

Sequence stratigraphy of the Jurassic of the Danish Central Graben

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A sequence stratigraphic framework is established for the Jurassic of the Danish Central Graben based primarily on petrophysical log data, core sedimentology and biostratigraphic data from about 50 wells. Regional seismic lines are used to assist in the correlation of some wells and in the construction of isochore maps. In the Lower Jurassic (Hettangian–Pliensbachian) succession, five sequences have been identified. The Middle Jurassic is subdivided into four sequences that together span the uppermost Aalenian/lowermost Bajocian to the Callovian. In the Upper Jurassic, better well coverage permits greater stratigraphic resolution, and 11 sequences are identified and mapped.

On the basis of the sequence stratigraphic correlation and the construction of isochore maps for individual sequences, the Jurassic basin history of the Danish Central Graben can be subdivided into seven discrete phases: (1) Shallow marine and offshore sediments deposited in a pre-rift basin extending from the North Sea to the Fennoscandian Border Zone (Hettangian–Pliensbachian). (2) Uplift and erosion in association with a Toarcian–Aalenian North Sea doming event. A major hiatus represents this phase in the study area. (3) Terrestrial and marginal marine sedimentation during initial rifting (latest Aalenian/earliest Bajocian – Late Callovian). (4) Early Oxfordian – Early Kimmeridgian transgression during and after a rift pulse. The sedimentary environment changed from coastal plain and marginal marine to fully marine. (5) Regression associated with a cessation or slowing of subsidence during a structural rearrangement that took place in the Late Kimmeridgian during a break in the main rift climax. Shallow to marginal marine sandstones were deposited above an erosion surface of regional extent. (6) Deep-water mudstones deposited in a composite graben with high subsidence rates related to rift pulses (latest Late Kimmeridgian – middle Middle Volgian). (7) Deposition of organic-rich mudstones and turbidite sandstones during the late Middle Volgian – Early Ryazanian. The main basin shallowed, became more symmetrical and experienced a decreasing rate of subsidence, recording the onset of the post-rift stage.

A relative sea-level curve is constructed for the Middle–Late Jurassic. It shows close similarity to published eustatic (global) and relative (North Atlantic area) sea-level curves in the latest Bathonian – late Early Kimmeridgian, but differs in the Late Kimmeridgian – Middle Volgian interval, probably due to the high rate of subsidence in the study area.

Keywords: Danish Central Graben, Jurassic, sequence stratigraphy, palaeogeography, basin evolution, sea-level changes

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The Jurassic deposits in the Danish Central Graben are of particular interest in relation to hydrocarbon exploration, as they include units of reservoir sandstones as well as oil- and gas-producing source rocks (Fig. 1).

Hydrocarbon discoveries have been made in both Middle and Upper Jurassic sandstones, and the Middle Jurassic Harald Field is under production. The distribution of both reservoirs and source rocks reflects the complex

tectonic evolution of the area and a depositional history strongly influenced by relative sea-level changes. Data from a large number of wells in the study area have been released in recent years, contributing impor-

tant new information to this study. The aim of this paper is to present a sequence stratigraphic model for the Jurassic deposits of the Danish Central Graben.

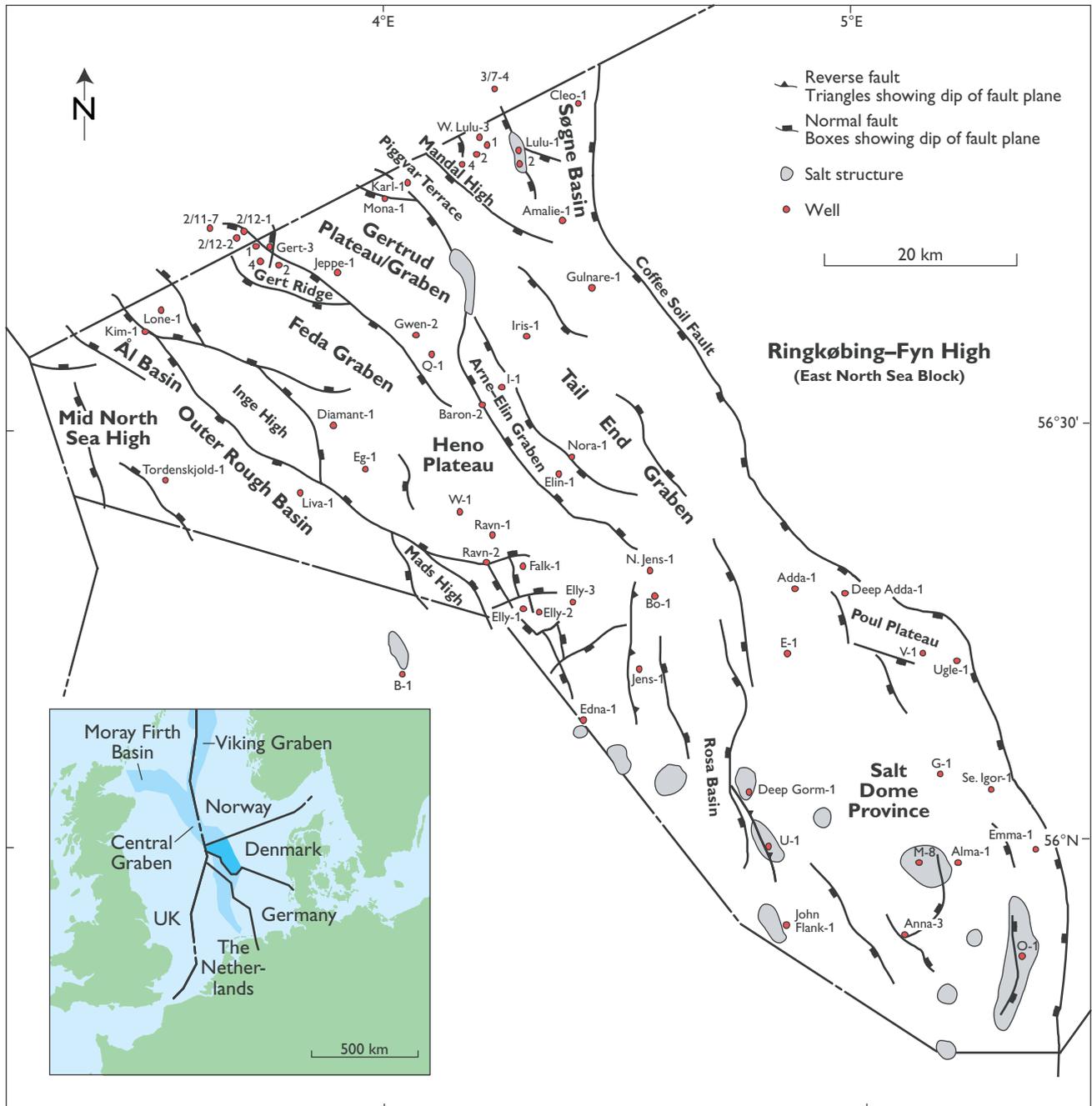


Fig. 1. Maps showing the location of the study area (dark blue) within the Jurassic North Sea rift system, the Jurassic structural elements in the Danish Central Graben and the position of the wells incorporated in the study; note that all wells forming the database for the isochore and palaeogeographic maps are indicated, although some are not specifically mentioned in the text or illustrated. The Middle Jurassic Harald Field is delimited by the location of the West Lulu-1, -2 and -3 wells; note that the structural feature denoted the Gertrud Plateau/Graben is so indicated to reflect the structural evolution of this region in the Late Jurassic, from a positive plateau in the early Late Jurassic to an actively subsiding graben from the Early Volgian.

Regional geology

The Central Graben forms the southern part of the North Sea rift system. Active rifting took place in this area both in the Triassic and in the late Middle Jurassic – Late Jurassic (Roberts *et al.* 1990; Ziegler 1990). The 500 km long Central Graben, approximately 150 km of which are situated in the Danish North Sea sector, consists of a NNW–SSE-trending complex of half-grabens, and a subordinate N–S-trending segment to the south extending into the German and Dutch sectors. The main bounding fault in the Danish sector is the Coffee Soil Fault, which forms the eastern margin of the Danish Central Graben (Fig. 1).

The Early Jurassic was characterised by a slow relative sea-level rise and deposition of marine muds of the Fjerritslev Formation, probably over most of the Danish area (Fig. 2; Larsen 1966; Michelsen 1978, 1989; Pedersen 1986; Michelsen *et al.* 1987).

In latest Early Jurassic – earliest Middle Jurassic times, the whole area was uplifted and most of the Lower Jurassic section was removed by erosion (Hallam &

Sellwood 1976; Ziegler 1982, 1990; Underhill & Partington 1993). The uplift may have occurred in response to the development of a rift dome extending 700 km in a north–south direction and 1000 km east–west across the central North Sea (Whiteman *et al.* 1975; Eynon 1981; Ziegler 1990; Underhill & Partington 1993). Domal uplift appears to have been accompanied by the development of a volcanic complex at the triple junction between the Viking Graben, the Central Graben and the Moray Firth Basin (Ziegler 1990). Sedimentation resumed in the Danish part of the Central Graben during the Middle Jurassic, with deposition of the sandstone-dominated Bryne and Lulu Formations and the mudstone-dominated Middle Graben Formation (Michelsen *et al.* 2003, this volume). These sediments are restricted to the Søgne Basin, the Tail End Graben and the Salt Dome Province (Vollset & Doré 1984; Jensen *et al.* 1986). The first marine transgression in the Danish Central Graben occurred during the Callovian–Oxfordian, probably reflecting the onset of domal collapse combined with eustatic sea-level rise (Ziegler 1990; Underhill & Partington 1993).

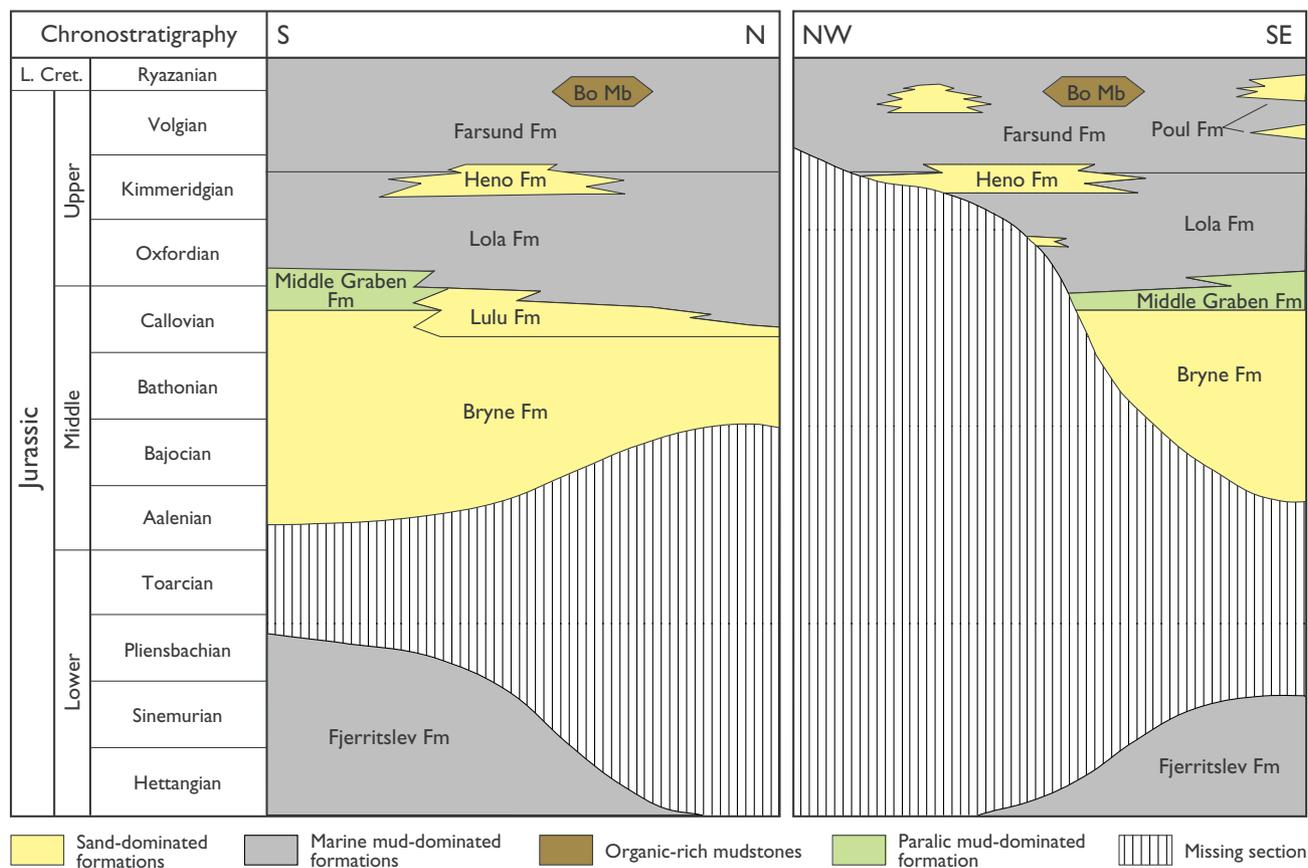


Fig. 2. Jurassic lithostratigraphy of the Danish Central Graben, based on Jensen *et al.* (1986) and Michelsen *et al.* (2003, this volume). **L. Cret.**, Lower Cretaceous.

During Late Jurassic time, the Feda Graben, the Heno Plateau and the Gertrud Graben became actively subsiding depositional basins, and the depositional area was later extended to the Outer Rough and Ål Basins (Fig. 1). Deep-water conditions were established in the Danish Central Graben during the Oxfordian–Kimmeridgian, when the marine mudstones of the Lola Formation were deposited. During the Late Kimmeridgian, shallow marine sandstones of the Heno Formation were deposited on plateau areas. The transgression culminated during the Volgian with deposition of the deep marine mudstones of the Farsund Formation (Vollset & Doré 1984; Jensen *et al.* 1986). The Early Cretaceous saw a change to more passive thermal subsidence (Roberts *et al.* 1990).

In addition to rift tectonics, basin development in the Central Graben was also strongly influenced by the presence of mobile Zechstein salt. Salt movements had a profound influence on the development of depocentres in the Søgne Basin, the Tail End Graben and the Salt Dome Province (Korstgaard *et al.* 1993; Sundsbø & Megson 1993).

Stratigraphic methods

The stratigraphic analysis is based on data from about 50 wells that penetrate Jurassic deposits in the Danish Central Graben, and a few wells from the Norwegian North Sea sector located near the boundary (Fig. 1). Well logs, including gamma-ray, sonic, neutron/density and resistivity logs, sedimentological core logs, lithology logs, and biostratigraphic data were used in the study. Furthermore, a number of interpreted regional seismic lines were included to support the generation of isochore maps.

Biostratigraphy

The biostratigraphic data used in this study are confined to palynomorphs because information from other groups of microfossils is too scarce in most wells to be useful for detailed correlation. The biostratigraphic correlations are based on events rather than recognition of biozones. The events used in our study are mainly last occurrence datums (i.e. first downhole appearance) of dinoflagellate cyst species (Fig. 3). The biostratigraphic information utilised in the study includes published data (Birkelund *et al.* 1983; Hoelstad 1986; Poulsen 1986, 1991; Heilmann-Clausen 1987; Johannessen *et al.* 1996), unpublished reports of the Geological Survey of

Denmark and Greenland and of industrial service companies, and the results of new analyses made specifically for this study.

Due to poor core coverage, being generally restricted to sandy successions of relatively few wells, most data were derived from cuttings samples. Although sample quality in the Lower Jurassic succession was variable, it was possible to establish a palynostratigraphic framework based on a small number of events and stratigraphically diagnostic palyno-assemblages (Fig. 3A). A detailed palynostratigraphical subdivision of the Middle Jurassic succession is precluded by the sparse occurrence of age-diagnostic palynomorphs (Fig. 3B); although scanty, the available records made it possible to give an approximate age for the sequences and major events in the Middle Jurassic. In the Upper Jurassic, the large sediment thicknesses, the somewhat better data quality, the favourable environment, and the higher diversity of dinoflagellate cyst species probably all contribute to a higher biostratigraphic resolution (Fig. 3C).

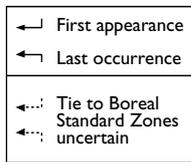
Sequence stratigraphy

The sequence stratigraphic terminology applied is that introduced by the Exxon group (Posamentier *et al.* 1988; Posamentier & Vail 1988; Van Wagoner *et al.* 1988, 1990). Sequences are subdivided into lowstand, transgressive and highstand systems tracts. A four-fold subdivision with the addition of the forced regressive systems tract (Hunt & Tucker 1992, 1995) is not attempted due to the scale of the study and the nature of the data.

Construction of the sequence stratigraphic framework is based on correlation of well log patterns, combined with biostratigraphic datings. The most conspicuous log patterns, supported by the most important biostratigraphic events, were used to produce a coarse grid, within which more detailed correlations were made. Most commonly the gamma-ray logs were used for high-resolution correlations but resistivity logs formed the primary tool in some organic-rich mudstone units. Distinct sonic log markers were occasionally used for correlation. Maximum flooding surfaces (MFS) are considered the most reliable correlation surfaces in successions dominated by marine mudstones, whereas sequence boundaries (SB), and occasionally flooding surfaces (FS), are of primary importance in silty and sandy units. Both SB and MFS key surfaces can normally be traced through most of the basin. In sections dominated by marine mudstones and siltstones, the SB is usually picked at the turnaround point between a coars-

C

Time in Ma (Gradstein <i>et al.</i> 1994)	Epoch	Age	Boreal Standard Zones	Bioevents (Palynology)				
				First appearance datums	Last occurrence datums			
140	Early Cretaceous	Valanginian	Paratollia					
			Ryazanian	Late	Albidum	←	[<i>Dingodinium spinosum</i> ⁴⁾ <i>Dichadogonyaulax culmula</i> ⁴⁾	
		Stenomphalus						
		Icenii						
		Early		Kochi	←	<i>Rotosphaeropsis thula</i> ⁴⁾		
				Runctoni	←	<i>Egmontodinium expiratum</i> ⁴⁾		
		145		Late Jurassic	Volgian	Late	Lamplughi	
							Preplicomphalus	
						Middle	Primitivus	←
			Oppressus Anguiformis Kerberus Okusensis Glaucolithus				←	<i>Muderongia simplex</i> *
Early	Albani		←		[<i>Scriniodinium irritabile</i> [<i>Gochteodinia mutabilis</i> <i>Senoniasphaera jurassica</i> ^{6)**}			
	Fittoni		←		<i>Occiscysta balia</i>			
Kimmeridgian	Late		Rotunda Pallasioides		←	<i>Perisseiasphaeridium pannosum</i> <i>Oligosphaeridium patulum</i>		
			Pectinatus		←	<i>Cribroperidinium longicorne</i>		
	Early		Hudlestoni					
			Wheatleyensis					
		Scitulus						
		Elegans	←	<i>Subtilisphaera? paeminosa</i>				
155	Late Jurassic	Kimmeridgian	Autissiodorensis	←	<i>Endoscrinium luridum</i>			
			Eudoxus					
			Mutabilis	←	[<i>Subtilisphaera? paeminosa</i> ³⁾ <i>S.? inaffecta</i> ³⁾			
		Oxfordian	Early	Cymodoce	←	<i>Stephanelytron scarburghense</i> <i>Endoscrinium galeritum</i>		
				Baylei				
			Late	Rosenkrantzi	←	[<i>Scriniodinium crystallinum</i> <i>Nannoceratopsis pellusida</i>		
				Regulare				
				Serratum				
				Glosense				
				Compositosphaeridium polonicum	←	<i>Rigaudella aemula</i>		
Middle	Tenuiserratum	←	<i>Wanaea</i> spp.					
	Densiplicatum	←	<i>Ctenidodinium continuum</i>					
	Early	Cordatium						
Mariae								



southern Feda Graben and the northernmost Heno Plateau. Most key surfaces in the Upper Jurassic succession are picked in marine mudstones and siltstones. Of these key surfaces, only one sequence boundary (the base Kimm-2 SB) shows any sign of significant erosion.

Unusually high organic contents in the Upper Volgian – Ryazanian mudstones of the Bo Member (Michelsen

et al. 2003, this volume) probably influence log responses and impede sequence stratigraphic interpretation (Ineson *et al.* 2003, this volume). In the uppermost sequence of this study, Ryaz-1, a maximum flooding surface is not identified. It might have been picked at the highest gamma-ray peak in accordance with conventional sequence stratigraphic concepts, but in this case all of

the Bo Member may form part of the transgressive systems tract as suggested for this type of deposits by Posamentier & James (1993). This level marks the top of the studied succession. In those wells where the uppermost part of the Farsund Formation is preserved, it is dated as Late Ryazanian.

