Structural evolution and basin architecture of the Traill Ø region, NE Greenland: A record of polyphase rifting of the East Greenland continental margin


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ABSTRACT

Fault block basins exposed along NE Greenland provide insights into the tectonic evolution of East Greenland and the Norwegian-Greenland Sea. We present a new geological map and cross sections of the Traill Ø region, NE Greenland, which formed the western margin of the Voring Basin prior to Cenozoic seafloor spreading. Observations support a polyphase rift evolution with three rift phases during Devonian–Triassic, Jurassic–Cretaceous, and Cenozoic time. The greatest amounts of faulting and block rotation occurred during Cenozoic rifting, which we correlate with development of the continent-ocean transition after ca. 56 Ma and the Jan Mayen microcontinent after ca. 36 Ma. A newly devised macrofaunal-based stratigraphic framework for the Cretaceous sandy mudstone succession provides insights into Jurassic–Cretaceous rifting. We identify a reduction in sedimentation rates during the Late Cretaceous; this corresponds to a transition from structurally confined to unconfined sedimentation that coincides with increased clastic sedimentation to the Voring and Møre Basins derived from East Greenland. With each rift phase we record an increase in the number of active faults and a decrease in the spacing between them. We attribute this to fault block rotation that leads to an excess build-up of stress that can only be released by the creation of new steep faults. In addition, we observe a stepwise migration of deformation toward the rift axis that we attribute to preexisting lithospheric heterogeneity that was modified during subsequent rift and post-rift phases. Such observations are not readily conformance to classic rift evolution models and highlight the importance of post-rift lithospheric processes that occur during polyphase rift evolution.

INTRODUCTION

Cenozoic uplift and exhumation of the East Greenland margin (Bonow et al., 2014) has exposed a series of east-west–extended fault blocks, defined largely by eastward-dipping faults (Fig. 1) developed during multiple rift phases since the Devonian (Surly, 1990; Price et al., 1997; Whitham et al., 1999; Lundin and Dore, 2002). The record of Carboniferous–Eocene sedimentation preserved within these basins is arguably one of the most important regional sources of geological information for understanding the fill of offshore basins on East Greenland and northwest European continental shelves (Kelly et al., 1998; Whitham et al., 1999). The exhumed Cenozoic sills, plutons, and plateau basalts intruded into and extruded onto these sediments provide an important record of the final stages of continental extension and subsequent initiation of ocean spreading (e.g., Larsen et al., 2014). Importantly, paleogeographic reconstructions prior to Cenozoic seafloor spreading place these basins against the western margins of the Voring and Møre Basins, both of which are major hydrocarbon exploration targets (Doré, 1991; Lundin and Dore, 1997; Olesen et al., 2007). Consequently, the exhumed basins of East Greenland provide a record of the post-Triassic structural and sedimentological evolution of the northeast Atlantic region and are an important analogue for submerged basins in the Norwegian-Greenland Sea (Doré and Jensen, 1998; Price and Whitham, 1997; Whitham et al., 1999). Here we present a new 1:100,000-scale geological map (Enclosure 1) and cross sections (Enclosure 2) of Traill Ø and Geographical Society Ø in NE Greenland (Fig. 2) that focus on the distribution of Jurassic and Cretaceous strata. A Google Earth compatible drape of the geological map of the Traill Ø region (.kmz file) is available in Supplemental Materials 1. An important mapping objective was subdivision of the Ryazanian–Campanian succession, a largely monotonous unit of mudstones. This was achieved through mapping the distribution of macrofaunas unique to different stratigraphic stages. The resulting map provides an improved record of the regional geology and new insights into the structural and sedimentological evolution of the region.
Figure 1. Geological map of NE Greenland: FRD—Fjord region detachment; GHF—Gauss Halve fault; MF—Månedal fault; WFZ—Western fault zone (after Escher and Pulvertaft, 1995). Locations of geophysical surveys on which Figure 14 is based on are shown: green line—KJF-3D-density cross section (Schmidt-Aursch and Jokat, 2005); blue line—seismic reflection profile 94320 (Schlindwein and Jokat, 1999); black line—seismic reflection profile AWI-20030050 (Voss and Jokat, 2007).
Cross sections through the Traill Ø Region

Enclosure 2. Geological cross sections through Traill Ø and Geographical Society Ø, NE Greenland. Gp.—group; Fm.—formation. To view Enclosure 2 at full size, please visit http://doi.org/10.1130/GES01382.07 or the full-text article on gsapubs.org.
## Enclosure 3: Lithostratigraphy

<table>
<thead>
<tr>
<th>Period, Epoch, and/or Stage</th>
<th>Lithostratigraphic Environment</th>
<th>Key Biostratigraphic Indicators in Traill Ø &amp; Geographical Society Ø</th>
<th>Mapped Units</th>
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<tbody>
<tr>
<td>Mesozoic</td>
<td></td>
<td></td>
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<tr>
<td>Triassic</td>
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<tr>
<td>Cenozoic</td>
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</tbody>
</table>

Enclosure 3. Lithostratigraphic framework for Traill Ø and Geographical Society Ø, NE Greenland. Abbreviations: Gp.—group; Fm.—formation; sst—sandstone; lst—limestone; mst—mudstone; cg—conglomerate; dol—dolomite. To view Enclosure 3 at full size, please visit [http://doi.org/10.1130/GES01302.58](http://doi.org/10.1130/GES01302.58) or the full-text article on www.gsapubs.org.
The Cretaceous strata that crop out in the Traill Ø region are of great regional importance. Not only do they cover large areas, but exposures of Cambrian strata in this area are the youngest exposed Cretaceous strata known in the Norwegian-Greenland Sea region, with the exception of Kangerlussuaq in southern East Greenland (Kelly et al., 1998, 2015). As part of the western margin of the Vøring and Møre Basins in the Cretaceous, the Traill Ø region also provides vital regional data for understanding how these basins were filled by sediments in the Late Cretaceous. Heavy mineral provenance analyses of sandstones from the Traill Ø region and the Vøring and Møre Basins indicate that large quantities of sediment sourced from the East Greenland rift margin entered the Vøring and Møre Basins during the Late Cretaceous (Morton et al., 2005).

Consideration of temporal and spatial variations in stratigraphic angular discordances, fault orientations and bedding dips, fault block widths, stratigraphic geometries, and offset marker horizons provides an insight into fault activity in the Traill Ø region. These observations define a polyphase rift evolution characterized by at least three distinct phases of rifting during Devonian–Triassic, Jurassic–Cretaceous, and Cenozoic intervals, separated by post-rift phases. Our record of faulting agrees with previous work (e.g., Price et al., 1997; Schindlwein and Jokat, 1999; Voss and Jokat, 2007, 2009) that describe an eastward migration (toward the rift axis) of the locus of concentrated extensional deformation and an increase in the number of active faults and a decrease in the spacing between them with each rift phase. This style of rift evolution is not readily conformable with classical strain localizing rift evolution models that display a temporal decrease in the number of active faults, and increase in the spacing between them (e.g., Gupta et al., 1998; Cowie et al., 2005). We suggest that fault block rotation during domino-style rifting (e.g., Jackson and McKenzie, 1983) led to a reduction in the efficiency of preexisting faults as their dips reduced during rotation. This led to a build-up in stress within the fault blocks that eventually caused the fault blocks to rupture and disperse through creation of new steeper faults (e.g., Jackson and McKenzie, 1983; Jackson and White, 1988).

Through consideration of the structural and rheological evolution of the East Greenland continental lithosphere, as determined from previous geo-physical analyses (e.g., Schindlwein and Jokat, 1999, 2000; Schmidt-Aursch and Jokat, 2005; Voss and Jokat, 2007, 2009; Hermann and Jokat, 2016), we suggest that the eastward migration of deformation and development of new faults may have been driven by evolving heterogeneities in lithospheric rheology. These heterogeneities initially developed during crustal thickening of the Caledonian orogen and were modified during subsequent rift and post-rift phases. Our findings highlight an important distinction between the evolution of progressive rift systems defined by only a single episode of rifting and polyphase rift systems in which post-rift lithospheric processes occurring between syn-rift phases can significantly influence the evolution of strain distribution. Our findings are comparable to those presented by Bell et al. (2014), who propose a similar polyphase rift evolution for the northern North Sea influenced by post-rift lithospheric modification between two distinct rift phases.

**MAP CONSTRUCTION**

The Traill Ø region as defined here consists of Traill Ø, Geographical Society Ø, and associated smaller islands covering an area of ~8000 km² (Fig. 2). The geological map and cross sections of this region are constructed primarily from data collected by CASP (University of Cambridge) geologists during field work in 1990–1996, 2001–2003, 2005–2006, and 2009–2011 (Fig. 3). For ease of reference, reduced-size versions of the cross sections, displayed with a 2:1 vertical to horizontal scale, are presented in Figure 4. For the full-size 1:100,000-scale geological map, 1:1 vertical to horizontal scale cross sections, and a lithostratigraphic column for the geology of the Traill Ø region, please refer to Enclosures 1–3. A Google Earth compatible drape of the geological map of the Traill Ø region (.kmz file) is available in Supplemental Materials 1 (see footnote 1).

The Traill Ø region has previously been mapped by Koch and Haller (1971), Harpøth et al. (1986), and Escher (2001). These maps form the main source of information in areas that have not been visited by CASP geologists (Fig. 3). Our mapping focuses on the distribution of Jurassic and younger sediments and the structure and distribution of faults and fault blocks. The mapped distribution of Devonian to Triassic sediments and the faults identified within them is largely unmodified from previously published maps.

Data collected by CASP geologists used to create the map and cross sections in this study include structural (e.g., bedding dip readings, fault plane orientations), lithostratigraphic, and biostratigraphic data. These data were supplemented with interpretations of aerial photographs and Ladsat images and additional information from Koch and Haller (1971), Harpøth et al. (1986), Escher (2001), and Seidler et al. (2004). Topographic data were provided by the United States Geological Survey (USGS). Rivers, lakes, and coast outlines for much of the region were provided by the Geological Survey of Denmark and Greenland (GEUS). The World Geodetic System 1984 was used as the reference system and geodetic datum for map projection (United States National Imagery and Mapping Agency, 2000). Structural data presented in Enclosure 1 can be found in Supplemental Materials 1.3.

The lithostratigraphic framework (Enclosure 3) is largely based on previously published sources for the pre-Ryazanian succession. A major challenge has been the subdivision and mapping of the lithologically monotonous Ryazanian–Campanian succession. This has been achieved using the distribution of macrofaunas observed and collected during field work combined with existing sources (see “Cretaceous Strata (Ryazanian–Campanian): Wollaston Fork and Hold with Hope Groups” section below for references used). The position of stratigraphic boundaries within the Ryazanian–Campanian succession is extrapolated from the locations of key macrofauna (Enclosure 3). In some areas faults are inferred to account for discrepancies between the distribution of macrofauna localities and dips of bedding. The locations of macrofossils identified by this study and from previously published sources are shown on the new map (Enclosure 1). A list of all macrofauna used for map construction, including those used for subdivision of the Ryazanian–Campanian succes-
Figure 2. Geological map of the Traill Ø region. Most dolerite intrusions are removed for clarity. Locations of field photographs presented in Figures 10–12 and their fields of view are outlined by black dash-lined boxes. Selected faults: GHF—Gauss Halve fault; LBF—Laplace Bjerg fault; MF—Månedal fault; MBF—Mols Bjerge fault; VF—Venildal fault. (For the full size map and cross sections, see Enclosures 1 and 2.) Gp.—group; Fm.—formation.
Figure 3. Map delimiting the areas mapped by individual geologists. All areas were mapped by foot traverses; the black dots indicate points where geological observations were recorded. Each numbered area was mapped at a scale of 1:50,000 by the following geologists:


A comprehensive outline of the lithostratigraphic framework implemented by this study with details of the published sources is presented in Supplemental Materials 4. Structural and biostratigraphic data collected during field work were located using global positioning system waypoints or located on aerial photographs and satellite images. Where possible, stratigraphic boundaries and faults were located in the field, supplemented by analysis of aerial photographs, Landsat images, field photographs, and information from previously published maps (Koch and Haller, 1971; Harpeth et al., 1986; Escher, 2001). Three classes of boundary accuracy are used on the map, applied to both stratigraphic contacts and faults: (1) observed; (2) inferred, well constrained; and (3) inferred, poorly constrained. Faults are drawn as red lines and stratigraphic contacts as black lines. Observed boundaries are boundaries that have been seen by a geologist and occur in areas of good exposure where the boundary can be placed to within 10 m. Inferred boundaries are used in areas of poor exposure and/or poor data coverage. These inferred boundaries are in many cases extrapolated from places where they are known across the map using the structure contour boundary projection method (e.g., Bennison et al., 2013), via the use of the CASP 3DPlanes Program (University of Cambridge). The two types of inferred boundary, well constrained and poorly constrained, reflect the degree of confidence and accuracy in the location of the inferred boundary.

