Crustal scale subsidence and uplift caused by metamorphic phase changes in the lower crust: A model for the evolution of the Loppa High area, SW Barents Sea from late Palaeozoic to Present

Kjetil Indrevær*, Sebastien Gac, Roy H. Gabrielsen & Jan Inge Faleide

The Research Centre for Arctic Petroleum Exploration (ARCEx), Department of Geosciences, Sem Sælands vei 1, 0371 Oslo, University of Oslo, Norway

*Correspondence (kjetil.indrevar@geo.uio.no)

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Abstract

The Loppa High area has been subject to several events of uplift and subsidence from the Late Palaeozoic to Present. The driving mechanisms behind the vertical movements, however, are not fully understood. We propose that uplift and subsidence were influenced by the combination of density changes caused by metamorphic phase changes in a 90x140 km wide mafic lower crustal body below the high and local (rift-related) and far-field stress. Through a numerical modeling approach we analyze the tectonically induced variations in pressure and temperature in the lower crust, its influence on phase changes in the mafic body and the affiliated vertical movements. Results show that i) densification of the mafic body caused by far-field compression associated with the late Triassic westward translation of Novaya Zemlya could cause surface subsidence, ii) heat and fluid influx provided by early Cretaceous rifting could trigger density reduction and surface uplift and iii) the present day geometry of the Loppa High as observed in seismic data can be reproduced by combining the modeled effect of rift flank uplift and phase changes in the mafic body. Phase change-driven vertical movements may also have affected other structural highs in the western Barents Sea, including the Stappen High.

The Loppa High is located in the southwestern Barents Sea. It is bordered by the late Jurassic-early Cretaceous Hammerfest Basin to the south and the very deep, mainly early Cretaceous, Tromsø and Bjørnøya basins to its west **(Figure 1;** Gabrielsen et al. 1990**)**. The Loppa High has a multi-stage tectonic history and is influenced by several phases of uplift and subsidence: Its core and predecessor, the Selis Ridge **(Figure 1 & 2a),** became gradually uplifted and eastward tilted in the late Carboniferous to middle Triassic (mainly mid-late Permian) (Riis et al. 1986; Wood et al. 1989; Gudlaugsson et al. 1998; Glørstad-Clark 2011). This event was succeeded by late Triassic subsidence of a wider area to form a sediment depocenter on top of the former high. Contemporaneously with the onset of accelerated lithospheric thinning in the neighboring Tromsø and Bjørnøya basins in the early Barremian, the depocenter was uplifted to form a sub-aerially exposed platform corresponding to the present day Loppa High (Wood et al. 1989; Gabrielsen et al. 1990; Faleide et al. 1993a,b; Glørstad-Clark 2011; Indrevær et al. 2017). Minor renewed uplift likely took place in Paleogene times as a part of regional uplift and tilting (Vorren et al. 1991; Riis & Fjeldskaar, 1992; Nyland et al. 1992; Riis, 1996; Dimakis et al. 1998; Cavanagh et al. 2006; Green & Duddy 2010).

The driving mechanisms behind the repeated uplift and subsidence of the Loppa High remain poorly understood. Previous work have suggested that the late Carboniferous to middle Triassic uplift of the Selis Ridge reflects rift-related footwall uplift along major extensional faults along the western flank of the Selis Ridge (Ziegler, 1978, Wood et al. 1989; Johansen 1994; Gudlaugsson et al. 1998) perhaps associated with depth-dependent extension (Glørstad-Clark 2011). Furthermore, it has been suggested that subsequent cooling of the thinned lithospheric mantle explains the late Triassic subsidence and the formation of the depocenter as a part of a post-rift sag-basin (Clark et al. 2014) and that the early Cretaceous uplift of the Loppa High was a direct consequence of accelerated lithospheric thinning in the Tromsø and Bjørnøya basins to the west (Glørstad-Clark 2011; Indrevær et al. 2017). This hypothesis was further strengthened by Indrevær et al. (2017) who linked the presence of early Barremian to early Aptian/middle Albian tectonic inversion structures along the flanks of the Loppa High to a distinct event of uplift of the high in this period. The uplift appears to be successfully modelled by lithospheric stretching and flexural isostasy (Clark et al. 2014), although the rift-flank-perpendicular wavelength of uplift of the Loppa High (>100 km) is greater than that typically associated with rift flank uplift for crustal extension of such magnitude ($\beta \approx 3$) (see Roberts & Yielding 1991; Kusznir & Ziegler 1992; Gabrielsen et al. 2005; Huismans & Beaumont 2005; Henk 2006) and therefore partly fails to explain the observations.

The mechanism for the formation of the Selis Ridge as caused by rift-related footwall uplift as suggested above appears to be kinematically valid. However, if the late Triassic subsidence and early Cretaceous uplift was related to cooling and heating of a rift flank, subsidence and uplift of similar amplitude and wavelength would be expected in the Hammerfest Basin located along the same rift flank just south of the Loppa High. This is not observed in the seismic record (cf. Gabrielsen 1984; Berglund et al. 1986; Gabrielsen et al. 1990) and indicates that additional mechanism(s) were involved in the evolution of the Loppa High.