Sequence stratigraphic subdivision and basin history

The Jurassic succession in the Danish Central Graben is subdivided into 20 sequences, referred to according to their age (i.e. Hettangian sequences: Hett-1, Hett-2). The main characteristics of each sequence – log expression of sequence boundaries, lithology, thickness, age – are summarised and illustrated below. The lower boundary is described for each sequence, the upper boundary being described under the succeeding sequence. Lithologies in the sequence descriptions have been interpreted primarily on the basis of gamma-ray logs, supported by other log types and by core data when available. In the sandstone-dominated intervals, increasing gamma-ray values are interpreted to reflect fining-upwards grain-size trends, and decreasing values to reflect coarsening-upwards trends. In the mudstone successions, changing gamma-ray values are interpreted to reflect changes in clay content; increasing gamma-ray values are thus interpreted to record increasing clay content and decreasing gamma-ray values to record decreasing clay content. Thin carbonate beds are represented by low-value gamma-ray spikes. Mudstone intervals with exceptionally high gamma-ray values are interpreted as being particularly rich in organic matter. Based on these inferred relationships between gamma-ray values and lithologies, increasing-upwards gamma-ray patterns are generally interpreted to be characteristic of the transgressive systems tract (TST) and decreasing-upwards gamma-ray patterns of the highstand systems tract (HST). The maximum flooding surface (MFS) is picked at the turnaround point where gamma-ray values are at a maximum between the increasing gamma-ray values of the TST below and decreasing gamma-ray values of the HST unit above.

The identified sequences are of approximately the same order of thickness. Due to variations in subsidence rate, the duration of the sequences recognised in the Lower and Middle Jurassic is somewhat longer than for the Upper Jurassic sequences. All sequences, however, probably correspond to 3rd order sequences as defined by Vail *et al.* (1977) and Van Wagoner *et al.* (1990).

The evolution of the Danish Central Graben can be considered in terms of seven depositional phases on the basis of the subdivision of the Jurassic deposits into sequences and systems tracts. The significant influence of tectonics on depositional patterns in rift settings has been demonstrated by a number of studies (Surlyk & Clemmensen 1983; Rosendahl 1987; Gabrielsen *et al.* 1990; Prosser 1993; Nøttvedt *et al.* 1995). However, we have attempted to relate the development of the depositional phases to the effects of both tectonics and eustasy.

The nomenclature and subdivision of rifting phases introduced by Prosser (1993) and adapted by Nøttvedt *et al.* (1995) are used in an attempt to relate the depositional phases of the present study to rifting stages. The proto-rift stage is characterised by flexural subsidence and interrupted by domal uplift. The subsequent rift stage is subdivided into rift initiation and rift climax: rift initiation is associated with fault block compartmentalisation and weak tilting whereas the rift climax is characterised by the maximum rate of fault displacement and strong tilting. The early post-rift stage is marked by the cessation of significant activity at the main faults and a declining rate of regional subsidence. It should be appreciated, however, that the three stage model of Nøttvedt *et al.* (1995) refers to the evolution of rift basins in general. In the present study, the model has been applied to minor basins in a large rift structure and the evolution of these basins may not be representative of the rift structure as a whole. This is emphasised by the diachronous onset of subsidence in neighbouring basins, as described below.

In the present study, several pulses of active faulting are inferred to have taken place between the initiation of the Middle Jurassic rift stage and the Early Cretaceous early post-rift stage. We have followed Surlyk (1978, 1989) and Blair & Bilodeau (1988) in correlating overall fining-upwards successions in rift basins with periods of active tectonic subsidence, and large-scale coarsening-upwards successions with periods of relative tectonic quiescence.

Phase 1. Pre-rift shallow marine deposition (Hettangian–Pliensbachian)

Sequence definitions

Hett-1 sequence

The Hett-1 sequence is present only in the Deep Gorm-1, M-8 and U-1 wells; it consists of marine mudstones. The lower sequence

boundary is located at the change from the low gamma-ray values of the underlying Triassic deposits to the high gamma-ray values, characteristic of the mudstones of the Hett-1 sequence (Figs 4, 5). The boundary may be abrupt or gradational. The thickness of the sequence attains a maximum of 23 m in M-8.

Stratigraphically useful bioevents have not been recorded from this unit. The sequence is thought to be of Hettangian age due to its stratigraphic position in the lowermost part of the Fjerritslev Formation, unconformably overlying Upper Triassic deposits.

Hett-2 sequence

The marine mudstones of the Hett-2 sequence have only been found in six wells but are considered to extend throughout the Salt Dome Province. The boundary between the Hett-1 and Hett-2 sequences is located at a shift to mudstones characterised by somewhat lower clay contents (Figs 4, 5). In wells where the Hett-2 sequence directly overlies pre-Jurassic sediments (e.g. Edna-1; Fig. 5), the lower sequence boundary is marked by an abrupt lithological shift. The thickness of the sequence shows very little variation, attaining a maximum of 64 m in Edna-1.

The abundance of bisaccate pollen of *Pinuspollenites minimus* within this sequence in the Edna-1, John Flank-1 and O-1 wells indicates the presence of the *Pinuspollenites-Trachysporites* Zone of Lund (1977). This zone is known from the Danish Basin, Skåne and northern Germany and is referred to the Hettangian (Lund 1977; Dybkjær 1991). A Hettangian age is further supported by the occurrence of the dinocyst species *Dapcodinium priscum* in the wells mentioned above and the absence of the pollen species *Cerebropollenites macroverrucosus* (Fig. 3A).

Sin-1 sequence

The marine mudstones of this sequence probably extend throughout the Salt Dome Province. The lower sequence boundary is typically located at a marked shift to mudstones with higher clay contents (Figs 4, 5). The thickness of the sequence attains a maximum of 63 m in Deep Gorm-1.

The last occurrence datum (LOD) of *Dapcodinium priscum* in the lower part of the sequence in the O-1 well, and the first appearance datum (FAD) of *Cerebropollenites macroverrucosus* in core samples from the middle part of the sequence in the Deep Gorm-1 well, indicate a latest Hettangian – Early Sinemurian age for the lower part of the sequence. In combination with the LOD of *Liasidium variabile* close to the upper boundary of this sequence in O-1, this indicates a latest Hettangian – Sinemurian age for the sequence (Fig. 3A).

Pliens-1 sequence

This sequence is only present in the Deep Gorm-1 and Edna-1 wells situated in the north-western part of the Salt Dome Province. In both wells, it is represented by marine mudstones. The lower sequence boundary is located at the top of the well-defined coarsening-upwards

HST of the Sin-1 sequence (Figs 4, 5). The thickness of the sequence attains a maximum of 63 m in Edna-1.

The common occurrence (in sidewall cores) of pollen referred to the genus *Chasmasporites* indicates the presence of the *Chasmasporites* Zone (Koppelhus & Nielsen 1994), defined in the Korsodde section, Bornholm, Denmark. The *Chasmasporites* Zone is referred to the Pliensbachian, an age assignment supported by the stratigraphic position of this sequence immediately above the LOD of *Liasidium variabile*. The acme of the small, spherical pollen referred to the genus *Spheripollenites*, characteristic of the uppermost Pliensbachian – Lower Toarcian deposits in the Danish Basin and known from many other locations in Northwest Europe (Dybkjær 1991), has not been reported from the Danish Central Graben. This may indicate that deposits of latest Pliensbachian – Toarcian age are absent in the study area.

Pliens-2 sequence

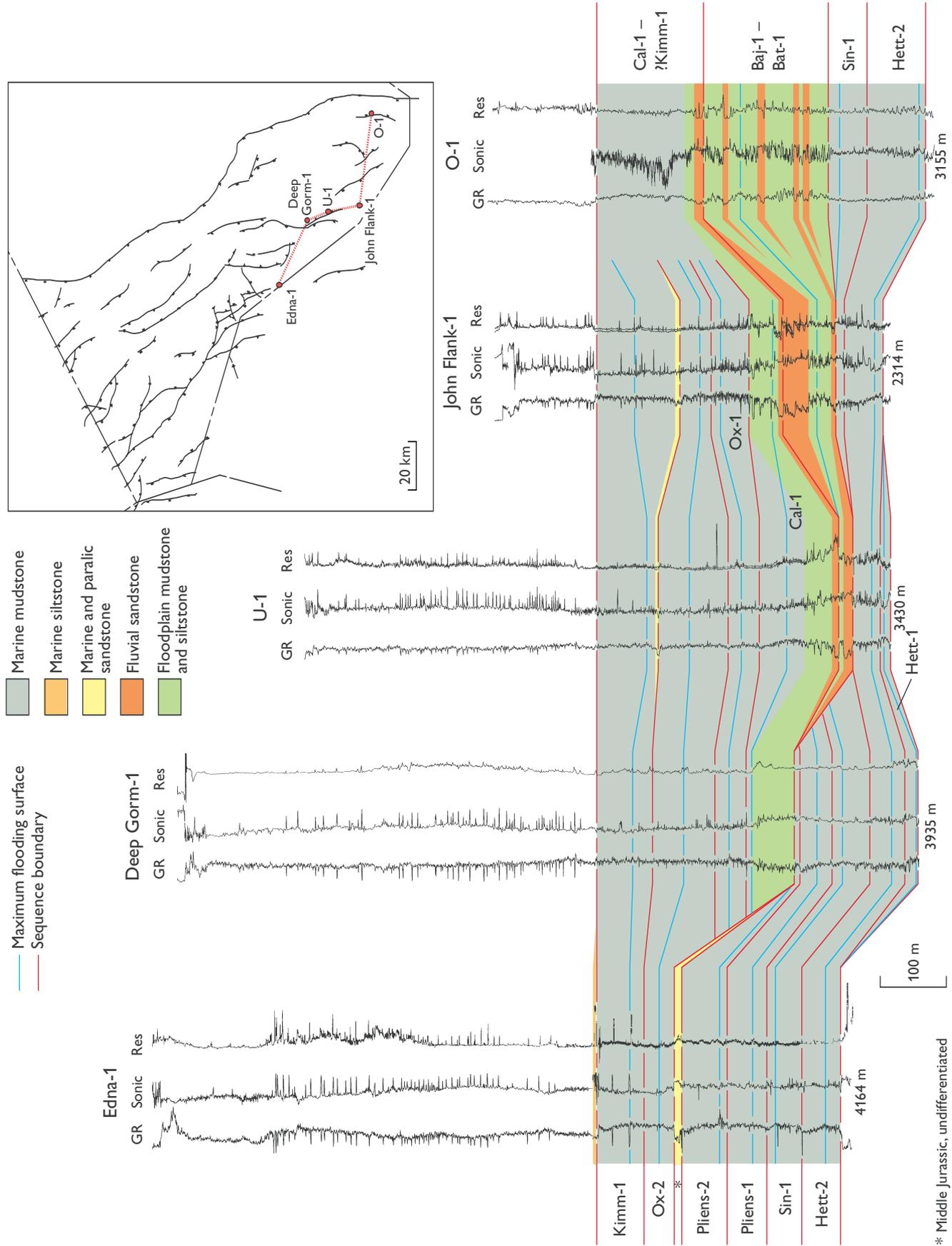
The marine mudstones of this sequence have only been recognised in the Deep Gorm-1 and Edna-1 wells (Fig. 4). The sequence boundary is positioned where the lithology changes abruptly from silty mudstones of the Pliens-1 HST to more clay-rich mudstones (Fig. 5). In the Deep Gorm-1 well, the sequence is represented by less than 10 m of clay-rich mudstone; in Edna-1, the sequence attains a thickness of 62 m.

The common occurrence (in sidewall cores) of *Chasmasporites* sp. pollen in Edna-1 and the absence of a *Spheripollenites* sp. acme and of other Toarcian bioevents (such as the FAD of the spore genera *Leptolepidites*, *Ischyosporites*, *Manumia* or *Staplinisporites*), indicate a Pliensbachian age for this sequence.

Basin evolution

The Early Jurassic was a tectonically quiescent period between phases of active rifting in the Triassic and in the Middle–Late Jurassic (Cartwright 1991; Nøttvedt *et al.* 1995). A eustatic rise in sea level during the Early Jurassic was proposed by Hallam (1981) and Haq *et al.* (1988), compatible with observations from the Danish Basin (Michelsen 1978, 1989; Pedersen 1986; Dybkjær 1988, 1991). Uniform lithologies, dominated by shelf mudstones, indicate that marine conditions existed across a major North Sea basin and extended into the Danish Basin (Michelsen 1978, 1982, 1989; Michelsen *et al.* 1987; Pedersen 1985; Nielsen 2003, this volume).

The Lower Jurassic marine mudstones within the study area have only been found in the Salt Dome Province. They were probably deposited over a wider area, but were removed by erosion during the regional uplift phase (phase 2) at the Early–Middle Jurassic transition (Andersen *et al.* 1982; Gowers & Sæbøe 1985; Underhill & Partington 1993). The extent of Lower



* Middle Jurassic, undifferentiated

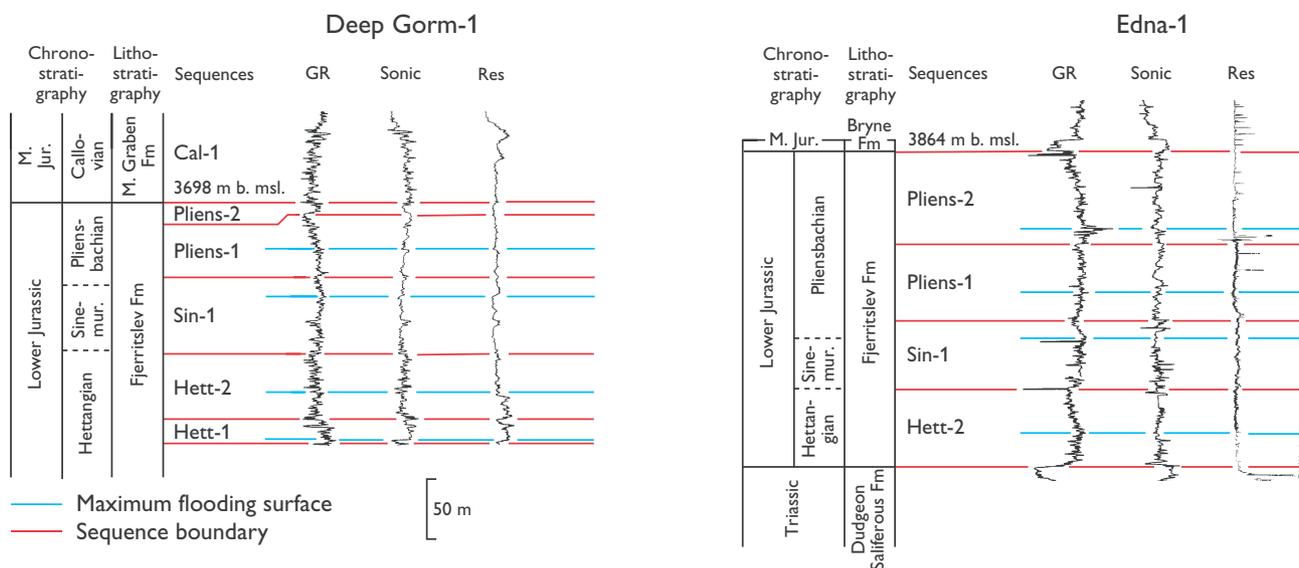


Fig. 5. Lower Jurassic sequences (Hett-1 – Pliens-1) in the Deep Gorm-1 and Edna-1 wells. For reference purposes, the depth (in metres below mean sea level) is indicated for a selected sequence boundary, typically in the upper levels of the illustrated section (see also Figs 6, 10, 12, 19, 22). **Jur.**, Jurassic; **M.**, Middle; **Sinemur.**, Sinemurian.

Jurassic deposits in the deepest parts of the Tail End Graben and the Søgne Basin is unknown due to the lack of wells penetrating the base of the Middle Jurassic in these areas.

Five transgressive–regressive cycles, corresponding to the Hett-1 – Pliens-2 sequences are identified in the Lower Jurassic succession (e.g. Deep Gorm-1, Edna-1; Fig. 4). Individual sequences show a remarkable lateral consistency in thickness (Fig. 4), indicating a uniform subsidence history in the study area during most of the Early Jurassic.

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Fig. 4. Sequence stratigraphic correlation of the Lower, Middle and lowermost Upper Jurassic in the Salt Dome Province. The Lower Jurassic succession is truncated by an unconformity at the base of the Middle Jurassic. The depth given beneath each well in this and subsequent correlation panels indicates the depth of the lowermost tieline (typically base-Jurassic section) below well reference level; for those few wells that did not penetrate the Jurassic section, the depth indicates the base of the logged section (i.e. well total depth). The log suite utilised in the study and illustrated in this and subsequent figures includes gamma-ray (**GR**), sonic velocity (**Sonic**) and resistivity (**Res**) logs.

Phase 2. Uplift and erosion in the proto-rift phase (latest Pliensbachian – latest Aalenian)

Basin evolution

A major unconformity separates Lower Jurassic deposits from the overlying Middle and Upper Jurassic deposits over much of the North Sea area. It has been suggested that this unconformity was caused by domal uplift in the Toarcian to Aalenian, centred at the triple junction between the Viking Graben, the Moray Firth Basin and the Central Graben (Whiteman *et al.* 1975; Hallam & Sellwood 1976; Eynon 1981; Ziegler 1982; Underhill & Partington 1993). The unconformity may therefore be an example of a proto-rift unconformity, described by Nøttvedt *et al.* (1995) as being typical of rift systems with thermally-induced domal uplift before or at the onset of active stretching.

As a result of the regional uplift, most of the Lower Jurassic deposits were removed. In the southern part of the Danish Central Graben, where Lower Jurassic deposits are preserved, differential erosion of the upper part of the succession may be due to local salt tectonics. Regional erosion caused a larger part of the succession to be removed over the top of the salt structures. The full Lower Jurassic succession shows thickness variations from about 60 m in U-1 to about 260 m in Edna-1 (Fig. 4).

Reworked Lower Jurassic palynomorphs recorded from the Upper Kimmeridgian sandstones of the Heno

Formation in the northern part of the Danish Central Graben (Gwen-2; Johannessen *et al.* 1996) supports the assumption that Lower Jurassic deposits were significantly more extensive prior to mid-Jurassic erosion.

Phase 3. Terrestrial and marginal marine deposition during the rift initiation stage (latest Aalenian/earliest Bajocian – Late Callovian)

Sequence definitions

Aalen-1 sequence

This sequence is encountered in most wells drilled in the Søgne Basin and in the Nora-1 well in the Tail End Graben (Figs 6, 7). It lies unconformably on either pre-Jurassic or Lower Jurassic deposits. The TST dominates the sequence and consists of a number of minor, backstepping, fining-upwards sandstone intervals, interpreted as stacked fluvial channels, with subordinate floodplain or lacustrine mudstones. The MFS is a distinct gamma-ray high, situated close to the strongly erosional upper sequence boundary. The thickness of the sequence attains a maximum of 69 m in West Lulu-1 (Fig. 6).

No stratigraphically useful bioevents were recorded from this sequence, but the lack of Toarcian bioevents (as discussed above with respect to the Pliens-1 and Pliens-2 sequences) and the LOD of *Kekryphalospora distincta* in the sequence above, may be seen as indirect evidence of an Aalenian or earliest Bajocian age.

Baj-1 sequence

The Baj-1 sequence extends throughout the Søgne Basin and the Tail End Graben (Nora-1), and further south to the northern and eastern parts of the Salt Dome Province (e.g. Alma-1, O-1). It may also be present in structurally deep locations elsewhere in the Salt Dome Province. The lower sequence boundary is a pronounced erosional surface (Figs 4, 6, 7). The lower part of the sequence typically consists of two laterally extensive channel sandstones separated by a fining-upwards/coarsening-upwards fine-grained interval (e.g. Amalie-1; Fig. 6). The channel sandstone interval may be interpreted as a low-stand systems tract (LST). Overlying this interval is a fining-upwards TST (Amalie-1, West Lulu-1; Fig. 7). The HST consists of floodplain sandstones and mudstones. The thickness of the sequence attains a maximum of 99 m in Amalie-1.

The LOD of *Kekryphalospora distincta* in the upper part of this sequence in the Alma-1 well indicates an Aalenian or earliest Bajocian age for the sequence (Fig. 3B).

Bat-1 sequence

This sequence is present in all wells that encountered Middle Jurassic deposits in the Søgne Basin, the Tail End Graben, the Salt Dome Province and the southernmost Heno Plateau (Figs 6, 7). In most wells,

the lower sequence boundary is located at the base of a fining-upwards channel sandstone (e.g. West Lulu-1; Fig. 6). The MFS is located in a mudstone interval, several metres thick. The HST is rarely present due to erosion; where present (e.g. West Lulu-1; Fig. 6), it consists of a coarsening-upwards interval of interbedded mudstones and sandstones. The thickness of the sequence attains a maximum of 64 m in West Lulu-1.

The occurrence (in a core sample) of *Adnatosphaeridium caulleryi* in the lower part of this sequence in West Lulu-1 indicates an age no older than Bathonian (Fig. 3B). In combination with the occurrence of *Impletosphaeridium varispinosum* immediately above the upper boundary of the sequence in West Lulu-3, this indicates a possible age range for this sequence of latest Bajocian – earliest Callovian.

