Differential vertical movements in the eastern North Sea area from 3-D thermo-mechanical finite element modelling

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The response of a heterogeneous lithosphere to a compressional stress field is studied using a three-dimensional thermo-mechanical finite element model. Weak zones in the lithosphere thicken and act as loads that pull down the lithosphere in regions around the weak zones. Strong zones are subjected to less lithospheric thickening than the surroundings and produce surface depressions and uplift in the surrounding areas.

The model is used to study the Late Cretaceous and Paleocene differential vertical movements in the eastern North Sea area. The Sorgenfrei-Tornquist Zone is assumed to be a pre-existing weak crustal zone, which inverts during compression and produces marginal basins by loading the litho-sphere. The area of the Silkeborg Gravity High is an example of a pre-existing strong crustal zone which subsides during compression. Moho topography in the area gives rise to lateral strength variations, which result in surface uplift where Moho is deep and subsidence where Moho is shal-low. These effects, together with the lateral variations of the thermal structure and the stress field, determine the overall Late Cretaceous-Paleocene distribution of vertical movements of the area. This has implications for the pattern of erosion, sediment transport and the distribution of sedi-ment facies.

Keywords: Sorgenfrei-Tornquist Zone, inversion, Silkeborg Gravity High, eastern North Sea, Moho, Cenozoic, finite element.

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The vertical displacement pattern of a heterogeneous lithosphere under compression depends on the mechanical properties of the lithosphere and on the applied stress field. In this study a three-dimensional thermo-mechanical finite element model is used to investigate the response of a heterogeneous lithosphere to a compressional stress field. In particular we wish to elucidate the interacting effects of crustal shortening and thickening and flexural effects in the lithosphere. In a lithosphere containing weak and strong zones, these effects are of significant importance to the vertical movements and thus to the deflections that can be observed at the surface.

The model is applied to the eastern North Sea area to investigate the effects of pre-existing weak and strong zones on the Cenozoic tectonic evolution of the area. This paper focuses on the pattern of differential vertical movements related to the Late Cretaceous-Paleocene compressional inversion.

Geological setting

The Cenozoic evolution of the easternmost part of the North Sea area is significantly influenced by the existence of the Sorgenfrei-Tornquist Zone (Fig. 1) (Liboriussen, Ashton & Tygesen 1987, EUGENO-S Working Group 1988, Mogensen & Jensen 1994, Nielsen & Hansen 2000; Hansen, Nielsen & Lykke-Andersen submitted). The Sorgenfrei-Tornquist Zone may be defined as a zone of Late Cretaceous/Paleocene inversion (Liboriussen et al. 1987, EUGE-NO-S Working Group 1988) which occurred due to the regional change of the stress field from extensional to compressional. This change in stress field was related to the Alpine orogeny (Ziegler 1990). The Sorgenfrei-Tornquist Zone has undergone deformation several times during the Phanerozoic (Michelsen & Nielsen 1993, Mogensen 1994) and can thus be considered to be a zone of prominent structural weakness.

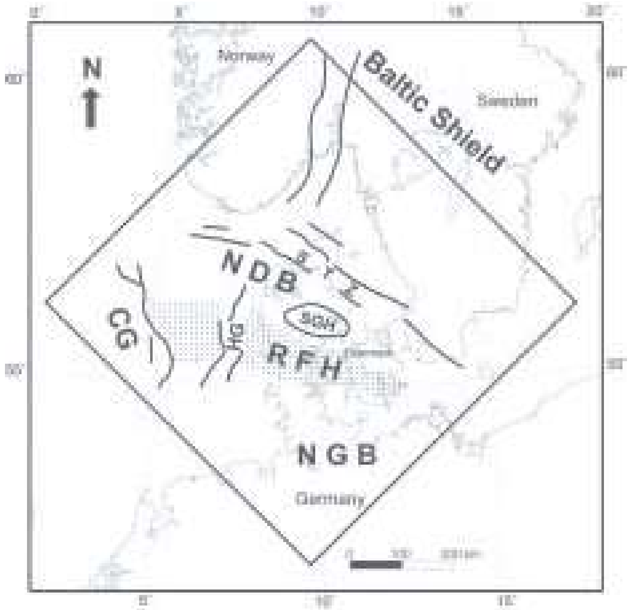

Fig. 1. Study area and main tectonic structures of the area. CG = Central Graben. NDB = Norwegian-Danish Basin. RFH = Ringkøbing-Fyn High. NGB = North German Basin. STZ = Sorgenfrei-Tornquist Zone. SGH = Silkeborg Gravity High. HG = Horn Graben.

An example of what is probably a relatively strong zone in the Danish area is the area of the Silkeborg Gravity High situated in the Norwegian-Danish Basin just north of the Ringkøbing-Fyn High (Fig. 1). The area is characterised by a positive gravity anomaly of 40–50 mGal, a positive seismic velocity anomaly and by a magnetic anomaly (Abrahamsen & Madirazza 1986, Thybo & Schönharting 1991; Thybo 2000). Presumably, these anomalies are due to intrusions in the crust, probably of Carboniferous-Permian age (Thybo & Schönharting 1991). Furthermore, the area is characterised by a higher Pre-Quaternary surface topography than the surroundings (Binzer, Stockmarr & Lykke-Andersen 1993) indicating a late Cenozoic uplift of this area. The high density and high velocity intrusion material is most likely indicative of a stronger crustal material than normal, for which reason the crust at the Silkeborg Gravity High may be assumed to be stronger than the surroundings.

