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Tectonophysics 373 (2003) 25–37

TECTONOPHYSICS

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Numerical models of tectonic deformation at the Baltica–Avalonia transition zone during the Paleocene phase of inversion

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Received 27 November 2001; accepted 23 September 2002

Abstract

Predictions from dynamic modelling of the lithospheric deformation are presented for Northern Europe, where several basins underwent inversion during the Late Cretaceous and Early Cenozoic and contemporary uplift and erosion of sediments occurred. In order to analyse the evolution of the continental lithosphere, the equations for the deformation of a continuum are solved numerically under thin sheet assumption for the lithosphere. The most important stress sources are assumed to be the Late Cretaceous Alpine tectonics; localized rheological heterogeneities can also affect local deformation and stress patterns. Present-day observations available in the studied region and coming from seismic structural interpretations and stress measurements have been used to constrain the model. Our modelling results show that lateral variation in lithospheric strength below the basin systems in Central Europe strongly controls the regional deformation and the stress regime. Furthermore, we have demonstrated that the geometry of the boundary between Baltica and Avalonia, together with different rheological characteristics of the two plates, had a crucial role on local crustal deformation and faulting regime resulting in the Baltica–Avalonia transition zone from the S–N Alpine convergence.

Keywords: Tectonic deformation; Baltica–Avalonia transition zone; Paleocene phase of inversion

1. Introduction

During the last 20 years, an extensive geological and geophysical database was collected in the area extending from the North Sea to the Mediterranean, in the southern part of Europe. In the 1980s, the European Geotraverse Project (EGT, 1990; Blundell et al., 1992) integrated much geological and geophysical information in Europe. This project paved the way to several new European projects during the 1990s, such as the EUROPROBE programme (EUROPROBE, 1996; Blundell, 1999) and one of its major scientific projects, the Trans-European Suture Zone project (TESZ) (Pharaoh, 1999), which enriched the already available database along the Trans European Suture Zone by new seismic experiments (MONA-LISA Working Group, 1997a,b; DEKORP BASIN Research Group, 1998, 1999; Guterch et al., 1999) and the acquisition of new potential field data (Wybraniec et al., 1998). All these new data provided a deeper knowledge of the structure of the crust and the mantle of Europe and encouraged controversial discussions about the origin and the extent of specific crustal structures. The interpretation of new seismic refraction data indicates a high velocity layer in the lower crust

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doi:10.1016/S0040-1951(03)00281-6

extending to the south as far as the Elbe Line (Thybo, 1990), to the east into eastern Germany and to the west into the North Sea (Rabbel et al., 1995; Abramovitz et al., 1999; Abramovitz and Thybo, 2000). However, the nature of the southern border of the ancient crust of the Baltic Shield and its transition into the younger crust of Western and Southern Europe (Avalonia) has not been understood yet. Possible advocated candidates are the Trans European Fault (TEF) (Berthelsen, 1992; Thybo, 1997), the Elbe-Odra Line (EL) (Tanner and Meissner, 1996; MONA-LISA Working Group, 1997a,b; Bayer et al., 2002) or the Caledonian Deformation Front (CDF) at shallower depths (Berthelsen, 1992). Fig. 1 shows the tectonic and geologic features of the region from the Baltic Shield to the Alpine Front that have a relevant role in our study.

In the last years, many studies pointed out that during their evolution a lot of basins were characterized by subsequent phases of fault reactivation which, in some cases, induced basin inversion with reverse movements of the faults under compressive stress. In particular, several areas in northern and central Europe (e.g. Danish Basin, North German Basin, Polish Trough) were subject to tectonic inversion in the Late Cretaceous–Early Cenozoic (Ziegler, 1990) as a response to compressive stresses caused by the collision of Europe and Africa and by the ridge push forces from the opening of the northern Atlantic. During the Alpine orogeny, the region embedded within the Sorgenfrei–Tornquist Zone and the Caledonian Deformation Front, and extending geographically from the Baltic Shield to the Danish and German sedimentary basins (Fig. 1) accommodated the main inversion structures, with earlier normal and transtensional faults being reactivated as reverse faults, before a relative stable phase. Consequently, the North Sea underwent region-

