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Intraplate deformation and 3D rheological structure of the Rhine Rift System and adjacent areas of the northern Alpine foreland

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Abstract The lithosphere of the Northern Alpine foreland has undergone a polyphase evolution during which interacting stress-induced intraplate deformation and upper mantle thermal perturbations controlled folding of the thermally weakened lithosphere. In this paper we address relationships among deeper lithospheric processes, neotectonics and surface processes in the Northern Alpine foreland with special emphasis on tectonically induced topography. We focus on lithosphere memory and neotectonics, paying special attention to the thermo-mechanical structure of the Rhine Graben System and adjacent areas of the northern Alpine foreland lithosphere. We discuss implications for mechanisms of large-scale intraplate deformation and links with surface processes and topography evolution.

Keywords Intraplate deformation · 3D rheological structure · Rhine Graben · Alpine foreland

Introduction

During Mesozoic and Cenozoic times, the lithosphere of the Alpine foreland has undergone repeated tectonic reactivation (Ziegler 1989a; Ziegler et al. 1995, 1998) and is indeed even at present still being deformed, as evidenced by significant intraplate seismicity and on-going differential vertical motions controlling the development of dynamic topography at large distances from plate boundaries (Cloetingh et al. 2003b) (Fig. 1). Increasing evidence is accumulating for widespread Neogene uplift and tectonics around the northern Atlantic (e.g. Japsen and Chalmers 2000; Chalmers and Cloetingh 2000), that are accompanied by accelerations of subsidence and sedimentation rates (Cloetingh et al., 1990), commencing too early to be attributed exclusively to the effects of climatic changes and post-glacial rebound of the North Atlantic borderlands. Results of geothermochronologic studies provide constraints on the timing and magnitude of uplift of the North Atlantic borderlands and document a polyphase record that commenced as early as in Oligocene times, interpreted as resulting from a combination of upper mantle perturbations, intraplate stresses and post-glacial rebound (Hendriks BWH et al. submitted).

Over the last few years, seismicity studies and geomorphologic evidence from Brittany (Bonnet et al. 2000), Normandy and the Channel and Dover Street areas (Lagarde et al. 2000; Van Vliet-Lanoë et al. 2000), southern England (Preece et al. 1990), the Ardennes-Eifel region (Demoulin et al. 1995; Meyer and Stets 1998; Van Balen et al. 2000), the Upper Rhine Graben (URG) (Nivière and Winter 2000) and the North German Basin (Ludwig 1995) demonstrate the important contribution of neotectonics to the topographic evolution of intraplate Europe (Fig. 1). Regional Neogene exhumation of the British Isles and the margins of the North Sea Basin (Japsen 1997) are documented by the analysis of petroleum industry data. Combined studies of geomorphology and the record of sedimentary basins, integrating results of subsidence analyses and geothermochronologic data (e.g. Ter Voorde et al. 2004), also demonstrate that neotectonics has a strong bearing on the development of topography and drainage patterns.

The origin of intraplate stress fields in continental lithosphere and their relationship to plate-tectonic driving forces has been the subject of a large number of observational (e.g., Van der Pluim et al. 1997; Marotta

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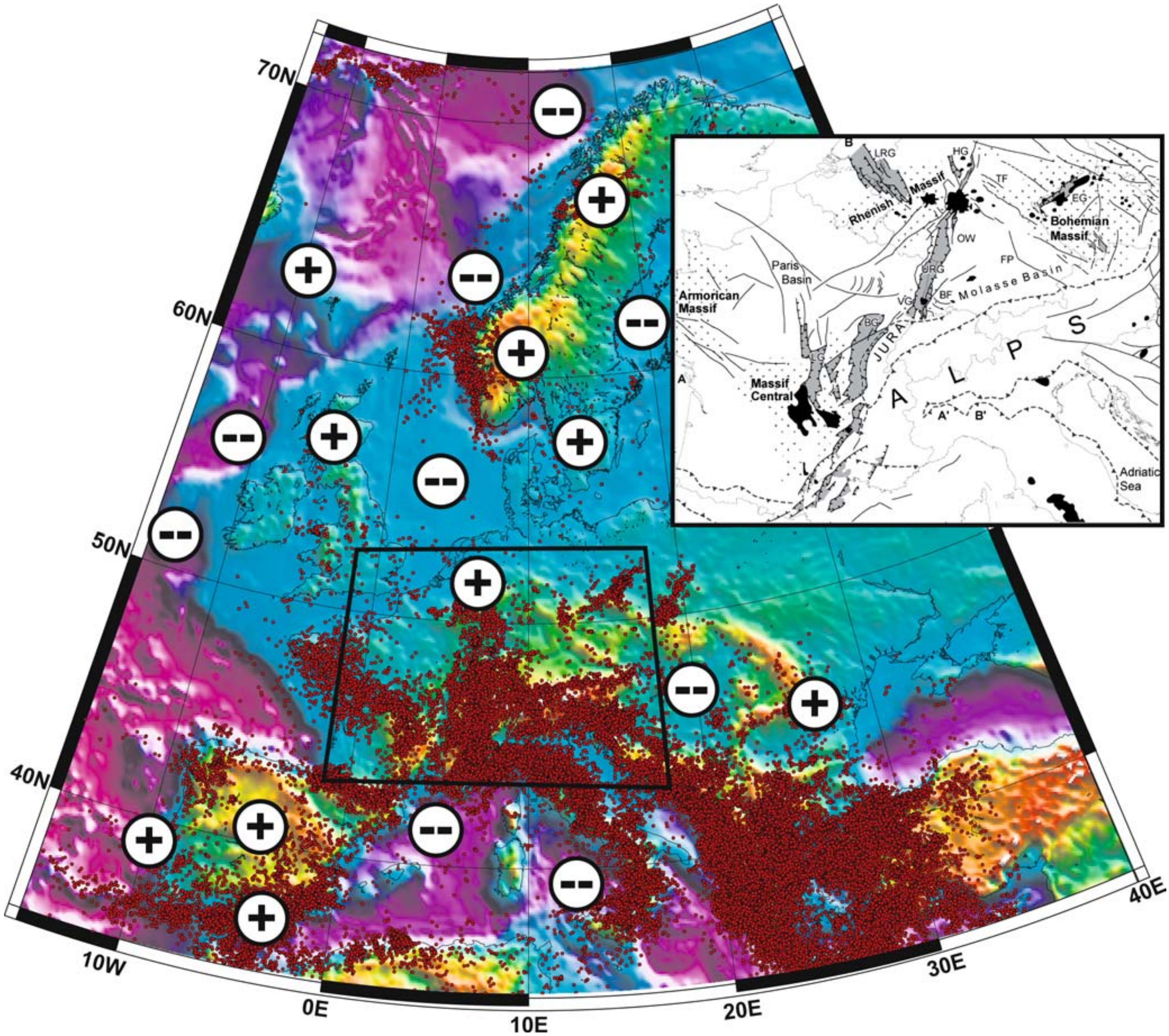

