

An overview of the Upper Palaeozoic–Mesozoic stratigraphy of the NE Atlantic region

M. S. STOKER^{1*}, M. A. STEWART¹, P. M. SHANNON², M. BJERAGER³, T. NIELSEN³, A. BLISCHKE⁴, B. O. HJELSTUEN⁵, C. GAINA⁶, K. McDERMOTT^{2,7} & J. ÓLAVSDÓTTIR⁸

¹*British Geological Survey, The Lyell Centre, Research Avenue South, Riccarton, Edinburgh EH14 4AP, UK*

²*School of Geological Sciences, University College Dublin, Belfield, Dublin 4, Ireland*

³*Geological Survey of Denmark and Greenland, Øster Vøldgade 10, DK-1350 Copenhagen, Denmark*

⁴*ÍSOR (Iceland Geosurvey), Department of Energy Technology, Rangarvellir, PO Box 30, 602 Akureyri, Iceland*

⁵*Department of Earth Sciences, University of Bergen, Allegaten 41, N-5007 Bergen, Norway*

⁶*Centre for Earth Evolution and Dynamics, University of Oslo, Postbox 1028, Blindern, N-0135 Oslo, Norway*

⁷*Present address: ION Geophysical, 1st Floor, Integra House, Vicarage Road, Egham TW20 9JZ, UK*

⁸*Jarðfeingi, Brekkutín 1, PO Box 3059, FO-110 Tórshavn, Faroe Islands*

*Corresponding author (e-mail: mss@bgs.ac.uk)

Abstract: This study describes the distribution and stratigraphic range of the Upper Palaeozoic–Mesozoic succession in the NE Atlantic region, and is correlated between conjugate margins and along the axis of the NE Atlantic rift system. The stratigraphic framework has yielded important new constraints on the timing and nature of sedimentary basin development in the NE Atlantic, with implications for rifting and the break-up of the Pangaeon supercontinent. From a regional perspective, the Permian–Triassic succession records a northwards transition from an arid interior to a passively subsiding, mixed carbonate–siliciclastic shelf margin. A Late Permian–earliest Triassic rift pulse has regional expression in the stratigraphic record. A fragmentary paralic to shallow-marine Lower Jurassic succession reflects Early Jurassic thermal subsidence and mild extensional tectonism; this was interrupted by widespread Mid-Jurassic uplift and erosion, and followed by an intense phase of Late Jurassic rifting in some (but not all) parts of the NE Atlantic region. The Cretaceous succession is dominated by thick basinal-marine deposits, which accumulated within and along a broad zone of extension and subsidence between Rockall and NE Greenland. There is no evidence for a substantive and continuous rift system along the proto-NE Atlantic until the Late Cretaceous.



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The NE Atlantic region incorporates the conjugate continental margins of NW Europe (between SW Ireland and northern Norway) and East Greenland, as well as the intervening NE Atlantic Ocean. The latter is an area of complex bathymetry and structure that includes both active and extinct spreading ridges, Iceland and its insular margin, the Greenland–Iceland–Faroe Ridge Complex, and the Jan Mayen microcontinent (JMMC) (Fig. 1; Table 1). To the north, the NE Atlantic Ocean is connected

to the Arctic Ocean via the Fram Strait, which developed as a result of transtensional shear between North Greenland and the western Barents Sea; to the south, the NE Atlantic Ocean is separated from the North Atlantic Ocean by the Charlie–Gibbs Fracture Zone.

For much of the Phanerozoic eon, the NE Atlantic region experienced a shared geological development since its amalgamation – the collision of Baltica and Laurentia to form Laurussia – during the

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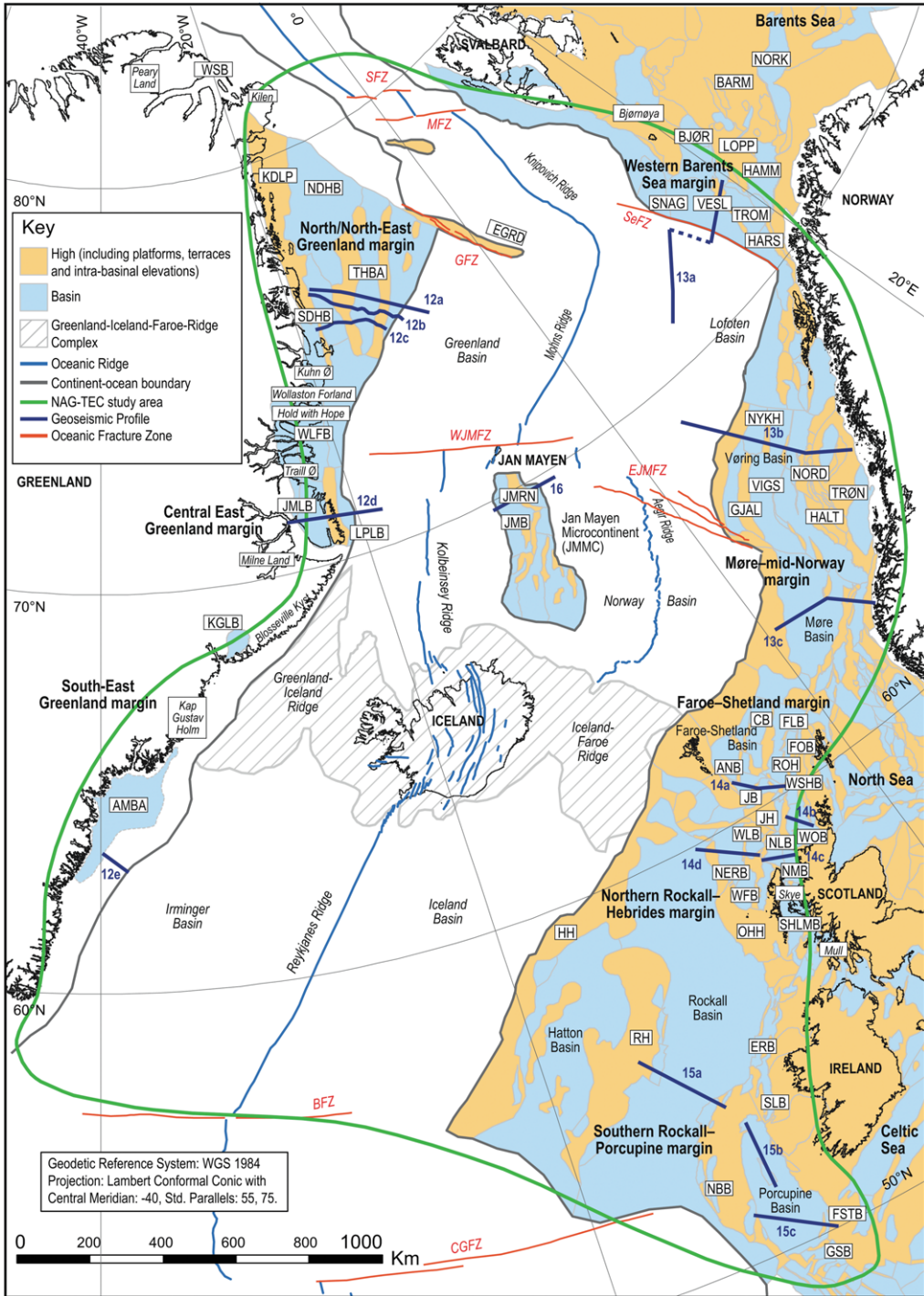


Fig. 1. Map showing location of NAG-TEC study area, with geographical regions, geoseismic profile locations and key structural elements referred to in the text (modified after Hopper *et al.* 2014). See Table 1 for key to structural element abbreviations. Map information: the Geodetic Reference System is WGS 1984; the Projection is Lambert Conformal Conic with Central Meridian of -40 , and standard parallels of 55 and 75.

THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

Table 1. Explanation of structural element abbreviations in Figure 1

<i>European margin</i>	
ANB	Annika Sub-basin
BARM	Bjarmeland Platform
BJØR	Bjørnøya Basin
CB	Corona Sub-basin
ERB	Erris Basin
FLB	Flett Sub-basin
FOB	Foula Sub-basin
FSTB	Fastnet Basin
GJAL	Gjallar Ridge
GSB	Goban Spur Basin
HALT	Halten Terrace
HAMM	Hammerfest Basin
HARS	Harstad Basin
HH	Hatton High
JB	Judd Sub-basin
JH	Judd High
LOPP	Loppa High
NBB	North Bróna Basin
NERB	NE Rockall Basin
NLB	North Lewis Basin
NMB	North Minch Basin
NORD	Nordland Ridge
NORK	Nordkapp Basin
NYKH	Nyk High
OHH	Outer Hebrides High
RH	Rockall High
ROH	Rona High
SHLMB	Sea of the Hebrides–Little Minch Basin
SLB	Slyne Basin
SNAG	Sørvestsnaget Basin
TROM	Tromsø Basin
TRØN	Trøndelag Platform
VESL	Veslemøy High
VIGS	Vigrid Syncline
WFB	West Flannan Basin
WLB	West Lewis Basin
WOB	West Orkney Basin
WSHB	West Shetland Basin
<i>East Greenland margin and Jan Mayen microcontinent</i>	
AMBA	Ammassalik Basin
EGRD	East Greenland Ridge
JMB	Jan Mayen Basin
JMLB	Jameson Land Basin
JMMC	Jan Mayen Microcontinent
JMRN	Jan Mayen Ridge North
KDLP	Koldewey Platform
KGLB	Kangerlussuaq Basin
LPLB	Liverpool Land Basin
NDHB	North Danmarkshavn Basin
SDHB	South Danmarkshavn Basin
THBA	Thetis Basin
WFLB	Wollaston Forland Basin
WSB	Wandel Sea Basin
<i>Oceanic fracture zones</i>	
BFZ	Bight Fracture Zone
CGFZ	Charlie Gibbs Fracture Zone

Table 1. Continued

EJMFZ	East Jan Mayen Fracture Zone
GFZ	Greenland Fracture Zone
MFZ	Molloy Fracture Zone
SFZ	Spitsbergen Fracture Zone
SeFZ	Senja Fracture Zone
WJMFZ	West Jan Mayen Fracture Zone

Late Silurian Scandic phase of the Caledonian Orogeny: its subsequent Late Palaeozoic incorporation into the Pangaeian supercontinent; the prolonged history of extension linked to the Mesozoic break-up of Pangaea; and, ultimately, the early–mid-Cenozoic opening of the NE Atlantic Ocean (Ziegler 1988; Doré *et al.* 1999; Roberts *et al.* 1999; Torsvik *et al.* 2002; Pharaoh *et al.* 2010). The assembly of Pangaea resulted from the convergence and collision of Laurussia with Gondwana (from the south), with final suturing – in the NE Atlantic region – marked by the Late Carboniferous–Early Permian Variscan Orogeny. However, the development of Permian–Triassic rift basins over a large part of the west European and NE Atlantic regions marks the onset of instability shortly after the assemblage of Pangaea. This instability was driven by a number of processes, including the late spreading collapse of the Variscan orogen, the southwards propagation of rifting emanating from the Arctic and the opening of the neo-Tethys (Williams & McKie 2009). This early rift phase instigated a long period of post-Variscan extension that eventually culminated in early–mid-Cenozoic break-up of this wide rifted region.

In general terms, the tectonic evolution of this region has been dominated by several phases of episodic extension, which are increasingly attributed to the interaction of a southwards-propagating ‘Arctic’ rift and a northwards-propagating ‘Atlantic’ rift (Doré 1992; Doré *et al.* 1999; Roberts *et al.* 1999; Lundin & Doré 2005a; Ellis & Stoker 2014). The break-up phase that led to the formation of the NE Atlantic Ocean was accompanied by magmatic activity, which some authors attribute to an ‘Iceland’ mantle plume (e.g. White 1988, 1989; White & McKenzie 1989; White & Lovell 1997; Smallwood *et al.* 1999; Skogseid *et al.* 2000; Smallwood & White 2002), whereas others argue that rifting and break-up can be wholly explained by plate tectonic mechanisms, lithospheric thinning and variable decompressive melting along the rifts (e.g. Foulger & Anderson 2005; Foulger *et al.* 2005; Lundin & Doré 2005a, b; Ellis & Stoker 2014).

Within the NE Atlantic region, Roberts *et al.* (1999) perceived the ‘Arctic’ rift to extend southward via the Norwegian–Greenland area into the West Shetland area and the North Sea, whereas

(Continued)

the 'Atlantic' rift comprises the SE Greenland, Hebrides, Rockall, Porcupine and the western Celtic Sea areas. The overall rift system is generally fairly simple in the Greenland–Norway region, but becomes more complex with different rift arms from the SE Greenland–Faroe–Shetland region southwards. Arguably, this is why Roberts *et al.* (1999) divided the overall system into the Arctic (simple, northern) and the Atlantic (more complex, southern) rifts. Whilst taking into consideration this north–south variation in structural complexity, we prefer to use the term 'NE Atlantic' for the entire rift system – excluding the North Sea – in this paper, which we suggest is a more accurate indicator of the geographical domain that we are focused upon, and does not force or perpetuate this subdivision without a more rigorous analysis of its history of rifting (see below).

According to Roberts *et al.* (1999), these northern and southern sections of the rift system remained largely separate entities until the Late Jurassic or Early Cretaceous; thereafter, the rift system is envisaged to have become a single entity represented by the Rockall, Faroe–Shetland, Møre and Vøring basins. By way of contrast, various palinspastic reconstructions of the NE Atlantic region depict a through-going NE Atlantic rift system along this line of basins since the Permian (e.g. Ziegler 1988; Cope *et al.* 1992; Doré 1992; Knott *et al.* 1993; Torsvik *et al.* 2002; Coward *et al.* 2003; McKie & Williams 2009; Pharaoh *et al.* 2010). Alternatively, Doré *et al.* (1999) have suggested that pre-Cretaceous extensional fault activity within the area between NW Britain/Ireland and SE Greenland was limited and discontinuous, and that the NE Atlantic rift system was not fully linked until the mid-Cretaceous. These authors describe a mosaic-like fragmentation of Pangaea during the Permian–Triassic, replaced by a more systematic, albeit restricted, east–west extension in the Jurassic, with the change to NW–SE extension in the Cretaceous. The latter extension vector was maintained through to plate break-up. The change from east–west Jurassic extension to NW–SE Cretaceous extension is attributed to a switch from a system dominated by north–south Tethyan rift propagation to one dominated by NE–SW rifting driven by northwards-propagating Atlantic spreading (Doré *et al.* 1999).

On the basis of these differing viewpoints, it is clear that the evolution of the NE Atlantic rift systems remains enigmatic. Despite the pioneering work of Ziegler (1988, p. vii), in his own words he described his palinspastic reconstructions as, 'generalised ... they may have serious shortcomings ... and should be regarded as working hypotheses'. Unfortunately, most subsequent authors that have considered the post-Caledonian evolution of the NE Atlantic region have largely continued to rely

on the maps of Ziegler (1988), thereby perpetuating all of their inherent limitations. This is particularly evident in that part of the NE Atlantic region between NW Britain/Ireland and SE Greenland – the critical rift tip overlap area in the dual rift model – where the Late Palaeozoic–Mesozoic rock record is very fragmentary and only sparsely known (cf. Ritchie *et al.* 2011a; Hitchen *et al.* 2013). Thus, reconstructions that currently show linked Permo-Triassic, Jurassic and Early Cretaceous rift basins across this area are largely without any robust foundation.

In view of the uncertainty relating to the pre-break-up configuration of the NE Atlantic rift system, this paper focuses on the Upper Palaeozoic–Mesozoic (Permian–Cretaceous) rock record preserved on the circum-NE Atlantic continental margins (Fig. 1). The description and distribution of the stratigraphic successions presented in this paper were compiled as part of the NAG-TEC project – a large-scale observational programme designed to better understand the tectonic development of the conjugate NE Atlantic margins and adjacent ocean basin (Hopper *et al.* 2014). The limit of the main NAG-TEC study area is indicated in Figure 1, and extends over 35° of latitude between northern Greenland and the Porcupine Basin. The main objective of the paper is to summarize our current understanding of the Permian–Cretaceous stratigraphy around the NE Atlantic region (within the NAG-TEC study area). We present the first detailed correlation of the various lithostratigraphies developed across this region, and highlight key major unconformities and changes in regional sedimentation patterns. On this basis, we are better able to appraise the history of development of the full NE Atlantic rift system; in particular, to provide a base level of observational data that we hope can be used to test more specific process-orientated studies. The implications of our observations for the putative Late Palaeozoic–Mesozoic palaeogeographical reconstructions will also be briefly discussed.

Data and methods

This stratigraphic summary presents a comprehensive review of the known record of Permian–Cretaceous rocks around the NE Atlantic region, with a focus on distribution, structural setting, stratigraphy, boundary relationships and depositional environments. The stratigraphic architecture is further illustrated in a series of geoseismic profiles from across the study area. The study area has been divided into several sub-areas that include: the North–NE Greenland margin; the western Barents Sea–Svalbard margin; the central East Greenland

margin; the Møre–mid-Norway margin; the SE Greenland margin (Cretaceous only); the Faroe–Shetland–northern Rockall–Hebrides margin; and the southern Rockall–Porcupine margin (including the NW Irish Shelf and the western part of the Celtic Sea region) (Fig. 1). The boundary between the NE Greenland and the central East Greenland sub-areas is taken at about 72° N, and between central East Greenland and SE Greenland at about 70° N: however, it should be noted that, for ease of description, the Permian–Triassic rocks between Traill Ø and Hold with Hope (the Wollaston Foreland Basin; 72–74° N) are included within the summary of the central East Greenland sub-area. The Jan Mayen microcontinent (JMMC) is also included as part of this review, and described separately. Although Svalbard lies outside of our main area of study, information from the rocks cropping out on the archipelago has been utilized as a guide to the composition of the adjacent offshore area. Note also that, whilst our stratigraphic distribution maps (Figs 2–4) extend into the northern North Sea, this region is beyond the scope of this study. For details of the North Sea region, the reader is referred to two major atlases that provide a comprehensive review of the stratigraphic framework, and which have been utilized in the production of our maps (Evans *et al.* 2003; Doornenbal & Stevenson 2010).

The decision-making involved in the compilation of the regional stratigraphic distribution maps, the stratigraphic correlation panels and the geoseismic profiles, whilst based in part on published data, was also informed by the intimate expertise and knowledge of the contributors responsible for each sub-area, including the provision and use of unpublished geological survey information – commonly derived from onshore outcrops and offshore boreholes – and released commercial well data. Those boreholes and wells perceived to contain the most significant stratigraphical information have been collated within a stratigraphic database, details of which are presented elsewhere (Hopper *et al.* 2014; Stoker *et al.* 2014). The main processes involved in the creation of the maps and correlation panels are described below.

Stratigraphic distribution maps

Stratigraphic distribution maps have been compiled for the Permian–Triassic, Jurassic and Cretaceous successions (Figs 2–4). These have been compiled largely on the basis of existing published and unpublished geological survey maps, supplemented by correlating chronostratigraphic divisions in key stratigraphic wells and boreholes (cf. Stoker *et al.* 2014), and by cross-checking against published/unpublished seismic data and regional cross-sections. The basic mapping was compiled in

ArcGIS, and for each chronostratigraphic division the following polygons have been delivered for each sub-area:

- ‘present’ and ‘not-present’ polygons, indicating the areas where a chronostratigraphic division is present or absent as proven by outcrops, wells and/or seismic interpretation; and
- ‘inferred’ polygons, indicating areas where the presence of a chronostratigraphic division has been assumed based on seismic interpretation or regional maps.

The information from each sub-area was subsequently merged to produce the regional stratigraphic distribution maps illustrated in Figures 2–4.

Stratigraphic correlation panels and geoseismic sections

A series of seven regional stratigraphic correlation panels (Table 2; Figs 5–11) has been constructed, each of which features a number of generalized stratigraphic columns that summarize the Permian–Cretaceous geological history of selected parts of the NE Atlantic region, both onshore and offshore. The details of their geographical extent and the key structural elements/domains that they represent are summarized in Table 2. Each correlation panel is arranged as a general east-to-west conjugate-margin stratigraphic transect.

The correlation panels combine chronostratigraphy with lithostratigraphic (where available) and other information, such as sedimentary environments and magmatic events. Lithostratigraphic information is presented mostly at ‘Group’ level. Each individual offshore stratigraphic column on the correlation panel has been constructed on the basis of information derived from a number of key wells and boreholes located in the area represented by that generalized column (cf. Stoker *et al.* 2014). This type of information is mostly located on the NW European continental margin. On the Greenland margin, the generalized stratigraphic columns (Figs 5–10) are mostly compiled from onshore data (Table 2). All of the generalized stratigraphic columns based on the onshore mapping data and the offshore well and borehole data present a moderate–high reliability in terms of the chronostratigraphy. By way of contrast, the age and composition of the succession in several of the basins offshore north, NE and central East Greenland, as well as the JMMC (Table 2), is inferred on the basis of seismic interpretation and structural position in relation to adjacent established basins. Whereas thick successions are observed on seismic reflection profiles across these basins, the stratigraphic reliability of the infill remains low at the time of this study.

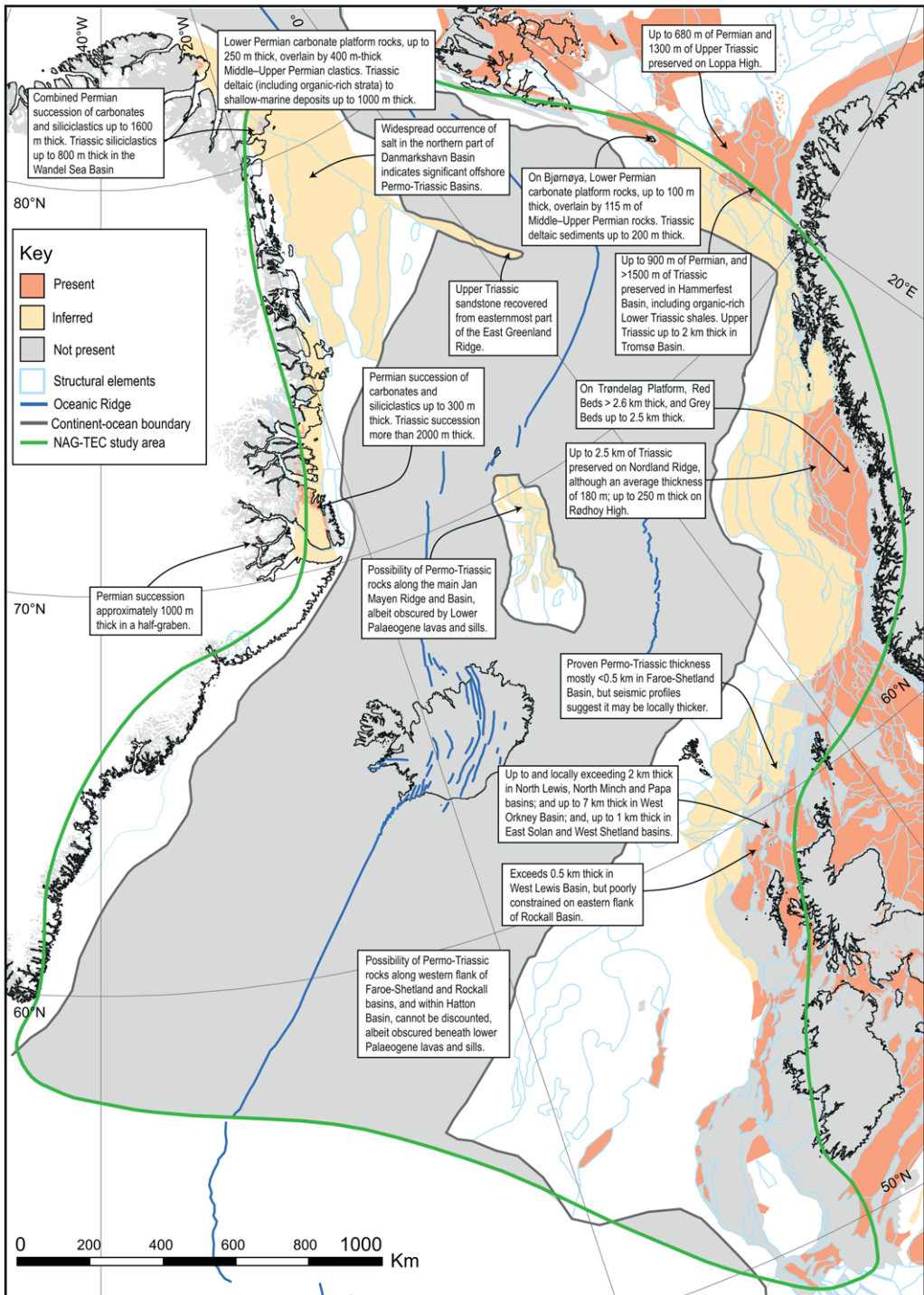


Fig. 2. Stratigraphic distribution of Permian–Triassic rocks (based on the compilation of Stoker *et al.* 2014). Map information: the Geodetic Reference System is WGS 1984; the Projection is Lambert Conformal Conic with a Central Meridian of -40 , and Std. Parallels of 55 and 75 .

THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

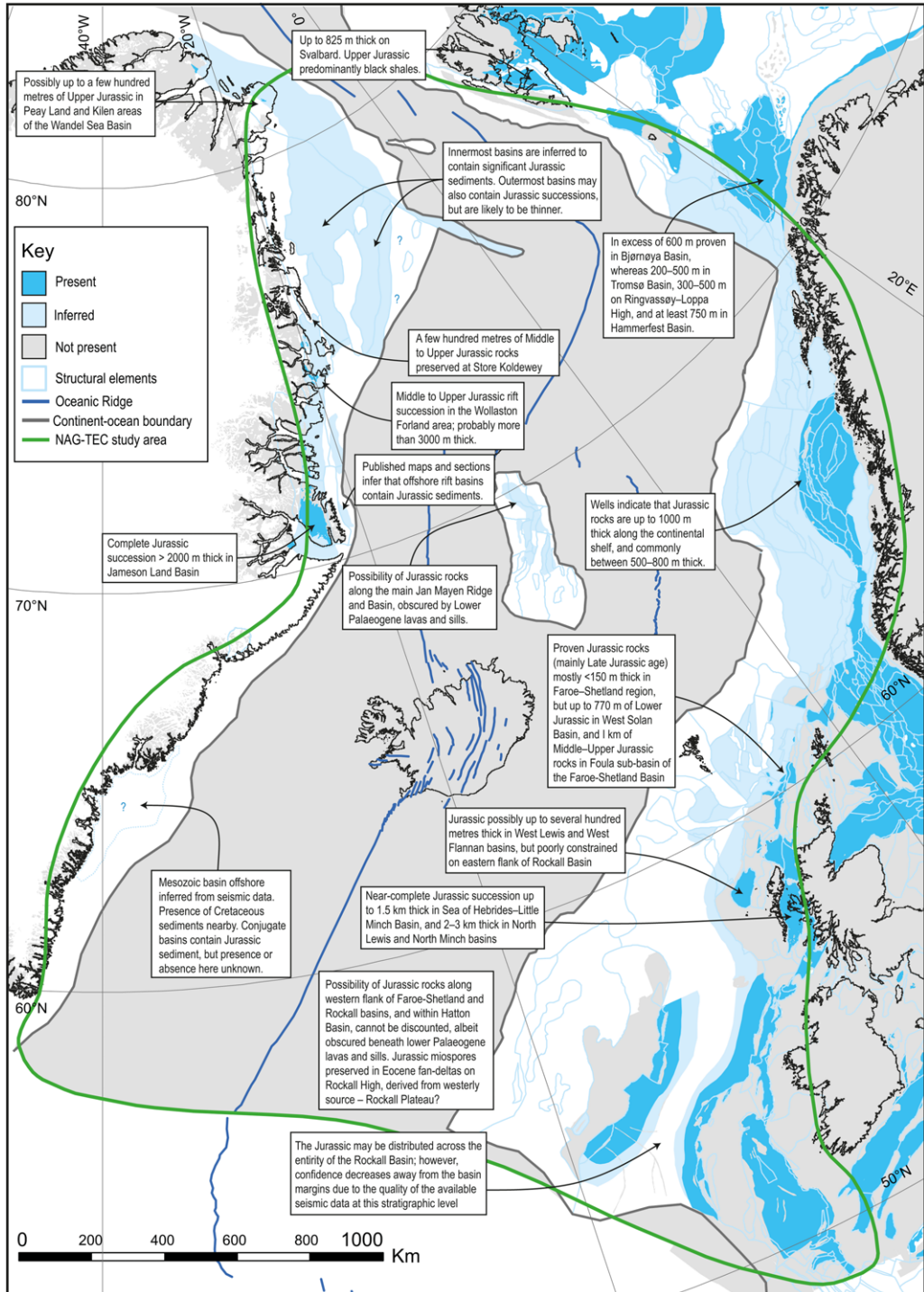


Fig. 3. Stratigraphic distribution of Jurassic rocks (based on the compilation of Stoker *et al.* 2014). Map information: the Geodetic Reference System is WGS 1984; the Projection is Lambert Conformal Conic with Central Meridian of -40 , and standard parallels of 55 and 75 .