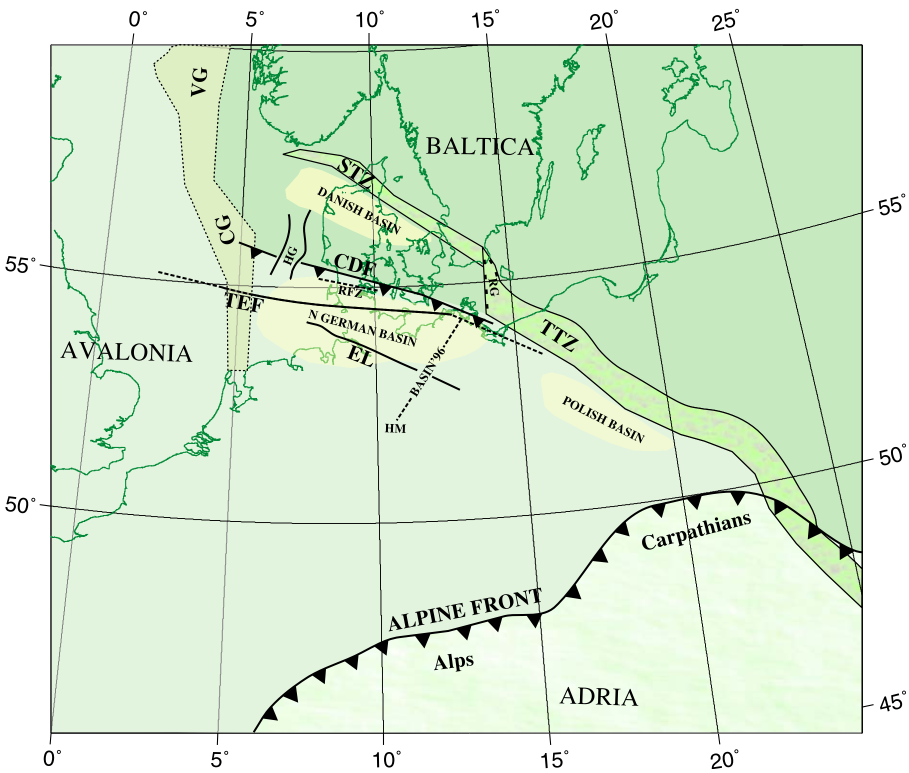

Fig. 1. Sketch map of the areas between the Alpine Front and the Baltic Shield with the main geological structure crucial in our study, according to Ziegler (1999) and Berthelsen (1992). Abbreviations: CDF, Caledonian Deformation Front; CG, Central Graben; EL, Elbe Line; HG, Horn Graben; HM, Hatz mountains; RFZ, Rømø Fault Zone; RG, Ronne Graben; STZ, Sorgenfrei–Tornquist Zone; TEF, Trans-European Fault; TTZ, Teisseyre–Tornquist Zone; VG, Viking Graben.

al subsidence, while uplift characterized the Baltic Shield (Ziegler, 1990). The main inversion features developed along the Sorgenfrei–Tornquist Zone (STZ) and the Teysseire–Tornquist Zone (TTZ).

Part of the Danish Basin, situated on the Baltic continental plate, was inverted during the Late Cretaceous–Early Cenozoic as a result of the Alpine orogeny and the ridge push from the North Atlantic (Thybo, 2001). Inversion continued during the Paleocene, characterized by two phases of uplift associated with the Paleocene Laramide and the Late Eocene to Early Oligocene Pyrenean orogenic phases and resulted into regional uplift and erosion of large quantities of sediments (De Lugt et al., 2001). Clausen et al. (2001) demonstrate that the mid-Paleocene development of the eastern North Sea Basin was dominated by the inversion of the Sorgenfrei–Tornquist Zone.

The Polish Trough, located at the boundary between the East and the West European Platforms, is another example of an inverted Permian basin, where tectonic inversion took place during the Maastrichtian–Paleocene as a result of SW–NE intra-continental compression. Lamarche et al. (2001) demonstrate that the induced uplift affects a narrower area than the original basin, corresponding to the area where the subsidence was most intense before the inversion took place.

A complex poly-phase history characterized the North German Basin (NGB) region, with three main stages: formation in the Upper Carboniferous (Breitkreuz and Kennedy, 1999), thermal subsidence lasted until the Late Triassic (Scheck and Bayer, 1999) when a second extension phase followed and inversion took place during the Late Cretaceous and Early Cenozoic, as also observed throughout NW Europe (Ziegler, 1990). The inversion structures are characterized by surface uplift in the inverted zone, including the uplift of the Hartz mountains (Fig. 1) and surface subsidence in the areas bordering the inverted zones, where sedimentation could occur. Increased subsidence followed during the Cenozoic in the North East German Basin. Recent studies about structural development of the inverted North German Basin by Kossow et al. (2001) clear that two distinct periods of inversion can be distinguished in the North East German Basin. A first phase of inversion occurred in the Late Jurassic to Early Cretaceous and was characterized by the growth of salt-cored anticlines, while a second inversion period occurred in the Late Cretaceous, with upthrusting along the Gardelegen Fault system that exerted a north-directed compression on the post-Zechstein sequence. In the present study, it is assumed that the Alpine tectonics is responsible for the Late Cretaceous inversion, while the Late Jurassic deformation is interpreted as a local intra-plate transpressional phenomenon.

In the presented modelling, it is assumed that during the process of basin inversion, sedimentary basins behave like relatively weak zones of the continental lithosphere due to the existence of pre-existing faults. During the last years, considerable improvement has been achieved in tectonic modelling studies applied to Central Europe (e.g. Kooi et al., 1991; Cloetingh and Kooi, 1992; Horwath and Cloetingh, 1996; Gruntal and Stromeyer, 1992; Bayer et al., 1999; Marotta et al., 2000; Marotta et al., 2001). These studies demonstrate the importance of taking into account both spatial and temporal variations of the thermo-mechanical properties of the geological systems for a correct prediction of the evolution of the basin systems. In this study, we model the horizontal and the vertical lithospheric deformations induced by Alpine tectonics in Central and North Europe in order to predict the evolution of the strain and stress fields stored in the crust, and to verify the occurrence of basin inversion during the orogenic deformation phase.

2. Model set up

We used the finite element approach developed in Marotta et al. (2001), within the viscous thin sheet scheme proposed by England and Mc Kenzie (1983), to model the tectonic deformation in Europe due to the Alpine orogene. Since the thin sheet approach is broadly described in literature, only the main points of its theory will be described. For the full description of the approach, the reader is referred to England and Mc Kenzie (1983).