Topographic lineaments identified from aerial photographs, Landsat images, and digital elevation models are also plotted on the map as potentially unidentified faults. These lineaments are plotted on the map with blue boundary lines (Enclosure 1), but are not included on cross sections as their significance and three-dimensional structure are unknown. It is suggested that these lineaments could form a point of interest for future field research in the region and highlight the possibility that current data sets may underestimate the amount of faulting in the region. Fault block rotation is based on the dip of the oldest strata assumed to be originally deposited as horizontal layers (i.e., ignoring clearly syn-rift packages). These dips were either measured in the field or extrapolated in the subsurface from cross sections. Variations in amounts of fault block rotation between different time periods are based on relative change in dip between underlying and overlying strata (i.e., a dip of 14° in Jurassic strata and 10° in overlying Cretaceous strata is interpreted as a fault block rotation of 4° during the Jurassic–Cretaceous). In some places the dip of igneous sills was taken as a rough estimate of bedding dip. For unconstrained fault plane orientations, prescribed fault dips are based on the tilt of bedding in the footwall and hanging wall, interpreted as a record of fault block rotation. A footwall dip of 10°, in the opposite direction to fault throw, corresponds to fault block rotation of 10° and a fault dip of 50°, assuming equal footwall and hanging-wall rotations and an initial fault plane dip of 60° (e.g., Anderson, 1951). Where fault plane orientation and footwall and hanging-wall strata orientation are well constrained yet do not correspond to the same amount of fault block rotation, internal deformation of fault blocks in the form of fault drag folds is inferred to allow cross sections to be restorable. The faults and fault blocks have been numbered for simplicity (Figs. 5 and 6) and are listed in Tables 1 and 2. A description of all numbered faults and fault blocks is provided in Supplemental Materials 5.

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Supplemental Materials 3. List of all macrofauna used for map construction, including those used for subdivision of the Ryazanian–Campanian succession. Please visit http://doi.org/10.1130/GES01382.S3 or the full-text article on www.gsapubs.org to view Supplemental Materials 3.


Supplemental Materials 5. Description of all numbered faults and fault blocks. Please visit http://doi.org/10.1130/GES01382.S5 or the full-text article on www.gsapubs.org to view Supplemental Materials 5.
Figure 4 (on this and following page). Structural cross sections through the Traill Ø region, NE Greenland, displayed at a vertical to horizontal scale of 1:1. Locations of cross sections are in Figure 2 (see also Enclosures 1 and 2; for full-size cross sections at 1:1 vertical to horizontal scale with inset maps showing cross-section locations, see Enclosure 2). Gp.—group; Fm.—formation.

Stratigraphic units

- Plateau basalts
- Sub-Basaltic Marine Beds
- Scophites Beds channel fill
- Scophites Beds
- Sphenoceras Beds
- Inoceramus lamordi Beds
- Inoceramus crippsi Beds
- Inoceramus anglicos Beds
- Inoceramus aucto Beds
- Aulcl Beds
- Bernbjerg Fm.
- Olympus Fm.
- Britzal Elv Fm. & Pelion Fm.
- Scoresby Land Gp.
- Foldvik Creek Gp.
- Traill Ø Gp.

Intrusive Igneous units

- Dolerite sills and dikes (mid Paleocene - early Oligocene)
- Syenite pluton (late Eocene - early Oligocene)

D. - Devonian; Ca. - Carboniferous;
Pe. - Permian; Tr. - Triassic; Pa. - Paleocene;
Th. - Thanetian; Eo. - Eocene; Yp. - Ypresian

Symbology

- Contact, mapped
- Contact, inferred, well constrained
- Contact, inferred, poorly constrained
- Fault, observed
- Arrow mark indicates direction of hanging wall-slip
- Fault, inferred
- Arrow mark indicates direction of hanging wall-slip
- Low-angle sub-surface structure identified from seismic reflection - structural nature unknown
- Bedding plane
- Topography
Figure 4 (continued).
Figure 5. Map of the Traill Ø region showing the locations of named faults, highlighted in red. Dip ticks point toward the hanging wall. Red solid line—observed normal fault; red dashed line, large dashes—inferrred normal fault, well constrained; red dashed line, small dashes—inferrred normal fault, poorly constrained; solid thick black line—line of section (A–F).

Selected faults: GHF—Gauss Halvø fault; LBF—Laplace Bjerg fault; MF—Månedal fault; MBF—Mols Bjerge fault; VF—Vælddal fault. Numbers refer to individually named faults. See Table 1 for a summary of all faults: 1—Gauss Halvø fault, Traill Ø; 2—Rubjerg Knude fault, Traill Ø; 3a—Bordbjerge north strand fault, Traill Ø; 3b—Bordbjerge south strand fault; 4—Svinhuvud Bjerge faults; 5a, 5b, 5c, 5d—Månedal fault zone; 5e—Månedal fault A; 5f—Månedal fault B; 6—Skelhøje fault; 7a, 7b—Mols Bjerge fault; 8a, 8b, 8c—Vælddal fault; 9—Gauss Halvø fault, Geographical Society Ø; 10—Rubjerg Knude fault, Geographical Society Ø; 11—Bordbjerge north strand fault, Geographical Society Ø; 12—Månedal fault, Geographical Society Ø; 13—Månedal fault north splay; 14—Lydal fault; 15—Lydal east branch fault; 16—Langbjerg fault; 17—Laplace Bjerg fault; 18—Laplace Bjerg branch fault; 19—Cambridge Bugt south fault; 20—Cambridge Bugt north fault; 21—Kap Mackenzie fault; 22a, 22b, 22c, 22d, 22e—Vega Sund cross faults.

Normal fault, observed
Normal fault, inferred, well constrained
Normal fault, inferred, poorly constrained
Figure 6. Fault blocks of the Trail Ø region. Numbers refer to individually named fault blocks. Fault blocks are defined by the footwall strata of faults listed in Figure 5 and are typically named after the fault to which they form the footwall. Selected faults: GHF—Gauss Halvø fault; LBF—Laplace Bjerg fault; MF—Månedal fault; MBF—Mols Bjerge fault; VF—Vælddal fault. See Table 2 for a summary of all fault blocks: 1—Rubjerg Knude block; 2—Bordbjerget block; 3—Månedal block, Trail Ø; 4—Svinhufvud Bjerge block; 5—Flakkebjerg block; 6a—Vælddal block; 6b—Vælddal fault slice 1; 6c—Vælddal fault slice 2; 7a—Månedal block 1; 7b—Månedal block 2; 7c—Månedal fault slice 1; 7d—Månedal fault slice 2; 7e—Månedal fault slice 3; 8—Ellemandsbjerg block; 9—Månedal block, Geographical Society Ø; 10—Lysdal block; 11a, 11b, 11c—Mid Geographical Society Ø blocks; 12—Laplace Bjerg horst; 13—Cambridge Bugt block; 14a—Kap Mackenzie block; 14b—Kap MacClintock block; 15—Fault Block 15.
<table>
<thead>
<tr>
<th>Fault number</th>
<th>Fault name</th>
<th>Location</th>
<th>Fault orientation</th>
<th>Displacement (m vertical throw)</th>
<th>Known activity*</th>
<th>Possible activity*</th>
<th>Data source†</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Gauss Halvø fault</td>
<td>Traill Ø</td>
<td>50/120</td>
<td>2000–3000</td>
<td>D, C, Cz</td>
<td>P, T, J–K</td>
<td>Koch and Haller (1971); field work (this study); Butler (1955); Peacock et al. (2000)</td>
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<td>2</td>
<td>Rubjerg Knude fault</td>
<td>Traill Ø</td>
<td>50/120</td>
<td>≥1000</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971)</td>
<td>Assumed to copy fault 1</td>
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<td>3a</td>
<td>Bordbjerget north strand fault</td>
<td>North central Traill Ø</td>
<td>50/090 to 115/115</td>
<td>1050–1150</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971); field work (this study); LandSat image; 3DPlanes Program</td>
<td>0° dip direction at Rold Bjerg; 115° dip direction at Grønne Bjerge</td>
</tr>
<tr>
<td>3b</td>
<td>Bordbjerget south strand fault</td>
<td>South central Traill Ø</td>
<td>50/055 to 110</td>
<td>1000–1250</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971); Price et al. (1997); field work (this study); LandSat image; aerial photographs</td>
<td>Exists as fault zone at coast</td>
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<td>4</td>
<td>Svinhufud Bjerge faults</td>
<td>Svinhufud Bjerge, Traill Ø</td>
<td>55/110</td>
<td>≥850 (cumulative)</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971); Price et al. (1997); field work (this study); LandSat image; digital photographs</td>
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<tr>
<td>5a</td>
<td>Månedal fault zone, Månedal fault</td>
<td>Månedal, Traill Ø</td>
<td>45/088 to 100/085</td>
<td>2400 on fault 5a, 5500 (cumulative) across faults 5a, 5b, 5c, 5d</td>
<td>J–K, Cz</td>
<td>–</td>
<td>Koch and Haller (1971); field work (this study); LandSat image; field slips; 3DPlanes Program</td>
<td>Fault trace is nonlinear; dip direction rotates toward east-southeast south of Månedal</td>
</tr>
<tr>
<td>5b</td>
<td>Månedal fault zone, splay 1</td>
<td>Månedal, Traill Ø</td>
<td>50/095</td>
<td>1950</td>
<td>Cz</td>
<td>–</td>
<td>Field work (this study); fossil localities; 3DPlanes Program; archival photographs; digital photographs</td>
<td>Nonlinear fault trace, roughly parallel with fault 5a</td>
</tr>
<tr>
<td>5c</td>
<td>Månedal fault zone, splay 2</td>
<td>Månedal, Traill Ø</td>
<td>50/110</td>
<td>500</td>
<td>Cz</td>
<td>–</td>
<td>Fossil localities; 3DPlanes Program; archival photographs; digital photographs</td>
<td></td>
</tr>
<tr>
<td>5d</td>
<td>Månedal fault zone, splay 3</td>
<td>Månedal, Traill Ø</td>
<td>50/110</td>
<td>750</td>
<td>Cz</td>
<td>–</td>
<td>Field work (this study); fossil localities; 3DPlanes Program; archival photographs; digital photographs</td>
<td>May be southward continuation of fault 14</td>
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<td>5e</td>
<td>Månedal fault A</td>
<td>Svinhufud Bjerge, Traill Ø</td>
<td>48/105 to 55/105</td>
<td>1600–2600</td>
<td>J–K, Cz</td>
<td>–</td>
<td>Koch and Haller (1971); field work (this study); aerial photographs</td>
<td>48° dip in north; 55° dip in south; segmented</td>
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<td>5f</td>
<td>Månedal fault B</td>
<td>Svinhufud Bjerge, Traill Ø</td>
<td>50/090 to 50/085</td>
<td>≥500</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971); field work (this study); aerial photographs; LandSat image; 3DPlanes Program</td>
<td>Poorly exposed; nonlinear fault trace</td>
</tr>
<tr>
<td>6</td>
<td>Skelheje fault</td>
<td>Månedal, Traill Ø</td>
<td>50/110</td>
<td>200–400</td>
<td>Cz</td>
<td>–</td>
<td>Field work (this study); fossil localities; aerial photographs; topography</td>
<td></td>
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<tr>
<td>7a</td>
<td>Mols Bjerge fault, north segment</td>
<td>Mols Bjerge, Traill Ø</td>
<td>36/090</td>
<td>≥2100</td>
<td>J–K, Cz</td>
<td>T</td>
<td>Koch and Haller 1971; field work (this study); 3DPlanes Program</td>
<td>305° dip on south face of Morris Bjerg; 270° dip on northeast face of Morris Bjerg</td>
</tr>
<tr>
<td>7b</td>
<td>Mols Bjerge fault, south segment</td>
<td>Mols Bjerge, Traill Ø</td>
<td>–50/097</td>
<td>≥2100</td>
<td>J–K, Cz</td>
<td>T</td>
<td>Koch and Haller 1971; field work (this study); 3DPlanes Program</td>
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</tr>
<tr>
<td>8a</td>
<td>Vælddal fault, south</td>
<td>Morris Bjerg, Traill Ø</td>
<td>60/305 to 60/270</td>
<td>≥1650</td>
<td>J–K, Cz</td>
<td>T</td>
<td>Koch and Haller 1971; field work (this study); digital photographs</td>
<td>Dip direction may rotate −5°–10° clockwise, northward along fault</td>
</tr>
<tr>
<td>8b</td>
<td>Vælddal fault splay</td>
<td>Vælddal, Traill Ø</td>
<td>60/290</td>
<td>≥300</td>
<td>J–K, Cz</td>
<td>T</td>
<td>Donovan, 1953; fossil localities; 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>8c</td>
<td>Vælddal fault, north</td>
<td>Lycett Bjerg, Traill Ø</td>
<td>60/275 to 55/295</td>
<td>≥550</td>
<td>J–K, Cz</td>
<td>T</td>
<td>Field work (this study); aerial photographs; 3D Planes Program</td>
<td>275° dip on western slopes of Vælddal; 295° dip in south; segmented</td>
</tr>
<tr>
<td>Fault number</td>
<td>Fault name</td>
<td>Location</td>
<td>Fault orientation</td>
<td>Displacement (m vertical throw)</td>
<td>Known activity*</td>
<td>Possible activity*</td>
<td>Data source†</td>
<td>Notes</td>
</tr>
<tr>
<td>--------------</td>
<td>------------</td>
<td>----------</td>
<td>------------------</td>
<td>---------------------------------</td>
<td>----------------</td>
<td>-------------------</td>
<td>---------------</td>
<td>-------</td>
</tr>
<tr>
<td>10</td>
<td>Rubjerg Knude fault</td>
<td>Geographical Society Ø</td>
<td>50/150 to 50/103</td>
<td>≥2000</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971); field work (this study); aerial photographs; LandSat images; 3DPlanes Program</td>
<td>150 dip direction on south Geographical Society Ø; 103 dip direction on north Geographical Society Ø</td>
</tr>
<tr>
<td>11</td>
<td>Bordbjerget north strand fault</td>
<td>Tvaerdal, Geographical Society Ø</td>
<td>50/096</td>
<td>Unknown</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971); 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>Månedal fault</td>
<td>Tvaerdal, Geographical Society Ø</td>
<td>45/116</td>
<td>1200–3100</td>
<td>J–K, Cz</td>
<td>–</td>
<td>Koch and Haller (1971); field work (this study); aerial photographs; fossil localities; 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>Månedal Fault North Splay</td>
<td>Tvaerdal, Geographical Society Ø</td>
<td>50/100</td>
<td>500</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Koch and Haller (1971); field work (this study); aerial photographs; fossil localities; 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>Lysdal fault</td>
<td>Geographical Society Ø</td>
<td>50/100 to 50/080</td>
<td>750–1200</td>
<td>Cz</td>
<td>–</td>
<td>Fossil localities; aerial photographs; field work (this study); 3DPlanes Program</td>
<td>Segmented; 100 dip direction on south segment; 080 dip direction on north segment</td>
</tr>
<tr>
<td>15</td>
<td>Lysdal East Branch fault</td>
<td>Lysdal, Geographical Society Ø</td>
<td>50/090 to 50/080</td>
<td>950</td>
<td>Cz</td>
<td>–</td>
<td>Fossil localities; aerial photographs; 3DPlanes Program</td>
<td>090 dip direction on south Geographical Society Ø; 080 dip direction on north Geographical Society Ø</td>
</tr>
<tr>
<td>16</td>
<td>Langbjerg fault</td>
<td>Geographical Society Ø</td>
<td>50/085</td>
<td>100–250</td>
<td>Cz</td>
<td>–</td>
<td>Aerial photographs; 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>17</td>
<td>Laplace Bjerg fault</td>
<td>Geographical Society Ø</td>
<td>50/090 to 54/103</td>
<td>≥2050</td>
<td>J–K, Cz</td>
<td>T</td>
<td>Koch and Haller (1971); field work (this study); 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>Laplace Bjerg Branch fault</td>
<td>North Geographical Society Ø</td>
<td>58/225</td>
<td>≥450</td>
<td>Cz</td>
<td>–</td>
<td>Field work (this study); aerial photographs; digital photographs; 3DPlane Program</td>
<td></td>
</tr>
<tr>
<td>19</td>
<td>Cambridge Bugt south fault</td>
<td>Freycinet Bjerg, Geographical Society Ø</td>
<td>50/105</td>
<td>≥1200</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Fossil localities; field work (this study); aerial photographs; 3DPlanes Program</td>
<td>Segmented</td>
</tr>
<tr>
<td>20</td>
<td>Cambridge Bugt north fault</td>
<td>Kap Mackenzie, Geographical Society Ø</td>
<td>50/115</td>
<td>≥1200</td>
<td>Cz</td>
<td>T, J–K</td>
<td>Fossil localities; field work (this study); aerial photographs; 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>21</td>
<td>Kap Mackenzie fault</td>
<td>Kap Mackenzie, Geographical Society Ø</td>
<td>45/115</td>
<td>Unknown</td>
<td>Cz</td>
<td>T, J–K</td>
<td>cf. Donovan (1955) and Hald (1996)</td>
<td>Inferred from the juxtaposition of west-dipping Eocene strata to the east of Cretaceous strata</td>
</tr>
<tr>
<td>22a</td>
<td>cross fault 1</td>
<td>Central Vega Sund</td>
<td>Unknown</td>
<td>Unknown</td>
<td>None</td>
<td>T, J–K, Cz</td>
<td>Not applicable</td>
<td>Inferred from geometric constraints</td>
</tr>
<tr>
<td>22b</td>
<td>cross fault 2</td>
<td>Central Vega Sund</td>
<td>Unknown</td>
<td>Unknown</td>
<td>None</td>
<td>T, J–K, Cz</td>
<td>Not applicable</td>
<td>Inferred from geometric constraints</td>
</tr>
<tr>
<td>22c</td>
<td>cross fault 3</td>
<td>Central Vega Sund</td>
<td>Unknown</td>
<td>Unknown</td>
<td>None</td>
<td>T, J–K, Cz</td>
<td>Not applicable</td>
<td>Inferred from geometric constraints</td>
</tr>
<tr>
<td>22d</td>
<td>cross fault 4</td>
<td>East Vega Sund</td>
<td>Unknown</td>
<td>Unknown</td>
<td>None</td>
<td>T, J–K, Cz</td>
<td>Not applicable</td>
<td>Inferred from geometric constraints</td>
</tr>
<tr>
<td>22e</td>
<td>cross fault 5</td>
<td>East Vega Sund</td>
<td>Unknown</td>
<td>Unknown</td>
<td>None</td>
<td>T, J–K, Cz</td>
<td>Not applicable</td>
<td>Inferred from geometric constraints</td>
</tr>
</tbody>
</table>