This means that tectonically the area of the Silkeborg Gravity High should behave as a relatively stable block. This is supported by the fact that Zechstein salt lies undisturbed in the area with a thickness of around 1 km and the occurence of major faults in the vicinity of the region (Madirazza 1986, Abrahamsen & Madirazza 1986).

The Moho topography of the area (Fig. 2) (Pedersen 1990, Hurtig et al. 1992, Kinck, Husebye & Larsson 1993, Thybo 1997, Balling pers. comm.) is characterised by Moho depths of up to 48 km beneath the Baltic Shield and about 35 km beneath the Ringkøbing-Fyn High. Beneath the Norwegian-Danish Basin, the North German Basin and the Central Graben, Moho is shallower, due to crustal thinning in Permian/Carboniferous time and in the Jurassic, respectively (Ziegler 1990, Vejbæk 1997).

In the following, the effects of the variations in Moho topography and of pre-existing weaknesses in the Sorgenfrei-Tornquist Zone and a strong zone at the Silkeborg Gravity High on the vertical deformation pattern are investigated.

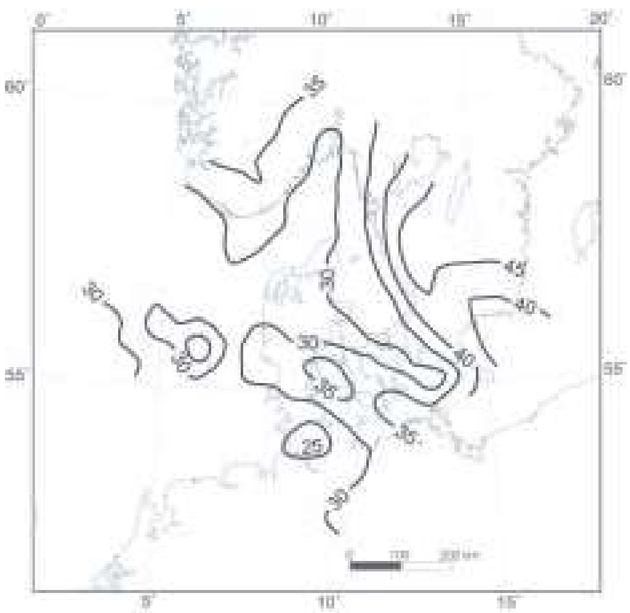

Fig. 2. Moho topography (km) of the study area. Sources of data are given in the text.

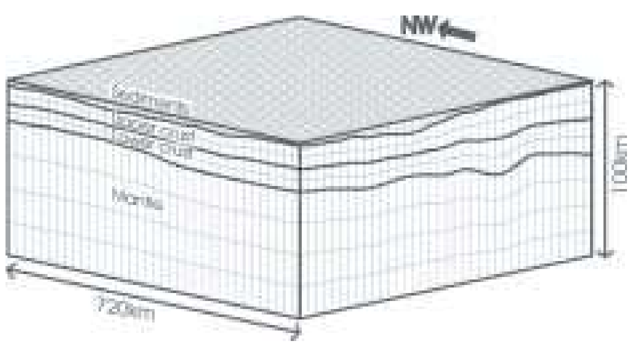

Fig. 3. 3-D finite element model. The heights of the elements are chosen so that the element boundaries coincide with the geological layer boundaries.

Numerical model

The deformation patterns of the area are investigated using a three-dimensional thermo-mechanical finite element model. The lithosphere is modelled as an elastic-visco-plastic continuum (Braun & Beaumont 1987), where the type of deformation taking place in the lithosphere is determined by the temperatures and the stress field. The lithosphere is divided into 36×36×10 eight node elements. The elements are adjusted vertically and assigned material properties, so that the upper two element layers represent the sediments, the four layers below comprise the upper and lower crust, and the upper mantle is represented by the lower four element layers (Fig. 3). The model is 720 km long and 720 km wide and has a vertical extent of 100 km.

The thermal model is the transient heat equation which reads (Carslaw & Jaeger 1959)

$$\nabla \cdot (\Lambda \nabla T) - \rho c \frac{\partial T}{\partial t} = -A \tag{1}$$

Λ is the thermal conductivity tensor, T is temperature, ρ is density, c is heat capacity, t is time and A is the heat production rate. Boundary conditions are a surface temperature of 8°C and a background heat flow that varies from 30 mW/m^2 in the north-eastern part of the area to 40 mW/m^2 in the south-western part of the area (Balling 1995; Gemmer & Nielsen 2000).