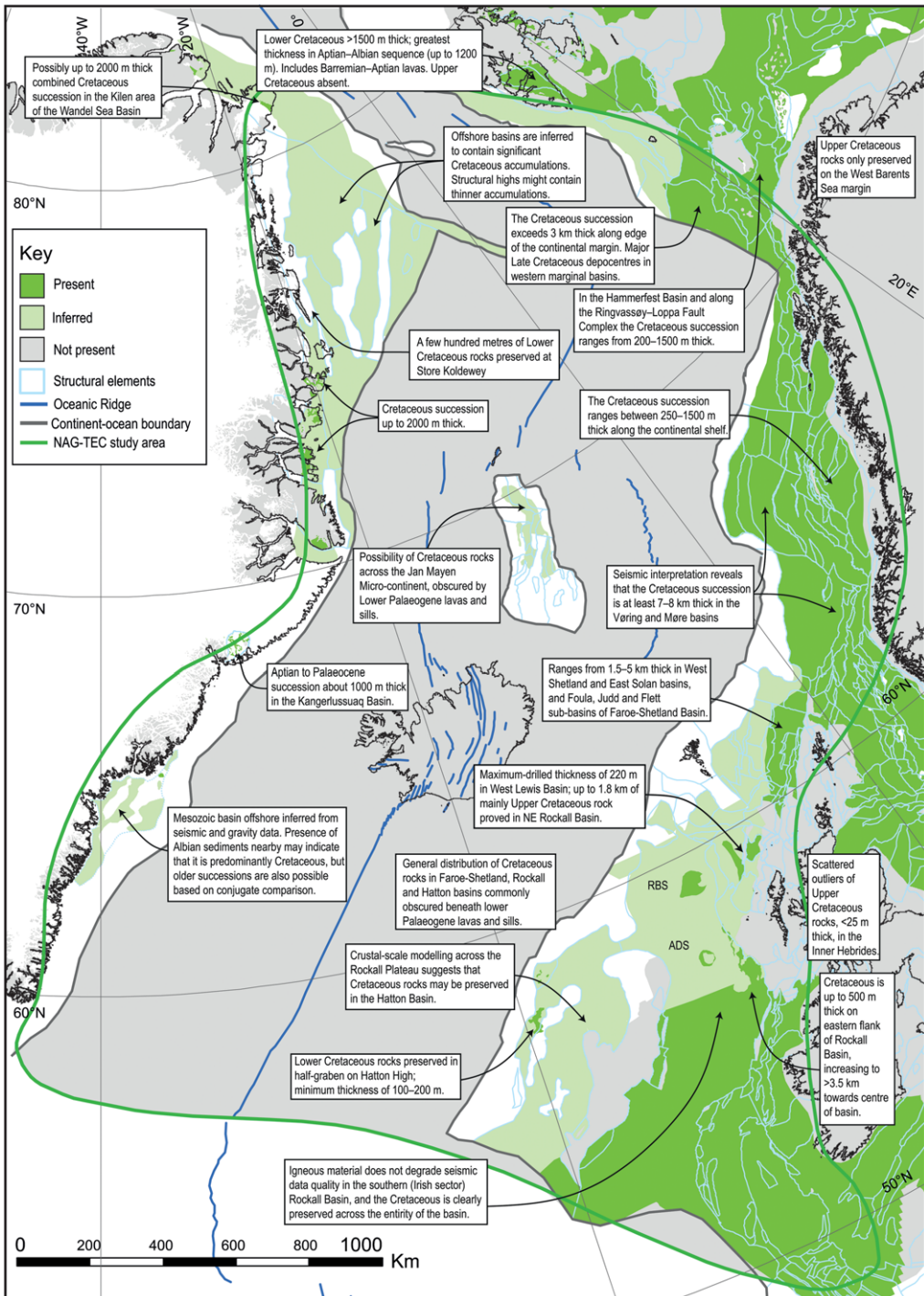


Fig. 4. Stratigraphic distribution of Cretaceous rocks (based on the compilation of Stoker *et al.* 2014). Abbreviations: ADS, Anton Dohrn Seamount; RBS, Rosemary Bank Seamount. Map information: the Geodetic Reference System is WGS 1984; the Projection is Lambert Conformal Conic with Central Meridian of -40 , and standard parallels of 55 and 75.

THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

Table 2. List of regional stratigraphic correlation panels (with figure numbers) arranged as conjugate-margin transects across the NAG-TEC study area

Fig. No.	Regional correlation panel	Key structural elements/domains
5	North Greenland–Svalbard–western Barents Sea region	Wandel Sea Basin*; northern Danmarkshavn Basin [†] ; Svalbard*; SW Barents Shelf, including the Loppa High and Hammerfest Basin
6	NE Greenland region	Wollaston Foreland–Kuhn Ø*; Koldewey Platform [†] ; southern Danmarkshavn Basin [†] ; Thetis Basin [†]
7	NE Greenland–Vøring region	Traill Ø–Hold with Hope*; northern and southern Vøring continental shelf
8	Central East Greenland–southern Jan Mayen–Møre region	Jameson Land*; Liverpool Land Basin [†] ; southern Jan Mayen microcontinent [†] ; Møre continental shelf
9	SE Greenland–Faroe–Shetland region	Kangerlussuaq Basin*; Faroe–Shetland Basin; West Shetland marginal basins
10	SE Greenland–northern Rockall Plateau–northern Rockall Basin–Hebrides region	Ammassalik margin; Hatton Basin; northern Rockall Basin; West Lewis Basin; Hebridean basins; Inner Hebrides*
11	Southern Rockall Basin–Porcupine Basin–NW Irish region	Southern Rockall Basin; Porcupine Basin; Erris Basin; Slyne Basin; Goban Spur; Fastnet Basin

*Onshore data.

[†]Seismic interpretation only.

The stratigraphic architecture across the conjugate continental margins is also displayed in a series of geoseismic profiles, which have been compiled from a variety of sources (see Figs 12–16).

Stratigraphic descriptions

On the basis of this database, the following sections describe the stratigraphy of the Permian–Triassic, Jurassic and Cretaceous successions. These descriptions are presented in order from north to south to illustrate the regional variations, and we have treated each of the sub-areas (outlined above) in a similar manner, addressing key aspects of distribution, structural setting, stratigraphy, boundary relationships and depositional environments.

Permian–Triassic stratigraphy of the NE Atlantic margins

North–NE Greenland margin

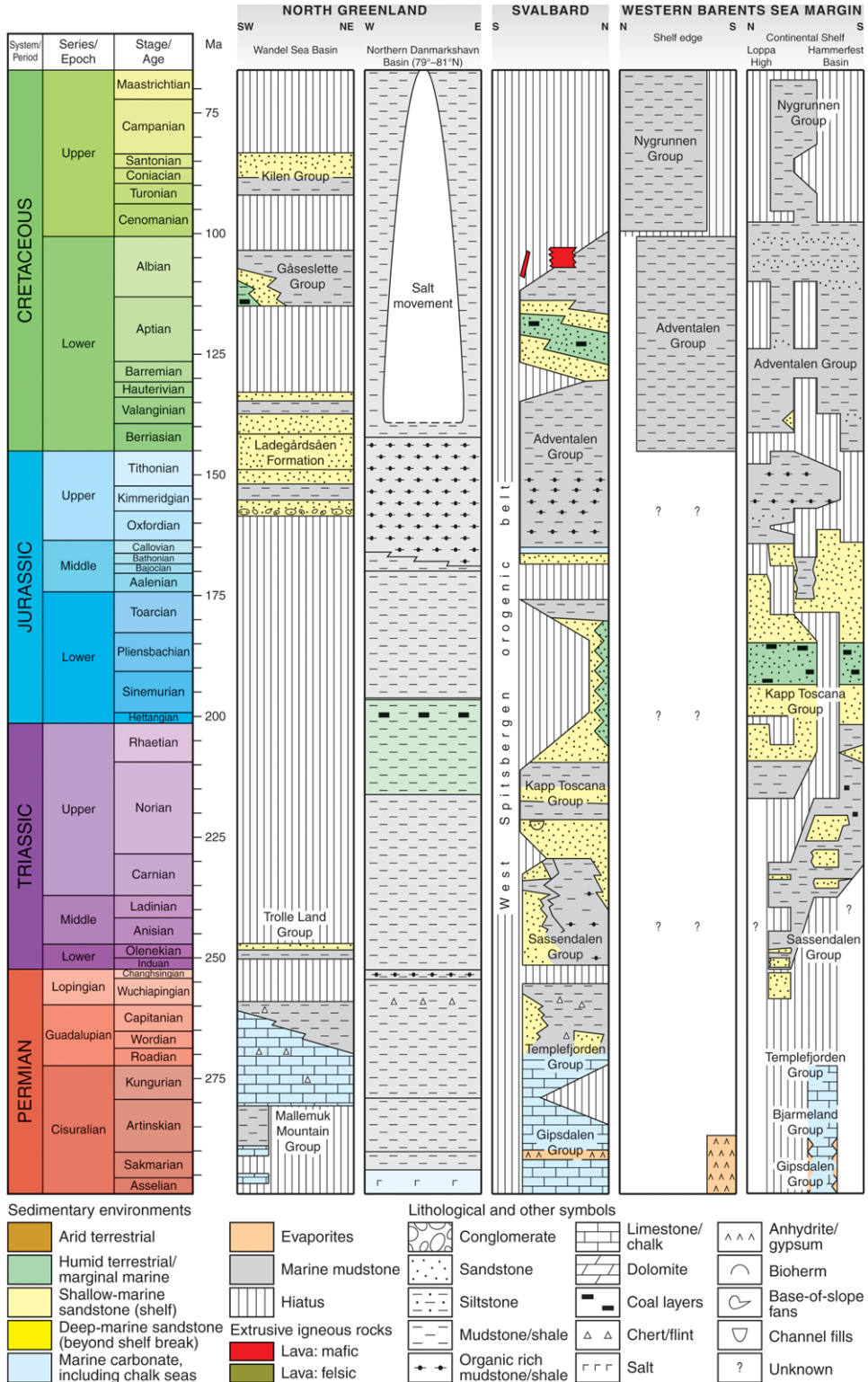
Distribution. Permian–Triassic strata are exposed onshore eastern North Greenland in the Wandel Sea Basin (79°–83° N) with an inferred continuation offshore to the north and east (Døssing *et al.* 2010) (Fig. 2). On the NE Greenland margin, the succession is interpreted to be present offshore on the Store Koldewey Platform and in the Danmarkshavn Basin based on seismic data (Hamann *et al.* 2005), and possible Permian–Triassic strata in the Thetis Basin are speculatively suggested by Dinkelman *et al.* (2010) and Jackson *et al.* (2012)

(Figs 2 & 12a). Rocks of Late Triassic age have been sampled on the East Greenland Ridge (Nielsen *et al.* 2014) (Fig. 2).

Structural setting. The Late Palaeozoic–Early Mesozoic rift basins trend approximately north–south along the NE Greenland margin. On the North Greenland margin, the Wandel Sea Basin forms a rift basin linking this north–south rift trend with a roughly NW–SE-orientated Palaeozoic rift trend between Greenland and Svalbard (Stemmerik & Håkansson 1991).

Stratigraphy. The Permian in the onshore part of the Wandel Sea Basin is represented by carbonates and mudstones overlain by shale, limestones and sandstones of the Mallemuk Mountain Group, up to about 1600 m thick (Stemmerik *et al.* 1996, 2000) (Figs 2 & 5). This succession is overlain by the Triassic Trolle Land Group of conglomerates, sandstones and mudstones, possibly up to 1 km thick (Håkansson *et al.* 1991; Stemmerik *et al.* 1996) (Fig. 5).

Offshore NE Greenland, a more complete rock record through the Permian–Triassic periods – as suggested by analogy to the conjugate Svalbard and western Barents Sea successions – is inferred for the Danmarkshavn and Thetis basins, and the Koldewey Platform (Figs 5 & 6). An Upper Carboniferous–Triassic succession up to 7 km thick is interpreted for the Danmarkshavn Basin (Hamann *et al.* 2005). The Permian rocks are inferred to comprise platform carbonate, basinal evaporite and marine mudstone deposits overlain by a Triassic



siliciclastic succession that is dominated by fine-grained marine and marginal-marine sediments (Gautier *et al.* 2011) (Figs 5 & 6). The deposition of halite evaporites is validated by widespread salt diapirism in the northern part of the basin (Fig. 5). A source-rock-prone interval may occur towards the top of the Permian sequence, and the boundary with the Triassic might be, at least locally, an erosional unconformity. A comparable sequence has been inferred for the Koldewey Platform, albeit with coarser clastic facies locally developed along its western margin (Fig. 6). The postulated Permian–Triassic succession in the Thetis Basin remains undivided (Dinkelman *et al.* 2010; Jackson *et al.* 2012). The Upper Triassic rock samples from the East Greenland Ridge consist of sandy mudstones (Nielsen *et al.* 2014).

Boundary relationships with underlying and overlying systems. The onshore part of the Wandel Sea Basin records an unconformity at the Carboniferous–Permian boundary, an intra-Permian unconformity and a major unconformity at the Permian–Triassic transition (Stemmerik *et al.* 1998). A major unconformity separates the Triassic from the Upper Jurassic rocks (Håkansson *et al.* 1991). In the offshore basins, the Carboniferous–Permian transition is inferred to be conformable, whereas an unconformity is inferred at the Permian–Triassic and Triassic–Jurassic boundaries in the southern Danmarkshavn Basin and on the Koldewey Platform.

Depositional environment. The Mallemuk Mountain Group in the Wandel Sea Basin represents the Early Permian shallow-water carbonate shelf. A marked flooding surface forms the base of the overlying succession, with mudstone deposition followed by intercalating carbonates, mudstones and siliciclastics deposited in shelfal environments (Stemmerik & Håkansson 1991). The Triassic Trolle Land Group includes fluvial and shallow-marine sandy to muddy shelf deposition. The basins offshore NE Greenland have an interpreted Upper Carboniferous–Permian marine carbonate platform, with an evaporite succession deposited during a phase of thermal subsidence; this was followed by deposition of marine basinal mudstones during the Triassic, which were succeeded by increasingly marginal-marine and terrestrial deposition during the Late Triassic (Hamann *et al.* 2005).

Western Barents Sea–Svalbard margin

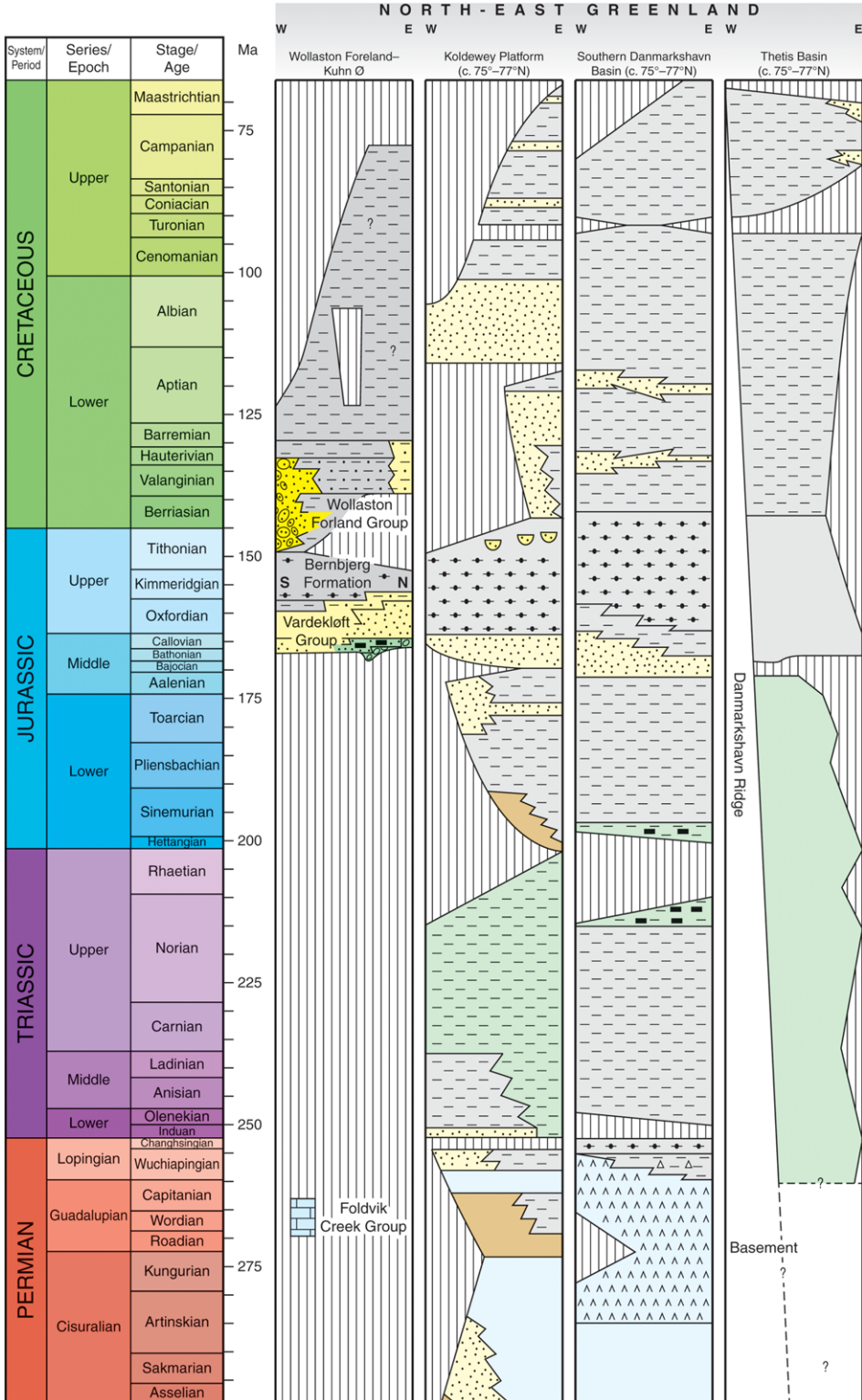
Distribution. Rocks of Permian–Triassic age crop out on Svalbard and Bjørnøya, and have been proved in numerous wells on and adjacent to the southern part of the western Barents Sea margin, and thus are inferred to underlie much of this part of the continental margin (Fig. 2).

Structural setting. A stable carbonate platform persisted from the Late Carboniferous into the Early Permian with only minor uplifts (late-stage Variscan orogenic activity), whereas the Mid–Late Permian was characterized by uplifts and erosional breaks (Steel & Worsley 1984; Brekke *et al.* 2001; Worsley 2008). The latter includes a major Mid–Late Permian unconformity that is widely developed across the Western Barents Sea–Svalbard region (Fig. 5). This is associated with Late Permian–Early Triassic rifting linked to the Uralian Orogeny, which gave rise to normal faulting and rapid subsidence: for example, the western margin of the Loppa High (Brekke *et al.* 2001; Faleide *et al.* 2010). However, for much of the Triassic period, the area was largely one of thermal relaxation (Doré *et al.* 1999; Roberts *et al.* 1999).

Stratigraphy. The Permian Gipsdalen, Bjarmeland and Templefjorden groups comprise a sequence of carbonate, evaporate and shallow-marine clastic rocks across the western Barents Sea–Svalbard margin, including Bjørnøya (Harland 1997; Dallmann 1999; Worsley 2008; Henriksen *et al.* 2011; NORLEX 2014) (Figs 2 & 5). On Svalbard, the Permian succession is up to 650 m thick, albeit punctuated by a late Early Permian unconformity, whereas up to 900 m of Lower Permian strata occur in the Hammerfest Basin, on the SW Barents Sea margin. The shallow-marine clastic succession, which includes mudstone, sandy bioclastic limestone and glauconitic sandstone with minor chert and silicified limestone, becomes increasingly dominant throughout the region in the Mid–Late Permian.

The Triassic succession comprises the Lower–Middle Triassic Sassendalen and the Upper Triassic–Middle Jurassic Kapp Toscana groups (Nøttvedt *et al.* 1993b; Harland 1997; Henriksen *et al.* 2011; NORLEX 2014) (Fig. 5). The Triassic rocks are commonly separated from the Permian by an unconformity marking a change from the cherts and siliciclastics of the Templefjorden Group to softer,

Fig. 5. Regional stratigraphic correlation panel showing the generalized Permian–Cretaceous succession preserved in the North Greenland (Wandel Sea Basin)–Svalbard–western Barents Sea region, and inferred (paler colourfill) on the basis of seismic interpretation for the area off NE Greenland (northern Danmarkshavn Basin). Correlation based on the compilation of Stoker *et al.* (2014). Timescale based on Gradstein *et al.* (2012).



darker, non-siliceous shales at the base of the Sassendalen Group (Harland 1997; Worsley 2008).

The Sassendalen Group is up to 800 m thick across the western Barents Sea–Svalbard region, and characteristically comprises alternating shale, siltstone and sandstone units. These units are commonly arranged as upwards-coarsening cycles and include a number of organic-rich formations (Dalland *et al.* 1988; Nøttvedt *et al.* 1993a, b; Harland 1997; Dallmann 1999; Worsley 2008; NORLEX 2014) (Fig. 5).

The Triassic formations of the Kapp Toscana Group are characterized by interbedded mudstone, shale and sandstone with sporadic thin coal beds arranged in small- to large-scale upwards-coarsening units (Dalland *et al.* 1988; Nøttvedt *et al.* 1993a, b; Dallmann 1999; Worsley 2008; NORLEX 2014). Whereas this succession continues into the Middle Jurassic across much of the region, no rocks younger than Triassic are found on Bjørnøya. The preserved thickness of the Upper Triassic succession is variable: on Svalbard, Bjørnøya and in the Sørvestnaget Basin, the succession is several hundred metres thick; on the southern Loppa High and in the Hammerfest Basin, it exceeds 1000 m in thickness; and in the Tromsø Basin, the Upper Triassic is >2000 m thick (Figs 1 & 2). In the SW Barents Sea region, the Upper Triassic is locally punctuated by unconformities (Fig. 5).

Boundary relationships with underlying and overlying systems. In the western Barents Sea basins and on Svalbard, the Carboniferous–Permian boundary is conformable (Fig. 5). The Triassic–Jurassic boundary is marked by seemingly continuous facies from Rhaetian to Toarcian in Svalbard; elsewhere, a local unconformity may exist (e.g. the Hammerfest Basin).

Depositional environments. The extensive and stable Late Carboniferous–Early Permian carbonate platform was replaced in the Mid–Late Permian by shallow-marine clastic deposits that transgressed the low-lying platform (Harland 1997; Steel & Worsley 1984; Worsley 2008; NORLEX 2014). By the Early Triassic, the deposits of the Sassendalen Group on Svalbard reflect a muddy distal marine shelf with coarser material being sourced from a delta that prograded from the west, possibly sourced by local exposed areas to the west and NW (Dallmann 1999; Riis *et al.* 2008; NORLEX 2014). In

the SW Barents Sea, siliciclastic coastal and deltaic sequences fringed a wide shelf. These environments persisted and expanded basinwards through the deposition of the Kapp Toscana Group, in the Mid- and Late Triassic. In the general Barents Sea region, the deltaic system was increasingly sourced from the Uralian Mountains in the east and the Fennoscandian Shield in the south to SE (Riis *et al.* 2008; Worsley 2008; Smelror *et al.* 2009; Henriksen *et al.* 2011). An overall shallowing of the Barents Sea region during the Mid–Late Triassic resulted from the input of large volumes of clastic sediment deposited by the westwards progradation of the deltaic system (Brekke *et al.* 2001; Riis *et al.* 2008; Glørstad-Clark *et al.* 2011; Høy & Lundschieen 2011; Anell *et al.* 2014).

Central East Greenland margin

Distribution. Permian rocks are exposed onshore in the Wollaston Forland area and, together with the Triassic, form a combined succession that extends southwards towards, and including, the Jameson Land Basin (Figs 1 & 2). Permian–Triassic strata may also be present in deep fault blocks in the offshore Liverpool Land Basin (Larsen 1990; Hamann *et al.* 2005) (Fig. 12d), but have not been recorded in basins, either onshore or offshore, further south (i.e. in SE Greenland basins, such as the Kangerlussuaq and Ammassalik basins) (Fig. 1).

Structural setting. As with the NE Greenland margin, the Late Palaeozoic–Early Mesozoic rift basins trend approximately north–south along the central East Greenland margin. The NE and central East Greenland margins experienced a Late Palaeozoic rift phase with the formation of half-graben that possibly persisted into the Early Permian (Surlyk *et al.* 1986; Stemmerik *et al.* 1991; Surlyk 1991). Marginal rift blocks in East Greenland were uplifted and peneplained during latest Carboniferous–Early Permian time. The uplift was associated with rotational block-faulting that tilted strata by up to 15° (Surlyk 1990).

The Lower–Upper Permian boundary marks a transition in tectonic style, with the initiation of regional subsidence by thermal contraction in the Late Permian. A Kazanian (Wordian–Capitanian) marine connection was established between the Barents Sea, the Jameson Land Basin and the European Zechstein Basin (Stemmerik 2000) (Figs 6–8).

Fig. 6. Regional stratigraphic correlation panel showing the generalized Permian–Cretaceous succession preserved onshore in NE Greenland (Wollaston Forland–Kuhn Ø), and inferred (paler colourfill) on the basis of seismic interpretation for the adjacent offshore area (Koldewey Platform; southern Danmarkshavn Basin; Thetis Basin). See Figure 5 for the key to colours and symbols. Correlation based on the compilation of Stoker *et al.* (2014). Timescale based on Gradstein *et al.* (2012).