We suggest that in the full analysis of the Loppa High, the presence of a ~90x140 km large body of assumed mafic composition in the lower crust below the high as indicated by magnetic, gravity, and seismic data must be taken into consideration. The geometry of this body is evident from the interpretation of reflection seismics and its lateral extent is constrained using gravity and magnetic data (Figure 1, Ritzmann & Faleide, 2007; Ebbing & Olesen 2010; Barrère et al. 2011; Gernigon et al. 2012; Marello et al. 2013; Clark et al. 2014). The body was probably emplaced during the Caledonian Orogeny (Ritzmann & Faleide, 2007) as the eastern margin of the Loppa High and Hammerfest Basin is believed to overlap the Caledonian suture zone (Torsvik & Cocks 2017). Depending on local conditions of pressure and temperature, such mafic bodies may be subject to metamorphic phase changes and associated density changes (Semprich et al, 2010). Compression may cause tectonic overpressure leading to densification and events associated with lithosphere heating (such as rifting or a plume) may supply enough heat to trigger phase changes to produce lighter mineral assemblages (Cloetingh & Kooi, 1992; Burov & Cloetingh, 2009; Cloetingh & Burov, 2011). Such density changes may in turn cause vertical surface deformation (Kaus et al. 2005; Semprich et al. 2010; Gac et al. 2013, 2014).

As the Palaeozoic to present geological evolution of the Barents Sea is associated with both orogenic events and rifting, metamorphic phase changes in the assumed mafic body below the Loppa High may have contributed to the repeated vertical motions of the high. This paper focuses on evaluating the effects of such phase changes in the mafic body located below the Loppa High. The model considers the effects of tectonically induced changes in pressure and temperature. We notably investigate whether or not Permian to late Triassic far-field east-directed contraction associated with the Polar Urals and thrusting of Novaya Zemlya **(Figure 1)** (e.g. Buiter & Torsvik 2007) may have caused densification of the mafic body and surface subsidence. Second, thermo-kinematic models of basin formation are applied in evaluating the effects of thermal uplift of the Loppa High as linked to the early Cretaceous rifting event and examine whether or not it is likely that this event generated sufficient heat to lower the density of the mafic body and hence cause additional uplift. We test the validity of the results by evaluating if the model can reproduce key geometries associated with the present day Loppa High as observed from seismic data.

2 Geological setting

The main structural elements surrounding the *Loppa High* that are relevant for the present work include the Hammerfest Basin in the south and the Tromsø and Bjørnøya basins to the west. These basins are shortly characterised in the following.

The Hammerfest Basin is separated from the Loppa High by the E-W-striking extensional top-to-the-south Asterias Fault Complex **(Figure 1)**. This basin is delimited in the south by the north- to northwest-dipping Troms-Finnmark Fault Complex and by the Ringvassøy-Loppa Fault Complex in the west that is characterised by a down-stepping array of normal faults to the deeper Tromsø Basin (Gabrielsen, 1984). To the east, the Hammerfest Basin gradually shallows and flexes to become the Bjarmeland Platform (Gabrielsen 1984; Gabrielsen et al. 1990). The Hammerfest Basin was subject to extension throughout the Carboniferous-Eocene, but particularly owes its present configuration to fault activity in the Late Jurassic to earliest Cretaceous. Fault activity in the Hammerfest Basin was interrupted in the early Barremian as fault activity focused to the Bjørnøya and Tromsø basins and the main subsidence of these basins (Rønnevik et al. 1982; Gabrielsen 1984; Berglund et al. 1986; Gabrielsen et al. 1990; Faleide et al. 1993b; 2008; Clark et al. 2014; Indrevær et al 2017).

The Bjørnøya and Tromsø basins **(Figure 1)** were initiated during Carboniferous and Permian-early Triassic rifting. Both basins are influenced by halokinesis affiliated with the late Carboniferous-early Permian evaporites. Late Jurassic - earliest Cretaceous extension was followed by accelerating subsidence and accumulation of very thick sediment sequences in the early Cretaceous as demonstrated by the down-faulting of Jurassic sediments to ~13 km depth in the Bjørnøya Basin across Bjørnøyrenna Fault Complex (Rønnevik et al. 1982; Gabrielsen et al. 1990; Faleide et al. 1993b; 2008; Clark et al. 2014). The axis defined by the Ringvassøy-Loppa and Bjørnøyrenna fault complexes marks the position of a major basement-involved, Caledonian zone of weakness (Rønnevik et al. 1982; Gabrielsen 1984; Gabrielsen et al. 1990; Faleide et al. 1993a,b; 2008; Ritzmann & Faleide 2007; Torsvik & Cocks 2017) and may explain why extension became focused in this zone.

2.1 Events of contraction

The SW Barents Sea has been exposed to several events of contraction or uplift in the Paleozoic - Mesozoic. Far-field events relevant to the present analysis include the evolution of the northernmost part of the Uralian Orogeny (Polar Urals) and thrusting in the Novaya Zemlya-region. These events were linked to the closure of the Uralian Ocean that eventually developed into a continent-continent collision between Baltica and Siberian cratons (Torsvik & Cocks 2017). The collision started in the sout in the Carboniferous and migrated northwards to reach the Barents Sea area in the middle Permian to early Triassic (Zonenshain et al. 1990; Puchkov, 1997, 2009; Brown & Echtler 2005; Gee et al. 2006). Novaya Zemlya was upthrusted in late Triassic – early Jurassic times (Otto & Bailey 1995; Bogatsky et al. 1996; Nikishin et al. 1996; Torsvik & Andersen 2002; Ritzmann & Faleide 2009) and is believed to mark the northwards (and delayed) continuation of the Polar Urals. Previous geodynamic modeling and plate reconstruction (Buiter & Torsvik, 2007; Torsvik & Cocks 2017) suggest Novaya Zemlya translated westward by ~100 km as a part of the up-thrusting. The tectonic events relevant for the present paper and their relative timing are summarised in Figure 3.