The mechanical properties of the model are calculated by solving the equations of motion (Ranalli 1995)

$$\frac{\partial \sigma_{ij}}{\partial x_j} + \rho X_i = 0 \qquad i, j=1,2,3 \tag{2}$$

At low deviatoric stress levels, rocks deform elastically (Braun & Beaumont 1987) following Hookes law (Ranalli 1995)

$$\sigma_{ij} = C_{ijkl}\varepsilon_{kl} \qquad i, j, k, l=1,2,3 \tag{3}$$

σ is the stress tensor, ε is strain, and C is a matrix that describes the mechanical properties of the material. In isotropic materials, C reduces to a matrix containing only two independent parameters for which Young's modulus, E, and Poisson's ratio, ν, are used. Thereby Hooke's law is expressed as (Ranalli 1995)

$$\varepsilon_{ij} = (1/E)[(1 + n)\sigma_{ij} - \nu\sigma_{kk}\delta_{ij}] \qquad i, j, k=1,2,3 \tag{4}$$

At higher deviatoric stresses, the lithosphere is subjected to more permanent kinds of deformation.

At high temperatures, it shows ductile behaviour which is modelled by a non-linear elasto-viscous rheology. The strain rate, $\dot{\varepsilon}$, for an elasto-viscous rheology is given as (Woodward 1980)

$$\dot{\varepsilon} = C^{-1}\dot{\sigma} + Q\sigma \tag{5}$$

C is the elastic matrix known from Hooke's law, $\dot{\sigma}$ is stress rate, Q is a matrix in which the components depend inversely proportionally on the effective viscosity, η (Braun & Beaumont 1987)

$$\eta = B^{*}\left[\dot{E}_{2D}^{1/2}\right]^{1-1/n^{*}} \exp\left(\frac{Q^{*}}{n^{*}RT}\right) \tag{6}$$

T is temperature, R is Boltzman's constant, $\dot{E}_{2D}$ is the second invariant of the deviatoric part of the strain rate vector and n*, B* and Q* are creep parameters that depend on the material.

At low temperatures, the lithosphere undergoes brittle failure (Braun & Beaumont 1987). This is modelled by an elasto-plastic rheology where the Murrell's extension of the Griffith criterion is used (Jaeger & Cook 1979; Braun & Beaumont 1987)

$$F = J_{2D} + 4T_0 J_1 \tag{7}$$

J_{2D} is the second invariant of the deviatoric part of the stress tensor, T_0 is tensile strength, and J_1 is the first

Table 1. Material parameters used. The values are partly taken from Jaeger & Cook (1979), Braun & Beaumont (1987), Balling (1995) and Hansen et al. (2000).

Parameter	Sediments	Upper crust	Lower crust	Mantle
Thermal conductivity (W/m°C	2.5	3.0	2.0	4.0
Heat production rate ($\mu W/m^3$	1.5	0.4–2.4	0.2	0.01
Heat capacity (W/kg/°C)	900	900	900	1000
Density (kg/m^3)	2300	2700	2900	3300
Young's modulus (Pa)	10^{11}	10^{11}	10^{11}	10^{11}
Poisson's ratio	0.25	0.25	0.25	0.25
Creep parameter, Q (kJ/mol)	135	135	220	500
Creep parameter, n	3.10	3.10	3.20	4.48
Creep parameter, B (MPa s^{ln})	200	200	12.3	0.26
Compressive strength (MPa)	17.5	26.2	26.2	52.4

invariant of the stress tensor. When F is larger than zero, brittle failure occurs, and the resulting stress is projected towards the failure envelope (Braun & Beaumont 1987)

$$\sigma = p^{tr} I + \left[\frac{-12 T_0 p^L}{J_{2D}^{tr}} \right]^{1/2} S^{tr} \qquad (8)$$

p^{tr} and S^{tr} are the pressure and the deviatoric stress obtained from the visco-elastic calculations, J_{2D}^{tr} is the second invariant of the deviatoric stress tensor obtained from the visco-elastic calculations and p^L is the lithostatic pressure. The applied material parameters are given in Table 1.

To study the thermo-mechanical properties of the eastern North Sea, the Moho and basement topographies of the area are included in the numerical model (Fig. 3), and the model is subjected to compression.

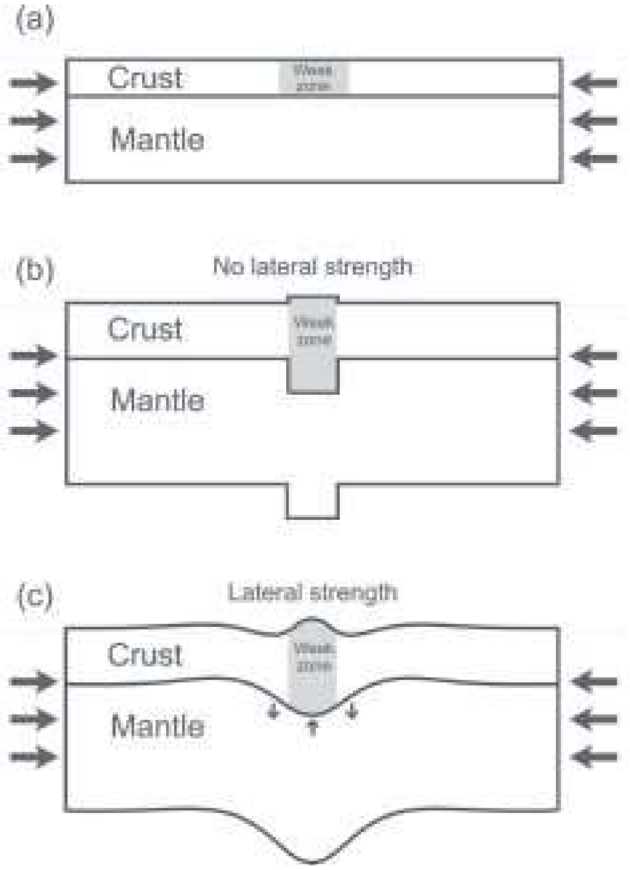

Fig. 4. Sketch of a crustal weak zone in a lithosphere under compression. (a) Initial model. (b) No lateral strength in the lithosphere (Airy isostasy). The weak zone is thickened more than the surroundings which produces surface uplift and subsidence at Moho relative to the surroundings. (c) Lateral strength in the lithosphere reduces the amount of Moho subsidence beneath the weak zone which causes additional surface uplift, and drags down the Moho in the marginal areas of the weak zone, resulting in basins at the surface.