Cal-1 sequence

This sequence is present in areas where the Bat-1 sequence is also found (Fig. 8A). In most wells, the lower sequence boundary is a very distinct erosional surface (see Amalie-1; Fig. 6), which locally shows truncation of at least 10–20 m (e.g. West Lulu-3; Fig. 7). The sequence boundary is overlain by a section of fluvial to estuarine channel sandstones up to 30 m thick (e.g. Amalie-1), which is interpreted as the fill of an incised valley. This interval probably represents the LST and the lowermost part of the TST. In the Søgne Basin and the Tail End Graben, the remainder of the TST consists of a succession of sandstone-dominated, paralic to shallow marine deposits, overlain by fully marine mudstones (Fig. 7). The paralic deposits include a number of coal beds, up to 3 m thick. The HST consists of a rather indistinct coarsening-upwards interval of marine mudstones. The sequence has a depocentre in the northern part of the Tail End Graben (Fig. 8A). In the Nora-1 well, where the sequence includes a section that contains tuffaceous sediments (4986–4932 m), it attains a thickness of 288 m. In other wells, the sequence does not exceed 100 m in thickness.

The LOD of *Impletosphaeridium varispinosum* in the lower part of this sequence in West Lulu-3 combined with the LOD of *Ctenidodinium continuum* in the sequence above (Ox-1) in U-1 and West Lulu-1 indicate an Early Callovian to earliest Oxfordian age for this sequence. The occurrence of *Wanaea acollaris* and *W. thysanota* in the middle part of the sequence in the John Flank-1 and West Lulu-1 wells supports this age assignment (Fig. 3A).

Basin evolution

Deposition resumed in the Danish Central Graben with the accumulation of uppermost Aalenian or lowermost Bajocian terrestrial sediments after the regional uplift and possible incipient collapse of the domal structure. Subsequent Middle Jurassic deposits rest unconformably on pre-Jurassic and Lower Jurassic sediments (Fig. 4).

The initiation of syn-rift subsidence is demonstrated by an asymmetric distribution of the sediment package in the Søgne Basin as shown by seismic evidence (Møller 1986, fig. 5; Cartwright 1987; Korstgaard *et al.* 1993, figs 39, 40) and well log data (Andsbjerg 2003, this volume). Salt tectonics also influenced sediment distribution and the location of local depocentres in the Søgne Basin, the Tail End Graben and the Salt Dome Province (Mogensen *et al.* 1992; Korstgaard *et al.* 1993).

Middle Jurassic deposits are preserved in the Søgne Basin, the Tail End Graben and the Salt Dome Province with a depocentre located to the east near the Coffee Soil Fault (Figs 4, 7). Sandstones and mudstones of sequences Aalen-1 – Bat-1 were deposited in a fluvially dominated environment during the early part of the Middle Jurassic (latest Aalenian/earliest Bajocian – Late Bathonian; Fig. 9A). Deeper parts of the basins were periodically inundated by lakes. Lacustrine conditions in the central parts of the basins were coeval with wet floodplain conditions in marginal locations.

The occurrence locally of rare marine palynomorphs and tidal indicators such as flaser bedding and double mud-drapes suggest that deposition took place in a coastal plain setting. However, during this period, fully marine conditions are only reported from the Dutch part of the

Central Graben (van Adrichem Boogaert & Kouwe 1993). A coastline must therefore have been located in the southern part of the Danish sector or in the German sector of the Central Graben. The regional transport direction in the Danish Central Graben was probably towards the coastline in the south, parallel to the basin axis.

Within the Middle Jurassic succession, several levels of well-developed, erosionally-based channels, prograding deltas and splays, and lacustrine and floodplain mudstone units form correlatable units over a wide area, comprising the Søgne Basin and the Tail End Graben. This suggests a common, external control on base level, such as rift tectonics, eustatic sea-level change, climate change or a combination of these factors.

In the Søgne Basin, the northern Tail End Graben and possibly in the Salt Dome Province, the deposits of the Aalen-1 – Bat-1 sequences are cut by a significant erosion surface, the basal sequence boundary of Cal-1 (Fig. 7). The development of this surface was caused by a major fall in relative sea level. At the time of formation of this surface, the regional slope had changed from a southwards to a northwards dip (Andsbjerg 2003, this volume). The erosional surface, which is commonly developed at the base of extensive fluvial or estuarine channel sandstones, is suggested to be the bounding

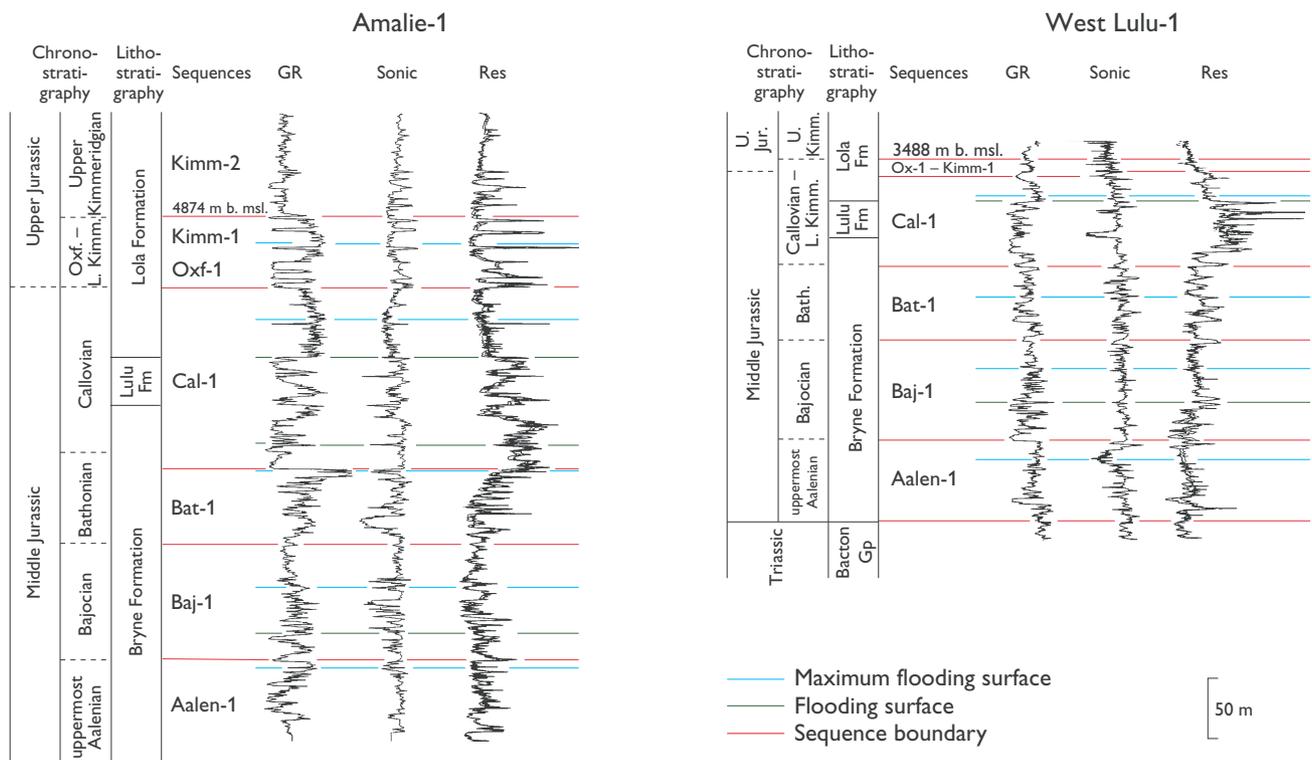
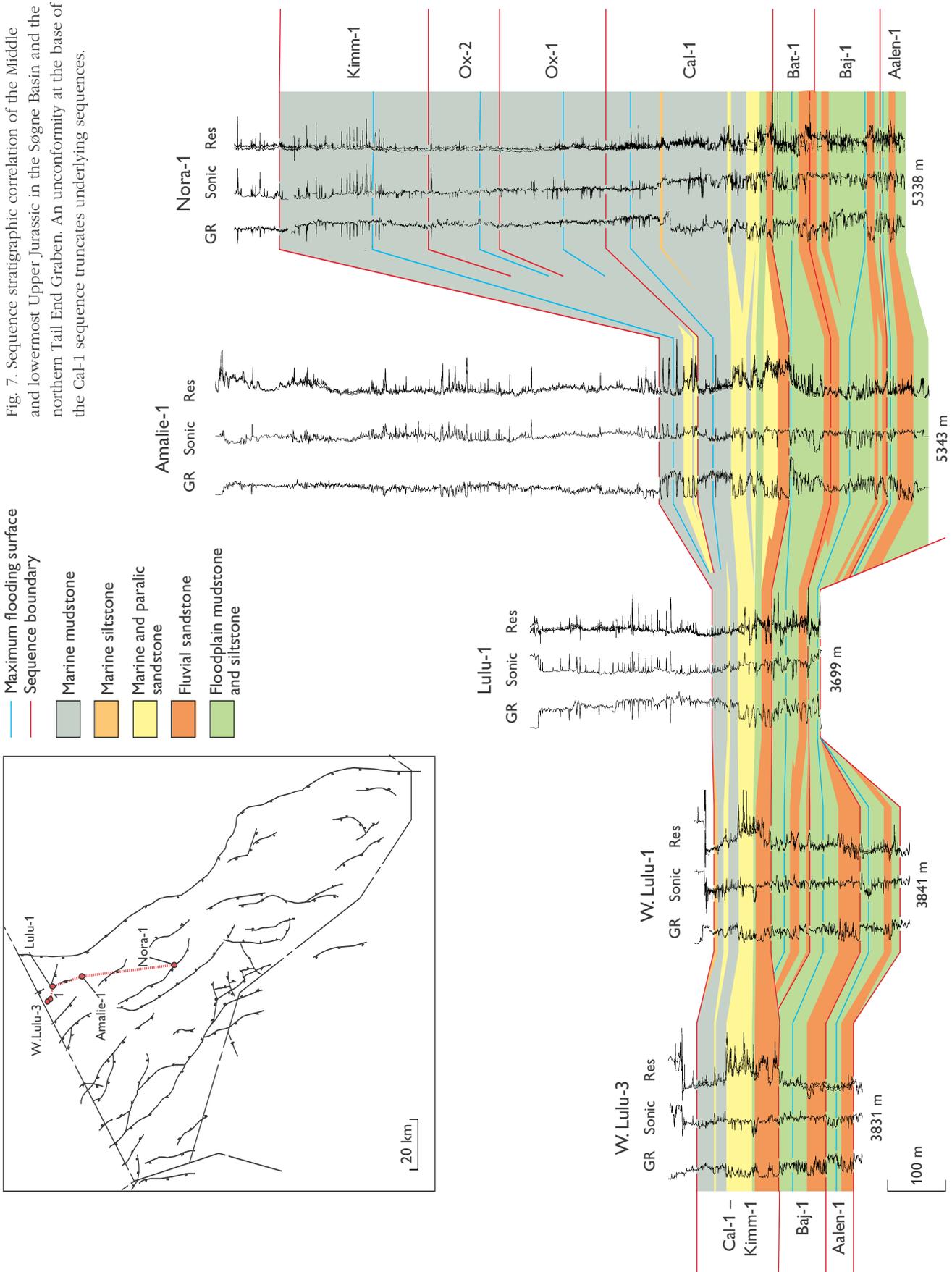


Fig. 6. Middle–Late Jurassic sequences in the Amalie-1 and West Lulu-1 wells illustrating, in particular, the characteristics of the Middle Jurassic (Aalen-1 – Cal-1) sequences. **Bath.**, Bathonian; **Jur.**, Jurassic; **Kimm.**, Kimmeridgian; **L.**, Lower; **Oxf.**, Oxfordian; **U.**, Upper.

Fig. 7. Sequence stratigraphic correlation of the Middle and lowermost Upper Jurassic in the Søgne Basin and the northern Tail End Graben. An unconformity at the base of the Cal-1 sequence truncates underlying sequences.



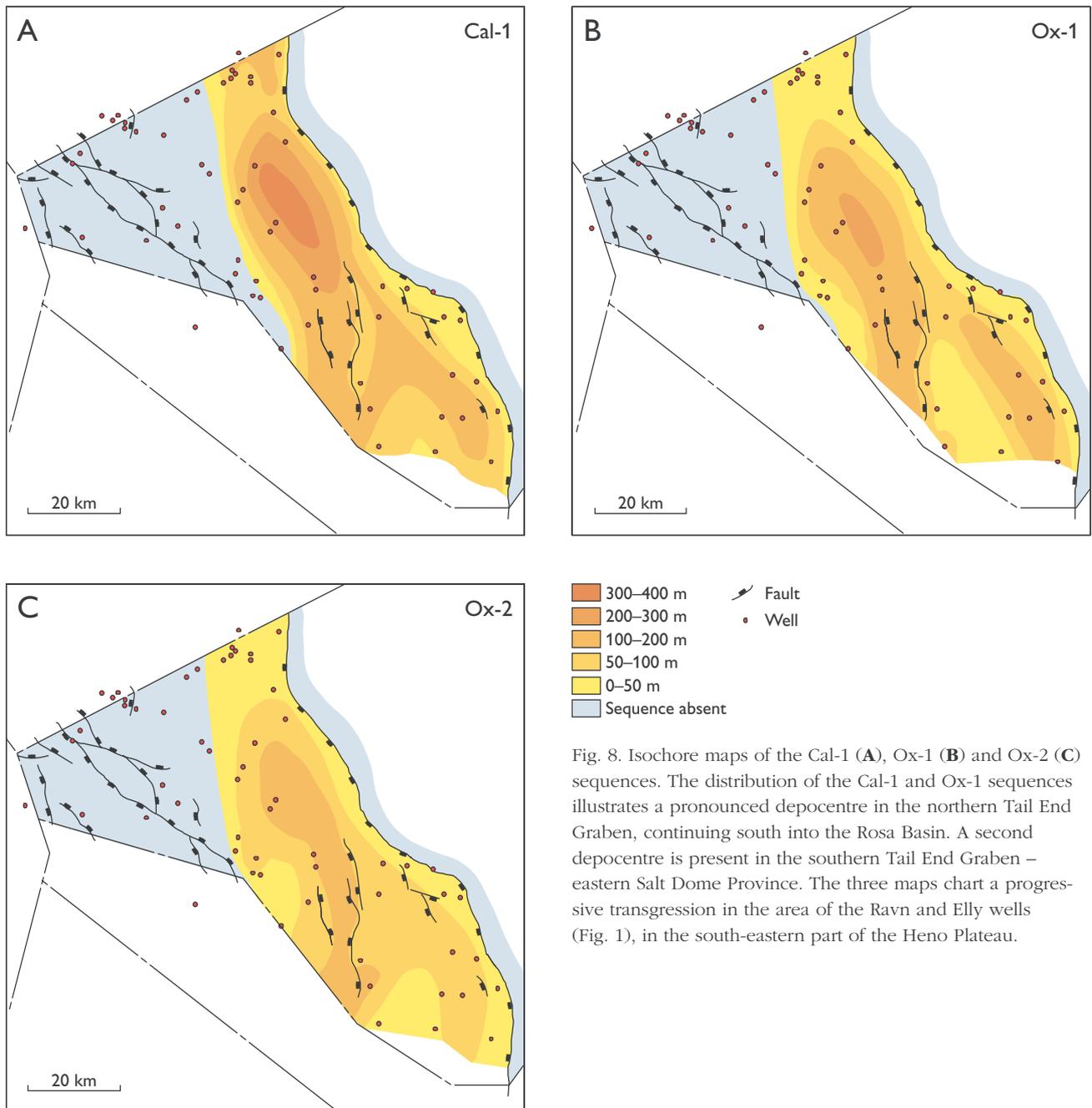


Fig. 8. Isochore maps of the Cal-1 (A), Ox-1 (B) and Ox-2 (C) sequences. The distribution of the Cal-1 and Ox-1 sequences illustrates a pronounced depocentre in the northern Tail End Graben, continuing south into the Rosa Basin. A second depocentre is present in the southern Tail End Graben – eastern Salt Dome Province. The three maps chart a progressive transgression in the area of the Ravn and Elly wells (Fig. 1), in the south-eastern part of the Heno Plateau.

surface of an incised valley (Andsbjerg 2003, this volume). The valley-fill is dated to the latest Late Bathonian – Early Callovian. In the Søgne Basin, the channel sands were initially deposited in straight or sinuous rivers, which show an increase in tidal influence with time. In the Salt Dome Province, channel sandstones are either stacked or isolated and may show fining-upwards trends typical of sinuous channel fills or blocky gamma-ray motifs, indicating deposition in a straight or braided river (e.g. John Flank-1, U-1; Fig. 4; Koch 1983). The

incised valley-fill of the Søgne Basin is capped by a coal bed up to 3 m thick.

Later in the Callovian, a low-energy coastal plain or delta plain characterised by small distributary channels, lagoons and coal swamps, was established in the southern part of the study area (Koch 1983). Contemporaneously, a barrier coast separated tidal lagoons at the margins of the Søgne Basin and the northern Tail End Graben from a wave-dominated marine bay in the central and eastern parts of these basins (Fig. 9B).

Phase 4. Rift pulse related transgression (Early Oxfordian – Early Kimmeridgian)

Sequence definitions

Ox-1 sequence

This sequence is restricted to the Tail End Graben, the deeper parts of the Søgne Basin, the Salt Dome Province and the flanks of the Heno Plateau (Fig. 8B). The lower boundary of the sequence is located at the top of a well-developed coarsening-upwards interval (e.g. Deep Gorm-1; Nora-1; Fig. 10). At the western margin of the

basin, a sandy LST up to 12 m thick may be present, erosively overlying pre-Jurassic deposits (e.g. Elly-2; Fig. 10). In most wells, the sequence consists of marine mudstones; only in Elly-2 are sandstones a major component of the sequence. The thickness of the sequence attains a maximum of 189 m in Nora-1. A significant depocentre occurs in the northern part of the Tail End Graben (Fig. 8B).

The LOD of *Ctenidodinium continuum* in the middle part of the sequence in U-1 indicates an Early Oxfordian age for at least the lower part of the sequence. Combined with the LOD of *Rigeaudella aemula* close to the upper boundary in the Falk-1 well, this indicates an Early–Middle Oxfordian age for the sequence (Fig. 3C).

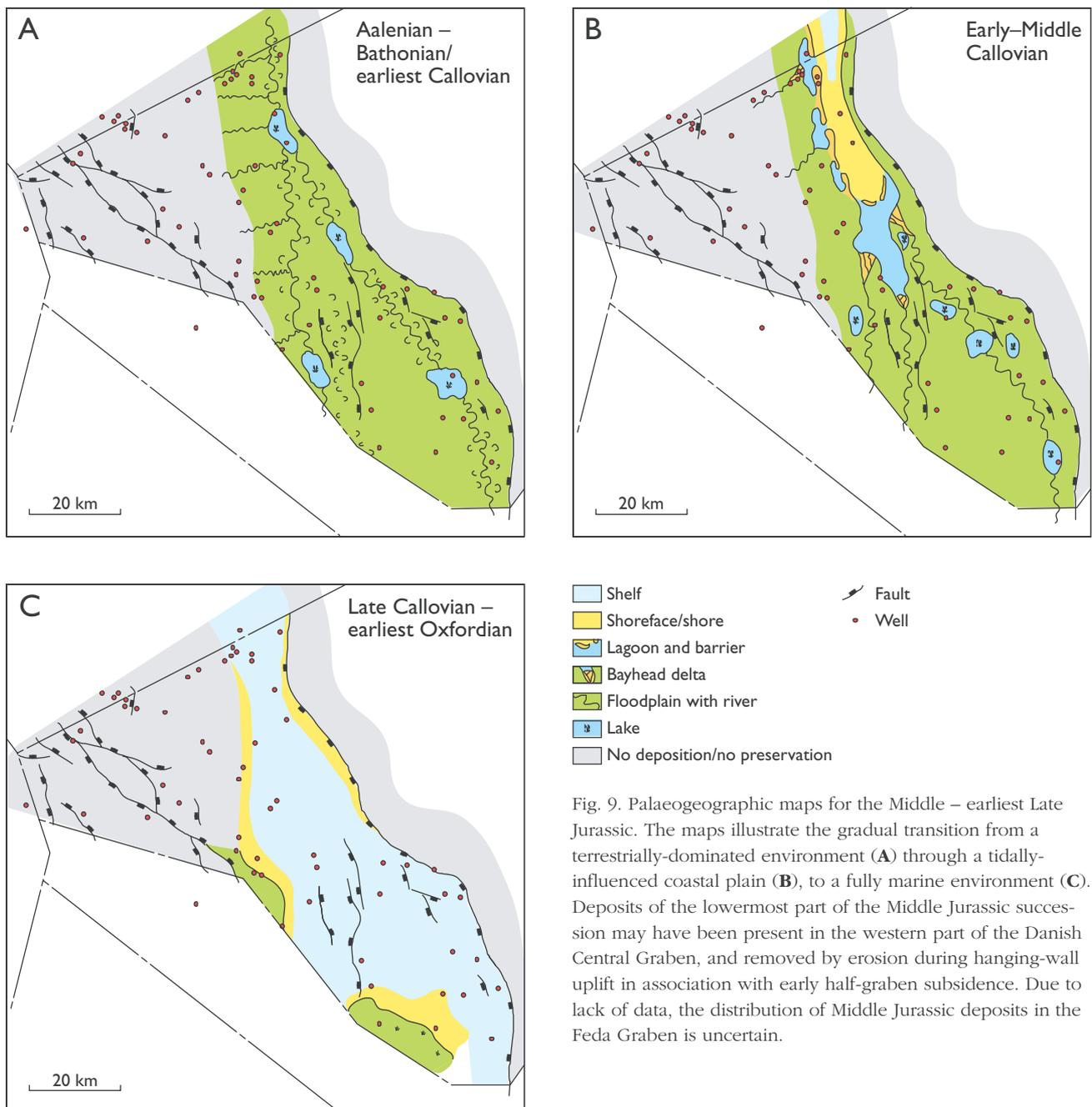


Fig. 9. Palaeogeographic maps for the Middle – earliest Late Jurassic. The maps illustrate the gradual transition from a terrestrially-dominated environment (A) through a tidally-influenced coastal plain (B), to a fully marine environment (C). Deposits of the lowermost part of the Middle Jurassic succession may have been present in the western part of the Danish Central Graben, and removed by erosion during hanging-wall uplift in association with early half-graben subsidence. Due to lack of data, the distribution of Middle Jurassic deposits in the Feda Graben is uncertain.

