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Basin modelling of the SW Barents Sea

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Abstract

The SW Barents Sea is an epicontinental platform consisting of N- to NNE-oriented basins separated by basement highs. The basins were formed during four distinct rift-phases in the Carboniferous, Late Permian, Late Jurassic-Early Cretaceous, and Paleocene-Eocene. Progressive rifting culminated in continental breakup and seafloor spreading along the North Atlantic axis approximately at 55 Ma. We present tectonic and thermal models of basin evolution along two seismic profiles crossing the SW Barents Sea. The thermal and isostatic history of basins is constrained through time-forward basin modelling based on an automated inverse basin reconstruction approach. We estimate the effects of continental breakup and near-margin processes (magmatic underplating and sill intrusions) on the thermal history and kerogen maturity below the Vestbakken Volcanic Province. Basin models are calibrated against well data. The results imply that both breakup and underplating alter the thermal and isostatic history of sediments along the margin. The hydrocarbon potential of source rocks modelled along the margin suggests breakup has a permanent thermal effect on the present-day sediments promoted by lateral heat flow, while heat conduction by underplating is more diffuse, inhibited to some degree by deeply-buried, low-conductivity shales and limestones.

34 **Keywords**

35 Basin modelling; SW Barents Sea; source rock maturity; rifting; magmatic underplate;
36 breakup

37

1 Introduction

The Barents Sea is a shallow sedimentary platform on the northwestern part of the Eurasian continental shelf (**Fig. 1**). It is bounded to the south by the Norwegian and Russian landmasses, to the north by the Arctic Ocean, to the east by Novaya Zemlya, and to the west by the North Atlantic Ocean. The SW Barents Sea is characterized by a series of NNE-SSW oriented rift basins filled with clastic/carbonate sediments separated by basement highs, formed by multiple rifting events since the collapse of the Caledonian Orogeny in Devonian time. Rifting culminated into seafloor spreading and continental breakup along a sheared margin in Eocene time (Rønnevik, 1981; Faleide et al., 1984, 1993, 1996, 2008, 2015; Dengo and Røssland, 1992; Gudlaugsson et al., 1998). Breakup was followed by passive margin development which continues to present day.

It is well known that the occurrence of oil and gas in a sedimentary basin depends on the maturation of source rocks which is in turn controlled by the thermal history of the sediments. The latter therefore needs to be constrained in order to reduce uncertainty. The thermal history of sediments depends on burial, sedimentation rate, and the thermal properties of the sediments, crust and mantle lithosphere (Theissen and Rüpke, 2009; Marcussen et al., 2010). It also depends on the amount of heat provided by the sub-lithospheric mantle during the basin formation. Basin formation is usually explained by lithosphere extension (McKenzie, 1978). During the rifting phase, extension of the lithosphere causes local thinning of the crust and mantle lithosphere providing large heat supply to sediments deposited in a fault controlled subsiding basin. The magnitude of extension is characterized by stretching factors (β factors) for the crust and mantle

lithosphere. The rifting phase is followed by a tectonic quiescent post-rift phase characterized by thermal cooling of the entire lithosphere and sag subsidence.

The thermal history of sediments in a specific basin can be inferred from modelling the thermal and tectonic evolution of the basin. The thermal and mechanical properties of sediments are usually known data. However, the extension parameters (stretching factors) have to be constrained. There exist two different approaches for constraining the stretching factors. Backstripping determines the basin evolution in a time-reversed manner, progressively removing each stratigraphic sequence, decompacting sediments, removing slip along faults, and restoring horizons to their interpreted water depths. The stretching factors are then calculated through analysis of the subsidence history (Fjeldskaar et al., 2004). However this method has disadvantages, notably the thinning of sediment layers during rift phases is neglected leading to biased stretching factors for older rift episodes (Wangen and Faleide, 2008). In contrast, time-forward modelling uses coupled thermal, kinematic, and isostatic calculations of lithospheric deformation to produce sedimentary basins progressively through time. β factors and water depths are then tuned to fit the predicted stratigraphy to the observed basin (Rüpke et al., 2008).

Basin modelling is “rare” in the SW Barents Sea publication record. Glørstad-Clark et al. (2011a) performed basin modelling along an E-W seismic transect crossing the Loppa High using a backstripping approach. However, their work was not published in the peer-reviewed literature. Clark et al. (2014) performed basin modelling along a regional NW-SE seismic transect crossing sedimentary basin structures and platforms in the SW Barents Sea using the two different approaches, backstripping and inverse. Their models

satisfactorily reproduced the present-day geometry of the basins along the profile assuming 4 rifting episodes. However, near-margin processes were not taken into account (breakup, seafloor spreading and underplating, and sills) in the reconstruction. These processes may have a large influence on basin reconstruction, subsidence and thermal history of sediments and petroleum systems, especially at the Vestbakken Volcanic Province at the NW end of the profile.

Here, we model basin evolution using the inverse approach in order to evaluate the influence of the aforementioned processes on the thermal history of sediments and maturity in the SW Barents Sea margin. Basin modelling is performed along the PETROBAR 07 profile (same as Clark et al., 2014) crossing the SW Barents Sea extended by the HB3-96A seismic profile crossing the Knipovich Ridge. Basin modelling is also done along a second profile oriented W-E and crossing the Loppa High (Glørstad-Clark et al., 2011a) (see **Fig. 1**). This paper summarizes the master thesis of Hansford (2014).

In the following we first summarize the geology of the SW Barents Sea. Second, we describe the data. We then present the methods. We use the commercial forward modelling software TecMod2D (Rüpke et al., 2008) to develop tectonic evolution models along two seismic transects crossing the SW Barents Sea. We test several models, each including various tectono-magmatic processes (breakup, seafloor spreading, underplating, and sills). Last, we discuss the tested models and the consequences of the different processes on the subsidence, thermal history of sediments and maturity.

2 Geological background

2.1 Geological evolution of the SW Barents Sea

The western Barents Sea is bounded to the west by the continent-ocean boundary resulting from the breakup of the North Atlantic (Faleide et al., 1993; Gudlaugsson et al., 1998) and, to the north, by the Eurasian Basin (**Fig. 1**).

The crystalline basement of the western Barents Sea consists of Caledonian igneous and metamorphic rocks (Ritzmann and Faleide, 2007; Gernigon and Brönnert, 2012; Marelli et al., 2013). The Caledonian orogeny resulted from the oblique continental collision of the Baltica and Laurentia cratons between Ordovician and Early Devonian times (Roberts and Gee, 1985; Roberts, 2003). The Caledonian grain set up a structural framework that has influenced later structural development in the area (Faleide et al., 1984; Gudlaugsson et al., 1998; Breivik et al., 2002; Ritzmann and Faleide, 2007). Following the collapse of the Caledonian orogenic belt in the Early Devonian, the western margin of Baltica (including the western Barents and Vøring margins) acted as a transfer zone linking extension occurring in the present-day North Atlantic and Arctic regions (Ziegler, 1988; Doré, 1991; Gudlaugsson et al., 1998; Clark et al., 2014).

The basins of the western Barents Sea (**Fig. 1**) were formed in response to four major rift phases, Late Devonian?-Carboniferous, Late Permian, Middle Jurassic-Early Cretaceous, and Early Cenozoic (Faleide et al., 2008; Faleide et al., 2015; Blaich et al., 2017; Serck et al., 2017). Generally, the locus of extension and sediment deposition migrated westwards (Dengo and Røssland, 1992; Faleide et al., 2008).

During the Late Devonian?-Carboniferous rift event, most of the Barents Sea was affected by crustal extension. Interconnected half-grabens were created with fan-shaped configurations along a N-NE trend (Faleide et al., 1984; Rønnevik and Jacobsen, 1984; Gabrielsen et al., 1990; Dengo and Røssland, 1992; Gudlaugsson et al., 1998). The Late Carboniferous to Early Permian post-rift phase was marked by deposition of evaporites in narrow basins covered by a shallow water carbonate platform (Faleide et al., 1984). Clastic sediment infill then increased throughout the Permian (**Fig. 2**).

The Late Permian rifting event provided the accommodation space required for large volumes of clastic sedimentation. The Ural orogenic belt, the Baltic Shield, and other locally uplifted areas are thought to be the source of clastic sediments (Faleide et al., 2015; Glørstad-Clark et al., 2010; Glørstad-Clark et al., 2011b). Through the Middle Triassic, the SW Barents Sea basin shallowed considerably from predominantly marine depositional conditions to continental and deltaic conditions (Glørstad-Clark et al., 2011b). The latest Triassic was characterized by a marine regression and erosion. Uplift of the Selis Ridge (paleo-Loppa High) is observed in the Early and Middle Triassic, evidenced by widespread erosional surfaces. However, throughout the Late Triassic the paleo-high was a depocenter (Glørstad-Clark et al., 2010). As the Barents Sea progressively filled with clastic sediments throughout the Triassic, the load on the Carboniferous evaporite deposits in the Nordkapp and Ottar basins triggered upwards salt mobilization (Gabrielsen et al., 1990). Clastic sedimentation continued into the Jurassic, with Early-Middle Jurassic sandstones found over the entire western Barents Sea.

A major rift event is recorded from the Middle Jurassic to Early Cretaceous. The event is characterized by two rift pulses in Late Middle Jurassic - earliest Cretaceous and Aptian (Faleide et al., 1984, 1993; Serck et al., 2017) giving rise to narrow and deep basins, such as the Harstad, Tromsø and Bjørnøya basins. Salt tectonics (halokinesis) was prevalent in the Tromsø Basin in the Late Cretaceous. The Loppa High was inverted in Early Barremian - Early Aptian (Indrevær et al., 2016), it was an island throughout the Cretaceous (Faleide et al., 1993).

