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Tectonophysics 387 (2004) 217–228

TECTONOPHYSICS

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Crustal thickening in an extensional regime: application to the mid-Norwegian Vøring margin

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Received 19 November 2003; accepted 12 July 2004

Abstract

The structure of the mid-Norwegian volcanic Vøring margin at the onset of the Maastrichtian–Paleocene extension phase reflects the cumulative effect of earlier consecutive rifting events. Lateral structural differences present on the margin at that time are a consequence of migration of the location of maximum extension in time between Norway and Greenland. The most important imprints (Moho depth, thermal structure) of these events on the lithosphere are incorporated in a numerical simulation of the final extension phase. We focus on a possible mechanism of formation of the Vøring Marginal High and address the relationship between spatial and temporal evolution of crustal thinning and thickening, uplift of the surface and strength of the lithosphere.

It is found that the Vøring Basin formed the strongest part of the margin which explains why the Maastrichtian–Paleocene rift axis was not located here but instead jumped westward with respect to the earlier rift axes locations. The modeling study predicts that local crustal thickening during extension can be expected when large lateral thermal variations are present in the lithosphere at the onset of extension. Negative buoyancy induced by lateral temperature differences increases downwelling adjacent to the rifting zone; convergence of material at the particular part of the margin is mainly taken up by the lower crust. The model shows that during the final phase of extension, the crust in the Vøring Marginal High area was thickened and the surface uplifted. It is likely that this dynamic process and the effects of magmatic intrusions both acted in concert to form the Marginal High.

Keywords: Lithosphere extension; Norwegian margin; Numerical model; Passive margin

1. Introduction

The Vøring margin is a volcanic passive margin, situated off Mid-Norway, NE Atlantic (e.g., Skogseid and Eldholm, 1987; Eldholm et al., 1989; Planke et al., 1991; Blystad et al., 1995) (Fig. 1). Before continental breakup took place between mid-Norway and East Greenland in the Early Eocene, at ~54 Ma (e.g.,

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doi:10.1016/j.tecto.2004.07.049

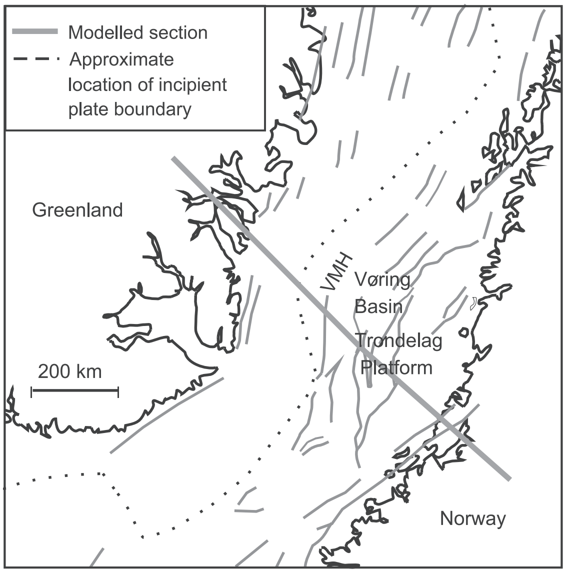


Fig. 1. Reconstruction of study area just prior to continental breakup, and distribution of major extensional structures (gray lines) (Price et al., 1997). The thick gray line denotes the modeled section. VMH is Vøring Marginal High.

Bukovics and Ziegler, 1985; Skogseid et al., 2000; Mosar et al., 2002), the area was affected by multiple extensional episodes since the end of the Caledonian orogeny (Bukovics and Ziegler, 1985; Ziegler, 1989; Ziegler and Cloetingh, 2004; Skogseid et al., 1992a; Hinz et al., 1993; Andersen, 1998; Mosar, 2003). This resulted in the formation of deep sedimentary basins positioned between the continent–ocean boundary and the Norwegian mainland and development of small basins in western Norway. Excessive magma production during a short (3 My, Skogseid et al., 1992b) period prior to rift–drift transition resulted in emplacement of magmatic rocks in the outer margin. The Early Tertiary magmatic event has frequently been linked to the Iceland mantle plume (e.g., Eldholm et al., 1989; Skogseid et al., 2000). Recent thermomechanical modeling studies show that the thermal influence of a hotspot is not required to explain the large volumes of melt generated (Anderson, 2000; Van Wijk et al., 2001).

The Jurassic–Cretaceous Vøring Basin in the offshore part (Fig. 1) is a deep sedimentary basin with maximum sediment thicknesses of probably ~6–10 km (e.g., Bukovics and Ziegler, 1985; Brekke, 2000; Osmundsen et al., 2002) and maximum crustal thinning factors of about 2.6 (Vågnes et al., 1998; Reemst and Cloetingh, 2000). Maximum crustal thinning factors of 1.6 are found for the Permo-Triassic basin developed on the Trøndelag Platform (e.g., Reemst and Cloetingh, 2000). The locus of highest extension on the mid-Norwegian margin moved westwards through time (Brekke, 2000; Reemst and Cloetingh, 2000; Walker et al., 1997), ultimately resulting in drifting; the breakup axis was located on the Greenland side of the extensional structures (Fig. 1).

