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Tectonic modelling of the Middle Jurassic synrift stratigraphy in the Oseberg–Brage area, northern Viking Graben

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ABSTRACT

A finite difference model, allowing for episodic movements along different faults, is used to examine the effect of tectonics on the stratigraphic signature in the Oseberg–Brage area in the northern Viking Graben. Constraints are provided by local exploration and production well data and 3-D seismic coverage, and a regional depth-converted seismic line.

In the modelling, we focus on the influence of varying rates of fault movement on stratigraphic signatures such as upflank unconformities and changes in layer thickness. We couple the basinwide features of the northern Viking Graben with the fault-block-scale features of the Oseberg–Brage area by using parameter constraints derived by large-scale modelling as input for the local-scale model. In addition, subsidence patterns resulting from the basinwide model were used as background subsidence for the fault block model of the Oseberg–Brage area.

The model results indicate that the alternating activation of different faults with varying extension rates can cause stratigraphic features such as unconformities, condensation and onlap/offlap patterns. Onlap occurs during periods of low extension rates. An increase in extension rate along a fault causes footwall uplift, resulting in condensation or upflank erosion yielding unconformities. This influence can also affect sub-basins further away from the fault. Downdip layer thickening reflects the local tilting of fault blocks.

The coupling of the local and regional scales turns out to be essential in explaining the stratigraphy of the Oseberg–Brage area: basinward and, notably, updip layer thickening as observed on some of the fault blocks can only be explained by activity of the boundary fault on the opposing, western margin of the northern Viking Graben.

INTRODUCTION

In actively extending rift systems, tectonics is likely to be of significant influence on the basin-fill stratigraphy (e.g. Frostick & Steel, 1993). Both accommodation space and the potential for erosion are to a large extent controlled by fault movements, by the creation of half-grabens and uplifted footwall islands, respectively. Other factors playing an important role are eustatic sea-level changes (Vail *et al.*, 1977; Van Wagoner *et al.*, 1990), sediment compaction (Bond & Kominz, 1984), flexural changes in the shape of the basin due to intraplate

stresses (Cloetingh, 1986) and sediment supply (Schlager, 1993), controlled, for example, by variations in climate (Crowley, 1983; Olsen, 1986).

Numerous authors recognized the significance of variations in stretching rates for basin stratigraphy (e.g. Gloppen & Steel, 1981; Blair & Bilodeau, 1988; Steel, 1988; Gordon & Heller, 1993), and several subdivisions of the rift evolution into phases with their own stratigraphic characteristics have been proposed, based on variations in the rate of extension (e.g. Hamblin & Rust, 1989; Lambiase, 1990; Prosser, 1993; Nøttvedt *et al.*, 1995; Ravnås & Bondevik, 1997). In all these studies, the stretching rate is assumed to increase to a maximum value, and subsequently decrease again. For example, Prosser (1993) proposed a four-fold division of the rift

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Fig. 1. In basin studies various features are considered, depending on the scale of the problem. Regional studies concentrate on structural features, whereas on the local scale attention is focused on stratigraphic patterns. However, the large-scale deformation strongly influences the small-scale stratigraphy.

evolution to characterize basin-fill stratigraphy: (1) the rift initiation stage, where the rate of displacement is relatively low and sedimentation keeps pace with subsidence; (2) the rift climax, where the rate of displacement has reached its maximum value and sedimentation is likely to be outpaced by subsidence; (3) the immediate postrift stage, where active tectonism has ended, only a (decreasing) regional subsidence due to thermal cooling is left and the infilling of the remnant topography starts; and (4) the late postrift stage, in which a slow peneplanation of the remaining low-relief topography takes place. Ravnås & Bondevik (1997), based on a case study of the Heather Formation in the Oseberg–Brage area, made a similar kind of subdivision but on a smaller scale. They first subdivided the rift evolution into short-period ‘rift-phases’. These were subdivided into a tectonic quiescence stage and a rotational tilt stage, of which the latter was further subdivided into an early synrift, a rift-climax and a late synrotational substage.

The above models are based on half-graben type basins, and can be applied on highly variable scales, both in time and in space. However, some complications arise. Local-scale events are influenced by the effects of regional-scale phenomena, such as the flexural response on crustal deformation in the surroundings, and movements along neighbouring master faults (see Fig. 1). Furthermore, most rift systems are composed of a series

of half grabens, and show a migration of rifting with time, the extension shifting from one fault to the other. This is described, for example, for the Mesozoic Southern Alps (Bertotti *et al.*, 1993), the East African rift system (Rosendahl, 1987; Bosworth, 1992) and the Viking Graben in the northern North Sea (Badley *et al.*, 1988). The mutual interaction of the resulting subbasins, both by the overlap of influenced areas and by flexural effects, complicates the idealized picture.

An example of an area where the stratigraphy is probably mainly controlled by extensional tectonic forces – rather than by eustatic sea-level changes – and where a strong interaction of various faults has taken place is the Oseberg–Brage area in the eastern part of the northern Viking Graben (Badley *et al.*, 1984; Ravnås *et al.*, 1996; Færseth & Ravnås, unpubl. obs.; see Fig. 2). Færseth & Ravnås (unpubl. obs.) investigated the Middle Jurassic to Early Cretaceous development of this area, and recognized periods of significant rotational faulting in the late Bajocian to late Bathonian, late Callovian, late Oxfordian–Kimmeridgian and middle Volgian. Each of these events was probably accompanied by footwall uplift, large enough to create upflank unconformities. The stratigraphic signature of contemporary downflank deposits (Ravnås *et al.*, 1996; Ravnås & Bondevik, 1997) suggests that the formation of these unconformities could not have formed as a result of relative sea-level lowering

Fig. 2. (a) The northern Viking Graben and its surroundings. Inset: the position of the Oseberg/Brage area. The bold line indicates the position of the seismic line used for the large-scale model. (b) Detailed map of the Oseberg-Brage area. The grey line connects the boreholes used for the construction of the well-log profile.

Fig. 3. (a) East-west correlation of the Tarbert Formation and Lower Heather from the Horda Platform, across the Oseberg Fault Block and towards the Viking Graben (after Ravnås *et al.*, 1996). (b) Schematic overview of the features to be explained by tectonic movements, using the numerical model.

alone. Wedge-shaped, divergent layered sediment packages, showing large variations in thicknesses between individual fault blocks, are thought to have been deposited during short periods of active rotational faulting, suggesting a pulsatory nature for the extensional faulting. The parallel-layered sediments deposited during the intervening tectonic quiescence stages display some basinward thickening, indicating the possible occurrence of some basinwide anticlockwise rotation as well (Ravnås *et al.*, 1996; Færseth & Ravnås, unpubl. obs.; see Fig. 3a).