A widespread unconformity marks the boundary between the Cretaceous and Early Cenozoic sediments. Dextral oblique-slip along the proto-breakup margin – the De Geer Zone – accommodated Late Cretaceous – Paleocene extension, and produced pull apart basins in the Wandel Sea (present-day NE Greenland) and the western Barents Sea (Vestbakken Volcanic Province and Sørvestsnaget Basin) (Faleide et al., 2008). The De Geer megashear zone, which linked the Norwegian-Greenland Sea with the Eurasia Basin in the north, consisted of two wrench-faulted segments; the Hornsund Fault Zone in the north (between Bjørnøya and Svalbard) and the Senja Fracture Zone in the south, separated by a releasing bend where rifting dominated. Continental breakup and seafloor spreading occurred along this predominantly sheared margin at approximately 55-54 Ma (Faleide et al., 2008). The different components of the De Geer system had different breakup histories. The Hornsund and Senja fracture zones developed by continent-continent shear followed by continent-ocean shear. Both margin segments have clearly identifiable continent-ocean boundaries, and have been passive since earliest Oligocene time (Faleide et al., 2008). Breakup-related magmatism and partly associated uplift is recorded in the Vestbakken Volcanic Province from Paleocene to Eocene. Interpretation

of deep seismic data suggests it is magmatic underplating (Faleide et al., 1991) or an intruded lower continental crust (Libak et al., 2012a). Based on the interpretation of gravity and magnetic data, Clark et al. (2013) suggested a 1-2 km thick sill layer within the lower sedimentary succession of the Vestbakken Volcanic Province. Breakup was followed by down-faulting and deposition of thick Eocene strata. At the end of the Eocene, a narrow oceanic basin existed between the western Barents Sea and NE Greenland continental margins. A plate reorganization event in the earliest Oligocene caused Greenland to move in a more westerly direction with respect to Eurasia, and reactivated faults and renewed magmatism in the Vestbakken Volcanic Province (Faleide et al., 1991).

Reactivation of faults and uplift is recorded in the Middle to Late Cenozoic. In the Late Miocene, a regional tectonic uplift event is recorded with increasing amplitude from west to east (Dimakis et al., 1998). Following the Neogene uplift, global climate change at 2.6 Ma brought about the Northern Hemisphere glaciation, marked by a basin-wide unconformity. Glacial erosion exacerbated by Plio-Pleistocene uplift removed kilometer-thick layers of Cenozoic sediments from the shelf and deposited huge prograding wedges of sandy/silty muds along the western margin (Faleide et al., 1996). Debate continues, however, in the literature (Dimakis et al., 1998; Green and Duddy, 2010; Henriksen et al., 2011; Laberg et al., 2012) regarding the relative contribution to the total Cenozoic erosion budget of the Neogene uplift and the Quaternary glaciations.

2.3 Hydrocarbon source rocks in the SW Barents Sea

Source rocks ranging from Carboniferous to Cretaceous in age have been identified within the SW Barents Sea (**Fig. 2**). The five main source rock groups within the SW Barents Sea include gas-prone coals and mudstones from the Early Carboniferous Billefjorden Group, oil-prone Late Permian carbonaceous mudstones and limestones of the Tempelfjorden Group, oil- and gas-prone Triassic marine shales, oil- and gas-prone Late Jurassic shales of the Hekkingen Formation, and Cretaceous source rocks with significant proportions of organic material (Vestbakken Volcanic Province and Sørvestsnaget Basin) (Larssen et al., 2002).

3 Data

3.1 2D seismic profiles

Three seismic profiles are used for the basin modelling (**Fig. 1**).

Loppa High profile

The Loppa High profile is 262 km long running in an approximate east-west direction from the Nordkapp Basin in the east to the Tromsø Basin in the west (**Fig. 3a**). The stratigraphy was compiled from 2D multi-channel seismic reflection surveys, depth-converted and interpreted by Glørstad-Clark et al. (2011a), following the velocity model of Clark et al. (2013).

PETROBAR-07 profile

The PETROBAR-07 profile (**Fig. 3b**) extends from the coastline outside Finnmark in the south-east to within 14 km on the continent-ocean boundary inside the Vestbakken Volcanic Province in the north-west (location map in **Fig. 1**). The PETROBAR-07 profile was depth-converted and interpreted by Clark et al. (2013, 2014). The oldest horizons are hardly interpretable because they are so deeply buried that they are into the low-grade metamorphic window. Thickness trends were just flattened out in decompacted sections to estimate the original thicknesses of those units (Clark et al., 2014).

HB-3-96A profile

The HB-3-96A short profile runs from the Vestbakken margin to the Knipovich Ridge (**Fig. 3b**). The horizons and faults are depth-converted and merged to extend the input stratigraphy of the PETROBAR-07 profile for some of the models.

The horizon interpretation of the seismic profile HB-3-96A was calibrated using the 7316/5-1 well located approximately 4.8 km north of the line (within the Vestbakken Volcanic Province; **Fig. 1**). A total of 7 horizons were identified and interpreted. The ages of these horizon range from 0 to 55 Ma (Seabed – Base Eocene). The horizons were picked based on well formation tops from the 7316/5-1 and a series of published references based on work along the continental margin (Eidvin et al., 1994, 1998; Faleide et al., 1996; Libak et al., 2012b; Solbu, 2013).

The Top Basalt/Base Eocene horizon was mapped across the profile and over the Knipovich Ridge. This horizon represents the top of volcanic extrusives and shallow intrusives associated with continental breakup.

The seabed is a distinct, strong, positive reflection across the HB-3-96A line.

Reflections are predominantly parallel below the seabed. The Base Pleistocene unconformity marks the onset of glacial erosion on the shelf and outbuilding of sedimentary fans towards the margin. The Base Oligocene unconformity pinches out at the Early Miocene unconformity within the Vestbakken Volcanic Province. Significant normal faulting with rotated fault block geometries can be identified throughout the Vestbakken Volcanic Province up to the Early Miocene and Base Pleistocene horizons.

The transition between the continental platform in the east and oceanic crust in the west is defined by large, deep, steeply west-dipping faults. This interpretation is in agreement with the work of Faleide et al. (1996); Libak et al. (2012b); and Clark et al. (2013). The COB is located approximately 152 km east of the Knipovich Ridge.

By comparing the TWT horizon intersections on the HB-3-96A seismic line with formation tops identified in the 7316/5-1 well (Eidvin et al., 1998), an estimate of average velocities between each identified horizon was made. Calibration of the velocity model was conducted by comparing the results with those derived from the PETROBAR-07 study (Clark et al., 2013), the BIN-2008 study (Libak et al., 2012a), the BIS-2008 study (Libak et al., 2012b), and Barents Sea continental margin seismic refraction compilations (Fiedler and Faleide, 1996). Each horizon was converted to depth in metres.

251

252 3.2 Well data

253 Data from a series of wells drilled within the vicinity of the Loppa High profile,
254 PETROBAR-07 profile and HB-3-96A profile (**Fig. 1**) were used to verify and calibrate
255 the models. The data used include vitrinite reflectance (R_o), bottom hole temperature, and
256 Drill Stem Test (DST) temperature measurements. They were acquired from the
257 Norwegian Petroleum Directorate. While the three wells are located within 4-12 km of
258 the sections B_short and B_long, there are only two wells located close to the section A
259 (7224/7-1 and 7226/11-1). The wells further west (7120/2-1 and 7121/111R) are located
260 over 40 km south of the section, so their value as calibration points is somewhat limited.
261 A series of pseudo-wells were also included along the sections for direct comparison with
262 interpreted paleo-water depths by Clark et al. (2014) and Glørstad-Clark et al. (2011a).
263 These are shown in **Fig. 4** (dashed lines).

264

265

266 4 Method

267 4.1 TecMod2D

268 TecMod2D is a basin modelling software package that automates sedimentary basin
269 reconstruction in 2D. It is based on an algorithm which couples a forward lithosphere
270 stretching model to an inverse scheme which automatically updates crustal- and mantle-

stretching factors (β and δ respectively) and paleo-water depths until the input stratigraphy is fitted to desired accuracy (generally within 5-10% error) (Rüpke et al., 2008, 2010). The details of the method are described in Rüpke et al. (2008).

Forward Model

The 2D forward model is based on pure shear kinematics (McKenzie, 1978) and allows for multiple rifting events of finite duration (Jarvis and McKenzie, 1980). The numerical domain is split into vertical columns. Each column is assigned a crustal- (β) and mantle (δ)-thinning factor. Along the vertical direction, the numerical domain is subdivided into units: air, water, sediments, upper crust, lower crust, lithospheric mantle, and asthenosphere. The numerical grid is deformed using an arbitrary Eulerian-Lagrangian approach (Fullsack, 1995), where the grid is moved horizontally according to crustal thinning and is fixed in the vertical direction.

The effects of flexural isostasy and depth of necking are included (Watts et al., 1982; Braun and Beaumont, 1989). Vertical loads due to both thinning and temperature changes act on an elastic plate, whose flexural rigidity depends on the effective elastic thickness, T_e . The velocity field derived from pure shear kinematics and crustal flexure is used to advect the temperature field. The time-dependent heat-transport equation includes advection and diffusion and is solved in the entire modeling domain.

Crustal radiogenic heat production decreases exponentially with depth (Jaupart et al., 1981; Turcotte and Schubert, 1982) and is assumed constant in the sediments. Water and sediments are included into the thermal solver to account for the effects of sediment

blanketing (Theissen and Rüpke, 2009). Sedimentation is controlled by paleo-water depth maps and/or sedimentation rates determined by the inversion scheme. The deposited sediments are compacted using empirical compaction laws (Royden and Keen, 1980; Sclater and Christie, 1980).

The boundary conditions for the thermal solver are fixed temperatures at the base and top of the numerical domain and zero horizontal heat flow at the sides. Density changes are computed from a reference density and the thermal expansion factor (McKenzie, 1978).