The Vøring Marginal High is located west of the Vøring Basin (Eldholm et al., 1989). It became a distinct structural unit of the margin when it was separated from the Vøring Basin by vertical movements along the Vøring Escarpment (Blystad et al., 1995). The thickened crust of the Vøring Marginal High is overlain by seaward dipping reflector wedges that consist of basalt flows with sediments interbedded (e.g., Planke, 1994; Planke and Eldholm, 1994). The dipping wedges are suggested to be formed in shallow water conditions just prior to breakup or during the initial stage of oceanic crust formation (White, 1988; Viereck et al., 1989; Desprairies and Laloy, 1989; Eldholm et al., 1989; Berndt et al., 2001; Callot et al., 2001). Suggested explanations for the thickened crust of the Vøring Marginal High include a larger volume of intrusions close to the continent–ocean boundary (Mjelde et al., 2001) and a jump of the axis of maximum extension of the final extensional phase (Mjelde et al., 2001; Van Wijk and Cloetingh, 2002). Tectonic subsidence curves show uplift around the time of rift–drift transition on the western side of the Vøring Basin and the Vøring Marginal High (Skogseid et al., 1992a). Suggested explanations for these structures include the addition of magmatic material to the crust and dynamic support of the Iceland mantle plume (e.g., Clift et al., 1995), ridge push forces (e.g., Doré and Lundin, 1996), differences in spreading rates and the accretion of oceanic lithosphere (Mosar et al., 2002) and basin migration-related uplift (Van Wijk and Cloetingh, 2002). Below the western part of the Vøring Basin and the Marginal High, an up to 7-km-thick (Mjelde et al., 2001) high-velocity lower crustal body is observed. This body was proposed to be of magmatic origin or heavily intruded lower crustal material, formed during the final phase of extension (e.g., White, 1988; Coffin and Eldholm, 1994; Mjelde et al., 2002). In an alternative explanation, the high-velocity body is attributed to inherited high-pressure granulite/eclogite rocks or pre-breakup mafic material (Gernigon et al., 2003, 2004). The location and assumed timing of formation suggest a causal relation between the thickened crust of the Vøring Marginal High and the multiple extensional events on the margin (e.g., Skogseid and Eldholm, 1988; Planke et al., 1991; Mjelde et al., 2002; Mosar et al., 2002).

Crustal thinning and tectonic subsidence of the volcanic Mid-Norway margin have been extensively studied (e.g., Grunnaleite and Gabrielsen, 1995; Walker et al., 1997; Ren et al., 1998; Reemst and Cloetingh, 2000). However, with rift models that did not include rheology or buoyancy forces, it appeared to be difficult to sufficiently explain the tectonic subsidence and uplift history and crustal thickness of the Vøring Marginal High (Ren et al., 1998). By using a numerical model for simulation of the final extension phase of the mid-Norwegian margin that includes temperature-dependent rheology and buoyancy forces, we investigate tectonic subsidence and crustal thickness of the Marginal High area in relation to structural and thermal imprints on the margin resulting from earlier extension phases. The relationship between the spatial and temporal evolution of crustal thinning and thickening, uplift of the surface and strength of the lithosphere and the timing and amount of melting is addressed. The focus of this study is a possible mechanism of formation of the Vøring Marginal High, of which the thickened crust and uplift history are still a matter of debate. The results of the simulations suggest a somewhat more complex deformation pattern than is predicted by previous modeling studies. A dynamical explanation is found in the model for crustal thickening during extension, and the fact that the predicted pattern of deformation resembles the situation at the Vøring margin rather well suggests that formation of the Marginal High may well have been influenced by lateral variations in lithosphere configuration present at the onset of the final extensional phase.

2. Modeling approach

The Maastrichtian–Paleocene phase of extension prior to continental breakup (~67–54 Ma) is modeled using a numerical model describing lithosphere deformation. The model is two-dimensional and each solution represents a vertical section of the lithosphere.

