Pre-breakup extension in the northern North Sea defined by complex strain partitioning and heterogeneous extension rates

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Key Points:

- Regionally extensive subsurface data are used to quantify basin-wide strain behaviour during early stages of continental rifting
- Variable magnitude and rate of extension-related strain affect the structural development of upper-crustal fault systems and host array
- Three-dimensional strain behaviour during initial continental rift phases might be more complex than previously assumed
Abstract

The early stages of continental rifting are accommodated by the growth of upper-crustal normal fault systems that are distributed relatively evenly across the rift width. Numerous fault systems define fault arrays, the kinematics of which are poorly understood due to a lack of regional studies drawing on high-quality subsurface data. Here we investigate the long-term (~150 Myr) growth of a rift-related fault array in the East Shetland Basin, northern North Sea, using a regionally extensive subsurface dataset comprising 2D and 3D seismic reflection surveys and 107 boreholes. We show that rift-related strain during the pre-Triassic to-Middle Triassic was originally distributed across several sub-basins. The Middle-to-Late Triassic saw a decrease in extension rate (~14 m/Myr) as strain localized in the western part of the basin. Early Jurassic strain initially migrated eastwards, before becoming more diffuse during the main, Middle-to-Late Jurassic rift phase. The highest extension rates (~89 m/Myr) corresponded with the main rift event in the East Shetland Basin, before focusing of strain within the rift axis and ultimate abandonment of the East Shetland Basin in the Early Cretaceous. We also demonstrate marked spatial variations in timing and magnitude of slip along-strike of major fault systems during this protracted rift event. Our results imply that strain migration patterns and extension rates during the initial, pre-breakup phase of continental rifting may be more complex than previously thought; this reflects temporal and spatial changes in both thermal and mechanical properties of the lithosphere, in addition to varying extension rates.

Key words: East Shetland Basin, continental rift, strain behaviour, extension rate, normal fault array, North Sea

1. Introduction

Continental rifting is accommodated by the growth of upper-crustal (i.e. top 5-10 km of the crustal structure) normal faults. Resolving the dynamics of continental rifting is important because normal faults control rift geomorphology and landscape development in time and space, and the erosion, transport and storage of sediment (Gawthorpe & Leeder, 2000). Our current understanding of continental rift dynamics is largely based on studies focused on examples that have proceeded to full plate rupture and continental break-up (e.g., Gibbs, 1984; Brun, 1999; Ziegler & Cloetingh, 2004; Huismans & Beaumont, 2007; Nagel & Buck, 2007; Péron-Pinvidic et al., 2013), supplemented by those concentrating on failed rifts. The latter tend to focus on specific aspects or time periods of the rifting process, such as local and regional migration of extension-related strain (e.g., Behn et al., 2002; Cowie et al., 2005; Corti et al., 2013; Bell et al., 2014; Naliboff & Buiter, 2015), the influence of crustal composition and (pre-existing) structures on fault and rift geometry (e.g., Paton, 2006; Whipp et al., 2014; Duffy et al., 2015; Phillips et al., 2016; Henstra et al., 2019), and/or the effect of the initial lithospheric conditions (e.g., crustal thickness and thermal state) on rift development (e.g., Buck, 1991, 2006; Odinsen et al., 2000; Corti et al., 2003).

The way in which strain is accumulated during lithospheric stretching (e.g., varying magnitude and rates), and how this relates to the overall geometry of the resultant rift, has also been extensively studied (e.g., England, 1983, Kuznir & Park, 1987; Bassi, 1995, Behn et al., 2002; Van Wijk & Cloetingh, 2002; Naliboff et al., 2017). Numerical and physical models of rift development, which simulate the formation of upper-crustal deformation, do not, however, commonly consider how strain behaves in three dimensions. This often reflects
the limited spatial and temporal resolution of such models, which allows them to only predict
the patterns of strain migration in two dimensions (e.g., towards or away from the rift axis)
(e.g., McClay, 1990; Cowie et al., 2000; Huismans et al., 2001; Behn et al., 2002; Ziegler &
Cloetingh, 2004; Nagel & Buck, 2007; Naliboff et al., 2017). However, observations from
individual faults or fault systems (i.e. a kinematically linked group of fault segments that are
several km to tens of km long) suggest that the overall accumulation of rift-related strain can
be rather complex in three dimensions due to, for example, fault segment interaction and/or
the composition and structure of the upper-crust (e.g. Cowie et al., 2000; Walsh et al., 2003;
Soliva et al., 2006; Putz-Perrier & Sanderson, 2008; Nixon et al., 2014; Whipp et al., 2014;
Duffy et al., 2015; Jackson et al., 2017). Recent studies of relatively young (<5 Myr old),
still-active rifts (e.g., Gulf of Corinth Rift, Bell et al., 2009; Ford et al., 2013; Nixon et al.,
2016) and inactive rifts (e.g., southern South African extensional system, Paton, 2006;
northern North Sea, Claringbould et al., 2017) suggest that the initial phase of upper-crustal
stretching is distributed across a wide zone. During this early phase of continental rifting,
diffuse extension is associated with the diachronous growth of individual fault systems that
make up the larger, rift-related fault array (i.e. a kinematically linked group of fault systems
that are tens to hundred km of length, and typically cover one margin of a rift). Péron-
Pinvidic et al. (2013) argue that rift-related strain migrates during the transition from diffuse
stretching and thinning of the upper-crust, to hyperextension and mantle exhumation, a
progression that may also be linked to an increase in extension rate (e.g., Brune et al., 2016;
Naliboff et al., 2017).

The way in which strain rate controls rift geometry is closely related to the way in
which heat is generated and transferred during extension. England (1983) shows that during
extension, the lithosphere increases in strength because rift-related continental thinning will
result in the lithosphere cooling as it is brought closer to the surface. If the extension rate is
relatively slow, this so-called synrift cooling will prevent further extension, causing the locus
of maximum strain to shift laterally, allowing the rift to widen (Bassi, 1995). However, when
the extension rate is relatively fast, synrift cooling will not occur, and necking and rift
narrowing will instead take place (Kuznir & Park, 1987). This proposed relationship between
extension rate and the resulting rift pattern has since been observed in numerous 2D,
lithospheric-scale models (e.g., Van Wijk & Cloetingh, 2002; Brune et al., 2016; Naliboff et
al., 2017; Tetreault & Butier, 2018). However, we have yet to use observations from natural
rifts to document how and the timescale over which early rift-related strain is recorded by the
growth of upper crustal fault arrays. Nor have we determined how changes in bulk extension
magnitude and rate affect the temporal evolution of rift-wide strain. The increased
complexity and inferred realism of relatively recent numerical models has essentially not
been matched by an increased level of observational details to test their predictions.

Determining the geometry and growth of crustal-scale (~10,000 km²) normal fault
arrays, as opposed to individual fault systems, requires extensive, high-quality, subsurface
data. To this end, we focus on the East Shetland Basin, northern North Sea (Figure 1). The
northern North Sea represents a failed rift basin that developed in response to protracted
extension spanning ~150 Myr (Færseth, 1996). The East Shetland Basin, located on the
western margin of the North Viking Graben, contains a large fault array that is part of the
wider northern North Sea rift system (Figure 1). We use a large subsurface dataset
comprising long (~75 km), deep-imaging (~8 s TWT) regional 2D seismic reflection profiles,
multiple, merged 3D seismic surveys (covering ~10,000 km²) that image to moderate depths
(4.5-6.5 s TWT), and 107 hydrocarbon exploration and production boreholes. Our dataset
allows a relatively high-resolution (i.e. as little as ~10 Myr temporal-scale, and a few
hundred-of-metres spatial-scale) examination of: (i) the long-term (~150 Myr) migration of
rift-related strain across a large fault array (~10,000 km²), and (ii) temporal changes in the magnitude and rate of extension at the sub-basin scale, which we can then compare to rift-scale variations in these parameters. Our analysis provides an improved understanding of how rift-related strain accumulates during the initial, pre-breakup phase of continental rifting, and the effect that heterogeneous extension magnitudes and rates have on the resulting rift geometry. We also use our results to infer how the thermal and mechanical properties of the lithosphere varied through time during protracted extension. Finally, our study of a natural rift allows us to critically test the predictions of physical and numerical models of continental extension.