In this paper we investigate to what extent the stratigraphic features observed in the Oseberg–Brage area can be explained by tectonic mechanisms. Emphasis is put on the influence of variations in fault activity and extension rates in reproducing stratigraphic surfaces such as upflank unconformities and lateral and vertical changes in layer thickness. We focus on a short time interval of approximately 10 Myr in which rotational faulting is initiated, from the latest Early Bajocian to the latest Bathonian – earliest Callovian.

A finite difference model is used, describing crustal deformation due to extensional faulting in the crust (ter Voorde & Cloetingh, 1995). A specific novel feature of the model is that various faults can be activated in different time intervals and that stretching rates can be changed with time. The incorporation of faults makes the model applicable not only for regional-scale but also for local-scale modelling. This probably will become of increasing importance in the future, as a result of the growing shift in exploration and production towards smaller fields (Gabrielsen & Strandenes, 1994).

Regional-scale modelling, concentrating on structural features of a basin, provides constraints on lithosphere properties such as rigidity, Moho depth and fault con-

figuration, whereas on the local-scale attention is focused on stratigraphic patterns (Fig. 1). The coupling of these different spatial scales is made by using parameter constraints and subsidence rates obtained by regional-scale modelling as input and background subsidence, respectively, for the local-scale modelling. This coupling, accomplished by combining data sets such as basinwide seismic sections and detailed well-log analysis, will turn out to be essential to obtain a good understanding of basin evolution (e.g. Cloetingh *et al.*, 1994).

STRUCTURAL SETTING

The northern Viking Graben

The northern Viking Graben forms part of the northern North Sea rift system, and is situated between 60° and 62° N (Fig. 2). It separates the Horda Platform in the east from the East Shetland Basin in the west. The Viking Graben structure is dominated by faults with N–S, NNW–SSE and NE–SW orientations, typically 15–20 km apart, defining large tilted fault blocks. As shown by, among others, Marsden *et al.* (1990), the structure is asymmetrical in the north, between the Tampen Spur and the Horda platform, where deformation took place mainly along east-dipping normal faults. To the south, between 61° and 60°30' N, the structure becomes more symmetrical, and returns to an asymmetric pattern further south.

Growing consensus exists about the occurrence of at least two rifting phases in the Viking Graben, the first one in the ?Late Permian – Early Triassic, the second one in the Middle Jurassic – Earliest Cretaceous. During the Early–Middle Jurassic, the North Sea basin experi-

enced the effects of both thermal cooling related to the Permo-Triassic event and the increased subsidence heralding the initiation of the Jurassic-Cretaceous event (e.g. Giltner, 1987).

The second rifting phase is generally assumed to have taken place from the late Bajocian – early Bathonian to the Ryazanian (Yielding *et al.*, 1992; Badley *et al.*, 1988; Marsden *et al.*, 1990), with a peak in the Oxfordian–Kimmeridgian. The onset of rifting was not synchronous over the area. On the Horda platform, for example, tilting did not occur until later in the Jurassic (e.g. Badley *et al.*, 1988). From the Bajocian, the effect of increased fault activity is noticeable in the basin (Gabrielsen *et al.*, 1990; Helland-Hansen *et al.*, 1992; Mitchener *et al.*, 1992), in terms of enhanced differential subsidence across faults and by early rotational uplift on some fault blocks. The time interval of our study thus includes this period of increased fault activity and the first phases of the rifting episode.

The Oseberg-Brage area

The Oseberg-Brage area comprises the western margin of the Horda Platform in the east, the Oseberg Fault Block and a number of smaller fault blocks in the south-west, forming the transition zone between the Horda Platform and the eastern flank of the northern Viking Graben (Fig. 2). The Brage area forms the updip part of a gently easterly rotated mega-fault block, the Horda Platform. The western part of this block is a narrow, asymmetrical horst, the Brage Horst, which is separated from the Oseberg Fault Block to the west by the N-S-trending, west-dipping Brage Fault. The Oseberg structure has a present-day width of c. 16 km in its central part, and narrows to the south and the north. It shows a tilt to the east, and consists of a series of smaller fault blocks, e.g. the α - and the γ -structures shown in Fig. 2(b). The Oseberg Fault Block is separated from the graben margin area in the west by the west-dipping, normal Oseberg Fault. The smaller fault blocks of the terrace area, representing the western part of the initial Oseberg mega fault block (Færseth & Ravnås, unpubl. obs.), consist from the east to the west of the ω -, the B- and the G-structures, which are separated by west-dipping faults (Fig. 2b).

During the early rift stage, from the late Bajocian to the late Oxfordian, movement on N-S-trending faults dominated, taken over by movement along NE-SW-trending faults during later stages (Ravnås & Bondevik, 1997; Færseth & Ravnås, unpubl. obs.). From the early Bajocian to the early Callovian, significant differences can be observed between the Brage Area, the Oseberg Fault Block and the graben margin area. While the Brage area was characterized by erosion, the Oseberg Fault Block was tilted gently to the east, resulting in the deposition of wedge-shaped, eastward-thickening sediment packages and the formation of upflank unconformities. In the terrace area faults separating smaller blocks influence the

sediment thicknesses, giving evidence of early rotational faulting, and suggesting the area became progressively more strongly segmented (Ravnås *et al.*, 1996).

From the early Callovian, periods of increased stretching alternated with periods of relative tectonic quiescence, and resulted in the deposition of a series of wedge-shaped packages separated by dark mudstones of off-shore origin.

In this paper, we will concentrate on the initial rift stage, from the latest Early Bajocian to the latest Bathonian – earliest Callovian.

DATA SET

The Oseberg-Brage area has been extensively investigated by drilling and seismic profiling. The Bathonian-Ryazanian synrift architecture is documented with exploration and production well data from blocks 30/3, 30/6, 30/9 and 31/4 and 3-D seismic coverage (Ravnås *et al.*, 1996; Ravnås & Bondevik, 1997; Færseth & Ravnås, unpubl. obs.). Figure 3 shows an east-west profile of the uppermost Ness, Tarbert and lower Heather Formations (Vollset & Doré, 1985) through the area, constructed from these data. The numbers in Fig. 3 indicate timelines, of which the ages are given in Table 1. The location of the well-log correlation profile is indicated in Fig. 2.