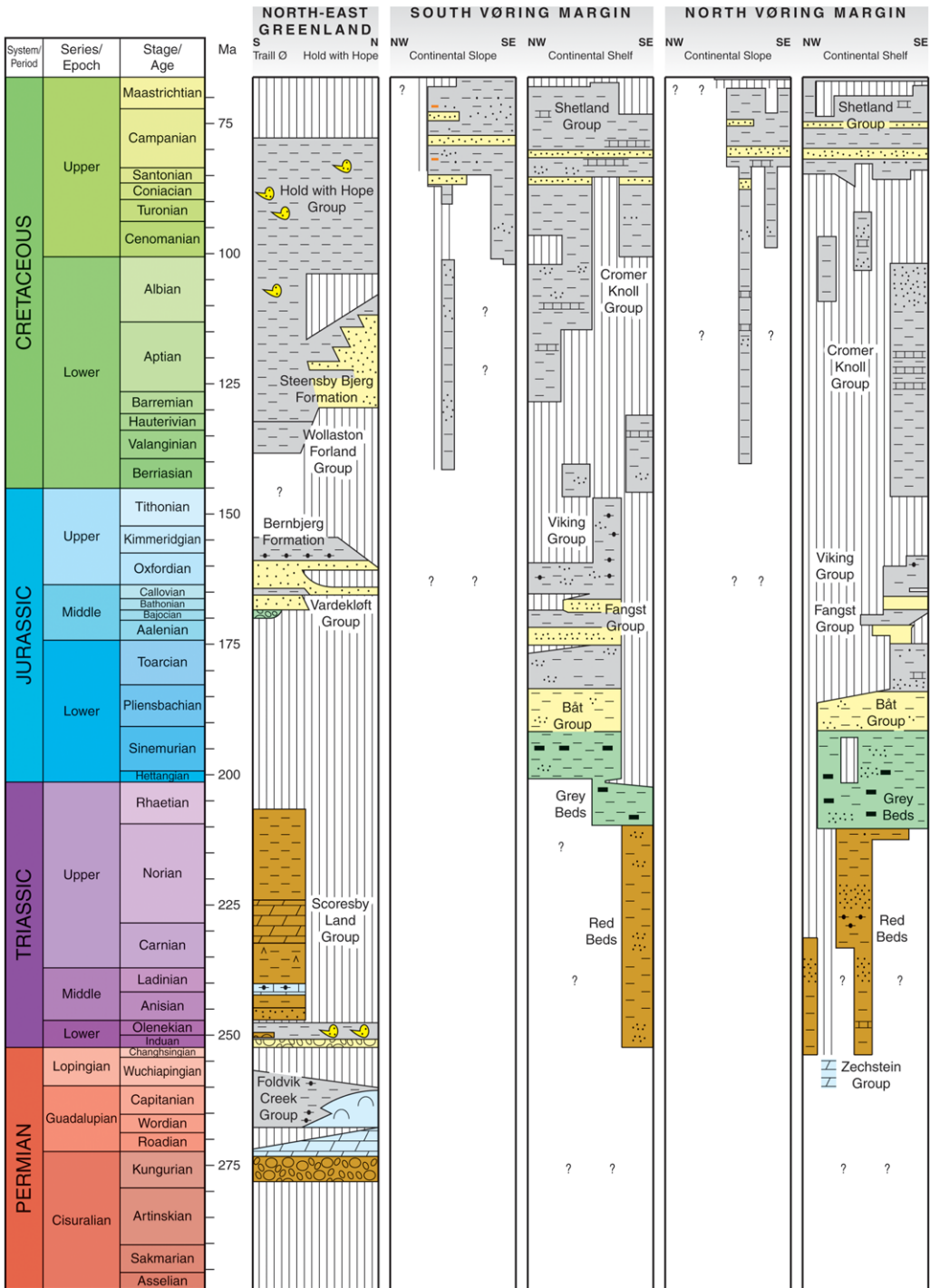


Fig. 7. Regional stratigraphic correlation panel showing the generalized Permian–Cretaceous succession preserved in the Central East Greenland–Vøring region. See Figure 5 for the key to colours and symbols. Correlation based on the compilation of Stoker *et al.* (2014). Timescale based on Gradstein *et al.* (2012).

THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

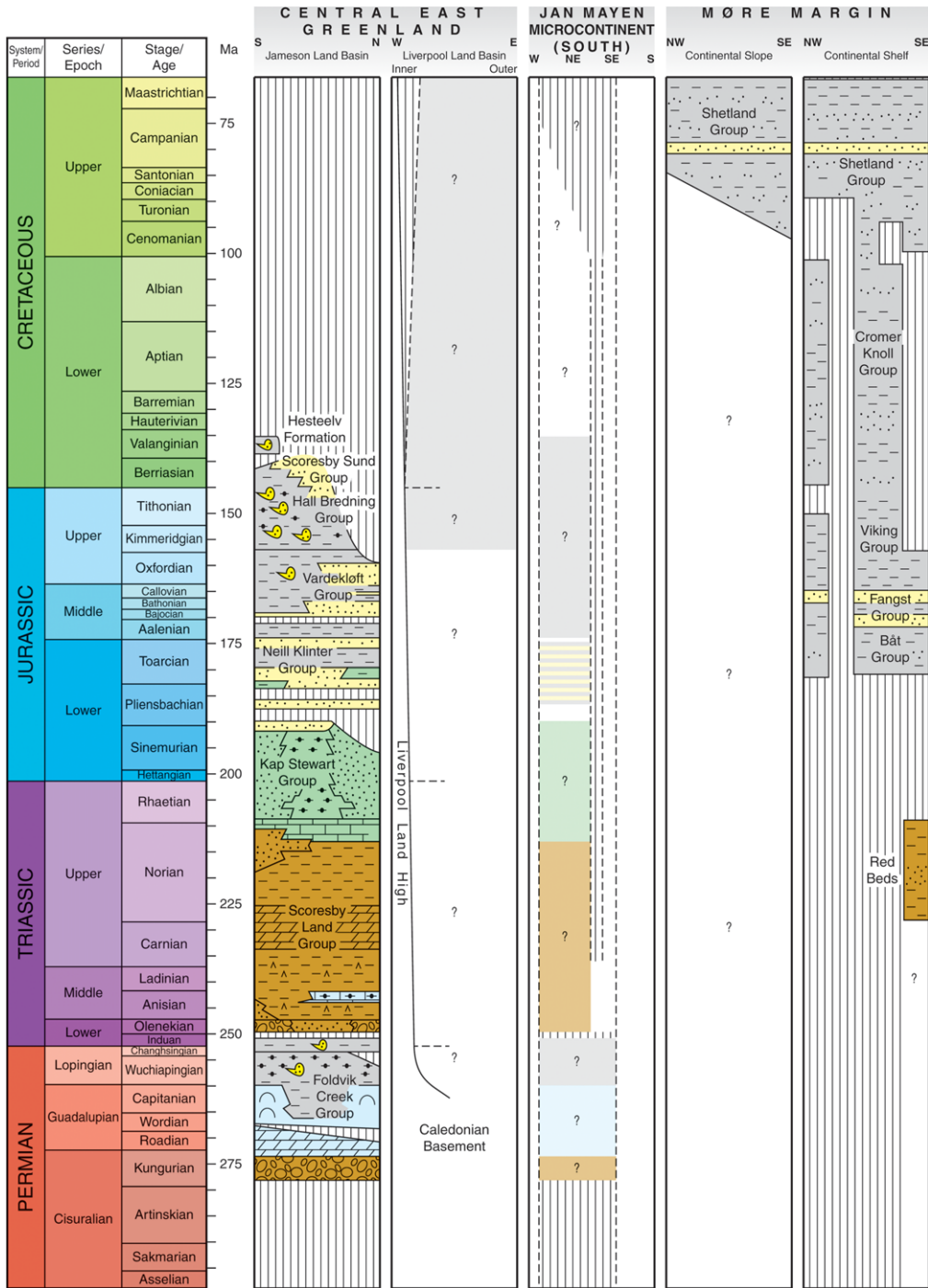
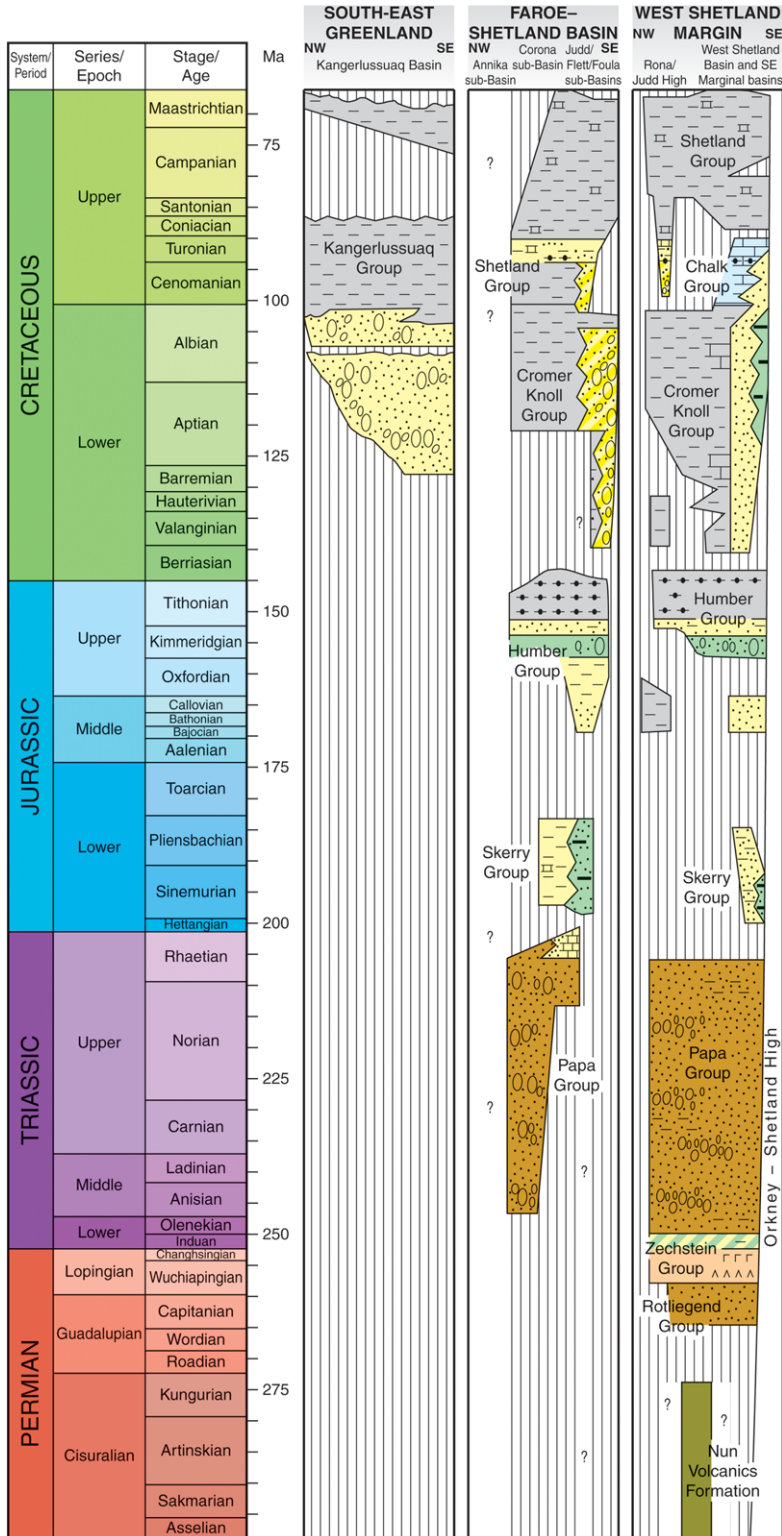


Fig. 8. Regional stratigraphic correlation panel showing the generalized Permian–Cretaceous succession preserved in the Central East Greenland (Jameson Land Basin)–Møre region, and the inferred (paler colourfill) stratigraphic range based on seismic interpretation for the Liverpool Land Basin (offshore Central East Greenland), and the southern Jan Mayen microcontinent. See Figure 5 for the key to colours and symbols. Correlation based on the compilation of Stoker *et al.* (2014). Timescale based on Gradstein *et al.* (2012).



THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

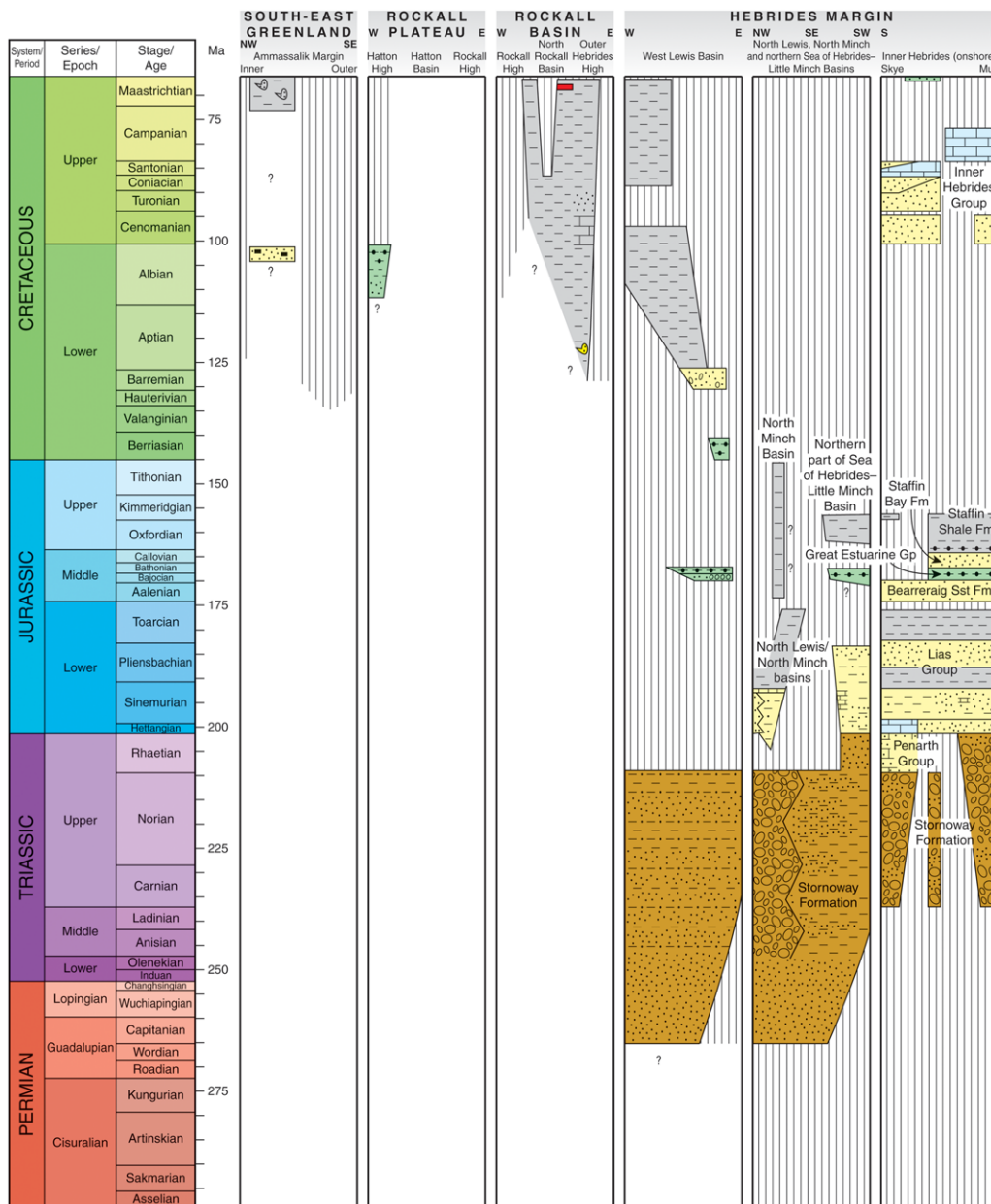
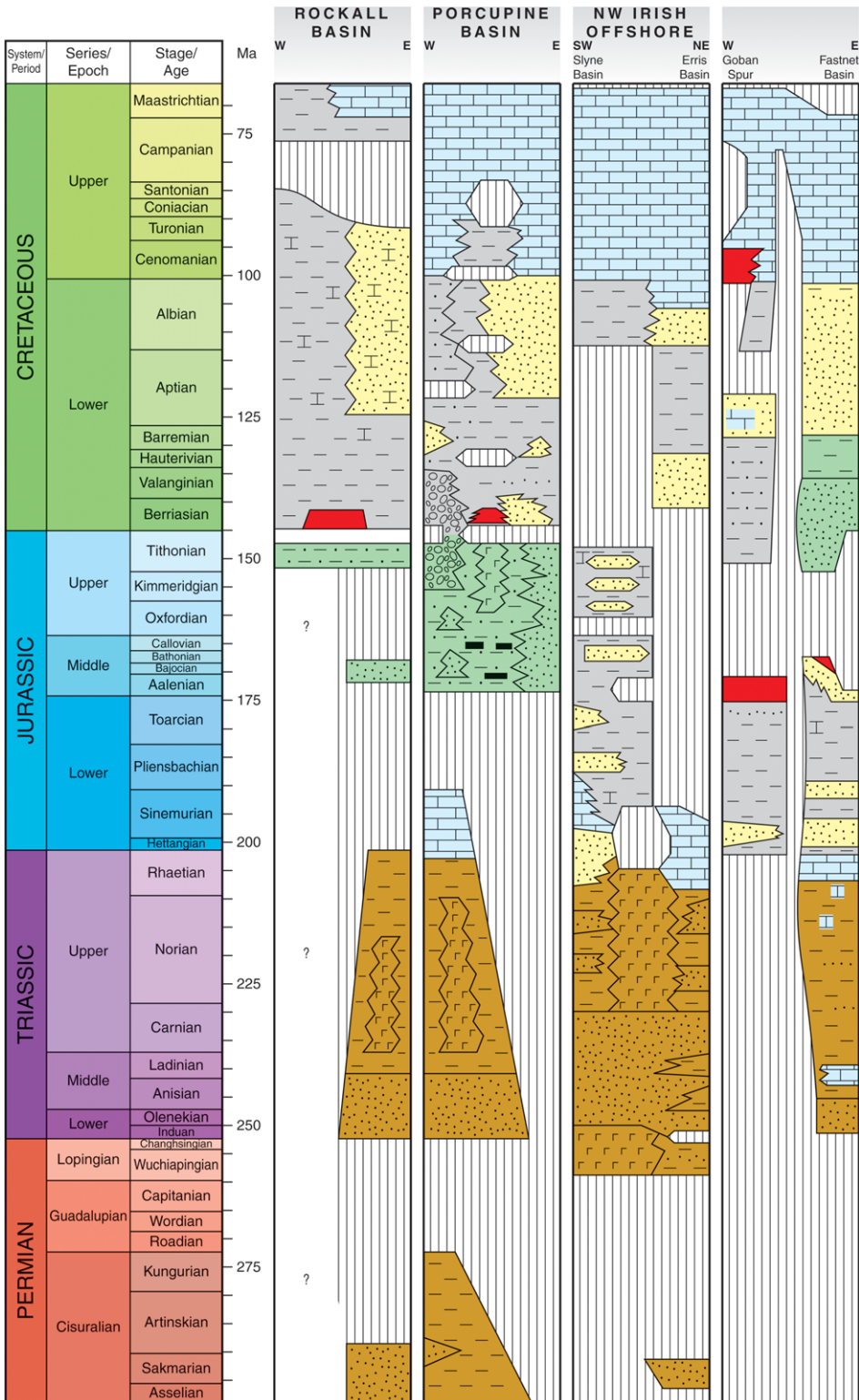


Fig. 10. Regional stratigraphic correlation panel showing the generalized Permian–Cretaceous succession preserved in the SE Greenland–northern Rockall Plateau–northern Rockall Basin–Hebrides region. See Figure 5 for the key to colours and symbols. The sequence in the Ammassalik Basin is based partly on seismic interpretation (Gerlings *et al.*, this volume, in review). Correlation based on the compilation of Stoker *et al.* (2014). Timescale based on Gradstein *et al.* (2012).

Fig. 9. Regional stratigraphic correlation panel showing the generalized Permian–Cretaceous succession preserved in the SE Greenland–Faroe–Shetland region. See Figure 5 for the key to colours and symbols. Correlation based on the compilation of Stoker *et al.* (2014). Timescale based on Gradstein *et al.* (2012).



THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

Minor latest Permian–earliest Triassic rifting and rotational block faulting was recorded in an overall east–west extensional regime, and several half-graben were formed in the Early Triassic in East Greenland (Seidler *et al.* 2004). A marine seaway between East Greenland and Norway was connected to the seas covering Svalbard, the Sverdrup Basin and NW Siberia in the Early Triassic (Mørk *et al.* 1992; Seidler *et al.* 2004; Müller *et al.* 2005; Bjerager *et al.* 2006).

Early Triassic rotational block faulting resulted in the formation of angular unconformities in East Greenland (Figs 7 & 8). The main rift event occurred in the late Scythian (Induan–Olenikian transition) along north–south-trending basin-margin faults, with minor rifting also in the Norian. NW–SE-trending transform fault zones probably controlled local differential uplift and subsidence patterns in the basins (Clemmensen 1980a). The latest Triassic (Rhaetian)–Mid-Jurassic (early Bajocian) was characterized by thermal subsidence in East Greenland.

Stratigraphy. The stratigraphy of the Permian–Triassic succession is herein described largely from the Wollaston Forland to Jameson Land area (Figs 1 & 2). An isolated Middle–Upper Permian Røde Ø Conglomerate succession, approximately 1 km thick, also occurs in the Rødefjord area, west of Milne Land, in central East Greenland (Stemmerik & Piasecki 2004); in contrast, the stratigraphy of the Liverpool Land Basin remains conjectural (Figs 2, 8 & 12d).

In the combined Jameson Land–Wollaston Forland basins, the Permian Foldvik Creek Group is up to about 300 m thick, and comprises a basal conglomerate that is succeeded by carbonates and evaporates, in turn overlain by carbonate build-ups and organic shales that are capped by siliciclastic deposits (Stemmerik 2001) (Figs 2, 6–8). The top of the Permian consists of basinal bioturbated mudstones and sandy turbidites in the Jameson Land Basin and southern Traill Ø (Kreiner-Møller & Stemmerik 2001). The Upper Permian succession includes a source-rock-prone interval (Christiansen *et al.* 1993).

The Triassic rocks are largely assigned to the Scoresby Land Group. The Lower Triassic sequence is up to approximately 1 km thick in the Traill Ø area, in the deepest part of the combined Jameson Land–Wollaston Forland Basin, and consists of shales, sandstones, conglomerates and minor carbonates (Seidler *et al.* 2004; Bjerager *et al.* 2006)

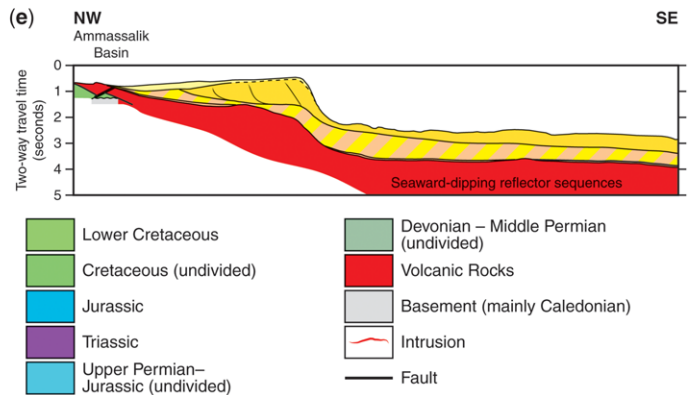
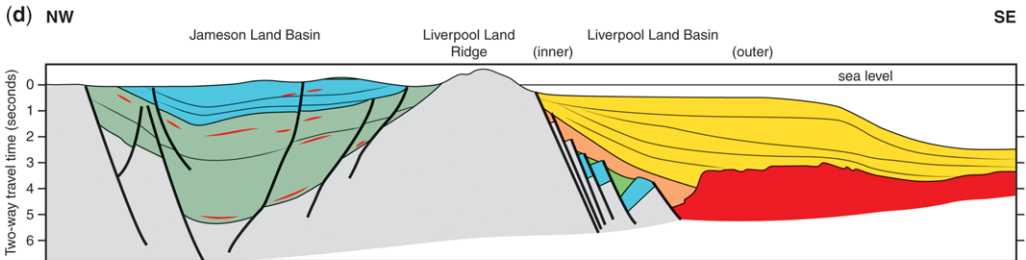
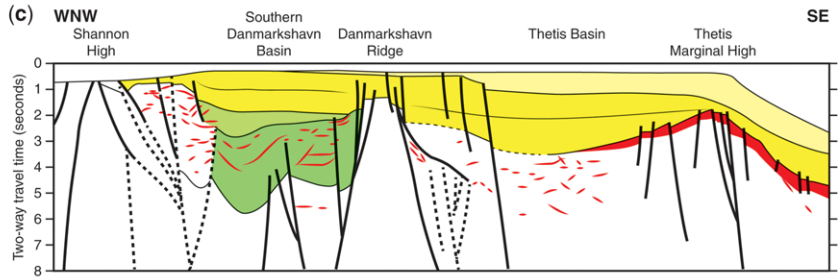
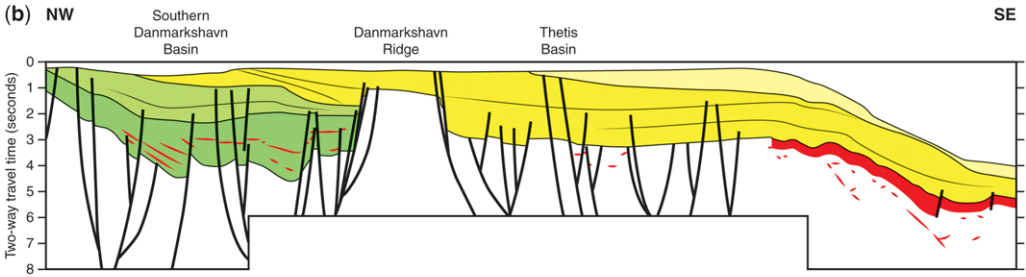
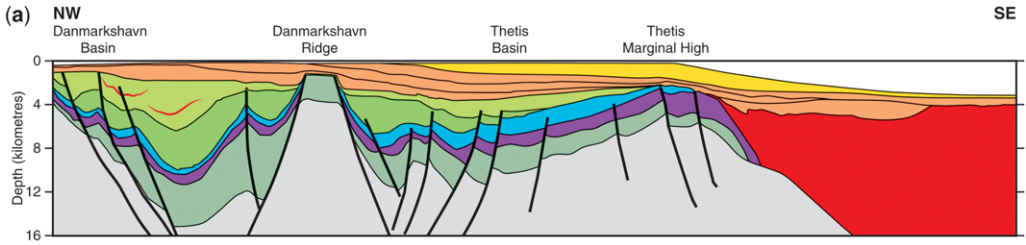
(Figs 7 & 8). These rocks are unconformably overlain by a series of Lower–Middle Triassic basal conglomerates and sandstones, up to 600 m thick, which grade upwards into fine-grained sandstones, mudstones, evaporites and carbonates of Mid–Late Triassic age (Clemmensen 1980a). The uppermost Triassic (Rhaetian) rocks in the Traill Ø–Hold with Hope area are missing, whereas in the Jameson Land Basin the Scoresby Land Group is overlain by organic-rich mudstones and sandstone of the Rhaetian–Sinemurian (Lower Jurassic) Kap Stewart Group (see below).

Boundary relationships with underlying and overlying systems. The regional Early–Mid-Permian erosional expanse of tilted Carboniferous–Devonian strata or Caledonian basement in central East Greenland is overlain by the Foldvik Creek Group (Figs 6–8). A prominent erosional unconformity at the Permian–Triassic boundary is recorded in basin-marginal positions. The associated hiatus expands towards the north with evidence of subaerial exposure, and much of the Upper Permian has been removed by erosion prior to deposition of the Triassic Scoresby Land Group (Surlyk *et al.* 1986) (Fig. 7). In the deeper basinal areas in the south, a complete marine Permian–Triassic boundary interval is recorded in a marine mudstone succession (Stemmerik *et al.* 2001) (Fig. 8). The Triassic succession is conformable with the Jurassic in the Jameson Land Basin, whereas the Upper Triassic is erosionally overlain by Jurassic and/or Cretaceous further north.

Depositional environment. The Foldvik Creek Group shows an overall development from fluvial into shallow-marine and hypersaline carbonate deposition. The successive transgression resulted in deposition of fully marine carbonate build-ups along shelf margins contemporaneously with the deposition of anoxic shales in basinal areas (Stemmerik *et al.* 2001) (Fig. 8). An increased siliciclastic influx is evident in the latest Permian that continued into the Triassic, with the overall upwards-shallowing marine shales and sandy turbidites at the base of the Scoresby Land Group succeeded by shoreface and coastal-plain sandstones and mudstones (Seidler *et al.* 2004; Bjerager *et al.* 2006) (Figs 7 & 8). A fluvial conglomerate was deposited locally in the Traill Ø area as the result of local block tilting.

Marine deposition terminated in the Early Triassic and was followed by the deposition of rift-related

Fig. 11. Regional stratigraphic correlation panel showing the generalized Permian–Cretaceous succession preserved in the southern Rockall Basin–Porcupine Basin–NW Irish region. See Figure 5 for the key to colours and symbols. Correlation based on the compilation of Stoker *et al.* (2014). Timescale based on Gradstein *et al.* (2012).



0 25 kilometres

- Pliocene–Holocene
- Neogene (undivided)
- Oligocene–middle Miocene
- Eocene
- Palaeogene (undivided)
- Tertiary (undivided)
- Upper Cretaceous

- Lower Cretaceous
- Cretaceous (undivided)
- Jurassic
- Triassic
- Upper Permian–Jurassic (undivided)

- Devonian – Middle Permian (undivided)
- Volcanic Rocks
- Basement (mainly Caledonian)
- Intrusion
- Fault

alluvial conglomerates and sandstones, which were sourced from the western (Post-Devonian Main Fault area) and the eastern (Liverpool Land) basin margins (Clemmensen 1980a). The succession shows a gradual upwards transition into fine-grained floodplain, freshwater, and saline lacustrine and aeolian deposits of Mid–Late Triassic age, although a thin Anisian unit of mudstones and carbonates reflect a shallow-marine interlude (Clemmensen 1980a).