Weak and strong zones in a lithosphere under compression

Lateral variations in lithospheric properties are of significant importance to the resulting deformation pattern of a lithosphere under compression. The effect of the presence of a crustal weak zone is sketched in Figure 4. The weak zone is subjected to a larger amount of thickening than the surroundings, which, in the case of isostatic equilibrium, will result in uplift of the surface and subsidence of Moho (Fig. 4b). However, lateral strength, particularly in the upper mantle, causes a smoother low-amplitude pattern of Moho subsidence, resulting in enhanced uplift of the surface of the weak zone and subsidence of the surface at the margins of the weak zone (Fig. 4c). The formation of inversion zones and marginal basins is discussed in detail by Nielsen & Hansen (2000).

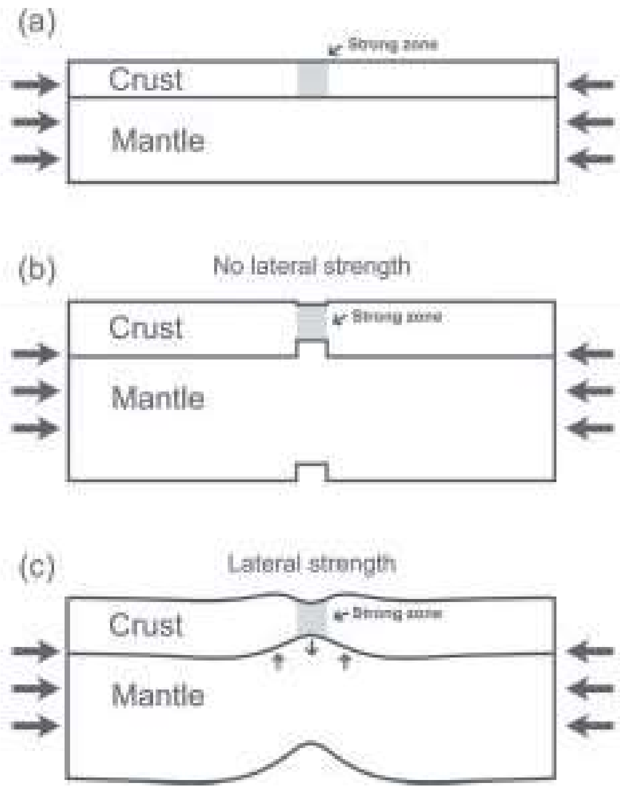

Fig. 5. Sketch of a crustal strong zone in a lithosphere under compression. (a) Initial model. (b) No lateral strength in the lithosphere (Airy isostasy). The strong zone is not subjected to deformation, and the surface level remains constant. The surrounding areas are thickening, and the surface is uplifted to obtain isostatic equilibrium. (c) Lateral strength in the lithosphere causes a drag-down of the strong zone, resulting in surface subsidence. At the margins of the strong zone, the flexural effects produce uplift of the lithosphere, creating surface uplift.

A strong zone in a lithosphere under compression is subjected to a smaller amount of thickening than the surroundings. In the case of isostatic equilibrium, this results in constant surface and Moho levels in the strong zone, whereas in the surroundings the surface is uplifted which is compensated by Moho subsidence (Fig. 5b). As in the case of the weak zone, the competent layers, particularly the one beneath the Moho, reduce the lateral variations in vertical movements. Thus, the strong zone is dragged down causing subsidence of the surface and uplift at the margins (Fig. 5c). This effect is used as the main explanation for the Neogene evolution of the Transylvanian Depression.

Synthetic model

To investigate the combined effects of pre-existing weak and strong zones on the deformation pattern, a synthetic model is studied. This model consists of 5 km sediments, a 12.5 km thick upper crust, a 12.5 km thick lower crust and 70 km upper mantle material. The material parameters of the different model layers are given in Table 1. All layer boundaries are horizontal, and the model is 600 km long and 400 km wide. A pre-existing weak zone is introduced in the crust by reducing the strength in a narrow zone at one side of the model (Fig. 6a). The average effect of a pre-existing strong zone (e.g. an intrusion) is modelled by introducing values of the material parameters that are intermediate between crustal and mantle values in a circular area, as indicated in Fig. 6a. The shapes of the anomalies and the distance between the two zones are chosen so that they simulate a simplified situation of the Sorgenfrei-Tornquist Zone and the area of the Silkeborg Gravity High. The model is subjected to compression at a strain rate of $5\times10^{-17}s^{-1}$ for a period of 20 Ma by moving the boundary opposite to the weak zone. This results in a total model shortening of about 3%. Fig. 6b shows the cumulated surface deflection after compression. The area of the weak zone inverts, and bordering marginal basins are formed. This effect is due to increased crustal thickening in the weak zone. The thickened area acts as a load on the lithosphere and thus causes it to subside in an area wider than the load due to lateral strength of the upper mantle (Fig. 4).