At a geological time scale, the Navier-Stokes equation for the deformation of a continuum is usually expressed in the form

$$\frac{\partial p}{\partial x_i} = \frac{\partial \tau_{ij}}{\partial x_j} - \rho g_i \tag{1}$$

with $(i, j = 1, 2, 3)$. τ_{ij} is the deviatoric stress tensor, p is the pressure, ρ is the density and $\overrightarrow{g} = (0, 0, g)$ is the gravity acceleration. Within the thin sheet approximation, the lithosphere is assumed to be a thin viscous plate of varying crustal and lithospheric thickness above a non-viscous asthenospheric half-space. This assures the absence of basal stress. Since the thickness of the lithosphere is assumed to be small compared to the wavelength of the applied load, the vertical gradients of the horizontal velocity components and of the deviatoric stress can be neglected in the analysis. Isostatic compensation is also assumed, which implies the absence of lateral strength and therefore, flexural phenomena are not reproduced by the model. All these assumptions lead the three equations expressed by Eq. (1) to be reduced, after the integration through the thickness of the lithosphere, to a set of two equations, only for the horizontal components of the velocity

$$\frac{\partial}{\partial x_j}\left[B\dot{E}^{\left(\frac{1}{n}-1\right)}\dot{\varepsilon}_{ij}\right]-\frac{\partial}{\partial x_i}\left[B\dot{E}^{\left(\frac{1}{n}-1\right)}\dot{\varepsilon}_{zz}\right]=\frac{g\rho_c(1-\rho_c/\rho_m)}{2L}\frac{\partial s^2}{\partial x_i} \tag{2}$$

with $(i, j=1, 2)$.

$$\bar{\tau}_{ij}=B\dot{E}^{(1/n-1)}\dot{\varepsilon}_{ij}=2\mu_{\text{eff}}\dot{\varepsilon}_{ij} \tag{3}$$

is the vertically averaged deviatoric stress, where B is the depth-averaged strength coefficient, $\dot{\varepsilon}_{ij}$ is the strain rate expressed in terms of the velocity components u_i as

$$\dot{\varepsilon}_{ij}=\frac{1}{2}\left(\frac{\partial u_i}{\partial x_j}+\frac{\partial u_j}{\partial x_i}\right) \tag{4}$$

$\dot{E}$ is the second invariant of the strain rate defined as

$$\dot{E}=(\dot{\varepsilon}_{ij}\dot{\varepsilon}_{ij})^{1/2} \tag{5}$$

and n is the power-law exponent. The parameter

$$\mu_{\text{eff}}=\frac{B\dot{E}^{(1/n-1)}}{2} \tag{6}$$

defines the effective viscosity of the lithosphere; the use of different values of μ_{eff} allows to implement rheological heterogeneities in the model. ρ_c and ρ_m are the mean crustal and mantle densities, 2800 and 3000 kg/m^3, respectively. S is the crustal thickness, g is the gravitational acceleration and L is the lithosphere thickness. Once an initial crustal thickness is assigned into Eq. (2), it is possible to solve for the corresponding velocity field.

The time variation of the crustal thickness, s, is obtained from the continuity equation, expressed in the form

$$\frac{\partial s}{\partial t}=-\frac{\partial}{\partial x_1}(su_1)-\frac{\partial}{\partial x_2}(su_2) \tag{7}$$

where u_1 and u_2 are the horizontal and vertical components of velocity obtained from Eq. (2).

In order to solve for the deformation field, the equations for the horizontal deformation and for the time variation of the crustal thickness are solved numerically, one after the other, using the Finite-Element technique. A set of plain triangular elements covers the modelled region (Fig. 2a). The velocity and the crustal thickness within each element are approximated by linear polynomials and numerical integration is carried out by Gaussian quadrature with seven integration points. At each time step, we updated the mesh assuming that the nodal points are advected according to the velocity flow resulting from the integration of Eq. (2). Once the velocity field is known, the updated crustal thickness is then retrieved from Eq. (4). We have verified, in the steady state, that sphericity does not play a significant role at the characteristic dimensions of the domain under study.

The study area extends from 45° to 60° latitude and from 0° to 25° longitude and is composed of two main structures, Baltica and Avalonia, that are both geologically and rheologically different, and one purely rheological heterogeneity, the North German Basin (Fig. 2a). We carried out a series of numerical experiments which differ with regard to relative effective viscosity of the rheological heterogeneities, initial conditions, as well as to geometry of the boundary between Baltica and Avalonia. Table 1 lists all the model types considered in this study. The grey colour enlightens the from now on called reference model (MODEL 1). The bracketed numbers indicate the models that reveal the most significant results and that, for this reason, will be discussed in details in the next section. For all the models in Table 1, the same boundary conditions have been considered.