Automated Inversion Methods

During the automated reconstruction, the misfit between observed and modelled stratigraphic data is to be minimized. The inversion consists of the iterative search for the optimal set of δ , β , and paleo-water depth values, which progressively reduce the misfit (Poplavskii et al., 2001; White and Bellingham, 2002). The first iterative forward run requires an initial guess for δ , β , and paleo-water depth. The initial guess for the thinning factors is based on analytical solutions for isostasy and thermal subsidence. The initial guess for paleo-water depth is based on the present-day water depth. After every forward run, the misfit between the input and computed stratigraphy is used to update the δ and β factors and paleo-water depth until the input stratigraphy is fitted to desired accuracy (generally within 5-10% error).

External processes

Extra features within the forward model can be applied for specific scenarios. Those among others include (i) Continental breakup, spreading ridges, and formation of new

oceanic crust; (ii) Generalized igneous events to model the thermal and structural effects of underplating, extrusive flows, and sill intrusions; and (iii) Forced sediment deposition and erosion. .

Breakup

TecMod2D provides the possibility to include multiple breakup events into forward models and into reconstruction runs. Continental breakup can be solved in different ways within TecMod2D. In the normal mode, breakup results in seafloor spreading inside the modelling domain. New oceanic crust is created and the split continent drifts apart in both directions. It is also possible to isolate the impact of breakup on one margin without the conjugate side. This is achieved by putting the spreading axis on the respective boundary (left or right) so that only half of the ridge system is simulated. The time of breakup and end of spreading parameters must be specified.

Igneous events

The following kinds of igneous events can be implemented: magmatic underplating, sills, and extrusive lava flows. Any number and combination of igneous events can be used but they cannot overlap in time. The time of intrusion, spatial coordinates and rock type must be defined. In case the igneous event is a sill, the intrusion level (this is the age of the sediments where the sill intrudes) must be specified. In case the igneous event is a magmatic underplate, the thickness and width of body must be specified. The magmatic underplate will locally replace mantle material beneath the Moho.

Erosion & Forced deposition

Sedimentation within TecMod2D runs is controlled by paleo-water depths maps and/or sedimentation rates determined by the inversion scheme. If accommodation space exists, sediments will be deposited until the resulting water depth matches the one from the paleo-water depth map. These maps are either user-prescribed or determined by TecMod2D's inverse algorithm. Sedimentation rates can further be restricted by limiters.

Sediment erosion is handled in TecMod2D through erosion events. An event is specified through a time and space range during which sediments that are located above a specified depth (erosion level) are eroded with a specified rate. TecMod2D allows for the forced deposition of sediments to compensate for later erosion event.

4.2 Input parameters

Basin modelling is performed along three stratigraphic sections built on the three previously described seismic profiles.

Section A

The stratigraphic section A is an interpretation of the Loppa High profile (**Fig. 4a**) (Glørstad-Clark et al., 2011a). It is oriented east-west, crossing the PETROBAR-07 profile on the Bjarmeland Platform (**Fig. 1**). The section A intersects the deep Tromsø Basin, Polheim Subplatform, Loppa High, and Bjarmeland Platform over a distance of 262 km.

Section B_short

This section solely includes the PETROBAR-07 profile (550 km long; **Fig. 4b**, from $x = 166$ to $x = 716$ km) published in Clark et al. (2014). This section is used for the reference model.

Section B_long

This section includes the PETROBAR-07 profile (Clark et al., 2014) merged with the HB-3-96A line to the Knipovich Ridge (approximately 716 km long; **Fig. 4b**). The continent-ocean boundary is set at $x = 152$ km. The base of the sedimentary succession below the Vestbakken Volcanic Province averages 12 km, and is relatively flat, as mapped by Clark et al. (2014). Deepest sediments follow an arbitrary base Eocene (55 Ma) horizon west of the COB.

Model parameters

Each of the three sections was loaded into TecMod2D for analysis. Rock properties were assigned to each stratigraphic layer. Assumptions regarding the lithology of each unit were made, choosing either the dominant lithology (eg. limestone) or mixtures (eg. 50% sand, 50% shale). Likewise, porosity-depth trends linked to mechanical compaction during burial were applied to each rock unit based on the assumed dominant lithology, while sand-shale mixtures were linearly interpolated based on their ratios. For simplicity, chemical compaction, diagenesis, and low-grade metamorphism were neglected (Clark et al., 2014). Rock properties assigned to each stratigraphic layer are presented in table 1.

Four rifting phases were defined that approximate the complex rifting episodes from Late Carboniferous to present: Late Carboniferous rifting (320 – 305 Ma), Late Permian

379 rifting (271 – 251 Ma), Late Jurassic – Early Cretaceous (165 – 145 Ma) and Paleocene -
 380 Eocene (65 – 55 Ma). The rift phases are summarized in table 2.

381 Model input parameters are summarized in table 3. Flexural isostasy is applied through
 382 an effective elastic thickness (T_e) of 5 km and a corresponding necking depth of 15 km.
 383 These values are difficult to constrain and vary spatially and temporally, but are the same
 384 as Clark et al. (2014) and are consistent with other published models for the Viking
 385 Graben and Vøring Basin (Fjeldskaar et al., 2004, 2009; Rüpke et al., 2008; Theissen and
 386 Rüpke, 2009).

387 Post-Caledonian crust and lithosphere varied in thickness along the profiles before rift
 388 initiation, but this geometry is unknown. Constant initial crustal thickness of 35 km (17.5
 389 km upper crust, 17.5 km lower crust) and a total lithospheric thickness of 120 km are
 390 therefore used (Clark et al., 2014).

391 Other main forward model parameters include temperature boundary conditions of 5°C at
 392 the seafloor, and 1300°C at the Lithosphere-Asthenosphere Boundary and a 2 $\mu\text{W}/\text{m}^3$
 393 radiogenic heat production in the crust. The numerical resolution of the finite element
 394 mesh is set at $n_x = 100$ and $n_y = 200$.

395 Potential low-grade metamorphism in the deep Bjørnøya and Tromsø basins is ignored.

396 *Erosion parameters*

397 Erosion estimates for the section A (Glørstad-Clark et al., 2011a) and the sections
 398 B_short and B_long (Clark et al., 2013) are utilized within this study. Glacial erosion in

the SW Barents Sea (1000-1100m) is larger than the pre-glacial erosion (600-700m), ~65-70% of eroded sediments were during the glaciation period (Fielder and Faleide, 1996; Laberg et al., 2012; Baig et al., 2016). In our modelling, the erosion is limited to the glacial-related erosion events in the Quaternary. Glacial erosion estimates of between 0 – 1000 m in the Tromsø Basin and 1000 – 1500 m in the Hammerfest Basin and Loppa High have been recorded (Laberg et al., 2012; Clark et al., 2014; Baig et al., 2016). For the sections B_short and B_long, the erosion level (400 m depth across most of the section, deepening towards the shelf) and erosion rate (maximum 10 mm/yr from 2.6 to 0 Ma) are the same as those interpreted by Clark et al. (2014). For the section A, the erosion level is 400 m across most of the profile, , while the erosion rate is maintained at 10 mm/yr from 2.6 to 0 Ma.

5 Results

Automated basin reconstructions were performed along the section A, the sections B_short and B_long crossing the SW Barents Sea. A series of models were run across the section B_long for testing the thermal and isostatic implications of continental breakup (and seafloor spreading), breakup with emplacement of an underplate below the Vestbakken Volcanic Province, and breakup with underplating and sill intrusions within the sediments of the Vestbakken Volcanic Province.

For each model the distribution of crustal stretching factors and the thermal history of the basin were computed and investigated. Models were calibrated against vitrinite

reflectance and bottom-hole temperature measurements from well data. Computed paleo-water depths were assessed.

5.1 Modelling along the section A

The automated basin reconstruction across the section A is presented in **Fig. 5**. Erosion is simulated from Late Pliocene to Present-day. The deposition of 1 – 1.5 km of sediments is forced from 120 to 2.6 Ma across most of the profile (maximum over the Loppa High, tapering in the NW through the Vestbakken Volcanic Province) to ensure they are available to be eroded from 2.6 Ma to present-day.

The convergence of the modelled stratigraphy to the input stratigraphy is good (approximately 5%) along the section after 11 iterations. A 350 m error in the fit for sediments over the Loppa High is observed.

A maximum cumulative β factor of 3.4 is modelled over the deep Tromsø Basin (**Fig. 6b**). Maximum stretching occurs during the Late Jurassic – Early Cretaceous rift phase (**Fig. 6e**), but this stretching is focused solely over the deep basin, and is absent from the Bjarmeland Platform further east. Elevated β -factors are modelled during the Early Carboniferous rift phase over the Bjarmeland Platform (**Fig. 6c**).

A trend can be seen in the β -factor distributions, with the main axis of extension slowly migrating westward until eventual Eocene. The basins and platform areas in the east, including the Bjarmeland Platform were the main focus of extension during the

Carboniferous (**Fig. 6c**) while milder Carboniferous extension is simulated below the future Tromsø Basin. Diffuse, low-magnitude rifting ensued during the Late Permian inferred rift event (**Fig. 6d**). Slowly, the main axis of extension migrated westward toward the Tromsø Basin in the Late Jurassic-Early Cretaceous (**Fig. 6e**).

Computed paleo-water depth outputs (black line, **Fig. 7**) are compared with interpreted paleo-water depths from Glørstad-Clark et al. (2011a) (dashed line). Their estimates are based on clinoform relationships from seismic sequence stratigraphy studies (Glørstad-Clark et al., 2010). In general, the modelled water depths (**Fig. 7a**) across the Tromsø Basin are much deeper during the Jurassic rift event when compared with the interpreted water depths. Low sedimentation rates (0,2-0,4 mm/yr) required for optimal reconstruction explain why each rifting phase is accompanied with a large increase in water depths. The Loppa High and Polheim-Sub-platform were subject to syn-rift flank uplift during the Late Permian rift phase, resulting in an island forming from 270-260 Ma (**Fig. 7c**).