The area of the strong zone coincides with the marginal basin of the inversion zone. Thus, the amount of subsidence is enhanced in this area due to a smaller amount of model thickening than in the surroundings and due to the pull-down effect of the lateral strength below Moho (Fig. 5). Furthermore, the lateral strength in the lithosphere prevents the bordering regions from subsiding, which results in uplift of the surface in the area surrounding the strong zone. This effect is particularly significant in the direction parallel to the direction of compression. The surface deflection pattern of Fig. 6b can be considered to be the superposition of the effects shown in Figs. 4 and 5.

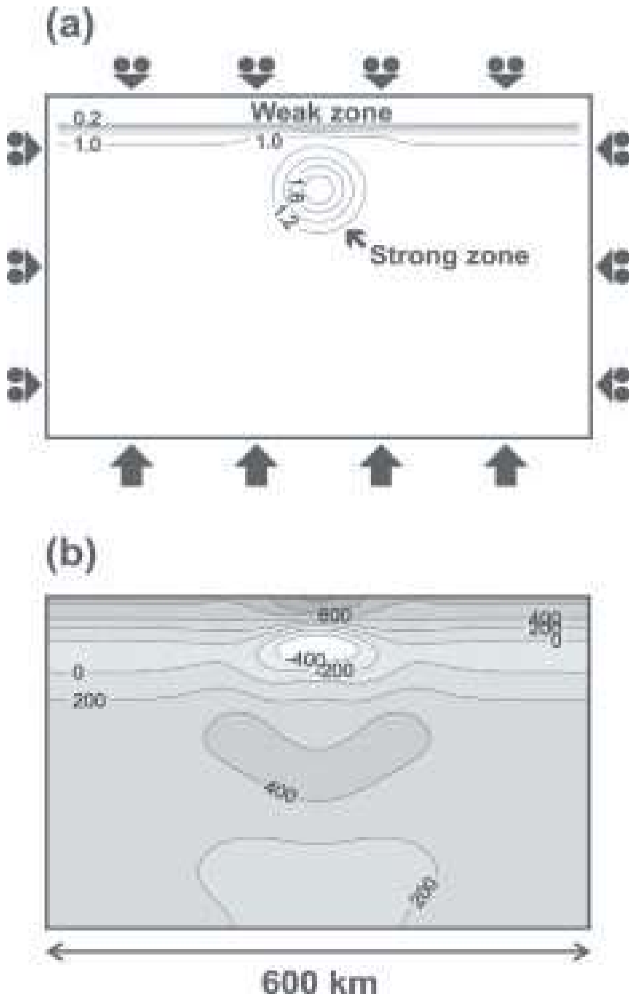

Fig. 6. Synthetic model with pre-existing weak and strong zones (see text for details). (a) Tensile strength in the crust as a fraction of the values given in Table 1. (b) Cumulative vertical movements (m) of the surface after 20 Ma compression at a strain rate of $5\times10^{-17}s^{-1}$.

Model of the eastern North Sea area

In the present model of the eastern North Sea area we study the effects of the Late Cretaceous/Paleocene compressional event on the differential vertical move-

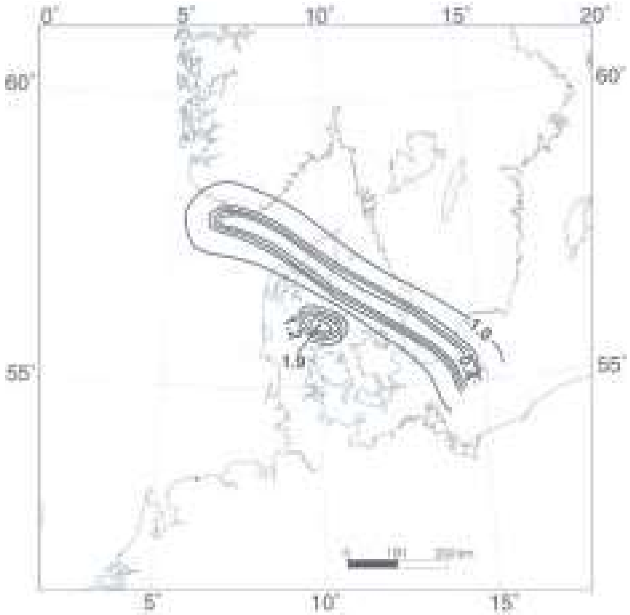

Fig. 7. Compressive strength in the crust as a fraction of the values given in Table 1, used in the model of the eastern North Sea area.

ments in the area. These effects contribute to the observed total deformation pattern for that period. The model does not consider the effects of the thermal subsidence related to cooling of the probable thermal remnant of the Jurassic plume in the Central North Sea (Ziegler 1990, Nielsen 2002), the regional uplift of Fennoscandia (Jensen & Michelsen 1992, Jensen & Schmidt 1992, Riis 1996, Stuevold & Eldholm 1996), the initial topography of the area, the variations in eustatic sea level and the surface processes. By including the effects of surface processes, the modelled displacements would be enhanced due to the loading and unloading effects of sedimentation and erosion, respectively.