Note: Fault plane orientation is given as dip and dip direction.

*Age abbreviations: D—Devonian; C—Carboniferous; Cz—Cenozoic; P—Permian; T—Triassic; J—Jurassic; K—Cretaceous.
†3D Planes Program of CASP, University of Cambridge.
### TABLE 2. SUMMARY OF FAULT BLOCKS

<table>
<thead>
<tr>
<th>Fault block number</th>
<th>Fault block name</th>
<th>Location</th>
<th>Observed stratigraphy</th>
<th>Dip and dip direction</th>
<th>Data source*</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Rubjerg Knude block</td>
<td>Traill Ø</td>
<td>Carboniferous</td>
<td>16/305</td>
<td>Field work (this study)</td>
<td>12/340 in the north; 10/341 in the south</td>
</tr>
<tr>
<td>2</td>
<td>Bordbjerget block</td>
<td>Traill Ø</td>
<td>Permian - Triassic</td>
<td>Subhorizontal to 06/317</td>
<td>Field work (this study)</td>
<td>12/340 to 10/341; 10/341 in the south</td>
</tr>
<tr>
<td>3</td>
<td>Månedal block, Traill Ø</td>
<td>Traill Ø</td>
<td>Permian–Triassic</td>
<td>Carboniferous</td>
<td>Field work (this study)</td>
<td>Subhorizontal in the north; 06/317 in the south</td>
</tr>
<tr>
<td>4</td>
<td>Svinhufud Bjerge block</td>
<td>South-central Traill Ø</td>
<td>Cretaceous</td>
<td>Triassic–Jurassic</td>
<td>Field work (this study)</td>
<td>Folded into northwest-plunging syncline; 12/065 on west limb; 15/246 on east limb</td>
</tr>
<tr>
<td>5</td>
<td>Flakkebjerg block</td>
<td>South-central Traill Ø</td>
<td>Cretaceous</td>
<td>07/251 to subhorizontal</td>
<td>Field work (this study)</td>
<td>07/251 in the west; subhorizontal in the east</td>
</tr>
<tr>
<td>6a</td>
<td>Vælddal block</td>
<td>Vælddal-Bjønedal, Traill Ø</td>
<td>Cretaceous</td>
<td>02/115</td>
<td>Field work (this study); 3DPlanes Program; digital photographs</td>
<td>15/100 in Vælddal; 12/093 in Bjønedal; 02/090 ForchammerDal</td>
</tr>
<tr>
<td>6b</td>
<td>Vælddal fault slice 1</td>
<td>Vælddal, Traill Ø</td>
<td>Cretaceous</td>
<td>14/063 to 24/124</td>
<td>Fossil localities; 3DPlanes Program</td>
<td>Appears folded into east-plunging syncline</td>
</tr>
<tr>
<td>6c</td>
<td>Vælddal fault slice 2</td>
<td>Vælddal, Traill Ø</td>
<td>Jurassic</td>
<td>12/093</td>
<td>Field work (this study)</td>
<td>Bedding varies due to internal deformation of fault slice</td>
</tr>
<tr>
<td>7a</td>
<td>Mols Bjerge block a, Skelhøje</td>
<td>Central and north Traill Ø</td>
<td>Cretaceous</td>
<td>10/274 to 12/284</td>
<td>Field work (this study)</td>
<td>10/274 in the eastern Mols Bjerge; 12/284 in Månedal</td>
</tr>
<tr>
<td>7b</td>
<td>Mols Bjerge block b, Skelhøje</td>
<td>Central and north Traill Ø</td>
<td>Cretaceous</td>
<td>14/274</td>
<td>Field work (this study)</td>
<td>10/274 in the eastern Mols Bjerge; 12/284 in Månedal</td>
</tr>
<tr>
<td>7c</td>
<td>Månedal fault slice 2</td>
<td>Månedal, Traill Ø</td>
<td>Cretaceous</td>
<td>08/190 to 03/152</td>
<td>Field work (this study)</td>
<td>Bedding varies across fault slice</td>
</tr>
<tr>
<td>7d</td>
<td>Månedal fault slice 3</td>
<td>Månedal, Traill Ø</td>
<td>Cretaceous</td>
<td>08/190 to 03/152</td>
<td>Field work (this study)</td>
<td>Bedding varies across fault slice</td>
</tr>
<tr>
<td>7e</td>
<td>Månedal fault slice 4</td>
<td>Månedal, Traill Ø</td>
<td>Cretaceous</td>
<td>04/125</td>
<td>Fossil localities; digital photographs; 3DPlanes Program</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>Eilemandbjerg block</td>
<td>Mols Bjerge–Eilemandsjø, Traill Ø</td>
<td>Cretaceous</td>
<td>11/231</td>
<td>Field work (this study)</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>Månedal block, Geographical Society Ø</td>
<td>Central Geographical Society Ø</td>
<td>Cretaceous</td>
<td>11/349 to 12/340</td>
<td>Field work (this study)</td>
<td>11/349 in the north east; 12/340 in the southwest</td>
</tr>
<tr>
<td>10</td>
<td>Lysdal block</td>
<td>Central Geographical Society Ø</td>
<td>Cretaceous</td>
<td>12/010 to 12/280</td>
<td>Fossil localities; field work (this study); 3Dplanes Program</td>
<td>12/010 in the southwest due to footwall uplift on a cross fault in Vega Sund; 12/280 in central Geographical Society Ø</td>
</tr>
</tbody>
</table>

(continued)
The mapped distribution, size, and geometry of Cenozoic dolerite intrusions (Enclosure 1) were obtained from Koch and Haller (1971); subsequent modifications and additions were made based on Landsat imagery. In the Cretaceous mudstones, these intrusions form as much as 40% of the stratigraphic thickness of the succession. However, due to the lack of constraints and for simplicity, many of these intrusions have been excluded from the cross sections because their three-dimensional geometries are not constrained. No attempt has been made to subtract the contribution made by these intrusions to the thickness of the units making up the Cretaceous succession. Consequently, the sediment thicknesses of the Cretaceous units described in this study may be overestimated by as much as 40%. To account for this, a modified structural thickness has also been calculated for the Cretaceous mudstone dominated units, based on the original measured thickness, minus 40% (i.e., minus the maximum thickness contribution of the Cenozoic sills to the measured thickness). In reality, the actual stratigraphic thickness of the Cretaceous units is likely to be somewhere between the measured and modified values.

### REGIONAL GEOLOGY

The post-Caledonian geology of East Greenland comprises a nearly complete succession of Carboniferous to Late Cretaceous strata deposited in continental to deep-marine environments (Donovan, 1957; Koch and Haller, 1971; Surlyk et al., 1973; Clemmensen, 1980; Harpøth et al., 1986; Surllyk, 1990; Kelly et al., 1998; Whitham et al., 1999; Escher, 2001; Seidler et al., 2004). Locally these strata are unconformably overlain by Paleocene–Eocene sediments (Jolley and Whitham, 2004; Nøhr-Hansen et al., 2011; Larsen et al., 2014), which are more regionally overlain by Eocene tholeiitic plateau basalts (Upton et al., 1981; Hald, 1996; Pedersen et al., 1997). The extrusion of tholeiitic plateau basalts was the precursor of the opening of the Norwegian-Greenland Sea in the early Eocene (36–35 Ma, magnetic anomaly C24; Talwani and Eldholm, 1977; Gradstein et al., 2012) and was also associated with the emplacement of basic intrusions (Price et al., 1997; Larsen et al., 2014). Alkaline basalt and syenite intrusions were emplaced in the late Eocene (ca. 36–35 Ma) coincident with separation of the Jan Mayen microcontinent (ca. 33 Ma, magnetic anomaly C13; Talwani and Eldholm, 1977; Larsen et al., 2014), which are more regionally overlain by Eocene tholeiitic plateau basalts (Upton et al., 1981; Hald, 1996; Pedersen et al., 1997). The extrusion of tholeiitic plateau basalts was the precursor of the opening of the Norwegian-Greenland Sea in the early Eocene (ca. 36–35 Ma). The extrusion of tholeiitic plateau basalts was the precursor of the opening of the Norwegian-Greenland Sea in the early Eocene (ca. 36–35 Ma).
The oldest strata in the Traill Ø region are exposed at the western end of Traill Ø and Geographical Society Ø (Fig. 2). These strata belong to the Kap Kolthoff and Celsius Bjerg Groups (Olsen and Larsen, 1993). They are mapped as a single unit in our compilation and have a combined minimum thickness of 2700 m. This unit comprises sandstones and subordinate conglomerates, siltstones, shales, and volcanics. The base of this unit is not observed (Surlyk, 1990; Olsen and Larsen, 1993; Clack and Neininger, 2000; Larsen et al., 2008). Apatite fission track analyses from East Greenland record the initiation of rapid exhumation ca. 20 Ma (Hansen, 1988; Johnson and Gallagher, 2000).

## LITHOSTRATIGRAPHIC FRAMEWORK

The lithostratigraphic framework used to construct the map is outlined in the following. With the exception of the Jurassic, Cretaceous, and Cenozoic stratigraphy, all other stratigraphic units are based entirely on previous work (cited at the end of each paragraph) and are therefore only briefly summarized in the main article. A description of our newly defined stratigraphic framework for the Cretaceous mudstones is given, along with a description of a newly recognized Thanetian unit. For full descriptions of all other units, including details of localities, stratigraphic subdivisions not presented on the map, lithology, depositional environment, fossil content, ages, and notes relating to map construction, refer to Supplemental Materials 4 (see footnote 4). A summary of the lithostratigraphic relationships is shown in Enclosure 3.