Along the profile, 500-600 m of erosion is modelled consistently from the Polheim Sub-platform and Loppa High, eastward along the Bjarmeland Platform (**Fig. 6a**). Approximately 100-200 m of erosion is modelled through most of the Tromsø Basin. The amount of erosion modelled is within the limits, or less than published estimates (eg. 0 – 1000 m glacial erosion estimate in the Tromsø Basin; Riis and Fjeldskaar, 1992; Laberg et al., 2012; Baig et al., 2016).

Model fit to data

Vitrinite reflectance (R_o), bottom hole temperature, and Drill Stem Test (DST) temperature measurements were acquired for 4 wells within 45 km of the section A (**Fig. 4a** for well locations). **Fig. 8** shows the R_o measurements plotted against the modelled R_o trends from the modelled Loppa High section and the temperature measurements plotted against the modelled geotherms.

Well control points show reasonable correlation when comparing against most of the bottom hole and DST temperatures, but with considerable variation at well 7226/11-1 (**Fig. 8d**). Bottom hole mud temperature measurements are not necessarily in equilibrium with the formation temperature, so the observed variability may be a function of the measurement, not the validity of the model.

The vitrinite reflectance values also show good correlation for the wells located close to the profile (**Fig. 8c** - 7224/7-1 and **Fig. 8d** - 7226/11-1), however the more distant wells (**Figs. 8a & b**) show discrepancies.

5.2 Modelling along section B_short: reference model B0

A reference model is required along the section B_short to allow for comparison with more detailed models incorporating continental breakup together with magmatic underplating. The reference model solely includes the 550 km long PETROBAR-07

profile (**Fig. 4b**). The automated basin reconstruction across the section is presented in **Fig. 9**.

Late Cenozoic erosion is applied. The deposition of 1 – 1.5 km of sediments is forced from 120 to 2.6 Ma across the most of the profile (maximum over the Loppa High, tapering in the NW through the Vestbakken Volcanic Province) to ensure they are available to be eroded later.

The convergence of the modelled stratigraphy to the input stratigraphy is good (approximately 5%) after 12 iterations.

A maximum cumulative β factor of 4.1 is modelled over the deep Bjørnøya Basin (black line, **Fig. 10b**). Maximum stretching along the profile occurs during the Late Jurassic – Early Cretaceous rift phase, but this stretching is focused solely over the deep basin, and is absent from the Bjarmeland Platform further east (**Fig. 10c**). Elevated β -factors are modelled during the Late Carboniferous rift phase (**Fig. 10c**), especially over the Bjarmeland Platform, including the Ottar and Nordkapp basins.

Like for section A, the main axis of extension migrated westward until eventual continental breakup. The basins and platform areas in the east, including the Nordkapp and Ottar basins were the main focus of extension during the Carboniferous (**Fig 10c**). Diffuse, low-magnitude rifting ensued during the Late Permian rift event (**Fig 10d**). The main axis of extension jumped westward toward the Bjørnøya Basin in the Late Jurassic- Early Cretaceous (**Fig 10e**). And finally, immediately prior to continental breakup,

elevated levels of stretching are recorded within 100 km of the margin in the Vestbakken Volcanic Province (**Fig 10f**).

Computed paleo-water depth output (black line, **Fig. 11**) from the reference model is compared with interpreted paleo-water depths (dashed line) from Clark et al. (2014). Their estimates are based on clinoform relationships from seismic sequence stratigraphy studies. In general, the modelled water depths (black line) along the Bjørnøya Basin are much deeper during the Jurassic rift event when compared with the interpreted depths (purple line; **Fig. 11b**). Like for the previous model along section A, low sedimentation rates (0,2-0,4 mm/yr) during the rifting phases explain the large increase in water-depths.

The modelled Loppa High was subject to syn-rift flank uplift during the Late Permian rift phase, resulting in an island forming from 270-260 Ma. Syn-rift flank uplift on the Loppa High is also observed during the Jurassic rift phase (**Fig. 11c**). These uplift episodes are the flexural response of the basement to rifting. Following the Jurassic rift phase, late post-rift subsidence is calculated, but the forced deposition fills the Loppa High from 120 Ma ensuring it remains at or above sea level until the Quaternary glacial period, which is largely in agreement with our current knowledge.

Significant amounts of erosion (500 – 700 m) are modelled over the Loppa High and east of the Nordkapp Basin (**Fig. 10a**). However, very little erosion could be modelled over the Ottar Basin, and west of the Loppa High through the Bjørnøya Basin and Vestbakken Volcanic Province.

The reference model matches general estimates of the topographic movement of the Selis Ridge/Loppa High.

Each rifting phase is also accompanied with an increase in basement heat flow (**Fig. 14**). The increase in basement heat flow is of similar amplitude for each rifting event. However, as the radiogenic crust becomes thinner through time, the total heat flow budget diminishes. This is highlighted by a long-term negative (decreasing) trend in heat flow.

Model fit to data

Vitrinite reflectance (R_o), bottom hole temperature, and Drill Stem Test (DST) temperature measurements were acquired for 3 wells located within 12 km of the profile B (**Fig. 4b** for well locations). **Fig. 12** shows the R_o measurements plotted against the modelled R_o trends and the temperature measurements plotted against the geotherms from the reference model. Well control points show reasonable correlation for the reference model when comparing against most of the bottom hole and DST temperatures, but with variation at well 7316/5-1 (**Fig. 12a**). Bottom hole mud temperature measurements are not necessarily in equilibrium with the formation temperature, so the observed variability may be a function of the measurement, not the validity of the model. The vitrinite reflectance values also show good correlation for the two wells intersecting the profile (**Figs 12b and c**), but poor correlation with the well in the Vestbakken Volcanic Province (7316/5-1, **Fig. 12a**).

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542 **5.3 Modelling along the section B_long: continental break-up and sea-floor**
543 **spreading (model B1)**

544 The position and timing of continental breakup along the sheared western Barents Sea
545 margin is well studied, based primarily on interpretations of seismic and potential field
546 geophysical data (Breivik et al., 1998; Breivik et al., 1999; Breivik et al., 2003; Faleide et
547 al., 1991; Faleide et al., 1993; Faleide et al., 2008; Libak et al., 2012a, 2012b). The
548 position of the continent-ocean boundary (COB) used in this study was mapped
549 accurately based on comprehensive gravity and seismic studies (Breivik et al., 1999;
550 Libak et al., 2012a, 2012b; and Clark et al., 2013). The models incorporating continental
551 breakup simulate replacement of continental crust by oceanic crust west of the COB. The
552 spreading axis is assigned to the left side of the modelling domain. Asthenosphere
553 temperatures are set to mantle values at the spreading axis (1300°C). A spreading velocity
554 of 0.55 cm.yr⁻¹ is used, consistent with published accounts (Faleide et al., 2008). Faleide
555 et al. (2008) reported the timing of final lithospheric breakup at 55-54 Ma along the
556 Norwegian margin. This is synchronous with the end of the final rifting episode between
557 65-55 Ma. Breakup is therefore set at 55 Ma.

558 The thickness of oceanic crust formed as a result of seafloor spreading is optimized to
559 provide the best fit for Cenozoic stratigraphy. The input oceanic crustal thickness is 14
560 km within models where breakup is included, with a density of 2900 kg.m⁻³.

The flexural boundary condition on the seaward boundary is assigned to approach a broken plate scenario, with $T_e = 0$ km, whereas the eastern flexural boundary condition remains at $T_e = 5$ km. This approach is similar to the work of Rüpke et al. (2010) along the Ghana transform margin.

Modelled crustal stretching factors (β) are higher for the Paleocene-Eocene rift event adjacent to the COB (red line, **Fig. 10f**) compared to the reference model (black line). This, in turn, equates to a higher cumulative β -trend at the COB, approaching $\beta = 4$ (**Fig. 10b**). The Late Jurassic-Early Cretaceous, Late Permian, and Late Carboniferous rift phase β -distributions are very similar to the reference model.

The paleo-water depths (red line, **Fig. 13**) are comparable to the reference model (black line) across most the profile, except within the Vestbakken Volcanic Province. Within 16 km ($x=182$ km; pseudo-well 'e') and 50 km (pseudowell 'f'; $x=216$ km) east of the COB, the water depths are average 600 m deep prior to the Paleocene-Eocene rift phase. During the final rift phase, approximately 250 m of additional subsidence is modelled. This is followed by 200-600 m of uplift associated with continental breakup. Breakup-related uplift is a response to the simulated broken-plate boundary condition ($T_e = 0$) applied to the left side of the model. A thermal uplift effect associated with lateral heat flow from the MOR is also a contributor. At 72 km (well 7316/5-1; $x=238$ km) from the COB, there are no obvious differences in the paleo-water depth trends between the breakup model B1 and the reference model B0, implying limited flexure related to continental breakup.

The heat flow pattern is generally similar to previous model B0. However, the continental breakup has a clear effect on Cenozoic heat flow close to the COB (16 km east; **Fig. 14a**

– red line), resulting in a 5-6 mW/m² increase in heat flow. At 50 km from the COB (**Fig. 14b**), a small 2 mW/m² heat flow increase is computed. The thermal effects of continental breakup are not recorded greater than 75 km away from the COB (**Fig. 14c**).