Fig. 1 Topographic map of Europe with superimposed distribution of seismicity (red dots), illustrating present-day active intraplate deformation. Also shown are intraplate areas of Late Neogene uplift (circles with plus symbols) and subsidence (circles with minus symbols). Background elevation image is extracted from the ETOPO2 data set. Earthquake epicenters were obtained from the NEIC data center. INSET: Location map of the European Continental Rift System (ECRIS) in the Alpine and Pyrenean foreland, showing Cenozoic fault systems (black lines), rift-related sedimentary basins (light gray), Variscan massifs (cross-pattern) and volcanic fields (black). Solid barbed line: Variscan deformation front; stippled barbed line: Alpine deformation front. BF Black Forest, BG Bresse Graben, EG Eger (Ohre) Graben, FP Franconian Platform; HG Hessian grabens, LG Limagne Graben, LRG Lower Rhine (Roer Valley) Graben, URG, OW Odenwald; VG Vosges (after Dèzes et al. 2004)

et al. 2001) and modeling studies (e.g. Bada et al. 1998; Gölke and Coblentz 1996). These have revealed the existence of consistently oriented first-order patterns of intraplate stress in, for example, the Northwest European platform (Fig. 2) and the North American craton. The effect of these stresses on vertical motions of the lithosphere, expressed in terms of, for example, apparent sea-level fluctuations (Cloetingh et al. 1985), development of foreland bulges (Ziegler et al. 2002), basin inversion (Ziegler et al. 1995, 1998) and lithosphere folding (Martinod and Davy 1994; Cloetingh et al. 1999), has been demonstrated to be an important element in the dynamics of continental interiors (Cloetingh 1988; Van der Pluim et al. 1997).

Stress propagation occurs in a lithosphere that can be significantly weakened by inherited structural discontinuities, but also by upper mantle thermal perturbations (e.g. Goes et al. 2000a, 2000b). Below we present first-order scale thermo-mechanical models for large-scale intraplate deformation, and discuss constraints on these models inferred from different studies carried out during the last few years on the rheology of the Northwest

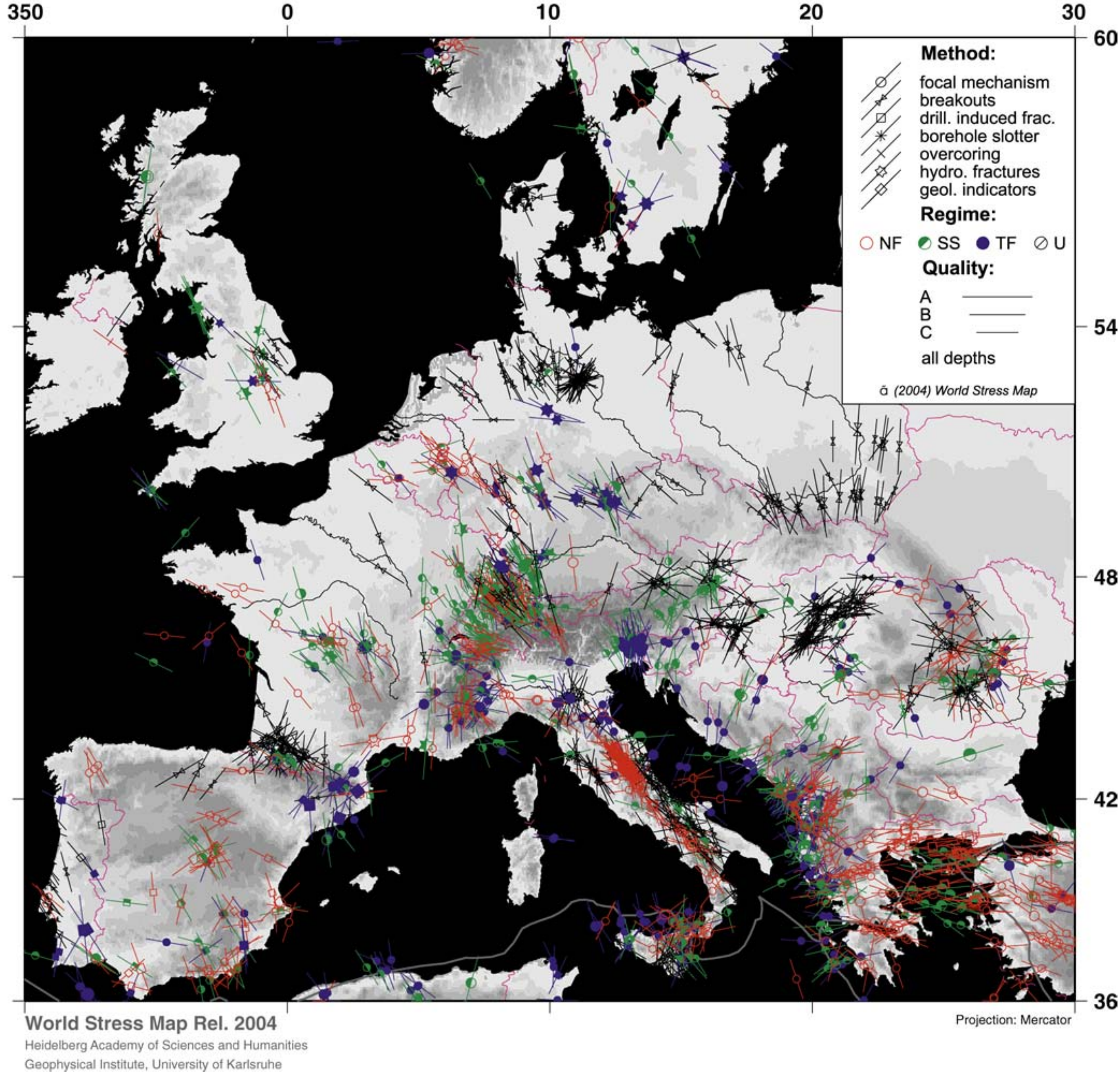

Fig. 2 Intraplate stress map for Europe, displaying the present-day orientation of the maximum horizontal stress (S_{Hmax}). Different symbols stand for different stress indicators, the length of the symbols represents the data quality, 'A' being of highest quality. Background shading indicates topographic elevation (darker is higher). Stress map is extracted from the World Stress Map database (http://www.world-stress-map.org/)