### Devonian to Triassic Stratigraphy

The oldest strata in the Traill Ø region are exposed at the western end of Traill Ø and Geographical Society Ø (Fig. 2). These strata belong to the Kap Kolthoff and Celsius Bjerg Groups (Olsen and Larsen, 1993). They are mapped as a single unit in our compilation and have a combined minimum thickness of 2700 m. This unit comprises sandstones and subordinate conglomerates, siltstones, shales, and volcanics. The base of this unit is not observed (Surlyk, 1990; Olsen and Larsen, 1993; Clack and Neininger, 2000; Larsen et al., 2008).

Carboniferous strata exposed in the western parts of Traill Ø and Geographical Society Ø belong to the Traill Ø Group (Vigran et al., 1999). They have a minimum thickness of 3000 m and are mapped as a single unit (Fig. 2). The unit consists largely of sandstones and subordinate mudstones, coals, and conglomerates. The basal contact is reported from northern Geographical Society Ø as conformable with Devonian strata (Christiansen, 1990; Surlyk, 1990; Stemmerik et al., 1991; Marshall et al., 1999; Vigran et al., 1999).

Permian strata belonging to the Foldvik Creek Group (Surlyk et al., 1986) are exposed on central Traill Ø and are mapped as a single unit with a thickness of 90–125 m (Fig. 2). This unit consists of conglomerates, sandstones, mudstones, carbonates, and evaporites, and in places contains a diverse marine macrofauna including brachiopods and bivalves. The basal contact has a 4°–12° discordance with the underlying Carboniferous strata (Surlyk et al., 1986; Christiansen, 1990; Surlyk, 1990; Stemmerik et al., 2001).

Triassic strata are found across central Traill Ø and in Tværdal and around Laplace Bjerg on Geographical Society Ø and are mapped as a single unit, the Scoresby Land Group (Fig. 2) (Clemmensen, 1980). This unit has a thickness of ≥1800 m; this is a minimum estimate because a complete section of the entire group has not been observed. This unit consists of gray-green mudstones and sandstones overlain by locally reddened and gypsiferous mudstones and sandstones. The lower part of the Scoresby Land Group contains marine macrofauna that include ammonoids and bivalves. The basal contact is observed on Traill Ø and is conformable with the underlying Permian strata (Clemmensen, 1980; Christiansen, 1990; Seidler et al., 2004; Bjerager et al., 2006; Andrews et al., 2014; Decou et al., 2016).

### Jurassic Strata: Jameson Land Group and Hall Bredning Group

Pre-Volgian (as used by Kelly et al., 2000, 2015) Jurassic strata in this region are divided from bottom to top into the Bristol Elv, Pelion, and Olypm Formations of the Jameson Land Group and the Bernbjerg Formation of the Hall Bredning Group (Surlyk, 1977; Engkilde and Surlyk, 2003; Therkelsen and Surlyk, 2004).

The Bristol Elv Formation (Therkelsen and Surlyk, 2004) and the Pelion Formation (Engkilde and Surlyk, 2003) are mapped as a single unit (Fig. 2), and are found across central Traill Ø and Geographical Society Ø, most commonly at the crests of tilted fault blocks. These formations have a combined maximum thickness of between 900 m (this study) and 1540 m (Price and Whitham, 1997) and consist of sandstones, mudstones, and subordinate conglomerates, shales, and coals. The Bristol Elv Formation was deposited in a fluvial environment and the Pelion Formation was deposited in a shallow-marine environment. Locally abundant assemblages of marine macrofauna are present in the Pelion Formation (Supplemental Materials 3 [see footnote 3]). The basal contact of this unit is erosive and cuts into the Scoresby Land Group (Donovan, 1953, 1955, 1957; Surlyk, 1973; Engkilde and Surlyk, 1993, 2003; Price and Whitham, 1997; Carr, 1998; Alsen et al., 2004; Therkelsen and Surlyk, 2004; Vosgerau et al., 2004; Alsen, 2015; Callomon et al., 2015; Kelly et al., 2015).

The Olypm Formation (Surlyk et al., 1973) is exposed on south-central and eastern Traill Ø (Fig. 2) and is 90–250 m thick. This formation consists of
coarsening-upward cycles of mudstones and sandstones containing a marine macrofauna (Supplemental Materials 3 [see footnote 3]). The basal contact is conformable with the underlying Pelion Formation. Variations in the thickness of this unit and the Pelion Formation may be related to syndepositional faulting (Surlyk et al., 1973; Surlyk, 1990, 2003; Engkilde and Surlyk, 1993; Price and Whitham, 1997; Larsen and Surlyk, 2003; Vosgerau et al., 2004; Kelly et al., 2015).

The Bernbjerg Formation (Surlyk, 1977) is mapped as a single unit exposed in eastern Traill Ø at the crests of tilted fault blocks (Fig. 2). Exposed sections have a maximum thickness of ~300 m, while cross sections suggest that the Bernbjerg Formation has a minimum thickness of 450 m (Fig. 4; Enclosure 2). This unit consists of black, micaceous, organic-rich shales with subordinate sandstones and contains marine macrofauna throughout (Supplemental Materials 3 [see footnote 3]). The basal contact is conformable with the Olymen Formation (Donovan, 1953, 1955, 1957; Surlyk, 1977; Birkelund and Callomon, 1985; Marcussen et al., 1987; Price and Whitham, 1997; Whitham et al., 1999; Vosgerau et al., 2004). This unit is predicted to be in the subsurface on cross-section A (Fig. 4; Enclosure 2), acting as the source rock for the hydrocarbons found in Jurassic sandstones in the region (cf. Price and Whitham, 1997). In Wollaston Forland (Fig. 1), macrofauna assemblages and palynology suggest that deposition of the Bernbjerg Formation continued during the Volgian and early Ryazanian at the crests of tilted fault blocks (Surlyk, 1977; Pauly et al., 2012, 2013). Ryazanian-aged Bernbjerg Formation is not known in the Traill Ø region and its absence at the crests of tilted fault blocks here is thought to be due to erosion during Volgian–Valanginian rifting. Syn-rift conglomerate deposition in Wollaston Forland commenced in the mid-Volgian (Surlyk, 1978). By analogy with the Wollaston Forland region, the mid-Volgian–Ryazanian parts of the Bernbjerg Formation in the Traill Ø region may thicken down the hanging-wall slopes of tilted fault blocks, grading laterally from mudstones into sandstones and then conglomerates. As a mudstone-dominated unit, the Bernbjerg Formation is likely to contain dolerite sills, as observed in the overlying Cretaceous mudstones.

**Cretaceous Strata (Ryazanian–Campanian): Wollaston Forland and Hold with Hope Groups**

Much of the Traill Ø region is composed of Cretaceous mudstones, within which sandstone and conglomerate units are minor components. We subdivided this thick, lithologically homogeneous unit into seven different Beds (Fig. 7), based on an earlier scheme devised by Donovan (1953, 1957) for subdivision of the Cretaceous mudstones.

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**Figure 7.** New stratigraphic framework for the Ryazanian–Campanian mudstone succession in the Traill Ø region. The dark gray regions in the Beds column show the stratigraphic range of index taxa in each unit.
Donovan (1957) subdivided these strata into four units: the early Albian to early Cenomanian Middle Cretaceous Shale Series, the late Turonian Inoceramus lamarcki Beds, the late Santonian Sphenoceramus Beds, and the late Campanian Scaphites Beds. Nehr-Hansen (1993) and Stemmerik et al. (1993) extended the base of the Middle Cretaceous Shale Series to include mudstones of Barremian and Aptian age and renamed them the Middle Cretaceous Sandy Shale Sequence. Valanginian strata in the Mols Bjerge, containing the bivalve Buchia along with a rich and diverse assemblage of ammonites, were discovered by Donovan (1953), but remained unnamed. Alsøn (2006) worked on these beds and assigned them to the Redryggen and Albrechts Bugt Members of the Palnatokes Bjerg Formation, which belongs to the Wollaston Forland Group (Surlyk, 1978). Immediately north of the Traill Ø region, in Hold with Hope (Fig. 1), Kelly et al. (1998) defined the Home Forland Formation (part of the Hold with Hope Group), a unit that consists of deep-marine mudstones and subordinate sandstones of mid-Albian to late Santonian age. Consequently, based on similarities in lithology, depositional environment, and age, we consider it appropriate to extend the Home Forland Formation to cover most of Ryazanian–Campanian mudstone succession in the Traill Ø region with the exception of strata already assigned to the Palnatokes Bjerg Formation of the Wollaston Forland Group (Surlyk, 1978) (Fig. 7).

While Donovan’s “beds” are by no means proper lithological divisions; given the general absence of lithological variation, the different macrofauna (Figs. 8 and 9) provide the only means by which subdivision and therefore mapping of the monotonous Ryazanian–Campanian succession can be achieved (Enclosure 1). We therefore subdivide this interval into seven mappable units (“Beds”) based on the occurrence of specific macrofauna (Figs. 8 and 9) in a manner similar to Donovan (1953, 1957). We list the beds from oldest to youngest as follows (Fig. 7): (1) Buchia Beds (new); (2) Inoceramus aucella Beds (new); (3) Inoceramus anglicus Beds (new); (4) Inoceramus crippsi Beds (new); (5) Inoceramus lamarcki Beds (Donovan, 1957); (6) Sphenoceramus Beds (Donovan, 1953); and (7) Scaphites Beds (Donovan, 1957). With the exception of the Buchia Beds, which are assigned to the Palnatokes Bjerg Formation (Wollaston Forland Group, Surlyk, 1978), all the other units (2–7) are assigned to the Home Forland Formation (Hold with Hope Group; Kelly et al., 1998). These units are summarized in the following, and described in detail in Supplemental Materials 4 (see footnote 4).

Implementation of these subdivisions reveals the geometry of different units and also provides an insight into Cretaceous sedimentation rates in the Traill Ø region. The Cretaceous mudstones contain up to 40% Cenozoic sills. The thicknesses of the individual Beds given below do not take into account the thickness of intrusions and are therefore overestimates. To acknowledge this, a minimum thickness assuming a 40% thickness attributed to Cenozoic sills is also given. The actual stratigraphic thickness will be between the measured structural thickness and the modified thickness, but the composite thickness of sills is generally greatest in the east and less in the west.

**Buchia Beds**

The Buchia Beds are indicated by the presence of the bivalves Buchia terebratuloides (Lahusen), B. inflata (Lahusen), and B. keyserlingi (Trautschold) (Figs. 8A, 8B) (Zakharov, 1981; Surlyk and Zakharov, 1982) and are poorly exposed on the crests of tilted fault blocks on Traill Ø, where the unit is ≤ 20 m thick (Fig. 2). Extrapolation of mapped boundaries indicates that the Buchia Beds should also be present in surface geology of southern Geographical So-
ciety Ø, although there is no outcrop exposure to confirm this (Fig. 2). In cross sections, the Buchia Beds thicken down the hanging-wall slopes of tilted fault blocks into the hanging-wall succession of the Månedal fault and possibly the Mols Bjerge and Laplace Bjerg faults (cf. Surylk, 1978; Whitham et al., 1999). Subsurface extrapolations of the Buchia Beds have a maximum thickness of 150–600 m (90–360 m where Cenozoic sills may account for as much as 40% of the thickness) (Fig. 4; Enclosure 2).

The Buchia Beds consist of gray (or red at the base of the unit) calcareous mudstones with calcareous concretions and contain a locally abundant and diverse marine macrofauna, including bivalves, ammonites, and belemnites (details in Supplemental Materials 4 [see footnote 4]). The basal contact corresponds to the base-Cretaceous unconformity (BCU), which onlaps the Jurassic Bernbjerg Formation on eastern Traill Ø with an angular unconformity of 4°–8°. This unconformity is erosive and relates to footwall uplift during fault block rotation associated with Volgian–Valanginian rifting (Surylk, 1978). A latest Ryazanian–early Hauterivian age (Fig. 7) is determined for the unit from ammonite and bivalve evidence (full fossil list in Supplemental Materials 3 [see footnote 3]). By analogy with equivalent strata on Wollaston Forland (Fig. 1), as the Buchia Beds thicken down the hanging-wall slopes of tilted fault blocks, they should grade laterally from mudstones into sandstones and then conglomerates in the hanging wall of the block-bounding fault. Syn-rift conglomerate deposition in Wollaston Forland ended in the Valanginian (Surylk, 1978).

Inoceramus aucella Beds

The Inoceramus aucella Beds are indicated by the presence of the bivalve Inoceramus aucella Trautschold (1865) (Fig. 8D) and the absence of Buchia spp. This unit is found across Traill Ø and on southern Geographical Society Ø (Fig. 2). Subsurface extrapolations of the Inoceramus aucella Beds indicate that it has a maximum thickness of 200–400 m (120–240 m where Cenozoic sills may account for as much as 40% of the thickness). Evidence for rifting between Barremian and Aptian time in Hold with Hope (Whitham et al., 1999) suggests that the Inoceramus aucella Beds probably thicken into the hanging walls of faults that were active during Jurassic–Cretaceous rifting (e.g., the Månedal fault and possibly the Mols Bjerge and Laplace Bjerg faults).