5.4 Modelling along the section B_long: continental break-up and magmatic underplating (model B2)

A 5 km thick magmatic underplate was inserted under the Vestbakken Volcanic Province at 55 Ma (at the end of the final rifting event from 65-55 Ma). This is based on the interpretation of deep crustal seismic transects by Faleide et al. (1991), who identified a reflector at ~ 17 km depth interpreted as the top of an underplated crustal unit, and a deeper reflector at ~ 22-23 km interpreted as the Moho. More recently, Libak et al. (2012b) interpreted this High Velocity Zone as intruded lower continental crust below the VVP. Despite differences, both interpretations have similar large-scale thermal and isostatic effects. The density of the magmatic underplate is set to 2900 kgm⁻³ in the model. This is the same density as the oceanic crust. The modelled underplate beneath the Vestbakken Volcanic Province is 90 km wide (x=152 to x=242km).

β -factor trends are significantly different within the Vestbakken Volcanic Province when comparing between the reference model B0 (black line), and the model B2 (green line) (**Fig. 10**). β -factors are considerably higher for the Late Carboniferous and Late Permian rift phases, especially toward the COB. This is due to the extra uplift associated with underplating, which is compensated for in the model by more crustal stretching.

Compared to the previous models, elevated stretching factors during the Late Carboniferous and Late Permian rift phases combined with low sedimentation rates (0,2-0,4 mm/yr) give significantly deeper paleo-water depths prior to underplate emplacement and breakup within the Vestbakken Volcanic Province (green line, **Fig. 13**). Significant uplift is recorded at 55 Ma, coinciding with the underplate emplacement: a 1000 m uplift close to the COB ($x=16$ km), an 800 m uplift in the Middle of the Vestbakken Volcanic Province ($x=50$ km) and a 600 m uplift near the eastern edge of the underplate ($x=72$ km).

Within the Vestbakken Volcanic Province, basement heat flow is considerably higher during the Late Carboniferous rift event, and during emplacement of the underplate at 55 Ma (green line, **Fig. 14**). The high heat flow in the Late Carboniferous is due to the higher levels of crustal stretching modelled during this period. This effect reduces with time, as the asthenosphere thermal anomaly cools.

A distinct peak in heat flow occurs 1 My after the emplacement of the underplate (**Fig. 14**), at 55 Ma, resulting in an 67% increase in heat flow within 16 km of the COB (pseudo-well 'e'; $x=182$ km). This effect diminishes away from the COB, averaging a 40% increase in heat flow 50 km (pseudo-well 'f'; $x=216$ km) and 72 km (well 7316/5-1; $x=238$ km) from the COB. This trend is due to the eastward dip of the underplate. The top of the underplate is shallower proximal to the COB, thereby delivering more heat to the overlying sediments close to the COB.

5.5 Modelling along the section B_long: continental break-up, magmatic underplating and sill intrusion (model B3)

Based on gravity and magnetic modelling, Clark et al. (2013) interpreted a 1-2 km thick sill layer within the lower sedimentary succession of the Vestbakken Volcanic Province. The thermal and isostatic implications of such an interpretation are tested here. A sill complex together with an extrusive volcanic flow unit are inserted at different depths at 55 Ma, contemporaneously with the initiation of underplating, and modelled together with continental breakup and underplate. The sills and volcanic flow were assigned a density of 2900 kgm^{-3} , temperature of $1200 \text{ }^{\circ}\text{C}$, and span 90 km within the Vestbakken Volcanic Province ($x=152$ to $x=242\text{km}$).

β -factor trends (blue line, **Fig. 10a**) are identical to model B2, but within the Vestbakken Volcanic Province cumulative β -factors are 3% less than the breakup and underplating only model. Paleo-water depth trends (blue line, **Fig. 13**) are a bit shallower than model B2.

An average reduction of 10 mW/m^2 in basement heat flow is observed upon sill intrusion (blue line, **Fig. 14**) compared to model B2 (break-up and underplating, green line). The observed decrease in heat flow is related to the position of the sills relative to the basement. The sills are above the basement (in the sedimentary infill) and therefore project a negative flow of heat down to the basement (heat flow, q , is always positive upward towards the surface of the earth). Therefore, sill intrusion has an immediate negative effect on the basement heat flow.

Sills have a significant effect on the thermal history of sediments adjacent to the intrusion (Fjeldskaar et al., 2008). However, with increasing distance from the sill, the thermal effect rapidly diminishes. The volcanic flow has limited thermal effect on the surrounding sediments at the well location 7316/5-1 (**Fig. 12a**; blue line, 3.7 km depth). The model hardly replicates the measured vitrinite reflectance trends. This suggests the number and thickness of sills and volcanic flows within the Vestbakken Volcanic Province is far more complex than modelled and/or the R_o measurements made on drilled cuttings from well 7316/5-1 are not entirely representative of the in-situ thermal maturity of the sediment column due to poor sample integrity. Note that the well 7316/5-1 is located near the well 7317/9-1 recently drilled further north (on the southern tip of the Stappen High). It penetrated shallow, though compacted, Jurassic-Triassic sandstones indicating uplift, and hence higher past temperatures.

6 Discussion

6.1 On the relevance of basin formation modelling.

The modelling of basin formation

The petroleum system concept embodies all the geological elements (source rock, reservoir rock and seal) and processes (hydrocarbon generation, migration and trap formation) required for effective oil and gas accumulation. Those elements and processes are controlled by the tectonic and thermal histories of the sedimentary basin which are in

turn controlled by the mechanism of basin formation. Hence, basin formation is a crucial boundary condition for petroleum systems.

However basin formation is in general not properly constrained. Basin formation and temperature history are generally constrained using backstripping. Yet backstripping introduces errors. Notably, in the case of multi-rift basins, early rift stretching factors are underestimated because thinning of sediment layers is not taken into account (Wangen and Faleide, 2008; Clark et al., 2014). Clark et al. (2014) applied both the backstripping and inverse/time-forward techniques along the PETROBAR-07 seismic transect. Their backstripping reconstruction predicts Late Carboniferous stretching factors much lower than for the inverse approach (almost 1 vs. 1.5 near the margin). This has implications on the thermal history and maturity of deeply-buried sediments. The backstripping reconstruction implies much less heat supply to deposited sediments at Late Carboniferous, hence a delayed maturation. If the basin reconstruction is biased this can give an erroneous exploration model (Cacace and Scheck-Wenderoth, 2016). It is therefore crucial to consistently model the formation and evolution of the sedimentary basin.

Near margin processes: breakup and magmatic underplating

Near-margin processes (Continental breakup, magmatic underplating and sill intrusion) have implications on the sediment thermal structure and maturity below the Vestbakken Volcanic Province.

Continental breakup causes a larger present-day thermal maturity of sediments approximately up to 70 km away from the COB. Breakup provides a lateral heat flow from the spreading ridge (as it moves along the sheared COB) directly into the adjacent sediments without heat dispersion through the continental crust and lowermost sediments within the Vestbakken Volcanic Province and is therefore measurable in the present-day thermal model through R_o trends.

Surprisingly, the emplacement of the magmatic underplate hardly changes the maturity of sediments at the margin. It even slightly reduces it! That is because the emplacement of the hot underplate is actually compensated by a thinner radiogenic crust (caused by higher levels of crustal stretching required in rift phases prior to 60 Ma) to maintain isostasy. Moreover, deeply buried Permian-Carboniferous limestones below Jurassic and Cretaceous shales have very low effective conductivities (1.2 W/m/K for limestones and 1.5 W/m/K for shales) such that they depress basement heat flow and thereby thermal conduction of the heat pulse from cooling of a magmatic underplate.

Modelling was performed assuming a conservative 5 km thickness for the underplate body. However, the interpretation by Libak et al, (2012) suggests the underplate body may be up to 10 km thick beneath the Vestbakken Volcanic Province. We have therefore tested a model whereby the underplate body is 10 km thick. The modelled Eocene heat flow peak, caused by the emplacement of the body, is much larger than for the 5 km thick model (model B2) (**Fig. 15**). The modelled heat flow is also larger during the Late Carboniferous rifting event, compared to model B2. This is because of the extra uplift caused by such a thick underplating, which must be compensated for in the model by

accentuated crustal stretching. However, thicker underplate hardly affects the modelled vitrinite reflectance near the margin.

6.2 Observed and computed stretching factors

Cumulative β -factors from models along the section B_long are compared with β -factors derived from observed crustal thickness (**Fig. 16**) inferred from velocity inversion studies (Clark et al., 2013, 2014). The observed values are based on assumed thinning of the crust relative to an idealized 35 km original crustal thickness.

The modelled cumulative β -factors are in general good agreement with the “observed” β -factors along the profile. The distribution of β -factors for the different models is identical east of the Vestbakken Volcanic Province. Modelled cumulative β -factors mostly agree with the work of Clark et al. (2014), peaking at approximately $\beta=4$ over the deep Bjørnøya Basin (**Fig. 16**). In the far east of the profile, around the Nordkapp Basin, the modelled β -factors are consistent with the crustal thickness estimates.

There are however a few differences. A magmatic underplate can be included as part of the crust within a velocity inversion, so any elevated levels of stretching required to accommodate the underplate may not be observable. So the elevated β -factors calculated for a modelled scenario incorporating breakup and underplating will be much higher than “observed” β -factors based on crustal thickness incorporating an underplate. In the area

between the Loppa High and Nordkapp Basin (around the Ottar Basin), modelled β -factors indicate a thinner-than-observed crust, suggesting a more complex mechanism than simple lithospheric thinning (Clark et al., 2014). Below the Ottar Basin, the lower crust shows almost no thinning (based on velocity modelling), yet up to 10 km of subsidence has been created to accommodate the overlying sediments (Breivik et al., 1995). Depth-dependent thinning, phase changes, or thermal downwelling in the mantle have been proposed as possible explanations for this discrepancy (Clark et al., 2014). Depth-dependent thinning could occur through asymmetric extension by simple shear mechanisms. This would require a detachment or series of detachments through the crust. Gernigon et al. (2014) proposed reactivation of crustal structures inherited from a Caledonian thrust within a pre-Permian extensional system across the Bjarmeland Platform (including the Ottar Basin). If these detachments facilitated further extension through the Permian, subsidence mechanisms would follow, creating the necessary accommodation space for thick Upper Carboniferous-Permian sedimentary successions in the Ottar Basin. The seismic resolution of Permian stratigraphy in the Bjarmeland Platform is poor (Glørstad-Clark et al., 2011a), and the lack of rift-related faulting noted in the Permian stratigraphy may be sub-seismic.