Møre–mid-Norway margin

Distribution. Rocks of Permian–Triassic age have been proved on the inner part of the Norwegian margin (Fig. 2), extending from the northern North Sea to the Lofoten archipelago, with the bulk of the rocks sampled in the Møre and Vøring regions being of Triassic age (Figs 7 & 8). However, by analogy with the East Greenland conjugate margin (see below), it has been assumed that the platforms and terraces that underlie the Norwegian margin comprise sedimentary rocks of Permian age atop crystalline basement (Blystad *et al.* 1995; Doré *et al.* 1999; Brekke 2000).

Structural setting. In the Møre and Mid-Norwegian areas, the Late Carboniferous–Early Permian was an active tectonic period, and might have generated a block-faulted terrain beneath the Trøndelag Platform and the Halten Terrace, on the inner Vøring margin (Blystad *et al.* 1995). Further rifting, faulting and fault-block rotation was instigated at the Permian–Triassic boundary and continued during the Early Triassic, whereas a reduction in fault activity and tectonic subsidence characterized the Mid- and Late Triassic (Müller *et al.* 2005).

Stratigraphy. On the Vøring margin, a Permian succession – including carbonates of the Zechstein Group – up to about 250 m thick has been proved in commercial wells (e.g. Nordland Ridge/Rødhoy High) (Figs 1, 2 & 7), whereas no wells on the Møre margin have penetrated Permian strata (Fig. 8). The Triassic has been drilled on both the Møre and Vøring margins, with exploration wells commonly recovering breccia, sandstone and claystone: however, the distribution and thickness of these rocks remains uncertain. The Triassic succession is often referred to as the ‘red’ and ‘grey’ beds,

which mark an upwards transition from arid to humid conditions in the Rhaetian (Figs 7 & 8). On the Trøndelag Platform, the red beds are >2.6 km thick, whereas the grey beds reach a thickness of up to 2.5 km.

Boundary relationships with underlying and overlying systems. Brekke (2000) reports a regional late Early Permian unconformity in the platform areas – dated by comparison with East Greenland – but the age of the underlying rocks remains ambiguous (NORLEX 2014; Norwegian Petroleum Directorate 2014). On the Vøring margin, the Triassic–Jurassic boundary appears to be largely conformable, whereas a hiatus is noted on the Møre margin. In the deeper parts of the Møre and Vøring basins, the boundary relationship remain unclear (Figs 13b, c).

Depositional environments. By analogy with the East Greenland margin, it has been suggested that alluvial and fluvial facies existed along the Møre–mid-Norway margin in the early Permian (Brekke *et al.* 2001) (Figs 7 & 8). A shallow carbonate platform became established in the East Greenland region in the Late Permian, and equivalent carbonate deposits (dolomite) of the Zechstein Group are recorded in wells in the NW part of the Trøndelag Platform (inner Vøring margin).

The Triassic rocks on the Møre–mid-Norway margin are dominated by continental fluvial and alluvial facies (Brekke *et al.* 2001; Müller *et al.* 2005). The Early Triassic was characterized by marginal-marine and terrestrial environments, with the latter becoming predominant during the Mid-Triassic. In the Late Triassic, evaporites and organic-rich shales formed in arid, isolated, marine sub-basins, which were replaced by fluvio-lacustrine depositional environments and, ultimately, conditions that were more humid in the latest Triassic – the change from red beds to grey beds.

Faroe–Shetland–northern Rockall–Hebrides margin

Distribution. Permian–Triassic rocks are commonly preserved as unconformity bounded deposits within Late Palaeozoic–Mesozoic basins that underlie the Hebrides and West Shetland shelf areas (Fyfe *et al.* 1993; Stoker *et al.* 1993; Quinn &

Fig. 12. Interpreted geoseismic profiles from the East Greenland margin showing the stratigraphic architecture of the Upper Palaeozoic–Mesozoic succession where proven (e.g. the Jameson Land and Ammassalik basins) and where inferred (e.g. the Koldewey Platform and Danmarkshavn, Thetis and Liverpool Land basins). Profile (a) based on Dinkelmann *et al.* (2010); profiles (b) and (c) based on Tsikalas *et al.* 2005; profile (d) based on Larsen (1990) and Hamann *et al.* (2005); and profile (e) based on Larsen *et al.* (1999). The locations of profiles are given in Figure 1.

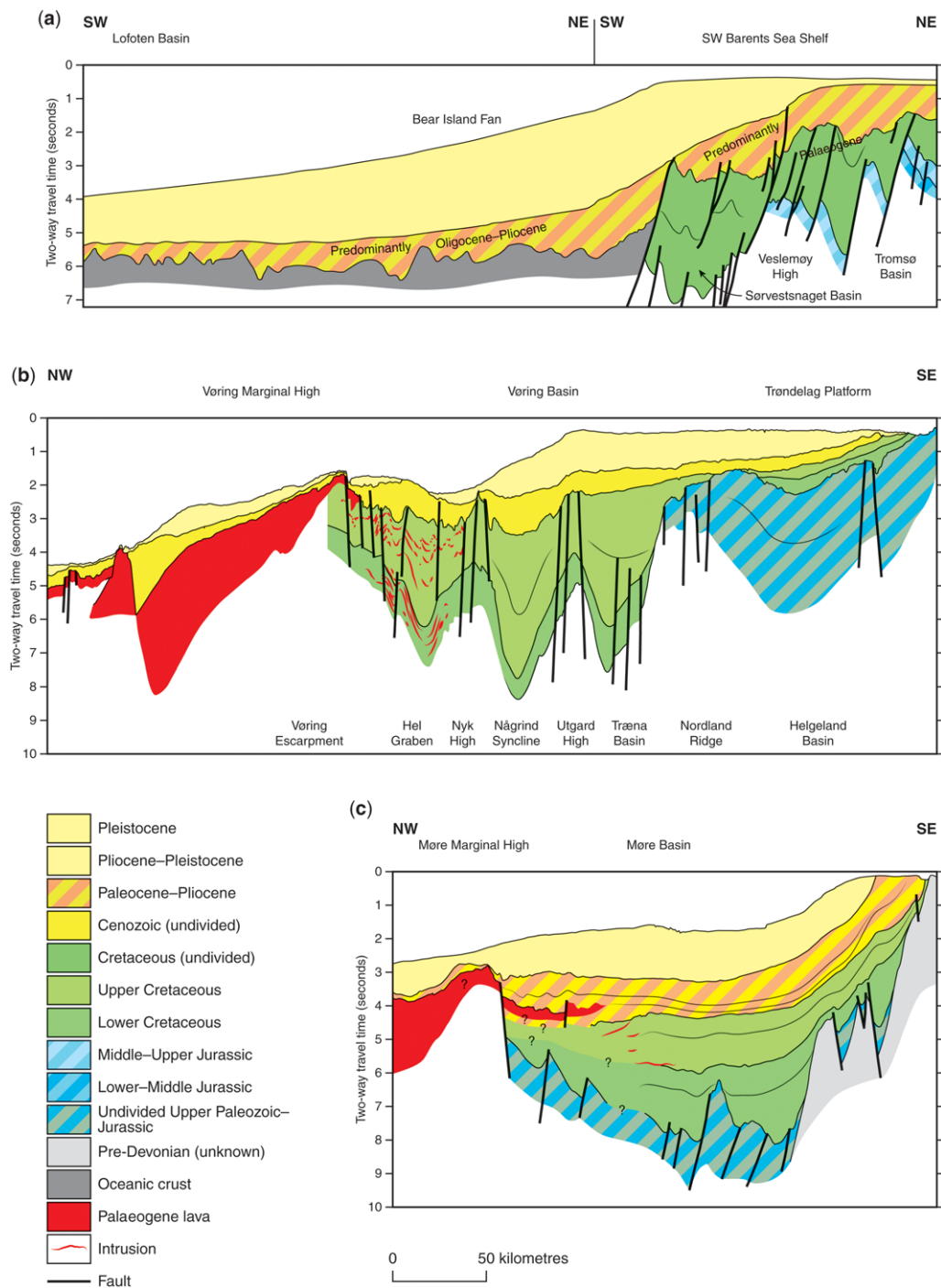


Fig. 13. Interpreted geoseismic profiles from the Mid Norway–SW Barents Sea margin showing the stratigraphic architecture of the Upper Palaeozoic–Mesozoic succession in basins that underlie the SW Barents Sea and mid-Norwegian margin, including the Møre and Vøring basins. Profile (a) based on Ryseth *et al.* (2003) and Hjelstuen *et al.* (2007); profiles (b) and (c) based on Faleide *et al.* (2010) and (c) Brekke (2000). The locations of profiles are given in Figure 1.

THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

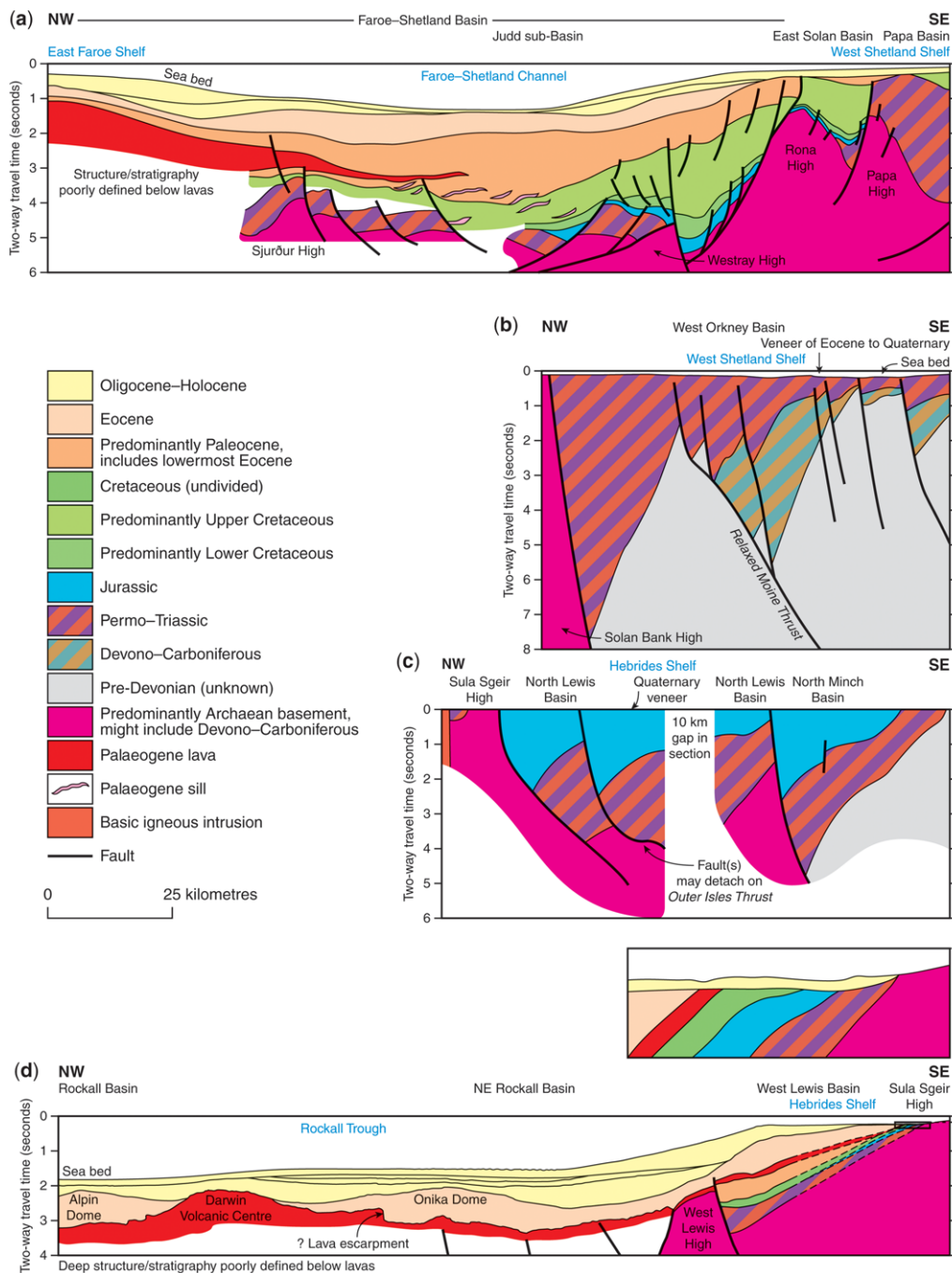


Fig. 14. Interpreted geoseismic profiles from the Faroe–Shetland–northern Rockall–Hebrides margin showing the stratigraphic architecture of the Upper Palaeozoic–Mesozoic succession in basins that underlie the West Shetland and Hebrides shelves, and in the adjacent deep-water Faroe–Shetland and Rockall basins. Profile (a) based on Lamers & Carmichael (1999); profile (b) based on Earle *et al.* (1989) and Ritchie *et al.* (2011a, b); profile (c) compiled from British Geological Survey (1989, 1990) mapping; and profile (d) based on Ritchie *et al.* (2013), with inset showing detail of West Lewis Basin based on Hitchen & Stoker (1993). The locations of profiles are given in Figure 1.

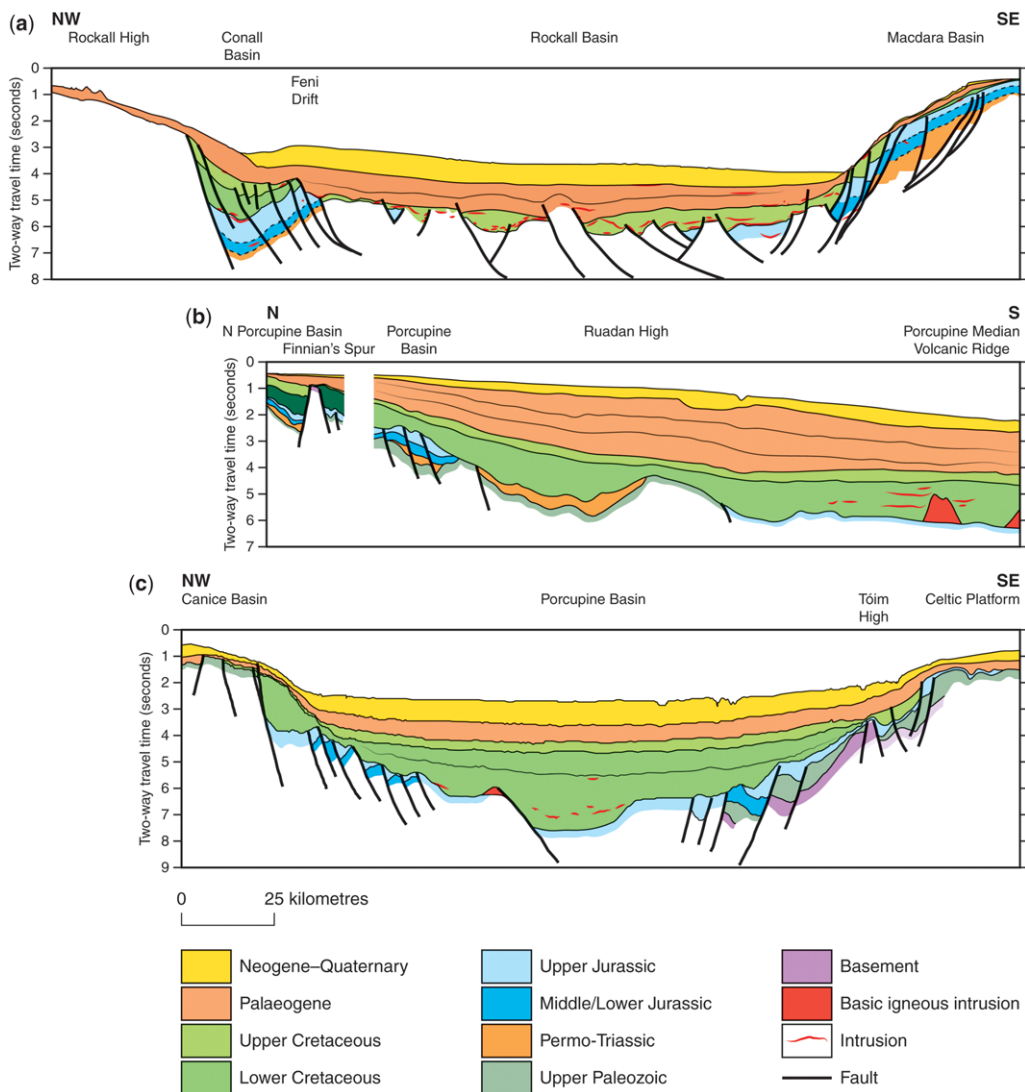


Fig. 15. Interpreted geoseismic profiles from the southern Rockall–Porcupine margin showing the stratigraphic architecture of the Upper Palaeozoic–Mesozoic succession in the Rockall, Porcupine and adjacent marginal basins. Profile (a) based on Naylor *et al.* (1999); profiles (b) and (c) based on Naylor *et al.* (2002). The locations of profiles are given in Figure 1.

Ziska 2011; Johnson & Quinn 2013) (Fig. 2). The thickest accumulations are found in the West Orkney, North Lewis, North Minch, Papa, East Solan and West Shetland basins (Fig. 14). Further west, the distribution is less well constrained. In the Faroe–Shetland Basin, Triassic rocks have been proved in the Judd Sub-basin, and on the Corona and Westray intrabasinal highs. In contrast, no rocks of Permian–Triassic age have been proven to occur in the northern Rockall and Hatton basins.

Structural setting. The Permian–Triassic rocks occupy a series of NE–SW-trending graben and half-graben (Stoker *et al.* 1993; Hitchen *et al.* 1995). The geometry of the preserved successions suggests that the rocks were deposited through a combination of local initial palaeotopographical infill, half-graben syndepositional fault-controlled geometries and uniformly thick sedimentary fill with no obvious syn-sedimentary control (Kirton & Hitchen 1987; Štolfova & Shannon 2009). In

THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

the Hebridean region, the fault-controlled basins (e.g. West Orkney Basin) were largely controlled by listric normal faults downthrowing to the SE: west of Shetland (e.g. West Shetland Basin), however, the faults downthrow to the NW. In some basins (e.g. North Rona Basin), the infill is parallel-bedded with no evidence of thickening towards the present faulted margin. This suggests that this faulting was initiated towards the end of the Permian–Triassic, preserving part of a sequence that originally might have had a greater regional extent, but has been subsequently eroded.

Stratigraphy. Subdivision of the Permian–Triassic succession is poorly constrained over much of the Hebrides–West Shetland shelf areas, where it comprises a largely unfossiliferous clastic assemblage of conglomeratic sandstone, siltstone and mudstone, with only sporadic carbonate, evaporitic and volcanic rocks (Fyfe *et al.* 1993; Stoker *et al.* 1993; Quinn & Ziska 2011; Johnson & Quinn 2013). On the Hebrides Shelf, clastic rocks predominate and are mainly assigned to the Stornoway Formation of Late Permian–Triassic age, although uppermost Triassic (Rhaetian) sandy limestone, preserved locally in the Inner Hebrides, is correlated with the Penarth Group (Storetvedt & Steel 1977; Emeleus & Bell 2005) (Fig. 10). In the North Minch and North Lewis basins, the Stornoway Formation is up to 2 km thick (Fig. 14c). In the West Lewis Basin, an unassigned clastic succession exceeds 0.5 km thick (Fig. 14d).

On the West Shetland Shelf, the Permian–Triassic succession is up to 1 km thick in the East Solan and West Shetland basins, up to 2 km thick in the Papa Basin (Fig. 14a), and up to 7 km thick in the West Orkney Basin (Fig. 14b). In general, a tripartite subdivision has been established for this area: (1) an upper clastic succession assigned to the Triassic Papa Group, which predominates across the region (Booth *et al.* 1993; Ritchie *et al.* 1996; Quinn & Ziska 2011); (2) a middle evaporite succession that has been assigned to the West Orkney Evaporite Formation of the Late Permian Zechstein Group (Ritchie *et al.* 1996); and (3) a lower sequence of sandstone and conglomerate of the Solan Bank Formation, originally assigned by Ritchie *et al.* (1996) to the Zechstein Group, although reassigned by Glennie (2002) to the Upper Rotliegend 2 unit, comparable with the Late Permian sequences in the North Sea Basin (Glennie *et al.* 2003) (Fig. 9).

In the Faroe–Shetland Basin, conglomerate, sandstone and minor mudstone occur in the Judd Sub-basin, and on the Corona and Westray intrabasin high, with sporadic limestone beds in the upper part of the succession (Quinn & Ziska 2011) (Fig. 14a). These rocks have been assigned to the

Middle–Upper Triassic Foula Formation of the Papa Group (Ritchie *et al.* 1996) (Fig. 9).

By way of contrast, trachyandesitic lavas assigned to the Margarita Volcanics and Nun Volcanics formations occur below the Zechstein Group on the East Shetland High and in the Papa Basin, respectively (Quinn & Ziska 2011). These lavas have been assigned an Early Permian age in common with the age range of most of the Rotliegend volcanics of Scotland (Hitchen *et al.* 1995; Ritchie *et al.* 1996; Glennie 2002) (Fig. 9).

Boundary relationships with underlying and overlying systems. The base of the Permian–Triassic succession is everywhere marked by an unconformity. West of Shetland, the youngest rocks below the basal unconformity are of late Early–early Late Carboniferous in age; in the Hebridean region, the underlying rocks are predominantly pre-Devonian basement. Despite a localized conformity between Triassic and Jurassic strata in the Inner Hebrides, an erosional hiatus separates these systems over the rest of the area.

Depositional environment. The rocks of the Upper Rotliegend 2 unit, Stornoway Group and Papa Group were deposited in a semi-arid, fluvial–lacustrine and/or alluvial braidplain setting (Stoker *et al.* 1993; Quinn & Ziska 2011; Johnson & Quinn 2013). The carbonates and evaporites of the Zechstein Group represent brackish–shallow-marine incursions during the Late Permian, whereas the shallow-marine limestones of the Penarth Group, and in the upper part of the Foula Formation, indicate a Late Triassic marine incursion in both the Inner Hebrides and West Shetland regions.

Southern Rockall–Porcupine margin

Distribution. Permian–Triassic strata are preserved throughout most of the Slyne, Erris and Donegal basins, the eastern part of the southern Rockall Basin, and the northern part of the Porcupine Basin (Naylor & Shannon 2011) (Figs 1, 2 & 15). They are likely to be present in the small half-graben basins lying on the eastern and western shoulders of the southern Rockall Basin, and on the eastern margin of the Hatton Basin (Naylor *et al.* 1999; Štolfová & Shannon 2009). Further south, Triassic and, probably, localized Permian strata occur in the Celtic and Fastnet basins (Robinson *et al.* 1981), and are likely to be present in the Goban Spur (Cook 1987).

Structural setting. The Permian–Triassic rocks throughout most of the region are preserved in NE–SW-orientated half-graben controlled by reactivated Caledonian structures (Shannon 1991) (Fig. 2). The successions in the Slyne and Erris basins

occur in narrow, interconnected depocentres that change fault polarity along strike, resulting in the formation of discrete half-graben basins. Croker & Shannon (1987) suggested that Permian–Triassic sediments are preserved in small rift basins along the SW and, possibly, the SE margins of the Porcupine Basin. Naylor *et al.* (1999) interpreted a Permian–Triassic sequence within the group of elongate ‘perched’ Mesozoic basins along the western margin of the Porcupine High (Fig. 2), and Štolfová & Shannon (2009) interpreted these strata, generally thick and parallel bedded, to have been deposited over a broader region than the current fault-preserved basins. Further south, in the Goban Spur–Fastnet–Celtic Sea region (Figs 1 & 2), Permian–Triassic successions are present in broadly east–west-orientated, fault-controlled basins, with evidence of growth faulting and thickening towards the controlling faults (Naylor *et al.* 2002). Rifting of the Variscan basement in the Goban Spur region probably began in Mid-Triassic time (Cook 1987).

Stratigraphy. Definitive Permian strata have not been encountered by drilling in the Porcupine Basin region. Tentatively dated Autunian mudstones with anhydrite have been drilled on the western margin margin of the Porcupine Basin (Croker & Shannon 1987) (Fig. 11). The Triassic is also little known and poorly dated in the north Porcupine Basin (Naylor & Shannon 2011). Here the Triassic succession, approximately 300 m thick, comprises clastic assemblages of sandstones, siltstones and mudstones overlain by evaporite-bearing shales (Fig. 11).

In the Fastnet–Celtic Sea region, no Permian strata have been proven, although clastic deposits may occur in small, isolated half-graben. Here, the Triassic, typically <600 m thick where drilled, consists of a sandstone-prone Sherwood Sandstone equivalent overlain by evaporitic mudstones of Mercia Mudstone equivalence. A similar Triassic stratigraphic succession is interpreted, on seismic evidence, from the Goban Spur (Fig. 11).

In the Slyne Basin, a massive evaporite sequence, assumed to be Zechstein equivalent, lies beneath the Triassic Sherwood Sandstone that is <400 m thick where drilled (Dancer *et al.* 1999, 2005) (Fig. 11). In the Erris Basin, Zechstein-equivalent marine mudstones, dolomites and anhydrites rest unconformably on the Carboniferous, while similar Upper Permian rocks also occur (Tate & Dobson 1989) resting on red-bed sandstones and shales of possible Early Permian age (Murphy & Croker 1992).

On the eastern margin of the southern Rockall Basin, approximately 180 m of Lower Permian (Asselian) sandstones and mudstones is overlain by almost 50 m of sandy, poorly dated redbeds

interpreted as probably Permian–Triassic in age (Tyrrell *et al.* 2010).

Boundary relationships with underlying and overlying systems. The base of the Permian–Triassic succession, which varies in age throughout the region, is everywhere marked by an unconformity. Throughout the Slyne, Erris and Donegal basins, and much of the northern Porcupine Basin, the youngest strata below the basal unconformity are of Upper Carboniferous (Namurian–Westphalian) age, with occasionally Permian–Triassic strata in the north Porcupine Basin resting on pre-Upper Carboniferous strata (Croker & Shannon 1987). Locally in the Celtic Sea region, putative Permian strata are unconformably overlain by the Triassic succession (Smith 1995). The Triassic–Lower Jurassic boundary is typically conformable (Fig. 11). In the eastern part of the southern Rockall Basin, the Permian–Triassic clastic succession is unconformably overlain by probable Middle Jurassic sandstones (Tyrrell *et al.* 2010).

Depositional environment. The Permian, where drilled in the southern Rockall Basin, consists of fluvial channel and aeolian sandstones with lacustrine mudstones, with overlying probably Triassic strata in broadly similar fluvial conglomeratic and sandstone facies. Zechstein-equivalent mudstones, dolomite and evaporitic sequences in the Erris and Donegal basins suggest a narrow marine incursion through these basins (Naylor & Shannon 2011).

Locally, on the western margin of the Porcupine Basin, possible Permian red claystones and siltstones interbedded with sandstones, anhydrites and limestones suggest marginal coastal swamp conditions giving way to brackish and marine conditions, succeeded by lagoonal to playa lake environments.

The Triassic in the Slyne Basin comprises a low-sinuosity braided fluvial sandstone succession with minor sand-flat and playa mudstones. In the Erris Basin, similar facies are recognized with a tuffaceous component seen in some of the sandstones. Within the northern part of the Porcupine Basin, the Triassic was deposited in an alluvial, coastal-plain and shallow-marine environment (Naylor & Shannon 2011).

Jurassic stratigraphy of the NE Atlantic margins

North–NE Greenland margin

Distribution. The Jurassic is exposed onshore in the Wandel Sea Basin and along the NE Greenland margin as far south as the Wollaston Forland (Fig. 3). Offshore, a Jurassic succession has been inferred

for the Danmarkshavn and Thetis basins (Hamann *et al.* 2005; Dinkelman *et al.* 2010; Jackson *et al.* 2012) (Figs 3 & 12a).