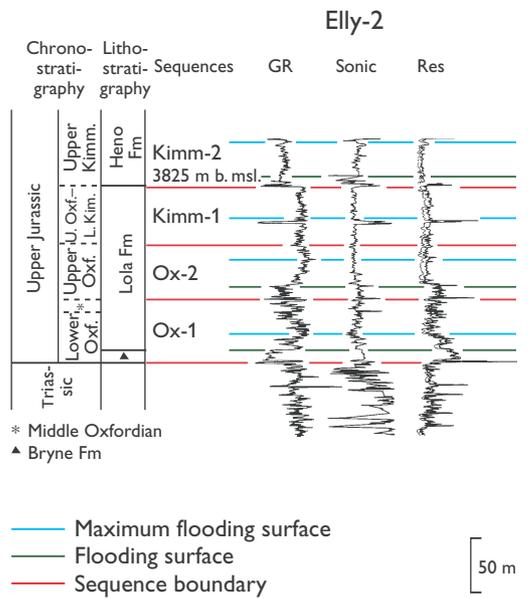
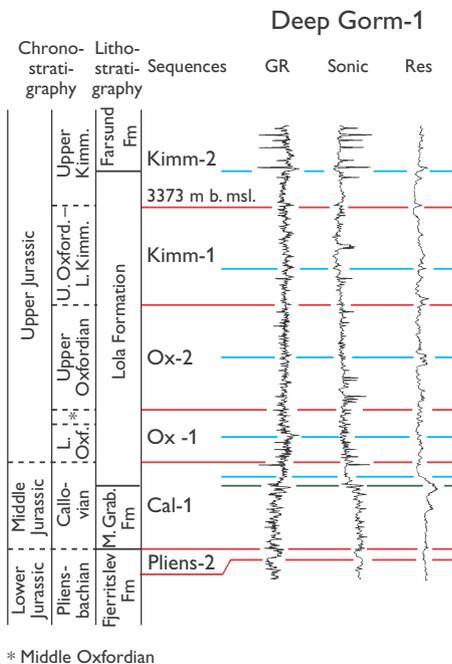
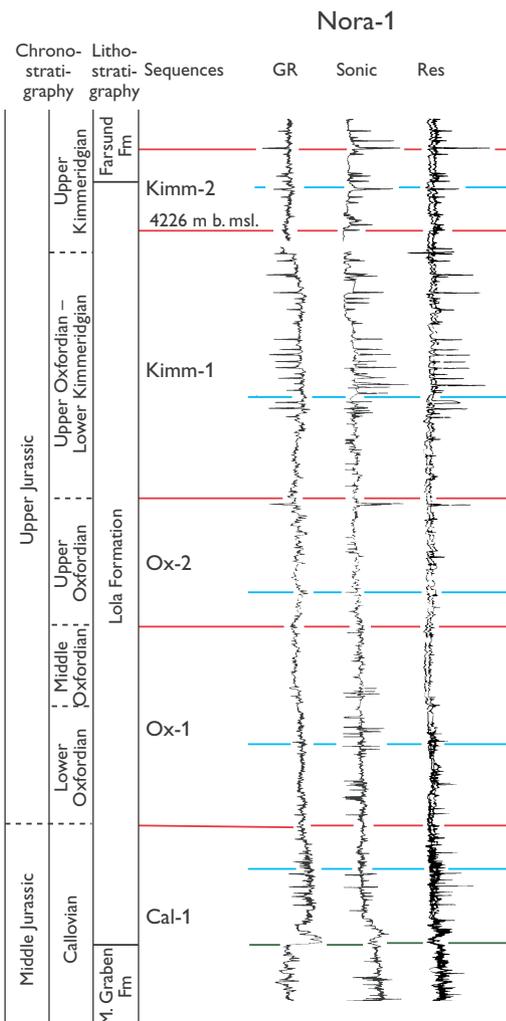


Fig. 10. Sequence stratigraphic breakdown of the lower Upper Jurassic in the Deep Gorm-1, Elly-1 and Nora-1 wells illustrating the features of the sequences Ox-1 – Kimm-1. **Kim./Kimm.**, Kimmeridgian; **L.**, Lower; **M. Grab.**, Middle Graben; **Oxf./Oxford.**, Oxfordian; **U.**, Upper.



Ox-2 sequence

The distribution of this sequence corresponds to that of Ox-1 (Fig. 8C). The lower sequence boundary is normally located at the top of a coarsening-upwards interval (e.g. Nora-1; Fig. 10). In a few wells situated at the western margin of the basin, the lower sequence boundary may be located at the sharp base of an up to 10 m thick sandstone unit referred to the LST (e.g. Elly-2). With the exception of these sandstones at the basin margin, the sequence consists of marine mudstones. The thickness of the sequence attains a maximum of 117 m in Nora-1. A depocentre for this sequence extends from the northern Tail End Graben to the Rosa Basin (Fig. 8C).

The LOD of *Rigeaudella aemula* immediately below the lower sequence boundary in Falk-1 combined with the LOD of *Compositosphaeridium polonicum* close to the upper boundary of this sequence in U-1 indicates an earliest Late Oxfordian age for this sequence (Fig. 3C).

Kimm-1 sequence

The sequence is missing from the north-western part of the study area (Fig. 11A). The lower boundary of the sequence is normally located at the top of a coarsening-upwards interval (e.g. Nora-1; Fig. 10). In the Heno Plateau area, the MFS is normally situated immediately above a marked gamma/sonic spike (e.g. Elly-2, Ravn-1; Figs 10, 12).

The sequence consists of marine mudstone. The maximum measured thickness is 250 m (Nora-1). The sequence has a well-defined depocentre in the central part of the Tail End Graben (Fig. 11A).

The LOD of *Compositumsphaeridium polonicum* immediately below the lower boundary of this sequence in U-1 and the LOD of *Scriniodinium crystallinum* and of *Nannoceratopsis pellucida* close to the upper boundary in a number of wells (e.g. Amalie-1, Anne-3, Emma-1, John Flank-1, M-8, U-1, West Lulu-2) indicate a Late Oxfordian – Early Kimmeridgian age for this sequence (Fig. 3C).

Basin evolution

In the Tail End Graben, subsidence along the Coffee Soil Fault accelerated in the Early Oxfordian resulting in a highly asymmetric distribution of the Oxfordian – Lower Kimmeridgian sediment package and in very large thicknesses of sediments in the central and eastern parts of the Tail End Graben (Figs 8A, 13; Fig. 13 faces page 294). This suggests that the Tail End Graben had entered the rift climax stage. During this phase, fault-controlled sub-

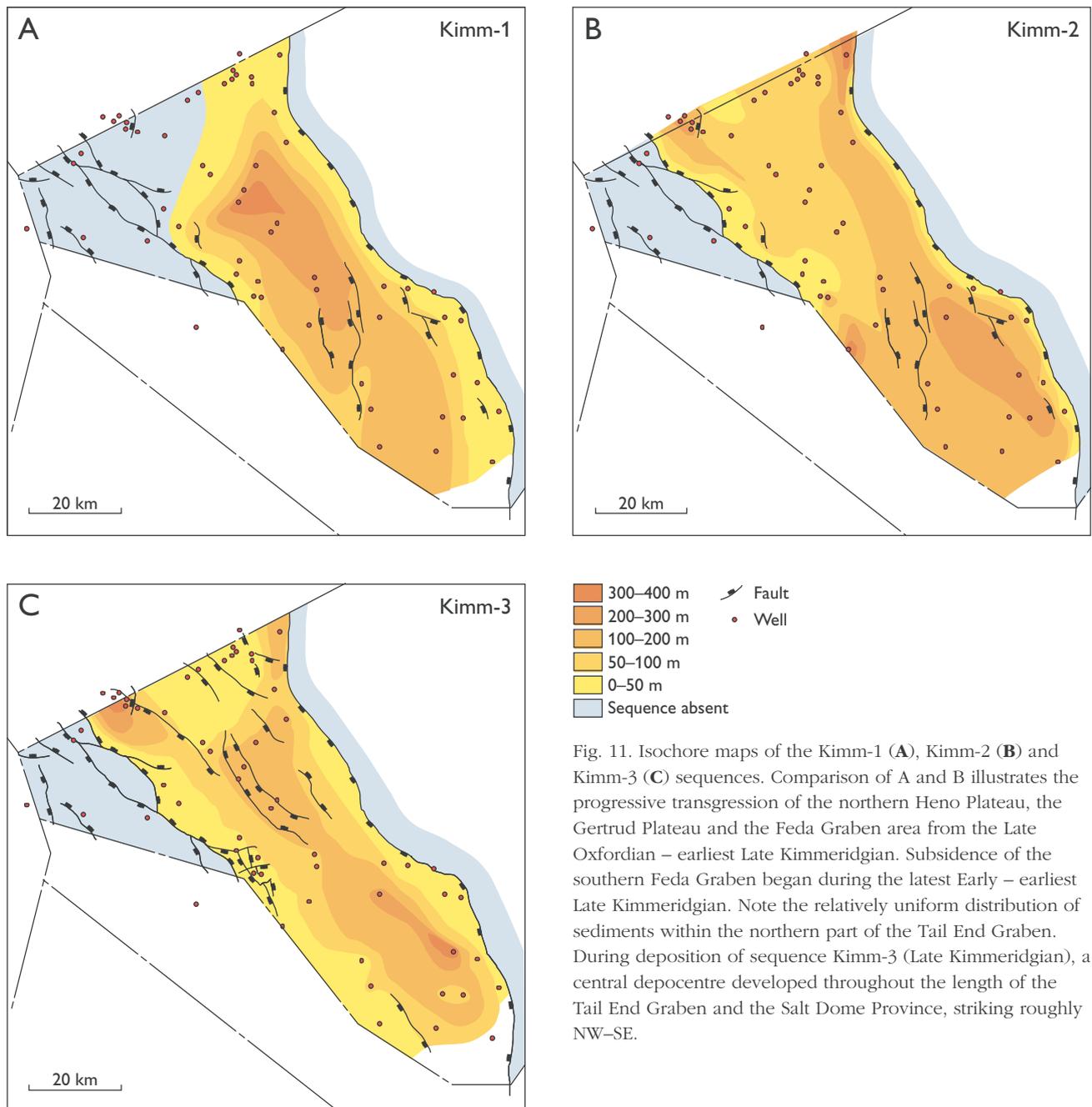


Fig. 11. Isochore maps of the Kimm-1 (A), Kimm-2 (B) and Kimm-3 (C) sequences. Comparison of A and B illustrates the progressive transgression of the northern Heno Plateau, the Gertrud Plateau and the Feda Graben area from the Late Oxfordian – earliest Late Kimmeridgian. Subsidence of the southern Feda Graben began during the latest Early – earliest Late Kimmeridgian. Note the relatively uniform distribution of sediments within the northern part of the Tail End Graben. During deposition of sequence Kimm-3 (Late Kimmeridgian), a central depocentre developed throughout the length of the Tail End Graben and the Salt Dome Province, striking roughly NW–SE.

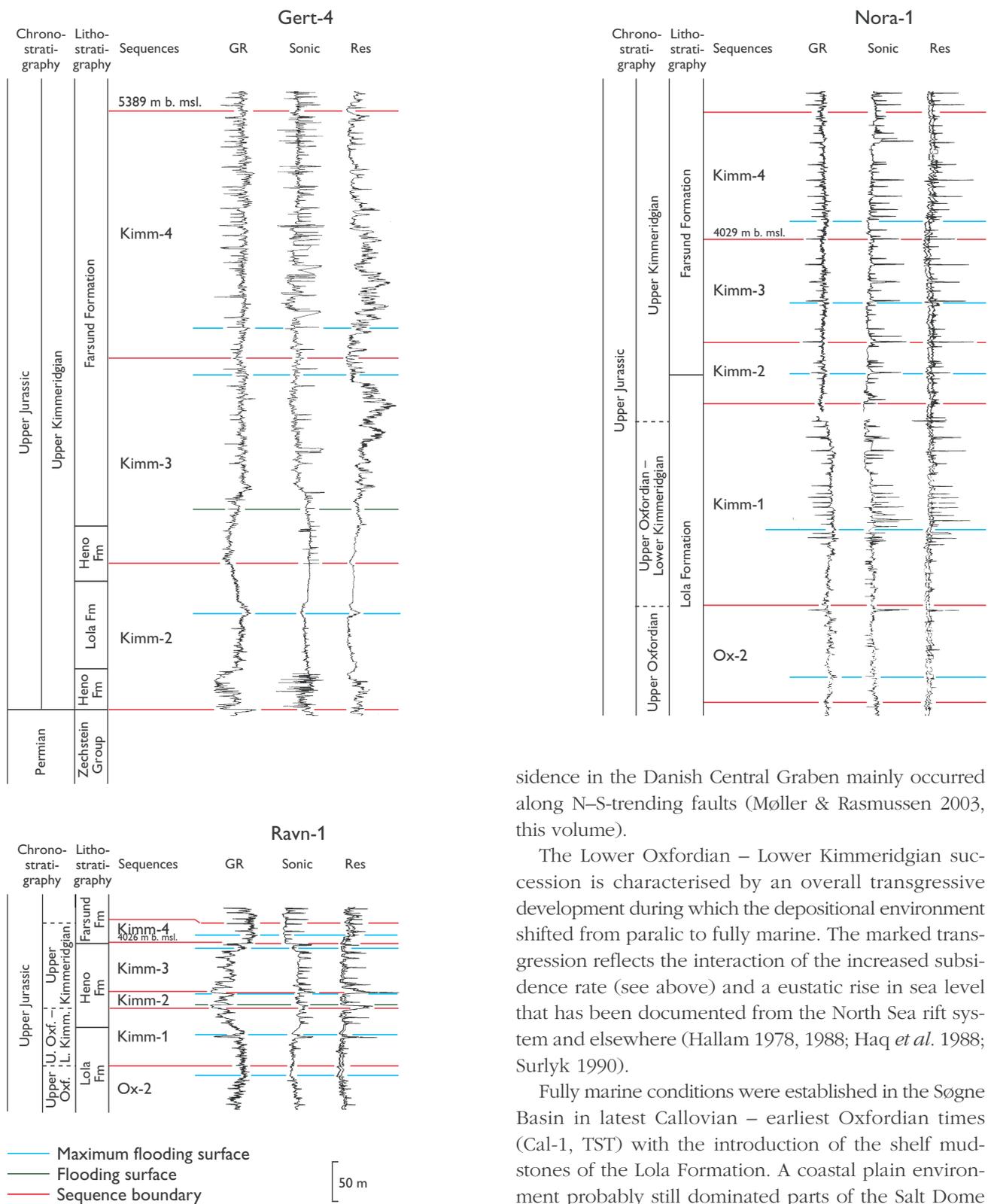


Fig. 12. Upper Jurassic stratigraphy of the Gert-4, Nora-1 and Ravn-1 wells illustrating, in particular, the characteristics of the Kimm-2 and Kimm-3 sequences. **Kimm.**, Kimmeridgian; **L.**, Lower; **Oxf.**, Oxfordian; **U.**, Upper.

sidence in the Danish Central Graben mainly occurred along N–S-trending faults (Møller & Rasmussen 2003, this volume).

The Lower Oxfordian – Lower Kimmeridgian succession is characterised by an overall transgressive development during which the depositional environment shifted from paralic to fully marine. The marked transgression reflects the interaction of the increased subsidence rate (see above) and a eustatic rise in sea level that has been documented from the North Sea rift system and elsewhere (Hallam 1978, 1988; Haq *et al.* 1988; Surlyk 1990).

Fully marine conditions were established in the Søgne Basin in latest Callovian – earliest Oxfordian times (Cal-1, TST) with the introduction of the shelf mudstones of the Lola Formation. A coastal plain environment probably still dominated parts of the Salt Dome Province to the south, indicating that the Callovian–Oxfordian transgression entered the Danish Central Graben from the north (Figs 9B, C).

During the Early Oxfordian, the Tail End Graben, the south-eastern marginal parts of the Heno Plateau,

the Salt Dome Province, and the Rosa Basin were also transgressed (Ox-1, Ox-2; Figs 8B, C, 9C, 13). The remainder of the Heno Plateau and possibly areas further west were slowly submerged during the Late Oxfordian – Early Kimmeridgian (Kimm-1; Figs 11A, 14; Fig. 14 follows page 294). Marginal marine sandstones were deposited locally on the southern part of the Heno Plateau during this transgression (e.g. Elly-2; Fig. 14) but marine mudstones of the Lola Formation

were deposited over most of the basin. A number of minor transgressive–regressive cycles can be discerned within the overall transgressive succession of marine mudstones (e.g. Nora-1; Fig. 13). The main basinal part of the study area was thus characterised by an offshore environment during this period. An inner shelf to shoreface environment developed, however, on the Gertrud Plateau and the northern part of the Heno Plateau at the end of this depositional phase (Fig. 15B).

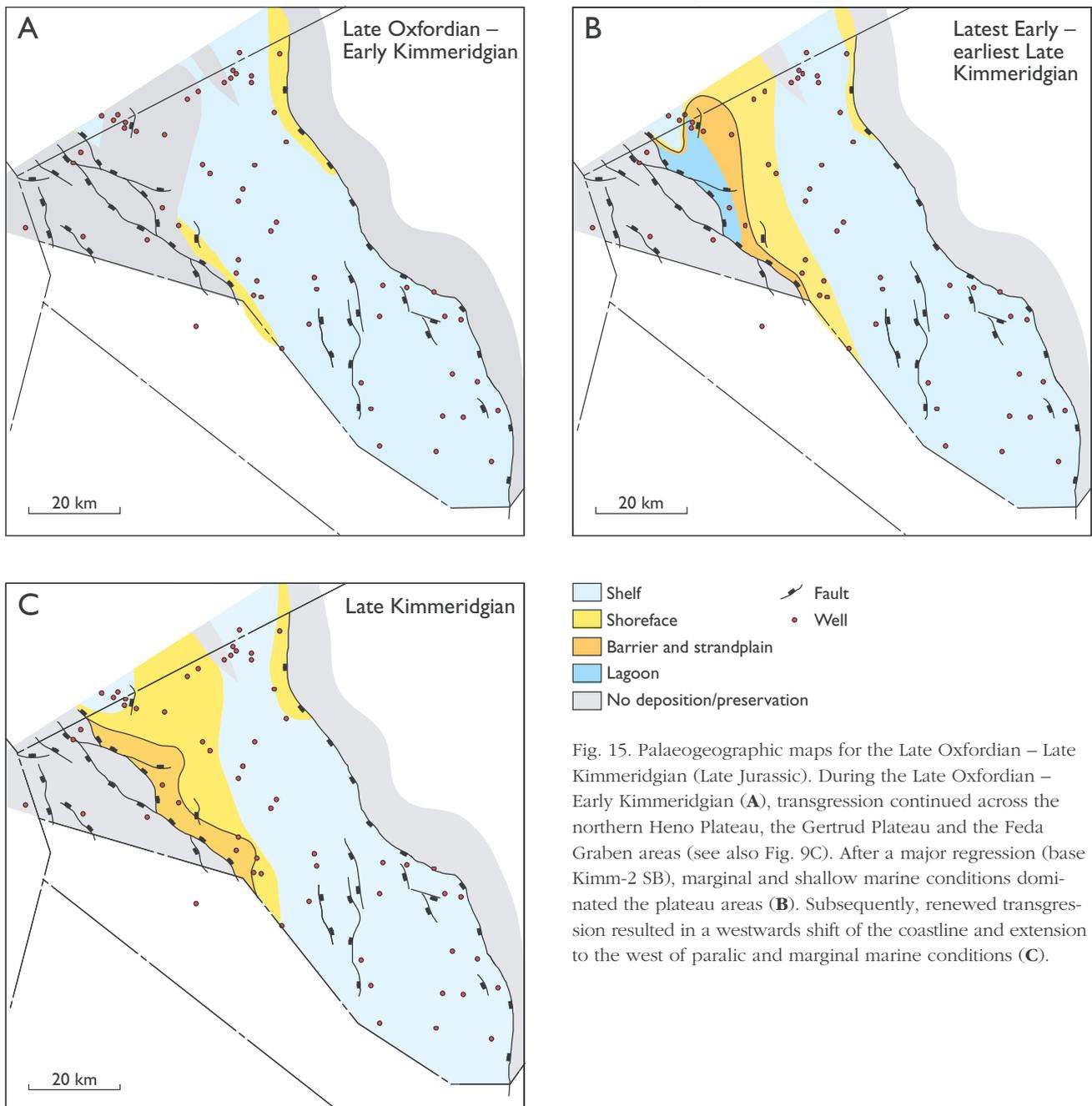


Fig. 15. Palaeogeographic maps for the Late Oxfordian – Late Kimmeridgian (Late Jurassic). During the Late Oxfordian – Early Kimmeridgian (A), transgression continued across the northern Heno Plateau, the Gertrud Plateau and the Feda Graben areas (see also Fig. 9C). After a major regression (base Kimm-2 SB), marginal and shallow marine conditions dominated the plateau areas (B). Subsequently, renewed transgression resulted in a westwards shift of the coastline and extension to the west of paralic and marginal marine conditions (C).