6.3 The Loppa High anomaly

The Loppa High is a remarkable elevated structure in the SW Barents Sea with a complex crustal structure and tectonic history which differ from the surrounding area. The interpretation of seismic, gravity and magnetic data suggests the presence of a ~100 km wide high-grade 'mafic' rock lower crustal body below the Loppa High (Ritzmann and Faleide, 2007). This dense body is supposed to originate from the Caledonian orogeny. The present day Loppa High is the result of several phases of uplift and subsidence. Its predecessor, the Selis Ridge (now located on the western side of the present day Loppa High) was uplifted in the Late Carboniferous to Middle Triassic (Riis et al., 1986; Wood et al., 1989; Gudlaugsson et al., 1998; Glørstad-Clark et al., 2010). The entire high then subsided in the Late Triassic forming a depocenter. In the earliest Cretaceous, a wider platform around the Selis Ridge became uplifted, causing the depocenter to form a sub-aerially exposed Loppa High (Wood et al., 1989; Gabrielsen et al., 1990; Faleide et al., 1993; Glørstad-Clark et al., 2011a; Indrevær et al., 2016). The uplift is estimated to have been of the order of 300 m (Indrevær et al., 2016). The latter event of uplift was contemporaneous with the onset of extreme lithospheric thinning in neighboring basins to the west. Erosion of the high and deposition of sediments along its flanks suggest gradual erosion and subsidence of the Loppa High in the Early Cretaceous, bringing the Loppa High to the same level as the wider Barents Sea shelf by the onset of the Late Cretaceous (Glørstad-Clark, 2011a). Gentle subsidence was interrupted by episodes of minor renewed uplift that affected the entire Barents Sea in Paleogene times (Vorren et al., 1991; Riis and Fjeldskaar, 1992; Nyland et al., 1992; Riis, 1996; Dimakis et al., 1998; Cavanagh et al., 2006; Green and Duddy 2010).

Our pure-extension model can roughly reproduce the vertical motions of the Loppa High from Late Paleozoic to present day, though amplitudes are exaggerated. The modelled vertical motions result from the flexural response of the basement to the different rifting episodes the area experienced since the Paleozoic. At the end of the Late Permian event, the basement uplifted by an ~800 meters. The Loppa High was then a depocenter until Late Jurassic. The high was again uplifted during the Late Jurassic rifting event by ~1800 m.

However, the pure-extension model fails to reproduce other observations. It requires a much thinner crust below the high to maintain isostasy. Hence, an extensional mechanism alone cannot explain the evolution of the Loppa High. Based on a thermal and petrological numerical model of continental lithosphere, Indrevær et al. (2018) proposed an alternative model for the tectonic evolution of the Loppa High. They propose that the phases of uplift and subsidence are the result of density changes accompanying phase transitions in the mafic body. The phase transitions result from pressure and temperature variations in the lower crust caused by geodynamic events that the SW Barents Sea experienced in Mesozoic times. These include: 1) Late Triassic far-field compression (associated with westward upthrusting of Novaya Zemlya) caused densification of the mafic body hence giving rise to subsidence and 2) The Early Cretaceous formation of the deep Tromsø and Bjørnøya basins provides large heat supply causing phase changes towards lighter material in the mafic body hence uplift. The phase of uplift was followed by subsidence driven by cooling and densification of the mafic body.

6.4 Petroleum systems

Several wells have recently been drilled in the SW Barents Sea. Some wells drilled in the vicinity of the Loppa High are successful oil and gas finds while others, drilled on the southern tip of the Stappen High are desperately dry. Well results are here interpreted in the light of our basin models.

Well 7220/11-1 – Alta Discovery

Located on the southern boundary of the Loppa High (see Fig. 1 for location), approximately 30 km south of the Loppa High profile (Glørstad-Clark et al., 2011a) is the 7220/11-1 Alta discovery. It is located very near the Loppa High profile (section A). With initial estimates of up to 300 million barrels of oil, this is a significant find in Permian reservoir carbonates (Ørn Formation/Gipsdalen Group; www.npd.no).

Oil was encountered at around 2 km depth within Ørn formation karstified carbonates (Middle Permian, 290-305 Ma). The trap appears to be a stratigraphic unconformity, with an Early Triassic shale acting as the seal (basal Kobbe Formation). Petroleum geochemistry studies on the 7120/2-1 well by Pestman et al. (2011) suggest the oil shows observed in Permian carbonates may derive from Ørret Formation (Tempelfjorden Group) source rocks, which date between approximately 250 and 258 Ma.

Our model A predicts that the Ørret Formation source rocks are within the oil window ($R_o = 0.5-1.5\%$) and that they could have charged the area with hydrocarbons since Triassic times until present day (**Fig 17a**).

However, remember that model A assumes the Loppa High evolution is driven by a series of rifting episodes. Notably, according to this model, the former High experienced a mild Late Permian rifting episode supplying heat to deposited sediments, hence accelerating the maturation of the Ørret formation source rocks. In contrast, according to the alternative metamorphic model of Indrevær et al. (2018), the Loppa High passively subsided in Triassic times owing to densification of the mafic body in the lower crust. According to this scenario, the Ørret formation source rocks should have then known a slightly colder evolution. However, the Ørret formation is relatively shallow (~4 km deep). The thermal evolution is therefore primarily controlled by the sediment burial history (Theissen and Rüpke, 2009; Marcussen et al., 2010) rather than by deep lithospheric processes (Gac et al., 2014). The maturity difference between the two scenarios is therefore expected to be negligible.

Well 7319/12-1 – Pingvin Technical discovery

The 7319/12-1 well discovered a 15 m gas column in reservoir sands belonging to the Early Paleocene Torsk Formation or Late Cretaceous Kveite Formation. The well is located on the eastern flank of the Bjørnøya Basin (**see Fig. 1 for location**), and although non-commercial, it is an active petroleum system. The discovery is located 19 km north

of the PETROBAR-07 profile (section B). The trap appears to lie within a Cretaceous-Paleocene anticline towards the eastern flank of the Bjørnøya Basin.

The highest quality source rocks studied in the western Barents Sea are Late Jurassic shales in the Hekkingen Formation. However, these source rocks are overmature in the Bjørnøya Basin before the Late Cretaceous-Paleocene trap is formed. Models B0-3 predict Early Aptian source rocks (123 Ma) are gas mature from approximately 100 Ma to 17 Ma on the eastern flank of the Bjørnøya Basin (**Fig. 17b**). Upon gas expulsion from the source rock interval, the gas could have migrated along the western boundary fault of the Bjørnøyrenna Fault Complex, and fed into the Late Cretaceous-Paleocene anticlinal trap. Shale in the Torsk Formation is the most attractive seal. Modelling suggests that the Late Cretaceous-Paleocene play in the Bjørnøya Basin is no longer active, since the source rocks have been overmature for the past 20 Ma (**Fig. 18b**).

Wells 7316/5-1 & 7317/9-1

A small dry gas discovery has been made at the calibration well 7316/5-1 (Knutsen et al., 2000). The drilled well has shown the presence of adequate reservoirs of Middle Eocene (Lutetian) age. However, the nearby well 7317/9-1 proved dry. This well was recently drilled a bit further north of well 7316/5-1, on the southern tip of the Stappen High (**see Fig. 1 for location**). The well penetrated shallow, though compacted, Late Triassic – Middle Jurassic sandstones.

Main source rocks in the VVP area are thought to be oil- and gas-prone Late Jurassic shales (Hekkingen Formation), and Cretaceous source rocks with significant proportions of organic material (Larssen et al., 2002). Model B3 predicts the Late Jurassic source rocks start to produce hydrocarbons at Early Cretaceous (blue line, **Fig. 17c**). They quickly become overmature following sill intrusion at Early Eocene, before the deposition of Eocene reservoir rocks at well 7316/5-1. However, the Cretaceous source rocks begin to generate oil and gas later, from Early Eocene onwards (red line, **Fig. 17c**). The small gas discovery in Eocene reservoir rocks at well 7316/5-1 could then originate from those Cretaceous source rocks.

Reservoir rocks at well 7317/9-1 were deposited at Late Triassic – Middle Jurassic, *i.e.* before generation of hydrocarbons in Late Jurassic and Cretaceous source rocks. However, the well is dry. One possible reason is erosion of trap rocks during Cenozoic uplift and erosion episodes in the Barents Sea (Dimakis et al., 1998; Green and Duddy, 2010; Henriksen et al., 2011; Laberg et al., 2012) as suggested by over-compaction of Triassic-Jurassic rocks at well 7319/9-1.

7 Conclusions

We have computed 2D thermo-kinematic models of basin evolution based on an inverse-scheme approach to simulate the thermal and tectonic history of SW Barents Sea basins along two seismic profiles. The isostatic and thermal effects of various near-margin

tectono-magmatic processes are evaluated. Those processes include continental breakup, magmatic underplating, and sill intrusion. The following conclusions are drawn:

1. The modelling of four rifting events (Late Carboniferous, Late Permian, Late Jurassic – Early Cretaceous and Paleocene – Eocene) satisfactorily reproduces the observed stratigraphy along the two profiles. This is an agreement with the work of Clark et al. (2014).
2. A 5 km thick magmatic underplate requires elevated levels of crustal stretching to compensate for the associated uplift in the Vestbakken Volcanic Province (VVP) during the Late Carboniferous and Late Permian rift phases.
3. Both continental breakup and magmatic underplating alter the thermal history of sediments within the VVP. Continental breakup has a permanent effect on the hydrocarbon maturation of source rocks in the VVP. Magmatic underplating has only a transient thermal effect, with no permanent change in hydrocarbon maturation indices.
4. Our modelling predicts Late Jurassic and Cretaceous source rocks were mature enough in the VVP area to generate oil. However, the well 7317/9-1 is dry, probably because Cenozoic erosion removed the traps.