Structural setting. The most important Mesozoic rifting phase was initiated in the late Bajocian, culminated in the Volgian–early Ryazanian (Tithonian–Berriasian) and waned in the Hauterivian (Early Cretaceous). The Wollaston Forland Basin underwent significant rotational block faulting in the westwards-tilted half-graben with rift-shoulder uplift and deep subsidence along north–south-trending faults (Surlyk 1978, 2003).

Stratigraphy. The Wandel Sea Basin records the middle Oxfordian–Valanginian Ladegårdsåen Formation, about 250 m thick, of sandstone and organic mudstones in the eastern part of Peary Land (Figs 3 & 5), and, further eastwards, a Kimmeridgian–Valanginian (Lower Cretaceous) succession up to 900 m thick is recorded (Håkansson *et al.* 1991; Dypvik *et al.* 2002). The lithostratigraphic subdivision of this region is currently being revised based on new data.

In the Traill Ø–Hold with Hope area, the basal Jurassic rocks of the Wollaston Forland Basin are represented by mainly sandstone, up to 560 m thick, of the Middle–Upper Jurassic Vardekløft Group (Figs 6 & 7). These rocks are overlain by black organic-rich mudstone of the Upper Jurassic Bernbjerg Formation, which shows prominent thickness variations from 10 to 700 m as a consequence of the rift-basin configuration (Surlyk 1977). The Jurassic–Cretaceous boundary in NE Greenland is incorporated within the Wollaston Forland Group, which consists of up to approximately 3 km thickness of conglomerates and sandstones that laterally pass eastwards into mudstones (Surlyk 1978, 2003) (Fig. 6).

In the offshore basins, the Jurassic succession within the Danmarkshavn Basin has an interpreted thickness of up to 3 km, and has been subdivided into three predominantly clastic megasequences bounded by unconformities (Hamann *et al.* 2005). By analogy with the Wollaston Forland and Jameson Land basins (cf. Figs 6–8), Hamann *et al.* (2005) suggested that the lower megasequence (Lower Jurassic) might be comparable to the Kap Stewart and Neill Klintner groups; the middle megasequence – in the southern Danmarkshavn Basin – has a basal unconformity, and is suggested as a sand-dominated equivalent to the Vardekløft Group; whereas the upper megasequence is tentatively correlated with Upper Jurassic source-rock-prone shales of the Bernbjerg Formation. An uppermost Jurassic–Lower Cretaceous succession is suggested to be equivalent to the Wollaston Forland Group in NE Greenland (Hamann *et al.* 2005) (Fig. 6). A

comparable Jurassic infill to the Thetis Basin is tentatively inferred (Hamann *et al.* 2005).

Boundary relationships with underlying and overlying systems. In the onshore part of the Wandel Sea Basin, Upper Jurassic strata unconformably overlie Triassic or older rocks, and the upper major unconformity is provisionally placed in the Lower Cretaceous and overlain by upper Lower Cretaceous rocks (Håkansson *et al.* 1991) (Fig. 5).

In the Wollaston Forland Basin, the lower boundary of the Middle Jurassic is represented by an erosional unconformity against the Triassic in the Traill Ø area; against the Permian or Caledonian basement further north in the Wollaston Forland; and against the Caledonian basement at Store Koldevey. The unconformity coincides with the initial transgressional base of the Middle Jurassic Vardekløft Group (Surlyk 2003; Stemmerik & Piasecki 2004). On the Koldevey Platform and in the southern Danmarkshavn Basin, the lower boundary is interpreted to be an erosional unconformity against the Triassic (Fig. 6).

The Jurassic succession is conformably overlain by the Cretaceous in the deep downfaulted parts of the Wollaston Forland Basin, whereas a major unconformity exists against Lower Cretaceous along uplifted footwall crests (Fig. 6). A major unconformity is also interpreted to mark the upper boundary on the Koldevey Platform, whereas the upper boundary in the both the northern and southern parts of the Danmarkshavn Basin is inferred to be conformable (Figs 5 & 6).

Depositional environment. In the Wandel Sea Basin, the basal Jurassic facies reflects an Oxfordian marine transgression in the eastern part of Peary Land (Fig. 2). In the Kilen area, the Kimmeridgian–Valanginian succession comprises restricted marine sediments (Håkansson *et al.* 1991; Dypvik *et al.* 2002).

In the Wollaston Forland Basin, the rift-related Vardekløft Group represents a major westwards and northwards expansion of shallow-marine deposition, with the facies belts moving northwards during the Bajocian–Oxfordian, along an echelon north–south-orientated relay embayments (Surlyk 2003). The Upper Oxfordian–Tithonian interval includes the maximum flooding zone in NE Greenland with deposition of black organic-rich mudstones of the Bernbjerg Formation. The Jurassic–Cretaceous Wollaston Forland Group encompasses the rift-culmination succession of deep-marine proximal gravity-flow conglomerates that pass laterally into more distal mudstones within westwards-tilted fault blocks (Surlyk 1978, 2003).

Comparable depositional environments inferred for the upper part of the Jurassic succession in the

Danmarkshavn Basin were probably preceded by Lower–Middle Jurassic lacustrine, deltaic and marginal-marine facies-equivalents of the Kap Stewart and Neill Klintner groups in the Jameson Land Basin of central East Greenland (see below) (Hamann *et al.* 2005) (Fig. 8).

Western Barents Sea–Svalbard margin

Distribution. Rocks of Jurassic age crop out on Svalbard and have been proven in numerous wells in basins on the southern part of the western Barents Sea margin. It is inferred that Jurassic rocks underlie much of the SW part of the Barents Sea continental margin (Fig. 3).

Structural setting. As for most of the Triassic period, Early–Mid-Jurassic deposition in the western Barents Sea–Svalbard region occurred within a tectonically quiet structural setting that was dominated by passive regional subsidence, and only minor fault activity and uplift (Brekke *et al.* 2001; Henriksen *et al.* 2011). A regional Mid-Jurassic unconformity (Fig. 5) marks the onset of a renewed phase of rifting that caused a marked rejuvenation of the topography into a system of tectonic highs and basins, including the Bjørnøya, Tromsø and Harstad basins (Faleide *et al.* 2010) (Figs 1 & 13a). Intra-Upper Jurassic unconformities are recorded in wells from the Hammerfest Basin and, together with major thickness variations in the Bjørnøya Basin and onshore in Svalbard, indicate further contemporary Late Jurassic tectonism along the western Barents Sea–Svalbard margin (Nøttvedt *et al.* 1993a; Doré *et al.* 1999; Worsley 2008; Henriksen *et al.* 2011). This rift phase culminated in the Early Cretaceous (see below).

Stratigraphy. The Upper Triassic–Middle Jurassic Kapp Toscana Group exceeds a thickness of 475 m on Svalbard, and consists of deltaic to shallow-marine shales, siltstones and sandstones (Harland 1997; Dallmann 1999; NORLEX 2014) (Fig. 5). On the western Barents Sea margin, the Kapp Toscana Group is up to 500 m thick (e.g. Hammerfest Basin), and is dominated by a paralic to shallow-marine succession of sandstone with subordinate shale and thin coal beds, which pass upwards into moderately to well sorted and mineralogically mature sandstones (Dalland *et al.* 1988; Dallmann 1999; NORLEX 2014).

The post-Bathonian Middle–Upper Jurassic succession is assigned to the Adventalen Group, which extends into the Early Cretaceous (Nøttvedt *et al.* 1993a; Dallmann 1999; NORLEX 2014). On Svalbard, the Jurassic component of this group is up to 350 m thick and is marked at its base by a conglomerate – the Brentskardhaugen Bed – a reworked

deposit that is overlain by a poorly sorted upwards-fining basal section of conglomerate, sandstone and siltstone, which pass upwards into dark grey to black organic-rich marine shales with subordinate thin siltstones (Fig. 5). This succession extends into the western Barents Sea where it is up to 470 m thick, and comprises predominantly dark-brown pyritic mudstones, with thin limestones and sporadic thin sandstone beds around basin margins. Thickness variations are observed in some basins (e.g. Hammerfest Basin) as a consequence of contemporary tectonic activity (Dallmann 1999).

Boundary relationships with underlying and overlying systems. In Svalbard, the Triassic–Jurassic boundary is marked, at least locally, by seemingly continuous facies from the Rhaetian to the Toarcian, although elsewhere (e.g. Hammerfest Basin) local unconformities exist (Fig. 5). The Jurassic–Cretaceous boundary is represented by an unconformity in the western Barents Sea, whereas a conformable boundary occurs on Svalbard (Nøttvedt *et al.* 1993a; Harland 1997; Dallmann 1999; Henriksen *et al.* 2011).

Depositional environments. The repeated clastic succession of mainly delta-related and coastal and shallow-marine shelf sedimentation that was instigated in the Late Triassic (the Kapp Toscana Group) persisted until the Mid-Jurassic (Bathonian), sourced from multiple provenances. At this time, the palaeogeography of the western Barents Sea–Svalbard region was probably dominated by a low-lying peneplain (Brekke *et al.* 2001). Extensive reworking within this succession has resulted in sandstones that are texturally and mineralogically mature (Worsley 2008; Henriksen *et al.* 2011). Following late Mid–Late Jurassic topographical rejuvenation, a renewed regional transgression led to the submergence of a large part of the western Barents Sea–Svalbard region, and the deposition of deeper-water, muddy marine shelf sediments of the Adventalen Group (Dallmann 1999; Henriksen *et al.* 2011). A fluctuating sea level in combination with the increased tectonic activity resulted in variable oxic and anoxic bottom-water conditions, which gave rise to favourable conditions for the accumulation of black shales (Brekke *et al.* 2001).

Central East Greenland margin

Distribution. Jurassic rocks are exposed in the onshore Jameson Land Basin (Figs 3 & 12d). No Jurassic is exposed onshore south of this basin, although seismic data show that the southwards-dipping Jurassic succession may extend below the volcanic sequence in the Blossville Coast (Larsen & Marcussen 1992) (Fig. 1). Seismic reflection

data suggest that Jurassic rocks might be present in the offshore Liverpool Land Basin (Hamann *et al.* 2005) (Fig. 12d): however, it remains unclear as to whether or not they are present in the Ammassalik Basin, offshore SE Greenland (Gerlings *et al.*, this volume, in review).

Structural setting. The latest Triassic (Rhaetian)–Mid-Jurassic (early Bajocian) was characterized by thermal subsidence in the Jameson Land Basin. The most important Mesozoic rifting phase was initiated in the late Bajocian, culminated in the Volgian–early Ryazanian (Tithonian–early Berriasian (earliest Cretaceous)) and waned in the Hauterivian (Early Cretaceous) (Surlyk 2003). The Jameson Land area only experienced minor tilting during this period, in comparison to NE Greenland (see above) (Surlyk 2003). The Lower Jurassic facies belts follow deep-seated Devonian NW–SE-striking faults in the northern part of Jameson Land (Dam *et al.* 1995).

Stratigraphy. The Jameson Land Basin preserves a near-complete Jurassic succession of up to about 2 km thick; the stratigraphy and basin evolution is comprehensively described by Surlyk (2003). The Rhaetian–Sinemurian Kap Stewart Group is up to 600 m thick, and consists of organic-rich mudstones and sandstones (Dam *et al.* 1995; Surlyk 2003) (Fig. 8). The Kap Stewart Group is overlain by alternating sandstones and mudstones of the Pliensbachian–early Bajocian Neill Klintner Group, which is 300–450 m thick and topped by an erosional unconformity (Dam & Surlyk 1998). This is succeeded by sandstones of the Vardekløft Group (up to about 650 m thick), which are overlain by Oxfordian–Tithonian black organic-rich mudstones and sandstones of the Hall Bredning Group (possibly up to 800 m thick), itself capped by coarse-grained sandstones (a few hundred metres thick) of the Scoresby Sund Group, which extends into the Berriasian (Early Cretaceous) (Surlyk 2003). In the Liverpool Land Basin, the presence of Upper Jurassic marine shales has been speculated upon (Gautier *et al.* 2011) (Fig. 8).

Boundary relationships with underlying and overlying systems. In the Jameson Land Basin, the base of the Kap Stewart Group is associated with a minor unconformity against the Triassic Scoresby Land Group along the SE margin of the Jameson Land Basin. The Jurassic–Cretaceous boundary is a minor erosional unconformity that extends into the Barremian (earliest Cretaceous), and is overlain by Valanginian rocks.

Depositional environment. The Triassic–Jurassic boundary marks a transition towards a temperate and humid climate in the Jameson Land Basin,

represented by the deltaic and lacustrine rocks of the Kap Stewart Group (Dam *et al.* 1995; Surlyk 2003) (Fig. 8). The sediments of the Neill Klintner Group represent marine-shoreface, restricted offshore and tidal-embayment environments deposited during an overall Pliensbachian–Early Bajocian transgression (Dam & Surlyk 1998). The Vardekløft Group represents mainly shallow-marine deposition that is succeeded by deposition of the Hall Bredning Group of basinal organic-rich mudstones and mass-flow sandstones during maximum sea level. The southwards progradation of prominent sandy shelf-edge deltas is represented by the sandstones of the Raukelv Formation (Surlyk 2003).

Møre–mid-Norway margin

Distribution. On the Norwegian mainland, Jurassic sedimentary rocks crop out on Andøya, northern Norway, where an approximately 350 m-thick sequence of sandstones, siltstones, shales and thin beds of coal is preserved (Øvrig 1960; Bøe *et al.* 2010). Offshore, Jurassic sediments have been proved in many of the wells drilled on the inner part of the Møre–mid-Norway margin (Figs 3 & 13b, c). On the continental shelf, most wells that prove Jurassic rocks have been drilled on the Trøndelag Platform and Halten Terrace, in the inner Vøring region. Where the wells penetrate the entire preserved Jurassic succession, thicknesses in excess of 1000 m are recorded: more commonly, the drilled succession ranges between 400 and 800 m.

Structural setting. It has been inferred that the mid-Norway margin, as well as the Møre region, was predominantly an area of uplift and erosion throughout the Early and Mid-Jurassic (Brekke *et al.* 1999; Doré *et al.* 1999). During the Late Jurassic–Early Cretaceous, rifting and differential vertical movements created sub-basins and highs in the Møre and Mid-Norwegian region (Faleide *et al.* 2010). Correspondingly, the thickness of the rocks may vary considerably as the sediments were deposited on a series of tilted fault blocks.

Stratigraphy. The Jurassic succession on the continental shelf comprises the Båt (latest Triassic (Rhaetian)–Early Jurassic), Fangst (Mid-Jurassic) and Viking (late Mid-Jurassic–earliest Cretaceous) groups (Dalland *et al.* 1988) (Figs 7 & 8). The Båt Group is best preserved on the inner Vøring margin where it locally exceeds 700 m in thickness, and consists predominantly of alternating sandstone and shale–siltstone units with thin coal beds. The upper part of the Båt Group is progressively truncated towards the crestal region of some of the inner-shelf highs (e.g. the Nordland Ridge), and younger strata within the group onlap directly onto

Precambrian basement. The Fangst Group is up to 250 m thick, and comprises a basal fine- to medium-grained sandstone unit, middle mudstone unit and an upper fine- to coarse-grained sandstone unit (Dalland *et al.* 1988). The Viking Group is best developed in the northern North Sea where it is up to 1 km thick: in contrast, it is more locally developed in the Møre–mid-Norway region, where a succession dominated by shales and mudstones up to 125 m thick has been drilled in commercial wells (Dalland *et al.* 1988). Well data also indicate a variable conformable to unconformable relationship with the underlying Fangst Group, as well as internal unconformities within the Viking Group.

Boundary relationships with underlying and overlying systems. On the inner Møre margin, pre-Toarcian rocks are locally absent, and the Jurassic sequence rests unconformably on Triassic and older strata (Fig. 8). On the inner part of the Vøring margin, a more conformable passage from Triassic to Lower Jurassic deposits is observed (Fig. 9). The boundary with the overlying Lower Cretaceous Cromer Knoll Group is commonly marked by an unconformity throughout the region (Dalland *et al.* 1988; Norwegian Petroleum Directorate 2014).

Depositional environments. The Jurassic rocks on Andøya represent lacustrine–lagoonal to shallow-marine depositional environments (Ørving 1960; Bøe *et al.* 2010). Deltaic to shallow-marine environments also characterize the latest Triassic to Mid-Jurassic succession on the Møre–mid-Norwegian margin, whereas an open-marine environment existed in the late Mid-Jurassic and Late Jurassic (Dalland *et al.* 1988). The observed variable conformable to unconformable relationship within the Fangst and Viking groups probably reflects the transgressive nature of the upper Middle–Upper Jurassic sequence at a time of transition into an increasingly unstable Cretaceous rifting episode (Blystad *et al.* 1995; Doré *et al.* 1999; Brekke *et al.* 2001).

Faroe–Shetland–northern Rockall–Hebrides margin

Distribution. Most proven occurrences of Jurassic rocks are confined to the basins that underlie the Hebrides and West Shetland shelves (Fyfe *et al.* 1993; Stoker *et al.* 1993; Ritchie & Varming 2011; Evans 2013) (Figs 1 & 3). A near-complete succession is preserved within the Hebridean chain of basins, including the Inner Hebrides, Sea of Hebrides–Little Minch, North Minch and North Lewis basins (Figs 10 & 14c). They have also been encountered in the West Lewis (Fig. 14d) and West Flannan basins. Upper Jurassic rocks are

widely present in the North Rona, Papa, West Shetland, West Solan, East Solan (Fig. 14a) and South Solan basins west of Shetland, as well as the Judd, Flett and Foula sub-basins in the Faroe–Shetland Basin (Fig. 9). Their occurrence on the Corona High might indicate a presence in the western half of the Faroe–Shetland Basin, but this remains conjectural. Middle and Lower Jurassic strata have a more restricted distribution across the Faroe–Shetland area. No rocks of Jurassic age have been proven to occur in the northern Rockall and Hatton basins, although the occurrence of Jurassic palynomorphs reworked into Eocene fan-delta deposits on the eastern flank of the Rockall High might reflect a Jurassic source on the Rockall Plateau (Hitchen 2004).

Structural setting. The Jurassic rocks are largely preserved within the same system of NE–SW-trending basins that was active during the Permian–Triassic (Fig. 3). In the Hebrides–West Shetland region, the geometry of the Jurassic basinal successions is asymmetrical and thickest adjacent to the footwall (e.g. North Lewis, North Minch, North Rona and East Solan basins) (Fig. 14) – a pattern of sedimentation that has been linked to pulsed episodes of footwall uplift within this Mesozoic basin system (Harris 1992; Morton 1992; Roberts & Holdsworth 1999). The isolated occurrences of Lower and Middle Jurassic rocks in the Faroe–Shetland Basin make it difficult to determine basin activity at this time. According to both Dean *et al.* (1999) and Doré *et al.* (1999), the general lack of variation in the thickness of the Upper Jurassic strata across the Faroe–Shetland region (see below) suggests that the effects of any contemporaneous rift activity were negligible.

Stratigraphy. On the Inner Hebridean islands of Skye and Raasay (part of the Sea of Hebrides–Little Minch Basin), the Jurassic succession is up to 1.5 km thick and is divisible into three unconformity bounded sequences: the Lower Jurassic Lias Group, comprising shallow-marine clastic rocks with limestone and sporadic ironstone formations; shallow-water sandstones and partly non-marine organic-rich rocks of the Middle Jurassic Berreraig Sandstone Formation and Great Estuarine Group, respectively; and shallow-marine Upper Jurassic sandstones and mudstones of the Staffin Bay and Staffin Shale formations (Fyfe *et al.* 1993; Hudson & Trewin 2002) (Fig. 10). A comparable succession of rocks ranging from 2 to 3 km in thickness has been proved in the North Lewis and North Minch basins, including Middle Jurassic organic-rich mudstone (Fyfe *et al.* 1993; Stoker *et al.* 1993; Evans 2013) (Fig. 10). Further west, up to several hundred metres of Middle and Upper Jurassic marginal- to shallow-marine sandstone and mudstone with

sporadic thin coal beds might be present in the West Flannan and West Lewis basins (Hitchen & Stoker 1993) (Fig. 14d).

In the Faroe–Shetland region, the Lower Jurassic sequence is separated from the Middle and Upper Jurassic sequences by a major late Early–early Mid-Jurassic (mid-Pliensbachian–Bajocian) hiatus (Ritchie *et al.* 1996; Ritchie & Varming 2011) (Fig. 9). The Lower Jurassic rocks have been grouped collectively within the Skerry Group, which comprises carbonaceous sandstone and mudstone, with maximum-drilled thicknesses of 770 m in the West Solan Basin and 340 m in the Foula Sub-basin (Faroe–Shetland Basin). The upper Middle–Upper Jurassic succession is assigned to the Humber Group, which has a maximum-drilled thickness of 1.05 km in the Foula Sub-basin, although more commonly it is <250 m thick across the Faroe–Shetland region. This group comprises sporadic shallow-marine sandstone and carbonaceous and pyritic marine mudstone of the Bajocian–Oxfordian Heather Formation overlain, locally unconformably, by the Kimmeridge Clay Formation, which is the most extensive Jurassic unit within the Faroe–Shetland region, and extends into the Early Cretaceous (Berriasian). Although the Kimmeridge Clay Formation is typically associated with dark grey to black organic-rich mudstone and minor siltstone, the base of the formation is characterized by sandstone, siltstone and conglomerate.

Boundary relationships with underlying and overlying systems. In the Inner Hebrides there appears to be conformity, at least locally, between the Upper Triassic and Lower Jurassic sequences (Fyfe *et al.* 1993; Hudson & Trewin 2002) (Fig. 10). Over the rest of the area, the base of the Jurassic succession mostly rests unconformably on Triassic and older strata. In the Faroe–Shetland region, the Kimmeridge Clay Formation extends into the early Berriasian (Cretaceous); however, it is almost everywhere separated from the overlying Cromer Knoll Group (Lower Cretaceous) by an unconformity (Ritchie & Varming 2011; Stoker & Ziska 2011). On the Hebrides Shelf, the top of the Jurassic is unconformably overlain by Cretaceous and younger rocks.

Depositional environment. The rocks of the Lias and Skerry groups were deposited in marginal- to shallow-marine environments. In the Hebrides region, interbedded mudstones, siltstones, sandstones and limestones of the Lias Group are arranged in transgressive–regressive facies cycles, whereas thicker, deeper-water mudstone formations are interpreted to reflect contemporaneous subsidence linked to a high clastic input (Fyfe *et al.* 1993; Hesselbo *et al.* 1998; Hudson & Trewin 2002).

Following the early Mid-Jurassic hiatus, marginal- to shallow-marine sedimentation was renewed in the Hebrides and was succeeded by fully marine sedimentation in the Late Jurassic, instigated by a major rise in sea level in the Callovian. The Faroe–Shetland region might have remained largely exposed until the widespread deposition of the Kimmeridge Clay Formation. According to Vestralen *et al.* (1995), the Kimmeridge Clay Formation represents the deposition of an overall transgressive succession marked by the transition from subaerial and shallow-marine coarse clastics to organic-rich basinal mudstones. This model contrasts with previous deep-water rift-related models in the Faroe–Shetland area (e.g. Haszeldine *et al.* 1987; Hitchen & Ritchie 1987; Meadows *et al.* 1987).

Southern Rockall–Porcupine margin

Distribution. Jurassic strata have a widespread distribution throughout this region (Figs 3 & 15). Lower Jurassic rocks occur through the Slyne and Erris basins and the Goban Spur, but have a very restricted distribution in the Porcupine region, encountered only in the north Porcupine Basin (Fig. 11). Middle Jurassic strata are widely present in the Slyne, Erris, Porcupine and Goban Spur basins. Upper Jurassic strata are present basinwide through the Porcupine Basin (Croker & Shannon 1987), locally preserved in the Slyne Basin (Corcoran & Mecklenburgh 2005), but generally absent through most of the Erris and Donegal basins. They are very locally present in the Fastnet Basin (Robinson *et al.* 1981) and the Goban Spur due to erosion, although seismic evidence (Cook 1987; Naylor *et al.* 2002) suggests Upper Jurassic strata may be preserved in places. The nature of Jurassic deposition within the southern Rockall and Hatton basins is speculative, although Naylor & Shannon (2005) suggested a widespread presence. Kimmeridgian–Portlandian (late Tithonian–earliest Berriasian (early Cretaceous)) marine sandstones and limestones have been drilled in the north Bróna Basin (Fig. 1) on the western flank of the Porcupine High (Haughton *et al.* 2005). Middle and Upper Jurassic strata are also interpreted from seismic data within the small perched basins along both margins of the southern Rockall Basin (Naylor *et al.* 1999) (Fig. 15a).

Structural setting. Croker & Shannon (1987) suggested that Early Jurassic deposition was limited through the Porcupine and southern Rockall region to isolated fault-controlled basins with a Caledonian trend, mirroring the pattern for the underlying Triassic. These NE–SW-trending basins are larger in the Slyne–Erris region, comprising linked half-graben with a reversal of fault polarity along the basin

trend. In the Goban Spur region, Jurassic sedimentation was controlled by extensional reactivation of structures at a high angle to the Caledonian fabric that controlled coeval sedimentation further north along the Porcupine–southern Rockall margin. Middle and Upper Jurassic sedimentation in the Porcupine Basin region was controlled by east–west rifting, resulting in a large fault-bounded basin system which transected the NE–SW caledonoid grain that predominated in the Slyne and Erris region to the north.

Stratigraphy. A fairly complete Lower Jurassic sequence, approximately 600 m thick, is preserved in the Slyne Basin, whereas the upper zones of the Lower Jurassic are missing, through erosion, in wells in the Erris Basin (Fig. 11). Hettangian–Toarcian limestones, siltstones and basal anhydrites are overlain by sandstones, siltstones and capped by up to 100 m of organic-rich shales. The stratigraphy is comparable to the Hebridean region described above (Trueblood & Morton 1991). A broadly similar succession has been encountered in the Goban Spur region.

The Middle Jurassic in the Slyne Basin comprises more than 1300 m of Bajocian–Bathonian claystones, limestones and sandstones with coal interbeds. The Middle and Upper Jurassic succession in the Porcupine Basin comprises coarse clastics in fining-up cycles of Bajocian–early Bathonian age, overlain in the north by sandstones and mudrocks (Croker & Shannon 1987). In the southern Rockall Basin, approximately 120 m of conglomeratic sandstones, overlain by thin marine sandstones and capped by a volcanic unit (Tyrrell *et al.* 2010), are thought to be of Middle Jurassic age. In the Porcupine Basin, the Oxfordian–Tithonian comprises mudstones with thin argillaceous siltstones and sandstones (Croker & Shannon 1987). Thicker sandstones occur towards the base of the succession. There are significant differences in facies and clastic geometries between the northern, central and southern margins of the basin that reflect different depositional settings. Up to 500 m of Upper Jurassic strata are locally preserved in the Slyne Basin, and comprise silty claystones with sandstone, limestone and coal interbeds (Corcoran & Mecklenburgh 2005).

Boundary relationships with underlying and overlying systems. Where the uppermost Triassic succession has been encountered in the region, it is marked by a rapid transition from redbed mudstones to a shallow-marine limestone–shale succession of Rhaetian age. The regionally developed Middle Jurassic sequences in the northern part of the Porcupine Basin (Fig. 11) rest unconformably on older, mostly Upper Carboniferous, strata (Naylor

& Shannon 1982; Croker & Shannon 1987). The Upper Jurassic rests conformably on the Middle Jurassic and is regionally unconformably overlain throughout the region by Lower Cretaceous strata.

Depositional environment. The Lower Jurassic succession is largely brackish to marine in nature. In contrast, the Middle Jurassic is predominantly a fluvio-estuarine succession dominated by sandstones and siltstones. In the northern and marginal parts of the Porcupine Basin, the Bajocian–early Bathonian braided river systems prograded southwards, and are succeeded by meandering fluvial and related environments. They pass southwards to shallow-marine sediments (Croker & Shannon 1987). The Late Jurassic was a period of differential subsidence in the basin and the overall facies pattern suggests a northwards encroachment of transgressive facies (Croker & Shannon 1987) along the basin axis, with active movement on marginal bounding faults. Marine influences increased, and nearshore marine conditions became dominant in Kimmeridgian–early Tithonian time. Away from the basin margins, proximal to distal submarine fan sandstones and mudstones are overlain by Tithonian marine shales (Croker & Shannon 1987).