2. Geological setting

The East Shetland Basin is located in the northern North Sea, offshore western Norway, on the western margin of the North Viking Graben (Figure 1). The present geometry of the basin is characterized by large (>25 km length), N-S- to NE-SW-striking, east-dipping normal fault systems that bound 15-25 km wide half-grabens (Figures 1c and 2) that developed during protracted, pre-Triassic-to-Late Jurassic rifting (~150 Myr) (e.g., Ravnås et al., 2000, Claringbould et al., 2017). Based on the interpretation of regional 2D seismic reflection lines, flexural backstrapping, and tectono-stratigraphic forward modelling, two main rift phases were classically identified; Permian-Triassic and Late Jurassic (e.g., Badley et al., 1988; Lee & Hwang, 1993; Thomas & Coward, 1995; Færseth, 1996), with the magnitude of extension varying between them (e.g., Roberts et al, 1993, 1995; Odinsen et al., 2000). However, seismic-stratigraphic analysis of borehole-constrained 3D seismic reflection datasets indicate that extension and active faulting actually continued into the Early Cretaceous (140-145 Ma; Valanginian-Berriasian), with strain eventually focusing on fault systems bounding the eastern margin of the East Shetland Basin (Cowie et al., 2005; see also Færseth et al., 1995; Bellingham & White, 2000; and McLeod et al., 2002). Strain localisation on these structures was associated with overall rift narrowing and ultimately abandonment of the East Shetland Basin (Cowie et al., 2005; Phillips et al., 2019).

The increasing availability of high-quality 3D seismic reflection data has permitted a more detailed analysis of the geometry and growth of the individual fault systems in the East Shetland Basin. Even then, most studies consider time-interval that are relatively short (e.g., Late Jurassic; ~18 Myr) given that, together, the various Permian-to-Early Cretaceous rift phases or pulses spanned ~150 Myr (e.g., Strathspey-Brent-Statfjord half graben, McLeod et al., 2000, 2002; Murchison-Statfjord North Fault, Young et al., 2001; eastern East Shetland Basin, Cowie et al., 2005; Triassic Ninian and Alwyn North fields, Tomasso et al., 2008). Because they focus on relatively small areas and/or for only a relatively short part of the much longer rift episode, these studies can also only show how strain accumulates during the development of individual fault systems; the longer-term (~150 Myr) dynamics of the larger host fault array remains unknown.

In this study we develop the ideas of Ravnås et al. (2000) and Claringbould et al. (2017), who show that rifting in the northern North Sea was protracted not punctuated. Ravnås et al. (2000) propose that the northern North Sea experienced Permian-to-Early Triassic and Middle-to-Late Jurassic rift episodes that were separated by an intervening, Middle Triassic-to-Middle Jurassic inter-rift period characterized by more diffuse extension. Claringbould et al. (2017) qualitatively describe the entire pre-Triassic-to-Late Jurassic evolution of the fault array in the East Shetland Basin. They argue that although pre-existing upper-crustal structures may have locally influenced the geometry and growth subsequent
rift-related structures, the more extensive, lithosphere-scale, thermal and rheologic heterogeneities served to somewhat dilute their control on the overall rift geometry. This study builds on Claringbould et al. (2017) by quantifying the ~150 Myr evolution of the East Shetland Basin fault array during eight time-intervals that individually span ~6–45 Myr. Because our seismic reflection dataset does not image structures associated with the very latest stage of extension (i.e. Early Cretaceous; post-145 Ma), we do not explicitly consider the detailed growth of fault systems most active at this time. We do, however, place our study within the more regional, late syn-rift-to-early post-rift tectono-stratigraphic framework erected by other authors (Bellingham & White, 2000; McLeod et al., 2000; Cowie et al., 2005; Phillips et al., 2019).

3. Data and methods

3.1 Seismic reflection and well data

We use an extensive dataset comprising 2D and 3D time-migrated seismic reflection surveys that were collected between 2006 and 2012 (Figure 1b). More specifically, we use four, partly overlapping, 3D seismic “merged-surveys”, which cover almost the whole East Shetland Basin (~10,000 km²). These data image to depths of 4.5 to 6.5 s TWT (6 – 8 km), and have a 12.5 × 12.5 m or 25 × 25 m in- and crossline spacing. We also use long (~75 km length), 2D seismic profiles that trend either NNE or WNW, image to depths of ~8 s TWT (~10 km), and have a line spacing of ~5 km (Figure 1b). Seismic data quality ranges from excellent for some of the 3D surveys to moderate for some of the 2D profiles. In addition to the seismic reflection data, we use 107 hydrocarbon exploration wells to determine the age of the basin-fill, of which 82 are tied to the seismic data through the construction of synthetic seismograms (Figure 3).

3.2 Seismic interpretation and fault system analysis

We interpret nine key seismic horizons across an area of ~6800 km² (pre-Triassic to the Base Cretaceous Unconformity; Figures 2 and 3). Our primary interpretation is based on the 3D surveys given they allow us to: (i) construct a detailed 3D view of the present basin structure; (ii) compile time-thickness maps that reveal fault-driven variations in subsidence and uplift; and (iii) extract throw and stratigraphic thickness measurements at any position along the fault systems forming part of the larger, rift-related fault array. We also use 2D seismic profiles to correlate key seismic horizons between the 3D surveys. With the exception of the pre-Triassic horizons, all of these horizons are tied to the wells (Figures 1b, 2 and 3). The three pre-Triassic horizons are picked based on their continuous, high-amplitude seismic character. Patruno and Reid (2017) use well data to identify Permian-Triassic to Devonian rift basins on the East Shetland Platform, which is located a few tens-of-kilometres SW of our study area (Figure 1a); however, we cannot directly constrain the ages of the pre-Triassic reflections in the East Shetland Basin, thus we name them Pre-Triassic 1, 2, and 3 (see Claringbould et al., 2017, for a full description of how these horizons were interpreted).

Our seismic mapping allows us to constrain the growth of major fault systems; these are defined as those that are >3 km long, offset at least pre-Triassic deposits, and have >200 m (>120 ms) of throw (Figure 1c). Such fault systems accommodate the majority of the rift-related strain (e.g., Fossen, 2010). Throw data are based on horizon cut-off information
collected on fault-normal seismic profiles that are spaced every ~625 m; this amounted to
>14,000 values along 34 fault systems, which have a combined length of 535 km (Figure 4).
This spatial resolution of analysis is considered sufficient to analyse strain accumulation
across the entire fault array during eight time periods that span 6-45 Myr over a ~150 Myr
time period (Figure 3). The horizon cut-off information is depth-converted using the average
time-depth relationship derived from 79 of our 107 wells.

3.3 Fault array analysis

Expansion indices (EI) are used to constrain temporal variations in fault system
activity and basin-wide extension magnitude (e.g., Thorsen, 1963; Cartwright et al, 1998;
Bouroullec et al., 2004; Jackson & Rotevatn, 2013; Lewis et al., 2013; Reeve et al., 2015;
Jackson et al., 2017) (Figure 4). EI represents the ratio between the vertical (i.e. stratigraphic)
thickness of time-equivalent hanging wall and footwall strata (see Supplementary material).
We also use throw backstripping to determine how strain accumulates along-strike of
individual fault systems and across the fault array (e.g., Jackson et al., 2017) (see
Supplementary material). With the exception of the pre-Triassic units, we also calculate slip
rates along the individual fault systems; this allows us to link temporal variations in local slip
rate to more regional variations in strain accumulation accommodated by the larger fault
array (see Supplementary material). The fault slip rate represents the backstripped
displacement over time and is quoted in m/Myr. Because lithostratigraphic horizons do not
necessarily represent chronostratigraphic surfaces (i.e. absolute time-lines), we use the
average absolute ages of the lithostratigraphic boundaries from the wells across the East
Shetland Basin to estimate horizon ages and thus fault slip rates (Figure 3) (see
Supplementary material).