Phase 5. Shallow marine deposits and changing structural patterns (Late Kimmeridgian)

Sequence definitions

Kimm-2 sequence

This sequence is found throughout the study area east of the Mads and Inge Highs (Fig. 11B). The lower boundary has been placed at a conspicuous surface that, even in the deeper parts of the basin, shows an abrupt shift from high to low gamma-ray values, probably indicating an erosion surface (Figs 10, 12). In parts of the study area, such as the Feda Graben, the northern part of the Gertrud Plateau and the northern part of the Heno Plateau, this surface coincides with the unconformity that separates the pre-Jurassic and the Upper Jurassic deposits (e.g. Gert-4; Fig. 12).

On the Heno Plateau, the LST is thought to be represented by a unit of conglomerates and coarse-grained sandstones less than 5 m thick (e.g. Ravn-1; Fig. 12) although it should be noted that Johannessen (2003, this volume) refers this interval to the lowermost TST. In more distal settings, the LST is represented by a 10–15 m thick sandstone or sandy siltstone unit (e.g. Elly-2; Fig. 10). In the proximal settings of the Heno and Gertrud Plateaus and the southern Feda Graben, the TST consists of paralic and shallow marine sandstones and mudstones. In basinal settings (e.g. Nora-1), the TST is represented by a well-defined fining-upwards interval (Fig. 12). This interval consists of silty claystone, siltstone or silty sandstone, becoming coarser grained towards the Heno Plateau. In the central part of the Heno Plateau (e.g. Ravn-1), most of the HST was removed by erosion during the subsequent lowstand. On the northern Heno Plateau, the Gertrud Plateau and in the southern Feda Graben, the HST consists of a section that coarsens upwards from sandy siltstones to gravelly sandstones of shoreface origin (Johannessen *et al.* 1996; Johannessen 2003, this volume). In basinal settings, the HST consists of a coarsening-upwards interval of marine mudstones, siltstones and sandstones. The thickness of the sequence attains a maximum of 319 m in Cleo-1. Major depocentres for this sequence are located in the southern Tail End Graben – eastern Salt Dome Province, in the northern Tail End Graben – Søgne Basin and in the Feda Graben (Fig. 11B).

A latest Early Kimmeridgian to earliest Late Kimmeridgian age is indicated by the LOD of *Endoscrinium galeritum* in the lower part of the sequence in Edna-1 and West Lulu-1 and the FAD of *Subtilisphaera? paeminosa* and *S.? inaffecta* in the lower part of the sequence in Gert-1 combined with the LOD in the sequence above of *Stephanelytron scarburghense* in the Gert-2, Jeppe-1 and Lulu-1 wells (Fig. 3C).

Kimm-3 sequence

The distribution of this sequence corresponds to that of the Kimm-2 sequence (Fig. 11C). In the central part of the Heno

Plateau (e.g. Ravn-1; Fig. 12), the lower sequence boundary is represented by an erosion surface overlain by shallow marine conglomerates or gravelly sandstones. Further north, on the Heno Plateau and in the Gertrud Plateau – Feda Graben area, the sequence boundary is marked by a thin conglomerate in several wells (Johannessen 2003, this volume). In more basinal areas, south and east of the Heno Plateau, the lower sequence boundary is located at the top of a well-defined coarsening-upwards interval (e.g. Nora-1; Fig. 12). In the Heno Plateau – northernmost Salt Dome Province area (e.g. Ravn-1), the TST is developed as a back-stepping set of parasequences that consist of marine silty sandstones, siltstones and claystones. In the Feda Graben – Gertrud Plateau area, the TST consists of shallow marine sandstones and siltstones that grade upwards to marine mudstones (e.g. Gert-4; Fig. 12). In some wells (e.g. Jeppe-1; Fig. 14), an abrupt change from marine sandstones to mudstones, interpreted as a flooding surface, replaces part of the coarsening-upwards section. In the Tail End Graben (e.g. Nora-1), the TST is represented by a thin fining-upwards interval of marine mudstones. The HST typically consists of marine mudstones. In the Feda Graben – Gertrud Plateau area and in the Søgne Basin, the HST terminates in a distinct coarsening-upwards interval, which includes silt- and sandstones (e.g. Gert-4; Fig. 12). The thickness of the sequence attains a maximum of 387 m in Cleo-1. Major depocentres are located in the Søgne Basin, the Feda Graben and the southern Tail End Graben – Salt Dome Province (Fig. 11C).

The LOD of *Stephanelytron scarburghense* in the lower to middle part of this sequence in Gert-2 and Jeppe-1 and the LOD of *Endoscrinium luridum* in the succeeding sequence (Kimm-4) indicate a Late Kimmeridgian age for this sequence (Fig. 3C).

Basin evolution

The basin configuration changed significantly during the late Early Kimmeridgian – Late Kimmeridgian. Fault activity and fault-related subsidence waned and at the end of the period a new pattern of dominantly NNW–SSE-trending faults was established (Møller & Rasmussen 2003, this volume). This development marked a pause between two rift pulses. At this time, the Tail End Graben, including the Søgne Basin and most of the Heno Plateau, formed one major half-graben with an eastwards-dipping hanging-wall slope. Accommodation space had been created in the Feda Graben prior to this phase. The Feda Graben was separated from the Tail End Graben by a transfer zone (Fig. 11B).

The lack of rift-related subsidence, probably combined with a regional fall in sea level in the beginning of this phase, caused a significant relative sea-level fall. This sea-level fall caused the development of a distinctive

erosional sequence boundary traceable over most of the Danish Central Graben (Figs 13,14); it can be seen on seismic sections as an onlap surface (Erik S. Rasmussen, personal communication 1997).

The succeeding transgression, caused by eustatic or regional relative sea-level rise, gave rise to a gradual flooding of the Heno Plateau, the southern part of the Feda Graben and the Gertrud Plateau area. Two higher-order sea-level cycles can be recognised during this overall transgression; the Kimm-2 sequence records the first cycle, the Kimm-3 sequence the second.

While marginal areas were characterised by deposition of sand in shallow marine to paralic environments, low-energy marine conditions prevailed in the deeper parts of the basin during the Kimm-2 cycle (Fig. 15B). Sand was deposited in a high-energy shoreface environment on the southern part of the Heno Plateau and in a back-barrier environment on the northern part of the Heno Plateau and in the area of the Gert Ridge (Johannessen *et al.* 1996; Johannessen 2003, this volume). The flooding of the Heno Plateau area forced the coastline back from its position near the eastern margin of the plateau at the beginning of the transgression to the western margin of the Heno Plateau or possibly further west at the time of maximum flooding. Subsequently, the shallow areas in the western part of the basin were separated from the marine basin to the east by a zone of sand-dominated prograding shoreface. The coastline was re-established in a position at the eastern edge of the Heno Plateau at the end of this cycle (Fig. 15C).

In latest Early Kimmeridgian time (lower Kimm-2 sequence), the relatively limited accommodation space in the central parts of the Heno Plateau was rapidly filled with deposits of the LST and TST, before maximum flooding was attained (e.g. Ravn-1, Falk-1; Fig. 14). A relatively thick TST was deposited in the Feda Graben due to a rapidly subsiding basin floor. After the time of maximum flooding, most remaining available accommodation space occurred in the Tail End Graben and the deeper parts of the Feda Graben where a thick HST was deposited (e.g. G-1, Gert-4; Fig. 14). By the end of the first sea-level cycle (Kimm-2), an area consisting of the southern Feda Graben and the Heno and Gertrud Plateaus was exposed or dominated by shallow-water conditions. Deep-water conditions only prevailed in the central parts of the Tail End Graben and the Søgne Basin.

The Kimm-3 sequence is characterised by a major depocentre in the Feda Graben where thick marine deposits are related to the early onset of the next phase of subsidence. A central depocentre developed through-

out the length of the Tail End Graben and the Salt Dome Province, striking roughly NW–SE (Fig. 11C).

Coarse-grained conglomeratic shoreface deposits were deposited immediately above the sequence boundary (the base Kimm-3 SB) on the Heno Plateau (Johannessen 2003, this volume). These deposits were overlain by backstepping parasequences of shallow marine sand and silt (e.g. Falk-1, Ravn-1; Figs 12, 13).

The beginning of the next phase of subsidence caused a gradual change from coarse- to fine-grained shallow marine deposits in basinal settings of the Feda Graben. An abrupt shift from shallow marine sandstone to offshore mudstone can be seen in marginal areas of the Feda Graben and on the Gertrud Plateau (e.g. Gert-2, Gwen-2, Jeppe-1; Fig. 14). When fully marine conditions were established in the Feda Graben, a very high rate of subsidence in this area impeded the re-establishment of paralic conditions during the highstand part of the cycle. In the Tail End Graben and Søgne Basin, the Kimm-3 sequence consists of a thin unit of marine mudstones.

At the beginning of Kimm-3 time, accommodation space on the Heno and Gertrud Plateaus was limited by an initial sea-level fall and a low rate of subsidence. The limited accommodation space available was mainly filled by deposits of the TST (Fig. 14). In the remainder of the study area, there was sufficient accommodation space for the development of the HST (e.g. Cleo-1, G-1; Figs 16, 17, following page 294). On the southern part of the Gertrud Plateau and adjoining parts of the Heno Plateau, the remaining accommodation space was filled with shallow marine HST deposits followed by exposure and bypass (e.g. Gwen-2; Fig. 14).

Phase 6. Deep-water mudstones in a composite graben: rift pulses and passive subsidence (latest Late Kimmeridgian – middle Middle Volgian)

Sequence definitions

Kimm-4 sequence

This sequence has the same areal distribution as the Kimm-2 and Kimm-3 sequences (Fig. 18A). On the Heno Plateau, the lower sequence boundary is typically placed where the lithology changes abruptly from the sandstones and siltstones of the underlying sequences to fully marine mudstones (e.g. Ravn-1; Fig. 12). In basinal areas, the lower sequence boundary is located at the top of a coarsening-upwards interval, which in the Feda Graben (e.g. Gert-4) and in the Søgne Basin (e.g. Lulu-1; Fig. 17) may be distinct. In the Feda Graben

– Gertrud Plateau area, the TST is well-developed (e.g. Gert-2, Gwen-2; Fig. 19), whereas over much of the Heno Plateau, the Salt Dome Province and in the Tail End Graben, it often occurs in a condensed form (e.g. Nora-1; Fig. 12). In general, the HST is much thicker than the TST (e.g. Gert-2, Nora-1; Figs 12, 19); both the TST and HST consist of marine mudstones. The thickness of the sequence attains a maximum of 318 m in Gert-4. The sequence has an elongate depocentre in the eastern part of the Tail End Graben and a secondary depocentre in the Feda Graben (Fig. 18A).

A Late Kimmeridgian age is indicated for the sequence based on the LOD of *Stephanelytron scarburghense* in the sequence below com-

bined with the LOD of *Endoscrinium luridum* close to the upper sequence boundary in Amalie-1 and Cleo-1 (Fig. 3C).

Volg-1 sequence

This sequence has the same areal distribution as the previous sequences (Fig. 18B). In the Feda Graben – Gertrud Plateau area and on the Heno Plateau, the lower sequence boundary is located at the top of a thin but conspicuous coarsening-upwards interval (Fig. 19). In the more basal settings of the Tail End Graben, the Søgne Basin, and the Salt Dome Province, the boundary is situated at the turnaround point between rather indistinct coarsening-upwards and fining-

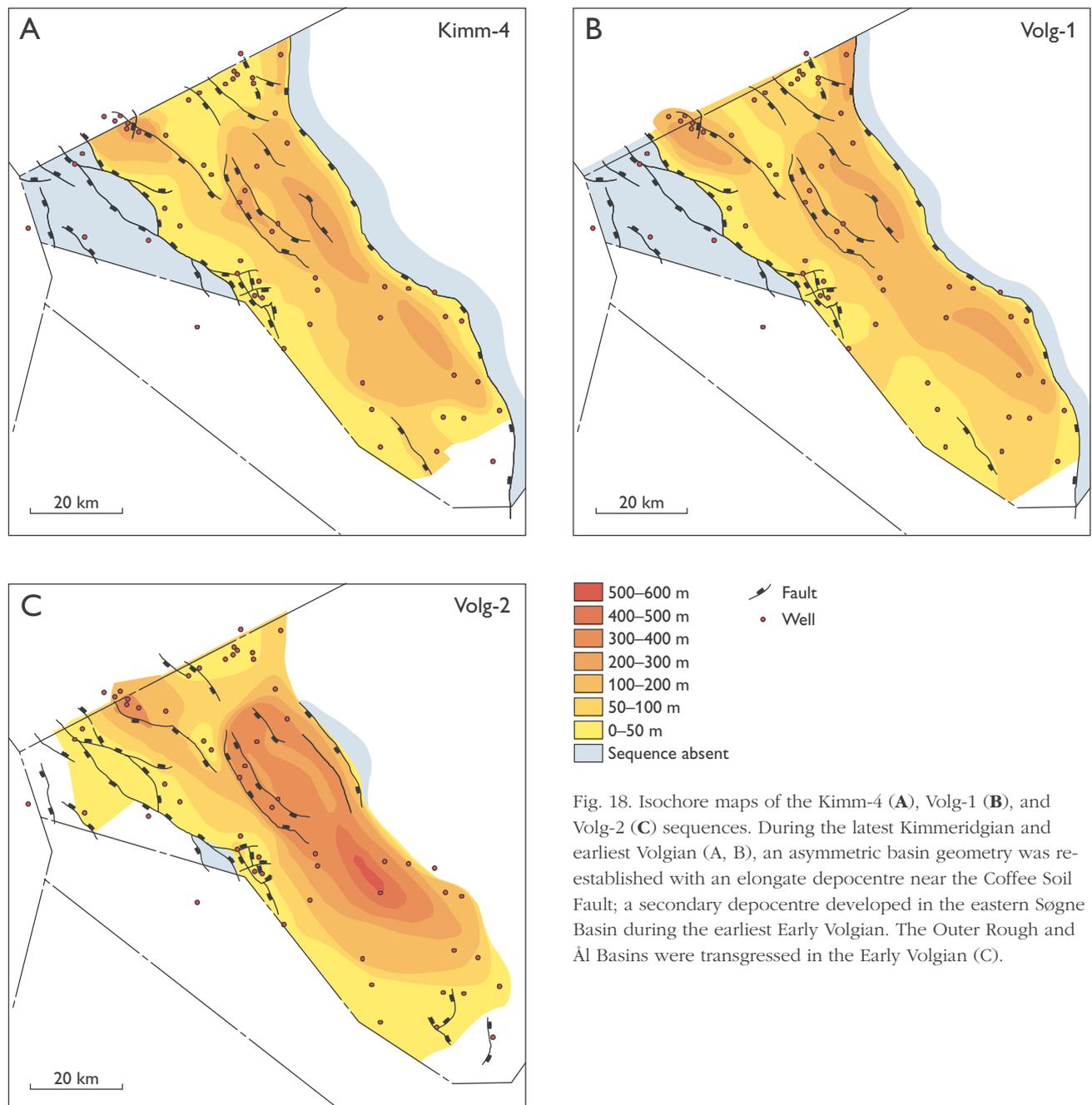
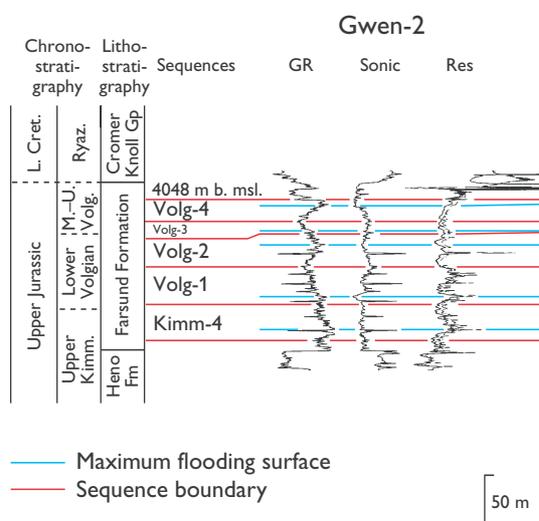
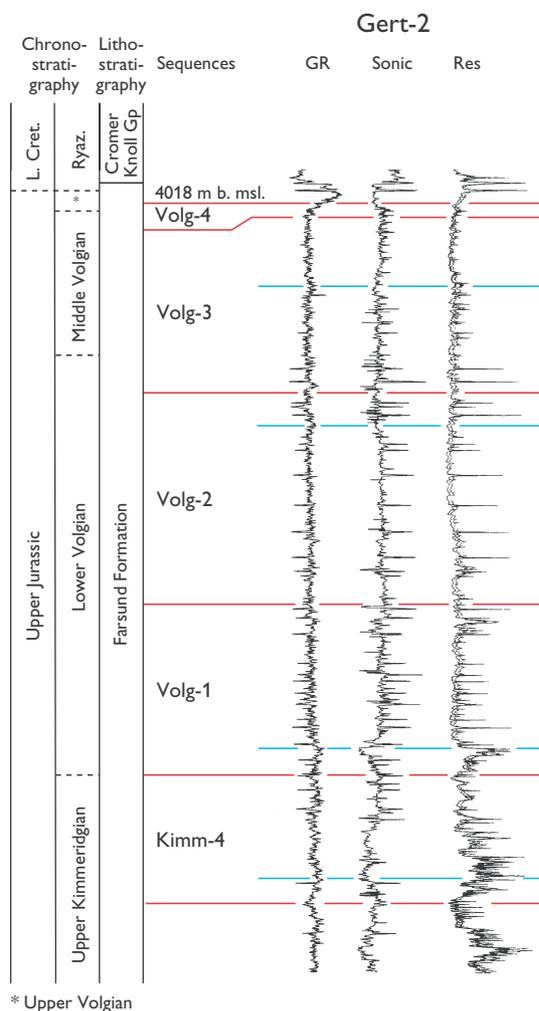
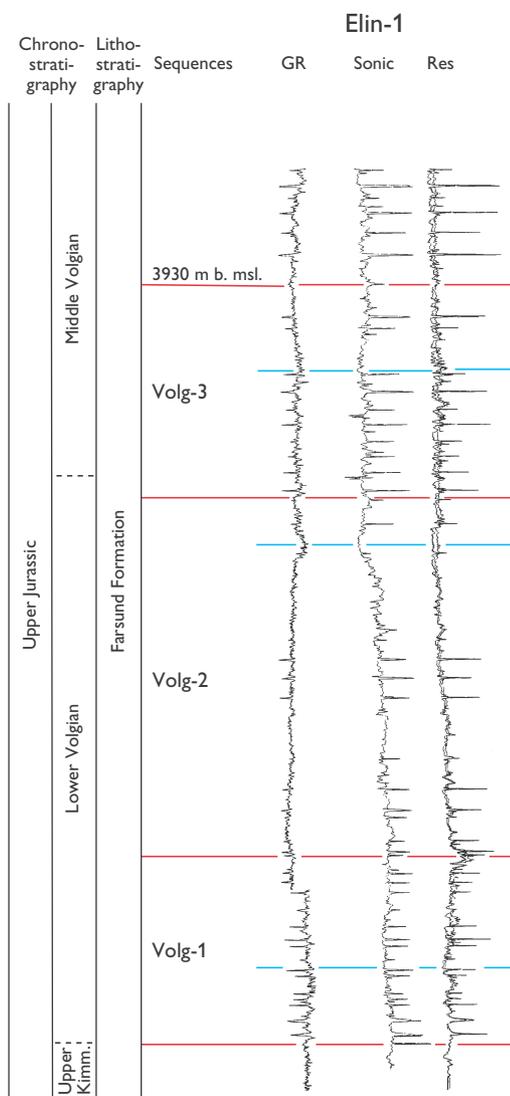


Fig. 18. Isochore maps of the Kimm-4 (A), Volg-1 (B), and Volg-2 (C) sequences. During the latest Kimmeridgian and earliest Volgian (A, B), an asymmetric basin geometry was re-established with an elongate depocentre near the Coffee Soil Fault; a secondary depocentre developed in the eastern Søgne Basin during the earliest Early Volgian. The Outer Rough and Ål Basins were transgressed in the Early Volgian (C).



upwards units (e.g. G-1, Nora-1; Fig. 16). In several wells in the Salt Dome Province, the TST is missing and the MFS amalgamates with the underlying sequence boundary (e.g. Deep Gorm-1; Fig. 14). The MFS is marked by a conspicuous peak on the gamma-ray, sonic and resistivity logs. Both the TST and HST consist of marine mudstones. The thickness of the sequence attains a maximum of 210 m in Gert-4. The main depocentre of the sequence is in the eastern part of the Tail End Graben and in the eastern Søgne Basin. A secondary depocentre is present in the Feda Graben (Fig. 18B).