3 Geological constraints for the numerical model

Several key observations made in the seismic reflection data need to be reproduced by the numerical model for the model to be deemed valid. Such observations were therefore used to calibrate model parameters and to evaluate model results. They include:

1. A distinct zone of gradual thickening of the upper Triassic strata towards the (palaeo)-basin center defines a sag-basin-like depression that resulted from late Triassic subsidence in the Loppa High area **(Figure 2a)**. Estimates on the amount of sediment thickening yield a thickness increase of ~1 km (corrected for later erosion and assuming 4 km/s velocity in late Triassic sediments) onto the present day Loppa High. The zone of thickening is concentric in map view and defines the southern and eastern limit of the late Triassic basin area **(Figure 1)**.

2. The early Cretaceous event of uplift, which elevated the area that at present defines the Loppa High, was contemporaneous with the onset of accelerated lithospheric thinning in the Tromsø and Bjørnøya basins initiating in the early Barremian (Indrevær et al. 2017). The eastern boundary of the uplifted area coincides with a monocline affecting the Triassic and Jurassic sequences **(Figure 2a)**. This structure also marks the zone of thickening of the upper Triassic sequence, suggesting that the eastern and southern boundary of the area that subsided in the late Triassic corresponds to the area that was later uplifted in the early Cretaceous **(Figures 1 & 2a)**. The renewed uplift had a rift-perpendicular wavelength of ~100 km and affected an area that extended ~150 km parallel to the rift axis.

3. The lateral extent of the mafic body located at the base of the Loppa High (Ritzmann & Faleide, 2007) coincides well with the lateral dimensions of the present-day Loppa High and thus also with the area that subsided in the late Triassic **(Figure 1 & 2)**.

4. Evidence of early Cretaceous uplift also exists to the south of the Loppa High, suggested by the westward thinning and subsequent thickening of the lower Barremian – Aptian strata along the western rim of the Hammerfest Basin **(Figure 2b)**. This is believed to represent a short-lived event of uplift occurred in the early Barremian. This event affected a ~40 km wide zone along the rift flank, which stands in strong contrast to the width of the area uplifted to form the Loppa High (> 100 km).

Although both values of wavelengths are within an expected range as caused by the mechanism of rift flank uplift (e.g. Kooi et al. 1992; van der Beek et al. 1994, 1995), it is the contrasting amount of uplift between the neighbouring Loppa High and Hammerfest Basin that strongly suggest the influence of additional uplift mechanism(s) for the Loppa High. We assume that the response of the Hammerfest Basin reflects the true effect of early Cretaceous rift flank uplift (thermal and isostatic) and that the Loppa High experienced rift flank uplift of similar wavelength due to early Cretaceous rifting. We accordingly use the response of the western flank of the Hammerfest Basin to calibrate the modeled rift flank uplift input parameters (including sedimentation, erosion, compaction, crust- and mantle thinning, heat transfer and flexural isostasy) for the Loppa High in order to determine the component of uplift of the high that can be attributed to rift flank uplift alone.

5. Density change in the mafic body could result from pressure and temperature changes caused by tectonic processes. For the late Triassic subsidence of the Loppa High to be caused by phase changes in the mafic body, an increase in pressure (or change in temperature) is required. Based on the contemporaneous timing, the Polar Urals and/or the westward up-thrusting of Novaya Zemlya in the east (**Figure 3**) are the only candidates for an enhanced stress situation.

To summarise, the conspicuous spatial correlation between the area subject to vertical movements through time and the lateral extent of the mafic body at depth strongly suggests an influencing role of the mafic body in the evolution of the Loppa High and is the main motivation for the modeling done in the present work. Furthermore, model also needs to reproduce the contrasting wavelength of uplift between the Loppa High and the Hammerfest basin for the model to be deemed valid.

4 Modeling

The modeling was performed in three stages:

1) First, a 2D plan-view elastic model covering the Barents Sea was generated to estimate the order-of magnitude of the potential compressional stresses that could affect the Loppa High area at lower crustal levels as a result of contractional events in the eastern Barents Sea (**Figure 4a**).
2) Thereafter the effects of the early Cretaceous extension in the southwestern Barents Sea and associated basin formation and rift flank uplift were simulated using the Tecmod2d thermo-kinematic modeling tool (**Figure 4b**).
3) The modeled pressures and temperatures from the above models were used as input in a phase change model to calculate expected changes in density in a mafic body located below the Loppa High **(Figure 4c)**. The densities were used as input to a 2D density-isostasy model, which modeled vertical movements at the surface as a result of phase changes in a mafic body at depth.

The different model set-ups, results and correlation with observations from seismic data are described in detail below.

4.1 Model M1 – Late Triassic compression modeling

To cover the sequential development of the Loppa High, the potential for transferring stress generated by west-vergent Carboniferous to late Triassic shortening from the eastern- to the western Barents Sea at lower crustal level was first modeled. In lack of better constraints on the interplay between the Polar Urals and Novaya Zemlya, we modeled the effect of the late Triassic thrusting of Novaya Zemlya only. For simplicity, we assumed that the Barents Sea crust is dividable into an upper and lower crust, with its yield strength determined by the mechanical properties of quartz and plagioclase (diorite), respectively. A dioritic lower crust can sustain differential stress up to 1000 MPa (Cloetingh & Burov 1996) thus leaving any lower differential stress to result in elastic deformation only. Since we are solely interested in stresses affecting the Loppa High at lower crustal levels, we thus used a purely elastic plate stress model to estimate the order of magnitude of potential horizontal compressive stresses propagating through the lower crust from Novaya Zemlya to the Loppa High at the time of subsidence (model M1, **Figure 5**).