The Inoceramus aucella Beds consist of poorly sorted, gray micaceous mudstones with an absence of sedimentary structures. The lower part of this unit commonly contains the index fossil Inoceramus aucella (Fig. 8D) and belemnites and rare ammonites. Barremian–Aptian dinocysts are found within the upper part of this unit (Nøhr-Hansen, 1993). The basal contact is not observed; however, bedding orientations suggest that the Inoceramus aucella Beds are conformable with the underlying Buchia Beds. Cross-section construction indicates that the Inoceramus aucella Beds must also onlap Jurassic strata (the BCU) on the west-dipping hanging-wall slopes of east-dipping normal faults (Fig. 4; Enclosure 2). A mid-Hauterivian–Aptian age is determined from molluscan assemblages (Fig. 7) and palynology (Supplemental Materials 3 [see footnote 3]).
Inoceramus anglicus Beds

The Inoceramus anglicus Beds are indicated by the presence of the bivalve I. anglicus Woods (1911) (Fig. 8C) and associated macrofauna (details in Supplemental Materials 4 [see footnote 4]). This unit is found across eastern Traill Ø and south-central Geographical Society Ø (Fig. 2), and has an estimated maximum thickness of 400–650 m determined from cross-section construction (240–390 m where Cenozoic sills may account for as much as 40% of the thickness) (Fig. 4; Enclosure 2). The Inoceramus anglicus Beds were partly deposited during a period of rifting that ended during the mid-Albian (Whitham et al., 1999), so extrapolations of this unit thicken down the hanging-wall slopes of tilted fault blocks.

The Inoceramus anglicus Beds are dominated by mudstone and strata are largely parallel bedded. The mudstone is interbedded with thin sandstone beds and laminae showing ripple cross-lamination and parallel lamination. Bedding is largely unaffected by bioturbation. Rare meter-scale lenticular sandstone beds fill channels. Lenticular, bedded, clast-supported conglomerates as much as 10.5 m thick are also rare (Donovan, 1955). Large matrix-supported glide blocks of Jurassic and Triassic sandstone are found adjacent to the degraded footwall slopes of some normal faults. These sediments were deposited in a deep-marine setting by a combination of sediment gravity flow processes and suspension settling with subsequent reworking by bottom currents (Whitham et al., 1999). In Tværdal, toward the base of the Inoceramus anglicus Beds, a 40-m-thick sandstone unit is found with coal and rootlet horizons toward its top. Unlike the rest of the unit, this interval was deposited in a continental fluvial setting.

The basal contact of the Inoceramus anglicus Beds has not been seen on hanging-wall slopes; however, bedding orientations suggest that this unit is conformable with the underlying Inoceramus aucella Beds. An angular unconformity is developed at the base of the unit on the submarine degraded footwall slopes of some tilted fault blocks, such as the Mols Bjerge fault block where it onlaps Triassic and Jurassic strata (the BCU). This unit also infills valleys that are incised into these footwall slopes (Figs. 10, 11, and 12A). The onlapping nature of the basal contact means that the Inoceramus anglicus Beds have a complex geometry, thinning onto the crests of tilted fault blocks and thickening into the hanging-wall successions. This indicates that the geometry and distribution of tilted-fault blocks controlled bathymetry in the Traill Ø region at the time (Whitham et al., 1999). Bathymetric relief confined the initial deposition of the Inoceramus anglicus Beds and older Cretaceous strata within rotated fault block intrabasins. The Inoceramus anglicus Beds overfilled some of these basins. An early–late Albian age (Fig. 7) is determined from ammonite and bivalve assemblages (full fossil list in Supplemental Materials 4 [see footnote 4]).

Inoceramus crippsi Beds

The Inoceramus crippsi Beds are indicated by the presence of the bivalve I. crippsi Mantell (1822) (Fig. 9A). This unit is found across large areas of Traill Ø and Geographical Society Ø and has a thickness of 600–1300 m (360–780 m where Cenozoic sills may account for as much as 40% of the thickness) (Fig. 2).
Figure 11. Photograph of the Månedal fault A (see Fig. 10 for location). Jurassic strata are found in the hanging wall of the fault, unconformably overlain by Cretaceous strata belonging to the Inoceramus crippsi Beds. The unconformity is a degraded footwall slope that is onlapped and covered by the Cretaceous mudstones. In places the unconformity must have been subhorizontal, as indicated by the presence of glide blocks of Jurassic strata overlying the unconformity. This exposure demonstrates that the degraded footwall slopes remained exposed on the seafloor until the Cenomanian (see Fig. 2 for location).
The lithology of the unit is similar to the *Inoceramus anglicus* Beds. It is dominated by mudstone, interbedded with thin sandstone beds and laminae showing ripple cross-lamination and parallel lamination. Strata are largely parallel bedded and mostly unaffected by bioturbation. These lithologies were redeposited as slide and slump deposits tens of meters thick in the hanging walls of normal faults and adjacent to submarine degrading footwall slopes, where they are associated with glide blocks of Jurassic and Triassic sandstone. In places this unit contains muddy debris flow deposits containing intraformational lithologies. These sediments were deposited in a deep-marine setting by a combination of sediment gravity flow processes and suspension settling with subsequent reworking in places by bottom currents (Whitham et al., 1999). A marine macrofauna dominated by *Inoceramus crippsi* Mantell (Fig. 9A) and the ammonite *Schloenbachia varians* (J. Sowerby) is observed throughout (details in Supplemental Materials 4 [see footnote 4]). A conformable basal contact with the underlying *Inoceramus anglicus* Beds is observed in hanging-wall successions such as in Tværdal (Fig. 2). An angular unconformity is developed at the
Inoceramus lamarcki Beds

The Inoceramus lamarcki Beds are indicated by the presence of Inoceramus ex gr. lamarcki Parkinson (1821) and associated macrofauna. The unit is found in north-central Traill Ø and across large parts of Geographical Society Ø (Fig. 2). The unit has an estimated minimum thickness of 400–450 m (240–270 m where Cenozoic sills may account for as much as 40% of the thickness). This unit is dominated by mudstone, interbedded with thin sandstone beds and laminae showing ripple cross-lamination and parallel lamination. The basal contact is conformable with the underlying Inoceramus crippsi Beds. A late Cenomanian age (Fig. 7) is indicated from inoceramid bivalves and ammonite assemblages. The late Cenomanian is probably present but has not been formally recognized (full fossil list in Supplemental Materials 3 [see footnote 3]).

Scaphites Beds

The Scaphites Beds are recognized by the presence of ammonites Hoploscaphites kobballi (Birkeland) and H. greenlandicus (Donovan) (Fig. 9B). This unit is found on Geographical Society Ø, to the south and east of Laplace Bjerg and on Traill Ø around Månedal (Fig. 2). The unit has an estimated minimum thickness of 400 m (240 m where Cenozoic sills may account for as much as 40% of the structural thickness).

This unit is dominated by mudstone and strata are largely parallel bedded. The mudstone is interbedded with thin sandstone beds and laminae showing ripple cross-lamination and parallel lamination; bedding is largely unaffected by bioturbation. In places there are rare muddy debris flow deposits containing intraformational lithologies. Rare carbonate methane vent deposits are found in the vicinity of normal faults. Around Leitch Bjerg, this unit contains a lenticular channel infilled by conglomerates and sandstones. The channel has a maximum thickness of 150 m and thins westward to 40 m over a distance of 3 km and forms a mappable unit within the Scaphites Beds, shown on the map as the Scaphites Beds channel fill. These sediments were deposited in a deep-marine setting by a combination of sediment gravity flow processes and suspension settling with subsequent reworking in places by bottom currents (Whitham et al., 1999).

This unit contains a marine macrofauna characterized by the index taxa plus other ammonites, belemnites, gastropods, and echinoids (details in Supplemental Materials 4 [see footnote 4]). The unit is conformable with the underlying Sphenoceramus Beds. A middle to late Campanian age (Fig. 7) is determined from ammonite assemblages (full fossil list in Supplemental Materials 3 [see footnote 3]).

Cenozoic Strata

Sub-Basaltic Marine Beds

The Sub-Basaltic Marine Beds compose a newly recognized unit found close to Leitch Bjerg, Geographical Society Ø (Fig. 2). The unit has a minimum thickness of 65 m. It consists of poorly lithified interbedded mudstones, sandy
mudstones, and sandstones with plant fragments (details in Supplemental Materials 4 [see footnote 4]). The presence of cross-bedded sandstone with mudstone drapes on the foresets suggests a tidal influence on sedimentation, while the presence of plant fragments suggests proximity to the shoreline. The basal contact is an angular unconformity with ≤4° discordance with the underlying Scaphites Beds. A Thanetian age is determined from the presence of the dinocyst Alloacosta margarita (Fig. S1 in Supplemental Materials 4 [see footnote 4]) in association with frequent specimens of the dinocysts Areoligera cf medusettiformis, Oligospheridium cf complex, and Spiniferites ramosus subp ramosus. All palynofloras from this interval are characterized by common to abundant reworked Late Cretaceous dinocysts. These are the oldest known Cenozoic strata in NE Greenland (additional information in Supplemental Materials 4 [see footnote 4]).

Plateau Basalts

Plateau basalts are found on eastern Geographical Society Ø at Kap Mackenzie (Hald, 1996) and near the summit of Leitch Bjerg (Fig. 2). They are as much as 150 m thick and consist of tholeiitic lava flows with rare volcanioclastic layers. At Leitch Bjerg they appear to directly overlie the Sub-Basaltic Marine Beds, but the contact is not seen. Observations in southern East Greenland indicate that faulting occurred prior to and during the initial stages of flood basalt extrusion (Whitham et al., 2004; Larsen and Whitham, 2005). Therefore, there may be an angular discordance at the base of the unit with the Sub-Basaltic Marine Beds. By analogy with other occurrences of plateau basalts in East Greenland, the basalts probably have a latest Paleocene to earliest Eocene age (Jolley and Whitham, 2004; Larsen et al., 2014).

Intrusive Rocks

Doleritic sills and dikes are found throughout the strata of the Traill Ø region, and increase in abundance from west to east. There is also a strong lithological control on dolerite intrusion form and abundance; sandstones typically promote the formation of dikes, whereas mudstone units promote the formation of sills. The thickest sills typically occur above the BCU, where they may be as thick as 300 m. In eastern Geographical Society Ø in areas of Cretaceous mudstone, dolerite sills compose as much as 40% of the structural thickness of Cretaceous mudstones. Most dolerite intrusions have a tholeiitic composition and a minority have an alkaline composition (Price et al., 1997). The tholeiitic dolerites were emplaced ca. 54 Ma and are related to the separation of the Jan Mayen microcontinent (ca. 33 Ma, magnetic anomaly C13; Taiwani and Eldholm, 1977; Mosar et al., 2002) from East Greenland (Brodie, 1995; Price et al., 1997).

The Kap Simpson and Kap Parry syenite plutons form the two eastern promontories of Traill Ø (Fig. 2) (Koch and Hailer, 1971; Price et al., 1997). K/Ar and apatite fission track analyses from the Kap Simpson and Kap Parry syenite plutons give approximate cooling ages of ca. 35 Ma (Noble et al., 1988; Thomson et al., 1999). Emplacement of alkaline acidic rocks is associated with the separation of the Jan Mayen microcontinent from the East Greenland continental margin (Price et al., 1997; Larsen et al., 2014).

STRUCTURE OF THE TRAILL Ø REGION

The Traill Ø region preserves a long and protracted record of sedimentation during multiple rifting events. Consequently, structure and sedimentation are inherently linked and fault and fault block geometries must be considered in order to accurately constrain syn-rift and post-pre-rift sedimentary package geometries and thicknesses. Tables 1 and 2 present a summary of major faults (>100 m vertical throw) and fault blocks identified in the mapped area and include orientation and displacement estimates. Major faults and fault blocks are numbered and labeled in Figures 5 and 6. A full description of each fault and fault block can be found in Supplemental Materials 5 (see footnote 5). A Google Earth compatible drape of the geological map of the Traill Ø region (.kmz file) is available in Supplemental Materials 1 (see footnote 1).

The structure of the Traill Ø region is dominated by east-dipping normal faults that bound fault blocks with west-dipping strata. A minority of westward-dipping normal faults are also identified (e.g., fault 8, Vælddal fault, and fault 18, Laplace Bjerg branch fault; Fig. 5). As shown by Price et al. (1997), most large-scale faults dip between 35° and 55°, while bedding surfaces dip between 5° and 25° (Tables 1 and 2). Assuming that normal faults originally ruptured with a dip of 60° (e.g., Anderson, 1951), these fault plane and bedding orientations (ignoring syn-rift packages) are consistent with a fault block rotation of ~5°–25°.

Vertical throw is measured from cross sections (Fig. 4; Enclosure 2) between stratigraphic boundaries of the same age, present in both the footwall and hanging wall of a single fault (typically top of the Carboniferous). As the stratigraphic record extrapolated above and below the surface on cross sections is sometimes incomplete, vertical throws are often minimum estimates. Vertical throw typically ranges between 500 and 1500 m (Table 1). The largest vertical throw of 5500 m is recorded on the Måndal fault zone (Fig. 5, faults 5a–5d), which extends from Geographical Society Ø to Traill Ø (Enclosures 1 and 2). Faults with larger displacements typically have shallower dips, reflecting greater amounts of fault block rotation (Table 2). Fault block widths increase southward; on Geographical Society Ø (Fig. 6) they are typically 4–16 km wide, whereas on Traill Ø, they are 4–33 km wide. A small number of narrower fault slices, 0.5–2 km wide are identified on both islands (Fig. 6). The general absence of closely spaced faults with small displacements on the map may be due to poor exposure in areas of Ryazanian–Campanian outcrops. On
the south coast of Traill Ø in the coastal cliffs of southern Svinhufvud Bjerge (Fig. 10), where exposure is excellent, a large number of normal faults are seen.