The subdivision of the uppermost Ness and Tarbert Formations and the lower Heather member into chronostratigraphic units is based on detailed quantitative biostratigraphy, showing that this interval was deposited during the *Humphriesianum* to *Discus* ammonite zones or the latest Early Bajocian-Bathonian. A series of timelines is recognized in the studied interval (Table 1; Ravnås *et al.*, 1996, their fig. 3) for which the boundary criteria are given in Helland-Hansen *et al.* (1992). The timelines have been tied to the Harland *et al.* (1990) time-scale, assuming equal lengths of ammonite zones within the stages. This was necessary in order to provide age determinations as input parameters for the numerical simulation. Hence, the ages given for each timeline in Table 1 should merely be regarded as guide-lines.

In Fig. 3, upflank unconformities can be observed on the B-, γ -, α - and Brage structures. Most of the sedimentary layers in the well-log profile show a thickening towards the east, which is explained by the geometry of

Table 1. Ages of the timelines indicated in Fig. 3. See text for how these ages are obtained.

Number	Age (Ma)	
9	170.7	Bajocian
10	169.3	Bajocian
11	167.7	Bajocian
12	166.1	Bajocian/Bathonian
13	164.9	Bathonian
14	161.3	Bathonian/Callovian
15	159.9	Callovian

the accommodation space, created by extension along west-dipping faults. These wedge-shaped sediment packages indicate synrift deposition in the fault-bounded half-grabens. However, between timelines 10 and 11 (thus between 169.3 and 167.7 Ma) a layer of lower delta plain deposits can be observed, clearly showing thickening towards the west. This basinward thickening of the lower delta plain deposits of the Tarbert Formation, in an *updip* direction on the present fault blocks, indicates anticlockwise rotation of the whole region.

In addition to the well-log profile, giving detailed information about the Oseberg–Brage area, we used a regional, depth-converted seismic line (NVGT 88-04), the position of which is depicted in Fig. 2. This line is shown in Fig. 4.

DESCRIPTION OF THE NUMERICAL MODEL

The finite difference model we use describes fault movements and their thermal and flexural response in an extensional regime. The lithosphere is represented by a two-layer model: an upper layer in which extension occurs by localized deformation (i.e. faulting) and a lower layer with a more distributed thinning (Fig. 5). The boundary between these two layers is assumed here to coincide with the level at which the faults sole out (the ‘detachment level’). In both layers, the condition of volume conservation is fulfilled (which implies that compaction is not included). The number of faults and their geometry are given as model input, and the sequence of fault activation as well as extension rates can be varied. The model is described extensively by ter Voorde & Bertotti (1994) and ter Voorde & Cloetingh (1995).

If the response on the mass redistribution caused by the fault movements is an upward state of flexure, this results in uplift of the footwall (e.g. Kusznir *et al.*, 1991; see Fig. 5b). The redistribution of loads due to the extension is dependent on the depth of the Moho, being a controlling factor for the density distribution before stretching, and the depth of the boundary between localized and distributed thinning. This boundary determines the ratio between thinning of the upper crust, where crustal material is replaced by sediments with, in general, low densities, and thinning of the lower lithosphere, where crustal material is replaced by dense mantle material. Therefore, the existence and amount of footwall

uplift is influenced not only by the size of the fault blocks and the amount of extension (Yielding, 1990) but also by the depth of the Moho, the depth of the localized-distributed deformation boundary, the sediment density and the lithosphere rigidity. These are important parameters that can be varied in the model. An increase in Moho depth results in a decrease in footwall uplift, whereas an increase in the depth of the detachment level causes an increase in footwall uplift (see ter Voorde & Cloetingh, 1995).

The position of the fault block with respect to the location of maximum thinning in the lower lithosphere determines the relative amount of mantle material below, and is thus also of great importance for the occurrence and amount of footwall uplift. The assumption that the amount of extension is equal for the lower layer and the upper layer is only realistic if we consider whole basins. If we use the model on a sub-basin or smaller scale (Fig. 5b,c), the amount and geometry of lower-crustal thinning is dependent on the position of the modelled region in the basin.

The Oseberg–Brage area is positioned in the eastern flank of the northern Viking Graben, where the lower-crustal thinning has a moderate, gently eastward-dipping geometry (Beach *et al.*, 1987; Christiansson *et al.*, 1997). Therefore, the condition of equal extension for the upper crust and lower lithosphere is not applicable. We assumed that no lower layer thinning occurred in the Oseberg–Brage area during this time, implying that only the brittle layer plays a role in the local-scale simulation runs described below. It should be kept in mind, however, that lithosphere deformation in the surroundings causes a flexural response. In this case this probably results in a mild tilting towards the west, which we thus imposed on the local scale model.

THE GENERATION OF UPFLANK UNCONFORMITIES

In the well-log correlation profile in Fig. 3(a), upflank unconformities can be observed, for example on the B-structure, on the γ -structure and on the α -structure. These unconformities are explained by the pulsating nature of rifting, an increase in stretching rate causing uplift of the footwall (e.g. Ravnås *et al.*, 1996). Figure 6(a) shows the idealized seismic expression of a half graben that evolved as a result of local tectonics, and in which

Fig. 4. Line drawing of seismic line NVGT 88-04.

Fig. 5. The finite difference model used in this study couples large-scale deformation to fault block or local-scale phenomena. (a) The lithosphere is represented by a two-layer model: an upper layer in which extension occurs by localized deformation and a lower layer with a more distributed thinning. The boundary between these two layers coincides with the level at which the faults sole out. (b) If the response on the mass redistribution caused by the fault movements is an upward state of flexure, this results in uplift of the footwall. (c) On the local scale, stratigraphic patterns can be modelled.

stratigraphic features like unconformities, condensation and onlap and offlap patterns are explained by changes in stretching rate. Also here, the upflank unconformity originates in the period of increasing and maximum rotation rate.

The observed upflank unconformities may be explained by the fact that sedimentation does not always keep up with the rate of basin subsidence: basins become underfilled if the rate of fault-related subsidence is high compared with the sedimentation rate or in cases of sediment input starvation (e.g. Gloppen & Steel, 1981; Blair, 1987; Prosser, 1993; see also Roberts *et al.*, 1993). For the simulation of the unconformities, the sedimentation rate was thus kept constant in time, with the limitation that the sediments could not rise above sea level. The sedimentation rate was chosen such that it was in exact equilibrium with the initial, low rate of basin subsidence (implying lateral variations in sedimentation rate). An increase in stretching rate then leads to an underfilled basin and an increase in flexural footwall uplift.

We simulated the creation of the unconformities by imposing a change in the rate of faulting during extension, starting with a low stretching rate, that first increased and then decreased. We assumed that the sea level was just below the initial basement surface and did not change with time. The modelling result is displayed in Fig. 6(c), showing a high rotation rate leading to footwall uplift, followed by a period of low rotation rate in which an overall subsidence takes place. The following features can be observed.