The map-view model dimension was set to 1000 x 1000 km, covering the entire Barents Sea **(Figure 4a).** The model used a Young's Modulus of 10 GPa and a Poisson ratio of 0.25. The values and composition used as input parameters for the model are all geologically reasonable and the model results are considered to represent a "best guess" estimate of lower crustal stress at the time of subsidence. We acknowledge that our model is a simplification as the Barents Sea region has an heterogenous and asymmetric lithospheric structure characterised by a thick, cold and stronger lithosphere in the eastern Barents Sea (including Novaya Zemlya) and a thinner, hotter and weaker lithosphere in the western Barents Sea (including the Loppa High area, Klitzke et al. 2015; Gac et al. 2016). The contrasting lithosphere thickness must have influenced how compressive stress originating from Novaya Zemlya propagated westward through the lithosphere but is no taken into account in the modeling presented herein.

Previous geodynamic modeling and plate reconstruction (Buiter & Torsvik, 2007) suggest Novaya Zemlya was translated westward by 100 km to its present location, with no later movement (total shortening is maintained). This is consistent with 100 km of westward contraction imposed on the east side of the model corresponding to the present-day shape of Novaya Zemlya.

The model shows that compressional stresses in the lower crust dissipate westward as the stresses radiate from the apex of stress at Novaya Zemlya. The model indicates that the present day Loppa High area experienced increased horizontal compressional stress ($\sigma_H = \sigma_1$) causing a differential stress of $\sigma_H - \sigma_V \approx 300$ MPa as a result of shortening caused by the eastward translation of Novaya Zemlya (**Figure 5)**. If the lithostatic pressure component is taken into account, the total horizontal stress amounts to ~1.3 GPa at Moho level.

Model M2 – Early Cretaceous extension

In this model, the effect of the onset of rifting in the Tromsø and Bjørnøya basins in the early Barremian was simulated in order to evaluate the modeled and observed amplitude and wavelength of rift flank uplift and heat influx as a result of early Cretaceous rifting **(Figure**

6a). We use “rift flank uplift” as a term that includes the effect of crust- and mantle thinning, thermal heating and flexural isostasy in the continued text.

Model M2 utilised the modeling software Tecmod2D **(Figure 4b),** which models lithospheric extension applying thermo-kinematic principles in a 2D-section parallel to extension (Rüpke et al. 2008) in pure shear (McKenzie, 1978; Jarvis & McKenzie 1980. The Tecmod2D’s forward model explores both lithosphere-scale (crust- and mantle thinning, heat transfer and flexural isostasy) and basin-scale (sedimentation, erosion and compaction) processes simultaneously. Crust and mantle lithosphere locally extend for a finite duration during the rift phase. The amount of extension is defined by thinning (β) factors for crust and mantle. The rift phase is characterised by basin formation and upwelling of hot asthenosphere. Extension is followed by a post-rift phase marked by cooling of the thermal anomaly supplying additional post-rift subsidence (McKenzie, 1978). Tecmod2d computes the subsidence during rifting and post-rift phase so that isostatic equilibrium is maintained throughout the simulation. The extensional modeling takes into account sedimentation and thermal blanketing effect of sediments.

The stratigraphic record in the Tromsø Basin indicates that accelerated early Cretaceous rifting took place from early Barremian to early Cenomanian (Indrevær et al 2017). The present thickness of the compacted post-Jurassic strata is ~12 km. A crustal stretching factor of ~3 is necessary to isostatically compensate this sediment pile. We therefore assume that the crust and mantle lithosphere thinned by a factor 3 in a time span of 30 Myr. Extension rate was set constant throughout the rifting phase. The crustal stretching factor was set to maximum at the center of the model laterally decreasing from the center outwards over a distance of 50 km following a sine curve. The model was calibrated to reproduce uplift wavelengths similar to that experienced by the Hammerfest Basin.

The model shows that rifting would cause an 8 km deep Tromsø Basin and a rift-flank uplift with an amplitude in the order of 200 m along the western flank of the Loppa High, progressively decreasing away from the basin over a distance of ~50 km (**Figure 6a**). This was followed by thermal relaxation giving an additional post-rift subsidence of ~4 km in the Tromsø Basin (which corresponds to the observed post-rift sediment thickness). At present, the effect of thermal flank uplift has largely receded (**Figure 6a**).

It is particularly emphasised that the amount of rift flank uplift (as modeled to mirror the actual effect of uplift of the western flank of the Hammerfest Basin) does not manage to reproduce the observed wavelength of early Cretaceous uplift as experienced by the Loppa High. Additional mechanism(s) must therefor be added in order to explain the present geometry of the high.

Model M3 - The phase change model

Model M3 was run to assess the effect of tectonically induced changes in pressure and temperature as obtained from model M1 and M2. A phase change model was coupled with a density-isostasy model in order to model vertical motions at the surface as a function of changing densities in the mafic body at depth **(Figure 4c).**

The density of mafic rocks was estimated from its assumed composition, calculated pressure (assuming a lithostatic reference state of stress) and temperature-dependent phase change model. The phase change model is computed with the Perple_X software (e.g. Connolly, 2005). It is based on Gibbs free energy minimization, which gives the proportion, compositions and thermodynamic properties of stable phases as a function of pressure and temperature. From the amounts and the densities of the phases predicted, the bulk rock density can be calculated. The thermodynamic calculations assume phase equilibrium.