The sequence is referred to the earliest Early Volgian (Fig. 3C) based on the LOD of *Endoscrinium luridum* in the sequence below combined with the LOD of *Subtilisphaera? paeminosa* in the upper part of the sequence in a number of wells (e.g. Amalie-1, Deep Gorm-1, Gert-2, -4, Gwen-2, Ravn-2, U-1).

Fig. 19. Upper Kimmeridgian – Volgian sequence stratigraphy of the Elin-1, Gert-2 and Gwen-2 wells, highlighting the features of the Kimm-4 – Volg-3 sequences. **Kimm.**, Kimmeridgian; **L. Cret.**, Lower Cretaceous; **M.–U. Volg.**, Middle–Upper Volgian; **Ryaz.**, Ryazanian.

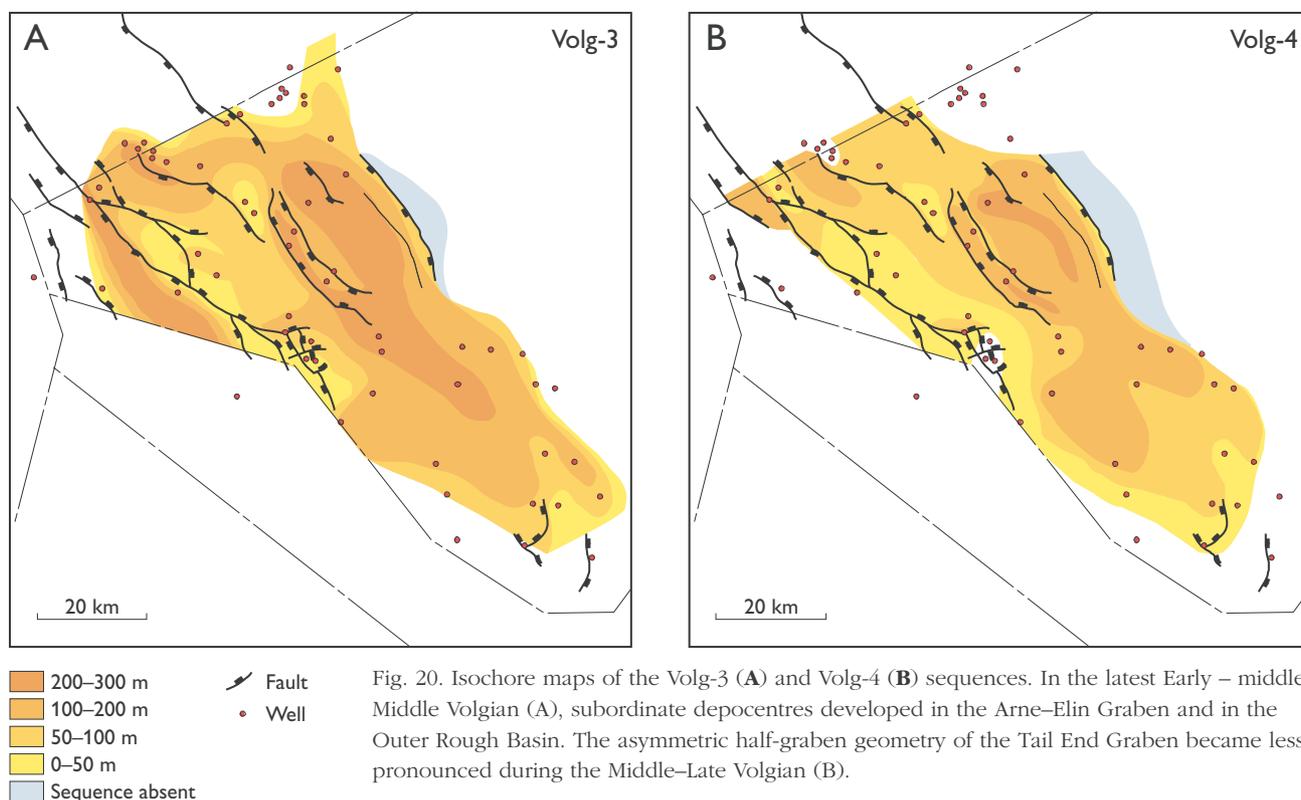


Fig. 20. Isochore maps of the Volg-3 (A) and Volg-4 (B) sequences. In the latest Early – middle Middle Volgian (A), subordinate depocentres developed in the Arne–Elin Graben and in the Outer Rough Basin. The asymmetric half-graben geometry of the Tail End Graben became less pronounced during the Middle–Late Volgian (B).

Volg-2 sequence

Although partly or completely eroded in most wells in the Salt Dome Province, this sequence is recognised throughout the remainder of the Danish Central Graben, including the Outer Rough and Ål Basins (Fig. 18C). The lower sequence boundary of this sequence is rather indistinct in most wells, being located at the top of a weak coarsening-upwards trend within the HST of the Volg-1 sequence below (Fig. 19). In basinal settings, where the sequence is relatively thick, the TST is four to five times thicker than the HST; in settings where the sequence is relatively thin, the TST and HST are of comparable thickness. The sequence consists of marine mudstones. The maximum thickness of the sequence measured in wells is 344 m (E-1) although seismic data indicate thicknesses in excess of 400 m in the prominent Tail End Graben depocentre. A secondary depocentre is present in the Feda Graben (Fig. 18C).

The LOD of *Subtilisphaera? paeminosa* in the sequence below (Volg-1) and of *Oligosphaeridium patulum* in the sequence above (Volg-3) indicate an Early Volgian age for this sequence. This is further supported by the LOD of *Cribroperidinium? longicorne* in the lower part of the sequence in the Eg-1 and Emma-1 wells (Fig. 3C).

Volg-3 sequence

The sequence is present over much of the study area but is missing locally in the south-western part of the Salt Dome Province and in the area around the Mandal High, probably due to erosion (Fig. 20A). The lower sequence boundary is positioned at the top of a coars-

ening-upwards interval (e.g. Elin-1, Gert-2; Fig. 19). In a number of wells where the section is relatively thin, the sequence boundary is placed at the base of a somewhat coarser bed (e.g. Gwen-2; Fig. 19). The Volg-3 sequence consists of marine mudstones. The thickness of the sequence attains a maximum of 259 m in E-1. The primary depocentre in the Tail End Graben branches into the Arne–Elin Graben. Secondary depocentres are present in the Gertrud and Feda Grabens and in the Outer Rough Basin (Fig. 20A).

The LOD of *Oligosphaeridium patulum* in the lower part of this sequence in a number of wells (e.g. Deep Gorm-1, Elly-2, Falk-1, Gert-2, Gert-4, I-1, M-8, U-1, V-1) combined with the LOD of *Occisucysta balia* in the middle to upper part of the sequence in the Bo-1, Gert-2, Gwen-2 and Ravn-2 wells indicate a latest Early Volgian – middle Middle Volgian age for this sequence. This is supported by the LOD of *Perisseiasphaeridium pannosum* in the upper part of the sequence in Gert-1 and Gert-2 (Fig. 3C).

Basin evolution

In this depositional phase, the occurrence of rift pulses is reflected by syn-rift successions with fining-upwards log patterns separated by successions with overall coarsening-upwards patterns, as described from the Mesozoic of East Greenland by Surlyk (1978). A rift pulse in Late Kimmeridgian time is indicated by the asymmetric sed-

iment distribution of the Kimm-4 deposits in the Tail End Graben (Fig. 18A). Another pulse in the earliest Early Volgian is reflected by fining-upwards log patterns in the lower part of the Volg-1 sequence (e.g. G-1, Gert-4; Fig. 14) and by seismic data from the Tail End Graben (Møller & Rasmussen 2003, this volume). A further rift pulse in the Early Volgian is interpreted from the abrupt change from low to high gamma-ray values immediately below the Volg-2 MFS (e.g. Gert-4, Nora-1; Figs 14, 16). The latter two rift pulses are recorded on seismic data from the Tail End Graben where they resulted in a rota-

tional tilt package (Møller & Rasmussen 2003, this volume, fig. 7).

The Tail End Graben and the Heno Plateau acted initially as one major asymmetric basin, with an elongate, NW–SE-oriented depocentre located in the Tail End Graben. During the Early Volgian, this depocentre expanded into the eastern part of the Søgne Basin, where subsidence increased significantly (Fig. 18A, B). The Feda Graben continued as an important depocentre. The Gertrud Graben and the Outer Rough and Ål Basins became actively subsiding elements at the time

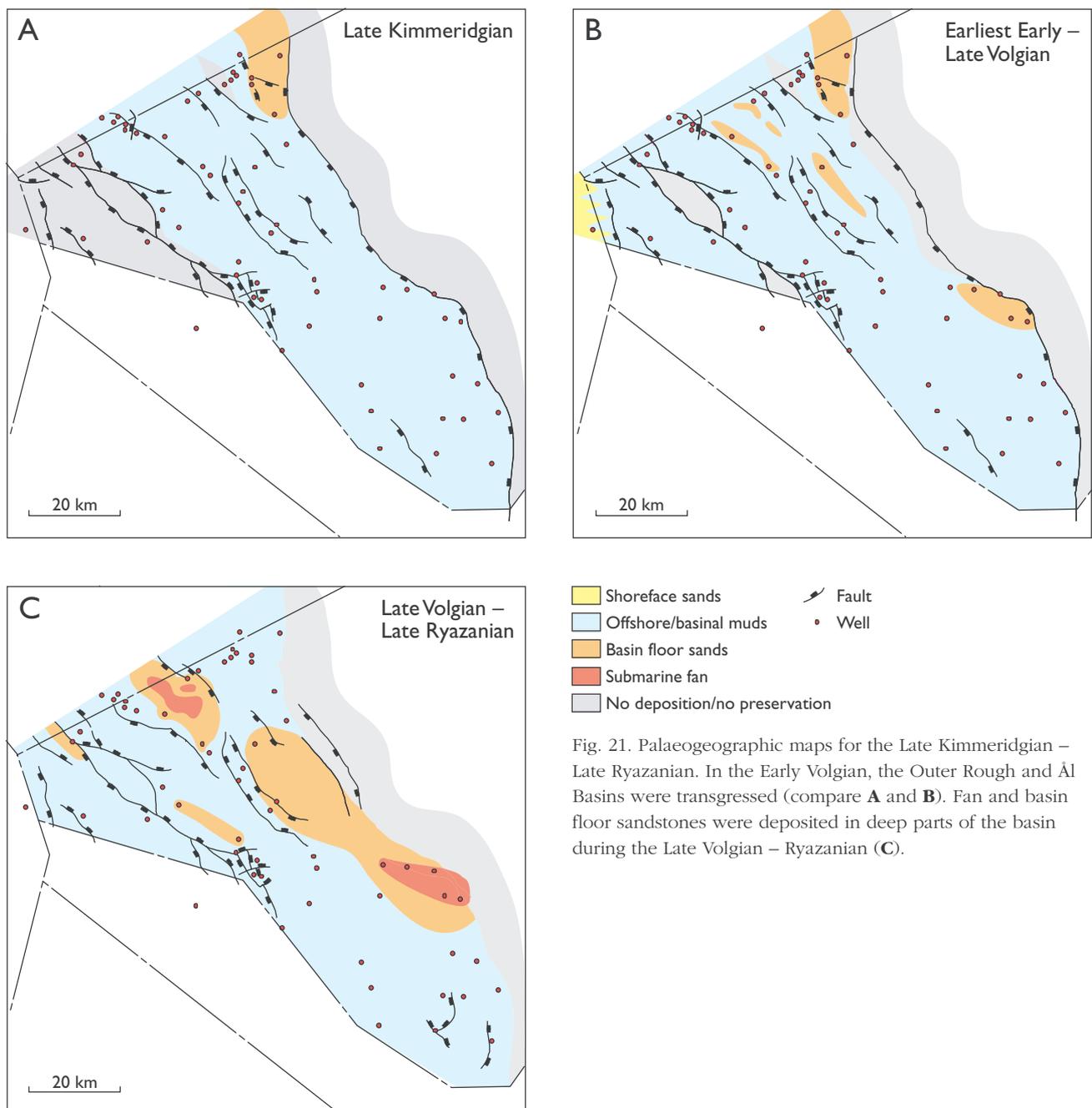


Fig. 21. Palaeogeographic maps for the Late Kimmeridgian – Late Ryazanian. In the Early Volgian, the Outer Rough and Ål Basins were transgressed (compare **A** and **B**). Fan and basin floor sandstones were deposited in deep parts of the basin during the Late Volgian – Ryazanian (**C**).

of deposition of the Volg-2 sequence (Figs 17, 18C). Later during this phase, increased fault activity resulted in the generation of several minor sub-basins, for example the Arne–Elin Graben which became a distinct depocentre in the early Middle Volgian during deposition of the Volg-3 sequence (Figs 18C, 20A).

Plateau areas draped by Heno Formation sandstones were drowned and a coarsening-upwards succession of marine siltstones and mudstones of the Farsund Formation was deposited above the Kimm-4 MFS (e.g. Ravn-1; Fig. 14). Thin storm- or gravity-derived sand layers were occasionally deposited in the Søgne Basin (Figs 17, 21A).

During the Early Volgian, the Outer Rough and Ål Basins were also transgressed, and fully marine conditions were established (Figs 17, 21B). Shoreface sands were deposited at the western margin of the Outer Rough Basin, as documented by data from the UK sector of the Central Graben (Mackertich 1996; Spathopoulos *et al.* 2000).

Phase 7. Organic-rich mudstones and basin axis turbidites (late Middle Volgian – Early Ryazanian)

Sequence definitions

Volg-4 sequence

This sequence has not been found in the Søgne Basin and Mandal High area; it is also missing from parts of the Outer Rough Basin and the Salt Dome Province. The sequence is present in all other parts of the study area (Fig. 20B). The lower boundary of the Volg-4 sequence is positioned at the top of the well-defined coarsening-upwards HST of the Volg-3 sequence below (e.g. Elin-1; Fig. 22). In a number of wells, the sequence boundary is further marked by the abrupt base of a slightly more silty or sandy unit (e.g. Jeppe-1).

In many wells (e.g. Iris-1, Jeppe-1), the lower part of the TST typically shows a fining-upwards/coarsening-upwards pattern; in V-1, this unit is developed as a 30 m thick sandstone unit (Fig. 23). In these wells, the remainder of the TST consists of a short, distinct fining-upwards mudstone section. In other wells, the TST is represented by a simple fining-upwards interval of marine mudstones (e.g. Bo-1, Elin-1; Fig. 22). The upper boundary of this sequence is commonly an erosion surface and the upper part of the HST is thus missing in many wells (e.g. Gwen-2, Jeppe-1; Figs 19, 22). Where a major part of the HST is preserved, it typically shows consistently high gamma-ray values (e.g. Bo-1, Elin-1). Due to the increasingly organic-rich nature of the sediments (see below), the gamma-ray log is not considered a fully reliable grain-size indicator. Despite the high gamma-ray readings, descriptions of the cuttings indicate the presence of sand-

and siltstone in this interval. It is therefore assumed that the high gamma-ray readings are caused by the presence of organic-rich 'hot shales' and not necessarily by a lack of sand- and silt-sized material. The sequence consists of marine, organic-rich mudstone with silt- and sandstone interbeds. The thickness of the sequence attains a maximum of 189 m in Elin-1. The sequence shows a more even distribution than the previous sequences, but depocentres are still recognisable in the Tail End Graben, the Arne–Elin Graben, the Feda Graben and the Outer Rough Basin (Fig. 20B).

The LOD of *Scriniodinium inritibile* in the lower part of the sequence in Bo-1 and of *Senoniasphaera jurassica* in the lower to middle part of the sequence in a number of wells (e.g. Bo-1, Deep Gorm-1, Elin-1, Gwen-2, I-1, Jeppe-1, Ravn-2, W-1) indicate a middle Middle Volgian age for the lower part of the sequence. In combination with the LOD of *Egmontodinium polyplacophorum* in the middle to upper part of the sequence in the Bo-1, Deep Adda-1, E-1, I-1, Lone-1 and V-1 wells, this indicates a middle Middle Volgian to Late Volgian age for the sequence. This age assumption is supported by the LOD of one or more of the dinocyst species *Dichadogonyaulax pannea*, *Glossodinium dimorphum*, *Muderongia simplex* (the form previously referred as *Muderongia* sp. A; Poulsen & Riding 2003, this volume) and *Gochteodinia mutabilis* within the sequence in a number of wells (e.g. Bo-1, Deep Gorm-1, E-1, Elin-1, Gwen-2, I-1, Iris-1, V-1; Fig. 3C).

Ryaz-1 sequence

The distribution of the Ryaz-1 sequence corresponds to the distribution of the Volg-4 sequence (see above). The development of the 'hot shales' of the Bo Member within this sequence and the transition to more calcareous sediments at the top of the Farsund Formation makes sequence stratigraphic analysis problematic. Neither a MFS nor an upper boundary of the sequence have been picked. The sequence is not necessarily limited to the marine mudstones of the Farsund Formation but may straddle the boundary to the overlying Cromer Knoll Group. The lower sequence boundary is normally distinct and often erosional (e.g. Bo-1, Elin-1, Jeppe-1; Fig. 22). Turbidite sandstones occur above the boundary in several wells (e.g. Iris-1, Jeppe-1; Fig. 23). In certain wells that lack such turbidite sandstones (e.g. E-1, Gert-2), the bioevents of the turbidite-bearing section are absent; this hiatus is interpreted to have resulted from erosion or sediment bypass, updip from areas in which turbidite sands accumulated (Ineson *et al.* 2003, this volume).

The turbidite interval may represent lowstand and possibly early transgressive deposits. In areas without active turbidite transport or deposition, this interval is characterised by a gradual upwards increase in gamma-ray values, probably representing the background sedimentation of the basin (e.g. Bo-1). This increase in the strength of the gamma-ray signal culminates in the highest gamma-ray values in the entire Jurassic studied section. This interval corresponds to the 'hot unit' of Jensen *et al.* (1986), formalised as the Bo Member (Farsund Formation) by Michelsen *et al.* (2003, this volume). The dominant lithology is

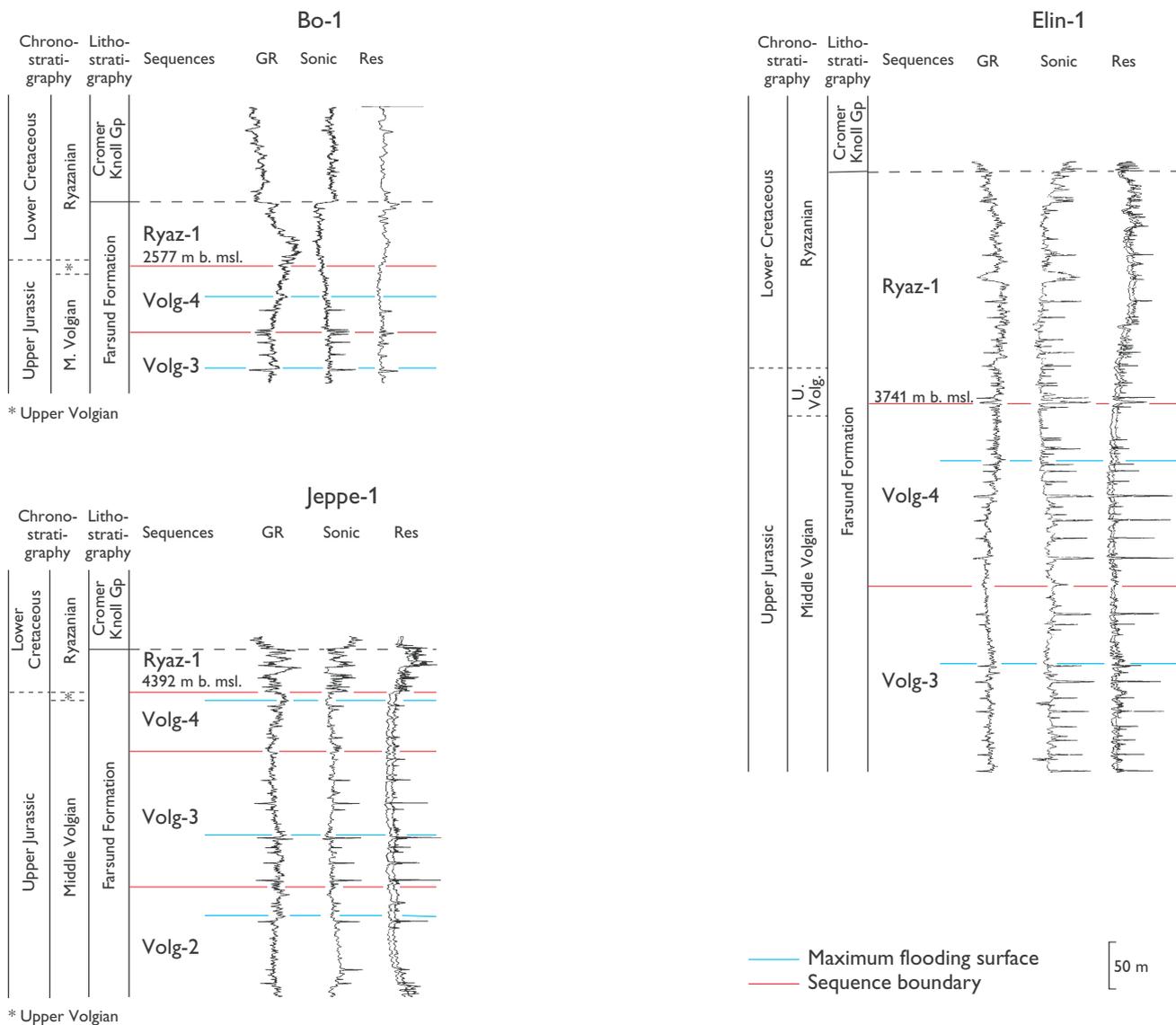


Fig. 22. Uppermost Jurassic to lowermost Cretaceous stratigraphy of the Bo-1, Elin-1 and Jeppe-1 wells illustrating the characteristic features of the Volg-4 – Ryaz-1 sequences. **M.**, Middle; **U. Volg.**, Upper Volgian.

organic-rich marine mudstone; the mudstones are locally interbedded with turbidite sandstones (Ineson *et al.* 2003, this volume). The thickness of this interval, from the basal sequence boundary to the top of the Farsund Formation, attains a maximum of 326 m in Iris-1.

