale. It is also clear, however, that local factors such as structural configuration and sediment influx in addition to short-term global variables (e.g. sea level, climate) controlled the degree of development in any one location. In the North Sea basin, the regional development of organic-rich shales in the Kimmeridge Clay Formation (and stratigraphic equivalents) shows marked diachronism, although peak developments may be more widespread and biostratigraphically correlatable. Doré *et al.* (1985), discussed at length the temporal and geographic distribution of 'hot shales' in the North Sea region. These workers proposed that deposition of organic-rich 'hot shales' was most extensive, temporally, in the Viking Graben of the northern North Sea where much of the Kimmeridgian and Volgian stages are represented by anaerobic, organic-rich claystones in distal basin-centre locations. According to Doré *et al.* (1985), such facies are not well-developed in the Ryazanian succession of the Viking Graben, however, in contrast to the Central Graben farther south.

As noted earlier, 'hot shales' occur at a number of levels in the Upper Jurassic of the North Sea basin, the Middle Volgian – Ryazanian succession under focus here forming the uppermost and best-developed example (Price *et al.* 1993). In the Norwegian sector of the Danish Basin, hot shales of Kimmeridgian – early Volgian age form a mappable unit defined as the Tau Formation (Hamar *et al.* 1983; Doré *et al.* 1985). As observed by Rawson & Riley (1982), the onset of anoxia in the Kimmeridgian marked by the appearance of the Tau

Formation organic-rich shales corresponds to the base of the oil shale facies in the Eudoxus Chronozone in the type Kimmeridge Clay Formation of the Wessex Basin (Tyson 1996). This event is marked by thin 'hot' log spikes within the Farsund Formation of the southern Norwegian (Doré *et al.* 1985) and Danish sectors of the Central Graben (Johannessen *et al.* 1996). According to Wignall (1991b), this is one of the most important flooding events recorded in the onshore Kimmeridge Clay Formation.

From the Outer Moray Firth through the Central Graben to the Netherlands sector (Fig. 1), the Volgian–Ryazanian succession is characterised, to a greater or lesser extent, by 'hot' organic-rich shales. On the basis of available published data, it appears that the onset of the development of this organic-rich facies may not have been wholly synchronous. In the UK sector of the Central Graben, Donovan *et al.* (1993) ascribed an Early Volgian age to this event whereas a middle–late Middle Volgian age is likely in the Norwegian and Danish sectors (Doré *et al.* 1985; Dybkjær 1998). The same event has been referred to the late Middle – Late Volgian in the northern part of the Netherlands sector (G.F.W. Herngreen, personal communication 1995). Although these sources suggest a crude north-to-south younging of the initiation of 'hot shale' sedimentation, caution should be exercised. Firstly, there is a scarcity of well-documented biostratigraphic information and secondly, watermass stratification may well have been regionally synchronous throughout the Central Graben, yet local factors (organic productivity, siliciclastic dilution, basin morphology etc) may have dictated when organic-rich mudstones began to be preserved at any one location. In any event, existing data suggest that there is considerable overlap in the age of the most organic-rich facies; true 'hot shales' accumulated from the latest Volgian to the mid-Ryazanian in most parts of the Central Graben and the Moray Firth.

Depositional and structural setting

The data from the Danish sector reinforce the regional interpretations of the Upper Jurassic mudstones in the Central Graben and onshore (Tyson *et al.* 1979; Miller 1990; Tyson 1996). Sedimentological core data, combined with palynological and geochemical data, attest to predominantly anoxic conditions on the sea floor; suboxic conditions may have prevailed periodically in certain locations. Regional stratigraphic data from the Danish Central Graben indicate that the mid-Volgian –

Ryazanian succession is sediment-starved relative to the Kimmeridgian – mid-Volgian (Andsbjerg & Dybkjær 2003, this volume). This period of reduced sediment supply to basinal settings has been correlated with an arid climatic episode between the early Volgian and the Late Ryazanian (Hallam 1985; Wignall & Ruffell 1990; Hallam *et al.* 1991; Ruffell & Rawson 1994), an event which is also reflected in the upwards decrease in the terrigenous component of the kerogen of the Farsund Formation in the Danish Central Graben (see above).