The effect of the assumed weak Sorgenfrei-Tornquist Zone is modelled by reducing the strength in the upper and lower crust and in the sediments of this area. The assumed relatively strong area at the Silkeborg Gravity High is modelled by introducing values of creep parameters and of the compressive strength that are intermediate between crust and mantle values in the crust of this area. The applied distribution of compressive strength in the crust is shown in Figure 7.

To simulate the compressional event of the area in Late Cretaceous/Paleocene, the model is subjected to NE-SW directed compression at a strain rate of $5\times10^{-17}\,s^{-1}$ in a period of 20 Ma. The total amount of model shortening is thereby about 3%, and the force used to perform the compression is 10^{12}–5×10^{12} Nm^{-1} which is comparable to the values of plate boundary forces (Bott 1993, Ranalli 1995).

The Moho topography in the eastern North Sea area (Fig. 2) varies and affects the strength of the lithosphere. Figure 8 shows strength and temperature profiles in the Baltic Shield, the Norwegian-Danish Basin and the Ringkøbing-Fyn High. The three temperature profiles are similar although the temperatures in the Baltic Shield are slightly lower than in the Nor-

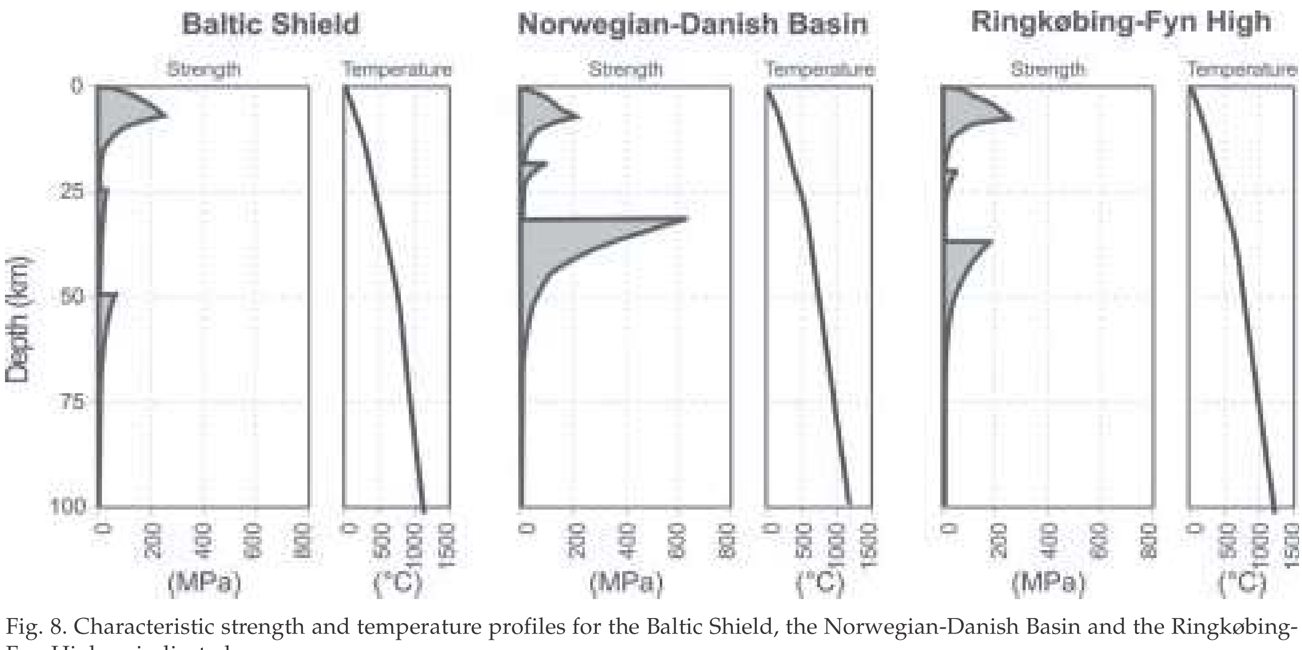

Fig. 8. Characteristic strength and temperature profiles for the Baltic Shield, the Norwegian-Danish Basin and the Ringkøbing-Fyn High as indicated.