European foreland with a focus on the Rhine Graben segment of the European Cenozoic rift system (ECRIS).

Structure of Europe's intraplate lithosphere

European Cenozoic rift system (ECRIS)

Development of the presently still active ECRIS (Fig. 1a), which extends from the Dutch North Sea coast to the western Mediterranean, commenced during the late Eocene (Dèzes et al. 2004). Its southern elements are the Valencia Trough, the graben systems of the Gulf of Lions, and the northerly striking Valence, Limagne and Bresse grabens; the latter two are superimposed on the Massif Central and its eastern flank, respectively. These grabens are linked via the Burgundy transfer zone to the northerly striking URG that bifurcates northwards into the northwest trending Roer Valley Graben and the north-easterly trending Hessian Grabens, which transect the Rhenish Massif. The northeast striking Eger Graben, which transects the Bohemian Massif, forms an integral part of ECRIS (Ziegler 1994). Localization of ECRIS involved the reactivation of Permo-Carboniferous and Mesozoic shear systems. Although the on-shore parts of ECRIS are characterized by relatively low crustal stretching factors, they are associated with a distinct uplift of the crust–mantle boundary. To what extent this can be attributed to Cenozoic rifting or whether Permo-Carboniferous processes have contributed to it, remains an open question (Dèzes et al. 2004; Ziegler and Dèzes 2005). Evolution of ECRIS was accompanied by the development of major volcanic centers in Iberia, on the Massif Central, the Rhenish Massif and the Bohemian Massif, particularly during Miocene and Plio-Pleistocene times. Seismic tomography indicates that mantle plumes well up beneath the Massif Central and the Rhenish Massif (Granet et al. 1995; Ritter et al. 2001), but not beneath the Vosges-Black Forest arch. Despite this, the evolution of ECRIS is considered to be a clear case of passive rifting (Dèzes et al. 2004; Ziegler and Dèzes this volume).

Under the present northwest directed stress field (Fig. 2), which came into evidence during the Miocene and intensified during the Pliocene, the URG is subjected to sinistral shear, and the Roer Valley Graben is under active extension (Dèzes et al. 2004), whereas the North Sea Basin experiences a phase of accelerated subsidence that can be related to stress-induced deflection of the lithosphere (Van Wees and Cloetingh 1996). Similarly, continued uplift of Fennoscandia is thought to be controlled by folding of the lithosphere under the present stress field that reflects a combination of Atlantic ridge push and collisional coupling of the Alpine orogen with its foreland (Gölke and Coblentz 1996).

Constraints on crustal and upper mantle structure

Ziegler and Dèzes (2005) have compiled the results of crustal studies that were carried out since the publication of Moho depth maps by Meissner et al. (1987), Ziegler (1990) and Ansorge et al. (1992) to obtain a better understanding of the present day crustal configuration of Western and Central Europe, and to analyze processes and their timing, which controlled the evolution of the crust in the different parts of Europe.

During the last decade, major improvements have been made in global travel time tomography. A new model parameterization technique and new 3D ray tracing algorithms (Bijwaard and Spakman 1999a) resulted in global mantle models that, for the first time, exhibit regional scale (60–100 km) detail (Bijwaard and Spakman 2000). Improved focusing on lower mantle structures led to the first evidence for a whole mantle plume below Iceland (Bijwaard and Spakman 1999b) and for up-welling of the lower mantle beneath Europe, which is proposed as underlying the Cenozoic Central European volcanism (Goes et al. 1999).

Goes et al. (2000a) developed a new inversion strategy that takes advantage of the fact that most of the upper mantle seismic wave velocity variations are caused by temperature variations. State-of-the-art seismological models and experimental data on the physical properties of mantle rocks were inverted for upper mantle temperatures, e.g. beneath Europe. Inferred mantle temperatures beneath Europe agree reasonably well with independent estimates from heat flow and general geological considerations (e.g. Dunai and Baur 1995).

Beneath ECRIS, mantle tomography images a system of upper asthenospheric low velocity anomalies that are interpreted as plume heads that have spread out above the 410 km discontinuity (Goes et al. 1999; Spakman and Wortel 2004; Sibuet et al. 2004). From these anomalies secondary plumes appear to rise up beneath the Eifel (Ritter et al. 2001), the Massif Central (Granet et al. 1995) and possibly the Bohemian Massif (U. Achauer, pers. com. 2003) but not beneath the Vosges-Black Forest arch (Achauer and Masson 2002). For instance, the upper mantle plume that rises up beneath the Massif Central (Granet et al. 1995) has a diameter of 100–300 km and involves material that is 100–200°C hotter than the ambient mantle. These and similar findings for the Eifel region (Ritter et al. 2001) support diapiric upwelling of small scale, finger-like convective instabilities from the deeper seated upper mantle anomalies. The latter presumably developed during the Paleocene, following activation of the Northeast Atlantic and Iceland mantle plumes, which rise up from the core–mantle boundary (Hoernle et al. 1995; Bijwaard and Spakman 1999b), and subsequently evolved further. This is compatible with volcanic activity in the ECRIS area that commenced during the Paleocene and persisted into the Quaternary (Wilson and Patterson 2001; Dèzes et al. 2004). As through time a shift in areas of main volcanic activity can be observed, it is likely that the supply of partial melts through secondary upper mantle plumes was not steady state but pulsated and entailed a shift in their location. In the ECRIS area this plume activity caused thermal weakening of the lithosphere, thus rendering it prone to deformation, but was not the driving mechanism of rifting. ECRIS appears to have evolved in response to passive rifting, mainly controlled by compressional stresses originating at the Alpine and Pyrenean collision zones (Achauer and Masson 2002; Dèzes et al. 2004).