With the exception of the easternmost portion of the Danish Central Graben, where increased siliciclastic supply is thought to have suppressed the organic signal, the Bo Member is recognised throughout the Danish Central Graben, even on Late Jurassic intrabasinal highs. Position within the complex segmented Central Graben clearly influenced the degree of development of the ‘hot shales’ (i.e. thickness, organic richness, timing of initiation etc. – see earlier discussion and Figs 4, 6). At its peak, however, watermass stratification and resultant bottom-water anoxia was sufficiently pervasive as to extend over most structural highs within the Danish Central Graben.

Role of sea-level variation

In assessing the possible contribution of sea-level variation to the development of a stratified watermass and bottom-water anoxia, it is important to recognise that the Volgian–Ryazanian succession records organic carbon burial and preservation at two different temporal scales. Enhanced preservation of organic carbon began in the middle–late Middle Volgian in the Danish sector and persisted until the early Late Ryazanian. The organic-rich shales of the Bo Member, however, record a discrete short-term pulse, essentially restricted to the Early Ryazanian, that was superimposed on the long-term trend. Referring, then, to published sea-level curves (Hallam 1988; Haq *et al.* 1988), it is apparent that the Middle Volgian – Late Ryazanian period was characterised, in broad terms, by a downward trend in sea level following the long-term sea-level rise that dominated most of the Late Jurassic and peaked in the Early Volgian.

Superimposed on this ‘second order’ fall in sea level, however, were a number of short-term sea-level events, as indicated by Rawson & Riley (1982, p. 2630) who referred to the Middle – Late Volgian period as a “strongly regressive phase with occasional transgression”. One such minor transgressive event is recorded, for example, by the quasi-marine Cinder Bed which records a

base Kochi Chronozone (Early Ryazanian) marine incursion into the largely non-marine Wessex Basin (Rawson & Riley 1982). An Early Ryazanian transgressive trend is also reflected in the regional onlap of Ryazanian ‘hot shales’ onto the flanks of the Central Graben in the UK sector (Gatliff *et al.* 1994), culminating in the deposition of Upper Ryazanian black laminated mudstones on Kimmeridgian strata in eastern England (Lott *et al.* 1986, 1989).

It is apparent, therefore, that conditions favourable for the deposition and preservation of organic-rich sediments in the Central Graben from the middle–late Middle Volgian to the early Late Ryazanian coincided broadly with a long-term fall in sea level. In contrast, the pulse of highly efficient burial of organic carbon in the Early Ryazanian, represented by the Bo Member, may correlate with a short-term sea-level rise; this is compatible with the evidence presented here from the lower boundary of the Bo Member of a marked decrease in sediment supply to the deep basin at this time.

Conclusions

It seems likely that the enhanced anoxia that developed in the Central Graben of the North Sea in the latest Jurassic and earliest Cretaceous was the result of a coincidence of several factors. Firstly, the Late Jurassic intracratonic seaway in the North Sea region was prone to stratification and the development of oxygen deficiency in bottom waters, probably as a result of regional palaeogeographic and partially related oceanographic factors (Tyson *et al.* 1979; Doré 1991). A number of additional factors combined to create conditions particularly favourable for the accumulation of highly organic-rich sediments in the late–middle Middle Volgian – early Late Ryazanian. An arid climate resulted in reduced sediment influx and, hypothetically, the development of saline waters in fringing evaporative bays that may have sourced the Central Graben bottom waters, thus enhancing watermass stratification (Miller 1990). A long-term ‘second order’ fall in sea level from the Middle Volgian to the Late Ryazanian, in combination with the complex tectonic topography of the North Sea, may have contributed to the poor circulatory system. It appears likely, however, that the Lower Ryazanian ‘hot shales’, represented by the Bo Member in the Danish sector, record a superimposed, short-term transgressive event (or events) that resulted in lateral expansion of the anoxic bottom waters onto the flanks of the Central Graben and over intra-basinal highs.

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