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Figure captions

Fig. 1: Tectonic map of the south-western Barents Sea. The PETROBAR-07 profile (red line; Clark et al. 2014) ties-in with the HB-3-96A seismic reflection line near the COB, extending the profile westward to the Knipovich Ridge. The regional seismic line of Glørstad-Clark et al. (2011a) is shown as a blue line. Major faults are shown as black lines (Faleide et al. 2008). Calibration well locations are labelled in green and other wells in orange. BB: Bjørnøya Basin, HB: Harstad Basin, HfB: Hammerfest Basin, LH: Loppa High, NB: Nordkapp Basin, OB: Ottar Basin, SB: Sørvestnaget Basin, SH: Stappen High, TB: Tromsø Basin, VVP: Vestbakken Volcanic Province.

Fig. 2: Lithostratigraphic column and petroleum systems for the south-west Barents Sea. Modelled rift phases shown in far right column, spanning the duration of rifting in TecMod2D. Time scale after Gradstein et al. (2004).

Fig. 3: Seismic profiles crossing the SW Barents Sea (see location on figure 1). The upper figure a) shows the West-East seismic profile crossing the Tromsø basin, Polheim subplatform, Loppa High and Bjarmeland platform. The profile was depth-converted and interpreted by Glørstad-Clark et al. (2011a), following the velocity model of Clark et al. (2013). The lower figure (b) shows the NW-SE seismic profile crossing the SW Barents Sea and the margin. This section includes the PETROBAR-07 profile (Clark et al. 2014) tied-in to the HB-3-96A line to the Knipovich Ridge. The PETROBAR-07 profile was depth-converted and interpreted by Clark et al. (2013, 2014).

Fig. 4: Input stratigraphic sections. The figure shows a) the stratigraphic section from the Loppa High profile (Glørstad-Clark et al. 2011a) and b) the stratigraphic sections from the PETROBAR-07 profile (after Clark et al. 2014) integrated with the HB-3-96A interpretation in the oceanic domain toward the Knipovich Ridge. The COB is located at $x = 152$ km. Locations of pseudo-wells for each section are shown as dashed lines (a-j); wells within the vicinity of the profile are indicated by solid lines.

Fig. 5: Modelled cross sections along the Loppa High profile (section A) at different time steps from Middle Carboniferous to Present-day.

Fig. 6: Glacial erosion and crustal stretching factors (β -factors) calculated along the Loppa High profile (section A). The top graph (a) represents the modelled erosion and (b) the cumulative β -factors, while graphs c-f represent the β -factors calculated for the individual modelled rift phases. Note the maximum period of stretching over the deep Bjørnøya and Tromsø Basins occurred during the Late Jurassic – Early Cretaceous rift phase.

Fig. 7: Water depths over time for four pseudo-well locations (see figure 4a for pseudo-well locations a-d) along the Loppa High profile (section A). Each sketch shows the modelled paleo-water depths (black line) and interpreted paleo-water depths (dashed line, Glørstad-Clark et al. 2011a) for each pseudo-well. Grey shaded areas represent modelled rift phases.

Fig. 8: Quality control for the model along the Loppa High profile (section A). Modelled vitrinite reflectance R_o (left column) and temperature (right column) are plotted against measured vitrinite R_o and well temperature data for each well (a-d). For well location see figure 1.

944

945 **Fig. 9:** Modelled cross sections along the PETROBAR-07 profile (section B_short) at different time steps
 946 from Middle Carboniferous to Present-day.

947

948 **Fig. 10:** Glacial erosion and crustal stretching factors (β -factors) calculated along sections B_short and
 949 B_long (the PETROBAR-07 profile tied-in to the short HB-3-96A profile). The top graph (a) represents the
 950 modelled erosion and (b) the cumulative β -factors, while graphs c-f represent the β -factors calculated for
 951 the individual modelled rift phases. Note the maximum period of stretching over the deep Bjørnøya Basin
 952 occurred during the Late Jurassic – Early Cretaceous rift phase. The black line shows the modelled
 953 stretching factors for the reference model (B0). The red line shows the modelled stretching factors for the
 954 break-up only model (B1). The green line shows the modelled stretching factors for the break-up and
 955 underplating model (B2). The blue line shows the modelled stretching factors for the break-up,
 956 underplating and sills model (B3).

957

958 **Fig. 11:** Water depths over time for four pseudo-well locations (see figure 4b for pseudo-well locations a-d)
 959 along the PETROBAR-07 profile (section B_short). Each sketch shows the modelled paleo-water depths
 960 (black line) and interpreted paleo-water depths (dashed line, Clark et al. 2014) for each pseudo-well. Grey
 961 shaded areas represent modelled rift phases.

962

963 **Fig. 12:** Quality control for the model along the PETROBAR-07 profile (section B_short). Modelled
 964 vitrinite reflectance R_o (left column) and temperature (right column) are plotted against measured vitrinite
 965 R_o and well temperature data for each well (a-c). The figure a) shows the modelled vitrinite reflectance for
 966 model B1 (breakup only model, black line) and for model B3 (breakup, underplate ad sill intrusion model,
 967 blue line). For well location see figure 1.

968

969 **Fig. 13:** Water depths over time for three pseudo-well locations (see figure 4b for well and pseudo-well
 970 locations) located in the Vestbakken Volcanic province along section B_long (PETROBAR-07 profile tied
 971 in to the HB-3-96A short profile). Each sketch shows the modelled paleo-water depths and interpreted
 972 paleo-water depths (purple line, Clark et al. 2014) for each pseudo-well. The black line shows the modelled
 973 water depths for the reference model (B0). The red line shows the modelled water depths for the break-up
 974 only model (B1). The green line shows the modelled water depths for the break-up and underplating model
 975 (B2). The blue line shows the modelled water depths for the break-up, underplating and sills model (B3).
 976 Grey shaded areas represent modelled rift phases.

977

978 **Fig. 14:** Basement heat flow data versus time for three pseudo-well locations (see figure 4b for well and
 979 pseudo-well locations) located in the Vestbakken Volcanic Province at different distances from the COB
 980 along section B_long (PETROBAR-07 profile tied in to the HB-3-96A short profile). Each sketch shows
 981 the modelled basement heat flow for each pseudo-well. The black line shows the modelled basement heat
 982 flow for the reference model (B0). The red line shows the modelled basement heat flow for the break-up
 983 only model (B1). The green line shows the modelled basement heat flow for the break-up and underplating
 984 model (B2). The blue line shows the modelled basement heat flow for the break-up, underplating and sills
 985 model (B3). Grey shaded areas represent modelled rift phases.

986

987 **Fig. 15:** Modelled basement heat flow versus time at pseudo-well 'e' located at the Vestbakken Volcanic
 988 Province. Heat flow is modelled for a model with a 5 km-thick underplate (black line) and a model for a 10
 989 km-thick underplate.

990

Fig. 16: β -factors based on crustal thickness (black line) together with uncertainty in the velocity inversion results (dark grey swath), and local uncertainty of up to 5 km of high-velocity sedimentary rocks included in the uppermost crystalline crust (after Clark et al. 2013). These are overlain by modelled cumulative β -factors within this study for reference (black line, model B0), breakup (red line, model B1), breakup and underplate (green line, model B2) and breakup, underplate and sills (blue line, model B3).

Fig. 17: Modelled evolution of the vitrinite reflectance R_o (%) for a) the Ørret formation source rocks deposited at 253 Ma on the Loppa High, b) for Early Aptian source rocks deposited at 123 Ma in the Bjørnøya Basin and c) for upper Jurassic (blue line) and Cretaceous source rocks (red line) at the Vestbakken Volcanic Province.

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Lithology	Density (kg/m ³)	Thermal Exp. (1/K)	Rad. Heating (W/m ³)	Heat Capacity (J/kg/K)	Grain Cond. (W/m/K)	Porosity	Inv.Comp. Length (1/km)
Water	1000	0	0	1000	1000	0	0
Peridotite	3340	3.2E-05	0	1000	3.5	0	0
Oceanic crust (Magmatic bodies)	2900	2.4E-05	0	1000	2	0	0
U. crust (Granite)	2700	2.4E-05	2E-06	1000	3	0	0
L. crust (Diabase)	2900	2.4E-05	2E-06	1000	3	0	0
Sandstone	2690	0	1E-06	1000	4.4	0.4900	0.27
Limestone	2600	0	1E-06	1000	2	0.4173	0.40
Shale	2720	0	1E-06	1000	1.5	0.6200	0.50
10ss90sh	2717	0	1E-06	1000	1.6704	0.6200	0.49
20ss80sh	2714	0	1E-06	1000	1.8602	0.6000	0.46
30ss70sh	2711	0	1E-06	1000	2.0716	0.5900	0.44
40ss60sh	2708	0	1E-06	1000	2.3069	0.5700	0.41
50ss50sh	2705	0	1E-06	1000	2.5690	0.5600	0.39
60ss40sh	2702	0	1E-06	1000	2.8609	0.5500	0.37
Shale (kerogen- rich)	2720	0	1E-06	1000	1.5	0.6200	0.50

Table 1. Rock properties assigned to the basin infill within the basin models for sections A and B.