In addition to analysing 34 major fault systems, we sum strain along three transect
lines to investigate basin-scale strain trends as rifting progressed (Figure 5). We did this on
three ~NW-trending transects drawn approximately orthogonal to the analysed fault systems;
these transects covered the North, Centre, and South of the basin (Figure 5a). Where a
transect line crosses one of the analysed fault systems, we calculate the horizontal extension
for each time period (see Supplementary material). These values are then summed per time
period along each transect (North, Centre, or South) and, additionally, by region (Western,
Central or Eastern), to show how strain accumulated in time and space (Figure 5b) (see
Supplementary material). Furthermore, we calculate the magnitude of extension (i.e.
extension factor or β-factor) along the three transect lines for each time period. With the
exception of the age-unconstrained pre-Triassic units, we calculated extension rates along the
three transects to again analyse how strain accumulated in time and space (see Supplementary
material) (Figure 5c-d). Similar to the approach used to constrain fault slip rates, we used the
average absolute ages of the lithostratigraphic boundaries from the wells across the basin to
estimate the extension rates (see Supplementary material) (Figure 5d).

Figures 6-9 show how EI and backstripped throw vary along strike of four of the
largest, longest-lived fault systems that accommodated most rift-related strain (Eider, Ninian-
Hutton, Cormorant, and Osprey faults systems; Figure 10); these data illustrate how the
growth of these systems relate to the overall, basin-scale pattern of strain accumulation across
the entire fault array (Figures 4 and 5). We also undertake throw-depth (T-z) analyses at
specific points along these fault systems to assess how strain accumulated along their lengths
(Figure 10) (e.g., Jackson et al., 2017).
Our fault analysis methods are based on several assumptions and are associated with some uncertainties. First, we note that sediment compaction may cause our measurements of fault throw to be 5–15% less than their true, near-surface, pre-burial values (Taylor et al., 2008; Giba et al., 2012). However, when considering our study area, we note that the thickness of sediment overburden above the analysed fault systems is fairly constant; we therefore believe that the overall patterns of present-day throw will be represented of their near-surface, pre-burial values (cf. Whipp et al., 2014 and Reeve et al., 2015). Second, our use of time-thickness maps (isochrons) to determine temporal changes in accommodation related to fault slip assumes that accommodation associated with the rifting was completely filled with syn-kinematic deposits. In the case of underfilled basins, syn-kinematic deposits are limited to the hanging wall depocentre; this can result in an underestimation of the rate of accommodation generation and associated fault slip (e.g. Jackson et al., 2017). However, our data indicate that many of the fault-bound sub-basins comprising the East Shetland Basin were overfilled during rifting; more specifically, seismic and, critically, well data demonstrate syn-kinematic deposits are preserved (and were thus deposited) in both the footwall and hanging walls of the faults (Figure 2). Finally, we recognize that our geometric and kinematic analysis, especially at deeper (i.e. pre-Triassic) structural levels and thus for older time-periods, are likely affected by the quality of and confidence we have in our seismic interpretation and depth conversion. However, converting values from ms TWT to metres (or feet) typically preserves the spatial patterns of fault throw and does not significantly impact the related kinematic analysis (Tvedt et al., 2013). Since we focus on basin-scale trends rather than specific, absolute measurements, we consider it appropriate to use data in ms TWT, extracted directly from our time-migrated seismic data.

4. Spatial and temporal strain variations across the East Shetland Basin

Temporal shifts in sediment depocentres across the East Shetland Basin reflect growth of the rift-related fault array (Claringbould et al., 2017) (Figure 4). Here we reconstruct growth of major fault systems comprising the larger fault array, as well as calculating how rift-related strain varied through time at the basin-scale (Figures 4 and 5).

4.1 Pre-Triassic-to-Middle Triassic (>245 Ma) (Units 1 and 2, and the Teist Formation) (Figures 4a-c)

During the deposition of pre-Triassic Units 1 and 2, and the Lower-to-Middle Triassic Teist Formation, several major fault systems in the Magnus, Tern, and Ninian sub-basins were active. These systems accumulated up to 1200 m of throw, corresponding to large expansion indices of 4 to 8 (Ninian West, Heather, and Cormorant faults, Figure 4a-c). During the deposition of pre-Triassic Units 1 and 2, summed extension values are highest in the Eastern region (up to 2063 m) and along the southern transect (up to 2696 m, with an extension factor of 1.042) (Figure 5b-c). In contrast, during deposition of the overlying Lower-to-Middle Triassic Teist Formation, most extension occurred in the Western region (1032 m) and along the central transect (1369 m, with an extension factor of 1.059). The average extension rate in the Early-to-Middle Triassic was ~129 m/Myr (Figure 5b-d). During this >50 Myr time period, extensional strain was diffuse and responsible for the formation of several sub-basins.
4.2 Middle-to-Late Triassic (ca. 245-201 Ma) (Lunde and Lomvi formations) (Figure 4d)

Over the next ~40 Myr, during deposition of the Middle-to-Upper Triassic Lomvi and Lunde formations, strain focused towards the southwestern part of the fault array (Figure 4d). During the early part (ca. 245-201 Ma) of the Middle Triassic-to-Middle Jurassic ‘inter-rift’ period (ca. 245-166 Ma), an up to 800 m thick sediment depocentre developed next to the southern end of the Eider Fault System, with only moderate activity (characterised by expansion indices <4) observed on some of the larger structures further east (Figure 4d). We calculate an average extension magnitude of 604 m for the Middle-to-Upper Triassic, which correlates to an average extension rate of 14 m/Myr (Figure 5b-d); this is significantly less than that defining the previous, pre-Middle Triassic time-interval (129 m/Myr) (Figure 5b).

4.3 Latest Triassic-to-Middle Jurassic (ca. 201-166 Ma) (Statfjord Formation, and Dunlin and Brent groups) (Figure 4e-g)

A significant shift in the locus of strain accumulation occurred during deposition of the uppermost Triassic-to-Lower Jurassic Statfjord Formation (Figure 4e). During this middle part (ca. 201-192 Ma) of the inter-rift period, moderately thick (~50 and ~250 m), relatively tabular sedimentary depocentres developed in the hanging wall of major fault systems in the eastern half of the basin (e.g., Ninian, Hutton, Alwyn, and Strathspey fault systems). These fault systems accumulated up to 300 m throw at this time, which was accompanied by EI value of 1.5-4 (Figure 4e). Most of this extension (385 m) accumulated in the Eastern region during this time (Figure 5b), with the highest extension rate (30 m/Myr) occurring in the north of the basin (i.e. along the northern transect; Figure 5d). The average extension rate had also increased slightly from the previous time-interval (from 14 to 23 m/Myr). Isochrons thus imply that strain was no longer focused on a single fault system in the western part of the basin (i.e. Eider Fault System; Figure 4d), but was now widely distributed across the eastern part of the basin, being accommodated by slip on several major fault systems.

During the latter part (ca. 192-166 Ma) of the inter-rift period, moderate amounts of throw (up to 300 m, associated with expansion indices of up to 6) accumulated on major fault systems in the east of the basin; this indicates that strain continued to be focussed here for another ~25 Myr, during deposition of the relatively thin (up to 300 m) Dunlin and Brent groups (Figure 4f and g). Most extension occurred in the Eastern region (599 m and 692 m) (Figure 5b), with the largest extension factor (up to 1.034) measured along the Central transect line (Figure 5c). The average extension rate initially decreased and then increased during deposition of the Dunlin (16 m/Myr) and Brent (41 m/Myr) groups in the Early-to-Middle Jurassic (cf. 23 m/Myr during the Early Jurassic; Figure 5d).