Previous studies of rifted continental margins suggest that extension may be accommodated by slip of fault blocks along a low-angle detachment into which fault block–bounding normal faults sole (e.g., Lister et al., 1986; Waltham et al., 1993) or by domino-style rotation of fault blocks above a middle to lower crustal shear zone (e.g., Jackson and McKenzie, 1983; Barr, 1987). Seismic reflection surveys in east Greenland identify a low-angle crustal interface that appears to extend from the Gauss Halvø fault (GHF; Fig. 1) eastward to a depth of ~6 km beneath the Mesozoic fault block basins of the Traill Ø region (Voss and Jokat, 2007, 2009; Hermann and Jokat, 2016). The nature of this interface is unknown and may correspond to a low-angle detachment horizon, a discrete ductile shear zone, or a diffuse zone of homogeneous stretching (e.g., Jackson and McKenzie, 1983; Peacock et al., 2000). Low β values (1.05–1.22) calculated for “pre-magmatic” and “post-magmatic” extension in the Traill Ø region (Price et al., 1997) and the dissection of rotated fault blocks by younger, steeper faults are comparable to domino-style fault block rotation in a pure shear rift setting (e.g., McKenzie, 1978; Jackson and McKenzie, 1983).

It is not possible to trace unit boundaries across Vega Sund from the north coast of Traill Ø to the north coast of Geographical Society Ø because the mapped position and area of stratigraphic units differ between shorelines, suggesting that a structural discontinuity must separate the two islands. To account for this discordance, a subsidiary set of northwest-southeast- and east-west–striking cross faults are postulated in the subsurface geology beneath Vega Sund (faults 22a–22e, Fig. 5). The exact positions of these faults and their geometries are poorly constrained and it is possible that some of these structures may also accommodate east-west strike-slip motion as well as north-south normal sense motion. However, evidence of strike-slip faulting has not been observed elsewhere in the Traill Ø region. Equivalent structural discontinuities have been proposed along Vega Sund and along Kong Oscar Fjord, south of Traill Ø and Kejser Franz Joseph Fjord, north of Geographical Society Ø (Surylk et al., 1973; Surylk, 1977, 1990; Schindwein and Jokat, 1999). It is beyond the scope of this study to assess the true nature of the structures beneath Vega Sund; however, the variation in outcrop pattern between the two islands necessitates their existence in one form or another.

## DISCUSSION

The map and cross sections (Enclosures 1 and 2) provide new insights into the geology of the area and the structural and sedimentological evolution of the wider region. This work provides a better understanding of the nature and significance of different rift events and sedimentation rates across NE Greenland during the Cretaceous; it also presents an opportunity to consider the lithospheric evolution of the East Greenland continental margin and assess the applicability of different models of lithospheric extension to the region.

### Structural and Sedimentary Evolution of the Traill Ø Region

#### Fault Activity and Rifting Events

Rifting leads to fault block rotation, erosion of footwall crests, and deposition of wedge-shaped sediment packages in the hanging walls of normal faults (e.g., Gawthorpe and Leeder, 2000). At fault block crests, post-rift thermal subsidence leads to the formation of an angular discordance between pre-rift and post-rift strata (e.g., Price and Whitham, 1997). Consideration of crosscutting relationships between normal faults and stratigraphic units and the spatial variations in stratigraphic geometries, angular discordances, and fault block rotations provides indications of when each fault was active and reveals the occurrence of multiple phases of rifting in the Traill Ø region, each with an approximately east-west extensional direction (Fig. 13).

Previous studies indicate that the earliest faulting is recorded on the GHF (faults 1 and 9; Peacock et al., 2000; also known as the Post Devonian Main fault; Vischer, 1943) during the Carboniferous and possibly during the Devonian (Supplemental Materials 5 [see footnote 5]) as indicated by spore-pollen assemblages in syn-rift hanging-wall strata (Fig. 13A) (Bütler, 1955; Stemmerik et al., 1991). Faulting may have also occurred during the mid-Permian, as indicated by an angular discordance of 4°–12° between the base of the Permian Foldvik Creek Group and underlying strata (Surylk et al., 1986), and during the Early Triassic as suggested by Seidler (2000) and Seidler et al. (2004), who interpreted the Early Triassic strata as syn-rift sediments (Fig. 13B; Supplemental Materials 5 [footnote 5]).

Faulting and fault block rotation during Volgian–mid-Albian rifting were documented previously (Surylk, 1978; Price and Whitham, 1997; Price et al., 1997; Whitham et al., 1999). In the Traill Ø region, our record of ~5°–8° angular discordance between the Cretaceous and Jurassic strata (Fig. 4; Enclosure 2; Supplemental Materials 5 [footnote 5]) is in agreement with previous studies from across the region (Surylk et al., 1973; Surylk, 1978, 1987; Price and Whitham, 1997; Price et al., 1997). Volgian–mid-Albian syn-rift strata observed on fault block crests and inferred in subsurface hanging-wall successions from cross-section construction (Fig. 4; Enclosure 2) indicate that movement occurred on the Månedal (faults 5 and 12), Mols Bjerge (fault 7), and Laplace Bjerg (fault 17) faults during this time (Fig. 13C; Supplemental Materials 5 [see footnote 5]). The fault block width between these faults was ~10–30 km. A decline in faulting in the Albian is indicated by observations of Albian–Cenomanian strata onlapping and covering degraded footwall slopes of the Månedal, Mols Bjerge, Laplace Bjerg, and Vælddal faults (Fig. 4; Enclosure 2).

Cenozoic faulting in the Traill Ø region was documented by Price et al. (1997), who recorded “post-magmatic” faulting which occurred after the emplacement of tholeiitic magmatic intrusions at ca. 54 Ma and before and after the emplacement of alkaline magmatic intrusions at ca. 36–35 Ma. Our observations are consistent with those of Price et al. (1997). On Geographic Society Ø, the steepest faults (~50°–60° dip), which are bound at the surface by Cretaceous strata with a dip of 10° in both the footwall and hanging wall, are
Figure 13. (A–D) Map showing the activity of faults in the region for different time periods (see Fig. 5 and Table 1 for fault numbers). The information used to construct these diagrams is summarized in Tables 1 and 2 and presented in detail in Supplemental Materials 5 (see footnote 5). Red indicates a fault known to be active; light gray tone indicates a fault that may have been active with no known previous activity; dark gray tone indicates a fault that may have been active and is known to have been active in previous rift events. Selected faults: GHF—Gauss Halvø fault; LBF—Laplace Bjerg fault; MF—Månedal fault; MBF—Mols Bjerge fault; VF—Vælddal fault.
thought to have been created during post-Cretaceous rifting, reducing fault block widths from ~10–30 km to ~4–10 km (Fig. 13D; Supplemental Materials 5 [see footnote 5]). These faults are the Månedal north splay (fault 13), Lysdal (fault 14), Lysdal east branch (fault 15), Langbjerg (fault 16), and Laplace Bjerg branch (fault 18) faults (Table 1; Fig. 13D; Supplemental Materials 5 [see footnote 5]). On Traill Ø, the Månedal splay faults (faults 5b, 5c, 5d) and Skelhøje fault (fault 6) were activated during this interval. In addition, observed displacements of post-Cretaceous igneous intrusions indicate that most, if not all, preexisting faults were reactivated during post-Cretaceous rifting.

Evidence for a previously undocumented phase of faulting in NE Greenland is seen at Leitch Bjerg, on Geographical Society Ø (Fig. 2). At this location an angular unconformity of ≤4° observed between the Sub-Basaltic Marine Beds and the *Scaphites* Beds indicates that rifting occurred in this region sometime between the late Campanian and Thanetian intervals. A rift event during this time interval is not unexpected; it is also documented from Kangerlussaq in southern East Greenland (Whitham et al., 2004; Larsen and Whitham, 2005) and in central Norway (Brekke, 2000).

Our findings support those of Price et al. (1997) and indicate that most faulting and fault block rotation in the Traill Ø region were the result of post-Cretaceous rifting. The total post-Cretaceous tilt of strata is typically ~10°, but locally as much as ~25°, as indicated by early Eocene lavas at Kap Mackenzie that dip 15°–25° toward the west (Fig. 2). The relatively minor angular discordance of ≤4° between Thanetian and Cretaceous strata suggests that most post-Cretaceous fault block rotation and extension occurred during “post-magmatic” faulting. This lends further weight to the hypothesis that the most significant phases of normal faulting and fault block rotation in the Cenozoic were “post-magmatic,” occurring between ca. 54 and 36 Ma and after ca. 36 Ma (cf. Price et al., 1997). The latter stage of post~36 Ma faulting occurred after the initiation of seafloor spreading in the Norwegian-Greenland Sea (Price et al., 1997; Olesen et al., 2007; Larsen et al., 2014) and has been ascribed to the separation of the Jan Mayen microcontinent from the East Greenland margin (Price et al., 1997; Scott et al., 2005; Larsen et al., 2014), which occurred ca. 33 Ma (magnetic anomaly 1C13; Talwani and Eldholm, 1977; Mosar et al., 2002). We note that as we have been unable to account for Cenozoic sills during cross-section construction, it is not possible to determine the amount of fault displacement that occurred during each rift phase. However, our interpretations are based on fault and bedding dips and fault block widths, which are not affected by sill thickness and therefore remain valid, despite not knowing the precise contribution of Cenozoic sills to structural thickness.

**Jurassic–Cretaceous Syn-Rift Sediment Thickness**

Volgian to mid-Ryazanian sediments have not been found in the Traill Ø region; however, comparison with the geology of Wollaston Forland (Fig. 1) would suggest that strata of this age are present in the subsurface hanging-wall successions of faults active during Volgian–Valanginian rifting (Price and Whitham, 1997). The absence of any strata of this age at the footwall crests of tilted faults is due to the fact that at the present-day level of exposure only the degraded footwall slopes of these features are exposed. The outcrops of the *Buchia* Beds at the crests of tilted fault blocks should have condensed sections of Volgian–Ryazanian strata sediments at their base, but so far these have not been detected based on the biotas collected.

Jurassic–Cretaceous rifting in NE Greenland continued until the mid-Albian (Whitham et al., 1999). Incorporating this information into cross-section constructions indicates that the subsurface hanging-wall succession of the Månedal fault contains a wedge of Volgian–late Albian syn-rift sediments comprising the *Buchia, Inoceramus aucella*, and *Inoceramus anglicus* beds. In addition, this wedge may or may not include early Volgian pre-rift sediments from the top of the Bernbjerg Formation.

Based on our cross sections, the subsurface Volgian–mid-Albian syn-rift wedge of the Månedal fault hanging-wall succession has a maximum thickness of 1050–1500 m (cross-sections A–D, Enclosure 2; Fig. 4). This is comparable to previous estimates of hanging-wall subsidence for the Mols Bjerge fault block (Fig. 6) of 1000–2500 m based on amounts of footwall crest erosion (Price and Whitham, 1997). Within this wedge of sediments, strata that are time equivalent to the uppermost Bernbjerg Formation and the *Buchia* Beds do not exceed a maximum thickness of 600 m. Equivalent Volgian–late Albian syn-rift wedges are also expected in the hanging-wall successions of the Mols Bjerge and Laplace Bjerg faults, although mapped outcrop patterns do not allow for extrapolation of these units in cross section. Given that these sediments are dominantly conglomerates and sandstones, they are unlikely to contain dolerite sills and so a modified thickness is not required for these strata.

We note that the estimated maximum thickness of 600 m of Volgian–Valanginian strata in the Traill Ø region is significantly thinner than the estimated thickness of ~2600 m of Volgian–Valanginian syn-rift strata recorded in Wollaston Forland (Fig. 1), as determined by Surlyk (1978).

**Late Cretaceous Post-Rift Sedimentation Rates**

The post-rift strata in the Home Forland Formation, belonging to the *Inoceramus crippsi*, *Inoceramus lamarcki*, *Sphenoceramus*, and *Scaphites* beds (Cenomanian–Campanian; Fig. 7), have a substantial thickness (~2500 m, including Cenozoic sills). It has been suggested that these units, dominated by deep-marine mudstones, were deposited during protracted thermal subsidence following the mid-Volgian–mid-Albian rift event (Whitham et al., 1999). The apparent rate of deposition of the *Inoceramus crippsi* Beds appears to be high, as much as 1300 m thickness of mudstones deposited over 6 m.y. Similar sedimentation rates are recorded by the Cenomanian strata on Hold with Hope (Whitham et al., 1999). A relative decrease in apparent sedimentation rate is recorded by the remaining units of the Home Forland Formation that overlie the *Inoceramus crippsi* Beds with a maximum thickness of 1100–1200 m for the next 17 m.y. A reduction in sedimentation rates is also seen in Hold with
Lithospheric Evolution of the East Greenland Continental Margin

Here we link the spatial and temporal distribution of faulting and sedimentation between the East Greenland Caledonides and the Traill Ø region to the lithospheric evolution of the East Greenland continental margin as previously determined from geophysical analyses, based on interpretations made by Schlindwein and Jokat (1999, 2000), Schmidt-Aursch and Jokat (2005), and Voss and Jokat (2007, 2009) from the KKF-3D-density cross section (Schmidt-Aursch and Jokat, 2005) and seismic reflection profiles 94320 (Schlindwein and Jokat, 1999) and AWI-20030050 (Voss and Jokat, 2007) (see Fig. 1 for locations of geophysical survey lines). This is followed by consideration of the spatial and temporal distribution of strain during rifting and the applicability of different models of lithospheric extension to the evolution of the East Greenland continental margin and the Norwegian-Greenland seafloor spreading center (see following discussion of Rift Evolution Models for East Greenland and Norwegian-Greenland Sea).