1 During the period of low extension rate, sedimentation keeps up with subsidence, and the basin is filled with sediments up to the sea level continuously. Due to the widening of the area that submerges below sea level, the sediments show an onlap pattern.

2 Since the sedimentation rate is constant, an increase in extension rate finally causes an underfilled basin, leading to upward flexural compensation. This results in enhanced uplift of the footwall and erosion of the upflank sediments. The wedge shape of the accommodation space causes divergent layers.

3 During lower extension rates the basin is again filled with sediments and subsides, causing onlap on the unconformity. If the fault is not activated again, deposition will continue until the topography is smoothed.

The pattern is independent on the scale of the problem; a multiplication of the stretching rates, the sedimentation rate and the depth of sea level by a constant results in a multiplication of the vertical scale of the figure by the same constant. The applied extension rates for this case are summarized in Table 2.

Figure 6 thus demonstrates that the model explains not only the unconformity but also the onlap pattern that develops during the slow extension phases. The same result was obtained by Hardy (1993), using a numerical model of domino fault block development. It should be emphasized that the *only* parameter we changed in the model was the rate of extension; all the features described thus can be explained even without taking into account the effects of eustatic sea-level changes, sediment supply and sediment compaction.

Fig. 6. Idealized seismic expression (a) and geological interpretation (b) of a half-graben, showing stratigraphic features caused by changes in rotation rate (after Ravnås & Bondevik, 1997). (c) Simulation run of extension with varying extension rate. Periods and rate of stretching are given in Table 2. All other model parameters were kept constant; the stratigraphic signatures are thus caused *only* by the variation in extension rate.

SMALL-SCALE EFFECTS CAUSED BY LARGE-SCALE PROCESSES: THE CHANGES IN LAYER THICKNESS

The Oseberg area should not be viewed as an isolated region, as it is strongly influenced by fault movements in the surroundings. The area can be subject to movements along neighbouring master faults or to the flexural

response on crustal deformation in the near environment. One of these mechanisms, or a combination of both, could be responsible for the anticlockwise rotation indicated by the basinward, updip layer thickening observed in the lower delta plain deposits between timelines 10 and 11 in Fig. 3.

Since the rift axis of the northern Viking Graben lies west of the modelled region (Fig. 2b), we expect the

Table 2. Extension rates used for the synthetic model shown in Fig. 6.

Period (Ma)	Extension rate (km Myr ⁻¹)
5.0–2.0	0.36
2.0–1.6	0.48
1.6–1.3	0.60
1.3–0.8	1.20
0.8–0.5	0.60
0.5–0.2	0.30
0.2–0.0	0.00

large-scale flexural response on the basin development to be a tilting towards the west. At the same time, the major fault forming the western boundary of the Viking Graben is dipping to the east, and therefore activation along this fault will also produce a tilting towards the west. Obviously, such a tilting generates layer thickening towards the west. To verify which mechanism is the most plausible, we modelled both scenarios.

Large-scale flexural modelling of the northern Viking Graben

Since the lithosphere acts as a stress guide, local deformations cause a regional response and thus exert influence on each other. Therefore, in order to obtain a better understanding of the small-scale development of the Oseberg–Brage area, the flexural response on movements in adjacent parts of the northern Viking Graben should be taken into account.

For this, we used the regional seismic line NVGT 88-04 shown in Fig. 4. Although this line is situated to the north of the well-log correlation profile (see Fig. 2), and thus gives a different basin configuration, we feel confident in using it for the calculation of the large-scale flexural deformation in the area. Large differences in this flexural deformation are not expected over this relatively short distance of approximately 25 km.

In order to constrain the values for the effective elastic thickness of the lithosphere and the depth of the brittle–ductile boundary, we first simulated the development of the northern Viking Graben in the entire Jurassic–Cretaceous time interval.

From the seismic line NVGT 88-04 (Fig. 4), the position of the Jurassic–Cretaceous sediments was extracted (Fig. 7a). The thickness of these sediments was determined and, as compaction is not included in the model, the decompacted thickness was calculated using an exponential porosity–depth relation

$$\phi = \phi_0 e^{-cz}$$

where ϕ is porosity, ϕ_0 is surface porosity, c is a porosity–depth coefficient and z is depth. We choose $\phi_0 = 0.56$ and $c = 0.39 \text{ km}^{-1}$, which are the values Sclater & Christie (1980) derived by averaging the values for shale and for sand in the North Sea. We thus obtained the decompacted

sediment thickness that could be used as a constraint for the large-scale modelling (Fig. 7b). The best fit with the data (Fig. 7c) was obtained by using the fault configuration of Fig. 7(d), and the parameters given in Table 3.

Subsequently, using the resulting values for EET, Moho depth and brittle–ductile boundary as model input, we determined what kind of fault movements could have produced the amount of westward tilting necessary to explain the layer thickening observed.

To obtain a reasonable fit between the basinward layer thickening that was measured and the modelling results, it turned out to be necessary to constrain the fault activity and its associated lower crustal thinning to the basin centre (Fig. 8, upper panel). If we assume that all the thermal relaxation occurs during rifting, a total extension of 6.9 km in the basin centre can explain the shape and amount of the observed layer thickening (see Fig. 8). From the constraints given by the timelines, we know that this rifting must have taken place in a timespan of maximum 1.6 Myr, yielding a very high extension rate of at least 4.3 km Myr^{-1} . However, as pointed out for example by Jarvis & McKenzie (1980) and by ter Voorde & Bertotti (1994), the assumption that thermal conduction keeps up with thermal advection (by which we mean the transport of heat with the lithospheric material) during the rift event is only valid for low extension rates, and thus cannot be held for this case.

To study the consequences of the inclusion of thermal effects, we considered the two possible end-members of the rifting process: instantaneous rifting where heat advection rules out heat conduction, and infinitely slow rifting where heat conduction rules out heat advection.

In Fig. 8 an overview is given of the total flexural subsidence after 6.9 km of extension, assuming an infinitely low extension rate (case a) and an infinitely high extension rate (case b). Different wavelengths for the thinning of the lower layer were considered. From the figure, it is evident that the inclusion of advective effects not only reduces the flexural subsidence but at high stretching rates even leads to flexural uplift. In Fig. 9 the flexural subsidence in time is given for the case in which the 6.9 km of extension is accomplished in 1.6 Myr, with a lower layer thinning as in Fig. 8(a). Apparently, this extension rate leads to flexural uplift.