4.4 Middle-to-Late Jurassic (ca. 166-145 Ma) (Viking Group) (Figure 4h)

In contrast to the inter-rift period (ca. 192-166 Ma), when strain was relatively focused in the east of the East Shetland Basin, strain is distributed across the whole basin during deposition of the Middle-to-Upper Jurassic Viking Group (Figure 4h). Up to 1200 m of throw accumulated on the major fault systems across the East Shetland Basin, forming thick (up to 900 m) depocentres that were associated with high expansion indices (6-8) (Figure 4h). Extension was distributed relatively evenly across the Western, Central, and Eastern regions in the basin (1849 m, 1908 m, and 1844 m, respectively) (Figure 5b). We observe an increase in both strain accumulation and extension magnitude in the East Shetland
Basin compared to the Early Jurassic. First, EI values that are locally <4 during the
deposition of the Lower Jurassic Dunlin Group (Figure 4f), increase to 6-8 across much of
the basin during deposition of the Middle-to-upper Jurassic Viking Group (Figure 4h).
Second, the average extension factor (1.025 compared to 1.021) and extension rate (89
compared to 16 m/Myr) are both significantly higher during the Middle-to-Late Jurassic
compared to the Early Jurassic (Figures 4f-h and 5c-d).

4.5 Early Cretaceous (ca. 145-140 Ma) (Cromer Knoll Group)

In this study we did not focus on the Early Cretaceous phase of extension in the East
Shetland Basin. However, several studies show that during the earliest Cretaceous (ca. 140-
145 Ma; i.e. Valanginian-Berriasian), active faulting migrated eastwards onto the fault
systems separating the East Shetland Basin from the deep rift-axis of the North Viking
Graben (so-called ‘Visund-Gullfaks fault’ of Cowie et al., 2005; see also Færseth et al. 1995,
McLeod et al. 2002, and Phillips et al. 2019). Strain localisation onto these structures, which
lie just east of the area imaged by our seismic reflection data (see Figure 1), was associated
with overall rift narrowing (from >200 km to ~50 km) and ultimately abandonment of the
East Shetland Basin (Cowie et al., 2005). To the best of our knowledge, extension factor and
magnitude have not been calculated for this specific period of late syn-rift strain localisation.
However, subsidence inversion (Newman & White, 1999) and geological observations
(Cowie et al., 2005) suggest that strain rate declined rapid (from 3 x 10^{-16} to 3 x 10^{-17} s^{-1})
from the Late Jurassic into the Early Cretaceous. Final abandonment of the North Viking
Graben (and the northern North Sea rift system in general) occurred later in the Cretaceous
(Phillip et al., 2019).

5. Relationship between fault system and fault array growth

During deposition of pre-Triassic and earliest Triassic, strain was distributed across
the basin, with most strain accommodated along the south transect line (see section 4.1). With
the exception of the Osprey Fault System (Figure 9a-c), which was located in the centre of
the basin (Figure 10), the presence of multiple throw and EI maxima along all major fault
systems during deposition of Unit 1 suggest these structures grew by the growth and linkage
of initially isolated fault segments (Figures 6a-c and 8a-c). Post-linkage, strain could have
migrated along-strike, as illustrated by the Eider Fault System, which saw an overall
southwestwards migration of activity through time (Figure 6a-c).

During the early part (ca. 245-201 Ma) of the inter-rift period, strain was primarily
focussed in the Western region, along the Eider Fault System. At the southwestern tip of the
Eider Fault System, up to 900 m throw accumulated at this time, decreasing to ~200 m along
strike towards the northeastern fault tip (Figure 6d). Relatively little strain was
accommodated along the Cormorant and Osprey fault systems in the basin centre at this time
(<500 m throw), with only some small segments of these structures being active (Figure 8d
and 9d). The Ninian-Hutton Fault System, which was located in the eastern part of the basin,
was inactive (Figure 7d; see also T-z plots in Figure 10b-c and e). We also observe an overall
decrease in slip rate from the previous, pre-Middle Triassic time-interval (from a maximum
of ~100 m/Myr to ~25 m/Myr) (Figure 11a-b); however, because the age of the lower
boundary of the Teist Formation is poorly constrained, these rates could be overestimated.
During the middle and latter part (ca. 201-166 Ma) of the inter-rift period, strain migrated from the western to eastern part of the basin. With the exception of its northeastern tip, the Eider Fault System was largely inactive, illustrated by vertical intervals on the corresponding T-z plots (Figure 10d). Strain was distributed relatively evenly along the length of the Cormorant, Osprey, and Ninian-Hutton fault systems during deposition of the Statfjord Formation (Figures 7e and 9e). Latest Triassic-to-Middle Jurassic slip rates increase from 20-30 m/Myr during deposition of the Statfjord Formation (Figure 11c) to 25-75 m/Myr during deposition of the Brent Group (Figure 11d-e). Reactivation of the Ninian-Hutton Fault System during deposition of the Statfjord Formation is clearly captured in the T-z plots. For example, between Pre-Triassic 1 and 2, activity occurred along two segments (i.e. 0---30 km and ~45--55 km; Figure 10e). Periods of fault inactivity along the entire length of the Ninian-Hutton fault are marked on T-z plots by vertical intervals, the tops of which defined by the Top Lunde horizon; above this, throw decreases, indicating fault reactivation (Figure 10e).

During the Middle-to-Late Jurassic, strain was distributed across the basin and accommodated by relatively rapid slip (up to 100 m/Myr; cf. Cowie et al., 2005) on many of the major fault systems (Figure 11). However, a key observation is that strain was not distributed evenly along these fault systems. For example, local throw minima that are associated with and thus define the position of now-breached relays, and which were inherited from earlier, possibly even pre-Triassic stage of rifting and fault linkage, persisted (Figure 6h).

Lastly, during the Early Cretaceous, strain focused on the Visund-Gullfaks fault (Fig. 1) as the rift narrowed (McLeod et al., 2002; Cowie et al., 2005). Despite an overall decrease in strain rate (Newman & White, 1999), maximum slip rates on the Visund-Gullfaks fault were apparently the highest that had ever occurred in the East Shetland Basin (~300 m/Myr) (Cowie et al., 2005).

6. Discussion

6.1. Temporal and spatial changes in the basin-scale distribution of rift-related strain

Despite rifting being rather protracted (~150 Myr), the northern North Sea region experienced only the early phases of continental rifting (i.e. the stretching and the onset of the thinning phase of Péron-Pinvidic et al., 2013); full plate rupture was not achieved. The East Shetland Basin forms part of the so-called proximal domain; this is a domain characterized by classical graben and half-graben basins filled with wedge-shaped, syntectonic sedimentary units mainly deposited during the initial stretching phase (Péron-Pinvidic et al., 2013). We propose that rift-related strain was partitioned in different parts of the basin and migrated through time. Figure 12 illustrates the qualitative strain distribution across the fault array in the East Shetland Basin from the pre-Triassic-to-Late Jurassic, based on our detailed, quantitative fault array analyses. Spatial variations in the timing and magnitude of slip occurred along-strike of major fault systems that make up the larger host-fault array. This reflects the heterogeneous nature of the early rift-related strain within the East Shetland Basin (Figure 12).

Strain migration along individual fault systems is common, typically reflecting fault growth by segment linkage, rheological differences in the deforming host rock, and/or the presence of pre-existing structures (e.g., Cowie et al., 2000; McLeod et al., 2000, 2002; Young et al., 2001; Walsh et al., 2003; Soliva et al., 2006; Putz-Perrier & Sanderson, 2008; Tomasso et al., 2008; Nixon et al., 2014; Whipp et al., 2014; Duffy et al., 2015; Jackson et
...Segment linkage and pre-existing structures played a role in the growth of several fault systems (e.g., segment linkage at Eider Fault System, Figure 6a-b; see also McLeod et al., 2002). However, we see no clear evidence that pre-existing structures dictated temporal variations in rift-related strain in the East Shetland Basin. This observation is consistent with Cowie et al. (2005) and Claringbould et al. (2017), who both propose that strain accumulation patterns during growth of upper-crustal normal fault systems, even during the relatively early stages of continental rifting, are controlled by the thermal and mechanical state of the entire lithosphere.