Strength and deformation mode of Europe's intraplate lithosphere

Lithospheric strength: basic concepts

The strength of continental lithosphere is controlled by its depth-dependent rheological structure (Fig. 3) in which the thickness and composition of the crust, the thickness of the mantle–lithosphere, the potential tem-

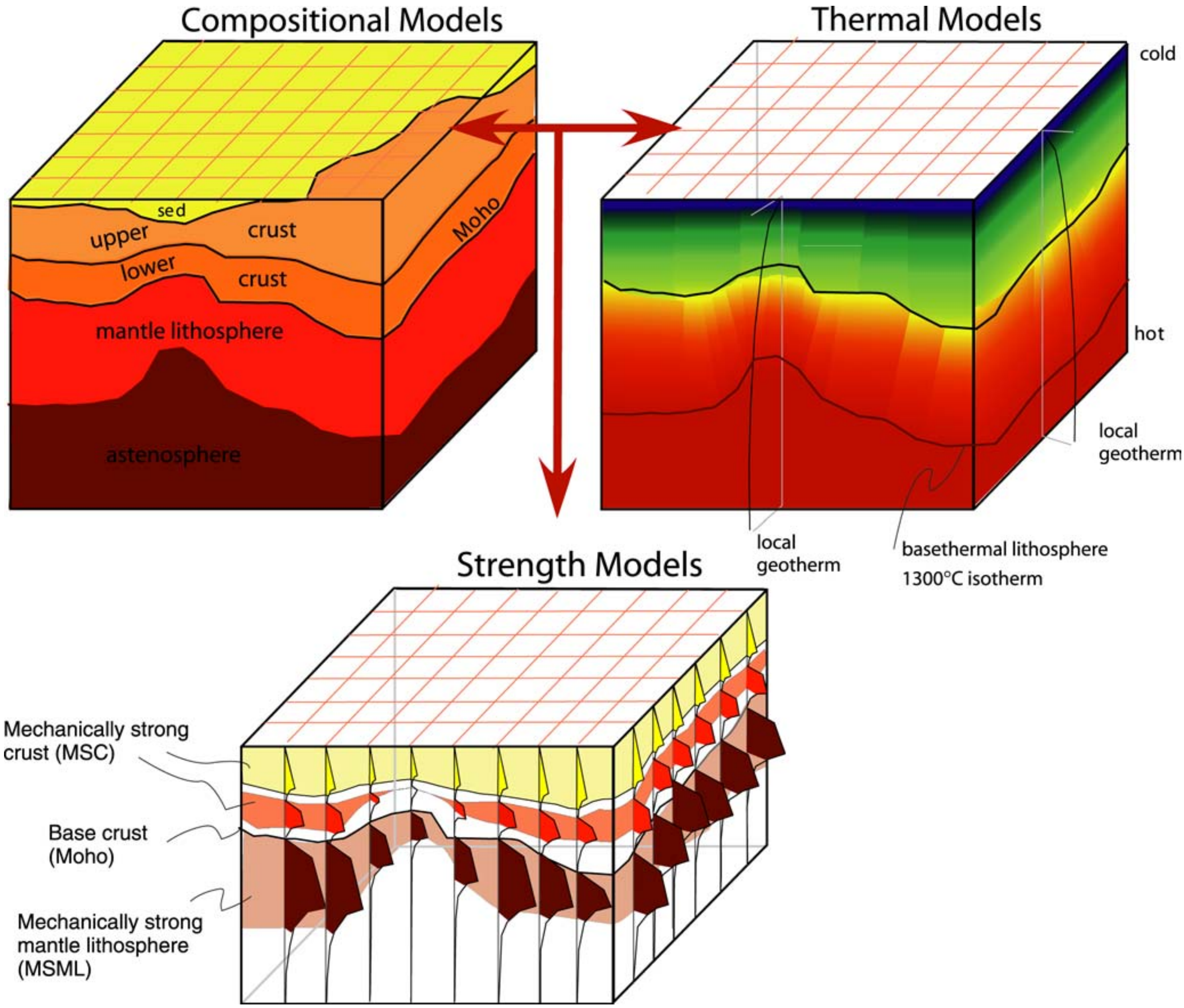

Fig. 3 From crustal thickness (top left) and thermal structure (top right) to lithospheric strength (bottom): conceptual make-up of the thermal structure and composition of the lithosphere, adopted for the calculation of 3D strength models

perature of the asthenosphere, the presence or absence of fluids, and strain rates play a dominant role (e.g. Carter and Tsenn 1987; Kirby and Kronenberg 1987). By contrast, the strength of oceanic lithosphere depends on its thermal regime, which controls its essentially age-dependent thickness (Panza et al. 1980; Kusznir and Park 1987; Stephenson and Cloetingh 1991; Cloetingh and Burov 1996).

Theoretical rheological models indicate that thermally stabilized continental lithosphere consists of the mechanically strong upper crust, which is separated by a weak lower crustal layer from the strong upper part of the mantle–lithosphere, which in turn overlies the weak lower mantle–lithosphere. By contrast, oceanic lithosphere has a more homogeneous composition and is characterized by a much simpler rheological structure. Rheologically speaking, thermally stabilized oceanic lithosphere is considerably stronger than all types of continental lithosphere. Atlantic-type continental margins mark the transition from oceanic to continental lithosphere, and are the sites of thinned continental lithosphere that was extended and heated during continental breakup. This has led to substantial lateral variations in the mechanical strength of the lithosphere that are controlled by complex variations in crustal thickness, composition of the lithospheric layers, and the thermal regime.

The strength of continental crust depends largely on its composition, thermal regime and the presence of fluids, and also on the availability of pre-existing crustal discontinuities. Deep-reaching crustal discontinuities, such as thrust- and wrench-faults, cause significant weakening of the otherwise mechanically strong upper parts of the crust. As such discontinuities are apparently characterized by a reduced frictional angle, particularly in the presence of fluids, and they are prone to reactivation at stress levels that are well below those required for the development of new faults.