Since the exact age of the timelines is not well constrained, no value can be attached to the exact amount of this flexural uplift. However, as shown by ter Voorde & Bertotti (1994), advective effects, reducing the amount of isostatic subsidence, play a role already for extension rates as low as 1 km Myr^{-1} .

We are thus confronted with the following problem: a high extension rate is required to explain the occurrence of layer thickening in a very short timespan, but a low extension rate is needed to cancel out the effects of thermal advection, counteracting the flexural subsidence. Obviously, the basinward layer thickening observed in the well-log profile calls for an alternative explanation.

Fig. 7. (a) The position of the Jurassic–Cretaceous sediments as extracted from the seismic line. (b) Thickness of the Jurassic–Cretaceous sediments. Dashed: decompacted; solid: not decompacted. (c) Decompacted thickness of the Jurassic–Cretaceous sediments. Dashed: modelled; solid: data. (d) Fault configuration used to model Jurassic–Cretaceous deformation.

Table 3. Parameters used for the large-scale modelling of the northern Viking graben.

Period of rifting	166.0–140.5
Total extension	24 km
Detachment depth faults	10 km
Initial Moho depth	29 km
Effective elastic thickness	5 km

Table 4. Parameters used for the small-scale modelling of the Oseberg Area.

Deformation time	170.7–159.9 Ma
Effective elastic thickness	5 km
Amount of extension along major fault (outside model)	0.6 km
Period of extension along major fault (outside model)	169.3–167.7 Ma

Extension along a major boundary fault

A major N–S-trending, east-dipping fault, indicated in Fig. 2 as the Hild–Brent–Statfjord (HBS) lineament (after Færseth & Ravnås, unpubl. obs.), bounds the Viking Graben on the western side at a distance of approximately 40 km from the Oseberg Fault. Extension along the HBS lineament will result in a westward tilting of the hangingwall, the size of which is determined by the fault geometry. Provided that the fault is wide and deep enough, this tilting will affect the Oseberg–Brage area, and thus could be the alternative cause of the westward layer thickening. From numerical modelling we established that 0.6 km of extension along a listric fault positioned at 40 km west of the Oseberg fault and detaching at 10 km depth can explain the shape and amount of layer thickening observed in the well-log profile. This is shown in Fig. 10(a).

The requirement that the master fault has to be deeper and wider than the faults in the Oseberg area to explain

Fig. 8. Total flexural subsidence after 6.9 km of extension, for two thermal end-members, and different wavelengths of lower layer thinning. Upper row: lithosphere configuration after stretching. Sediments are dark grey, crust is light grey, mantle is white. From left to right, the thinning of the lower layer occurs over a wider area. Middle row: an infinitely low extension rate is assumed, implying the thermal field is only influenced by conduction (case A). Lower row: an infinitely high extension rate is assumed, implying the thermal field is only influenced by advection (case B). Black dots indicate the thickness of the layer of lower delta plain deposits that thins to the east, as measured in the boreholes 30/9-13S, 30/9-7, 30/9-8 and 30/9-2 (see Fig. 3).

Fig. 9. Solid lines: flexural subsidence in period during 1.6 Myr of extension with a rate of 4.3 km Myr^{-1} . Every timeline represents 0.08 Myr. Long dashes: total flexural subsidence if conduction is ignored. Short dashes: total flexural subsidence if advection is ignored.

the basinward layer thickening supports the notion that these smaller faults were formed as antithetic structures on the HBS lineament, as proposed by Færseth & Ravnås (unpubl. obs.), although they appear to have been developed into larger faults subsequently (see Fig. 7). This agrees well with the largely asymmetric structure of the basin both north and south of line NVGT 88-04, where most of the extension seems to have occurred along east-dipping faults, and the eastern margin is relatively gradual

(see, for example, Marsden *et al.*, 1990, their fig. 12.2). The final fault configuration used for the modelling of the Oseberg–Brage area is depicted in Fig. 10(b). Although there may have been continuous faulting along the western graben-bounding lineaments (Færseth & Ravnås, unpubl. obs.), we simulated activation between timelines 10 and 11 only, because of the lack of constraints for the variations in rotation rate along these faults.

NUMERICAL SIMULATION OF THE TEMPORAL EVOLUTION OF THE OSEBERG-BRAGE AREA

We have demonstrated above that the upflank unconformities observed in the Oseberg–Brage area can be caused by changes in extension rate, and that variations in layer thicknesses are affected by the shape of the bounding fault and the extension rate, but also by movements along a deeper and wider fault at a larger distance. The next step is to combine these interpretations, to simulate a tectonic scenario to be compared with the stratigraphic patterns observed in Fig. 3.

Ravnås *et al.* (1996) suggested that the late Bajocian–Bathonian Tarbert Formation and lower Heather member in the Oseberg–Brage area jointly formed three regressive–transgressive landward-stepping shoreline prisms. The uppermost Ness Formation was argued to be the transgressive part of yet another of these shoreline prisms.

Fig. 10. (a) Crustal configuration after 0.6 km of extension along a listric fault, at 40 km west of the Oseberg fault. Inset: the Oseberg area in detail. Black dots indicate the thickness of the layer of lower delta plain deposits that is thinning to the east, as measured in wells 30/9-13S, 30/9-7, 30/9-8 and 30/9-2 (see Fig. 3). (b) Possible fault configuration in the northern Viking Graben, assuming that the smaller faults in the Oseberg–Brage Area form antithetic structures on the Hild–Brent–Statfjord lineament west of the basin axis.

The retreat and drowning of the Brent delta system was related to sediment supply not being able to keep pace with increasing basinal subsidence rates related to the effects of early Middle Jurassic stretching (see also Helland-Hansen *et al.*, 1992; Mitchener *et al.*, 1992). Variations in the rate of extension were favoured as the main controlling factor on the formation of the regressive–transgressive wedges. The regressive segments were suggested to have formed during periods with lower rates of extensional faulting, whereas the transgressive segments were correlated with periods with high stretching rates (Ravnås *et al.*, 1996). Since our purpose is to constrain to what extent the stratigraphic features can indeed be explained by tectonic movements, rather than making a perfect fit of the model with the data, we have restricted ourselves to variations in rates and periods of fault activity.

After an iterative process of adapting extension rates, the rates given in Fig. 11 were found to give the best result. The modelled scenario can be divided into five stages, each representing a period between two timelines, the ages of which are given in Table 1.