Complex strain migration patterns during the early phases of continental rifting, such as those identified in the East Shetland Basin, could be caused by the emplacement of magmatic bodies (e.g., Corti et al., 2003; Wolfenden et al., 2005; Buck, 2006; Stab et al., 2016). However, in the East Shetland Basin we see no clear evidence for significant rift-related magmatism, suggesting the emplacement of igneous bodies did not control how rift-related strain accumulated. Other studies link strain migration to flexural downbending of the crust (e.g., Bayona & Thomas, 2003; Bell et al., 2014). Indeed, on the eastern margin of the northern North Sea, Bell et al. (2014) observe that the strain migrates away from the Middle- to Early Cretaceous rift axis (North Viking Graben, Figure 1a), after a phase of Permian-Triassic rifting and Early Jurassic tectonic quiescence. However, flexural downbending of the upper-crust is typified by the overall migration of strain either towards or away from the principle rift axis, making it unlikely that this is the cause for the far more complex patterns observed in the East Shetland Basin (Figure 12). Cowie et al., (2005) show that Middle-to- Early Cretaceous rift-related strain in the East Shetland Basin migrated towards the rift axis during the rift maximum, relating this to a change in the geometry of the underlying thermal perturbation associated with the initial phase of rift narrowing (i.e. an increase of vertical thermal gradient towards the rift axis; e.g., Huismans et al., 2001; Behn et al., 2002; Nagel & Buck, 2007). However, their study is spatially restricted to the eastern part of the East Shetland Basin and considered only Middle-to-Early Cretaceous rifting. By considering the entire basin, which essentially represents the entire western margin of the northern North Sea rift system, and the full, ~150 Myr duration of rift activity, we show a much more complicated history comprising temporal and spatial changes in both strain distribution (Figure 12), and extension magnitude and rate (Figures 4, 5d, and 11). Our study thus highlights that a full understanding of the early stages of continental breakup requires analysis of a sufficiently large study area (at least one margin of the rift system), and needs to consider a sufficiently long period of rift development.

6.2. Variation in extension magnitude and rate during rifting

We show that extension magnitudes and rates in the East Shetland Basin vary in space and time (Figure 5). For example, extension and fault slip rates decrease and stay relatively low (≤30 m/Myr) for ~70 Myr during the Middle Triassic-to-Middle Jurassic inter-rift period (Figure 5d and 11a-c). From the Middle Jurassic onwards, slip rates increase for ~30 Myr (Figure 5d) to ≥50 m/Myr (Figure 11d-f). Maximum slip rates of ~300 m/Myr eventually occur during the Early Cretaceous on very large fault systems bounding the East Shetland Basin (Cowie et al., 2005).

We propose that changing extension rates may account for the strain distribution trends we observe, consistent with the predictions of lithospheric-scale numerical models (e.g., England, 1983; Houseman & England 1986; Kuznir & Park, 1987; Bassi, 1995; Van Wijk & Cloetingh, 2002; Péron-Pinvidic et al., 2013; Naliboff & Buiter, 2015; Brune et al.,...
2016; Naliboff et al., 2017). The distributed faulting that defined the pre-Triassic period (Figure 12a-c) may reflect the relatively slow, pre-Jurassic extension rates. This is consistent with the 2D lithosphere-scale numerical modelling results of Naliboff et al. (2017), who suggest that a relatively slow (<5000 m/Myr) extension rate during the initial stage of rifting is associated with uniform lithospheric thinning and distributed upper-crustal faulting.

Subsequently, Triassic strain focussed on a small number of fault systems, while elsewhere in the basin minimal to no fault growth activity took place for ~45 Myr. We suggest that, during this inter-rift period, the relatively slow and decreasing extension rate induced local synrift cooling and was associated with limited fault activity (Figure 5d, 11c and 12d-e) (e.g., England, 1983; Kuznir & Park, 1987; Bassi, 1995; Van Wijk & Cloetingh, 2002; Naliboff & Buiter, 2015). This interpretation is consistent with the predictions of previous 2D lithosphere-scale models (e.g., Van Wijk & Cloetingh, 2002; Naliboff et al., 2017). Van Wijk and Cloetingh (2002) model synrift cooling in a region that initially extended at a relatively slow rate (<8000 m/Myr). Subsequently, a shift in the locus of maximum extension occurs: the “old” rifted sub-basin is abandoned, and extension concentrates in other areas of the larger rift system. Van Wijk and Cloetingh (2002) compare their modelling results to several continental margins, including the Mid-Norwegian Margin, which is of comparable size and has a similar extension history to the East Shetland Basin. They find that strain migration patterns between their slow extension rate models (<8000 m/Myr) and natural example are similar; i.e. the gap between successive rifting events, during which time the locus of strain migrates, is of a similar magnitude (~20-60 Myr).

In contrast to initially slow and decreasing extension rates, we suggest that immediately post-Triassic patterns of faulting in the East Shetland Basin are controlled by the increasing rate of lithospheric extension (Figure 5d). We propose that eastwards migration of strain during the latest-Triassic-to-Early Jurassic reflect the initial phase of lithospheric necking and rift narrowing (e.g., Huismans et al., 2001; Behn et al., 2002; Cowie et al., 2005; Nagel & Buck, 2007; Péron-Pinvidic et al., 2013) (Figure 12e-f). The main Middle Jurassic-to-Early Cretaceous rift phase is ultimately characterized by the highest extension and fault slip rates (Figures 5d and 11f). This phase of rifting is also defined by distributed faulting, which involves the reactivation of some pre-Jurassic faults, as well as the growth of new faults (Figure 12h). Our observations are consistent with the numerical model predictions of Naliboff et al. (2017) who show that when extension rate increases (>5000 m/Myr), strain localises near a heated and weakened rift-axis as the advective heating of the lithosphere exceeds the conductive cooling; this can drive rift narrowing. Naliboff et al. (2017) also predict that when the extension rate increases after an initial period of relatively slow extension (<5000 m/Myr), the upper-crustal rift pattern is characterized by the growth of new faults and the reactivation of the earlier developed, more widely distributed normal faults. Corti et al. (2013) also show that a relatively high rate of extension is associated with an overall inward migration of faulting towards the rift axis. However, their lithosphere-scale, centrifuge sand-box experiments imply that inward migration of faults during rifting are also subject to other factors such as the thickness of brittle and ductile layers, the width of the weak zone that localizes extension, and the degree of rift obliquity (Corti et al., 2013).

6.3 Comparing extension magnitudes and rates in relation to rift pattern evolution

Depending on which numerical models is used, the absolute velocity threshold for which synrift cooling will or will not occur ranges between 1500 and 8000 m/Myr, (e.g., Bassi, 1995; Van Wijk & Cloetingh, 2002; Naliboff et al., 2017). This likely reflects variations in the initial conditions used by the different models, given that Bassi (1995)
shows that this transition velocity is highly dependent on the rheology of the rifted
lithosphere (see also Bassi, 1991; Buck, 1991). It is therefore difficult to directly compare
different models or natural rift systems, or models to natural examples.

We note however a marked discrepancy between the extension rates we calculate in
the East Shetland Basin (10-225 m/Myr) and those determined in active rift systems by
tectono-stratigraphic forward models to calculate the extension factor across the East
Shetland Basin for the Permian-Triassic (1.15) and Jurassic (1.15) rift phases. Using a similar
method, Odinsen et al. (2000) suggest slightly different values (i.e. 1.29 and 1.11 for the
Permian-Triassic and Jurassic phases, respectively). However, in contrast to these and other
workers, we argue for a single, protracted phase of rifting; i.e. we do not identify two,
discrete periods of rifting separated by a phase of tectonic quiescence. For the entire Pre-
Triassic-to-Jurassic period, we calculate an average extension factor of 1.11 along three
transects across the East Shetland Basin (Figure 5c). This slight difference between our
extension factors and those previously calculated (i.e. Roberts et al., 1993, 1995, and Odinsen
et al., 2000) likely arises due to the different time-intervals considered, and the different
extents and locations of the fault-normal transects used to calculate the extension factors.