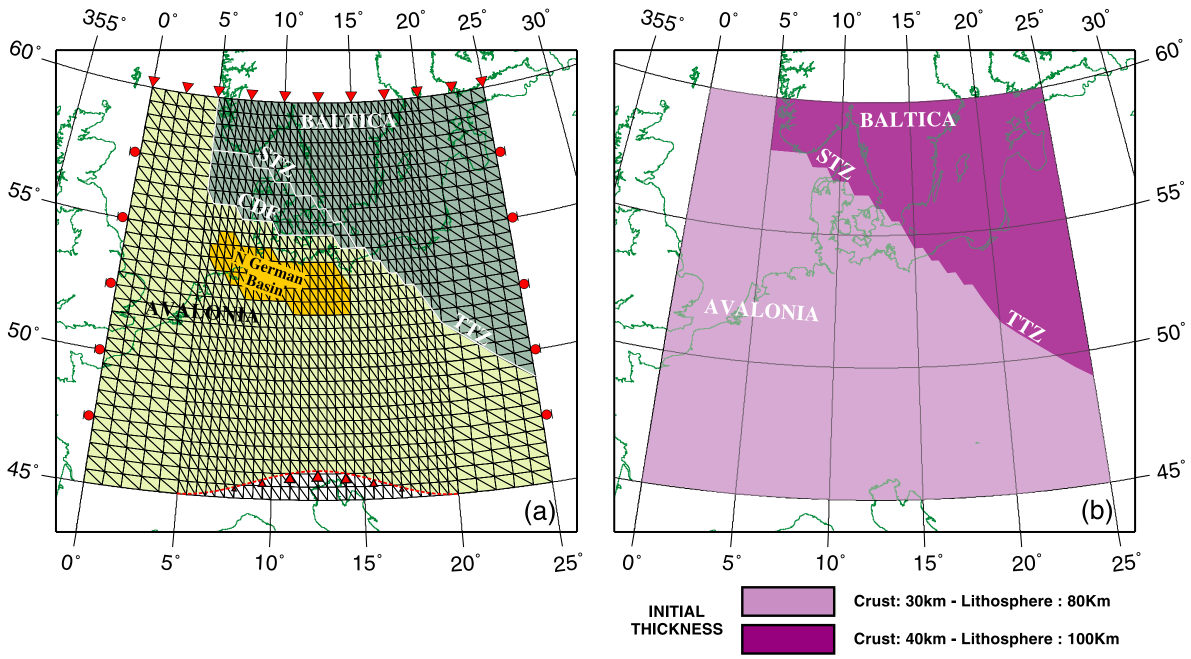

Fig. 2. (a) Finite Element Mesh and assumed boundary conditions (red symbols) for the study region. On the northern boundary of the model, no slip conditions (that is both components of velocity equal to zero) are applied. The lateral boundary is closed and free of shear. At the lower boundary (Alpine Front), we fixed both components of velocity in order to simulate the indentation of Adria towards Europe. Crustal thickness is kept fixed on all the boundaries of the model. Different colours for the elements indicate the different rheological entities considered in the analysis, as described in details in the model set-up paragraph and listed in Table 1. (b) Initial average crustal and lithospheric thickness assumed for Avalonia and Baltica.

2.1. Rheology

The model integrates lateral rheological heterogeneities matching particular geological structures in Central Europe and in the Baltic Shield. These rheological heterogeneities are implemented by assigning special values of averaged effective viscosity μ_{eff} to all the elements of the grid used to match the

Table 1
List of model types considered in the analysis

Normalized viscosity	Baltica/Avalonia boundary	Initial thickness (km)	North German Basin normalized viscosity	Model number
EUROPE = 1 BALTICA = 100	CDF	Crust = 30 Lithosphere = 80	1	(1)
			10	(2)
		Crust = 40 Lithosphere =100	1	3
			10	(4)
	STZ	Crust = 30 Lithosphere = 80	1	5
			10	(6)
		Crust = 40 Lithosphere =100	1	7
			10	8

The normalized viscosity refers to the value of $\mu = 10^{25}$ Pa s. Notation: CDF = Caledonian Deformation Front; STZ = Sorgenfrei–Tornquist Zone. The grey colour enlightens the reference model. The bracketed numbers indicate the models that will be discussed in details.

rheological heterogeneities (Fig. 2a). In all numerical experiments (models 1–8), the effective viscosity of Avalonia is assumed to be equal to 10^{25} Pa s, while Baltica is two orders of magnitude stiffer than Avalonia. For what concern the choice of a stronger rheology for the lithosphere in the North German Basin, respect to the surrounding area, it is based on recent conclusions from flexural modeling by Marotta et al. (2000) along the deep seismic profile DEKORP BASIN'96 crossing the North East German Basin (Fig. 1).

2.2. Initial conditions

The initial conditions are given in terms of crustal thickness. At the beginning of the compression phase, before the occurrence of inversion, the crustal thickness is assumed homogeneous through the two main plates that define the model (Fig. 2b). In particular, we implement two different cases. In the first one we assume an initial homogeneous crustal thickness of 30 km for the entire study region (models 1–2 and 5–6) that fits rather well the average crust in the European Plate but underestimates the thicker crustal layer seismically observed under the Baltic Shield. Although not completely in agreement with observations, addressing this simplified model allows to stress the role played by the thickness of the geological structures on the deformation and stress fields. We also implemented the case that differs for an initial crustal thickness of 40 km under the Baltic Shield (models 3–4 and 7–8), thus accounting for the significant discontinuity in crustal thickness observed between central and northern Europe (Blundell et al., 1992; Thybo, 2000, 2001; Cloething and Burov, 1996).

2.3. Geometry of Baltica/Avalonia boundary

Two scenarios were considered. In the first one the Baltica–Avalonia border is located at the Caledonian Deformation Front (Berthelsen, 1992) (Models 1–4).

Since the southern border between Baltica and Avalonia is not unambiguously defined yet (see previous section), we carried out a set of numerical experiments (Models 5–8) in which we assumed that this border is coincident in the west with the Sorgenfrei–Tornquist Zone. Although this is an extreme scenario, it maximizes the effects associated with a change of the geometry of this boundary on the local deformation and stress fields, allowing us to better enlighten them.

2.4. Boundary conditions

The boundary conditions are given in terms of velocities and crustal thickness (Fig. 2a). The northern boundary of the model is situated within the cratonic Baltic Shield and is kept fixed, i.e. the velocities are forced to be zero for all the models. The lower boundary coincides with the Alpine Front where the Adriatic Plate is forced to advance at 4 mm/year (Schmid et al., 1996, 1997; Bonini et al., 1999) as a rigid indenter toward the north, into the European plate, being responsible for the thickening of the European crustal margin. The assumption of rigid rather than deformable indenter for the advancing Adriatic Plate is not crucial for our study, since our analysis does not consider the internal Alpine deformation as part of the study. The western and eastern boundaries of the model are assumed closed and free of shear stress. For what concerns the crustal thickness, it was kept fixed at all the boundaries of the model. We are aware that this is a very strong constraint and it does not allow to predict particular crustal characteristics close to the borders of the model, such as the observed thinning of the crust in the Central Viking Graben System, near the western boundary of our model. However, this was necessary in order to assure the numerical stability of the analysis. On the other hand, the area of interest for this study is located in the centre of the model, sufficiently far from the sides of the model to assure that the main results of this study are not affected by this particular boundary condition.