Stage 1: 170.7–169.3 Ma

In this period the lower delta plain sediments of the uppermost Ness Formation were deposited, indicating that the basin was filled up to sea level continuously. Since this was an interval of Brent delta backstepping (correlating with the transgressive stage S0 of Ravnås *et al.*, 1996), stretching rates were assumed to be relatively high. Activity was simulated along the Oseberg Fault, the γ – α fault, the Brage Fault and the Brage East Fault only; the G-, B- and ω -structures were considered to form one single fault block. We imposed the sedimentation to keep up with subsidence during 1 Myr, after

which the fast extension continued but the sedimentation rate decreased, leading to slightly underfilled basins (Fig. 12a). This was based on the presence of the marine sandstones of the basal Tarbert Formation, resting upon the Ness Formation, and resulted, by slight uplift and erosion, in the oldest unconformity on the α -structure.

Stage 2: 169.3–167.7 Ma

This is the phase in which the layer thickening towards the west is observed. The subsidence caused by extension along a major master fault positioned west of the modelled region (the HBS lineament in Fig. 2 and Fig. 10) is superposed on the results of the smaller-scale model. This phase correlates with the first of the Tarbert Formation regressive–transgressive sequences (SI) of Ravnås *et al.* (1996). Stretching rates were low during the progradational stage, when the shallow-marine and paralic sediments were deposited, and high during the backstepping stage when there was a return to a marine depositional environment.

At the beginning of the regressive period, during deposition of the shallow-marine sediments, we simulated movements along the Oseberg, γ – α and Brage faults, resulting in the disturbance of the smooth thinning pattern caused by the major-fault activity. Sediments are assumed to fill the basin up to sea level again. These shallow-marine sediments are represented by the lowermost dark layer in Fig. 12(b). We assumed an absence of fault activity along the G–B, Oseberg and γ – α faults during the period in which the lower delta plain sediments were deposited, in order to simulate the generation of the nondisturbed, continuous basinward layer thickening as observed in the well-log profile. This period of nonactivation is indicated by the white layers in Fig. 12(b).

Fig. 11. Extension rates along the faults in the Oseberg–Brage area with time. Periods of fast extension are alternated with periods of slow extension. See text for discussion.

In the transgressive stage, the Oseberg, γ - α and Brage faults were reactivated. This was the onset of separation between the G- and B ω -structures, caused by fast movements along the G–B fault and leading to flexural uplift of the B ω -structure. Since this flexural uplift was accompanied by relatively fast tectonic subsidence along the Oseberg Fault, the net result is not necessarily an erosional unconformity. From well 30/9-7, we infer that the sediment surface was at or near sea level, which is supported by the modelling results (Fig. 12c). The fast movement along the G–B fault and the Oseberg faults continued for a short while (i.e. 0.25 Myr) into stage 3, causing an underfilled basin on the G- and ω -structures. As a result, uplift and erosion of the γ - and α -structures occurred, again leading to upflank unconformities (Fig. 12c,d).

Stage 3: 167.7–166.1 Ma

This period can also be divided into a regressive and a transgressive stage, correlating with sequence SII of Ravnås *et al.* (1996). During the regressive period extension rates along the G–B and Oseberg faults were low, and the underfilled basins that were created at the end of the previous stage were now filled with sediments. This phase of relative tectonic quiescence is observable in the parallel-layered sediment packages on these structures. The transgressive stage was characterized by high stretching rates along the Oseberg and Brage faults, resulting in footwall uplift and erosion, and in the creation of slightly underfilled basins on the ω - and α -structures (Fig. 12d). The higher subsidence of the G-structure produced a deeper basin. Still, only minor exposure is inferred for the B-structure.

Stage 4: 166.1–164.9 Ma

This phase corresponds to the regressive stage of sequence SIII of Ravnås *et al.* (1996). During this stage we modelled only slow extension along all faults, leading to very thin layers on the α - and γ -structures, and slightly thicker layers on the B- and G-structures. The B- and G-structures were still underfilled at the beginning of this period, and filled up with sediments in the course of time (Fig. 12e).

Stage 5: 164.9–161.3 Ma

This is the final time slice covered by our simulation, and a continuation of stage 4, with slow albeit slightly increased movements along all the faults, leading to sediment condensation during this stage. However, this condensation could also be caused by a general decrease in sediment supply (Ravnås *et al.*, 1996). During this phase, the site of major extension is believed to have shifted westwards to a master fault located at *c.* 2°30' E and to the Brage Fault (Færseth & Ravnås, unpubl. obs.). In the model, we imposed a relatively high extension rate along the Brage Fault, in order to create a basin depth comparable with that inferred from Fig. 3. However, in the absence of constraints on the exact position of timeline 14 in this basin, the basin subsidence during this stage may have been much lower and continued after the simulated period.

It appears that simulation of the stratigraphic features recognizable in the well correlation of Fig. 3 is possible by assuming changes in the rates of tectonic fault movements alone. An increase in extension rate along a fault causes footwall uplift, and as a result erosion of the sediments in the neighbouring basin. This influence can go further than one sub-basin; in the Oseberg–Brage area unconformities above the α -structure could even be explained by an increasing rotation rate of the G-structure. The basinward and, notably, updip thickening observed in the lower delta plain deposits is simulated

by activity of the western Viking Graben margin fault. The continuous character of the layer of delta plain deposits of the Tarbert Formation, thickening towards the basin, could only be simulated by imposing a period of zero extension along the three western faults of the Oseberg–Brage area. The final result is shown in Fig. 12(f) and 12(g), in which the changes in layer thickness as well as the most pronounced unconformities in the region can be recognized.

DISCUSSION AND CONCLUSIONS

The purpose of this study was to examine the effect of varying extension rates on stratigraphic features, in order to obtain a better understanding of the evolution of extensional basins, in this case the Oseberg–Brage area in the northern North Sea. However, to explain the evolution of a basin, more parameters should be taken into account, such as sea-level variations and sediment supply (e.g. Reynolds *et al.*, 1991; Hardy, 1993). We have not included these effects in our modelling, as our objective was to consider the possible effects of tectonics in isolation from other mechanisms.

We conclude that extensional tectonic forces in themselves can cause stratigraphic features such as unconformities, condensation, and onlap and offlap. The alternating activation of different faults causes a complex stratigraphic signature, especially as movement along one fault influences not only the most adjacent basins but also basins at a larger distance, depending on the flexural wavelength and thus on the flexural rigidity of the lithosphere. In the Oseberg–Brage area a sudden movement along the G–B fault may explain unconformities above the B-, γ - and α -structures.