6.4 Extension rate variability during rifting

Despite the absolute discrepancy between our and previously observed and predicted
extension and strain rates (section 6.3), our results are qualitatively consistent with the
predicted effect of changes in extension and strain rates on the rift pattern evolution (e.g.,
Bassi, 1995; Van Wijk & Cloetingh, 2002; Naliboff et al., 2017). We show that extension
rate decreases during the Triassic and increases throughout the Jurassic in the East Shetland
Basin (Figure 5d). We suggest that these changes are responsible for the observed patterns of
rift-related faulting and overall rift geometry (Figure 12). Although difference in lithospheric
characteristics between natural rift systems (e.g., rheology) complicate a direct comparison of
extension rates, and its resultant effect on rift geometry, changes in relative extension rate
during rifting have been observed in natural rift systems elsewhere (e.g., Ford et al., 2013;
Brune et al., 2016). Based on plate reconstructions, Brune et al. (2016), show that an abrupt
acceleration in extension rate ~10 Myr before break-up is apparent in the South Atlantic,
Central North Atlantic, North America-Iberia, and Australia-Antarctica rifts, as well as
during opening of the and South China Sea. Brune et al. (2016) argue that this is the result of dynamic rift weakening; i.e. as long as the rift is strong, the extension rate is low, but with continued deformation the rift axis weakens, and extension accelerates due to crustal necking and strain softening. Ford et al. (2013) use field data to calculate a significant increase in extension rates during the development of the young (<5 Myr), still-active Corinth Rift, whereas Nixon et al. (2016) use field and subsurface data to illustrate a relatively rapid (i.e. over a 300 kyr period) transition from a structurally complex, northward migrating rift to a predominantly asymmetric rift. They argue that this rapid change in rift structure over a relatively short period of extension can reflect multiple parameters, including an increase in extension rate (Corti et al., 2013).

Our results suggest that relative changes in extension rate play an important role in the basin-wide strain behaviour we observe in the East Shetland Basin (and the northern North Sea in general) during pre-Triassic-to-Late Jurassic rifting. We propose that lack of a clear direction for strain migration, especially during pre-Jurassic extension, shows that the early stages of continental rifting is complex due to a range of an underlying controlling factors such as variation in extension rate, evolving geometry of underlying thermal perturbation, and the influence of faults developed during the initial stage of rifting. It is possible that the limited spatial and temporal dimensions used by the previous studies in the northern North Sea meant details of this heterogeneous strain distribution and complex rift pattern evolution were missed (e.g., Badley, et al., 1988; Lee & Hwang, 1993; Roberts et al., 1993, 1995; Thomas & Coward, 1995; Færseth, 1996; Odinsen et al., 2000; Cowie et al., 2005; Tomasso et al., 2008; Bell et al., 2014). Therefore, high-resolution observations and analyses across at least a fault array, and over a considerable period of the rift event, are necessary to fully resolve the dynamics of continental rift development. Moreover, these details of three-dimensional strain behaviour during rift-related extension and its effect on the rift pattern evolution should be considered in future numerical and physical models.

7. Conclusions

Using an extensive, high-resolution subsurface dataset, we observe complex strain partitioning and varying extension rates during the ~150 Myr rift development of the East Shetland Basin, northern North Sea. Comprehensive quantitative fault growth analyses across the entire width of the basin enable us to document the development of a fault array on one margin of a failed rift system and analyse the related strain accumulation pattern over time (Figure 12). Our results highlight the complicated three-dimensional behaviour of strain in the upper-crust during the early stages of continental rifting.

For extended periods of time (>20 Myr) we find that strain is distributed across the full width of the basin where it accumulates and localizes at different parts, while during other time-intervals we observe minimal to no fault growth (Figure 12). Furthermore, we calculate varying extension magnitudes and rates across the basin over time. Average extension factor ranges between 1.020 and 1.034, and average extension rates range between 14 and 129 m/Myr (Figure 5c-d). This variation marks different time-intervals of relatively low and high rift activity during rifting in the East Shetland Basin. Fault segment linkage and prior rift structures affect the localization of strain within major fault systems (Figure 6a-b), however it is unlikely that these dictate strain behaviour across the larger fault array. Instead our results suggest that during the early stages of rifting changes in extension rate have significant control on strain behaviour. We argue that relatively low and decreasing extension rates (14 m/Myr) lead to an inter-rift period that is characterized by distributed faulting and
local synrift-cooling (Pre-Triassic-to-Late Triassic) (Figure 12a-d). Relatively high and increasing extension rates (from 16 m/Myr, Early Jurassic, to 89 m/Myr, Middle-to-Late Jurassic, Figure 5d) lead to a heterogeneous strain distribution and, in the case of the East Shetland Basin, the gradual transition from lithospheric stretching to thinning, and ultimately to rift narrowing (Figure 12e-h). Our results are qualitatively consistent with the previous results from natural rifts and predictions of rift models that investigate the effect extension rate on the rift pattern development.

This study illustrates the importance of the detailed analyses using regionally extensive, high-resolution 3D subsurface data over a considerable period of basin development, which results provide observations that can be compared with numerical rift analogues. Studies that propose a simple or multiphase rift evolution with a homogeneous strain distribution or directional strain migration pattern based on less extensive analyses across the full extent of the basin possibly overlook fault array development and local strain accumulations, especially during periods of relatively less rift activity. Heterogeneous three-dimensional strain behaviour during the initial phases of continental rifting as a result of varying extension rate and magnitude are not typically generated in simple rift models, yet can be a significant aspect of rift dynamics.
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The data used for this study are publically available for download via the UK National Data Repository (NDR) (https://ndr.ogauthority.co.uk) for the United Kingdom side, and the DISKOS online portal (Diskos) (https://portal.diskos.cgg.com) for the Norwegian side.

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Figure captions

Figure 1. a) Major tectonic elements of the northern North Sea (after Færseth, 1996; Bell et al., 2014). b) Outlines of dataset used for this study. All wells are tied to the seismic data and contain stratigraphic data for the Jurassic (blue), Jurassic and Top Triassic (purple), and Jurassic and Triassic (red). c) Time-structure map of the Top Lunde Formation with major structural elements and faults systems: Alw = Alwyn Fault System, Bre = Brent Fault System, Cor = Cormorant Fault System, Eid = Eider Fault System, ESP = East Shetland Platform, Hea = Heather Fault System, Hud = Hudson Fault System, Hut = Hutton Fault System, MSB = Magnus Sub-basin, Mur = Murchison Fault System, Nin = Ninian Fault System, NSB = Ninian Sub-basin, Osp = Osprey Fault System, Pel = Pelican Fault System, Sta = Statfjord Fault System, Str = Strathspey Fault System, TER = Tern-Eider Ridge, Ter = Tern Fault System, TSB = Tern Sub-basin, Thi = Thistle Fault System, Tor = Tordis Fault System, W–M = West Margin Fault System. The faults systems and structural features are named after the adjacent hydro-carbon bearing fields. Modified after Claringbould et al., 2017.

Figure 2. Uninterpreted and interpreted time-migrated seismic reflection profiles crossing the study area in the a) north, b) centre, and c) south. The seismic profiles including well penetrations and major faults and structural features. See Figure 1b for locations. Modified after Claringbould et al. (2017).

Figure 3. Stratigraphic column of the pre-Triassic-to-Cretaceous in the East Shetland Basin showing lithology (after Færseth, 1996), lithostratigraphic groups/formations and ages, and the interpreted horizons and synthetic well ties (modified after Claringbould et al., 2017). Depth = TVD, GR = Gamma Ray, RHOB = Density, DT = Sonic, RC = Reflection Coefficient, AI = Acoustic Impedance.

Figure 4. Isochrons overlain by fault polygons that offset the top surface (left) with line drawing of faults over outline of 3D seismic data coverage (grey polygons) overlain by the calculated backstripped throw (middle), and expansion index (right) during the deposition of a) Unit 1, b) Unit 2, c) Teist Formation, d) Lomvi and Lunde formations, e) Statfjord Formation, f) Dunlin Group, g) Brent Group, and h) Viking Group. Isochron colours are based on the maximum and minimum thickness value in ms TWT per isochron. Contour interval on all the isochrons is 100 ms TWT. Hatched areas show locations where the upper horizon is eroded. See caption of Figure 1 for abbreviated fault systems and structural features. See Figure 1c for location.