3. Results and discussion

We first discuss the results of the reference model (Model 1) in which we consider an initially constant thickness of the lithosphere and the crust throughout the modeled region and we assume the southern limit of the Baltic Shield to be located at the Caledonian Deformation Front. Subsequently, we discuss the effects of a different structural configuration of the system.

Fig. 3a shows the velocity field after 35 Ma of Alpine convergence at a rate of 4 mm/year for MODEL 1 (Table 1). For this type of numerical experiments, the lithosphere and the crust are assumed to be of constant thickness, 80 and 30 km, respectively. The reference viscosity is $\mu = 10^{25}$ Pa s, with the Baltic Shield two orders of magnitude stronger than the European plate. The velocity field spreads around the indenter in a uniform way, slackening progressively to the north where its intensity is strongly reduced due to the rheologically strong Baltic Shield. Fig. 3b shows the direction of the maximum horizontal stress S_{Hmax}. Different colors indicate the stress regime (blue = thrust faulting; green = normal faulting; red = strike slip) deduced by comparing the predicted three components of principal stress, σ_1, σ_2 and σ_3. Widespread thrust faulting, dominantly in the south–north direction, characterizes Central Europe during this analysis time span, with strike-slip deformation confined to the external boundary of the Alpine Front and to the southern limit of the strong Baltic Shield, where it occurs predominantly along the boundary itself. The green region of panel 3b indicates that a predominant NNE extensional stress regime characterizes the Baltic Shield region further to the north. Fig. 3c shows the predicted vertical strain rate for the study region. Due to the compressive stress regime, Central Europe undergoes crustal thickening at an almost homogeneous rate of the order of about 1×10^{-16} s^{-1}, with deviations from this general pattern in the south, close to the sides of the Alpine indenter (Fig. 3c). The northern part of the Baltic Shield shows crustal thinning, associated with the lateral material extrusion at the southern edge of the Baltic plate, where an important lateral variation in the stiffness of the lithosphere exists.

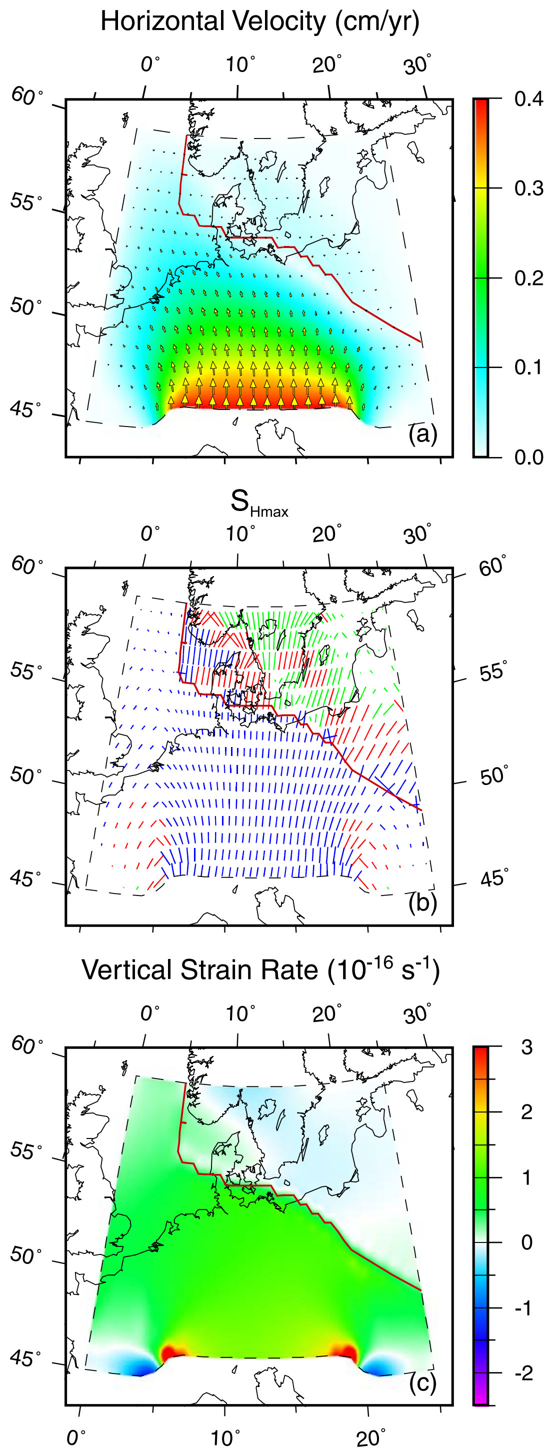

Fig. 3. Modelling results after 35 Ma of Alpine compression for Model 1. The boundary between Avalonia and Baltica is located at the Caledonian Deformation Front and a viscous uniform lithosphere is assumed below Avalonia. (a) Deformation Velocity Field. (b) Direction of S_{Hmax} and stress regime. Different colours are used to represent the stress regime, calculated by comparing the three components of principal stresses, σ_1, σ_2 and σ_3; trust faulting, blue colour; normal faulting, green colour; strike slip faulting, red colour. The length of the hyphens is proportional to the intensity of S_{Hmax}. (c) Vertical Strain Rate. Negative values indicate thinning while positive values indicate thickening.