2.1. Equations describing lithosphere deformation and thermal evolution

Approximately the upper half of the thermal lithosphere behaves elastically on geological time scales, while in the lower half, stresses are relaxed by viscous deformation. This viscoelastic behavior of the lithosphere is described by a Maxwell body (Turcotte and Schubert, 2002), resulting in the following constitutive equation for a Maxwell viscoelastic material:

$$\mathrm{d}\varepsilon/\mathrm{d}t = \sigma/2\mu + \mathrm{d}\sigma/E\mathrm{d}t \tag{1}$$

in which $\mathrm{d}\varepsilon/\mathrm{d}t$ is strain rate, μ is dynamic viscosity, σ is stress and E is Young's modulus (Turcotte and Schubert, 2002). For a Newtonian fluid, the dynamic viscosity μ is constant. In the lithosphere, however, nonlinear creep processes prevail (Ranalli, 1995; Carter and Tsenn, 1987), and the relation between stress and strain rate can be described by

$$e = A\sigma^n \exp(-Q/RT) \tag{2}$$

where A, n and Q are experimentally derived material constants (Ranalli, 1995; Carter and Tsenn, 1987), n is the power law exponent, Q is activation energy, R is the gas constant and T is temperature.

The state of the stress field is constrained by the force balance:

$$\partial\sigma_{ij}/\partial x_j + \rho g_i = 0 \tag{3}$$

where g is gravity and ρ is density. In this model, it is assumed that mass is conserved and the material is incompressible. The continuity equation following from the principle of mass conservation for an incompressible medium is:

$$\mathrm{div}\ v = 0 \tag{4}$$

Only thermal buoyancy is considered in the model, and density is dependent on the temperature following a linear equation of state:

$$\rho = \rho_0(1 - \alpha T) \tag{5}$$

where ρ_0 is the density at the surface, α is the thermal expansion coefficient and T is temperature. Buoyancy in the mantle arises also from the presence of melt (melt-retention buoyancy) and mantle depletion (e.g., Scott and Stevenson, 1989; Sotin and Parmentier, 1989). Extraction of melt results in accumulation of a depleted mantle residue, which is buoyant relative to undepleted mantle. The amount of melt present in the mantle during rifting is probably very limited (~3%, McKenzie and Bickle, 1988) and its influence on the density is neglected in this study.

Processes of fracture and plastic flow play an important role in deformation of the lithosphere. This deformation mechanism is active when deviatoric stresses reach a critical stress level. In this model, the Mohr–Coulomb criterion is used as yield criterion to define the critical stress level:

$$|\tau_n| \leq c - \sigma_n \tan\varphi \qquad (6)$$

where τ_n is the shear stress component, σ_n is normal stress component, c is cohesion of the material and φ is the angle of internal friction (Vermeer and De Borst, 1984). Stresses are adjusted every time step when the criterion is reached. Frictional sliding and fault movement are not explicitly described in the model. The displacement field is obtained by solving Eqs. (1)–(6).

The temperature field of the lithosphere is calculated every time step using the heat flow equation:

$$\rho c_{\mathrm{p}} \mathrm{d}T/\mathrm{d}t = \partial_j k \partial_j T + H \qquad (7)$$

where the density ρ is described by Eq. (5), c_{p} is specific heat, k is conductivity and H is heat production in the crust. The thermal and mechanical parts are coupled through the temperature-dependent power law rheology and buoyancy forces. Temperatures are calculated on the same grid as the velocity field, and advection is accounted for by the nodal displacements. Melt generation due to decompressional partial melting is calculated using the empirically derived expressions for dry mantle peridotite of McKenzie and Bickle (1988). This approach allows for variation in depth and degree of partial melting. Melt volumes are calculated assuming that all melt migrates vertically and either underplates or intrudes the crust. The emplaced melt generated during one time step is stretched during subsequent time steps, and the resulting cumulative melt thickness is calculated following the approach of Pedersen and Ro (1992). Thus, predicted melt volumes should be considered first-order estimations. Eqs. (1)–(7) are solved with the finite element method using 2560 straight-sided triangular elements. The Lagrangian formulation is used, and the finite element grid is periodically remeshed.

2.2. Lithosphere configuration and initial and boundary conditions

The location of the section that was modeled is indicated in Fig. 1. Fig. 1 shows the situation just prior to breakup, which is the situation at the end of the extensional phase that we modeled. The initial lithosphere configuration adopted for the modeling is shown in Fig. 2. The initial thickness of the lithosphere (at 67 Ma) was assumed to be 125 km, in accordance with the present thickness of the lithosphere of the west Norway Atlantic coastline (Suhadolc et al., 1990; Oleson et al., 2002; Skilbrei et al., 2002). The top surface of the upper crust is

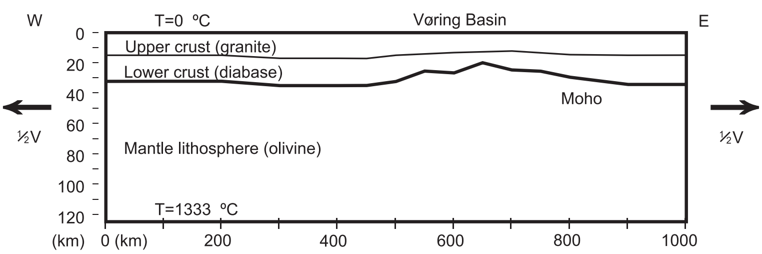


Fig. 2. Model setup. West (W) is Greenland side, East (E) is Norwegian side. The restored Moho topography below the Vøring Basin is taken from Skogseid et al. (2000). V is total extension velocity. Top surface of the upper crust is placed at sea level (z=0 km); the sediment infill of the basins and water depth are not included in the simulations.