Extension of stabilized continental crustal segments precludes ductile flow of the lower crust and faults will be steep to listric and propagate towards the hanging wall, i.e. towards the basin centre (Bertotti et al. 2000). Under these conditions, the lower crust will deform by distributing ductile shear in the brittle–ductile transition domain. This is compatible with the occurrence of earthquakes within the lower crust and even close to the Moho (e.g. URG: Bonjer 1997; East African rifts: Shudofsky et al. 1987). In young orogenic belts, which are characterized by a crustal thickness of up to 60 km and an elevated heat flow, the mechanically strong part of the crust is thin and the mantle–lithosphere is also weak. Extension of this type of lithosphere can involve ductile flow of the lower and middle crust along pressure gradients away from areas lacking upper crustal extension towards zone of major extensional unroofing of the upper crust, involving the development of core complexes (Bertotti et al. 2000).

The strength of the continental mantle–lithosphere depends to a large extent not only on the thickness of the crust but also on its age and thermal regime. Generally, the upper mantle of thermally stabilized, old cratonic lithosphere is considerably stronger than the strong part of its upper crust (e.g. Moisio et al. 2000). However, the occurrence of upper mantle reflectors, which generally dip in the same direction as the crustal fabric and probably relate to subducted oceanic and/or continental crustal material, suggests that the continental mantle–lithosphere is not necessarily homogenous but can contain lithological discontinuities that enhance its mechanical anisotropy (Vauchez et al. 1998; Ziegler et al. 1995, 1998). Such discontinuities, consisting of eclogitized crustal material, can potentially weaken the strong upper part of the mantle–lithosphere. These factors contribute to weakening of former mobile zones to the end that they present rheologically weak zones within a craton, as evidenced by their preferential reactivation during the breakup of Pangea (Ziegler 1989b; Janssen et al. 1995; Ziegler et al. 2001).

From a rheological point of view, the thermally destabilized lithosphere of tectonically active rifts, as well as of rifts and passive margins that have undergone only a relatively short post-rift evolution (e.g. 25 Ma), is considerably weaker than that of thermally stabilized rifts and unstretched lithosphere (Ziegler et al. 1998; Ziegler and Cloetingh 2004). In this respect, it must be realized that during rifting, progressive mechanical and thermal thinning of the mantle–lithosphere and its substitution by the upwelling asthenosphere is accompanied by a rise of the geotherms, causing progressive weakening of the extended lithosphere. In addition, permeation of the lithosphere by fluids causes its further weakening. Upon decay of the rift-induced thermal anomaly, rift zones are rheologically speaking considerably stronger than unstretched lithosphere. However, thermal blanketing through the accumulation of thick syn- and post-rift sedimentary sequences can cause a weakening of the strong parts of the upper crust and mantle–lithosphere of rifted basins (Stephenson 1989). Moreover, as faults permanently weaken the crust of rifted basins, they are prone to tensional as well as compressional reactivation (Ziegler et al. 1995, 1998, 2001, 2002; Ziegler and Cloetingh 2004). The reactivation potential of such discontinuities depends essentially on their orientation with respect to the prevailing stress field (Ziegler et al. 1995; Brun and Nalpas 1996).

In view of its rheological structure, the continental lithosphere can be regarded under certain conditions as a two-layered visco-elastic beam (Reston 1990; Ter Voorde et al. 1998). The response of such a system to the build-up of extensional and compressional stresses depends on the thickness, strength and spacing of the two competent layers, on stress magnitudes and strain rates and the thermal regime (Zeyen et al. 1997). As the structure of continental lithosphere is also areally heterogeneous, its weakest parts start to yield first once intraplate stress levels equate their strength (e.g., Brun 2002; Handy and Brun 2004). In this respect, the presence of crustal and mantle–lithospheric discontinuities, which can significantly reduce the strength of the lithosphere, play an important role.

On the other hand, oceanic lithosphere behaves as a single-layer beam that is thinner than the competent parts of thick cratonic continental lithosphere. However, in view of the high strength of the mature oceanic lithosphere, its deformation requires considerably higher stress levels than the deformation of continental lithosphere (Cloetingh et al. 1989). This suggests that tectonic stresses transmitted through mature oceanic lithosphere can be large enough to cause failure of the continental lithosphere forming part of the same plate, without at the same time causing deformation of the oceanic lithosphere (Ziegler et al. 1998, 2002).

Regarding the localization of rift zones, the strength of the mechanically strong upper part of the mantle–lithosphere plays an important role (Ziegler and Cloetingh 2004). Furthermore, lateral thickness heterogeneities of the lithosphere can play an important role in the localization of rifts (e.g. Oslo Graben: Pascal et al. 2002). On the other hand, at the scale of the crust, the width and deformation mode of evolving rifts depend on the thickness of the mechanically strong upper parts of the crust and on the availability of pre-existing crustal discontinuities, which can be tensionally reactivated under the prevailing stress field.

Evidence for tensional reactivation of rifts, which had been abandoned millions of years ago, suggests that crustal-scale faults permanently weaken the lithosphere to the degree that rifts are prone to tensional and compressional reactivation (Ziegler et al. 1995, 2001, 2002). Rifted basins are marked by a relatively low crustal strength throughout their syn- and post-rift evolution.

By contrast, the strength of the mantle–lithosphere is strongly reduced during rifting, followed by a steady increase due to post-rift cooling, ultimately leading to a configuration in which a strong basin is flanked by weaker margins (see Cloetingh et al. 2003a). Models adopting zero strength for the continental mantle (e.g. Jackson 2002; see also Watts and Burov 2003) predict on-going deformation in the central region of rifted basins, which in this case are the weakest part of the basinal system. These findings are, however, incompatible with observations (see Cloetingh et al. 2003a; Cloetingh and Van Wees 2005). Intraplate stresses (Cloetingh and Burov 1996), fluids and shear zones (Handy and Brun 2004) may further reduce the actual strength of the lithosphere, but they do not affect the first order patterns of the inferred temporal and spatial evolution of lithospheric strength in the European foreland.