Calculated densities were used as input in a 2D density-isostasy model of continental lithosphere to compute the vertical motions caused by metamorphic phase changes in the lower crustal mafic body **(Figure 4c)**. Computations were performed in a 500 km wide and 120 km thick continental lithosphere E-W-section crossing the Loppa High and the Tromsø Basin. The model consisted of two horizontal layers: a 36 km thick crust overlying an 84 km thick mantle lithosphere. Those values are consistent with the average calculated depths of Moho and lithosphere-asthenosphere boundary (LAB) in the southwestern Barents Sea (Klitzke et al. 2015). In the model, a mafic body was positioned in the crust. The dimensions of the mafic body were determined from free-air gravitational, reflection seismic and magnetic data (Ritzmann & Faleide 2007) and are set to be 100 km wide and 10 km thick in the model. The top of the body was set at a depth of 26 km and its base at 36 km deep (Moho level).

Acknowledging the uncertainties affiliated with the mafic body beneath the Loppa High, we made several assumptions in the modeling. These are discussed in the following.

Composition of the mafic body: We modeled the pressure-temperature-dependent density of the mafic body assuming a wet mafic gabbroic composition characterised by low SiO_2 content (Rudnick & Fountain 1995; Semprich et al. 2010) (**Figure 4c**). Minor differences in assumed composition would, however, result in considerable variations in calculated densities within the mafic body at given changes in pressure and temperature (e.g. Connolly 2005).

Fluids: The efficiency of phase changes and the density of the resulting mineral assemblage depend strongly on the presence or influx of fluids. In our modeling, we assumed that phase changes occurred during events of prograde metamorphism and during retrograde metamorphism only in the case where fluid influx was likely (e.g. during rifting). We further assumed that the amount of fluids during phase changes was kept constant and that phase changes were efficient and instantaneous throughout the mafic body.

Volume changes: Density changes in mafic rocks during metamorphism are generally accompanied by changes in rock volume. The effect of volume changes on the modeled amplitudes of uplift and subsidence are in the present work calculated to be in the order of 10 meters or less, even when considering extreme cases where volume loss/gain is accommodated along the vertical axis only. As these values are below the sensitivity of the phase change model, we assume a constant volume in the modeling.

Sedimentation and erosion would modify pressure and temperature in the lower crust through loading/unloading and thermal blanketing. This, in turn, would alter the density of the mafic body and cause additional isostatic adjustments.

Although these mechanisms certainly have had a significant influence on the absolute amplitudes of uplift/subsidence, they would only amplify the effects of uplift/subsidence related to phase changes in the mafic body, and thereby not influence the general trends in the modeling. Sedimentation and erosion would likely not affect the wavelength of phase change-induced uplift/subsidence as the wavelength is primarily controlled by the lateral extent of the mafic body. Hence, we do not take sedimentation/erosion into account in the phase change model.

To summarise, uncertainties related to composition, presence of fluids, volume changes and sedimentation/erosion as discussed above would all amplify (or not significantly limit) the amplitudes of phase change-induced uplift or subsidence. In this sense, the presented absolute values of densities and *amplitudes* of uplift/subsidence can be considered tentative only, although the general trends suggested by the modeling would not be affected. The modeled *wavelengths* of uplift/subsidence associated with phase changes, however, would not be significantly influenced by the uncertainties mentioned above. We thus consider the model to be valid for the purpose of modeling contrasting wavelengths of uplift and subsidence in the Loppa High area, as is key to the present analysis.

A modeled lithostatic pressure of ~0.98 GPa and a temperature of ~489°C at 36 km depth (base of the mafic body) was estimated for the situation prior to late Triassic contraction (**Figure 6b**). These values correspond to an average density of 3178 kg.m^{-3} for a wet mafic rock modeled to be present below the Loppa High.

Model M3 suggests that the late Triassic compression affiliated with the thrusting of Novaya Zemlya (Model M1) increased the pressure at Moho level to 1.3 GPa, triggered prograde metamorphism in the mafic body. The computed average density in the body increased to 3194 kg.m^{-3} causing ~200 m subsidence in the area situated above the mafic body (**Figure 6b**). The model suggests that the subsidence generated a basin-dipping monocline that formed atop the outer boundary of the mafic body at depth (**Figure 6b**). The modeled basin depth (~200 m) is less than the observed depth from the seismic data (~1 km). However, the effect of sediment loading is not taken into account, an effect that would have deepened the basin further. The modeled depocentre remained deep throughout the Jurassic, in harmony with that observed in the seismic data**.**