1

Rift Phase	Rift Interval (Sections B_short and B_long)	Rift Interval (Section A)
Late Carboniferous	320-300	320-305
Late Permian	270-251	271-251
Late Jurassic – Early Cretaceous	165-145	165-145
Paleocene – Early Eocene	65-55	65-55

2

3

4

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Table 2. Rift phases defined with the basin models for sections A and B.

1

Model Parameter	Value	Units
Forward Parameters		
Lithosphere thickness	120	km
Upper crust	17.5	km
Lower crust	17.5	km
Top and bottom temperature	5 and 1300	°C
e-fold length radiogenic heating (crust)	20	km
e-fold length correction	f(crust)	
Thermal conductivity (crust and mantle)	f(T)	W/m/K
Density (upper/lower crust and mantle)	2700/2900 and 3340	kgm ⁻³
Specific heat (upper/lower crust and mantle)	1000/1000 and 1000	J/kg/K
Thermal expansion co-efficient (crust and mantle)	2.4E-05 and 3.2E-05	K ⁻¹
Radiogenic heat (crust)	2E-06	W/m ³
Matrix conductivity	f(T)	W/m/K
Pore fluid conductivity	f(T)	W/m/K
Effective conductivity	Geometric	W/m/K
Numerical resolution of finite element mesh (nx/ny)	200/100	
Effective elastic thickness (T_E)	5	km
T_e Boundary Conditions – Left (section B only)	0	km
T_e Boundary Conditions – Right (section B only)	5	km
Necking depth	15	km

2

3

4

Table 3. Input physical forward parameters and inversion control parameters used in this study. These parameters are the same for all modelled scenarios.

FIGURE 1: 1.5 column

Sedimentary basins

Late Cretaceous - Palaeocene
Late Jurassic - Early Cretaceous
Late Palaeozoic

Wells

1	7316/5-1
2	7220/6-1
3	7120/2-1
4	7226/11-1
5	7224/7-1
6	7125/1-1
7	7121/1-1R
8	7220/11-1
9	7319/12-1
10	7317/9-1

Fig. 1: Tectonic map of the south-western Barents Sea. The PETROBAR-07 profile (red line; Clark et al. 2014) ties-in with the HB-3-96A seismic reflection line near the COB, extending the profile westward to the Knipovich Ridge. The regional seismic line of Glørstad-Clark et al. (2011a) is shown as a blue line. Major faults are shown as black lines (Faleide et al. 2008). Calibration well locations are labelled in green and other wells in orange. BB Bjørnøya Basin, HB: Harstad Basin, HfB: Hammerfest Basin, LH: Loppa High, NB: Nordkapp Basin, OB: Otta Basin, SB: Sørvestnaget Basin, SH: Stappen High, TB: Tromsø Basin, VVP: Vestbakken Volcanic Province.

FIGURE 10: 1.5 column

Fig. 10: Glacial erosion and crustal stretching factors (β -factors) calculated along sections B_short and B_long (the PETROBAR-07 profile tied-in to the short HB-3-96A profile). The top graph (a) represents the modelled erosion and (b) the cumulative β -factors, while graphs c-f represent the β -factors calculated for the individual modelled rift phases. Note the maximum period of stretching over the deep Bjørnøya Basin occurred during the Late Jurassic – Early Cretaceous rift phase. The black line shows the modelled stretching factors for the reference model (B0). The red line shows the modelled stretching factors for the break-up only model (B1). The green line shows the modelled stretching factors for the break-up and underplating model (B2). The blue line shows the modelled stretching factors for the break-up, underplating and sills model (B3).

Fig. 11: Water depths over time for four pseudo-well locations (see figure 4b for pseudo-well locations a-d) along the PETROBAR-07 profile (section B_short). Each sketch shows the modelled paleo-water depths (black line) and interpreted paleo-water depths (dashed line, Clark et al. 2014) for each pseudo-well. Grey shaded areas represent modelled rift phases.

Fig. 13: Water depths over time for three pseudo-well locations (see figure 4b for well and pseudo-well locations) located in the Vestbakken Volcanic province along section B_long (PETROBAR-07 profile tied in to the HB-3-96A short profile). Each sketch shows the modelled paleo-water depths and interpreted paleo-water depths (purple line, Clark et al. 2014) for each pseudo-well. The black line shows the modelled water depths for the reference model (B0). The red line shows the modelled water depths for the break-up only model (B1). The green line shows the modelled water depths for the break-up and underplating model (B2). The blue line shows the modelled water depths for the break-up underplating and sills model (B3). Grey shaded areas represent modelled rift phases.

Fig. 14: Basement heat flow data versus time for three pseudo-well locations (see figure 4b for well and pseudo-well locations) located in the Vestbakken Volcanic Province at different distances from the COB along section B_{long} (PETROBAR-07 profile tied in to the HB-3-96A short profile). Each sketch shows the modelled basement heat flow for each pseudo-well. The black line shows the modelled basement heat flow for the reference model (B0). The red line shows the modelled basement heat flow for the break-up only model (B1). The green line shows the modelled basement heat flow for the break-up and underplating model (B2). The blue line shows the modelled basement heat flow for the break-up underplating and sills model (B3). Grey shaded areas represent modelled rift phases.

Fig. 16: β -factors based on crustal thickness (black line) together with uncertainty in the velocity inversion results (dark grey swath), and local uncertainty of up to 5 km of high-velocity sedimentary rocks included in the uppermost crystalline crust (after Clark et al. 2013). These are overlain by modelled cumulative β -factors within this study for reference (black line, model B0), breakup (red line, model B1), breakup and underplate (green line, model B2) and breakup, underplate and sills (blue line, model B3).

FIGURE 2: 2 column

Fig. 2: Lithostratigraphic column and petroleum systems for the south-west Barents Sea. Modelled rift phases shown in far right column, spanning the duration of rifting in TecMod. Time scale after Gradstein et al., (2004).

Fig. 3: Seismic profiles crossing the SW Barents Sea (see location on figure 1). The upper figure a) shows the West-East seismic profile crossing the Tromsø basin, Polheim subplatform, Loppa High and Bjarmeland platform. The profile was depth-converted and interpreted by Glørstad-Clark et al. (2011a), following the velocity model of Clark et al. (2013). The lower figure (b) shows the SW-NE seismic profile crossing the SW Barents Sea and the margin. This section includes the PETROBAR-07 profile (Clark et al. 2014) tied-in to the HB-3-96A line to the Knipovich Ridge. The PETROBAR-07 profile was depth-converted and interpreted by Clark et al. (2013, 2014).

FIGURE 4: 2 column

Fig. 4: Input stratigraphic sections. The figure shows a) the stratigraphic section from the Loppa High profile (Glørstad-Clark et al. 2011a) and b) the stratigraphic sections from the PETROBAR-07 profile (after Clark et al. 2014) integrated with the HB-3-96A interpretation in the oceanic domain toward the Knipovich Ridge. The COB is located at x = 152 km. Locations of pseudo-wells for each section are shown as dashed lines (a-j); wells within the vicinity of the profile are indicated by solid lines.

FIGURE 5: 1.5 column

Fig. 5: Modelled cross sections along the Loppa High profile (section A) at different time steps from Middle Carboniferous to Present-day.

FIGURE 6: 1.5 column

ACCEPTED MANUSCRIPT

Fig. 6: Glacial erosion and crustal stretching factors (β -factors) calculated along the Loppa High profile (section A). The top graph (a) represents the modelled erosion and (b) the cumulative β -factors, while graphs c-f represent the β -factors calculated for the individual modelled rift phases. Note the maximum period of stretching over the deep Bjørnøya and Tromsø Basins occurred during the Late Jurassic – Early Cretaceous rift phase.

FIGURE 7: 1.5 column

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Fig. 7: Water depths over time for four pseudo-well locations (see figure 4a for pseudo-well locations a-d) along the Loppa High profile (section A). Each sketch shows the modelled paleo-water depths (black line) and interpreted paleo-water depths (dashed line, Glørstad-Clark et al. 2011a) for each pseudo-well. Grey shaded areas represent modelled rift phases.

FIGURE 9: 1.5 column

Fig. 9: Modelled cross sections along the PETROBAR-07 profile (section B_short) at different time steps from Middle Carboniferous to Present-day.

FIGURE 12: 1.5 column

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Fig. 12: Quality control for the model along the PETROBAR-07 profile (section B_short). Modelled vitrinite reflectance R_o (left column) and temperature (right column) are plotted against measured vitrinite R_o and well temperature data for each well (a-c). The figure a) shows the modelled vitrinite reflectance for model B1 (breakup only model, black line) and for model B3 (breakup, underplate ad sill intrusion model, blue line). For well location see figure 1.

Fig. 15: Modelled basement heat flow versus time at pseudo-well 'e' located at the Vestbakken Volcanic Province. Heat flow is modelled for a model with a 5 km-thick underplate (black line) and a model for a 10 km-thick underplate.

Fig. 17: Modelled evolution of the vitrinite reflectance R_o (%) for a) the Ørret formation source rocks deposited at 253 Ma on the Loppa High, b) for Early Aptian source rocks deposited at 123 Ma in the Bjørnøya Basin and c) for upper Jurassic (blue line) and Cretaceous source rocks (red line) at the Vestbakken Volcanic Province.

FIGURE 8: 1.5 column

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Fig. 8: Quality control for the model along the Loppa High profile (section A). Modeled vitrinite reflectance R_o (left column) and temperature (right column) are plotted against measured vitrinite R_o and well temperature data for each well (a-d). For well location see figure 1.

1 **Highlights**

- 2 • The tectono-thermal evolution of SW Barents Sea sedimentary basins is modelled.
- 3 • The effects of near-margin processes on maturity of source rocks are simulated.
- 4 • Continental breakup has a permanent effect on the maturation of source rocks.
- 5 • Magmatic underplating has a transient effect on the maturation of source rocks.