Rift Phase 1: Devonian–Triassic Post-Orogenic Collapse and Continental Rifting

The Caledonian orogeny of East Greenland culminated in the Silurian (Torsvik et al., 1996) and produced a crustal root with a present-day maximum depth of 49 km (Fig. 14A) (Schmidt-Aursch and Jokat, 2005). Orogenic collapse initiated in the Devonian, affecting supracrustal rocks of the Eleonore Bay Supergroup found in the inner fjord region, west of Traill Ø (Suryk, 1990; Bengaard, 1991). Faulting was restricted to a region bound to the west by the Fjord region detachment (FRD) (Hartz and Andresen, 1995; Voss and Jokat, 2009). Continued crustal thinning driven by orogenic collapse during the Devonian subsequently led to the formation of the Western fault zone (WFZ), east of the FRD (Larsen and Bengaard, 1991; Voss and Jokat, 2009). Upper crustal extension was accompanied by crustal thinning, resulting in a Moho depth of ~30 km beneath the WFZ (Fig. 14A) (Schlindwein and Jokat, 2000).

Between the Late Devonian and Carboniferous, extensional faulting and syn-rift deposition migrated eastward of the FRD, leaving the Eleonore Bay Supergroup close to its present-day configuration between the FRD and WFZ (Hartz and Andresen, 1995; Voss and Jokat, 2009). At that time, extensional deformation concentrated on the WFZ and GGH (Fig. 13B), and produced ~80–100-km-wide fault blocks (Bütler, 1955; Larsen and Bengaard, 1991; Stemberik et al., 1991; Price et al., 1997). In the subsurface, this eastward migration of deformation preserved an upwarped Moho underlying the Eleonore Supergroup at 30 km depth (Fig. 14B) (Hartz and Andresen, 1995; Voss and Jokat, 2009). Extensional faulting in the Traill Ø region during this time is recorded on the GGH (Fig. 13A) (Bütler, 1955; Stemberik et al., 1991) and may have also occurred during Permian–Triassic time (Fig. 13B) (Suryk et al., 1986; Seidler, 2000).

Rift Phase 2: Jurassic–Cretaceous Continental Rifting

A second eastward migration of concentrated extensional deformation and subsidence occurred during Jurassic–Cretaceous rifting (Voss and Jokat, 2007) and is supported by the absence of surface exposures of Devonian sediments mapped east of the GHF (e.g., Koch and Haller, 1971; Escher and Pulvertaft, 1990). To the north, within the Traill Ø region, Jurassic to Cretaceous deposition of clastic rocks and mudstone-bearing black shales and coal beds occurred (Figs. 16–19). This was associated with an increase in sedimentation rates as the hanging walls of active faults were subaerially exposed (Fig. 19). The resulting sedimentary basins, which onlap the northern part of the Eleonore Bay and Eleonore Fjord region (Fig. 13B), are unconformably overlain by Jurassic–Cretaceous black shales and coal beds. The deposition of these black shales and coal beds is correlated with the development of a post-rift thermal subsidence event (Fig. 19). The post-rift thermal subsidence event is thought to have been caused by the cooling of the post-rift crust, which resulted in the formation of a regional flexure (Fig. 13B). The flexure is thought to have had a significant impact on the distribution of sedimentary basins in the region, as it allowed for the deposition of thick sequences of black shales and coal beds.
Figure 14. Schematic model of crustal evolution for the East Greenland continental margin. (A) Devonian orogenic collapse of the Caledonian orogen. Extensional deformation concentrated east of the Fjord region detachment deforming the supracrustal Eleonore Bay Supergroup. (B) Late Devonian–Triassic rift interval (Rift Phase 1). Extensional deformation was concentrated in the crust east of the Western fault zone. (C) Jurassic–Cretaceous rift interval (Rift Phase 2). Extensional deformation was concentrated in the crust east of the Gauss Halvø fault. (D) Paleocene–Oligocene rift interval (Rift Phase 3). Extensional deformation was concentrated in the crust east of Månedal fault overlying the continent-ocean transition. Crustal structure drawn from Schlindwein and Jokat (2000, Fig. 4 therein), Voss and Jokat (2007, Fig. 10 therein), and Voss and Jokat (2009, Figs. 2 and 5 therein). Cited figures are based on interpretations of the KF JF–3D-density cross section of Schmidt-Aursch and Jokat (2005), and seismic reflection profiles 94320 (Schlindwein and Jokat, 1999) and AWI-20030050 (Voss and Jokat, 2007). Locations of these geophysical survey sections are presented in Figure 1.
The majority of faulting at the time was concentrated in basins east of the GHF, preserving the Devonian basins in the footwall of the GHF to the west (Fig. 14C). In the Traill Ø region, major faulting and syn-rift deposition occurred in the hanging walls of the Gauss Halvø, Månedal, Mols Bjerge, and Laplace Bjerg faults during this time, dissecting fault blocks into smaller widths of ~10–30 km (Fig. 13C). Active faulting in this region ended during Albian–Cenomanian time, and was followed by a phase of post-rift thermal subsidence during the Late Cretaceous (Whitham et al., 1999). This eastward migration of extensional deformation was mirrored by crustal thinning at lower crustal depth, resulting in a second west-dipping Moho slope beneath the GHF, rising to a depth of ~25 km below the Jurassic–Cretaceous basins (Fig. 14C) (Schlindwein and Jokat, 1999; Schmidt-Aursch and Jokat, 2005).

**Rift Phase 3: Post-Campanian Continental Rifting and Post–50 Ma Seafloor Spreading**

Cenozoic rifting can be divided into at least three episodes of faulting. The first, based on the discordance between the Cretaceous Hold with Hope Group and the Thantetian Sub-Basaltic Marine Beds, is a newly recognized episode of “pre-magmatic” faulting that occurred between the late Campanian and Thantetian. This was followed by “post-magmatic” faulting between ca. 54 and 36 Ma, and after ca. 36 Ma (Price et al., 1999).

Pre-Cenozoic stepwise migration of extensional deformation across the WFZ and GHF, during Rift Phases 1 and 2, respectively, was determined from the surface geology (e.g., Surlyk, 1990; Price et al., 1997; Schlindwein and Jokat, 1999). We recognize a third migration of the zone of major extensional deformation and subsidence in the Traill Ø region, concentrated in the crust east of the Månedal fault during Rift Phase 3 (Figs. 13D and 14D).

Faults east of the Månedal fault accommodated a significant amount of post-Cretaceous extension, as indicated by the juxtaposition of Late Cretaceous hanging-wall strata against Carboniferous footwall strata and maximum fault block rotations of 10°–25° (Fig. 4; Enclosure 2). In contrast, fault blocks west of the Månedal fault show only minor amounts of fault block rotation (<10°) and in some locations, Permian and Triassic strata on Traill Ø west of the fault are subhorizontal. During this time, fault blocks were dissected to widths of ~4–10 km. It is noted that mapping of the Svinhufvud Bjerge cliffs (Fig. 10) indicates that appreciable amounts of Cenozoic faulting did occur west of the Månedal fault. Consequently, the proposed post-Cretaceous migration of extension east of the Månedal fault was not a complete migration. Nevertheless, differences in fault block rotation and fault displacement support the hypothesis that Cenozoic extension and subsidence was greater to the east of the Månedal fault relative to the region to the west.

North of the Traill Ø region, Cretaceous hanging-wall strata are juxtaposed against Carboniferous Permian–Triassic footwall strata along a line of faults east of the GHF on Hold with Hope and Wollaston Forland (Fig. 1). This outcrop pattern may correspond to the northward continuation of the zone of major Cenozoic faulting and subsidence as defined by the rotated strata east of the Månedal fault in the Traill Ø region.

Through consideration of the subsurface structure we suggest that the third eastward migration of extensional deformation across the Månedal fault during the Cenozoic may correspond to the contemporary development of a continent-ocean transition zone prior to the initiation of continental break-up (Voss and Jokat, 2007, 2009). Magmatic underplating beneath the Traill Ø region has been inferred from the coincidental positions of a high-velocity lower crustal body and a strong negative magnetic anomaly identified beneath the Mesozoic basins east of the GHF at 16–30 km depth (Fig. 14D) (Schlindwein and Jokat, 1999; Voss and Jokat, 2009; Hermann and Jokat, 2016). Previous studies suggest that the inferred lower crustal magmatic underplating and emplacement of middle and upper crustal magmatic intrusions occurred contemporaneously with the emplacement of theleitic intrusions found across the Traill Ø region ca. 56–54 Ma, and correspond to the development of a continent-ocean transition prior to the initiation of continental break-up and ocean spreading along the Aegir Ridge ca. 49.9 Ma (Fig. 14D) (Price et al., 1997; Voss and Jokat, 2007; Gaina et al., 2009; Larsen et al., 2014). This was accompanied by faulting and syn-rift sedimentation in the overlying continental margin crust that continued at least to the early Oligocene during ocean spreading along the Kolbeinsrey Ridge and separation of the Jan Mayen microcontinent (ca. 33 Ma, magnetic anomaly C13; Taiwani and Eldholm, 1977; Mosar et al., 2002) from the East Greenland margin (Price et al., 1997; Scott et al., 2005; Gaina et al., 2009; Voss and Jokat, 2009).

**Rift Evolution Models for East Greenland and the Norwegian-Greenland Sea**

Classic models of continental rifting commonly describe a migration and localization of deformation toward the rift axis with time (e.g., Gupta et al., 1998; Gawthorpe and Leeder, 2000; Cowie et al., 2005). These models, which have been applied to the East African Rift system (e.g., Hayward and Ebinger, 1996; Bilham et al., 1999; Ebinger and Casey, 2001), the Gulf of Suez (Gupta et al., 1998), the North Sea (Gupta et al., 1998; Cowie et al., 2005), and other rift systems, also define a progression from many small densely spaced active faults with small fault displacements to fewer larger widely spaced active faults with large fault displacements. By contrast, fault activity in the Traill Ø region (Fig. 13) indicates that with each new rift phase, the number of active faults increased and the spacing between them decreased, thus suggesting a delocalization of strain over time. However, with each new rift phase we and others (e.g., Price et al., 1997; Schlindwein and Jokat, 1999; Voss and Jokat, 2007, 2009) document a stepwise eastward migration of the area of concentrated extensional deformation, thus suggesting a preferential localization of strain eastward toward the rift axis with time. These observations are not consistent with the aforementioned rift evolution models, and further consideration of strain localizing and strain delocalizing processes active during rifting is required.
The temporal evolution of strain distribution within continental rift systems is governed by the rheological state of the lithosphere prior to and during rifting (e.g., Bassi, 1995; Lavie et al., 2000; Wijns et al., 2005; Yamazaki et al., 2006; Yamazaki and Gernigon, 2009; Chenin and Beaumont, 2013). This reflects a balance between competing strain hardening and strain weakening mechanisms, the controls of which include pore fluid pressure (e.g., Sibson, 1990), frictional heating (e.g., Montési and Zuber, 2002), mineral transformations (e.g., phylloïdification; Wallis et al., 2015), strain rate, geothermal gradient, accretion of material during mantle upwelling and/or magmatic underplating (e.g., England, 1983; Kuszénr and Park, 1987; Yamazaki and Gernigon, 2009), fault geometry and cohesion (e.g., Jackson and McKenzie, 1983; Jackson and White, 1989; Forsyth, 1992; Lavie and Manatschal, 2006), and flexural loading (e.g., Jackson and White, 1989; Petit and Ebinger, 2000). The pre-rift structure of the lithosphere places an important control on rifting, as lateral and vertical variations in lithospheric composition and thickness result in a heterogeneous rheology that preferentially localizes deformation in rheologically weak zones (e.g., Gilbert and Sheehan, 2004; Louie et al., 2004; Yamazaki et al., 2006; Yamazaki and Gernigon, 2009; Chenin and Beaumont, 2013). In addition, preexisting faults and shear zones can form weaknesses in the lithosphere that preferentially localize deformation if such structures are favorably oriented relative to the principal stress directions (e.g., Muir et al., 2000; Reeeve et al., 2013). In essence, strain localization on a given fault or shear zone during rifting requires the structure in question to remain weaker than its surrounding lithosphere and as weak as or weaker than other active faults (Jackson and McKenzie, 1983; Jackson and White, 1989; Forsyth, 1992). If at any point, the strength of these structures exceeds the strength of the surrounding lithosphere, then the lithosphere will mechanically fail and a new fault will rupture, thus promoting strain delocalization (Jackson and McKenzie, 1983; Jackson and White, 1989; Forsyth, 1992). This requirement remains true for the reactivation of preexisting structures (e.g., Reeve et al., 2013; Bell et al., 2014).

The nature of faulting and basin architecture in the Traill Ø region is comparable to a domino-style model of fault block rotation during rifting (e.g., Jackson and McKenzie, 1983). Fault block rotation during extension leads to a reduction in fault dip, which effectively strengthens a fault (increased cohesion) as the orientation of vertical stress (i.e., gravity) relative to the fault plane rotates toward the fault plane normal (Jackson and McKenzie, 1983; Jackson and White, 1989). Consequently, as fault rotation continues, it eventually becomes more efficient to rupture a new steep fault than to continue faulting on an older low-angle fault, thus leading to a dissection of the fault block (Jackson and McKenzie, 1983). As fault blocks rotate, a saw-toothed topographic relief will develop that exerts a gravitational shear stress to the upper crust (Jackson and White, 1989). As the amplitude of the saw-tooth topography increases, so does the gravitational shear stress (Jackson and White, 1989). If this shear stress exceeds the stress drops that occur during slip along the fault-block bounding faults, then the fault block will rupture internally, dissecting into two or more smaller fault blocks (Jackson and White, 1989). Strong crust with a larger elastic thickness can sustain larger gravitational shear stresses without rupturing, and thus support wider fault blocks with greater topographic relief (Jackson and White, 1989). Changes in elastic thickness during extensional fault block rotation can lead to changes in fault block width (Jackson and White, 1989). A dissection of fault blocks similar to that observed in the Traill Ø region is also reported from Hold with Hope, from pre-Cenozoic widths of ~25 km to post-Cretaceous widths of ~5 km (Fig. 1) (Price et al., 1997; Peacock et al., 2000). It is likely that these strain delocalization processes associated with domino-style rifting played an important role in the spatial and temporal distribution of strain during the rifting of the East Greenland continental margin.