placed at sea level (z=0 km), so there is no sediment infill of the Vøring and Trøndelag Basins. Numerical studies showed that surface processes may affect evolution of a rift basin and can change the strength of the lithosphere (Burov and Cloetingh, 1997; Burov and Poliakov, 2001). Due to computational intensity, sedimentation and erosion are not included in our simulations. A first-order approximation of the Moho configuration at the beginning of the Maastrichtian–Paleocene rift phase of the Vøring Basin was taken from Skogseid et al. (2000), who restored the Moho to the situation before this rift phase. On both sides of the Vøring Basin, however, such reconstructions of the Moho are not available; therefore, a simple Moho topography was adopted there. The lithosphere was extended with a total extension velocity of 25 mm/year. Reconstructed spreading rates just after breakup were slightly lower (~20 mm/year, Mosar et al., 2002). This resulted in continental breakup after about 13 My of stretching. Continental breakup was defined to occur when the maximum crustal thinning factor reached 20, which corresponded to extension of the model domain of about 25–30%. Thermal and initial boundary conditions included a zero heat flow through the sides of the domain and a constant temperature at the surface and the base of the model (Fig. 2). Thinning of the Vøring Basin ceased before the final extension phase, modeled in this study, began. In order to investigate whether the Jurassic–Cretaceous extension phase that formed the Vøring Basin (Doré et al., 1999) still resulted in a thermal anomaly at the onset of the final Maastrichtian–Paleocene rift phase on the base of the model domain, a test was performed in which the Vøring Basin formation thinning was simulated. We found no significant (i.e. >5 °C) temperature variations at the onset of the Maastrichtian–Paleocene rift phase deep in the lithosphere, which justifies the initial constant thermal boundary conditions selected. Although anomalously high mantle temperatures caused by the proto-Iceland mantle plume have frequently been suggested for this area and period of time (e.g., Pedersen and Skogseid, 1989; White and McKenzie, 1989), normal values were adopted here (Van Wijk et al., 2001). Parameters that were used are listed in Table 1. The effect of various lithosphere rheologies on the rift evolution was not studied here; we refer to Bassi (1991, 1995) for studies on this effect.

Table 1
Material parameter values (Ranalli, 1995; Turcotte and Schubert, 2002)

Parameter		u.c.	l.c.	m.l.
Density (kgm^{-3})		2700	2800	3300
Thermal expansion (K^{-1})	1×10^{-5}			
Crustal heat production (W m^{-3})		1×10^{-6}	1×10^{-6}	
Specific heat (J kg^{-1} K^{-1})	1050			
Conductivity (W m^{-1} K^{-1})		2.6	2.6	3.1
Bulk modulus (Pa)		3.3×10^{10}	3.3×10^{10}	12.5×10^{10}
Shear modulus (Pa)		2×10^{10}	2×10^{10}	6.3×10^{10}
Power law exponent n		3.3	3.05	3.0
Activation energy Q (kJ mol^{-1})		186.5	276	510
Material constant A (Pa^{-n} s^{-1}]		3.16×10^{-26}	3.2×10^{-20}	7.0×10^{-14}
Friction angle	30°			
Dilatation angle	0°			
Cohesion factor (Pa)	20×10^{6}			

U.c. is upper crust, l.c. is lower crust and m.l. is mantle lithosphere.

3. Results and discussion

Extension of the lithosphere resulted in localization of deformation west of the Vøring Basin and continental breakup after about 13 My of stretching. Fig. 3A and B shows the thermal structure of the lithosphere during the Maastrichtian–Paleocene extension phase. Because of crustal thinning as a result of earlier extension phases that formed the sedimentary Vøring Basin, temperatures in the lithosphere below this basin were low at the onset of the Maastrichtian–Paleocene extension phase, and they stayed low in the Vøring Basin area during the entire extensional phase. West of this area, where continental breakup even-