Variations in the rate of extensional faulting provide a plausible explanation for the interpreted regressive–transgressive cycles in the late Bajocian–Bathonian uppermost Ness Formation, Tarbert Formation and lower Heather member in the Oseberg–Brage area (Ravnås *et al.*, 1996). These regressive–transgressive landward-stepping shoreline prisms, each being observed in a stratigraphic interval less than 100 m thick, and thought to be deposited in a period of less than 2 Myr, are comparable with the third-order regressive–transgressive cycles of Vail *et al.* (1977).

The coupling of local and regional scales is a key element in the explanation of the stratigraphy of the Oseberg–Brage area. Evidently, this is true for other regions as well: small areas should not be seen as isolated regions but are always part of a larger system, connected by deep and wide structures that influence broader regions, or by the flexural strength of the lithosphere. Therefore, in basin modelling, the scales on which regional tectonics and local, fault-controlled crustal movements operate should be further integrated.

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REFERENCES

- BADLEY, M. E., EGEBERG, T. & NIPEN, O. (1984) Development of rift basins illustrated by the structural evolution of the Oseberg features, Block 30/6, offshore Norway. *J. Geol. Soc. Lond.*, **141**, 639–649.
- BADLEY, M. E., PRICE, J. D., RAMBECH DAHL, C. & AGDESTEN, T. (1988) The structural evolution of the northern Viking Graben and its bearing upon extensional modes of basin formation. *J. Geol. Soc. Lond.*, **145**, 455–472.
- BEACH, A., BIRD, T. & GIBBS, A. (1987) Extensional tectonics and crustal structure: deep seismic reflection data from the northern North Sea Viking Graben. In: *Continental Extensional Tectonics* (Ed. by M. P. Coward, J. F. Dewey and P. L. Hancock), *Spec. Publ. Geol. Soc., Lond.*, **28**, 467–476.
- BERTOTTI, G., PICOTTI, V., BERNOULLI, D. & CASTELLARIN, A. (1993) From rifting to drifting: tectonic evolution of the South Alpine upper crust from the Triassic to the Early Cretaceous. *Sediment. Geol.*, **86**, 53–76.
- BLAIR, T. C. (1987) Tectonic and hydrologic controls on cyclic alluvial fan, fluvial, and lacustrine rift-basin sedimentation, Jurassic–lowermost Cretaceous Todos Santos Formation, Chiapas, Mexico. *J. sedim. Petrol.*, **57**, 845–862.
- BLAIR, T. C. & BILODEAU, W. L. (1988) Development of tectonic cyclothems in rift, pull-apart, and foreland basins: Sedimentary response to episodic tectonism. *Geology*, **16**, 517–520.
- BOND, G. C. & KOMINZ, M. A. (1984) Construction of tectonic subsidence curves for the early Paleozoic miogeocline, southern Canadian Rocky Mountains: implications for subsidence mechanisms, age of breakup, and crustal thinning. *Geol. Soc. Am. Bull.*, **95**, 155–173.
- BOSWORTH, W. (1992) Mesozoic and early Tertiary rift tectonics in east Africa. *Tectonophysics*, **209**, 115–137.
- CHRISTIANSSON, P., BERGE, A. M. & FALÉIDE, J. I. (1997) An integrated geophysical study of the crustal structure across the northern North Sea. *Spec. Publ. Geol. Soc., Lond.*, in press.
- CLOETINGH, S. (1986) Intraplate stress: a new tectonic mechanism for fluctuations in relative sealevel. *Geology*, **14**, 617–620.
- CLOETINGH, S., SASSI, W. & TASK FORCE TEAM (1994) The origin of sedimentary basins: a status report from the task force of the International Lithosphere Program. *Mar. Petrol. Geol.*, **11**, 659–683.
- CROWLEY, T. J. (1983) The geological record of climatic age. *Rev. geophys. Space Phys.*, **21**, 828–877.
- FROSTICK, L. E. & STEEL, R. J. (1993) *Tectonic Controls and Signatures in Sedimentary Successions*. *Spec. Publ. Int. Ass. Sediment.*, **20**. Blackwell Scientific Publications.
- GABRIELSEN, R. H., FÆRSETH, R. B., STEEL, R. J., IDIL, S. & KLOVJAN, O. S. (1990) Architectural styles of basin fill in the

Fig. 12. (a)–(f) Simulated development of the Oseberg–Brage area with time, from the Bathonian to the Callovian. Vertical exaggeration = 32.8 (f) Result of numerical simulation of the development of the Oseberg area from the Bathonian to the Callovian. Upflank unconformities, changes in layer thickness and condensation can be explained by the model. Vertical exaggeration = 32.8. (g) Close-up of the unconformities on the α -structure.

- northern Viking Graben. In: *Tectonic Evolution of the North Sea Rifts* (Ed. by D. J. Blundell and A. D. Gibbs), pp. 158–179. Oxford Science Publications.
- GABRIELSEN, R. H. & STRANDENES, S. (1994) Dynamic basin development – a complete scientific tool for basin analysis. In: *Proceedings of the 14th World Petroleum Congress*, pp. 13–21. J. Wiley & Sons.
- GILTNER, J. P. (1987) Application of extensional models to the northern Viking Graben. *Norsk Geologisk Tidsskrift*, **67**, 339–352.
- GLOPPEN, T. G. & STEEL, R. J. (1981) The deposits, internal structure and geometry in six alluvial fan-delta bodies (Devonian, Norway): a study in the significance of bedding sequences in conglomerates. In: *Recent and Ancient Non-Marine Depositional Environments: Models for Exploration* (Ed. by F. G. Ethridge and R. Flores), *Spec. Publ. Soc. Econ. Paleont. Miner.*, **31**, 49–69.
- GORDON, I. & HELLER, P. L. (1993) Evaluating major controls on basinal stratigraphy, Pine Valley, Nevada: Implications for syntectonic deposition. *Geol. Soc. Am. Bull.*, **105**, 47–55.
- HAMBLIN, A. P. & RUST, B. R. (1989) Tectono-sedimentary analyses of alternate-polarity half-graben basin-fill successions: Late Devonian–Early Carboniferous Horton Group, Cape Breton Island, Nova Scotia. *Basin Res.*, **2**, 239–255.