When the North German Basin (Marotta et al., 2000) is included as a rheologically strong zone (MODEL 2—Table 1), significant differences are introduced into the deformation pattern (Fig. 4 compared to Fig. 3). The local rheological heterogeneity has the double effect to concentrate the strongest velocity gradients near the Alpine Front, while it reduces the velocity field in Central Europe, south of the North German Basin. However, a peculiar effect can be observed below the western part of the North German Basin. There, the particular rheological configuration of the system enhances a local velocity increase, but with a reverse direction with respect to the general south–north trend induced by the Alpine Front advancement in the southern part of the study region (Fig. 4a). This effect induces the biggest gradients in the crustal thickness through the study area, with significant thickening in Central Europe, in the region between the Alpine Front and the North German Basin, due to the compression. Note the localized thickening at the southern margin of the North German Basin in the region where the Hartz Mountain is geographically located. Thinning is strongly enhanced in the transition zone between the stiff North German Basin and the Baltic Shield, including the North Sea and part of the North East German Basin, due to the local velocity gradients between the Baltic Shield and the North German Basin, as shown by Fig. 4a. On the contrary, small thickening affects the northern portion of Baltica (Fig. 4b).

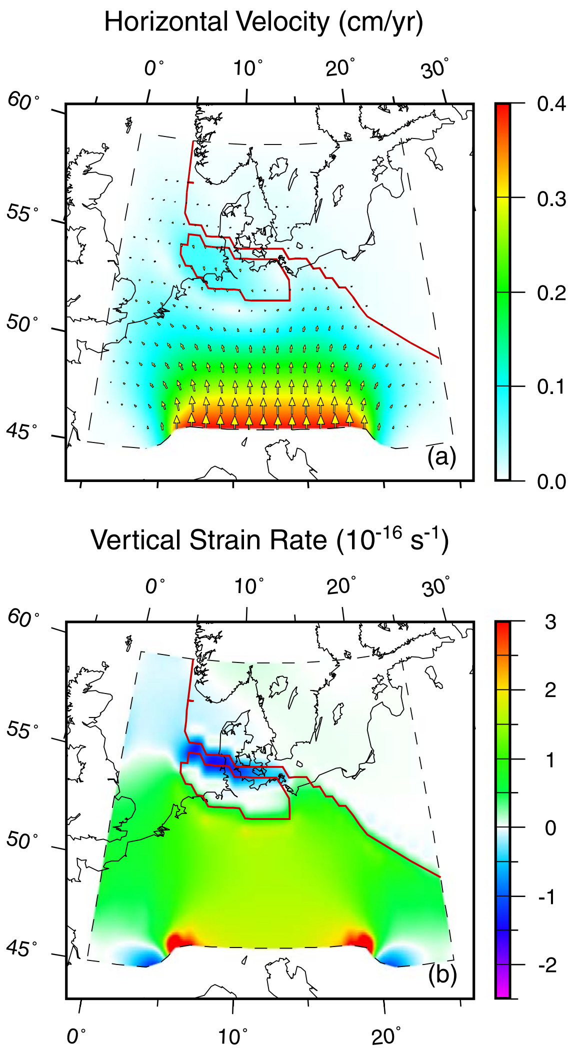

Fig. 4. Modelling results after 35 Ma of Alpine compression for Model 2. The boundary between Avalonia and Baltica is located at the Caledonian Deformation Front and a rheologically strong inclusion represents the North German Basin. (a) Deformation Velocity Field. (b) Vertical Strain Rate. Negative values indicate thinning while positive values indicate thickening. See text for discussion.

Fig. 5 shows the regional stress regime at the beginning of the Alpine Orogeny and after 35 Ma of compression for MODEL 2. While the areas close to the Alpine Front remain under a general compressive regime, several primary features can be distinguished in the stress field predicted for the northern part of the modelled region, with respect to the homogeneous model showed in Fig. 3. The relative intense, south-directed local flow below the North German Basin prevents the propagation of thrust faulting toward the southern border of the Baltic Shield as it occurs in the homogeneous model. In this model, the area between the Sorgenfrei–Tornquist Zone and the North German Basis exhibits a predominant extensional stress regime, while a strike slip characterizes the surrounding region. In conclusion, the effect of the rheological gradients in the area between the strong North German Basin and the strong Baltic Shield is to reduce and to restrain the areas where the deformation is accommodated by normal faulting to the south, and to enhance strike slip faulting along two main regions which coincide geographically with the Elbe Line in the south and the Sorgenfrei–Tornquist Zone and the Teysseire–Tornquist Zone in the north. It is worth to

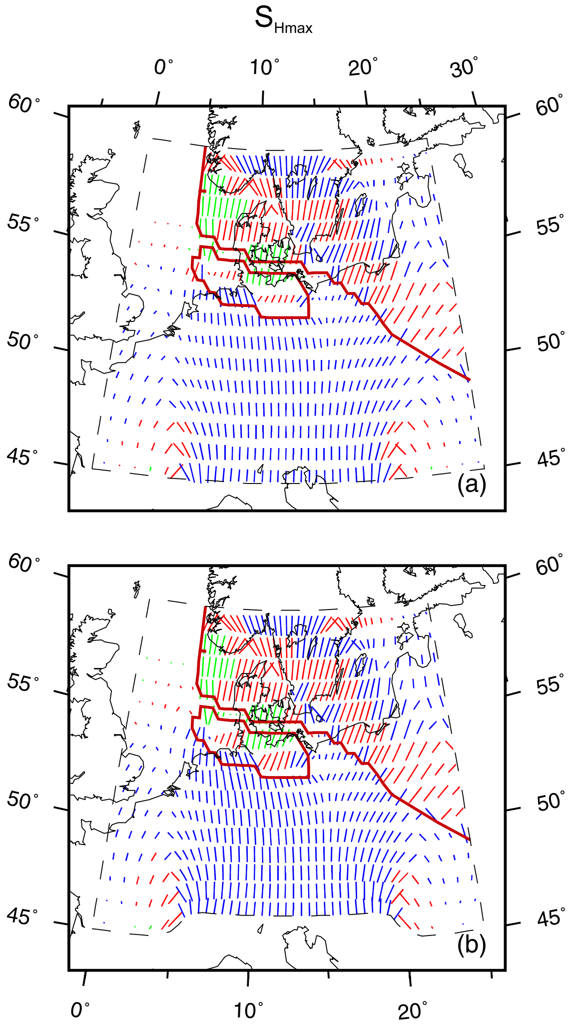

Fig. 5. Predicted direction of S_{Hmax} and stress regime at 0.1 Ma (a) and 35 Ma (b) for Model 2, the same as in Fig. 4. The boundary between Avalonia and Baltica is located at the Caledonian Deformation Front and a rheologically strong inclusion represents the North German Basin. The same notation as in Fig. 3 is used to represent the stress field.