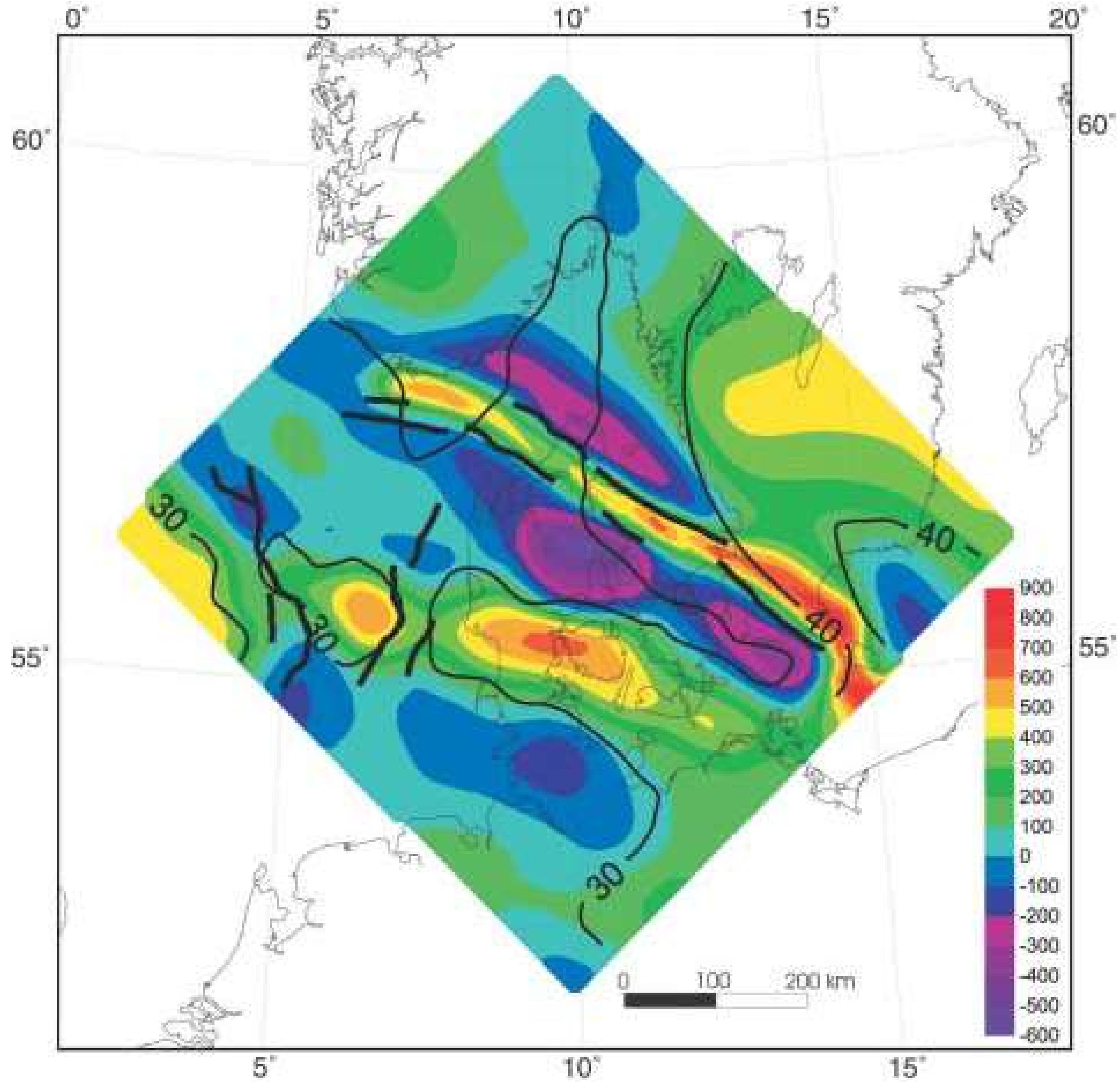

Fig. 9. Modelled, cumulative surface deflections (m) in the eastern North Sea area after 20 Ma compression at a strain rate of $5 \times 10^{-17} s^{-1}$. Contours indicate Moho depths (km). Bold lines indicate main tectonic structures.

wegian-Danish Basin and in the Ringkøbing-Fyn High, due to lower heat flow values in the Baltic Shield (Hurtig et al. 1992, Balling 1995). The strength profiles differ more significantly. In the Baltic Shield and in the Ringkøbing-Fyn High the strengths in the upper mantle are 3-10 times lower than in the Norwegian-Danish Basin. This is caused by the larger crustal thicknesses in the Baltic Shield and at the Ringkøbing-Fyn High, resulting in higher temperatures in the uppermost part of the mantle. The strength of the lithosphere decreases with increasing temperature therefore the upper mantle is weak in areas of a thick crust.

The deformation pattern resulting from compression (Fig. 9) reflects the lateral variations in lithospheric strength. In the Baltic Shield and at the Ringkøbing-Fyn High, the surface is uplifted because of lithospheric thickening in the areas where the mantle is relatively weak.

In the area of the Sorgenfrei-Tornquist Zone the surface is uplifted due to extra thickening of the introduced pre-existing weak crust. Around the Sorgenfrei-Tornquist Zone significant areas of surface subsidence appear. The areas of subsidence are most prominent where Moho is shallow. Like in the synthetic example, these marginal basins arise due to the lateral strength of the lithosphere, particularly in the relatively strong uppermost mantle. The thickened crust in the weak zone acts as a load that pulls down the lithosphere in the area surrounding the weak zone.

The modelled surface uplift in the Sorgenfrei-Tornquist Zone can be correlated with the observed Late Cretaceous-Paleocene basin inversion of this area, and the modelled marginal basins south of the weak zone may be correlated with the observed coeval development of depositional centres (Liboriussen et al. 1987, Michelsen & Nielsen 1993, Mogensen & Jensen 1994). Also the modelled basin north-east of the weak zone south of Sweden can be correlated with geological observations of subsidence and chalk deposition in the Late Cretaceous (Norling & Bergström 1987). The modelled basin north-east of the weak zone in Kattegat may be correlated with occurrences of Upper Cretaceous deposits in this area (Ziegler 1990, Thomsen 1995). The model suggests that a more prominent Upper Cretaceous/Paleocene basin has existed at this location. However, this basin would be eroded today due to the deep Cenozoic erosion of the Kattegat-Skagerrak area (Jensen & Michelsen 1992, Jensen & Schmidt 1992), which is not part of the present model.