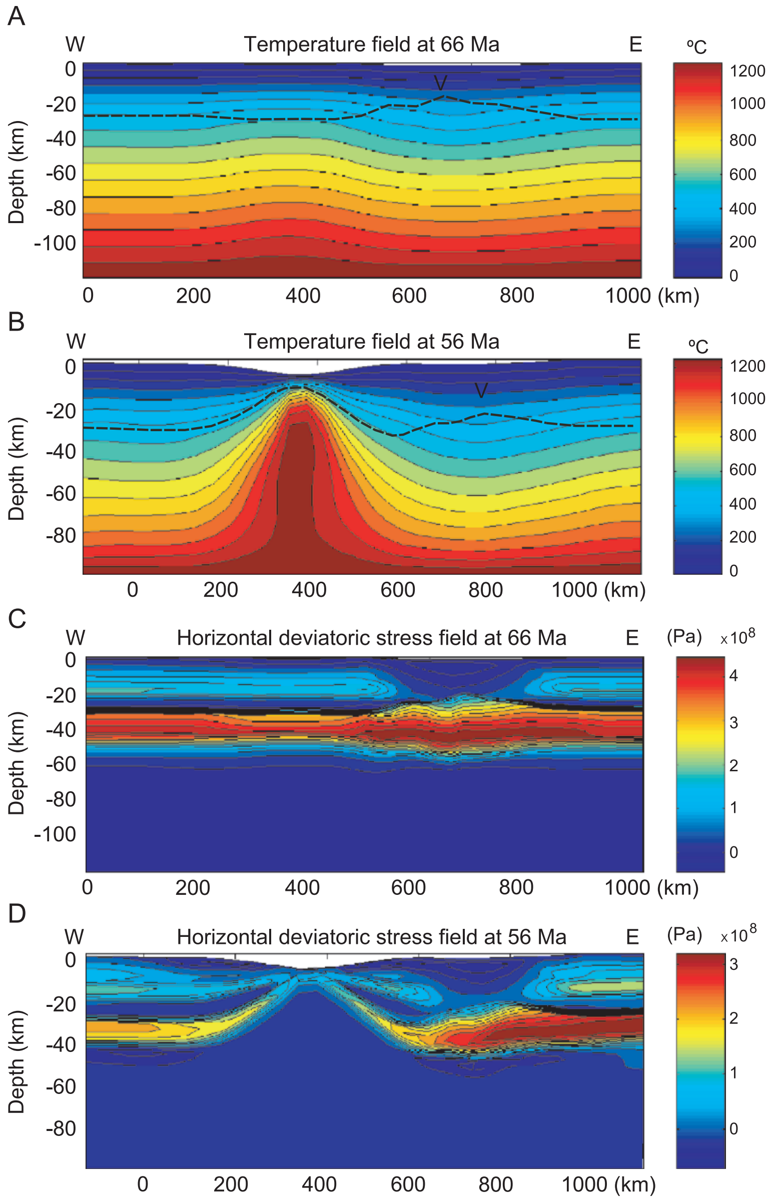


Fig. 3. (A and B): Thermal structure of the lithosphere during the Maastrichtian–Paleocene extension phase. Temperatures in °C. V denotes location of Vøring Basin. West (W) is Greenland side, East (E) is Norwegian side. Black dashed line denotes Moho. (C and D): Horizontal deviatoric stress field.

tually occurred, upwelling of warm mantle material was predicted by the model (Fig. 3A and B).

Decompressional partial melting was predicted to occur in the 4 My preceding breakup (i.e. ~58–54 Ma). This is in agreement with observations (e.g., Eldholm et al., 1989; Skogseid et al., 1992b). Melt was generated in a ~175-km-wide area, mainly in the head of the upwelling mantle material, at a depth of ~20–50 km. The total predicted melt volume was about 950 km^3 (per kilometer along strike of the margin), which is about 1/3 less than estimated melt volumes based on observations (Eldholm and Grue,

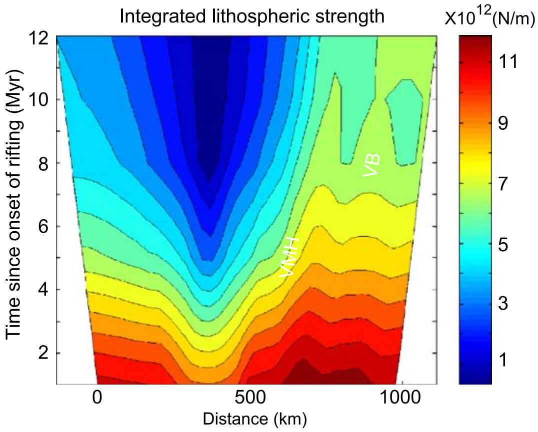


Fig. 4. Evolution of integrated lithospheric strength. VMH is Vøring Marginal High, VB is Vøring Basin. Lithosphere is strongest in Vøring Basin area.

1994). These estimated melt volumes assume the high velocity lower crustal body to consist mainly of magmatic material. It is a matter of debate (Gernigon et al., 2004) to what extent the high velocity body consists of mafic material, and the magma production could have been less than thought. An increase in mantle temperature of 50–100 °C, which is a thermal anomaly frequently associated with mantle plumes (e.g., Pedersen and Skogseid, 1989; White and McKenzie, 1989), would probably generate considerably more melt than observed (Van Wijk et al., 2001). Also, higher extension velocities would cause an increase in the melt volume (e.g., Bown and White, 1995; Van Wijk et al., 2001). We performed tests with higher plate velocity boundary conditions and found that a slightly higher extension velocity of 30 mm/year resulted in continental breakup after only ~9 My. This is a much smaller final rift period than estimated from observations (e.g., Skogseid et al., 2000), and because higher extension velocities are also not supported by seafloor spreading rates shortly after the rift to drift transition (Mosar et al., 2002), these simulations are not considered here. Lower plate velocity boundary conditions would decrease the predicted melt volume. The model does not predict melt generation directly below the western part of the Vøring Basin; (horizontal) migration of melt that is not included in the model would be necessary to explain the intrusions of melt observed at those locations.