Figure 5. Strain summation across the East Shetland Basin. a) The location of each sample location along three transect lines (North, Centre, South) is shown along with the outline of the three regions (Western, Central, and Eastern). b) Summation of extension [m] for each time period. Values are subdivided per region, average, and transect lines. c) Extension factor per period along each transect line and average, and total pre-Triassic-Jurassic extension factors. d) Triassic-Jurassic extension rate [m/Myr] for each period along each transect line and average across the basin. Darker shades represent relative larger values.

Figure 6. Expansion index (dashed) and backstripped throw [m] (continuous) along the Eider Fault System per time-interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic. See Figure 10 for location of Eider Fault System.

Figure 7. Expansion index (dashed) and backstripped throw [m] (continuous) along the Ninian-Hutton Fault System per time-interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-
Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic. See Figure 10 for location of Ninian-Hutton Fault System.

Figure 8. Expansion index (dashed) and backstripped throw [m] (continuous) along the Cormorant Fault System per time-interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic. See Figure 10 for location of Cormorant Fault System.

Figure 9. Expansion index (dashed) and backstripped throw [m] (continuous) along the Osprey Fault System per time-interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic. See Figure 10 for location of Osprey Fault System.

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Figure 11. Line drawing of faults over outline of 3D seismic data coverage (grey polygons) overlain by Triassic-Jurassic fault slip rates across the East Shetland Basin per time-interval: a) Early-to-Middle Triassic, b) Middle-to-Late Triassic, c) Latest Triassic-to-Early Jurassic, d) Early Jurassic, e) Middle Jurassic, and f) Middle-to-Late Jurassic.

Figure 12. Basin-wide strain distribution across the East Shetland Basin per time-interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic.
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c) Early-to-Middle Triassic (ca. 251-245 Ma) (Teist Formation)

Figure 4. -continued-

d) Middle-to-Late Triassic (ca. 245-201 Ma) (Lomvi and Lunde formations)
Figure 4. -continued-
g) Middle Jurassic (ca. 175-166 Ma) (Brent Group)

h) Middle-to-Late Jurassic (ca. 166-145 Ma) (Viking Group)

Figure 4. -continued-
**Figure 5.** Strain summation across the East Shetland Basin. a) The location of each sample location along three transect lines (North, Centre, South) is shown along with the outline of the three regions (Western, Central, and Eastern). b) Summation of extension [m] for each time period. Values are subdivided per region, average, and transect lines. c) Extension factor per period along each transect line and average, and total pre-Triassic-Jurassic extension factors. d) Triassic-Jurassic extension rate [m/Myr] for each period along each transect line and average across the basin. Darker shades represent relative larger values.

| Time (Ma) | Age (Formation/Group) | Western (n=9) | Central (n=6) | Eastern (n=6) | Average | Transect line | Transect line | Transect line | Transect line | Average | Transect line | Transect line | Transect line | Average | Transect line | Transect line | Transect line | Average |
|----------|-----------------------|---------------|---------------|---------------|---------|---------------|---------------|---------------|---------------|---------|---------------|---------------|---------------|---------|---------------|---------------|---------------|---------|
| 145      | Middle-to-Late Jurassic (Viking Group) | 1849          | 1908          | 1844          | 1867    | 1245          | 2350          | 2007          | 1.018         | 1.028          | 1.028          | 1.025          | 59                | 112          | 96                | 89                |
| 166      | Middle Jurassic (Brent Group) | 232           | 179           | 692           | 368     | 214           | 544           | 346           | 1.003         | 1.031          | 1.027          | 1.020          | 24                | 60           | 38                | 41                |
| 175      | Early Jurassic (Dunlin Group) | 84            | 150           | 599           | 278     | 316           | 392           | 126           | 1.004         | 1.034          | 1.025          | 1.021          | 19                | 23           | 7                 | 16                |
| 192      | Latest Triassic-to-Early Jurassic (Statfjord Formation) | 136           | 111           | 385           | 211     | 266           | 145           | 222           | 1.004         | 1.035          | 1.028          | 1.022          | 30                | 16           | 25                | 23                |
| 201      | Middle-to-Late Triassic (Lomvi and Lunde formations) | 711           | 537           | 564           | 604     | 418           | 877           | 516           | 1.006         | 1.046          | 1.034          | 1.029          | 10                | 20           | 12                | 14                |
| 245      | Early-to-Middle Triassic (Teist Formation) | 1032          | 827           | 461           | 773     | 450           | 1369          | 501           | 1.007         | 1.059          | 1.037          | 1.034          | 75                | 228          | 83                | 129               |
| >251     | Pre-Triassic 3 - Pre-Triassic 2 (Unit 2) | 1655          | 1420          | 2011          | 1695    | 902           | 1688          | 2495          | 1.013         | 1.022          | 1.037          | 1.024          | 24                | 96           | 62                | 162               |
|          | Pre-Triassic 2 - Pre-Triassic 1 (Unit 1) | 1772          | 1765          | 2063          | 1866    | 1416          | 1487          | 2696          | 1.021         | 1.019          | 1.042          | 1.028          | 1078              | 1.115         | 1.139              | 1.111             |

\[n\] is the number of locations where the extension is summed per region and transect line for each time period.
Figure 6. Expansion index (dashed) and backstripped throw [m] (continuous) along the Eider Fault System per time interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic. See Figure 10 for location of Eider Fault System.
Figure 7. Expansion index (dashed) and backstripped throw [m] (continuous) along the Ninian-Hutton Fault System per time interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic. See Figure 10 for location of Ninian-Hutton Fault System.
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Figure 9. Expansion index (dashed) and backstripped throw [m] (continuous) along the Osprey Fault System per time interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic. See Figure 10 for location of Osprey Fault System.
Figure 10. a) Map of the East Shetland Basin, showing fault systems that cross the Top Brent horizon. The grey outlines the seismic data coverage. The location of the detailed analysed fault systems are highlighted in blue with the locations of throw-depth plots marked along the length of the fault system: b) Cormorant Fault System, c) Osprey Fault System, d) Eider Fault System, and e) Ninian-Hutton Fault System. Green shaded areas are interpreted to represent fault growth activity, while red shaded areas represent inactive fault growth. BCU = Base Cretaceous Unconformity, TB = Top Brent Group, TD = Top Dunlin Group, TS = Top Statfjord Formation, TL = Top Lunde and Lomvi formations, TT = Top Teist Formation, PT3 = Top Unit 2, PT2 = Top Unit 1, PT1 = Bottom Unit 1.
Figure 10. -continued-
Figure 11. Line drawing of faults over outline of 3D seismic data coverage (grey polygons) overlain by Triassic-Jurassic fault slip rates across the East Shetland Basin per time interval: a) Early-to-Middle Triassic, b) Middle-to-Late Triassic, c) Latest Triassic-to-Early Jurassic, d) Early Jurassic, e) Middle Jurassic, and f) Middle-to-Late Jurassic.
Figure 12. Basin-wide strain distribution across the East Shetland Basin per time interval: a) Pre-Triassic 2 – Pre-Triassic 1, b) Pre-Triassic 3 – Pre-Triassic 2, c) Early-to-Middle Triassic, d) Middle-to-Late Triassic, e) Latest Triassic-to-Early Jurassic, f) Early Jurassic, g) Middle Jurassic, and h) Middle-to-Late Jurassic.
Supporting Information for

Complex strain partitioning and heterogeneous extension rates during early rifting in the East Shetland Basin, northern North Sea

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Introduction

This supporting information contains supplemental material for 3.3 Fault array analyses and references, and includes a detailed data availability statement.
**S1. Expansion index and backstripped throw**

The expansion index represents the relative vertical stratal thickness ratio between the hanging wall and adjacent footwall (i.e., the hanging wall vertical stratal thickness divided by the footwall vertical stratal thickness). Where the expansion index is >1, syn-depositional fault activity is interpreted to have occurred, and when the expansion index ≤1, the fault is interpreted to be inactive during deposition of that stratal unit (Thorsen, 1963). We use the expansion index to constrain strain trends across the fault array over the different examined time-intervals. The calculated expansion indices across the East Shetland Basin are shown in Figure 4 for each examined time-interval, where warm colour represents low values (pink for values <1.25) and cold colours represent high values (up to 8). Figures 6-9 highlight the expansion indices (dashed) along four key fault systems per each examined time-interval.