Like in the synthetic example, the effects on the deformation pattern of an introduced strong zone in the area of the Silkeborg Gravity High coincide with the marginal basin south-west of the weak Sorgenfrei-Tornquist Zone. However, the effect of the strong zone seems to be an enhancement of the surface subsidence in Central Jutland. This may be correlated with significantly thicker occurrences of Paleocene and Eocene deposits in this area (Dinesen, Michelsen & Lieberkind 1977, Thomsen 1995, Clausen et al. 2000). The present day elevated levels of the Pre-Quaternary in this area may reflect later uplift of the area as a consequence of the change in the dominant compressional regime to a NW-SE direction in connection with the opening of the North Atlantic.

The modelled uplift of the Ringkøbing-Fyn High agrees with observations of erosion in this area (Thomsen 1995, Clausen & Huuse 1999, Huuse 1999). Huuse (1999) studied the offshore area of the Ringkøbing-Fyn High and observed a significantly smaller thickness of the Chalk Group to the west of the Horn Graben. This may be correlated with the modelled gradient in surface deflection from about 300 m uplift to the east of the Horn Graben to about 400 m uplift to the west. However, the modelled amplitudes of surface deflection at the Ringkøbing-Fyn High are in general relatively large. A smaller amplitude of uplift could, for instance, be obtained by introducing stronger crustal material in the Ringkøbing-Fyn High. A possible physical phenomenon that would validate this is phase transformations in the lower crust, which may have increased the strength (Mengel & Kern 1991). Furthermore, the model does not take into account lateral variations in initial surface topography. Before compression, the Ringkøbing-Fyn High was below sea level (Ziegler 1990). Thus, part of the uplift brought the Ringkøbing-Fyn High to the sea level without causing erosion.

As in the case of the Ringkøbing-Fyn High, it can be argued that the strength in the Baltic Shield should be larger, due to the possible existence of e.g. eclogites in the lower crust (Balling 1995). Thus the deflections in the Baltic Shield caused by the Late Cretaceous/Paleocene compressional event may be lower than indicated by the model results.

Conclusion

The relationships between lithospheric heterogeneity and compression-induced vertical deflections have been investigated. A pre-existing weak zone in the crust results in local thickening of the weak zone. The thickened crust acts as a load that pulls down the lithosphere and thus affects the surface deformation pattern on a regional scale. Pre-existing strong zones are subjected to a smaller amount of thickening than the surroundings, which become elevated relative to the strong zone. The lateral strength of the lithosphere causes further subsidence of the strong zone and uplift of the adjoining areas.

In the eastern North Sea area, the Sorgenfrei-Tornquist Zone acted as a pre-existing crustal weak zone and the area of the Silkeborg Gravity High acted as a pre-existing strong zone during the Late Cretaceous/Paleocene compressional event. This produced basin inversion in the Sorgenfrei-Tornquist Zone and coeval development of marginal basins. The area of the Silkeborg Gravity High subsided, enhancing the subsidence of the marginal basins associated with the inversion zone.

The Moho topography is of significant importance to the lateral variations in upper mantle strength. In areas of a relatively deep Moho (the Ringkøbing-Fyn High and the Baltic Shield) the lithosphere is relatively weak causing increased lithospheric deformation and corresponding surface uplift. This explains, for instance, the observed erosion of the Ringkøbing-Fyn High in the middle Paleocene.

The described correlations between the modelled surface deflections and actual observed vertical movements in the area indicate that several of the observed Late Cretaceous/Paleocene uplift and subsidence events may be explained by the large scale mechanisms of lithospheric compression dealt with in this paper.

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Dansk sammendrag

En 3D termomekanisk finite element model er benyttet til at undersøge de differentielle vertikale bevægelser, der opstår, når lithosfæren udsættes for et kompressivt stressfelt. Svage zoner i lithosfæren bliver fortykket og forårsager hævning af overfladen lokalt over den svage zone og indsynkning af de tilgrænsende områder. Stærke zoner i lithosfæren udsættes for mindre deformation end omgivelserne og bevirker lokal overfladeindsynkning og hævning af overfladen i området omkring den stærke zone.

Modellen er anvendt på det østlige Nordsøområde til at studere betydningen af den Sen Kridt/Paleocene kompressive begivenhed for de vertikale bevægelser i området. I dette område er Sorgenfrei-Tornquist Zonen et eksempel på en svag zone, mens området omkring Silkeborg Tyngde Maksimet formodes at være et eksempel på en stærk zone. Moho topografien i området bevirker laterale variationer i lithosfærestyrke, hvilket resulterer i overfladehævning, hvor Moho er dyb, og indsynkning, hvor Moho er tættere på overfladen. Disse effekter bestemmer sammen med de laterale variationer i den termiske struktur og stressfeltet den overordnede Sen Kridt-Paleocene fordeling af differentielle vertikale bevægelser i området.

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