Fig. 3C and D shows the influence of the presence of the Vøring Basin on the horizontal deviatoric stress pattern of the mid-Norwegian margin. The crust is weaker and the mantle part of the lithosphere is stronger. The main phase of crustal thinning in the Vøring Basin had ended before the final extension period on the margin began. Consequently, temperatures are low below the basin, and weaker crust material is replaced by stronger mantle material. The lithosphere has locally become more resistant to deformation. As the progress of extension in a tectonic setting is influenced by the evolution of strength of the lithosphere (England, 1983), the center of deformation shifts to a newly preferred location. In Fig. 4, the integrated lithospheric strength (Ranalli, 1995) is shown. The term 'lithospheric strength' is used here in a nonstandard sense, and 'lithospheric strength' is obtained by integrating deviatoric stress values over a vertical column in the lithosphere (Ranalli, 1995). The model predicts that the lithosphere is strongest below the Vøring Basin during the whole Maastrichtian Paleocene extensional phase. This explains the location of the breakup axis further to the west.

Thinning of the lithosphere and continental breakup occurred westward of the Vøring Basin and Marginal High (Fig. 5). Model predictions show that crustal thinning affects a zone landward of the

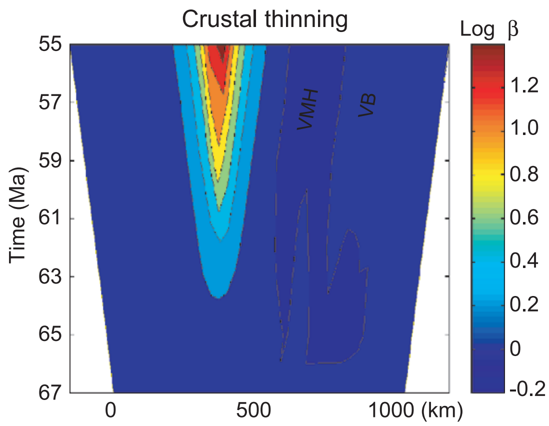


Fig. 5. Evolution of crustal thinning factor β with respect to situation at 67 Ma. VMH is Vøring Marginal High, VB is Vøring Basin. Crustal thinning factor β is defined as the ratio between the initial thickness of the crust and the current thickness. Large thinning factors are west of the VMH. Crustal thickening (log β<0) in VMH area.

continent–ocean boundary, about 100 km wide. Estimations from seismic data and backstripping analyses indicate a 150–200 km wide zone affected by stretching (Skogseid et al., 2000). The more localized deformation in our simulations may be caused by our choice for rheology and the absence of surface processes like sedimentation and erosion. The crust in the Marginal High region between the western Vøring Basin and the breakup axis experienced thickening. In this area, thickening of the crust with several kilometers occurred while the lithosphere model was extended. Crustal thickening mainly concentrated in the lower part of the lower crust and was limited in the upper crust. Ductile flow in the low-strength layer of the lower crust may account for this deformation. This result is supported by observations; strong variations in thickness of the lower crust indicated on Ocean Bottom Seismograph profiles through the Vøring margin suggest that crustal deformation was concentrated in the lower crust (Mjelde et al., 2001). Local thickening of the crust is expected to cause isostatic uplift of segments of the margin. The model-predicted maximum uplift in the western Vøring Basin and the Vøring Marginal High area is about 0.5 km, and the timing is middle to late syn-rift (~63–56 Ma).

Absolute thickening of the crust was not predicted for the Greenland side of the rift axis during extension (Fig. 5). The main difference in the situation between the conjugate Greenland and Norwegian sides was the pre-rift lithosphere configuration; due to the presence of the Vøring Basin, there was a large temperature gradient across the outer part of the margin and a varying crustal thickness on the Norwegian side of the rift axis. This resulted in a strong negative buoyancy in that particular part of the domain adjacent to the rifting zone, increasing downwelling in the area. The vertical velocity component shows negative values at the crust–mantle boundary in the Vøring Marginal High area (Fig. 6). While on the Greenland side upwelling due to stretching of the lithosphere com-