- HARDY, S. (1993) Numerical modelling of the response to variable stretching rate of a domino fault block system. *Mar. Petrol. Geol.*, **10**, 145–152.
- HARLAND, W. B., ARMSTRONG, R. L., COX, A. V., CRAIG, L. E., SMITH, A. G. & SMITH, D. G. (1990) *A Geologic Timescale*. Cambridge University Press, Cambridge.
- HELLAND-HANSEN, W., ASHTON, M., LOMO, L. & STEEL, R. (1992) Advance and retreat of the Brent delta: recent contributions to the depositional model. In: *Geology of the Brent Group* (Ed. by A. C. Morton, R. S. Haszeldine, M. R. Giles and S. Broom, S.), *Spec. Publ. Geol. Soc. Lond.*, **61**, 109–127.
- JARVIS, G. T. & MCKENZIE, D. P. (1980) Sedimentary basin formation with finite extension rates. *Earth Planet. Sci. Lett.*, **48**, 42–52.
- KUSZNIR, N. J., MARSDEN, G. & EGAN, S. S. (1991) A flexural-cantilever simple shear/pure shear model of continental lithosphere extension: applications to the Jeanne d'Arc Basin, Grand Banks and Viking Graben, North Sea. In: *The Geometry of Normal Faults* (Ed. by A. M. Roberts, G. Yielding and B. Freeman), *Spec. Publ. Geol. Soc. Lond.*, **56**, 41–60.
- LAMBIASE, J. J. (1990) A model for tectonic control of lacustrine stratigraphic sequences in continental rift basins. In: *Lacustrine Basin Exploration: Case Studies and Modern Analogs* (Ed. by J. Katz), *Am. Ass. Petrol. Geol. Mem.*, **50**, 265–276.

- MARSDEN, G., YIELDING, G., ROBERTS, A. M. & KUSZNIR, N. J. (1990) Application of a flexural cantilever simple-shear/pure shear model of continental lithosphere extension to the formation of the northern North Sea basin. In: *Tectonic Evolution of the North Sea Rifts* (Ed. by D. J. Blundell and A. D. Gibbs), pp. 241–256. Oxford Science Publications.
- MITCHENER, B. C., LAWRENCE, D. A., PARTINGTON, M. A., BOWMAN, M. B. J. & GLUYAS, J. (1992) Brent Group: sequence stratigraphy and regional implications. In: *Geology of the Brent Group* (Ed. by A. C. Morton, R. S. Haszeldine, M. R. Giles and S. Broom, S.), *Spec. Publ. Geol. Soc. Lond.*, **61**, 45–80.
- NOTTVEDT, A., GABRIELSEN, R. H. & STEEL, R. J. (1995) Tectonostratigraphy and sedimentary architecture of rift basins, with reference to the northern North Sea. *Mar. Petrol. Geol.*, **12**, 881–901.
- OLSEN, P. E. (1986) A 40 million year lake record of early Mesozoic orbital climatic forcing. *Science*, **234**, 842–848.
- PROSSER, S. (1993) Rift-related linked depositional systems and their seismic expression. In: *Tectonics and Seismic Sequence Stratigraphy* (Ed. by G. Williams and A. Dobb), *Spec. Publ. Geol. Soc. Lond.*, **71**, 35–66.
- RAVNÅS, R. & BONDEVIK, K. (1997) Architecture and controls on Bathonian–Kimmeridgian shallow-marine synrift wedges of the Oseberg–Brage area, northern North Sea. *Basin Res.*, in press.
- RAVNÅS, R., BONDEVIK, K., HELLAND-HANSEN, W., LOMO, L., RYSETH, A. & STEEL, R. J. (1996) Sedimentation history as an indicator of rift initiation and development: the late Bajocian–Bathonian evolution of the Oseberg–Brage area, northern North Sea. *Norsk Geologisk Tidsskrift*, in press.
- REYNOLDS, D. J., STECKLER, M. S. & COAKLEY, B. J. (1991) The role of sediment load in sequence stratigraphy: the influence of flexural isostasy and compaction. *J. Geophys. Res.*, **96**, 6931–6949.
- ROBERTS, A. M., YIELDING, G. & BADLEY, M. E. (1993) Tectonic and bathymetric controls on stratigraphic sequences with evolving half-graben. In: *Tectonics and Seismic Sequence Stratigraphy* (Ed. by G. D. Williams and A. Dobb), *Spec. Publ. Geol. Soc. Lond.*, **71**, 87–121.
- ROSENDAHL, B. R. (1987) Architecture of continental rifts with special reference to East Africa. *Am. Rev. Earth Planet. Sci.*, **15**, 445–503.
- SCHLAGER, W. (1993) Accommodation and supply – a dual control on stratigraphic sequences. *Sediment. Geol.*, **86**, 111–136.
- SCLATER, J. G. & CHRISTIE, P. A. F. (1980) Continental stretching: an explanation of the post-mid-Cretaceous subsidence of the central North Sea basin. *J. Geophys. Res.*, **85**, 3711–3739.
- STEEL, R. J. (1988) Coarsening-upward and skewed fan bodies: symptoms of strike-slip and transfer fault movement in sedimentary basins. In: *Fan Deltas: Sedimentology and Tectonic Settings* (Ed. by W. Nemec and R. J. Steel), pp. 75–83. Blackie, Glasgow.
- VAIL, P., MITCHUM, JR, R. M. & THOMPSON, S. (1977) Global cycles of relative changes of sealevel. *Am. Ass. Petrol. Geol. Mem.*, **26**, 83–97.
- VOLLSET, J. & DORÉ, A. (1985) *A revised Triassic and Jurassic lithostratigraphic nomenclature for the North Sea*. *NPD Bulletin*, **3**, Oljedirektorat, Stavanger.
- VAN WAGONER, J. C., MITCHUM, R. M., CAMPION, K. M. & RAHMANIAN, V. D. (1990) Siliciclastic Sequence Stratigraphy in Well Logs, Cores and Outcrops. *Am. Ass. Petrol. Geol. Meth. Expl. Series*, **7**.
- TER VOORDE, M. & BERTOTTI, G. (1994) Thermal effects of normal faulting during rifted basin formation. 1: A finite difference model. *Tectonophysics*, **240**, 133–144.
- TER VOORDE, M. & CLOETINGH, S. (1995) Numerical modelling of extension in faulted crust: effects of localized and regional deformation on basin stratigraphy. In: *Modern Developments in Structural Interpretation, Validating and Modelling* (Ed. by P. Buchanan and D. Nieuwland), *Spec. Publ. Geol. Soc. Lond.*, **99**, 283–296.
- YIELDING, G. (1990) Footwall uplift associated with Late Jurassic normal faulting in the northern North Sea. *J. Geol. Soc. Lond.*, **147**, 219–222.
- YIELDING, G., BADLEY, M. E. & ROBERTS, A. M. (1992) The structural evolution of the Brent Province. In: *Geology of the Brent Group* (Ed. by A. C. Morton, R. S. Haszeldine, M. R. Giles and S. Broom, S.), *Spec. Publ. Geol. Soc. Lond.*, **61**, 27–40.

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