The LOD of *Egmontodinium expiratum* in the lower part of this sequence in a number of wells (e.g. Bo-1, Edna-1, Gert-2) and of *Rotosphaerotopsis thula* immediately above the 'hottest' part of the Bo Member (e.g. in Anne-3, Bo-1, E-1, Edna-1, Gert-2, Jeppe-1) indicate a Late Volgian – Early Ryazanian age for this part of the sequence. The uppermost part of the Farsund Formation is of Late Ryazanian age. This is indicated by the LOD of *Dingodinium spinosum* which coincides broadly with the upper boundary of the Farsund Formation in a number of wells (e.g. Deep Gorm-1, E-1, Edna-1, Jens-1, Jeppe-1,

Kim-1, Lone-1, V-1; Ineson *et al.* 2003, this volume) and by the LOD of *Dichadogonyaulax culmula* at the same stratigraphic level in the E-1, Edna-1 and Gert-2 wells (Fig. 3C).

Basin evolution

The rate of subsidence decreased in the Tail End Graben and the Salt Dome Province during this phase; in combination with decreased sediment supply, this resulted in stratigraphic condensation. According to the time-scale of Gradstein *et al.* (1994), the succession deposited during phase 7 represents a period of *c.* 5.5 million years,

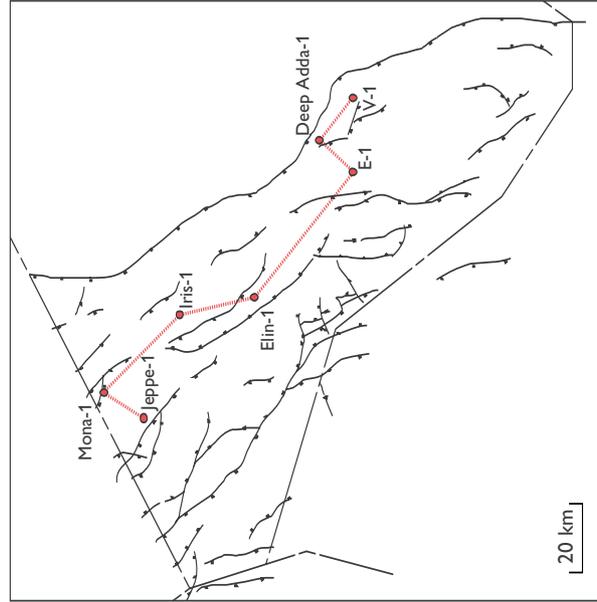
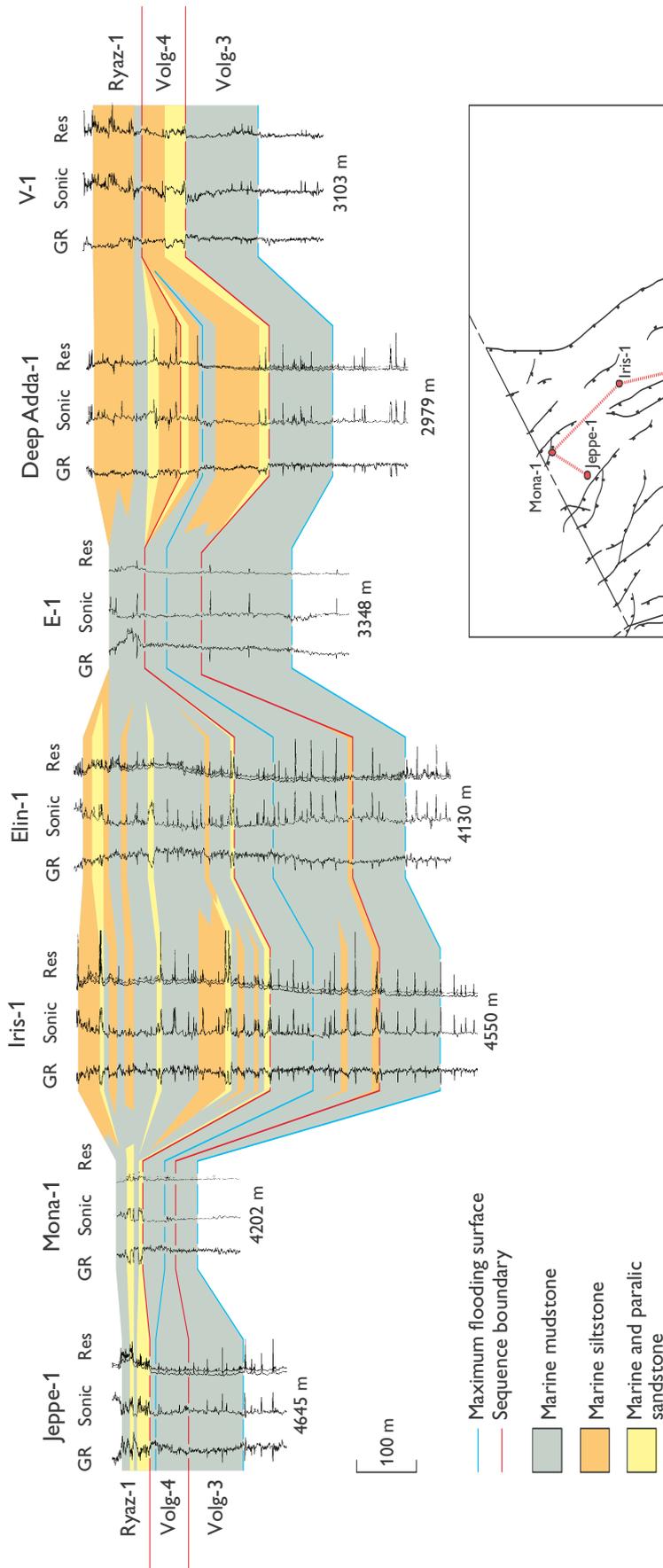


Fig. 23. Sequence stratigraphic correlation of the uppermost Upper Jurassic and lowermost Cretaceous in the Tail End Graben and the Gertrud Graben. Fan and basin floor sandstones are common in the Upper Volgian – Ryazanian succession.

approximately equal to the time represented by the 2–10 times thicker succession of phase 6.

Fault activity ceased along large segments of the main boundary fault (Møller & Rasmussen 2003, this volume), and this caused the geometry of the Tail End Graben to change from an asymmetric rift to a more symmetrical saucer-like basin with a relatively uniform sediment distribution (Fig. 20B). These changes are the first indications in the Tail End Graben of the beginning of an early post-rift stage. Minor depocentres continued to exist in the northern part of the Tail End Graben and in the Arne–Elin Graben. Subsidence continued in the Feda Graben and in the Outer Rough Basin.

The deposits of phase 7 consist of marine mudstones of the Farsund Formation. They generally show a gradual increase in gamma-ray and TOC values culminating in the ‘hot shales’ of the Bo Member in the uppermost part (Figs 17, 23; Jensen *et al.* 1986; Ineson *et al.* 2003, this volume). Below the Bo Member, a well-developed sequence boundary separates the Volg-4 and Ryaz-1 sequences. This sequence boundary is strongly erosional in most wells on the northern part of the Heno Plateau and along the southern margin of the Feda and Gertrud Grabens (e.g. Jeppe-1; Fig. 17). Above the sequence boundary, sandstones are present in a number of wells along the basin axes of the Tail End Graben and the Gertrud Graben (Figs 21C, 23). The sandstones were deposited from turbidity currents (Iris-1) and debris flows (Jeppe-1) and may have originated from a number of sources, such as the East North Sea Block of the Ringkøbing–Fyn High, the Mandal High and the Gert Ridge (Nielsen 1985; Ineson *et al.* 2003, this volume). Their occurrence may be related to changed drainage patterns in the hinterland after subsidence had ceased along major segments of the main boundary fault, and to erosion of inverted structures or crests of rotated fault blocks.

Relative sea-level changes

On the basis of the sequence stratigraphic framework presented here (Fig. 24), a relative sea-level curve can be constructed for the Middle–Late Jurassic time interval and compared with eustatic curves proposed by Haq *et al.* (1988) and Hallam (1988) based on North Sea and global data and with the relative sea-level curve of Surlyk (1990) for East Greenland (Fig. 25). The Lower Jurassic has only been penetrated by wells in a small part of the study area, and no attempt has been made to construct a sea-level curve for this part of the suc-

cession. Data are lacking from the Toarcian and most of the Aalenian due to the regional uppermost Aalenian or lowermost Bajocian unconformity. Good biostratigraphic datings become available from the uppermost Bathonian–Callovian and upwards; the sea-level curve thus has its starting point at this level. Wells from the basin centre and intermediate positions were preferred in selecting data for the construction of the curve.

The relative sea-level changes have been interpreted mainly from lithological changes (e.g. changes in mud content in marine sediments) reflecting bathymetrically related changes in energy level. Lithological changes were interpreted from well log patterns and cores. Sedimentary facies, palynofacies and the extent of marine flooding surfaces have also been used.

High uranium contents in some marine mudstones, related to organic matter content rather than clay proportion, can cause large deflections on the well logs, thus hampering lithological interpretation. For that reason, sea-level interpretations were not attempted in the ‘hot shales’ of the Bo Member.

The coastal plain deposits of latest Bathonian – earliest Callovian age were transgressed during the Callovian – earliest Oxfordian in most of the Danish Central Graben. A relative sea-level rise caused a rapid, step-wise transgression that expanded the basin laterally and drowned previous sediment sources. The sea-level rise and transgression continued throughout the Oxfordian and culminated in the Early Kimmeridgian. This development, from the latest Bathonian to the late Early Kimmeridgian, is almost identical to the sea-level curves of Haq *et al.* (1988) and Hallam (1978, 1981, 1988), based on North Sea and global data, and the curve of Surlyk (1990) for East Greenland. The gradual collapse of the North Sea Dome (Hallam & Sellwood 1976; Ziegler 1982; Underhill & Partington 1993) and the initiation of rifting in the Danish Central Graben may have influenced transgressive trends in the region during this period. However, the resemblance between the relative sea-level curve of this study and the curves of Haq *et al.* (1988), Hallam (1988) and Surlyk (1990) may indicate that a common causal factor of super-regional or global extent also influenced sea-level changes during this interval.

The uppermost Lower Kimmeridgian – Upper Kimmeridgian of the Danish Central Graben is characterised by a double lowstand peak. The other sea-level curves are close to their maximum at this level. The lowstand in the Danish Central Graben probably evolved as a result of a sea-level fall during a pause in rift-induced subsidence (phase 5).

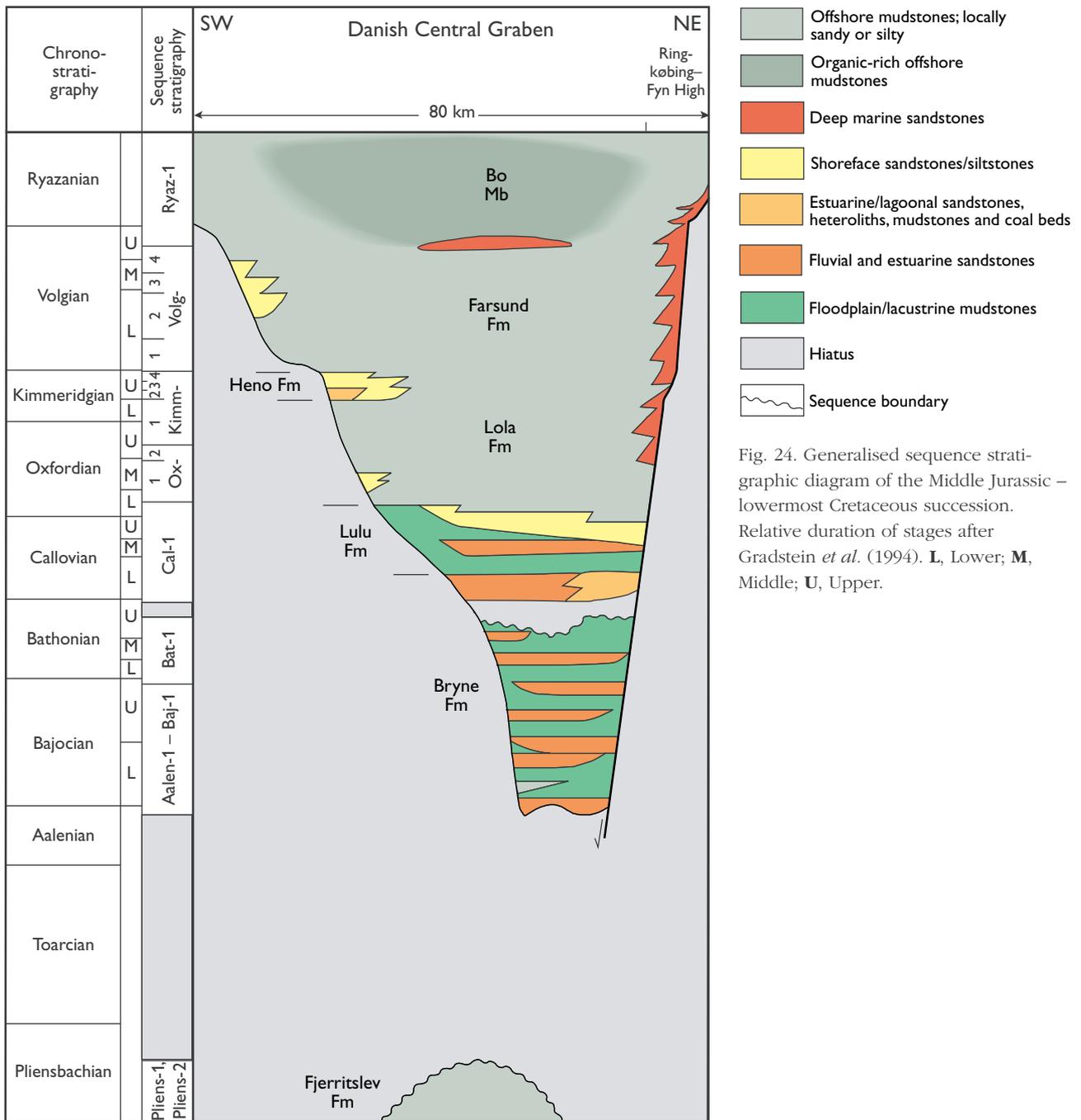


Fig. 24. Generalised sequence stratigraphic diagram of the Middle Jurassic – lowermost Cretaceous succession. Relative duration of stages after Gradstein *et al.* (1994). L, Lower; M, Middle; U, Upper.

After the lowstand event, the Late Kimmeridgian – late Middle Volgian interval shows an overall rise in relative sea level, interrupted by a number of minor falls. This overall trend differs from the fall seen in most of the other sea-level curves, probably due to the continued high subsidence rate in the Danish Central Graben, which neutralised the effect of the global sea-level fall indicated by the other curves. The minor relative sea-level falls in the interval may reflect pauses between periods of active subsidence.

Lithology prediction

An important reason for undertaking a sequence stratigraphic study of a hydrocarbon-producing basin is to improve lithology and reservoir prediction. The predictive potential in syn-rift units is much less than in post-rift units. This is due to the problems for orderly sediment dispersal caused by the creation of tilted fault blocks and sub-basins, by the continuous presence of accommodation space on the lower hanging-wall slopes of tilted fault blocks, and by the local supply of sand

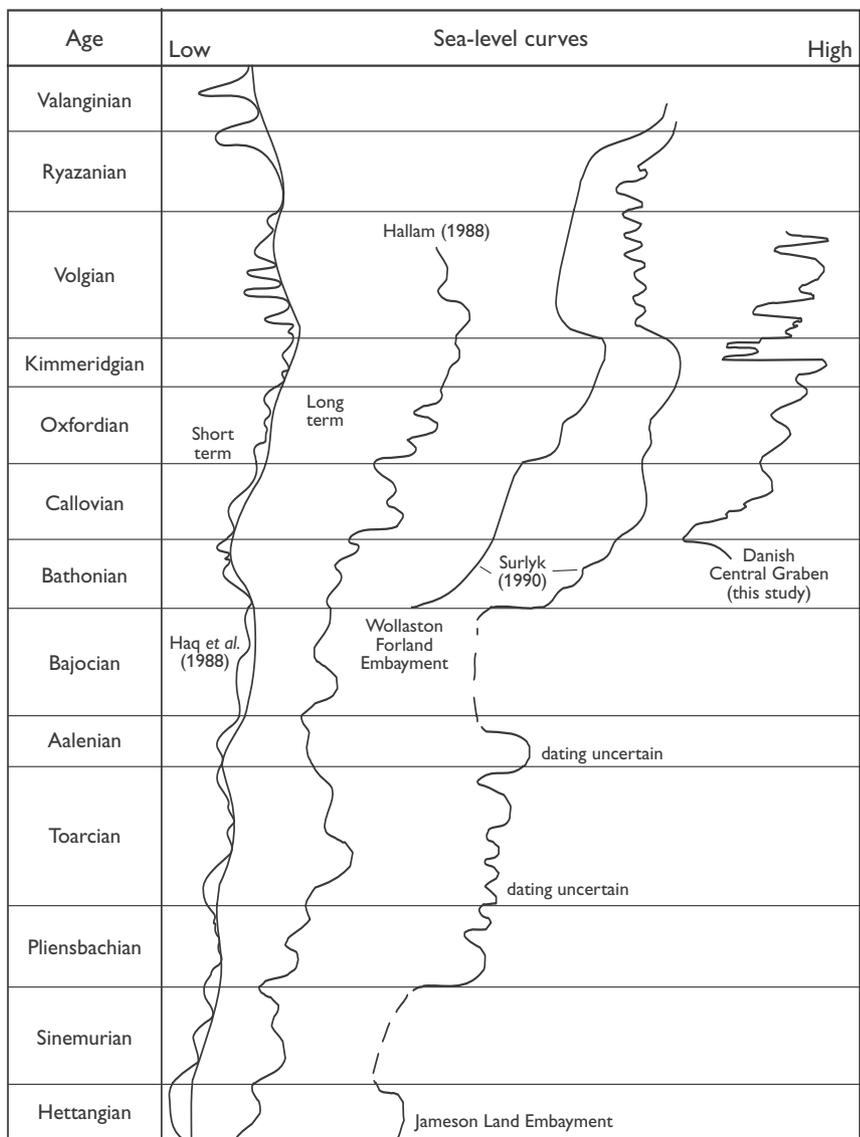


Fig. 25. Comparison of the relative sea-level curve of this study with the sea-level curves presented by Haq *et al.* (1988), Hallam (1988) and Surlyk (1990). Figure modified after Surlyk (1990), according to the Gradstein *et al.* (1994) time-scale.

from erosion of uplifted footwall shoulders. However, in the Middle to Upper Jurassic syn-rift deposits of the Danish Central Graben, the sandstone units seem to show a systematic distribution, which may be related to the combination of sea-level changes and periods of little tectonic subsidence.

The Bathonian–Callovian sandstones, that are widely distributed in the Søgne Basin, the Tail End Graben and the Salt Dome Province, resulted from a generous supply of sand from the active denudation of the North Sea dome area and a relatively low and only slowly increasing accommodation space in the rift initiation stage. The best reservoir sandstones occur in the lowstand and transgressive systems tracts of the Baj-1, Bat-1 and Cal-1 sequences (Figs 4, 7), in the form of laterally extensive fluvial channel sandstones and tidal and shoreface sandstones in the uppermost part of the sandstone unit,

deposited prior to the final transgression. The desirable combination of large fluvial and tidal channels and extensive wave-reworked sandstones occurs in the deeper parts of the Søgne Basin and the Tail End Graben. In these areas, subsidence was sufficiently fast to create accommodation space for successions of wave-influenced sediments up to 30 m thick during the final Middle Jurassic transgression (Andsbjerg 2003, this volume).

An overall transgressive development is interpreted for the Late Jurassic. The largest concentration of reservoir sandstones in this succession (the Heno Formation) was deposited during a pause in the otherwise rapid subsidence. The sand is concentrated on the Heno Plateau, the Gertrud Plateau and along the southern margin of the Feda Graben. The Heno Plateau constituted an upper hanging-wall slope of a major half-graben, the Tail End Graben, and the Gertrud Plateau probably

comprised a transfer zone between the Tail End and Feda Grabens (Figs 11, 13, 14). Possible source areas for the sandstones were to the north-east, where the Ula Formation and Bryne Formation sandstones were located, and to the west, on the Mid North Sea High (Figs 15B, C). Fluvial systems and/or marine current and wave activity may have facilitated sediment transport to the Gertrud and Heno Plateaus. A similar scenario was envisaged by Howell *et al.* (1996) for the southern UK Central Graben.