The effect of heat and fluid influx associated with early Cretaceous mantle thinning and rifting accompanying the formation of the deep Tromsø and Bjørnøya basins was modeled by introducing a second phase change in the mafic body (**Figure 6b**). According to the model, the pressure was reduced to lithostatic due to the onset of E-W extension. The modeled Moho temperature increased from 489°C to 514°C on the western side of the Loppa High. This caused prograde metamorphism in the western half of the mafic body. Heating in the eastern half of the mafic body was very limited, but we assume that fluid influx associated with extension and basin formation did also affect the eastern half of the mafic body promoting retrograde phase changes due to the pressure decrease. The average density of the mafic body was 3160 kg.m^{-3} at the termination of rifting, with a more pronounced reduction in density in the western part of the Loppa High (3139 kg.m^{-3}) compared to the eastern part (3182 kg.m^{-3}). According to the model, the phase changes in the mafic body inverted the late Triassic basin infill, uplifting the upper Triassic sequence ~200 m above the pre-subsidence baseline (top Permian, **Figure 6b)**. Because of the additional heating of the western part of the mafic body, the western part was uplifted an additional ~100 m close to the rift flank. The uplift inverted the late Triassic monocline resulting in a spatial overlap between the late Triassic zone of thickening, the outer boundary of the mafic body at depth and the newly formed monocline facing away from and defining the outer boundary of the newly formed high. The spatial overlap between the three matches what is observed in seismic data **(Figure 1 & 2a)**. As later fault activity related to the opening of the North Atlantic ocean localised further west (e.g. Faleide et al. 2008), we assume that no later retrograde metamorphism occurred in the mafic body until present, thus preserving the lighter mineral assemblages and maintaining the Loppa High as a positive structure.

For reference, the combined effect of rift flank uplift and phase changes are given in **Figure 6c.**

Discussion

For the total model for the development of the Loppa High to be considered successful, it must reproduce the main geological observations as seen in the present Loppa High area and its vicinity. Our modeling results show that phase changes caused by pressure and temperature variations due to tectonic processes (i.e. late Triassic compression from Novaya Zemlya and early Cretaceous rift-induced heat and fluid influx) are sufficient to cause vertical motions of a magnitude and lateral distribution to that seen in the reflection seismic data. The model thus satisfactorily explains the spatial overlap between the subsided area in the late Triassic, the uplifted area in the early Cretaceous and the lateral extent of the mafic body at depth **(Figure 6)**. Comparing the modeled and the observed wavelength of early Cretaceous uplift as experienced by the Loppa High (**Figure 7**), it is evident that rift flank uplift as a response to accelerated extension in the Tromsø and Bjørnøya basins alone cannot explain the observed wavelength of uplift **(Figure 7a).** Rift-flank uplift has definitely affected the Loppa

High in the early Cretaceous in a similar fashion to that observed at the western rim of the Hammerfest Basin (**Figure 7b**). However, it is only by adding the effect of phase changes in the mafic body that the model manage to reproduce values of wavelength for uplift in accordance with the geometry of the present day Loppa High. It is emphasised that the modeled effect of rift flank uplift as calibrated from the response of the western rim of the Hammerfest Basin corresponds well to the area affected by Carboniferous – middle Triassic formation of the Selis Ridge, supporting previous work that suggested the Selis Ridge formed through the mechanism of rift flank uplift along the present day Ringvassøy-Loppa and Bjørnøyrenna fault complexes (Wood et al. 1989; Glørstad-Clark 2011; Clark et al. 2014).

A conceptual model of the Late Palaeozoic – present evolution of the Loppa High area

Based on the above model results and discussion and previously published observations and analyses, we present a unified conceptual model for the evolution of the Loppa High from late Palaeozoic to present-day (**Figure 8**). The emplacement of the mafic body was probably associated with the Silurian-Devonian Caledonian Orogeny, and it can be speculated that it represents the lower part of the suture zone separating rocks of Laurentian and Baltican origin (Ritzmann & Faleide, 2007; Gernigon et al. 2012; Torsvik & Cocks, 2017) and comprised of subducted gabbroic oceanic crust. Subsequent orogenic collapse, moderate extension and regional subsidence transformed the Barents Sea into a shallow epicontinental sea characterised by large evaporitic basins by the Carboniferous and Permian (Gabrielsen et al. 1990; Faleide et al. 1993a,b) **(Figure 8, I-III)**. The Carboniferous to middle Triassic (mainly mid-late Permian) stages of formation of the Tromsø and Bjørnøya basins was accompanied by uplift and rotation of the Selis Ridge in the Loppa High area, generating an elongated, ~100 km long ridge stretching along strike of the rift flank with a rift-perpendicular wavelength of ~40 km **(Figure 8, II)**. We have not included in full the formation of the Selis Ridge in our present model approach, but the width of the Selis Ridge fits well with the expected wavelength of rift flank uplift for the early Cretaceous phase, supporting that the formation of the Selis Ridge is indeed a result of this mechanism as suggested by others (Wood et al. 1989; Glørstad-Clark 2011; Clark et al. 2014).

In the Late Triassic, the Loppa High area subsided and formed a depocenter with a southern and eastern boundary corresponding to a zone of thickening of the late Triassic sequence (**Figure 1 & 2**) (Glørstad-Clark et al. 2010). This event of subsidence has been suggested by Clark et al. (2014) to be linked to a subsequent cooling of the thinned lithospheric mantle associated with late Carboniferous to middle Triassic rifting. However, if this was the dominating mechanism, the depocenter should be expected not to be limited to the present day Loppa High, but also include the Hammerfest Basin located along the same rift flank, which is not the case (**Figure 1**). According to our model results, we thus suggest that subsidence was influenced by prograde phase changes in the mafic body at the base of the Loppa High and that these phase changes were promoted by increasing compressional stress in the Barents Sea related to the reported ~100 km westward migration of Novaya Zemlya (Buiter & Torsvik 2007) **(Figure 8, IV)**. Due to tectonic quiescence in the Jurassic, the modeled depocenter remained deep throughout the Jurassic, which corresponds to observations (**Figure 8, V**).