It is also important to make a distinction between models of progressive rifting and polyphase rifting when considering the temporal and spatial distribution of strain during rifting (e.g., Bell et al., 2014). Progressive rifting describes a single, continuous, and perhaps protracted episode of rifting that may develop into a seafloor spreading center or failed rift. Polyphase rifting describes the occurrence of multiple distinct phases of rifting, each separated by a period of tectonic quiescence prior to the renewal of extension in a subsequent rift phase. Most models of strain localization during rifting typically describe a progressive rifting evolution, during which continuous extension promotes growth and linkage of fault segments. This leads to a rift climax stage characterized by a concentration of deformation on fewer but larger active fault segments, surrounded by smaller, less active or inactive fault segments (e.g., Gupta et al., 1998; Gawthorpe and Leeder, 2000; Cowie et al., 2005). In the case of polyphase rifting, the period of time between each rift phase may be sufficient for the lithosphere to modify through post-rift thermal relaxation and subsidence and addition of material to the base of the crust through magmatic underplating and accretion (e.g., England, 1983; Buck, 1991; Lavie and Manatschal, 2006; Yamazaki et al., 2006; Yamazaki and Gernigon, 2009; Bell et al., 2014). These processes will modify the state of the lithosphere such that rheological heterogeneities present during the previous rift phase will differ from those present at the onset of a subsequent rift phase. Consequently, the distribution of deformation at the onset of a subsequent rift phase may not simply reactivate preexisting faults that were active at the end of the previous rift phase, and may instead require the activation of new faults.

The rift evolution of the East Greenland continental margin constitutes that of a polyphase evolution with at least three distinct phases of rifting during Devonian–Triassic (Rift Phase 1), Jurassic–Cretaceous (Rift Phase 2), and Cenozoic (Rift Phase 3) intervals. With each rift phase, the increase in the number of active faults can be attributed to the dissection of larger fault blocks into smaller fault blocks (e.g., Jackson and McKenzie, 1983). With each distinct rift phase and proceeding post-rift phase, modification to the structure of the lithosphere may have preferentially localized deformation toward the east in the subsequent rift phase. This stepwise migration of deformation may be attributed to evolving lateral heterogeneities in lithospheric thickness and structure (and thus rheology) that initially developed during the Caledonian orogeny and were subsequently modified during proceeding rift phases (Fig. 14) (e.g., Schindwein and Jokat, 1999, 2000; Schmidt-Aursch and Jokat, 2005; Voss and Jokat, 2007, 2009). A preferential reduction in elastic thickness of the
crust closest to the rift axis may have been responsible for the eastward migration of the area of major extensional deformation (and fault block dissection) in the subsequent rift phases (cf. Jackson and White, 1989). It is also likely that multiple phases of post-rift thermal relaxation and subsidence with or without magmatic accretion and/or underplating altered the strength of the crust sufficiently to influence the delocalization of faulting and the eastward migration of deformation. A similar polyphase rift evolution, influenced by post-rift lithospheric modification between two distinct rift phases, was described by Bell et al. (2014) for the northern North Sea.

Our proposed model suggests that strain hardening mechanisms may have outpaced strain weakening mechanisms during Rift Phases 1 and 2, such that these rift phases were unable to mature to seafloor spreading centers. This is probably due, at least in part, to the initial thickness of the Caledonian orogenic root, which effectively strengthened the lithosphere and hindered the localization of deformation during extension (e.g., Sonder and Jones, 1999; Lavér et al., 2000). The lithospheric structure of the Lofoten-Vesterålen continental margin in northeast Norway (Tsikalas et al., 2005) approximately represents the conjugate rift margin to East Greenland prior to seafloor spreading (Olesen et al., 2007), and is remarkably similar to that of East Greenland (Voss and Jokat, 2007, 2009). This indicates that the rift phases in East Greenland reflected that of a pure shear rift model and/or a wide rift model in which strain delocalization prevails with extensional deformation of the lower lithosphere being evenly distributed, rather than localized to a discrete crustal-scale detachment (e.g., McKenzie, 1978; Buck, 1991).

The outlined rift evolution of the East Greenland continental margin suggests that it was not until post-Campanian rifting (Rift Phase 3) that strain weakening mechanisms were sufficient to outpace strain hardening mechanisms and initiate seafloor spreading. This transformation may have occurred in response to contemporaneous magmatic underplating (e.g., Voss and Jokat, 2007, 2009; Hermann and Jokat, 2016) and increased extensional magmatism (e.g., Olesen et al., 2007; Gernigon et al., 2009; Larsen et al., 2014; Hermann and Jokat, 2016), which would have intensified rheological heterogeneities in the lithosphere and promoted strain localization (e.g., Ebinger, 2005; Lavér and Manatschal, 2006; Yamazaki et al., 2006; Yamazaki and Gernigon, 2009).

The subsurface identification of seafloor spreading–associated low-angle extensional structures and seaward-dipping reflectors outboard of the East Greenland coastline (Berger and Jokat, 2008) are more indicative of a simple shear rift model (e.g., Wernicke, 1985) characterized by high heat fluxes and associated magmatism, thermal weakening, and localized deformation on a crustal-scale low-angle detachment (e.g., Lister et al., 1986; Klepeis et al., 2007). However, the dissection of domino-style fault blocks during rifting observed across East Greenland does not indicate or require the presence of a lower to mid-crustal low-angle detachment, and low P values previously estimated from for Cenozoic rifting in the Traill Ø region (\(P = 1.05 \pm 0.01\) to \(1.22 \pm 0.06\); Price et al., 1997) are more comparable to a pure shear rift model. As such, it is unclear which model best describes the evolution of Rift Phase 3.

However, differences between rift phases suggest that changes in the rheological boundary conditions that governed the style of rifting in East Greenland occurred during, or immediately prior to, Rift Phase 3.

In summary, the rift evolution of the East Greenland continental margin and Norwegian-Greenland Sea as described here and by others (e.g., Surlyk, 1990; Price et al., 1997; Lundin and Dore, 2002; Olesen et al., 2007; Voss and Jokat, 2007, 2009; Gernigon et al., 2009) was probably influenced by the pre-rift structure of the lithosphere and the subsequent modifications made to it during multiple syn-rift to post-rift cycles. The proposed evolution highlights an important distinction between the evolution of progressive rifting and polyphase rifting. A similar polyphase rift evolution was described by Bell et al. (2014) for the Horda platform in the northern North Sea; they identified two distinct phases of rifting during Permian-Triassic and Middle Jurassic–Early Cretaceous time and noted an increase in the number of active faults and decrease in the spacing between them in the second phase of rifting, relative to the first. Bell et al. (2014) attributed this to a change in rheological state of the lithosphere and change in the orientation of principal stresses that occurred between rift phases, promoting the initiation of new faults in the second phase of rifting over the complete reactivation of preexisting faults. It is from this same rift system that Cowie et al. (2005) define a model of strain localization from the East Shetland Basin during Late Jurassic rifting that preferentially localizes on preexisting Permian-Triassic normal faults, leading to a decrease in the number of active faults and an increase in the spacing between them. Comparison of these studies highlights the importance of additional lithospheric modifications that occur during the inter-rift phases of a polyphase rift evolution (i.e., post-rift phases between multiple rift phases) that may prevent or hinder preexisting structures from accommodating most or all of the subsequent extensional deformation.

**CONCLUSIONS**

A new geological map and cross sections have been constructed for the Traill Ø region of NE Greenland. Stratigraphic morphologies, angular discordances, and crosscutting relationships between strata and faults indicate distinct intervals of rifting during the Devonian–Triassic, Jurassic–Cretaceous, and Cenozoic.

The distribution and geometry of units within the Cretaceous (Ryazanian–Campanian) succession has been determined in order to investigate faulting and sedimentation during Jurassic–Cretaceous rifting. Based on the stratigraphic distribution of specific macrofauna identified across the Traill Ø region, the Cretaceous strata have been divided into seven new mappable units: the *Buchia* Beds (latest Ryazanian–Valanginian), *Inoceramus auctell* Beds (early Hauterivian–Aptian), *Inoceramus anglicus* Beds (early–late Albian), *Inoceramus crippsi* Beds (early–late Cenomanian), *Inoceramus lamarki* Beds (Turonian–mid-Coniacian), *Sphenoceramus* Beds (late Coniacian–early Campanian), and *Scaphites* Beds (middle to late Campanian).
The base-Cretaceous unconformity (BCU) in the Traill Ø region is well developed near the crests of tilted fault blocks. Comparison with time-equivalent strata in Wollaston Forland suggests syn-rift sedimentation in the hanging-wall successions between Volgian to Ryzanian time may have been continuous. Consequently, there may be no discontinuity between Jurassic and Cretaceous strata in these syn-rift sediment packages. Construction of cross sections indicates that Volgian to mid-Ryzanian syn-rift strata exist in subsurface hanging-wall successions across the Traill Ø region. Based on a Jurassic-Cretaceous angular unconformity of 4°–8° at the crests of tilted fault blocks, an estimated thickness of ≤600 m of Volgian–Valanginian syn-rift strata is predicted in the subsurface hanging-wall sequence of the Manédal fault in the Traill Ø region.

Following Volgian–mid-Albian rifting, an apparent reduction in sedimentation rate is identified between the Inoceramus crippsi Beds (1300 m thickness in 6 m.y.) and the Inoceramus lamarcki Beds (1100–1200 m thickness in 17 m.y.). This change in sedimentation rate corresponds to a shift that occurred between Cenomanian and Turonian time from structurally confined deposition within active fault bound intrabasins to relatively unconfined deposition within a larger, wider shelf margin basin with little or no internal bathymetric relief. This later unconfined sedimentation capped the underlying infilled intrabasins and reflects a transition from Early Cretaceous active faulting to Late Cretaceous post-rift subsidence, as East Greenland became a constructive margin to the Vøring and Møre Basins. The change in apparent sedimentation rates between the Inoceramus crippsi and Inoceramus lamarcki beds coincided with an increase in sediment input into the Vøring and Møre Basins during Turonian–Maastrichtian time, sourced from the East Greenland margin, and suggests that the aforementioned changes in basin confinement may correspond to a larger regional reorganization of basin morphology and sediment routing systems.

Synthesis of our record of fault activity in the Traill Ø region with geophysical studies reveals a three-stage, stepwise eastward migration of the area of concentrated extensional deformation, and a stepwise decrease in fault block width and stepwise increase in the number of active faults following orogenic collapse of the Caledonides in the Devonian. Three distinct rift phases are recognized during Devonian–Triassic (Rift Phase 1), Jurassic–Cretaceous (Rift Phase 2), and Cenozoic (Rift Phase 3) time. During Rift Phase 1, extensional deformation concentrated in the region east of the Western Fault Zone (WFZ) between ~80–100-km-wide fault blocks. During Rift Phase 2, extensional deformation concentrated in the region east of the Gauss Halvø Fault (GHF) between ~10–30-km-wide fault blocks. In the Traill Ø region, ~4°–8° fault block rotation occurred during Rift Phase 2. We recognize a third eastward migration of extension concentrating east of the Manédal fault during Rift Phase 3 (Cenozoic), as indicated by spatial variations in fault block rotation and the juxtaposition of Late Cretaceous hanging-wall strata against Carboniferous footwall strata. This mapping exercise indicates that more than 10° fault block rotation occurred during Rift Phase 3 and has identified several new faults that initiated during this rift phase, dissecting fault blocks to widths of 4–10 km. These findings suggest that most of the observed amounts of faulting, fault block rotation, and extension occurred during the Cenozoic rift interval.

Cenozoic rifting can be divided into at least three episodes of faulting, the first of which is a newly recognized episode in NE Greenland. “Pre-magmatic” faulting occurred between the late Campanian and Thetanian. This was followed by “post-magmatic” faulting after the emplacement of tholeiitic dolerite intrusions ca. 54 Ma, and before and after emplacement of alkaline dolerite intrusions ca. 36 Ma. We postulate that in the Traill Ø region, the Manédal fault largely defines the western limit of major upper crustal extensional deformation during Rift Phase 3, associated with development of the continental-ocean transition and the Aegir spreading center after ca. 56 Ma, and later formation of the Jan Mayen microcontinent and the Kolbeinsey spreading center after ca. 36 Ma.

In summary, the structural evolution of the East Greenland continental margin is defined by an eastward stepwise migration of the area of concentrated extensional deformation and an increase in the number of active faults and decrease in the spacing between them that occurs with each rift phase. The strain delocalizing characteristics of this polyphase rift evolution are not readily conformable with classic rift evolution models driven by strain localizing mechanisms. We suggest that strain delocalization was driven by domino-style fault block rotation, which caused a build-up in stress within the fault blocks that eventually required new steep faults to form that dissect the fault blocks to smaller widths. The stepwise eastward migration of deformation may have been driven by evolving lateral heterogeneities in lithospheric rheology that were initially developed during crustal thickening of the Caledonian orogen and subsequently modified during each proceeding rift and post-rift phase.

Our findings highlight an important distinction between the evolution of progressive rift systems defined by only a single episode of rifting and polyphase rift systems in which post-rift lithospheric processes occurring between syn-rift phases can play an influential role on the evolution of strain distribution. Importantly, the structural evolution of a polyphase rift system should not always be expected to progress toward the rift climax stage, as often displayed by mature progressive rift systems dominated by strain localization processes.

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