Model setup

Strength profiles and lithospheric effective elastic thicknesses have been calculated over the last few years for a number of locations in Europe (e.g. Cloetingh and Burov 1996). Most of these strength profiles and estimates of integrated strength were calculated along available deep seismic crustal cross sections, such as the European Geotraverse (Cloetingh and Banda 1992) and the TransAlp deep seismic profile (Willingshofer and Cloetingh 2003). So far, lithospheric strength maps have been calculated for restricted areas of Europe only, including the Pannonian Basin-Carpathian region (Lankreijer et al. 1999) and the Baltic shield (Moisio et al. 2000), but are not available on a regional scale for intraplate Europe.

Drawing on the newly compiled Moho map of Europe of Dèzes and Ziegler (2004), on constraints on the thermal lithospheric structure from heat flow studies and estimates of the lithospheric thickness from seismological studies (Plomerova et al. 2002), we constructed a 3-dimensional strength map for the lithosphere of a large part of Europe. The underlying strength model is based on a 3D multi-layer compositional model (Table 1) including one upper mantle layer, two to three crustal layers and a sedimentary cover layer (Fig. 3) (Hardebol et al. 2003).

The temperature structure of the lithosphere below Europe inferred from seismic tomography (Goes et al. 2000a, 2000b) has only limited resolution in the mechanically strong part of the lithosphere. Therefore, in this study the temperatures in the lithosphere were calculated analytically, using Fourier's law for heat conduction. Thermal rock properties (listed in Table 1) were taken from Cloetingh and Burov (1996), whereas the thermal boundary conditions were extracted from Babuska and Plomerova (1992, 1993) and Plomerova et al. (2002), or, where available, from higher quality regional or local studies. A comparison of the calculated thermal cube with temperature structures inferred from seismic tomography studies of upper mantle below Europe (Goes et al. 2000a, 2000b) shows a first order overall agreement at depths corresponding to the lithosphere-asthenosphere boundary, as further discussed below.

Model resolution

In several areas of Europe the rheological structure of the lithosphere will deviate locally and/or regionally from our first order 3D model, and consequently will affect the estimated strength to some degree. For instance, local variations in crustal composition and crustal architecture (e.g. caused by faults offsetting parts of the crust) were not incorporated in our model, as these hardly affect the first order patterns of integrated strength. On the other hand, the orogenic zones of the Alps and Pyrenees, with a substantially thickened and complex crustal architecture, as well as areas close to

Table 1 Rheological model parameters (Carter and Tsenn 1987; Brace and Kohlstedt 1980)

Parameter	Symbol	Units	Sediment	Upper crust	Lower crust	Upper mantle
Composition			Various	Quartzite (dry)	Diorite (wet)	Olivine (dry)
Density	ρ	kg m^{-3}	2400	2650	2900	3300
Thermal conductivity	k	W m^{-1} K^{-1}	2.3	2.5	2.8	3.5
Specific heat	C_P	J kg^{-1} K^{-1}	1050	1050	1050	1050
Heat production	H	μW.m^{-3}	1.5	2.0	0.5	0
Exponential decay rate of heat production	z_H	km	20	20	20	0
Power law exponent	n	–	–	2.72	2.4	3.0
Power law activation energy	E_P	kJ mol^{-1}	–	134	212	510
Power law strain-rate	A_P	Pa^{-N} s^{-1}	–	6.03e−24	1.26e−16	7.0e−14
Dorn law activation energy	E_D	kJ mol^{-1}	–	–	–	535
Dorn law strain-rate	A_D	s^{-1}	–	–	–	5.7e11
Dorn law stress	σ_D	Pa	–	–	–	8.5e9
Creep equations						
Power law creep	$\sigma_{\text{creep}} = (\dot{\varepsilon} \cdot A_P)^{1/n} \cdot \exp[E_P/nRT]$					
Dorn law creep	$\sigma_{\text{creep}} = \sigma_D\left\{1 - [1 - (RT/E_D)\ln(\dot{\varepsilon}/A_D)]^{1/2}\right\}$					

plate boundaries, should not be included in any interpretation. Other second order regional/local processes may influence the strength estimates, such as the presence of water or serpentinite in the subcrustal lithosphere, both of which reduce the strength of the mantle (e.g. Bassi 1995; Pèrez-Gussinyé and Reston 2001), or the removal of melts, which strengthens the mantle (Van Wijk and Cloetingh 2002).

A depth-varying rheology such as employed in this study depends on several parameters, of which the most important are crustal thickness, composition and temperature. Spatial variations of these parameters across Europe are described by first order models, as explained above, thus neglecting local, second order scale deviations. The strength of the ductile and viscous layers in the lithosphere also depends on strain-rate, for which we have adopted a value of 10^{-16} s^{-1}, which is characteristic for the long-term and first order bulk deformation of intraplate Europe. However, just as for the other rheological parameters, it is likely that local deformation mechanisms are better described by higher strain-rates, such as, for instance, in the ECRIS areas. In these tectonically active areas, our strength predictions may underestimate the true strength of the lithosphere.

Summarizing, we want to emphasize that the 3D strength cube for the European intraplate lithosphere at its present state, as given in this paper, is based on first order variations in the geometry, composition and temperature of the lithosphere. Thus, interpretations and conclusions reflect only first order variations in (integrated) strength across Europe. Any interpretation of the estimated strength has to be first order and preferably qualitative—comparing different parts of intraplate Europe in terms of being relatively (much) stronger or weaker.

Lithospheric strength maps for intraplate Europe

Fig. 4a shows the integrated compressional strength of the entire lithosphere of Western and Central Europe, whereas Fig. 4b and c display the integrated strength of the mantle and crustal parts of the lithosphere, respectively. As evident from Fig. 4a, Europe's lithosphere is characterized by major lateral mechanical strength variations, with a pronounced contrast between the strong lithosphere of the East-European Platform east of the Teisseyre-Tornquist line and the relatively weak lithosphere of Western Europe. A clear strength contrast occurs also at the transition from strong oceanic lithosphere to the relatively weak continental lithosphere of Western Europe. Within the Alpine foreland, a pronounced northwest—southeast trending weak zone is evident, which coincides with the Mesozoic Sole Pit and West Netherlands Basins, the Cenozoic Rhine Rift System and the southwestern margin of the Bohemian Massif. Furthermore, a broad zone of weak lithosphere characterizes the Massif Central and surrounding areas, as well as the Alps. Higher-strength zones are associated with the central parts of the North German Basin, the British Isles and parts of the Armorican and Bohemian Massifs, all of which are characterized by moderate seismicity (Fig. 1).