Cretaceous stratigraphy of the NE Atlantic margins

North–NE Greenland margin

Distribution. In north Greenland, the Cretaceous is exposed in Peary Land and at Kilen in the Wandel Sea Basin (Figs 1 & 4). Further south, Cretaceous rocks are exposed along the NE Greenland margin from Store Koldewey to Traill Ø, whereas the offshore Danmarkshavn and Thetis basins have both been interpreted to contain Cretaceous successions, several kilometres thick (Hamann *et al.* 2005).

Structural setting. The Wandel Sea Basin was probably subdivided into several fault-bounded sub-basins in the Cretaceous that were increasingly controlled by strike-slip forces (pull-apart basins) (Håkansson *et al.* 1991; Dypvik *et al.* 2002). The Danmarkshavn Basin shows a relatively symmetrical basin configuration in the northern part and an eastwards-tilted half-graben against the Danmarkshavn Ridge in the south (Figs 12a–c). In contrast, the Thetis Basin is interpreted to consist of a westwards-tilted half-graben against the Danmarkshavn Ridge (Fig. 12a). Post-Valanginian basin infills along the NE Greenland margin, including the Wollaston Forland Basin, reflect partly inherited Jurassic–Cretaceous basin configurations. In addition, several episodes of fault reactivations

have been suggested to explain the local deposition of coarse-grained units (e.g. Surlyk & Noe-Nygaard 2001).

Stratigraphy. The post-Valanginian Cretaceous siliciclastics in the Wandel Sea Basin comprise the 650 m-thick Gåseslette Group (Aptian–Albian), unconformably overlain by the Kilen Group (Turonian–Santonian), which is probably more than 1900 m thick (Håkansson *et al.* 1991; Dypvik *et al.* 2002).

The post-Valanginian Cretaceous succession exposed from Traill Ø to Store Koldewey is up to a few kilometres thick and is dominated by mudstone, although several coarse-grained units can be identified, including the Barremian–Aptian Steensby Bjerg Formation at Hold with Hope (Whitham *et al.* 1999; Larsen *et al.* 2001), as well as the Albian Rold Bjerge Formation, the Turonian Månedal Formation and the Coniacian Vega Sund Formation at Traill Ø (Surlyk & Noe-Nygaard 2001), all of which are included within the Hold with Hope Group (Fig. 7).

In the offshore basins, the southern part of the Danmarkshavn Basin is interpreted to contain a shale-dominated Cretaceous succession up to about 4 km thick, which is divided by an internal unconformity that roughly separates the Lower and Upper Cretaceous sequences (Tsikalas *et al.* 2005) (Figs 6 & 12a–c). The Cretaceous succession in the northern Danmarkshavn Basin is affected by salt movement (Fig. 5). In the Thetis Basin, a Cretaceous succession several kilometres thick is inferred (Figs 6 & 12a). The succession thickens markedly to the north of the profile (shown in Fig. 12a) and it is possibly analogous to the Vøring Basin offshore mid-Norway (Hamann *et al.* 2005; Dinkelman *et al.* 2010) (Fig. 7).

Boundary relationships with underlying and overlying systems. The onshore part of the Wandel Sea Basin records an unconformity bounded Lower Cretaceous (Aptian–Albian) succession overlying Valanginian or older Mesozoic–Palaeozoic strata (Fig. 5). The Upper Cretaceous comprises unconformity bounded Turonian–Santonian units: Paleocene deposits occur onshore, but the contact relationship with the Cretaceous is not known (Håkansson *et al.* 1991; Dypvik *et al.* 2002).

The post-Valanginian sequences between Traill Ø to Store Koldewey display lower unconformities against Caledonian basement, Carboniferous, Permian, Triassic, Jurassic or lowermost Cretaceous strata. The upper boundaries are controlled by the results of post-Cretaceous uplift events: in some places, the Cretaceous is erosionally overlain by Paleocene sediments and Palaeogene volcanics (e.g. Nøhr-Hansen *et al.* 2011).

The Danmarkshavn and Thetis basins have an interpreted conformable lower boundary against the Jurassic, and an upper boundary that is marked by an angular unconformity: the latter is overlain by Quaternary deposits to the west, and pre-break-up Palaeogene strata to the east along and flanking the Danmarkshavn Ridge (Hamann *et al.* 2005; Tsikalas *et al.* 2005) (Fig. 12a).

Depositional environment. The Ladegårdsåen, Gåseslette and Kilen groups of the Wandel Sea Basin all represent siliciclastic rocks deposited in a restricted marine setting (Håkansson *et al.* 1991; Dypvik *et al.* 2002). Several episodes of rift reactivation have been suggested to explain local deposition of coarse-grained material as preserved by the shallow to deeper-water coarse-grained units, including gravity-flow deposits between Hold with Hope (Whitham *et al.* 1999; Larsen *et al.* 2001) and Traill Ø (Surlyk & Noe-Nygaard 2001) (Figs 6 & 7).

The Danmarkshavn Basin is interpreted to have been a mudstone-dominated marine basin (Figs 5 & 6). Deep-marine gravity-flow sandstones sourced from the west are inferred to have been deposited exclusively in the southern part of the basin during two episodes in the Early and mid-Cretaceous. The Cretaceous succession in the Thetis Basin is similarly interpreted to have been dominated by marine-basinal mudstone deposition (Hamann *et al.* 2005) (Fig. 6).

Western Barents Sea–Svalbard margin

Distribution. Cretaceous rocks crop out in Svalbard and are extensively preserved on the southern part of the western Barents Sea margin: their occurrence west of Svalbard remains uncertain (Figs 4, 5 & 13a). Whereas an accumulation of up to 3 km of Lower–Upper Cretaceous rocks has been identified on the SW Barents Sea margin, most of the greater Barents Sea region, including Svalbard, preserves only Lower Cretaceous rocks (Faleide *et al.* 1993; Breivik *et al.* 1998; Brekke *et al.* 2001).

Structural setting. Increasing tectonic activity throughout the Late Jurassic culminated with rifting in the Early Cretaceous, and the establishment of the present-day structural configuration of basins and highs in the western Barents Sea region (Gabrielsen *et al.* 1990). This included the rapid subsidence and development of deep basins, such as the Harstad, Tromsø, Bjørnøya and Sørvestnaget basins (Figs 1 & 13a), which became decoupled from the rest of the Barents Sea shelf through a series of rift episodes in the Early–mid-Cretaceous (Aptian–Albian, Cenomanian?) (Smelror *et al.* 2009) (Fig. 5). In the Late Cretaceous, much of the NW Barents

Sea region, including Svalbard, was uplifted (Brekke *et al.* 2001), whereas the southern western Barents Sea margin continued to subside (Dallmann 1999; Smelror *et al.* 2009; Faleide *et al.* 2010).

Stratigraphy. The Lower Cretaceous succession on Svalbard and across the western Barents Sea margin is assigned to the upper part of the Adventalen Group (which also includes the Upper Jurassic, see earlier) (Dallmann 1999; NORLEX 2014): Upper Cretaceous rocks on the western Barents Sea margin are assigned to the Nygrunnen Group (Dalland *et al.* 1988) (Fig. 5). An unconformity separates the Adventalen and Nygrunnen groups over much of the western Barents Sea margin (Fig. 5).

On Svalbard, the Cretaceous rocks of the Adventalen Group exceed 1500 m in thickness, and consist predominantly of dark shale with clay-ironstone and siltstone nodules that coarsen upwards into siltstone and sandstone with thin coal beds topped by a wedge-shaped accumulation of interbedded sandstones and shales (Dallmann 1999; Brekke *et al.* 2001; Worsley 2008) (Fig. 5). Between central and southernmost Svalbard, this clastic wedge thickens from 190 m to over 1000 m, increasing further in thickness on the western Barents Sea margin where the rocks mainly include dark marine mudstones with sporadic thin beds of limestone, dolomite, sandstone and siltstone (Dalland *et al.* 1988; Dallmann 1999; NORLEX 2014). In the Tromsø Basin, in excess of 2 km of unconformity bounded Lower Cretaceous rocks have been identified, whereas 700–1000 m of Lower Cretaceous strata have been reported from the Hammerfest Basin. Further west, thicknesses exceeding 2 km can be inferred from seismic data beneath the Veslemøy High and Sørvestsnaget Basin beneath the western Barents Sea shelf (Ryseth *et al.* 2003; Henriksen *et al.* 2011) (Fig. 13a).

Early Cretaceous igneous activity is evidenced by the occurrence of intercalated basalt lava flows of Barremian–Aptian age in Kong Karls Land (eastern Svalbard) (Harland 1997; Dallmann 1999; Henriksen *et al.* 2011; NORLEX 2014) (Fig. 5), with doleritic intrusions within Triassic–Lower Cretaceous shales. Sporadic tuffs are noted in the Hammerfest Basin and on the Senja Ridge.

On the western Barents Sea margin, the Upper Cretaceous Nygrunnen Group exceeds 2 km in thickness in the Tromsø and Sørvestsnaget basins, beneath the outer margin (Henriksen *et al.* 2011) (Fig. 13a). This group consists mostly of greenish-grey to grey shales and mudstones with thin interbeds of limestone, and a tuffaceous component is locally preserved in the Tromsø Basin (Dallmann 1999; NORLEX 2014). These fine-grained clastic deposits pass eastwards into condensed calcareous to sandy units (Dallmann 1999). These

strata are commonly truncated to the east (Henriksen *et al.* 2011).

Boundary relationships with underlying and overlying systems. The Jurassic–Cretaceous boundary is represented by a distinct unconformity in the western Barents Sea, whereas continuous sedimentation prevailed across this boundary on Svalbard (Fig. 5). The Cretaceous–Paleocene boundary is characterized by a significant erosional unconformity throughout much of the western Barents Sea–Svalbard region (Henriksen *et al.* 2011).

Depositional environments. On Svalbard, the Lower Cretaceous rocks represent a paralic to shallow-marine shelf succession that records an overall regressive succession deposited under oxic conditions in an open-marine shelf setting. The clastic wedge resulted from a relative sea-level fall and delta progradation from the north in response to a major uplift of the NW Barents Sea area associated with the break-up of the Amerasian Basin (present Arctic Sea) (Nøttvedt *et al.* 1993a, b; Dallmann 1999; Brekke *et al.* 2001; Worsley 2008). This was accompanied by magmatism in Kong Karls Land and Franz Josef Land (Arctic Ocean, NE of Svalbard) (Steel & Worsley 1984; Harland 1997; Dallmann 1999; Henriksen *et al.* 2011; NORLEX 2014). An Aptian regional sea-level rise cut off most of the coarse clastic supply: however, northern uplift persisted and the southwards-thickening prodelta to distal marine mudstone-dominated wedge was deposited (Dallmann 1999; Brekke *et al.* 2001; Worsley 2008). Large-scale clinoforms prograding from the north are associated with this wedge (Worsley 2008; Henriksen *et al.* 2011).

Thick sequences of Lower Cretaceous rocks in the western Barents Sea were deposited in a partially restricted marine-shelf setting atop rapidly subsiding basins (Dalland *et al.* 1988; Dallmann 1999). In contrast, a sequence of condensed platform carbonate deposits accumulated further east in the Barents Sea (Bjarmeland Platform).

During the Late Cretaceous, much of the Barents Sea shelf was uplifted, and areas to the east were either transgressed only at times of maximum sea level and/or display only condensed calcareous to sandy units (Dallmann 1999). The basins beneath the western Barents Sea margin continued to subside and the Nygrunnen Group was deposited in a well-oxygenated deep-marine setting (Faleide *et al.* 2010).

Central East Greenland margin

Distribution. The lowermost Cretaceous is exposed in the southern part of the Jameson Land Basin and in Milne Land (Surlyk *et al.* 1973; Birkelund *et al.*

1984), and the southwards-dipping strata may continue below the Palaeogene volcanic rocks of the Bløseville Kyst (Larsen & Marcussen 1992) (Figs 1 & 4). Cretaceous rocks are also interpreted to occur within the inner Liverpool Land Basin (Hamann *et al.* 2005) (Fig. 12d).

Structural setting. The structural setting in the Early Cretaceous is considered a continuation from the Jurassic basin configuration in Jameson Land. In the Liverpool Land Basin, the interpreted Cretaceous drapes a westwards-rotated fault-block relief (Fig. 12d).

Stratigraphy. In the Jameson Land Basin, the Tithonian–Berriasian Scorseby Sund Group is unconformably overlain by Valanginian mudstones and sandstones of the 120 m-thick Hesteev Formation (Surlyk *et al.* 1973) (Fig. 8). To the west, Valanginian and Hauterivian outcrops in Milne Land are up to about 300 m thick and dominated by sandstones of the Hartz Fjeld Formation (Birkelund *et al.* 1984). In the Liverpool Land Basin, Gautier *et al.* (2011) inferred the presence of Cretaceous marine shales.

Boundary relationships with underlying and overlying systems. The base of the Cretaceous succession forms an erosional unconformity against the Upper Jurassic–lowest Cretaceous rocks in the Jameson Land Basin (Surlyk 2003): its upper boundary is a poorly constrained post-Valanginian–Hauterivian erosion surface (Fig. 8). Palaeogene basalts erosionally overlie the Upper Jurassic–Lower Cretaceous rocks in Milne Land (Larsen *et al.* 2003).

Depositional environment. The Jameson Land Basin succession is represented by marginal- to shallow-marine deposits, basal mudstones and gravity-flow sandstones (Surlyk *et al.* 1973). The Milne Land succession is of shallow-marine origin in its lower part, becoming increasingly marginal marine in the upper part (Birkelund *et al.* 1984).

Møre–mid-Norway margin

Distribution. On the Norwegian mainland, Lower Cretaceous sediments are preserved on Andøya, in northern Norway (e.g. Bøe *et al.* 2010). Offshore, the Cretaceous succession is preserved as thick accumulations within the Møre and Vøring basins (Figs 1, 4 & 13b, c).

Structural setting. In the Early Cretaceous, major rifting occurred along the Møre–mid-Norway continental margin (Doré *et al.* 1999). On seismic profiles, the onset of rifting is marked in both the Møre and Vøring basins by a very pronounced horizon that is observed to onlap the eastern flank of the

basins, adjacent platforms and highs, and which is interpreted as the base Cretaceous unconformity (Brekke 2000; Faleide *et al.* 2010) (Fig. 13b, c). The subsequent Cretaceous subsidence was driven by flexuring of the basin flanks rather than faulting, and resulted in exceptionally thick basin fills. Brekke (2000) further describes a ‘top Cenomanian’ unconformity, which on seismic profiles in the Vøring Basin is identified as a tilted and faulted surface overlapped by younger (post-Cenomanian) strata along both flanks of the basin. This unconformity reflects a phase of tectonism at the end of the Cenomanian, which included compressional deformation and the formation of the Gjallar Ridge (Fig. 1), and eastwards tilting of the westernmost parts of the Vøring Basin (Blystad *et al.* 1995; Lundin & Doré 2011). In the Vøring Basin, the Upper Cretaceous rocks are overprinted by further phases of post-Cenomanian deformation, including the development of anticlines, such as the Nyk and Utgard highs (Figs 1 & 13b), that was instigated in the latest Turonian, and the inversion of the Turonian Vigrid syncline during Maastrichtian–late Paleocene compression (Blystad *et al.* 1995; Brekke 2000; Lundin & Doré 2011). These phases of compression were interrupted by an episode of Campanian extension (Ren *et al.* 2003). By way of contrast, the Møre Basin was tectonically quiet in the Late Cretaceous and underwent passive thermal subsidence (Brekke *et al.* 1999; Brekke 2000).

Stratigraphy. The Cretaceous succession comprises the Cromer Knoll and the Shetland groups, which are separated by the ‘top Cenomanian’ unconformity. In the Møre–mid-Norway region, the former group is assigned a Berriasian–Turonian age, whereas the latter is designated as Turonian–Maastichtian (Dalland *et al.* 1988). This contrasts with the North Sea, where the Cromer Knoll and Shetland groups represent the Lower and Upper Cretaceous, respectively.

On Andøya, the Lower Cretaceous sequence is up to 550 m thick. Offshore, in the Møre and Vøring basins, the Cromer Knoll Group is 2–4 km thick, and is composed predominantly of claystones and interbedded marls, carbonates, and sandstones, with the latter becoming more common towards the top of the sequence (Dalland *et al.* 1988; Brekke 2000) (Figs 7 & 8). The overlying Shetland Group is 3–4 km thick, and predominantly comprises claystones with subordinate amounts of carbonate and sandstone. Many wells in the Møre and Vøring basins penetrate into the Upper Cretaceous succession, in which some wells are terminated having penetrating over 3.5 km of Shetland Group claystones. A thinner Cretaceous succession, ranging from 250 to 1500 m, is present outside of the rift basins.

Boundary relationships with underlying and overlying systems. The Cromer Knoll Group rests with a widespread unconformity on Jurassic and older rocks. The boundary between the Upper Cretaceous and the Paleocene successions is marked by a regional erosional unconformity in the Vøring Basin, and along the flanks of the Møre Basin (Brekke 2000) (Figs 7 & 8).

Depositional environments. The Lower Cretaceous Cromer Knoll Group comprises a mixed clastic–carbonate assemblage that was deposited in shallow-to deep-marine environments, and with increased sandstone input towards the top of the sequence, which probably reflects the onset of mid-Cretaceous deformation (Dalland *et al.* 1988; Brekke 2000). The predominance of claystones in the overlying Shetland Group, with only subordinate amounts of carbonate and sandstone, suggests deposition within an open-marine basin, albeit subjected to episodic phases of compression, uplift and extension.

SE Greenland margin

Distribution. Cretaceous rocks are exposed in the Kangerlugssuaq Basin and small outcrops with possible exposed Cretaceous–Palaeogene sediments occur at Kap Gustav Holm further south (Myers *et al.* 1993) (Figs 1, 4 & 9). Offshore, Cretaceous sediments have been sampled in the Ammassalik Basin (Vallier *et al.* 1998; Thy *et al.* 2007) (Figs 4, 10 & 12e).

Structural setting. The Kangerlugssuaq Basin is a fault-bounded Cretaceous–Palaeogene basin juxtaposed with crystalline basement. The structure of the Ammassalik Basin is poorly known, but also appears to be fault-bounded (Gerlings *et al.*, this volume, in review).

Stratigraphy. The Kangerlugssuaq Group represents the 1 km thick Cretaceous–Paleocene succession in the Kangerlugssuaq Basin (Fig. 9). Aptian sandstones and conglomerates are overlain by Albian–Coniacian sandstones and mudstones, which in turn are overlain by Campanian–Maastrichtian mudstones (Larsen *et al.* 2005b). The succession in the Ammassalik Basin may be several kilometres thick, although the stratigraphy remains largely unknown. The only samples taken from the basin were recovered in cores from near the top of the succession, and include Lower Cretaceous (Albian?) sandstone with coal flasers and Upper Cretaceous–Lower Paleocene siltstone and sandstone (Vallier *et al.* 1998; Thy *et al.* 2007) (Fig. 10). The contact between these two units, and thus the extent of the stratigraphic gap, has not been established.

Boundary relationships with underlying and overlying systems. The base of the Cretaceous against the crystalline basement is not exposed in detail in the Kangerlugssuaq Basin: the erosional upper boundary is overlain by Danian fluvial conglomerates and Palaeogene volcanics of the Blossville Group (Larsen *et al.* 1999). Boundary relationships within the Ammassalik Basin remain unknown (Gerlings *et al.*, this volume, in review).

Depositional environment. The Kangerlugssuaq Basin comprises Aptian alluvial units succeeded by Albian–Coniacian shoreface and offshore deposits. The uppermost Cretaceous is represented by marine basinal mudstones and turbidites (Larsen *et al.* 1999). In the Ammassalik Basin, the Albian sandstones are of shallow-marine origin (Thy *et al.* 2007), whereas the Upper Cretaceous–Lower Paleocene rocks represent marine siltstones and turbidites overlain by a thin, sandy fluvial unit (Vallier *et al.* 1998).

Faroe–Shetland–northern Rockall–Hebrides margin

Distribution. Cretaceous strata are probably widespread throughout the Faroe–Shetland–northern Rockall region, although their full extent in the deep-water basins is obscured by Palaeogene volcanic rocks (Figs 4 & 14). The northern Rockall Basin also includes the Late Cretaceous volcanic seamounts of Rosemary Bank and Anton Dohrn (Jones *et al.* 1974; Morton *et al.* 1995) (Fig. 4). The Cretaceous is only sparsely preserved on the Hebrides margin. On the Rockall Plateau, Lower Cretaceous rocks occur in half-graben on the Hatton High (Hitchen 2004) (Fig. 4), but the presence of Cretaceous in the adjacent Hatton Basin remains ambiguous (Shannon *et al.* 1999).

Structural setting. The Cretaceous succession was deposited in a synrift setting (Dean *et al.* 1999). The Lower Cretaceous sequence is largely confined within the various basins and sub-basins, with the thickest accumulations juxtaposed against the footwall of the adjacent basement/intrabasinal high (Dean *et al.* 1999; Grant *et al.* 1999; Goodchild *et al.* 1999; Lamers & Carmichael 1999; Ritchie *et al.* 2011b, 2013) (Fig. 14a, d). The Upper Cretaceous sequence is more widely developed and by the end of the Cretaceous period most of the major basement highs, including the Rona High, had been isolated or drowned (Dean *et al.* 1999; Stoker & Ziska 2011). In the Faroe–Shetland region, an angular unconformity separates folded and eroded Turonian and older strata from Coniacian–Maastrichtian rocks in the West Shetland, North Rona, East Solan and West Solan basins, as well as the

Foula Sub-basin (Booth *et al.* 1993; Dean *et al.* 1999; Goodchild *et al.* 1999; Grant *et al.* 1999; Stoker 2016). In other West Shetland Shelf basins, much of the Cenomanian–Turonian section is absent, and the unconformity essentially separates Upper and Lower Cretaceous (Fig. 9). Several other intra-Lower and Upper Cretaceous unconformities occur in the West Shetland Shelf and Hebrides Shelf basins, ranging from the Hauterivian to the Campanian in age (Figs 9 & 10), and indicate the persistence of differential uplift and subsidence throughout the Cretaceous. In the northern Rockall Basin, the intrusion of the Rosemary Bank and Anton Dohrn seamounts is linked to Late Cretaceous extension (Ritchie *et al.* 1999). This pattern of coeval extension and compression is consistent with regional strike-slip associated with transtension and transpression (Roberts *et al.* 1999).

Stratigraphy. A total Cretaceous thickness of 5 and 3.5 km has been estimated for the Faroe–Shetland and northern Rockall basins, respectively (Stoker & Ziska 2011; Smith 2013), the bulk of which is of Late Cretaceous age. In the Faroe–Shetland region, the Lower Cretaceous Cromer Knoll Group is dominated by a punctuated coarse clastic assemblage (Ritchie *et al.* 1996; Stoker & Ziska 2011; Stoker 2016) (Fig. 9). Up to 1 km of sandstones with subordinate conglomerates, mudstones, limestones and argillaceous coal is preserved in the West Shetland Basin, although the sequence is generally much thinner (<0.5 km) in other shelf basins (e.g. the North Rona Basin and the West Solan Basin). In the Faroe–Shetland Basin, the Cromer Knoll Group is predominantly marine mudstone, with sporadic coarse clastic deposits preserved adjacent to the basin margin and intrabasinal highs. A maximum-drilled thickness of 1.5 km occurs in the Foula Sub-basin. The Upper Cretaceous consists of argillaceous limestones and mudstones, with rare organic-rich pyritic mudstone (Chalk Group) of Cenomanian–Turonian age in the North Rona, East Solan and West Shetland basins, which pass northwards and upwards into Cenomanian–Maastrichtian calcareous mudstones (Shetland Group) in the Faroe–Shetland Basin (Ritchie *et al.* 1996; Harker 2002; Stoker & Ziska 2011; Stoker 2016) (Fig. 9). The Chalk Group is generally less than 250 m thick, whereas the Shetland Group possibly exceeds 3.5 km in thickness in the Foula, Judd and Flett sub-basins.

In the Inner Hebrides region, Lower Cretaceous rocks are absent, but a thin (<25 m thick) succession of Upper Cretaceous shallow-marine deposits, including chalk and limestones, comprise the Inner Hebrides Group (Mortimore *et al.* 2001; Hopson 2005; Waters *et al.* 2007) (Fig. 10). The West Flannan Basin contains organic-rich Berriasian

mudstones, which are probably equivalent to the Upper Jurassic–lowest Cretaceous Humber Group. In the West Lewis Basin, these rocks are unconformably overlain by thin (<10 m thick) Barremian–Cenomanian shallow-marine sandstones and mudstones, which are, in turn, unconformably overlain by over 200 m of Coniacian–Maastrichtian marine mudstones (Hitchen & Stoker 1993; Smith 2013).

In the NE Rockall Basin, a predominantly Upper Cretaceous mudstone sequence has been proved, with a maximum-drilled thickness of about 1.8 km, although the section is locally intruded by up to 700 m of basic igneous sills of Late Cretaceous or Paleocene age (Archer *et al.* 2005; Smith 2013). On the eastern flank of the central northern Rockall Basin, a 500 m-thick Cretaceous sequence comprises late Barremian–early Aptian to Maastrichtian mudstones (on a possible basal conglomerate) that overlie Archaean basement (Smith 2013). A prominent, largely unfaulted reflector that separates the Lower and Upper Cretaceous sequences is marked by a thick Cenomanian limestone (Musgrove & Mitchener 1996). It is estimated that the Upper Cretaceous is up to 2.5 km thick beneath the central part of the northern Rockall Basin, with the Lower Cretaceous in excess of 1 km (Smith 2013).

The volcanic seamounts of Rosemary Bank and Anton Dohrn have been dated as Late Cretaceous on the basis that upper Maastrichtian limestone with basalt clasts is preserved on the top of Rosemary Bank (Morton *et al.* 1995), whereas unmetamorphosed Maastrichtian chalk is preserved in vesicles in lavas from the Anton Dohrn Seamount (Jones *et al.* 1974).

On the Rockall Plateau, Albian organic-rich mudstones of paralic origin are preserved within half-graben on the Hatton High (Hitchen 2004) (Fig. 10).

Boundary relationships with underlying and overlying systems. The Cretaceous is largely separated from Jurassic and older strata by a widespread unconformity (Figs 9 & 10). On the Hatton High, the base of the sequence has not been sampled. The Cretaceous–Paleocene boundary is sometimes conformable in the deeper parts of basins, but, more commonly, it is marked by a widespread unconformity (Stoker & Ziska 2011; Smith 2013).

Depositional environment. In the basins underlying the West Shetland Shelf, the Cromer Knoll Group was deposited in marginal- to shallow-marine and basinal marine environments that were subject to fluctuating aerobic/anaerobic bottom waters (Ritchie *et al.* 1996; Harker 2002; Stoker & Ziska 2011). Common intraformational unconformities imply contemporary rifting, albeit localized and intermittent (Dean *et al.* 1999; Harker 2002; Larsen

et al. 2010; Stoker 2016). An expansion of rift activity across the entire Faroe–Shetland–northern Rockall–Hebrides region probably occurred in the late Barremian/Aptian–Albian, with paralic to marine clastic deposition in all basins, including on the Hatton High.

The Chalk and Shetland groups were largely deposited in an aerobic, open-marine, shelf to basinal setting. Sediment thickness and accumulation rate increased during the Late Cretaceous as both the Faroe–Shetland and northern Rockall basins accumulated an expanded Cretaceous sequence (e.g. Fig. 14a). However, the increased subsidence of these basins was periodically interrupted by intermittent compression, uplift and erosion, which persisted throughout the Late Cretaceous (Dean *et al.* 1999; Doré *et al.* 1999; Roberts *et al.* 1999; Larsen *et al.* 2010; Stoker & Ziska 2011; Stoker 2016). In the northern Rockall Basin, the abundant Late Cretaceous–Paleocene sills throughout the Upper Cretaceous sequence, together with the intrusion of the axial volcanic seamounts, are probably associated with the crustal thinning.

Southern Rockall–Porcupine margin

Distribution. Cretaceous strata have a widespread occurrence in the southern Rockall–Porcupine region. Lower Cretaceous clastic-dominant (sandstones, siltstones and mudstones) successions occur in all the basins, with Upper Cretaceous chalk in all the basins and covering the intervening basement highs (Figs 4 & 15).