Fault throw backstripping involves the sequential subtraction of throw across successively older horizons: this is equivalent to the difference in across-fault vertical stratal thickness for each time period (i.e., the hanging wall vertical stratal thickness minus the adjacent footwall vertical stratal thickness) (e.g., Jackson et al., 2017). As we are investigating how strain is distributed across the basin we are using the original method for fault throw backstripping, rather than the modified method that is more appropriate when one is concerned with the detailed growth and linkage of individual segments comprising the larger fault systems (Jackson et al., 2017). Our calculated backstripped throw values are related to the accompanied expansion indices and are able to put these in perspective if the backstripped throw value is below the vertical seismic resolution (~28 m) (Claringbould, 2015): e.g., when the backstripped throw is below the vertical seismic resolution, a high expansion index can be the result of a picking error. As the examined time-intervals are not equal in time duration (ranging between 6 and 45 Myr) the backstripped throw values are interpreted to indicate a quantitative measure that shows along strike variation in syn-depositional fault activity within a specific time-interval. Figure 4 displays the backstripped throw measurements across the East Shetland Basin for each time period. Warm colours represent low values, while cold colour represents high values. However, as the duration of the examined time-intervals are unequal, the scale bars are not normalized to present the detailed results in the widest colour range possible. Figures 6-9 highlight the backstripped throw measurements (continuous lines) along four key fault systems per each examined time-interval.

**S2. Extension, extension rate, and extension factor**

Extension, extension rate, and extension factor are commonly used to analyse the strain distribution history across basins (e.g., Færseth, 1996; Odinsen et al., 2000; Bell et al., 2011). Three, transect lines are drawn striking ~NW-SE across the basin (North, Centre, South), approximately orthogonally crossing major structural elements. We calculate the upper-crustal extension, extension rate, and extension factor at each location where one of the three transect lines crosses one of the 34 analysed faults within the fault array.

The upper-crustal extension at these points represents the amount of fault heave that developed at the fault during a certain time period. We calculate extension using the backstripped displacement and maximum fault dip angle. We use the
maximum fault angle even though the listric nature of the analysed fault is limited: the average difference between the maximum and minimum fault dip angle measured at each sample location along the analysed faults is 13.2 degrees, with a maximum of 18.7 degrees. We assume that the maximum fault dip angle is most representative of the fault dip angle during fault growth at an analysed time period. Furthermore, using the maximum fault angle also limits the potential effect of footwall erosion, which decreases the fault dip updip. The maximum fault dip angle is subsequently corrected for block-rotation if the fault is located in a rotated hanging wall of a neighbouring fault, or when the fault dip angle is affected by younger fault cross-cutting the analysed fault.

Per time period, the extension calculated at each sample location is summed along the three transects (North, Centre, South) and, additionally, within the three regions (Western, Central, and Eastern). These regions divide the East Shetland Basin in approximately equal parts, to investigate strain migration orthogonal to the main rift axis (i.e. towards or away from the rift axis). Subdividing the amount of extension per transect line and by region allows us to investigate the distribution of strain across the East Shetland Basin over time. Subsequently, the extension rate per transect line is determined for the Triassic and Jurassic time periods to analyse the absolute strain distribution history. Since lithostratigraphic horizons do not necessarily represent chronostratigraphic surfaces (i.e. absolute time-lines), the absolute ages used to estimate the extension rate are based on the average absolute age of the lithostratigraphic boundaries from the wells. The absolute ages of the lithostratigraphic boundaries based on biostratigraphic analyses of the well data. Due to the large extent of the East Shetland Basin (~10,000 km²) the average ages have a maximum difference of ±4 Myr across the fault array, but the time interval of deposition is relatively similar within the basin (±2 Myr). Due to the large time interval of analysis (~150 Myr), we consider using the average absolute age for the lithostratigraphic boundaries to be sufficient to analyse strain accumulation trends across the entire fault array.

Lastly, the extension factor is calculated along each transect line per time period to constrain the relative strain distribution across the fault array during rifting. During rifting the basin extends along the active faults in the upper-crust, increasing the fault heaves and therefore increasing the initial length of the transect line. The extension factor represents relative length ratio of the transect line between two time-horizons: younger over older transect line length. The calculated extension factors are >1, as the length of the transect line increases over time due to rifting (e.g., Bell et al., 2011). Similar to throw and displacement backstripping, the length of the transect line is based on the sequential subtracting of extension amounts of younger time-horizons and the extension of the analysed time-horizon from the current transect length at each point where it crosses an analysed fault.

S3. Throw-depth plots

Throw-depth plots can be used to determine the depth at which faults nucleate and how they propagate vertically (e.g., Hongxing & Anderson, 2007; Jackson & Rotevatn, 2013; Bell et al., 2014; Reeve et al., 2015; Jackson et al., 2017). Hongxing and Anderson (2007) show that throw-depth plots in which throw is constant or decreases with depth (and thus horizon age) are typically associated with post-depositional faulting, whereas, throw-depth plots in which throw increases with depth
and horizon age are indicative of syn-depositional faulting. In this study we are concerned with the varying strain accommodation along the fault (i.e. along strike fault growth evolution). If the character of the throw-depth profiles (i.e., the geometry of the throw-depth plot) are similar along the length of a fault, a laterally consistent growth evolution is assumed (e.g., Jackson & Rotevatn, 2013). However, if throw-depth profiles vary in geometry (i.e., gradient variation between the same horizon-nodes) along the length of the fault, the fault growth is assumed to be laterally diachronous, and therefore reflecting heterogeneous strain distribution across the basin over time.

**S4. Fault slip rate**

Similar to the expansion indices and backstripped throw values, we calculate fault slip rate along each major fault for every time period, with the exception of the pre-Triassic units as the age of these is unconstrained (Units 1 and 2) (Figure 11). The fault slip rate represents the backstripped displacement over time in m/Myr, and shows the variation in strain distribution along strike of each major fault and across the fault array as rifting progressed. Similar to backstripped throw, displacement backstripping involves the sequential subtractions of displacements on successively older horizons; where the displacement is calculated using the throw and heave of a faulted horizon (e.g., Childs et al., 1993; Ten Veen & Kleinspehn, 2000; Walsh et al., 2002; Taylor et al., 2004, 2008; Bell et al., 2014; Jackson et al., 2017). Similar to the extension rates the absolute ages used to estimate the fault slip rates are based on the average absolute age of the lithostratigraphic boundaries across the East Shetland Basin (see Supplemental material section 2). Fault slip rates are displayed in Figure 11 and since only a few measurements are >100 m/Myr the colour ranges from 0 to >100 m/Myr. Of the 634 measurements made for each time period, 29 (maximum of 212 m/Myr, Teist Formation), 12 (maximum of 333 m/Myr, Brent Group), and 9 (maximum of 127 m/Myr, Viking Group), are >100 m/Myr.
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Data availability statement

The data used for this study are publically available for download via the UK National Data Repository (NDR) (https://ndr.ogauthority.co.uk) for the United Kingdom side, and the DISKOS online portal (Diskos) (https://portal.diskos.cgg.com) for the Norwegian side.

2D seismic reflection lines:
Diskos: NRS06
NDR: NSR06 (NP062D) (TGS NOPEC)

3D seismic reflection surveys:
Diskos: ST07M06 (ST07M06_mega_south & ST07M06_mega_n) (Equinor ASA)
NDR: MC3D_NNS14 (PGS Exploration UK LTD), which is a merge of:
   Survey name (Survey alias)
   PP113DGESB (MC3DG11UK_ESB)
   PP093DGESB (MC3D_ESB2009)
   PP123DGHBR (MC3DG12UK_HBR)
   PP103DGESB (MC3D_ESB2010)
   PP123DGDUN (MC3D_DUN2012)
   PP133DGDUN (MC3D_DUN2013)

Borehole data:
Diskos: NO blocks 31, 32, and 35
NDR: UK quadrants 210, 211, 1, 2, and 3