The turbidite and debris flow sandstones of the Upper Volgian – Ryazanian constitute potential reservoirs (Fig. 23). Most of the sands were deposited during depositional phase 7, when subsidence had decreased, and parts of the Danish Central Graben had broken up into fault-bounded sub-basins. Inactivity of some Coffee Soil Fault segments (Møller & Rasmussen 2003, this volume) may have increased the drainage area in the marginal hinterland and opened new sediment transport routes, causing an increased sand supply. Sediment dispersal in the receiving basin was still hampered by a topography dominated by tilted fault blocks and many small sub-basins. The most extensive sandstones are present immediately above the Ryaz-1 basal sequence boundary.

Sandstones have not been encountered in the few released wells drilled in the Danish part of the Outer Rough and Ål Basins. However, the presence of Volgian reservoir sandstones of a type similar to the Heno Formation is reported from wells nearby in the UK sector (Mackertich 1996; Spathopoulos *et al.* 2000).

Conclusions

The depositional history of the Danish Central Graben in the Jurassic shows a relationship to contemporaneous rift stages and to relative and eustatic sea-level changes. The basin development in the Jurassic is subdivided into seven depositional phases that reflect this interplay:

Phase 1 (Hettangian–Pliensbachian). During phase 1, pre-rift marine mudstones were deposited in a stable epicontinental sea, that stretched across a major part of the North Sea region. The deposits were removed from large parts of the central North Sea region by subsequent erosion (phase 2).

Phase 2 (latest Pliensbachian – latest Aalenian/earliest Bajocian). This corresponds to the tectonic proto-rift stage, being characterised by domal uplift and regionally extensive erosion.

Phase 3 (latest Aalenian/earliest Bajocian – Late Callovian). The Middle Jurassic sandstones owe their presence to erosion of the North Sea dome, and to slow generation of accommodation space during initial rifting (initiation of the tectonic rift stage). Subsidence rates increased in the Callovian and Oxfordian, heralding the onset of the climax of the tectonic rift stage. This resulted in marine flooding and cessation of sand deposition.

Phase 4 (Early Oxfordian – Early Kimmeridgian). In the Oxfordian – Early Kimmeridgian, the Tail End Graben entered the rift climax resulting in the development of a half-graben. Marine mudstones were deposited during this phase.

Phase 5 (Late Kimmeridgian). Extensive sand deposition took place in the early Late Kimmeridgian during a halt in subsidence between two rift pulses. Sand was deposited over a large area with low accommodation space, which was created on the hanging-wall plateau. Exposed highs and land areas outside the graben supplied the sand. Due to the decrease in subsidence in the Tail End Graben, input of fine-grained sediment could catch up with accommodation space generation, probably resulting in a flat topography at the end of this phase.

Phase 6 (latest Late Kimmeridgian – middle Middle Volgian). During this phase, renewed rifting caused rapid subsidence in the study area. Several rift pulses occurred during this interval. Due to the high overall rate of subsidence, sand deposition associated with relative sea-level falls did not occur. The Danish Central Graben began to break up into minor sub-basins during this phase.

Phase 7 (late Middle Volgian – Ryazanian). Subsidence decreased over large areas of the basin, which had probably entered an early post-rift stage. Major segments of the Coffee Soil Fault became inactive, resulting in a shallower, more symmetrical basin. Active subsidence continued in a number of sub-basins. Highly organic-rich mudstones are prominent in the deposits of this phase. The mudstones are interbedded with turbidite and debris flow sand deposits locally, in particular along the Tail End Graben – Gertrud Graben basin axis and along the eastern margin of the Tail End Graben.

Comparison of the sea-level curve constructed for the Danish Central Graben with published sea-level curves indicates that deposition in the periodically very rapidly subsiding rift basin was strongly influenced by tectonics. In the latest Bathonian – late Early Kimmeridgian, the overall deepening trend resulted from both eustatic sea-level rise and rift-related subsidence. During the rest of the Late Jurassic, the deepening trend was primarily related to the high subsidence rate associated with rifting. Most sediments were deposited between rift pulses in accommodation space that was generated during the rift pulses or by passive subsidence; only a relatively minor proportion of the sediments was deposited during the rift pulses.

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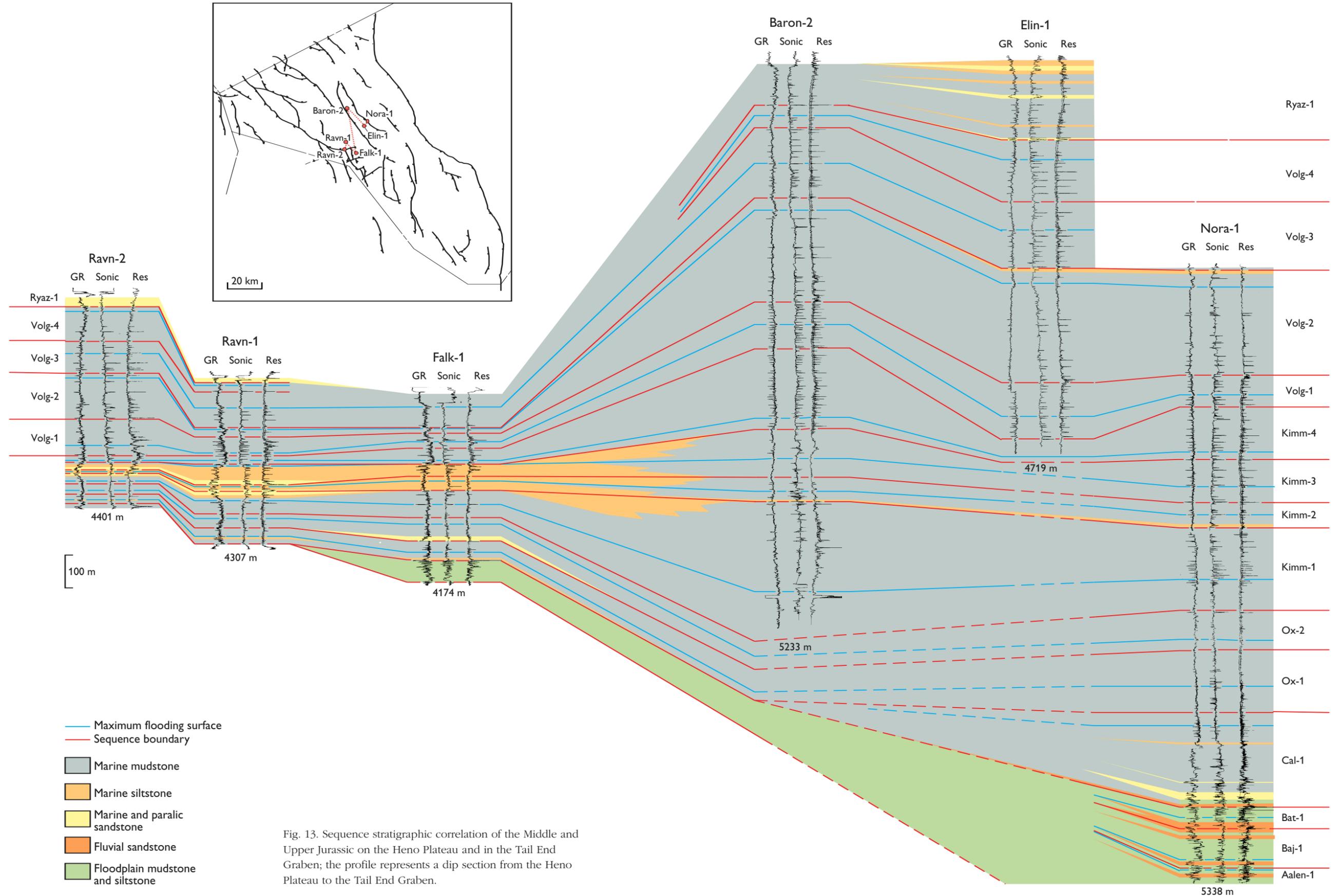
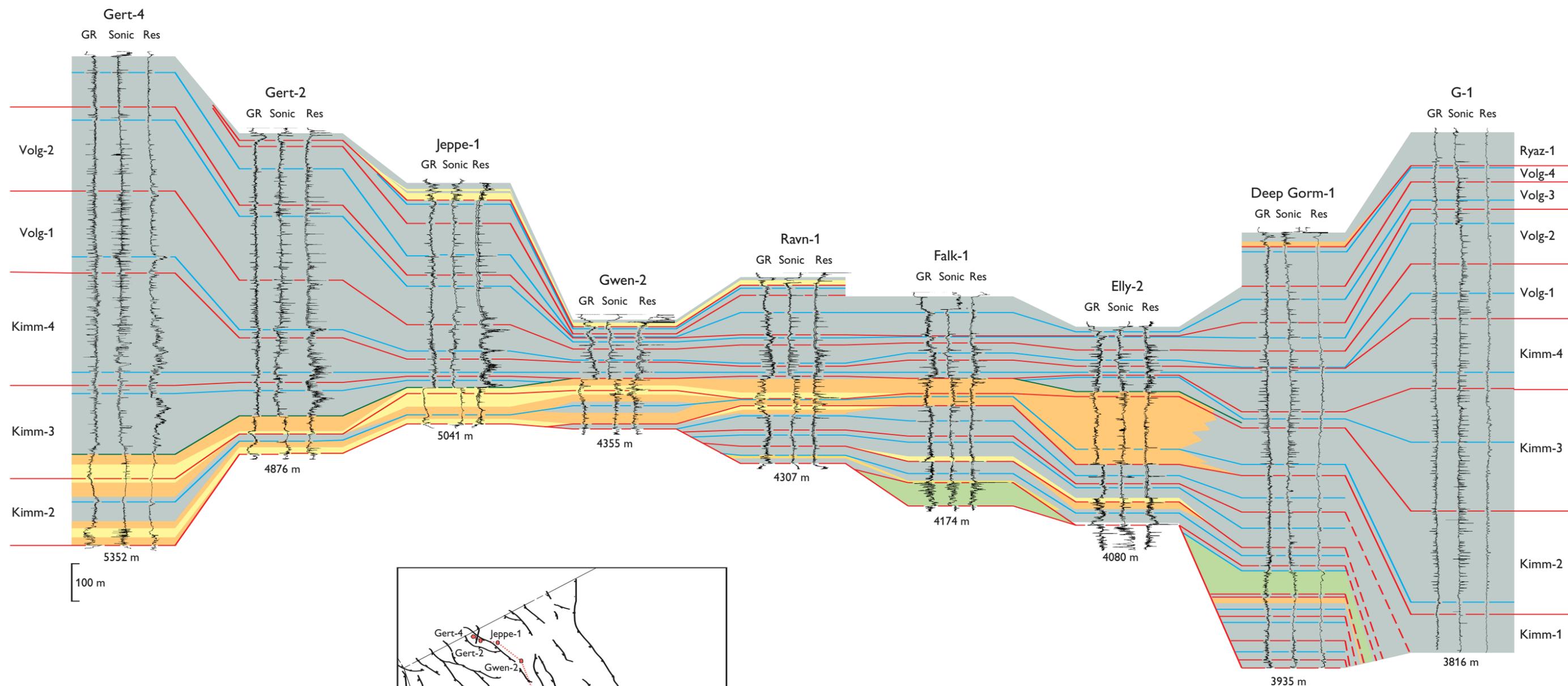


Fig. 13. Sequence stratigraphic correlation of the Middle and Upper Jurassic on the Heno Plateau and in the Tail End Graben; the profile represents a dip section from the Heno Plateau to the Tail End Graben.



- Maximum flooding surface
- Flooding surface
- Sequence boundary
- Marine mudstone
- Marine siltstone
- Marine and paralic sandstone
- Fluvial sandstone
- Floodplain mudstone and siltstone

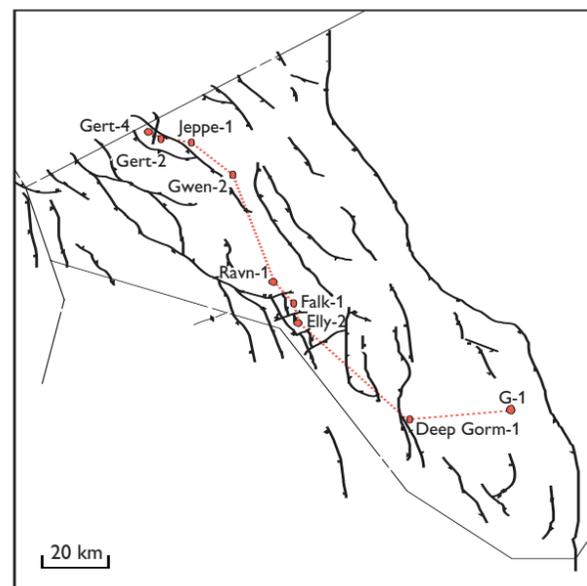


Fig. 14. Sequence stratigraphic correlation of the Upper Jurassic succession from the Feda Graben across the Heno Plateau to the southern Tail End Graben. Onlap of the base Jurassic unconformity can be seen on the southern part of the Heno Plateau (i.e. from Deep Gorm-1 to Ravn-2). The base Kimm-2 sequence boundary exhibits a distinct log break, interpreted as an erosional surface that cuts into underlying marine mudstones in the southern part of the study area. The upper part of the Kimm-3 sequence and possibly the lowermost part of the Kimm-4 sequence onlaps the Kimm-3 flooding surface in the Feda Graben and on the southern Heno Plateau.

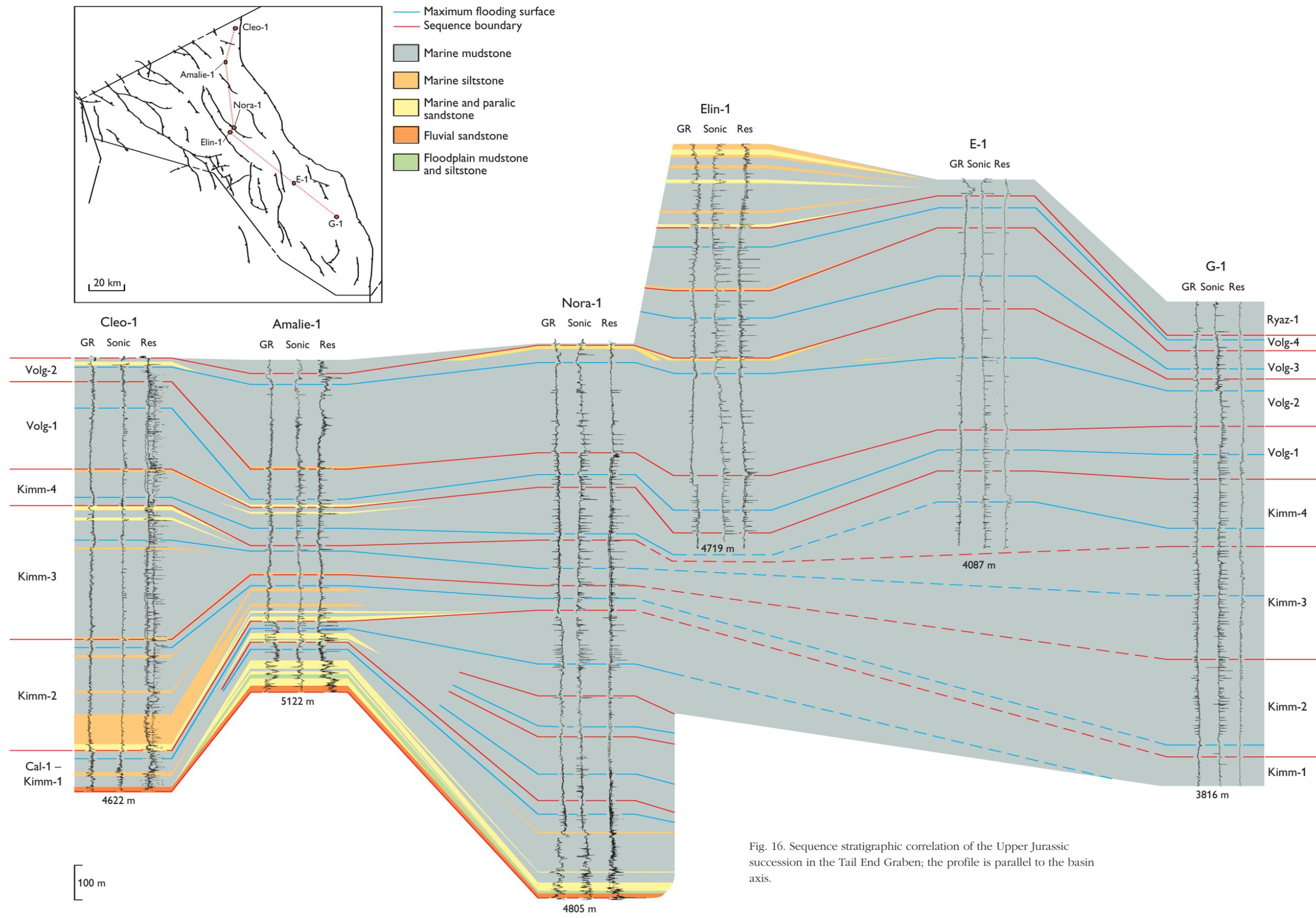


Fig. 16. Sequence stratigraphic correlation of the Upper Jurassic succession in the Tail End Graben; the profile is parallel to the basin axis.

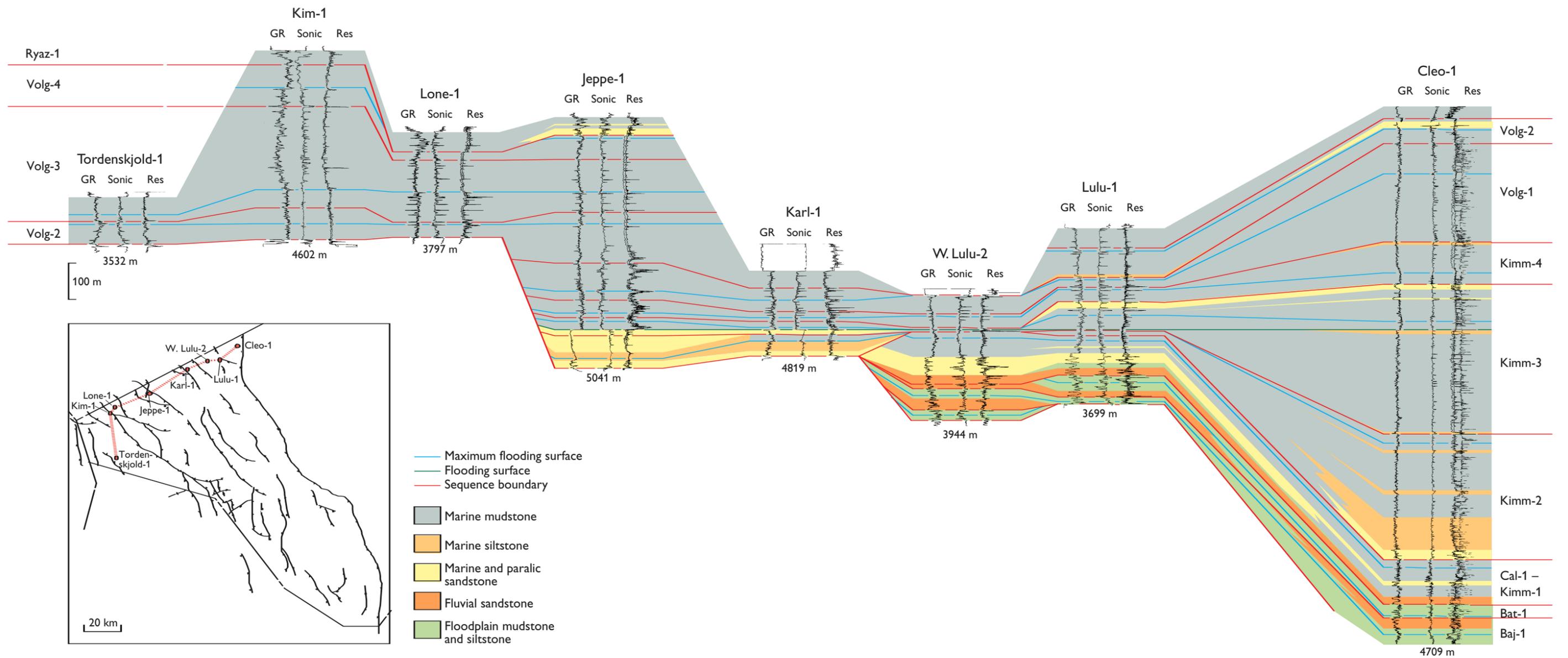


Fig. 17. Sequence stratigraphic correlation of the Upper Jurassic in the northern part of the Danish Central Graben. The profile is perpendicular to the main structural elements from the Outer Rough Basin in the west to the Søgne Basin in the east. The sub-basins were transgressed in a stepwise manner from east to west from the Middle Jurassic to the earliest Early Volgian.