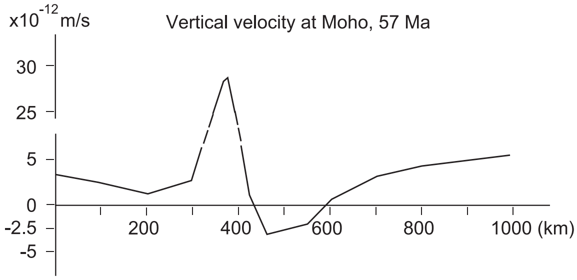
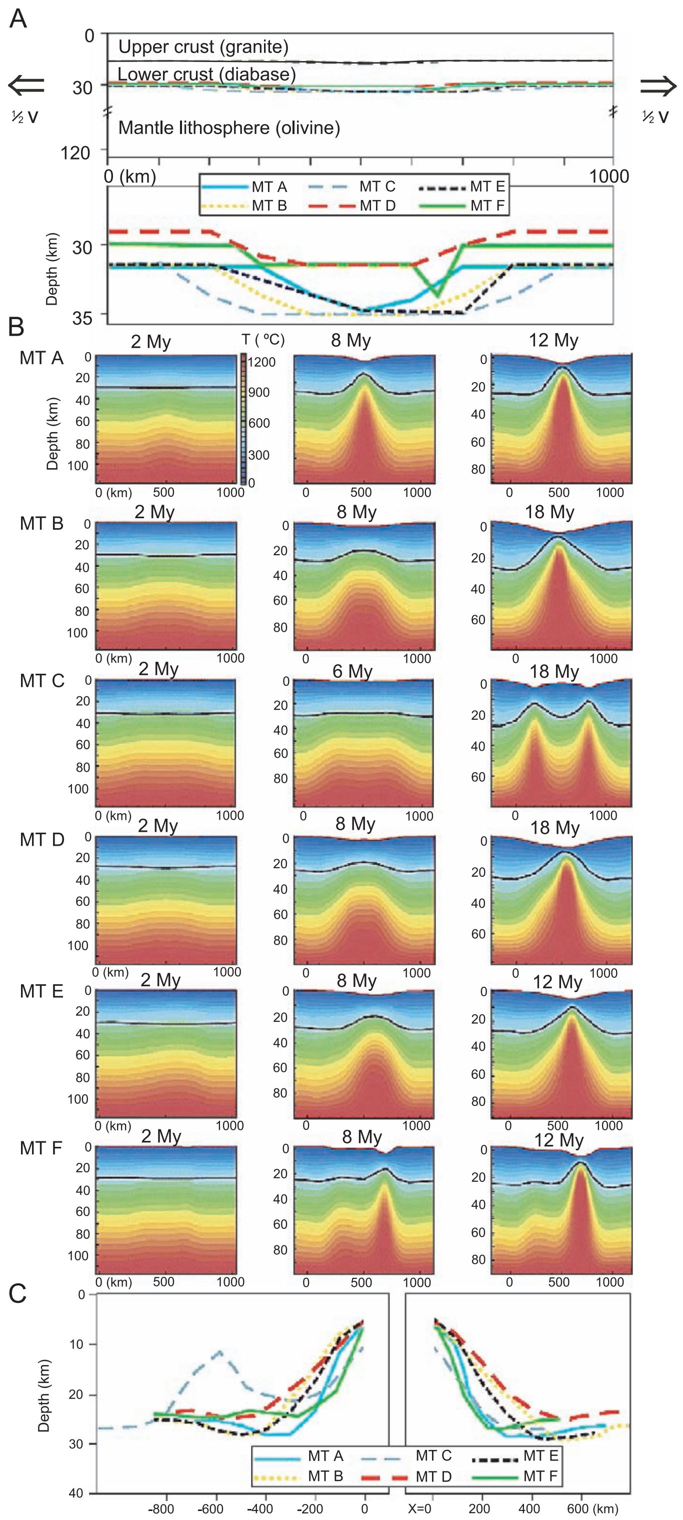


Fig. 6. Vertical component of velocity field at Moho depth, 57 Ma. Strong upwelling is present in rifting zone, and downwelling in Vøring Marginal High area adjacent to rift zone. Absolute downwelling is absent at the Greenland side of the rift axis.

pensated downwelling due to small-scale convection, on the Norwegian side, the downwelling component appeared to be larger.

To further investigate the role that the pre-rift temperature field of the lithosphere played in the rifting process, supplemental tests were performed with different initial Moho topographies, i.e. different initial temperature fields and perturbations in lithospheric strength (also Corti et al., 2003). The extension velocity was 32 mm/year; all other parameters were kept the same (see Table 1). Fig. 7 shows the initial Moho configurations tested (Fig. 7A) and the resulting thermal evolution and Moho depths (Fig. 7B and C). The differences between the initial Moho configurations were not very large; such small variations are very likely to exist almost everywhere. Between Moho tests A–F, the width of the depression, the depth of the Moho and the slope of the Moho on the (right) side of the perturbations varied. We found that this latter factor influenced the evolution of crustal thinning and thickening during extension like observed in the Vøring margin test. Moho test F, where the steepness of the slope was closest to that of the Vøring margin (Fig. 2) and large lateral temperature gradients were present, experienced in the region of the slope very limited crustal thinning and, with respect to its surroundings, relative crustal thicken-