In the early Cretaceous, accelerated thinning of the lithosphere resulted in the deepening of the Tromsø and Bjørnøya basins west of the Loppa High (Faleide et al., 1993a,b; Glørstad-Clark 2011; Clark et al. 2014). The upwelling of the asthenosphere supplied heat and fluids to the base if the Loppa High and, according to our modeling, caused rift flank uplift of the western flank of the Loppa High with a wavelength of ~40 km that was superimposed on a phase change-driven uplift with a wavelength of ~100 km caused by a transition to lighter mineral assemblages in the mafic body at depth (**Figure 7a)**. The uplift inverted the late Triassic depocentre as the Loppa High, outlined to the south and east by a monocline (**Figure 8, VI**).

From the middle Cretaceous to the Present **(Figure 8, VII-IX)**, subsequent cooling following early Cretaceous rifting caused a post-rift basin development west of the Loppa High, overstepping the rims of the syn-rift Tromsø and Bjørnøya basins. Gentle subsidence was interrupted by episodes of minor renewed uplift that affected the entire Barents Sea in the Paleogene (Vorren et al. 1991; Riis and Fjeldskaar, 1992; Nyland et al. 1992; Riis, 1996; Dimakis et al. 1998; Cavanagh et al. 2006). At present, the effect of thermal rift-flank uplift has largely receded due to cooling, leaving behind only uplift as the effect of the early Cretaceous phase change in the mafic body.

Similar effects of phase change-driven vertical movements may also be valid for other structural highs in the Barents Sea. One example may be the Stappen High, which is located northwest of the present study area. It is positioned along a left-stepping segment of the same rift axis as the Loppa High, and is flanked to the west by the Vestbakken Volcanic Province, similar in depths to the Tromsø and Bjørnøya basins (Gabrielsen et al. 1990) (**Figure 1**). The southern part of the Stappen High is characterised by magnetic and gravity anomalies comparable to that seen beneath the Loppa High (Skilbrei et al. 2001; Ritzmann & Faleide 2007; Gernigon et al. 2014) and shows a similar structural evolution: The Stappen High experienced late Carboniferous – early Permian uplift, subsidence in the Triassic and renewed uplift in the early Cenozoic (Wood et al. 1989; Gabrielsen et al. 1990; Faleide et al. 1993a,b; Worsley et al. 2001; Blaich et al. 2012, 2017). This renders the possibility that the Stappen High experienced similar effects of lower crustal phase changes as a part of its tectonic history. The fact that the latter phase of uplift of the Stappen High post-dates the early Cretaceous event of uplift of the Loppa High strengthens this hypothesis, as the main phase of formation Vestbakken Volcanic Province was in the Eocene (e.g. Faleide et al. 1993, 2008; Ryseth et al. 2003). This indicates that the early Cenozoic uplift of the Stappen High was linked to heat and fluid influx associated with basin formation in the west, in a similar fashion that the early Cretaceous uplift of the Loppa High was associated with contemporaneous rifting in the Tromsø and Bjørnøya basins.

7. Conclusions

The Loppa High has been subject to several events of subsidence and uplift as is reflected by the complex geometry of the high in the seismic record. Using a forward thermo-mechanical modeling approach coupled with a phase change model for mafic rocks we propose a new model of evolution for the Loppa High from late Palaeozoic to present day.

We propose that the evolution of the Loppa High area is strongly influenced by changes in density as a result of phase changes in a mafic body located at the base of the Loppa High. Our model results show that late Triassic far-field compression caused by the westward translation of Novaya Zemlya in the eastern Barents Sea likely contributed to densification of the mafic body leading to subsidence and the formation of a depocentre in the Loppa High area. Further, the results show that early Cretaceous rift activity could provide an influx of heat and fluids to the mafic body causing phase changes towards lighter mineral assemblages and subsequent uplift of the high.

The model successfully reproduces wavelengths of repeated uplift and subsidence as observed in the seismic data, including the spatial correlation between late Triassic depocentre, the area that became uplifted to form the early Cretaceous Loppa High and the lateral extent of the mafic body at depth. Furthermore, the model explains the contrasting wavelengths of early

Cretaceous uplift of the Loppa High (~100 km) compared to uplift of the western rim of the Hammerfest Basin (~40 km). We conclude that these relationships cannot be reproduced by modeling the effect of thermal and isostatic rift flank uplift mechanisms alone.

Similar effects of phase change-driven vertical movements may also be valid for other structural highs in the Barents Sea and beyond. One candidate is the Stappen High, located northwest of the Loppa High, which shows a similar structural evolution and magnetic and gravity anomalies comparable to that of the Loppa High.

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Figure captions:

Figure 1: Overview of the study area showing the main structural elements in the study area. Location of key seismic lines used in the paper is given in addition to the location of a zone of thickening of upper Triassic strata and an early Cretaceous monocline (see legend for details).

Colors show the magnetic anomaly pattern in the study area interpreted to reflect the presence of a mafic body at the base of the Loppa High (modified from Ritzmann & Faleide 2007). Note how the zone of thickening, the position and geometry of the monocline and the southern and eastern extent of the magnetic anomaly overlaps. Structural element map modified from Norwegian Petroleum Directorate (npd.no). FC, fault complex.

Figure 2: **a)** Interpreted seismic line running from the Bjarmeland Platform in the east, across the Loppa High, the Ringvassøy-Loppa Fault Complex and into the Tromsø Basin (see Figure 1 for location). Note the distinct thickening of upper Triassic strata from the Bjarmeland Platform and onto the Loppa High and the presence of an early Cretaceous monocline in the same area. **b)** Interpreted seismic line running from the Hammerfest Basin, across the Ringvassøy-Loppa Fault Complex and into the Tromsø Basin (see Figure 1 for location). Note the westward thinning and subsequent thickening of the lower Barremian to lower Aptian strata that is interpreted to be the result of a short-lived rift-flank uplift of the western margin of the Hammerfest Basin in the early Barremian. One-arm arrows indicate onlaps.