Structural setting. The Cretaceous succession was deposited in a late synrift to early post-rift setting. The earliest Lower Cretaceous succession in the Porcupine Basin marks the waning phase of major Tithonian rifting (Crocker & Shannon 1987; Moore 1992), broadly coeval with North Sea Late Cimmerian events and with a change in regional plate movements. This was succeeded by a tectonically quiescent, post-rift period interrupted locally in the Porcupine, Slyne and Erris basins by an Aptian–Albian rift phase. Early Cretaceous volcanism is suggested in the central part of the Porcupine Basin (Crocker & Shannon 1987; Tate & Dobson 1988; Naylor *et al.* 1999) (Fig. 15b, c). The Late Cretaceous saw thermal subsidence and a major sea-level rise resulting in transgression following the commencement of seafloor spreading in the North Atlantic.

Stratigraphy. The total Cretaceous section in the Porcupine Basin is up to 4 km thick. In the northern part of the Porcupine Basin, Valanginian–late Aptian sandstones, siltstones and mudstones occur, with Aptian–Albian sandstones prevalent along parts of the eastern basin margin (Fig. 11).

Basinwards, sandy and silty units are encapsulated in shales (Shannon 1993). Lower Cretaceous extrusives and tuffs occur through the Barremian–Albian succession (Tate & Dobson 1988; Naylor *et al.* 2002). The Cenomanian–Danian chalk sequence in the Porcupine Basin ranges from 400 m near the margins of the basin to more than 1000 m in the basin centre (Moore & Shannon 1995).

In the northern Slyne Basin, Albian claystones and glauconitic sandstones pass upwards to sandy claystones with occasional limestones (Dancer *et al.* 2005). In the Erris Basin, late Berriasian–Valanginian claystones and sandstones are overlain by thick (>250 m) late Valanginian–Hauterivian sandstones, followed by claystones and Albian argillaceous limestones (Fig. 11). Upper Cretaceous chalk is present in the northern Slyne Basin (Dancer *et al.* 1999) and throughout the Erris Basin (Chapman *et al.* 1999).

On the eastern margin of the southern Rockall Basin, thin (<40 m) Lower Cretaceous mudstones are overlain by thick Upper Cretaceous mudstones, marls and chalks (Tyrrell *et al.* 2010) (Fig. 11). Further south along the margin, an Early Cretaceous condensed ('brownsand') succession (Haughton *et al.* 2005) is followed by a thin upper Cenomanian–middle Turonian succession, which is overlain unconformably by an upper Cenomanian–upper Maastrichtian mixed clastic–carbonate succession.

In the Fastnet Basin, up to 700 m of Lower Cretaceous strata have been encountered in wells. Berriasian–Hauterivian sandstones and shales are followed by Barremian–Albian sandstones and overlain by latest Albian–Turonian chalk (Robinson *et al.* 1981). To the west, in the Goban Spur, Neocomian sandstones and cherty limestones are overlain by Barremian–lower Aptian limestones and claystones (Cook 1987). Late Aptian–Maastrichtian strata are predominantly chalk (Fig. 11).

An important igneous episode occurred in the Early Cretaceous, represented by the Barra (south Rockall Basin) and Porcupine Volcanic Ridge systems (Naylor & Shannon 2005) (Fig. 11). The Barra system remained upstanding until the Eocene, whereas the Porcupine system was overlapped and buried by the mid-Cretaceous.

Boundary relationships with underlying and overlying systems. Regionally, the Cretaceous rests with angular unconformity on the underlying Jurassic (Figs 11 & 15). Sometimes this is a single unconformity, but locally a set of composite unconformities occurs. Several intra-Lower Cretaceous unconformities occur, with a major Aptian–Albian rift-related unconformity identified in all of the basins (Crocker & Shannon 1987; Chapman *et al.* 1999; Dancer *et al.* 2005). The top of the Cretaceous succession is sometimes conformable, but more often

is defined by an unconformity or disconformity with variable amounts of erosion (Fig. 11).

Depositional environment. Early Cretaceous deposition in the Porcupine Basin was primarily in a marine shale facies, interrupted locally by deltaic sandstones reflecting minor Aptian–Albian rifting (Fig. 11). Rift-generated deltaic deposits prograded from the northern and SE basin margins, with a series of offshore-bar sandstones rimming the basin. Basinwards, the clastic pulses produced sandy and silty basin-floor fans (Shannon 1993). The Upper Cretaceous (Cenomanian–Danian) chalk sequence in the Porcupine Basin reflects marine deposition with little terrigenous input.

A marine setting dominated the Cretaceous of the Slyne and Erris basins. In the Erris Basin, a thick (> 120 m) Valanginian–Hauterivian submarine-fan sandstone complex is preserved (Murphy & Croker 1992; Naylor & Shannon 2011) (Fig. 11). The Upper Cretaceous, like that of the Porcupine Basin, represents a deep-water, non-terrigenous depositional environment.

On the eastern margin of the southern Rockall Basin, the Lower Cretaceous represents a shallow-marine shelf setting (Haughton *et al.* 2005), giving way basinwards to deep-water facies (Tyrrell *et al.* 2010). The Upper Cretaceous on the basin margin is in marine facies, with fluctuations in water depths, with more uniform pelagic deposition in relatively deep water.

The Cretaceous depositional environment in the Goban Spur is predominantly marine with a lower clastic (Lower Cretaceous) and an upper carbonate (Upper Cretaceous) setting reflecting a deepening of the basin. However, in the Fastnet Basin, the Lower Cretaceous comprises a fluvio-deltaic depositional setting, with a shallow-marine transgression in Aptian–Albian times and capped by deep-water Upper Cretaceous chalk facies (Robinson *et al.* 1981) (Fig. 11).

Upper Palaeozoic–Mesozoic stratigraphy of the Jan Mayen microcontinent

Jan Mayen is a volcanic island located at the northern end of the Jan Mayen Ridge (Fig. 1). The latter is a north–south-trending submarine feature that extends about 400 km southwards from Jan Mayen towards Iceland and its insular margin. The Jan Mayen Ridge is a continental remnant derived from the break-up of Norway and Greenland, and was left out in the ocean as a ‘microcontinent’: the Jan Mayen microcontinent (JMMC) (Blischke *et al.*, this volume, in press).

The JMMC comprises an area larger than the Jan Mayen Ridge (Fig. 1). Its northern limit is ambiguous, and there is uncertainty as to whether it extends beneath Jan Mayen Island, or whether it is located just to the south of the island; its southern limit is also unclear and might extend further towards

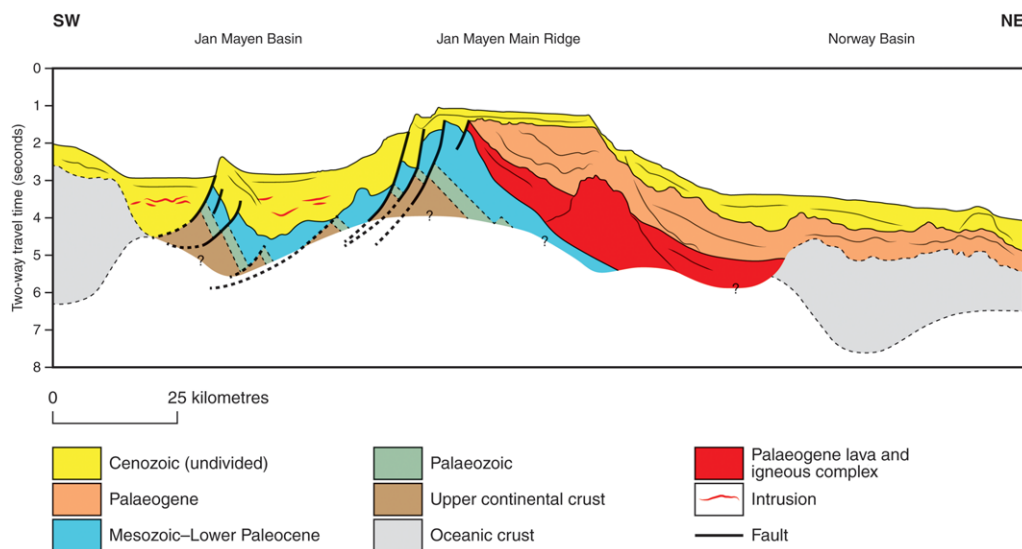


Fig. 16. Interpreted geoseismic profile across the Jan Mayen microcontinent showing the inferred stratigraphic architecture of probable Palaeozoic–Mesozoic rocks beneath the Jan Mayen Ridge and the Jan Mayen Basin. Interpretation based on Peron-Pinvidic *et al.* (2012a, b). Location of profile given in Figure 1.

Iceland than is shown in Figure 1. Consequently, the stratigraphy and structure of the JMMC is poorly known, particularly at deeper levels: thus, the following description of the Upper Palaeozoic–Mesozoic geology remains preliminary (cf. Blischke *et al.*, this volume, in press).

Distribution. The Upper Palaeozoic–Mesozoic succession is mainly inferred for the JMMC on the basis of its position to its conjugate margins in Norway and Greenland (Figs 2–4). Whereas the possibility of Mesozoic and older Palaeozoic rocks has been interpreted at depth beneath the Jan Mayen Ridge (e.g. Peron-Pinvidic *et al.* 2012a, b), a major problem in their recognition is caused by lower Palaeogene lavas and sills that commonly obscure the pre-Cenozoic sequences on seismic reflection data (Fig. 16).

The pre-Palaeozoic basement, which is tentatively imaged on seismic reflection data, is herein compared with the granite-intruded Caledonian basement (Gilotti *et al.* 2008; Kalsbeek *et al.* 2008) that is preserved on the Liverpool Land Ridge in East Greenland (Fig. 12d). Seismic refraction and reflection data further suggest that the sub-basalt strata might be comparable to the Upper Palaeozoic–Lower Mesozoic successions that are exposed in the Jameson Land Basin and Traill Ø–Hold with Hope area in East Greenland, as well as the Møre and Vøring basins offshore Norway (Figs 7 & 8).

Structural setting. The structural setting of the JMMC is best compared to the structural framework along the central East Greenland coastal area, where the Jameson Land Basin, in particular, provides a potential analogue. This narrow, north–south-trending basin includes a Permian–Lower Cretaceous succession (Fig. 8) that is intruded by Cenozoic igneous rocks (e.g. thick sills) (Larsen & Marcussen 1992; Hald & Tegner 2000). On the JMMC, it is possible that deeper-level structures and rocks, which are visible through windows in the igneous section, might represent comparable Palaeozoic–Mesozoic strata (Blischke *et al.* 2011). As previously noted (see above), the Mesozoic succession in the Jameson Land Basin might continue southwards beneath the Palaeogene volcanic succession in the Blossville Kyst area of central East Greenland (Larsen *et al.* 2005a, 2013), which would have located this structure right across the central part of the JMMC – a scenario previously proposed by Surlyk (1990, 2003). The observation that a thinner and more condensed pre-Cenozoic succession is preserved on the JMMC in comparison to the East Greenland and Møre–mid-Norway areas might imply that the JMMC formed a structurally higher and shallower platform between the

adjacent Jameson Land and Møre basins, possibly along strike with the north–south-trending Liverpool Land High (Blischke *et al.*, this volume, in press).

Stratigraphy. The presence and extent of Palaeozoic–Mesozoic rocks along the Jan Mayen Ridge and within the Jan Mayen Basin remains ambiguous on the basis of refraction and reflection seismic interpretations. An inferred stratigraphic range for the southern JMMC is indicated in Figure 8, based on a direct comparison with the conjugate Jameson Land, Møre and Vøring basins: the northern part of the JMMC – formerly juxtaposed with the Vøring margin – has been subjected to intense Palaeogene volcanism that has obscured much of the succession on seismic profile data.

Whereas the recovery of Upper Permian–Lower Triassic and Lower Cretaceous rocks, including limestone, from the Jan Mayen Ridge was initially interpreted by the Norwegian Petroleum Directorate (2012) to indicate the presence of Upper Palaeozoic–Mesozoic rocks on the JMMC, the likelihood is that these samples represent ice-rafted detritus derived and dumped from East Greenland icebergs (Norwegian Petroleum Directorate 2013). Nevertheless, the presence of Cretaceous rocks across the JMMC seems highly probable based on the interpretation of seismic reflection data (Fig. 16), together with evidence from the conjugate Vøring and Møre basins to the east, and the Traill Ø–Hold with Hope area of NE Greenland to the west (Figs 7 & 8).

Boundary relationships with underlying and overlying systems. No boundary relationships can be firmly established: however, potential boundaries indicated in Figure 8 relate to reflecting surfaces observed on seismic data that might correlate with key unconformities preserved in the adjacent conjugate basins.

Depositional environment. In the absence of a proven Late Palaeozoic–Mesozoic stratigraphic record for the JMMC, the conjugate Jameson Land, Møre and Vøring basins are the best providers of information concerning potential depositional environments in this area.

Regional stratigraphic framework

The stratigraphic distribution maps and correlation charts presented in Figures 2–11 provide a strong observational basis for the establishment of a unified Late Palaeozoic–Mesozoic stratigraphic framework for the NE Atlantic region. Any attempt to interpret this framework in terms of the break-up of the Pangaeian supercontinent has to take into

consideration the general lithostratigraphic pattern and correlation along the entire length of the NE Atlantic rift system. In this section, we address this issue by utilizing a rift-axial correlation chart (Fig. 17) that links basins along the NE Atlantic rift system. In the context of the entire Pangaeon supercontinent, this chart represents a transect from the northern (Boreal) marine margin to the centre of the Pangaeon plate, as depicted on recent plate-wide reconstructions (e.g. Doré *et al.* 1999; Pharaoh *et al.* 2010; Golonka 2011; Lawver *et al.* 2011). Reference to the southern (Tethyan) marine margin is also included in the text where necessary. The rift system has been arranged into five geographical segments: (1) the North/NE Greenland–Svalbard–western Barents Sea segment; (2) the NE/central East Greenland–mid-Norway–Møre segment; (3) SE Greenland–Faroe–Shetland segment; (4) the southern Greenland–northern Rockall–Hebrides segment; and (5) the Inner Hebrides–NW Irish–Porcupine–southern Rockall segment. This arrangement retains the conjugate placement of basins and margins whilst, at the same time, presenting an along-axis view of the stratigraphic framework. Almost all of the stratigraphic columns shown on this chart represent a proven rock record.

From the stratigraphic information described above, and collated on the rift-axial chart (Fig. 17), we present the regional Late Palaeozoic–Mesozoic stratigraphic framework of the NE Atlantic region as it is currently known from the observed rock record. Even from a cursory inspection of the stratigraphic chart, it is obvious that regional changes in stratigraphic development largely coincide with the Permian–Triassic, Jurassic and Cretaceous successions. In the ensuing discussion, we summarize the large-scale pattern of sedimentation and basin development associated with these successions, as well as the character of the major bounding surfaces. Potential correlation to tectonic events that were affecting the Pangaeon plate throughout the Late Palaeozoic–Mesozoic will also be considered, together with established views regarding faunal provinciality (Boreal–Tethyan) across this interval.

Carboniferous–Permian boundary

The northernmost part of the study area is characterized by a largely conformable Carboniferous–Permian boundary. In the North/NE Greenland–Svalbard–western Barents Sea region, a passively subsiding margin formed in the Late Carboniferous and persisted, with only minor interruptions, until the Mid–Late Permian (Stemmerik & Håkansson 1991; Surlyk 1991). A mixed siliciclastic–carbonate shelf sequence of Late Carboniferous–Permian age is preserved in basins throughout this region (see below).

By way of contrast, over most of the study area, the Carboniferous–Permian boundary is marked by a regionally extensive unconformity (Fig. 17). On the western side of the NE/central East Greenland–mid-Norway–Møre region, a regional peneplain formed by Early Permian uplift and erosion is well constrained in outcrops in the Traill Ø–Hold with Hope and Jameson Land areas where rocks of late Early–Mid-Permian age overlie Upper Carboniferous strata (Surlyk *et al.* 1986). The event is also recognized in the North/NE Greenland–Svalbard–western Barents Sea region ('minor interruptions'), albeit within the overall context of a subsiding margin (Stemmerik 2000; Hamann *et al.* 2005).

Further south, the Permian–Triassic succession is generally more poorly dated, making it difficult to assess the precise nature of the Carboniferous–Permian boundary: however, the duration of the erosional hiatus is generally more expanded. Whereas locally there is evidence (e.g. Porcupine Basin) of a rapid transition from humid to arid conditions and the onset of red-bed deposition, elsewhere (SE Greenland–NW Irish region) there is a widespread unconformity, with rocks ranging in age from Mid-Permian to Early Triassic variably overlying Upper–Lower Carboniferous rocks. In the North Sea Basin, a comparable regional hiatus is known to be an amalgamation of several unconformities rather than a single event (Glennie *et al.* 2003; Gast *et al.* 2010). It has been suggested that the Carboniferous–Permian boundary, particularly in the southern part of the study area, formed through regional uplift and erosion driven by a regional pattern of wrench tectonics and crustal thinning associated with Variscan orogenic collapse (Pharaoh *et al.* 2010).

Permian–Triassic succession

There is a marked north–south contrast in the overall character of the Permian–Triassic succession, with a marine-shelf assemblage of carbonate and siliciclastic rocks preserved in the northernmost (Boreal) part of the study area, whereas a predominantly arid setting prevailed elsewhere (Fig. 17). The most complete Permian–Triassic successions have been proved from Svalbard and the SW Barents Sea. Further south, in the area between NE Greenland/mid-Norway and the Porcupine Basin, Middle/Upper Permian–Triassic sequences predominate, with Triassic strata being the thickest and most widespread unit. In some parts of North and NE Greenland, Permian and/or Triassic rocks are absent, and in other areas, such as SE Greenland, the Rockall Plateau and the north Rockall Basin, no Permian–Triassic rocks have so far been proved.

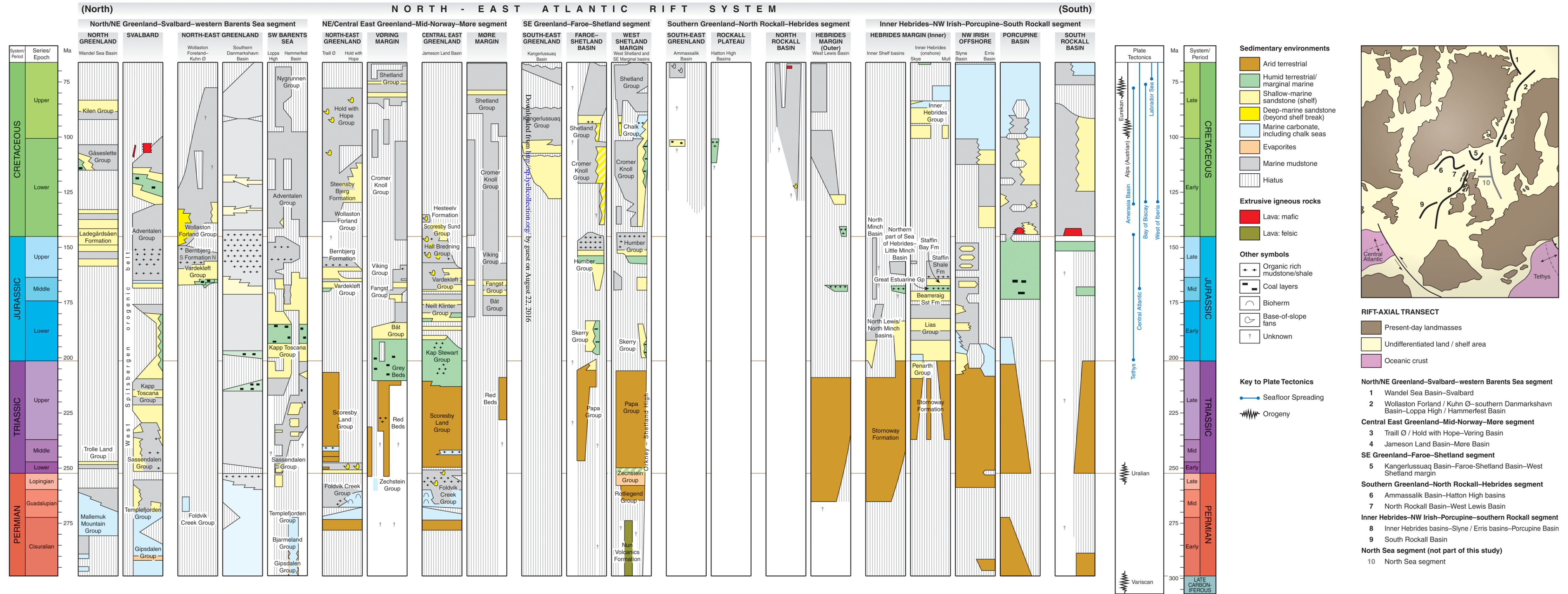


Fig. 17. Regional Permian-Cretaceous stratigraphic correlation panel showing selected columns taken (and simplified) from Figs 5-11, and rearranged into a general north-south transect along the axis of the NE Atlantic rift system. The chart is intended to provide a link between basins along the entire NE Atlantic rift system, extending from the northern margin to the centre of the Pangaea plate, with an emphasis on sedimentary environments. Inset map shows the general location of the stratigraphic columns that represent specific geographical segments of the rift zone (see the text for details); the configuration of the Pangaea plate is based on Late Jurassic-Early Cretaceous reconstructions of Ziegler (1988) and Doré *et al.* (1999). Timescale based on Gradstein *et al.* (2012).

The bulk of the known Permian rocks are preserved in central East Greenland, North and NE Greenland, Svalbard, and the western Barents Sea. In the Wandel Sea Basin, the Late Carboniferous–Permian marine-shelf succession comprises sandstone, shale and carbonate rocks (Stemmerik & Håkansson 1991; Surlyk 1991). A comparable assemblage is also preserved on Svalbard and in the Hammerfest Basin (SW Barents Sea), and is inferred to be present in the Danmarkshavn Basin (Gautier *et al.* 2011).

In central East Greenland, subsidence was instigated in the Mid-Permian as a consequence of thermal contraction, with minor rifting in mainly westerly tilted half-graben (Surlyk *et al.* 1986). The deposition of an overall transgressive Middle–Upper Permian succession in the Traill Ø–Hold with Hope and Jameson Land areas was initiated by fluvial siliciclastics, followed by marine and hypersaline carbonates and evaporites that are unconformably overlain by fully marine carbonate build-ups along basin margins, together with organic-rich basinal mudstones (Surlyk *et al.* 1986; Surlyk 1990; Stemmerik 2000) (Fig. 17).

Further south, a fragmentary Lower Permian record includes alkaline lavas in the West Shetland region, and ?Lower Permian terrestrial sandstones and mudstones in the Porcupine, southern Rockall and Erris basins, offshore Ireland. The Early Permian volcanism and sedimentation in these areas are linked to tectonism associated with Variscan orogenic collapse.

A significant pulse of rifting affected the NE Atlantic region during the Late Permian–earliest Triassic, which coincided with the final phase of the Uralian Orogeny (Fig. 17) in the eastern Barents Sea region (Smelror *et al.* 2009), to the NE of the study area. Variable large-scale basin architectures are preserved, including half-graben and broad sheet-like geometries. This is interpreted as reflecting a depositional setting controlled by palaeotopographical infill, faulted depocentres (early localized rifts along reactivated Variscan and Caledonian lineaments) and wide-rift extensional processes (e.g. Štolfová & Shannon 2009). Mid–Late Triassic time was generally characterized by post-rift thermal subsidence.

In the northern part of the study area (Greenland–Norway conjugate), the Late Permian–earliest Triassic rifting controlled marine-dominated deposition in mainly westerly tilted half-graben. The Permian–Triassic boundary in this region is a prominent unconformity at basin margins. A coeval rise in sea level resulted in a marine passageway – the Zechstein Sea (cf. Glennie *et al.* 2003) – that transgressed southwards between Greenland and Norway (Seidler *et al.* 2004; Bjerager *et al.* 2006). Upper Permian–Lower Triassic shallow-marine

evaporites in the Vøring Basin and in the West Shetland region are most probably linked to this shallow-marine incursion, which extended into the North Sea (Glennie *et al.* 2003), although faunal evidence indicates that there was no connection with the Tethyan realm at this time (Smith 1980; Doré 1992). Equivalent rocks are not present on the Hebrides margin, but are preserved in basins offshore NW Ireland (Fig. 17). However, in the latter area, deposition may have been entirely terrestrial and related to a relative rise of the regional water table rather than marine flooding (Glennie *et al.* 2003).

The Triassic was a time of mainly arid continental deposition with thick coarse clastic successions of red beds preserved in various basins located between NE/central East Greenland–mid-Norway and the Porcupine and southern Rockall basins, offshore Ireland (Fig. 17). Thicknesses in excess of 2 km are preserved in basins offshore NW Scotland, on the inner Vøring margin and onshore central East Greenland (Fig. 2). Recorded Triassic deposits in central East Greenland decrease progressively north of Jameson Land and reappear in the North Greenland Wandel Sea Basin, where Lower Triassic marine siliciclastic rocks are preserved. In contrast, Triassic deposits of presumed marine shale and sandstone are inferred to be present in the Danmarkshavn Basin (Hamann *et al.* 2005), and which may have been sourced from an uplifted western basin margin (Middle and Upper Triassic rocks are absent onshore North and NE Greenland). Triassic deposits are also inferred to be present in the Thetis Basin (Hamann *et al.* 2005; Dinkelman *et al.* 2010). The Triassic siliciclastic succession is essentially restricted to the northern margin of the study area, which was a wide shelf basin at this time. On Svalbard and in the western Barents Sea, a succession of alternating marine, deltaic and coastal facies prograded into this basin (Brekke *et al.* 2001; Riis *et al.* 2008; Worsley 2008; Smelror *et al.* 2009; Henriksen *et al.* 2011). The extensive Mid–Late Triassic westwards progradation of deltaic systems was a consequence of the progressive uplift and erosion in the Uralian highlands, as well as the northern Fennoscandian Shield.

A brief marine incursion from the Boreal Sea in the Mid-Triassic brought shallow-marine mudstones and carbonates to the North/NE and central East Greenland part of the NE Atlantic rift system (Fig. 17). It has been speculated that this incursion might correlate with Tethyan marine incursions into the North Sea (e.g. Muschelkalk Formation) (Jacobsen & van Veen 1984). However, it is stressed that palaeogeographical and faunal evidence do not support the likelihood of a through-going marine connection between the Boreal and Tethyan realms in Mid- and Late Triassic times (Doré 1992).

Triassic–Jurassic boundary

The character of the Triassic–Jurassic boundary is highly variable across the NE Atlantic region. It is marked by a major unconformity in North and NE Greenland, the Møre Basin, and across much of the Faroe–Shetland–Hebrides region. By way of contrast, shallow-marine and paralic sedimentation persisted, at least locally, from the Triassic into the Early Jurassic on Svalbard, in the western Barents Sea, and in the Vøring and Jameson Land basins, whereas, in the Inner Hebrides region, a transition from an arid to a shallow-marine setting, which began in the latest Triassic (Rhaetian) and continued into the Early Jurassic, is locally observed. Further south, the Triassic–Jurassic boundary where observed (e.g. in the northern part of the Porcupine Basin, NW Irish offshore basins and the Fastnet Basin) and interpreted (e.g. Goban Spur, perched basins on the flanks of the southern Rockall Basin) is more widely conformable and marks a rapid transition, through a marine transgression, from continental arid to shallow-marine shelf limestones. The marine transgression commenced in the Rhaetian, and was a regionally extensive event, occurring rapidly and at the same time throughout all of the Irish offshore basins. There is no evidence of any significant tectonism (rifting) at the Triassic–Jurassic boundary in this area.

This variability in the character of the Triassic–Jurassic boundary might be a consequence of events occurring on the flanks of, and within, the Pangaeian plate. Along the Boreal margin of the plate, it is suggested that sedimentation in the Svalbard–western Barents Sea region was still controlled by compression and uplift generated by the Uralian Orogeny, as well as a renewal of fault-related tectonic activity at the end of the Triassic (Smelror *et al.* 2009). The latter tectonism is probably linked to an acceleration of rifting and subsidence within the northern part of the NE Atlantic rift system (Greenland–Norway) during the Late Triassic, as it started to propagate southwards towards the Central Atlantic domain and generated contemporaneous uplift and erosion (and, thus, unconformities) on the flanks of the rift system (Pharaoh *et al.* 2010). On the southern edge of the Pangaeian plate, incipient ocean-floor spreading was instigated in the Tethys to the SE and in the proto-central Atlantic to the SW (Doré *et al.* 1999) (Fig. 17). According to Doré (1992), marine flooding (?Rhaetian transgression) of the Permian–Triassic rift basins occurred as rifting breached Pangaea.