Fig. 7. (A) Initial Moho configurations of Moho tests A–F used for additional tests. Except for the total extension velocity V (32 mm/year) and the initial Moho configuration, all parameters and the model setup were the same as for the Vøring margin test (see Fig. 2). The tested Mohos varied in width of depression, depth and steepness of the side of the depression. (B) Thermal evolution of lithosphere during extension for the different models. The black line in each panel shows the Moho. (C) Final predicted Moho topography at margin pairs. From Corti et al. (2003).


A
Upper crust (granite)
Lower crust (diabase)
Mantle lithosphere (olivine)
½ V
½ V
0
30
120
0 (km)
1000
MT A
MT B
MT C
MT D
MT E
MT F
Depth (km)
30
35
B
MT A
2 My
8 My
12 My
T (°C)
1200
900
600
300
0
MT B
2 My
8 My
18 My
MT C
2 My
6 My
18 My
MT D
2 My
8 My
18 My
MT E
2 My
8 My
12 My
MT F
2 My
8 My
12 My
Depth (km)
0 (km)
500
1000
C
Depth (km)
0
10
20
30
40
-800
-600
-400
-200
0
X=0
200
400
600 (km)

ing during extension as the crust continued to thin in other areas. Thinning in this area ceased after about 6 My of extension with the parameters chosen. The large lateral gradient in temperature initially present in the lithosphere seemed to be a required condition for this behavior. These simulations (Fig. 7, Corti et al., 2003) indicate that inherited Moho topography affects the thermal structure during and after rifting and hence the amount of decompressional partial melting, as well as margin width and structure.

Also, previous studies have shown that the inherited lithosphere structure influences rifting dynamics. For example, the style of rifting and physiography of resulting conjugate margin pairs are affected by the nature of preexisting weaknesses (Dunbar and Sawyer, 1989a,b). Buck (1991) has found a relation between crustal thickness and margin width. The thermal structure of the lithosphere and hence melt generation during extension are suggested to be dependent on the preexisting lithosphere structure (Harry and Bowling, 1999; Corti et al., 2003). From these studies and the present study follows that variations in the thickness of the crust are probably quite important in shaping deformation on the margin, which implies that formation of the Vøring Marginal High and surface uplift may be somewhat more complex than previously suggested. A possible scenario supported by observations includes the influence of fossil imprints resulting from earlier rifting events, besides the isostatic and thermal effects of magmatic intrusions and underplating. It is likely that both processes acted in concert to create the observed thickened crust and uplift. Although the exact locations and dimensions most likely depended on distribution and orientation of older structures (e.g., faults) on the margin (Bukovics and Ziegler, 1985; Vågnes et al., 1998; Mosar et al., 2002), the model results show that the outer margin area was a favorable location for crustal thickening and uplift.

4. Summary and conclusion

A critical component for the understanding of the development of volcanic passive margins relies on the understanding of the role played by fossil imprints in the lithosphere during rifting and continental breakup. From this study and previous studies follows that geodynamics of the rifting process are strongly influenced by the pre-rift lithosphere structure, also away from the central rift zone. On the Vøring margin, the structure of the lithosphere at the onset of the Maastrichtian–Paleocene extension phase reflects the cumulative effect of earlier consecutive rifting events. The lateral differences in temperature and crustal thickness were present as a consequence of migration of the location of maximum extension in time between Norway and Greenland. The most important imprints (Moho depth, thermal structure) of these events on the mid-Norwegian margin were incorporated in the simulation of the final extension phase. It was found that the pre-rift tectonic history is reflected in the strength distribution over the margin; the Vøring Basin formed the strongest part of the margin which explains why the Maastrichtian–Paleocene rift axis was not located here but instead jumped westward with respect to the earlier rift axes locations. The model furthermore predicts that local crustal thickening during extension can be expected when significant lateral thermal variations are present in the lithosphere at the onset of the extensional phase. Remainders of earlier multiple extensional phases on the Vøring margin thus seem to have contributed to the formation of the Vøring Marginal High. Material converged at the particular location on the Norwegian side of the rift axis, resulting in a thickened crust and surface uplift. The model furthermore predicts generation of large amounts of melt during the last part of the extensional phase that may cause additional uplift and crustal thickening of the Marginal High and western Vøring Basin areas.

Acknowledgments

We thank F. Beekman for help with figures and R.S. Huismans and Y.Y. Podladchikov for their large contribution to the numerical code. Constructive reviews by Jon Mosar and an anonymous reviewer helped us to improve the manuscript. This is publication no. 2004.06.04 of the Netherlands Research School of Sedimentary Geology.

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