Figure 3: Timeline summarizing tectonic events and their timing in relation to vertical movements in the Loppa High area. The period of formation of the Selis Ridge is characterised by regional extension in the western Barents Sea and contractional events (Polar Urals) in the eastern Barents Sea. The late Triassic subsidence is coeval with the western translation of Novaya Zemlya, while the early Cretaceous uplift correlates in time with accelerated lithosphere thinning associated with the main phase of formation of the neighboring Tromsø and Bjørnøya basins.

Figure 4: Schematic overview of the different modeling steps used in the present work. **a)** A 2D plan-view elastic model is computed to assess compressional stresses at mid-crustal level as caused by the westward migration Novaya Zemlya. **b)** An early Cretaceous extension phase and associated basin formation is simulated using the Tecmod2d thermo-kinematic modeling tool. Parameters are calibrated to reproduce rift flank uplift as observed in along the western margin of the Hammerfest Basin **c)** The output of expected pressures and temperatures from a) and b) respectively, are used as input to a phase change model that gives the density of rocks with a wet mafic composition as a function of pressure and temperature (modified from Semprich et al. 2010). A 2D density-isostasy model calculates vertical movements at the surface as an effect of phase changes in the mafic body at depth.

Figure 5: Model M1 - Plan-view model showing the mid-crustal lithospheric stress in the Barents Sea as an effect of the up-thrusting of Novaya Zemlya in the late Triassic. **a)** Situation prior to thrusting. **b)** Same model showing the increase in average horizontal lithosphere stress in the Barents Sea after applying 100 km of shortening simulating the up-thrusting of Novaya Zemlya. The modeled increase in horizontal stress in the Loppa High area is ~300 MPa.

Figure 6: Results of the 2D thermo-kinematic extension model (model M2) and the phase change and 2D density-isostasy model (model M3) showing the effect of tectonic-induced pressure and temperature changes to the Loppa High area in the late Triassic, early Cretaceous and at Present. Note that each figure shows the modeled topography on the eastern half of the models (from basin center to the east side of the model) as we are solely interested in the geological evolution of the eastern margin of the Tromsø Basin. **a)** Modeled early Cretaceous rifting in the Tromsø Basin caused a thermal rift flank uplift with an amplitude of ~200 m and a wavelength of ~50 km. The effect is largely receded at present day. **b)** The late Triassic

compressional event as induced by the westward thrusting of Novaya Zemlya caused according to the model a densification of the mafic body that resulted in ~200 m of subsidence at the surface. Note that the width of the modeled basin corresponds to the width of the mafic body at depth. Early Cretaceous heat- and fluid influx allowed for a reduction in density in the mafic body. The eastern part of the Loppa High was uplifted back to the pre-subsidence base level (due to the pressure reducing back to its lithostatic component. The western part of the Loppa High saw additional phase change-driven uplift due to increased heating. Note that a monocline has formed at the surface above the eastern limit of the mafic body. As we assume that no further retrograde phase changes occurred, the effect of early Cretaceous phase changes is preserved to present day in the mafic body. **c)** The combined effect of thermal rift flank uplift and phase changes. At present, the effect of thermal rift flank uplift has largely receded, and the remaining net uplift is mainly due to the preservation of a lighter mineral assemblage in the mafic body.

Figure 7: Comparison of observed and modeled uplift wavelengths. **a)** The modeled effect of rift flank uplift alone cannot reproduce the early Cretaceous uplift of the Loppa High with respect to wavelength. It is only by adding the modeled effect of uplift caused by phase changes in the mafic body that the geometry of the present day Loppa High is successfully reproduced. Seismic profile is from Figure 2a. Note that the models have been stretched to account for the orientation of the composite seismic profile. **b)** Wavelength of modeled thermal rift flank uplift as a result of accelerated lithospheric thinning in the Tromsø Basin. Because the response of the western rim of the Hammerfest Basin has been used to calibrate the thermal rift flank uplift model, the model result correlates well (profile from Figure 2b). Amplitudes of modeled profiles are arbitrary.

Figure 8: Conceptual model of the late Palaeozoic to present day evolution of the Loppa High area. **I, II and III:** From Late Carboniferous to middle Triassic, the southwestern Barents Sea was subject to moderate extension and subsidence, rift flank uplift and rotation of the Selis Ridge and early stages of formation of the Tromsø and Bjørnøya basins. **IV and V:** The late Triassic ~100 km westward migration of Novaya Zemlya caused a build-up of compressional stresses and subsequent phase changes in the mafic body below the Loppa High. This again caused subsidence and the formation of a depocenter at the surface with similar lateral extent as the mafic body at depth. **VI:** Heat from early Cretaceous accelerated thinning of the lithosphere in the Tromsø and Bjørnøya basins caused thermal rift flank uplift and uplift as a result of phase changes in the mafic body. This uplifted the late Triassic depocenter to form the subaerially exposed Loppa High. A monocline formed along the eastern and southern boundary of the Loppa High, coinciding with the eastern extent of the mafic body at depth. **VII, VIII and IX:** Throughout Cretaceous until present, cooling caused a partial, gentle subsidence of the Loppa High and sagging subsidence in the Tromsø and Bjørnøya basins, interrupted by episodes of uplift in the Paleogene.