mention that the viscosity heterogeneity of the North German Basin induces a progressive rotation in the direction of the principal stress axes from S–N/NW in

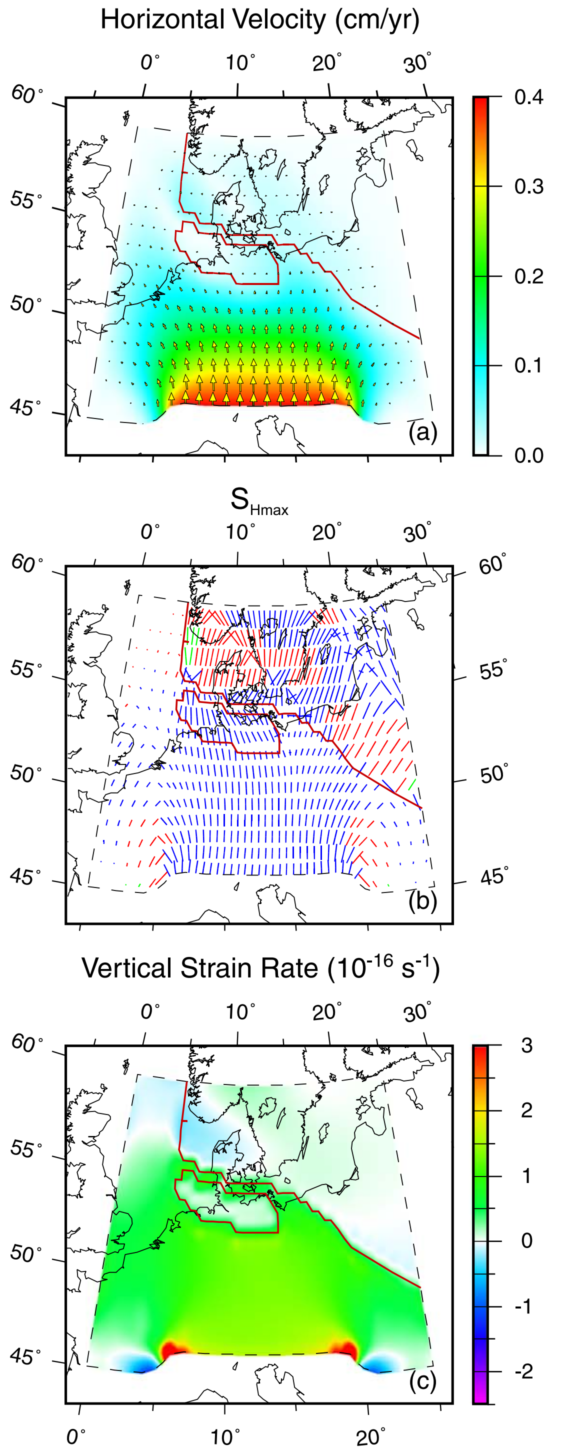

Fig. 6. Modelling results as in Fig. 3 but for Model 6. Different initial crustal and lithosphere average thicknesses for the Avalonia and Baltica plates are considered. (a) Deformation Velocity Field. (b) Direction of S_{Hmax} and stress regime. The same notation as in Fig. 3 is used to represent the stress field. (c) Vertical Strain Rate. Negative values indicate thinning while positive values indicate thickening.

the western side to S–N/NE in the east side of the study region, thus causing a fan-like stress pattern in North Central Europe. This result is in agreement with previous results obtained by means of a similar approach by England and Houseman (1985) for the Tibetan plateau and by Marotta et al. (2001) who modeled the effects induced by the Alpine collision on the deformation at a smaller scale, below the North German Basin.

Fig. 6 shows the results for MODEL 4 that differ from MODEL 2, shown in Figs. 4 and 5, because different values of initial average crustal and lithosphere thickness are considered for the northern Baltica and Avalonia, as indicated in Fig. 2b. The deformation pattern in the area extending from the North German Basin and the Sorgenfrei–Tornquist Zone is remarkably different from that predicted by MODEL 2. With respect to Fig. 5a, the southerly directed velocity region is displaced to the north (Fig. 6a) due to the deviatoric stressed set up by lateral thickness variations in the crust. The main consequence is that transpressional deformation characterizes the entire area north of the North German Basin, with crustal thickening predicted at the northern border of the North German Basin (Fig. 6a), instead of the thinning predicted in MODEL 2 and shown in Fig. 4b. The elongated region predicted to be undergoing thickening is slightly south of the Rømø Fracture Zone (Cartwright, 1990), where a crustal thickness remarkably higher than the surrounding region is observed in the seismic data (Thybo, 1997). The thinning in the North Sea and the thickening in the Baltic Shield are intensified. These features agree with the observed subsidence in the North Sea and uplift of the Baltic Shield (Ziegler, 1990). However, MODEL 4 does not predict the observed deformation pattern in the North German Basin as well as MODEL 2, presented in Figs. 4 and 5. No significant changes are predicted in the stress and deformation fields in the region between the Alpine Front and the North German Basin, with