Jurassic succession

The distribution and stratigraphic range of the Jurassic succession is variable across the NE

Atlantic region (Fig. 17). The most complete successions are preserved in Svalbard and the western Barents Sea, the Jameson Land Basin, the inner Vøring margin, the inner Hebrides margin, and offshore NW Ireland (e.g. Slyne Basin), although unconformities are present in these areas, particularly a widespread Mid-Jurassic break. Fragmentary Jurassic sequences are preserved elsewhere (e.g. onshore North and NE Greenland, the Faroe–Shetland Basin, the outer part of the Hebrides margin, the Porcupine Basin, and the southern Rockall Basin), and it was not until the Mid–Late Jurassic that a more regionally extensive succession was deposited. Areas of uncertainty remain, especially the basins offshore NE Greenland, where relatively complete successions are inferred, but remain untested. In some areas, such as SE Greenland, the Rockall Plateau and the northern Rockall Basin, no Jurassic rocks have so far been proved.

In Svalbard, the SW Barents Sea, the Jameson Land Basin and the inner Vøring margin, the repetitive clastic succession of deltaic, paralic and shallow-marine shelf sedimentation initiated in the Triassic, persisted into the Early Jurassic. These areas were characterized by passive thermal subsidence, with only minor fault activity and uplift (Blystad *et al.* 1995; Brekke *et al.* 2001; Surlyk 2003; Henriksen *et al.* 2011). In contrast, the adjacent areas of North and NE Greenland, as well as parts of the Norwegian margin (e.g. Møre region), were areas of more extensive uplift and erosion (Brekke *et al.* 1999; Doré *et al.* 1999). Comparable coastal and marine deposition occurred in the Faroe–Shetland and Inner Hebrides regions, and offshore NW Ireland. Limited extensional fault activity is interpreted to have occurred in the Inner Hebrides region (Morton 1989).

This general pattern of Early Jurassic thermal subsidence and mild extensional tectonism was interrupted in the Mid-Jurassic by widespread uplift and shallowing, which is commonly reflected in the stratigraphic record as an unconformable boundary between Lower and Middle Jurassic rocks. In the North Sea, this boundary is termed the ‘Mid-Cimmerian Unconformity’, and the uplift has been attributed to a hotspot-related dome (Underhill & Partington 1993; Glennie & Underhill 1998). However, the palaeogeographical extent of the Mid-Jurassic unconformity (Fig. 17) might suggest that this dome was one of a family of uplifts that extended across the European–NE Atlantic region (Doré *et al.* 1999). A major impact of uplift at this time was the lack of marine continuity along what was to become the Atlantic margin. Indeed, the acute faunal provinciality (Callomon 1984) established between the Boreal and Tethyan realms from the late Bajocian onwards is an indication of the

existence of a substantial palaeogeographical barrier (Doré 1992).

An intense phase of rifting occurred in some parts of the NE Atlantic region in late Mid–Late Jurassic time, and there is a general trend towards marine conditions. However, continuity of any sea-way between NW Europe and the Arctic was probably highly complex, at the very least, as reflected in the variable intensity of the Late Jurassic tectonic activity, as well as the high degree of faunal (ammonite) provinciality even within the Boreal realm (Casey 1971; Callomon & Birkelund 1982; Doré 1992). In central East, NE and North Greenland, major rifting was initiated in the late Bajocian and culminated in the earliest Cretaceous. Middle Jurassic flooding and the northwards propagation of the rifting resulted in the formation of a series of connected, elongate, marine basins in platform and half-graben settings (Surlyk 2003). Whereas the Jameson Land Basin experienced minor tilting during this interval, the Wollaston Forland Basin (Traill Ø–Hold with Hope area) underwent significant rotational block faulting in the westwards-tilted half-graben. In Svalbard and the western Barents Sea, Mid-Jurassic–Early Cretaceous rifting caused a marked rejuvenation of the topography into a system of structural highs and basins (e.g. Bjørnøya Basin, Tromsø Basin), and triggered the development of a marine connection across the Barents Sea shelf (Smelror *et al.* 2009; Faleide *et al.* 2010). In the northern part of this area, some of this tectonic activity may have been related to extension in the Amerasia Basin, as a precursor to the opening of the Arctic Ocean (Smelror *et al.* 2009). Shallow-shelf to deep-marine sedimentation prevailed over a large part of this northern area, including the accumulation of black shales in many of the basins. In the mid-Norway–Møre region, variable thicknesses of marine sediment accumulated on a series of tilted fault blocks, particularly on the Vøring margin (Faleide *et al.* 2010).

Paralic and marginal-marine sedimentation prevailed in the Faroe–Shetland–Hebrides region until latest Jurassic–earliest Cretaceous time when the area was transgressed and marine mudstones, including organic-rich black shales, were deposited. It has been suggested that the general lack of variation in the thickness of the Upper Jurassic rocks across the Faroe–Shetland region is an indication that the effects of any contemporaneous rift activity in this area was negligible (Dean *et al.* 1999; Doré *et al.* 1999).

Offshore Ireland, the Middle Jurassic marks a major change in depositional basin shape and size with the onset of rifting. Some rifting (e.g. NW Irish offshore basins) reactivated existing lineaments (e.g. of Caledonian age), while in other

cases (e.g. Porcupine Basin) a new, large north–south basin geometry resulted. The switch from a marine-shelf (Lower Jurassic) to a humid fluvial (Middle Jurassic) environment reflects this tectonic change (i.e. the onset of rifting). Rifting reached its nadir in Late Jurassic times with the formation of tilted fault blocks and major facies variations (Shannon 1991; Sinclair *et al.* 1994). In the central and southern Porcupine Basin, and in the southern Rockall Basin, crustal stretching was extreme (hyperextension) in Late Jurassic time, resulting in significant differential subsidence towards the basin centres. The Upper Jurassic succession is characterized by facies variations, ranging from fluvial to basin-floor sandstones, siltstones and mudstones to shallow-water limestones. This marks a general upwards trend from non-marine to marine conditions.

In the North Sea Basin, a major phase of extensional activity, accelerated subsidence and basin deepening in the Late Jurassic is well established (Glennie & Underhill 1998). Pharaoh *et al.* (2010) have attributed the intensification of the North Sea rift system to a wrench-dominated tectonic regime. Doré *et al.* (1999) suggested that Late Jurassic rifting, in general, throughout NW Europe was driven by an east–west extensional stress regime influenced predominantly by seafloor spreading in the Tethys, and cite close to northerly-trending structural elements, such as the Porcupine Basin, Halten Terrace, and the basins of central East and NE Greenland, as unequivocal Jurassic rift basins.

Jurassic–Cretaceous boundary

In North and NE Greenland, the Jurassic–Cretaceous boundary is transitional and marks the culmination of rifting in this region, with associated basin segmentation, block tilting and footwall uplift. Prominent deep-marine coarse clastic sediments were deposited along rift margins, which pass distally into mudstones (Surlyk 1978). This tectonic activity triggered major salt movements in the northern part of the Danmarkshavn Basin (Hamann *et al.* 2005). A transitional boundary is also observed on Svalbard, whereas an unconformity is developed in the SW Barents Sea.

Over most of the rest of the NE Atlantic region, extending southwards from the central East Greenland–mid-Norway–Møre region, the Jurassic–Cretaceous boundary is predominantly an unconformity, albeit commonly an intra-Berriasian (earliest Cretaceous) event. Significantly, perhaps, this coincides with the most pronounced phase of Ammonite provincialism, with Boreal and Tethyan communities completely cut off from each other during this interval (Rawson & Riley 1982). The unconformity is especially well developed in the

Faroe–Shetland–Hebrides region where it marks the termination of the Upper Jurassic–lowest Cretaceous black shales. Offshore Ireland, the Jurassic–Cretaceous boundary is always marked by one or more unconformities. Unconformities coalesce and individual unconformities do not extend regionally through a basin: however, the overall boundary is characterized by an unconformable nature. Some (diminishing) rifting prevailed across the boundary. In the Faroe–Shetland region, the Upper Jurassic–lowest Cretaceous black shales are locally overlain by paralic deposits, which suggest that there was subaerial exposure, at least locally. In contrast, the Upper Jurassic and lowest Cretaceous strata offshore Ireland is marine in nature, suggesting that there was no regional subaerial exposure in this region.

In the North Sea, the intra-Berriasian unconformity is generally termed the Late Cimmerian Unconformity (Base Cretaceous Unconformity) (Oakman & Partington 1998). This is described as a response to an eustatic sea-level lowstand combined with stress-induced deflection of the lithosphere, which led to earliest Cretaceous emergence and erosion of large parts of western and central Europe (Pharaoh *et al.* 2010). It also marks the approximate end of rifting in the North Sea as the influence of Tethyan spreading waned and Atlantic spreading intensified (Fig. 17) in response to the change in regional (plate-scale) stress directions from east–west in the Jurassic to NW–SE in the Early Cretaceous (Doré *et al.* 1999).

Cretaceous succession

Cretaceous rocks are widespread across the region and, whilst the stratigraphic range of the preserved succession is variable, its thickness is commonly of kilometre-scale (Fig. 4). In several areas, such as the Jameson Land Basin and the inner Hebrides margin, the Cretaceous is largely absent: whereas, in SE Greenland, the Rockall Plateau and the north Rockall Basin, Cretaceous rocks represent the first record of any Upper Palaeozoic–Mesozoic deposition in these areas. The succession mainly comprises marine facies dominated by mudstones, although carbonate rocks become increasingly common in Upper Cretaceous sequences south of the Faroe–Shetland region (Fig. 17), which coincides with the northern limit of the chalk sea in the North Sea region (Oakman & Partington 1998). According to Casey & Rawson (1973), an open-marine seaway had been established by Albian time connecting the Boreal and Tethyan realms via the North Sea. The timing of a fully marine connection with the developing Central Atlantic Ocean via the NW British–SE Greenland region remains equivocal (see below).

As Atlantic spreading propagated northwards, a broad zone of extension and subsidence developed stretching from the southern Rockall Basin to the western Barents Sea (Doré *et al.* 1999; Roberts *et al.* 1999). Along the northernmost edge of the study area, much of the rifting between North Greenland and the Svalbard–western Barents Sea region was progressively taken up by strike-slip movement within the De Geer Zone – a mega-shear system that linked the North Atlantic and Arctic regions prior to break-up (Eldholm *et al.* 2002) – leading to the formation of pull-apart basins in North Greenland and the SW Barents Sea (Håkansson *et al.* 1991; Smelror *et al.* 2009). In North Greenland, the Wandel Sea Basin developed as a pull-apart basin, which rapidly accumulated thick sequences of paralic and marine deposits during discrete, relatively short-lived pulses of rifting in the Early and Late Cretaceous. Activity in this strike-slip zone was terminated close to the Cretaceous–Paleocene boundary by compression associated with the Eureka Orogeny (Håkansson *et al.* 1991). In the SW Barents Sea, episodic rifting phases led to rapid subsidence and the development of a series of deep basins, such as the Harstad, Tromsø, Bjørnøya and Sørvestsnaget basins, which became decoupled from the rest of the Barents Sea shelf during the rifting. These basins accumulated thick sequences of predominantly marine mudstone.

At the same time, the northern part of the Barents Sea area was uplifted in the Early Cretaceous, which resulted in the deposition of a regressive clastic wedge on Svalbard. This Early Cretaceous uplift was associated with a major volcanic event in the Barents Sea region, including Svalbard, which coincided with the onset of break-up of the Amerasian Basin (Smelror *et al.* 2009). Uplift prevailed in the north, and by the Late Cretaceous large parts of the Barents Sea shelf – including Svalbard – had been uplifted.

The basins of NE and central East Greenland were mainly characterized by thermal subsidence and/or infilling of inherited rift relief, throughout the Cretaceous, accompanied by the deposition of predominantly marine basinal mudstones. This style of sedimentation and basin development was locally and sporadically interrupted by minor fault episodes that generated coarse-grained siliciclastic deposits (Whitham *et al.* 1999; Larsen *et al.* 2001; Surlyk & Noe-Nygaard 2001). Towards the end of the Cretaceous many of the onshore basins were uplifted and eroded, whereas the offshore Danmarkshavn and Thetis basins continued to subside and accumulate sediments (Hamann *et al.* 2005). In the Møre–mid-Norway region, the Vøring and Møre basins also underwent significant subsidence throughout the Cretaceous,

and accumulated thick sequences of predominantly marine mudstone. However, whereas the Møre Basin experienced mainly continuous subsidence, the Cretaceous development of the Vøring Basin comprised an early thermal subsidence phase and a post-Cenomanian phase of tectonically driven subsidence involving intermittent phases of normal faulting and compression and folding (Brekke 2000). This deformation has been linked to tectonic activity on structures, such as the Jan Mayen Lineament – the continental extension of the East Jan Mayen Fracture Zone (Blystad *et al.* 1995; Brekke 2000) (Fig. 1). Both the Møre and Vøring basins were uplifted at the end of the Cretaceous.

The Cretaceous development of the basins of the Faroe–Shetland region involved an Early Cretaceous rifting phase that became regionally expansive in the Aptian–Albian, and, as with the Vøring Basin, a post-Cenomanian–Turonian tectonically driven subsidence involving intermittent phases of normal faulting, compression and folding. The accumulation of upper Berriasian–Barremian paralic and shallow-marine deposits in basins on the West Shetland Shelf marks the instigation of rifting: however, it was not until the Aptian–Albian that marine clastic rocks became more widespread across the region as connectivity increased between basins. The development of the Faroe–Shetland Basin as an integrated, wide marine basin occurred in the Turonian–Maastrichtian interval, although the region as a whole was subjected to differential uplift and subsidence, which has been interpreted to be indicative of wrench tectonics (Dean *et al.* 1999; Roberts *et al.* 1999; Stoker 2016). A large part of the region was uplifted and eroded at the end of the Cretaceous, although depositional continuity may have prevailed in the deeper parts of the Faroe–Shetland Basin.

On the Hebrides Shelf, the Cretaceous succession is commonly absent due to uplift and erosion, although a fragmentary Upper Cretaceous shallow-marine clastic and carbonate sequence is preserved in the Inner Hebrides. On the Outer Hebrides Shelf, Barremian–Aptian and younger Cretaceous marine clastic deposits are preserved in the West Lewis Basin, and have also been proved in the adjacent northern Rockall Basin. Comparable sequences of paralic to marine clastic rocks are present in half-graben on the north Hatton High (outer Rockall Plateau), and in the Kangerlussuaq and Ammassalik basins of SE Greenland. In the northern Rockall Basin–Rockall Plateau–SE Greenland region, the Cretaceous rocks represent the first observable indications of rift activity in this area. The folded, faulted and eroded character of the Cretaceous rocks on the north Hatton High suggests that they may have been more extensive prior to

the break-up between SE Greenland and the Rockall Plateau.

The Cretaceous succession is extensively preserved offshore Ireland, and is predominantly marine in nature. The Porcupine Basin developed mainly by thermal subsidence, initially following Late Jurassic rifting and, subsequently, in response to Late Cretaceous seafloor spreading in the southernmost North Atlantic. This general pattern of subsidence was interrupted in the Aptian–Albian by a phase of rifting, which also affected the NW Irish basins. The southern Rockall Basin may have undergone a similar development. The Lower Cretaceous succession is predominantly clastic in nature (sandstones and mudstones), and was deposited typically in an outer-shelf to deep-marine setting, with ponded turbidites in residual rift topographical lows and in basin-floor settings. Localized Aptian–Albian deltaic sandstones occur on the flanks of the Porcupine Basin and represent the deposits of minor rifting. The Upper Cretaceous is chalk-dominant, and is present in all basins covering Jurassic footwalls and extending onto basement highs and beyond the edges of the rift basins. Whereas an unconformity separates Cretaceous and Paleocene rocks on the flanks of the Porcupine and southern Rockall basins, the transition in deeper water was more continuous.

It may be no coincidence that the pulse of Aptian–Albian rifting that is widely recorded across the southern part of the study area correlates with the onset of break-up between Newfoundland and Iberia (Doré *et al.* 1999) (Fig. 17). The increasing focus of crustal extension within the Rockall–Faroe–Shetland and Hatton–SE Greenland regions reflects this change in regional stress directions (as described above). This was accompanied by Early Cretaceous volcanism in the central part of the Porcupine Basin and in the southern Rockall Basin (areas of crustal hyperextension according to Lundin & Doré 2011), and extrusives and tuffs occur through the Barremian–Aptian in the Porcupine Basin. The tectonic pulses in Late Cretaceous (post-Cenomanian) times, observed across a large part of the study area, were probably coupled to the opening of the Labrador Sea and the associated anti-clockwise rotation of Greenland (Brekke 2000). However, the different structural expressions (faulting, folding and inversion) indicate that the intra-plate stress regime changed through time, and the build-up of compressional stresses as part of the evolving Alpine orogen might also have played an important role in modulating stresses throughout the plate (Pharaoh *et al.* 2010). Volcanism continued to prevail in the Rockall Basin with the intrusion of the Late Cretaceous Anton Dohrn and Rosemary Bank igneous centres in the northern part of the basin.

Implications for palaeogeographical reconstruction of the NE Atlantic region

We acknowledge all existing palaeogeographical syntheses that have been established for the NE Atlantic region in terms of their important contribution to the on-going interpretation and discussion of the evolution of this area. Ideas are, however, being constantly revised. As such, the distribution and stratigraphic range of the Permian–Cretaceous rock record that are presented in this paper enable us to test the viability of putative palaeogeographical reconstructions in terms of the accuracy of the reconstructed plate-tectonic base maps, and the history of basin development. Notwithstanding the limitations commonly inherent in plate-tectonic reconstructions (e.g. continental shortening, problems of de-stretching the crust), our basic stratigraphic distribution maps help to illustrate two key problems that are common to many of the existing reconstructions:

- Early Mesozoic plate reconstructions based on palaeomagnetic interpretations (e.g. Ziegler 1988; Coward *et al.* 2003; Pharaoh *et al.* 2010; Torsvik *et al.* 2012) result in a re-fit of the conjugate margins of NE Greenland and Norway that is too tight (i.e. overlapping), leaving little or no space available to accommodate the offshore sedimentary basins that are depicted in Figures 1–4. Whereas the Mid-Jurassic plate reconstruction of Nøttvedt *et al.* (2008) acknowledges the distribution of the Mesozoic basins along the conjugated margins of East/NE Greenland and Norway, the new NAG-TEC data are better constrained and show more detailed basin distributions (Figs 2–4).
- The reconstruction of the area between SE Greenland and NW Britain–Ireland is a long-standing problem. Not only is the amount and timing of extension in this area controversial (e.g. Roberts *et al.* 1988, 1999; Doré 1992; Shannon *et al.* 1995, 1999; Doré *et al.* 1999), the stratigraphic record (as shown in the NAG-TEC data) and, thus, the basin history remain largely unknown (see below) (Figs 2–4 & 17). Despite this, an arrangement of structural highs and basins, as well as inferred rifts, such as the ‘Hatton–Greenland rift’, comparable to the modern-day structural and bathymetric framework is commonly utilized in all reconstructions back to, and including, the Permian (e.g. Ziegler 1988; Cope *et al.* 1992; Doré 1992; Knott *et al.* 1993; Torsvik *et al.* 2002; Coward *et al.* 2003; McKie & Williams 2009; Pharaoh *et al.* 2010).

Classic plate reconstructions do not address this complexity: thus, existing palaeogeographical maps

of this region should be regarded as highly speculative.

In terms of basin history, existing palaeogeographical syntheses have been, and continue to be, updated and modified as more knowledge becomes available. This is especially the case in the North/NE Greenland–western Barents Sea–Svalbard region, where a series of studies (e.g. Doré 1992; Roberts *et al.* 1999; Torsvik *et al.* 2002; Nøttvedt *et al.* 2008; Smelror *et al.* 2009) show a progressive refinement of the palaeogeography. The stratigraphic data presented in this paper should add to this growing database.

By way of contrast, the geological history of the southern part of the study area, especially between SE Greenland and NW Britain, remains poorly understood. For reasons noted above, the structural development of this region remains equivocal and, coupled with the fragmentary nature of the stratigraphic record, challenges the basis of those maps (e.g. Ziegler 1988; Doré 1992; Knott *et al.* 1993; Roberts *et al.* 1999; Torsvik *et al.* 2002; Coward *et al.* 2003; Pharaoh *et al.* 2010) that detail one or several Late Palaeozoic–Early Mesozoic through-going rift systems: that is, along the axis of the Faroe–Shetland and Rockall basins, as well as the inferred ‘SE Greenland–Hatton rift’. Inspection of Figures 2–4 indicates that, on the basis of the available evidence (or lack of), the long-term existence of the latter rift is totally without foundation. Currently, the only proven Mesozoic rocks on the Hatton and SE Greenland margins are of Cretaceous age. Moreover, the distribution of Cretaceous strata across a large part of this region remains inferred. Whilst the Cretaceous rocks on the Hatton High have been subjected to later folding, uplift and erosion, which implies a greater former extent, the relatively isolated occurrence of the basins necessitates caution in applying interbasinal connectivity across this region.

The issue of interbasinal connectivity is also pertinent with regard to the validity of a long-lived Faroe–Shetland–Rockall rift. On most reconstructions, the depiction of this proposed rift zone largely mimics the current bathymetric expression of the continental margin off NW Britain and Ireland: however, this morphological expression is largely a late-stage response to Cenozoic subsidence and post-break-up ‘passive margin’ tectonics (Naylor & Shannon 2005; Praeg *et al.* 2005; Ritchie *et al.* 2011b, 2013). It is also increasingly apparent that the Faroe–Shetland and Rockall basins are both segmented by NW-trending lineaments, at least some of which may have been inherited from earlier structures (Kimbell *et al.* 2005). For example, the major offset (and change in trend) between the northern and southern parts of the Rockall Basin (Fig. 1) is related to the Anton Dohrn Lineament

Complex (cf. Kimbell *et al.* 2005), which represents a transfer zone that may have originated as a Precambrian terrane boundary. Doré *et al.* (1999) have suggested that the current offset relates to an Early Cretaceous event; in which case, the Late Palaeozoic–Early Mesozoic development of the southern Rockall Basin might be linked more to basins in the Hebridean region. In this scenario, the status of the northern Rockall Basin, and even the Hatton Basin, remains unclear. Thus, any southern connection with the Faroe–Shetland region might have been via the Hebridean basins, although the latest Triassic–Jurassic stratigraphic record (Fig. 17) indicates sporadic and intermittent deposition of mainly paralic and shallow-marine deposits in both these areas. More specific stratigraphic information includes the near-absence of deposits associated with either the Zechstein or Rhaetian transgressions offshore NW Britain (Fig. 17), which further questions the validity of widespread marine connections in this region, at this time (e.g. Ziegler 1988; Knott *et al.* 1993; Roberts *et al.* 1999; Coward *et al.* 2003; Pharaoh *et al.* 2010).

Of the published plate reconstruction models, the scenario that most accurately complements the proven Late Mesozoic rock record, as presented here, is that of Doré *et al.* (1999), who linked changes in basin development across the NW European and NE Atlantic regions to the change in regional extensional stress directions: that is, from east–west (Jurassic) to NW–SE (Cretaceous) in response to plate break-up. Whilst the possibility of a more extensive Jurassic basin development in the area between NW Britain and SE Greenland remains to be fully tested, the currently available data suggest that it was not until the Late Cretaceous that a substantive rift system linking the Arctic and NE Atlantic regions across the SE Greenland–NW British region was developed. This is consistent with the long-standing view, based on studies of faunal provinciality (as described above), that there was a general lack of continuity between the Boreal and Tethyan realms during much of the Mesozoic.

Conclusions

An overview of the Upper Palaeozoic–Mesozoic rock record around the NE Atlantic region has produced a better understanding of the regional stratigraphic response to the break-up of the Pangaeian supercontinent. By combining the distribution and stratigraphic range of the preserved Permian–Triassic, Jurassic and Cretaceous successions we have been able to identify both spatial and temporal changes in the large-scale pattern of sedimentation and basin development throughout the region, and

link them to established tectonic events within and on the margins of the Pangaeian plate. In particular:

- The Carboniferous–Permian boundary is largely conformable on the northern margin of the study area, where a passively subsiding margin formed in the late Carboniferous and persisted until the Mid–Late Permian. Over most of the rest of the NE Atlantic region, the boundary is a regionally extensive unconformity.
- The Permian–Triassic succession records a northwards transition from an arid, terrestrial facies to a shallow-marine shelf facies that dominates along the northern margin of the study area. Whereas the arid, terrestrial facies is mainly Late Permian–Triassic in age, the shelf facies comprises Permian carbonate rocks that pass upwards into Upper Permian–Triassic siliclastic deposits. A pulse of rifting affected the entire NE Atlantic region during the Late Permian–earliest Triassic interval that generated variable large-scale basin architectures both in the interior and on the northern margin of the plate.
- The Triassic–Jurassic boundary is variable in character: in the north, paralic to marine clastic deposition persisted locally, although elsewhere the boundary is an unconformity; in the central part of the study area, the boundary is mainly an unconformity; and, in the south, a transitional passage from the arid Triassic into marine Jurassic represents the Rhaetian transgression. This variation is inferred to be a response to tectonic events occurring on the flanks of the Pangaeian plate, with a southwards propagation of the rift system between Greenland and Norway, and marine flooding of the southern margin in response to incipient seafloor spreading in the Tethys and the proto-central Atlantic.
- The Jurassic succession preserves a record of Early Jurassic thermal subsidence and mild extensional tectonism, with a fragmentary succession of coastal and shallow-marine deposits predominating across the entire region. This general pattern of development was interrupted in the Mid-Jurassic by widespread uplift and shallowing, and is commonly reflected in the stratigraphic record as an unconformity between Lower and Middle Jurassic rocks. An intense phase of rifting occurred in some (but not all) parts of the NE Atlantic region in late Mid–Late Jurassic time, and there is a general trend towards marine conditions. This phase of rifting may have been driven by seafloor spreading in the Tethys; on the northern margin some of the tectonic activity might also have been related to extension in the Amerasia Basin.
- The Jurassic–Cretaceous boundary is transitional in North and NE Greenland, and marks

THE UPPER PALAEOZOIC–MESOZOIC OF THE NE ATLANTIC

the culmination of rifting in this region. Elsewhere, the boundary is predominantly an unconformity, albeit commonly an intra-Berriasian (earliest Cretaceous) event. Its origin has been linked to a eustatic sea-level lowstand combined with stress-induced deflection and uplift of the lithosphere and/or a change in regional stress directions as Tethyan spreading waned and Atlantic spreading intensified.

- The Cretaceous succession is the most widely developed system of rocks proven to occur in the NE Atlantic region. Marine rocks dominate the observed record, with Upper Cretaceous chalks restricted to the southern part of the study area. As Atlantic spreading propagated northwards, a broad zone of extension and subsidence developed stretching from the southern Rockall Basin to the western Barents Sea. Basin development within this zone is variable: strike-slip movement and pull-apart basins characterize the northern margin of the area, whereas thermal subsidence prevailed in a number of basins in the central and southern part of the region. Regional Aptian–Albian rifting included the instigation of Mesozoic basin development in the area between SE Greenland and NW Britain. A post-Cenomanian phase of tectonically driven subsidence involving intermittent phases of normal faulting, compression and folding is recognized, especially in the successions of the Faroe–Shetland region and the Vøring Basin, and has been linked to wrench tectonics within the Pangaeian plate. It has been inferred that the intra-plate stress regime at this time may have been modulated by a combination of Atlantic spreading (extension) and the evolving Alpine orogen (compression).

It is clear from the stratigraphic distribution maps that substantially more information remains to be gathered from around the NE Atlantic region: indeed, many areas remain that need to be fully explored. Nevertheless, we feel that there is scope for the development of a new and more comprehensive model of NE Atlantic rifting that takes into account the stratigraphic information presented in this paper. A key question remains: what was the degree of continuity along the proto-NE Atlantic during the Mesozoic? Our stratigraphic data suggest either a lack of continuity or, at least, a degree of complexity to any continuity that is currently not part of any reconstruction.

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