The presence of thickened crust in the area of the Teisseyre-Tornquist suture zone gives rise to a pronounced mechanical weakening of the crustal parts of the lithosphere, whereas the mantle–lithosphere retains a moderate strength (Fig. 4b, c). Whereas the lithosphere of Fennoscandia is characterized by relatively high strengths, the North Sea rift system corresponds to a zone of weakened lithosphere. A pronounced strength contrast is evident between the strong Adriatic indenter and the weak Pannonian Basin, the Apennines and the Alps.

Comparing Fig. 4a–c reveals that the lateral strength variations of Europe's intraplate lithosphere are primarily caused by variations in the mechanical strength of the mantle–lithosphere (MSML), whereas the contribution from crustal strength variations appears to be more modest. The variations in MSML are primarily related to variations in the thermal structure of the lithosphere, reflecting upper mantle thermal perturbations imaged by seismic tomography, with lateral changes in crustal thickness playing a secondary role, apart from Alpine domains that are characterized by deep crustal roots. For instance, the strong lithosphere of the East-European Platform, the Bohemian Massif, the London-Brabant Massif, and the Fennoscandian Shield can be explained by the presence of old, cold lithosphere, whereas the European Cenozoic Rift System coincides with a major axis of weakened lithosphere within the Northwest European Platform. Similarly, weakening of the lithosphere of southern France can be attributed to the presence of a tomographically imaged plume rising up under the Massif Central (Granet et al. 1995; Wilson and Patterson 2001).

Strength transects of the Rhine Rift System and adjacent segments of ECRIS

Zooming in on spatial lithospheric strength variations in the wider ECRIS area and its tectonic setting in the Alpine foreland, we examine below a number of regional rheological lithosphere-scale transects across the Massif Central and the Bresse graben, the Rhine Rift System and its flanks, and the North Sea Basin and its underlying grabens (Fig. 5).

Rheological transect along the Rhine Rift System (Amsterdam-Basel)

Figure 6 presents a NNW-SSE transect that extends from Amsterdam across the Ardennes, the Vosges-Black Forest Arch and ends at Basel. Along this transect important lateral variations in the strength of

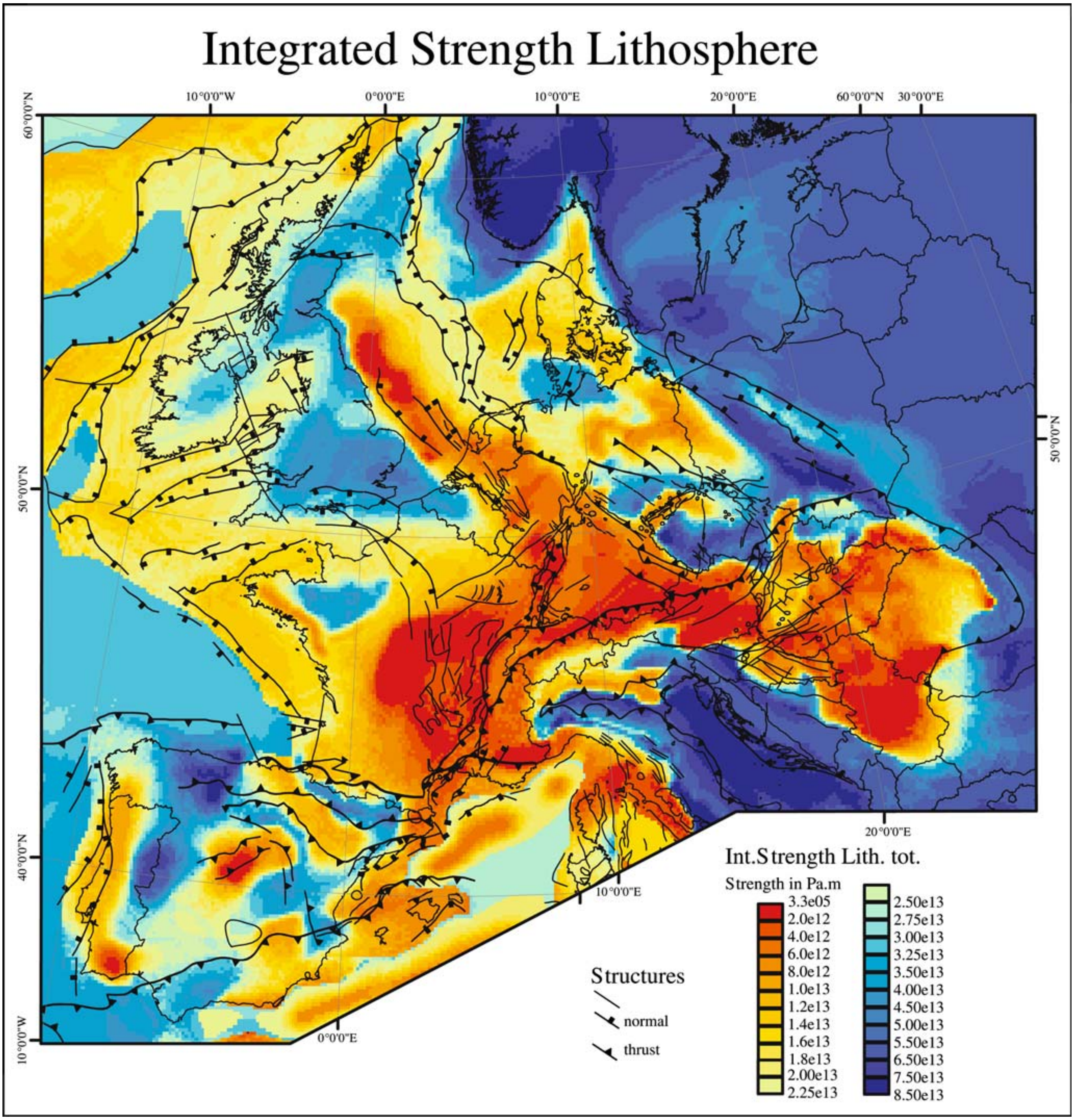

Fig. 4 Integrated strength maps for intraplate Europe. Adopted composition for upper crust, lower crust and mantle is based on a wet quartzite, diorite and dry olivine composition, respectively (see Table 1). Rheological rock parameters are from Carter and Tsenn (1987). The adopted bulk strain-rate is 10^{-16} s^{-1} . Contours represent integrated strength in compression for (a) total lithosphere, (b) mantle, and (c) crust. The main structural features of Europe are superimposed on the maps (Dèzes et al. 2004; Ziegler 1988)