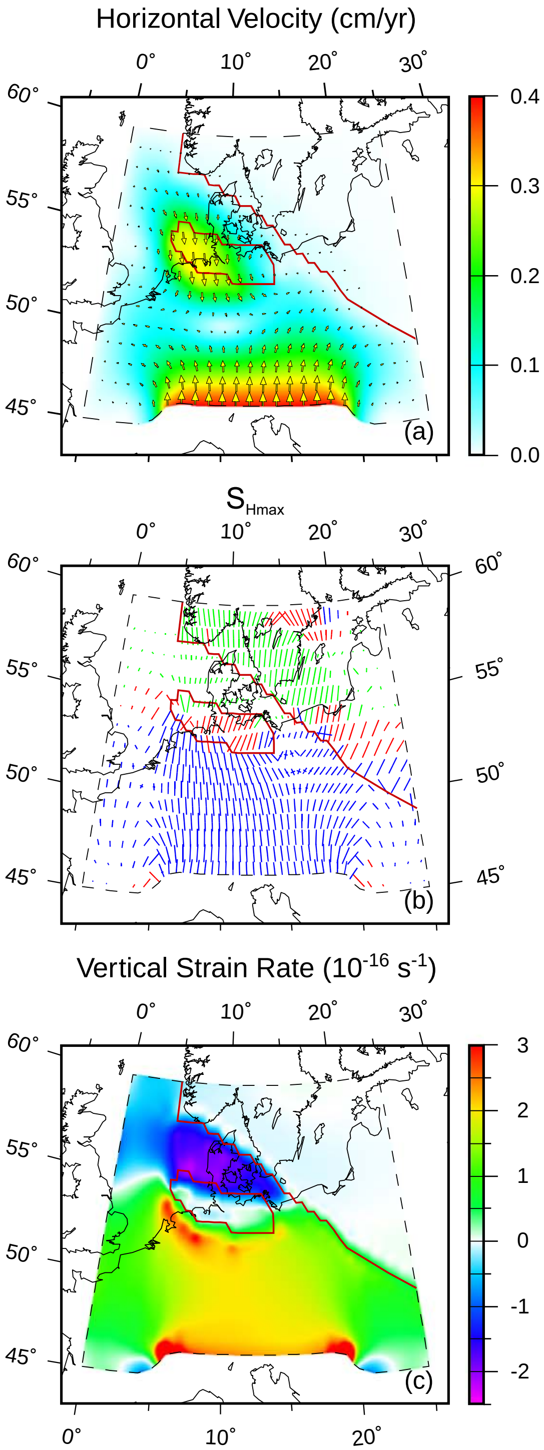

Fig. 7. Modelling results as in Fig. 3 but for Model 8. The boundary between Avalonia and Baltica is now located at the STZ. (a) Deformation Velocity Field. (b) Direction of S_{Hmax} and stress regime. The same notation as in Fig. 3 is used to represent the stress field. (c) Vertical Strain Rate. Negative values indicate thinning while positive values indicate thickening.

respect to MODEL 2, although the crustal thickness gradients decrease due to the accommodation of compression over a wider region.

Fig. 7 shows the results for Model 6. The deformation pattern shows that the velocity field in the western part of the rheologically strong North German Basin is significantly intensified due to the wider softer fan-like region between the basin and the Sorgenfrei–Tornquist Zone (Fig. 7a). The intensification in this local south-directed velocity field induces a remarkable increase of the maximum compressive stress south of the strong North German Basin (Fig. 7b). Consequently, remarkable crustal thickness variations are now predicted in Central Europe, with inhibited thickening below the strong basin, in contrast with the increased thickening between the Alpine Front and the North German Basin. In addition, thinning is enhanced in the softer region between the strong North German Basin and the Baltic Shield (Fig. 7c). However, the intensification in the local south-directed velocity field in the north eastern part of the model can be so large that it causes in the Baltic Shield a tectonic regime opposite to that predicted when the border between Baltica and Avalonia is located at the CDF, a feature that is in contrast with the observations (Ziegler, 1990). The results of MODEL 6 seem to indicate that the Sorgenfrei–Tornquist Zone is not a suitable candidate as border between Baltica and Avalonia, and that this border should be located at one of the geological structures further to the south, as supported by MODEL 2.

4. Conclusions

We have used a viscous thin sheet model to study the tectonic regime at the border between Baltica and Avalonia during the beginning of the Alpine Collision. We have demonstrated that the geometry of the boundary between Baltica and Avalonia has a crucial role on the predicted local crustal deformation and faulting regime resulting in the area extending from the North German Basin to the Sorgenfrei–Tornquist Zone from the S–N Alpine convergence, provided that a rheologically strong North German Basin exists to transmit the compressive stresses from the south to the north. The predictions of modeling approach are in agreement with the observations when:

(1) The Baltica–Avalonia border is located in proximity of the Caledonian Deformation Front.
(2) Baltica and Avalonia have different rheological characteristics.
(3) Different average crustal and lithospheric thicknesses are taken into account.

In particular, both MODEL 2 and MODEL 4 succeed to predict the observed deformation field, such as the subsidence in the North Sea and the uplift of the Baltic Shield, while MODEL 4 does not predict the observed deformation pattern in the North German Basin as well as MODEL 2.

The modeled orientation of the principal horizontal compressive stresses is in agreement with the present-day stress pattern observed in North Central Europe (Zoback, 1992; World Stress Map, 2000) and with previous studies (Marotta et al., 2001).

Acknowledgements

This work received financial support from the Italian Ministry for the University and Research (project: 'Active deformation at the Northern boundary of Adria'). The authors would like to thank Ulf Bayer for his valuable comments to the first draft of the manuscript and the reviewers Dr. Ole Rønø Clausen and the anonymous reviewer for their comments and suggestions. All illustrations were made using GMT of Wessel and Smith (1998).

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