the crust but mainly at the level of the mantle–lithosphere, are evident. In the area of the Vosges-Black Forest Arch, the MSML is apparently rather thin (about 20 km). This arch, which involves a 25–28 km thick crust and about a 100 km thick lithosphere (Babuska and Plomerova 1992), has an amplitude of 2.5 km and in a N–S direction, a wavelength of 200 km. Its Mio-Pliocene development is attributed to lithospheric folding (Dèzes et al. 2004) that was accompanied by partial melting of the lithospheric thermal boundary layer and the upper asthenosphere, as evidenced by volcanic activity within and outside the Upper Rhine Graben spanning 18–7 Ma (Jung 1999). However, there is no evidence for thermal

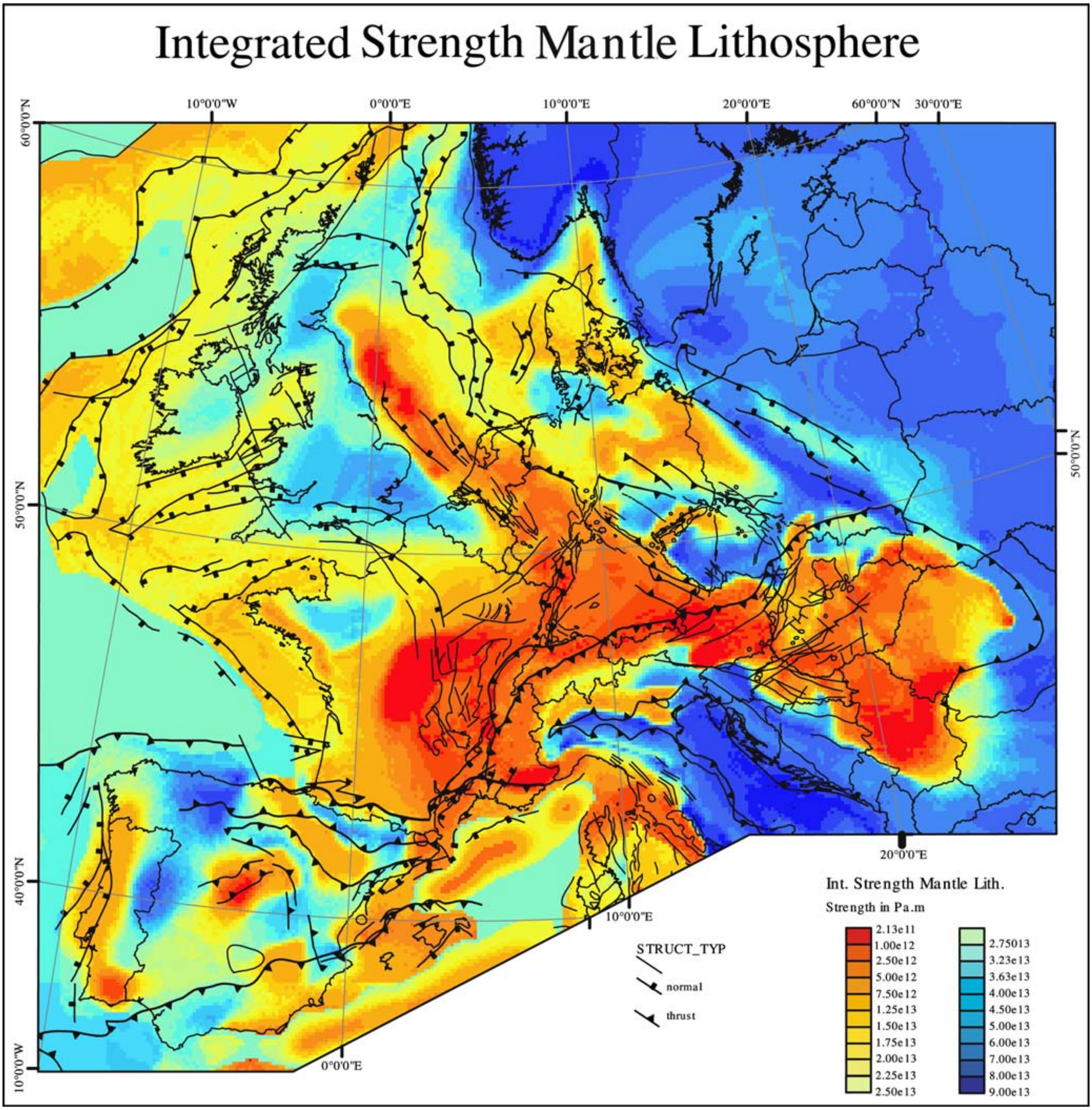

Fig. 4b (Contd.)

thinning of the lithosphere (Achauer and Masson 2002). Towards the Rhenish Massif the thickness of the MSML increases to about 40 km. Beneath the Ardennes, which flank the volcanic fields of the Eifel, plume-related thermal weakening of the MSML and the lower crust is evident (see also Fig. 9). Further northward, the thickness of MSML increases and reaches a maximum of 50 km at the transition to the North Sea Basin near Amsterdam where the thermal thickness of the lithosphere is of the order of 120–150 km (Goes et al. 2000a, b).

Rheological transect through the Massif Central and Bresse Graben

Figure 7a presents a transect that extends from the Atlantic coast of France across the Massif Central to the Western Alps (see Fig. 5 for location). Seismic tomography indicates that beneath the Massif Central the thermal thickness of the lithosphere has been reduced to 50–60 km (Sobolev et al. 1997) and increases away from it to about 110 km beneath the Western Alps (Lippitsch et al. 2003) and perhaps to 120 km

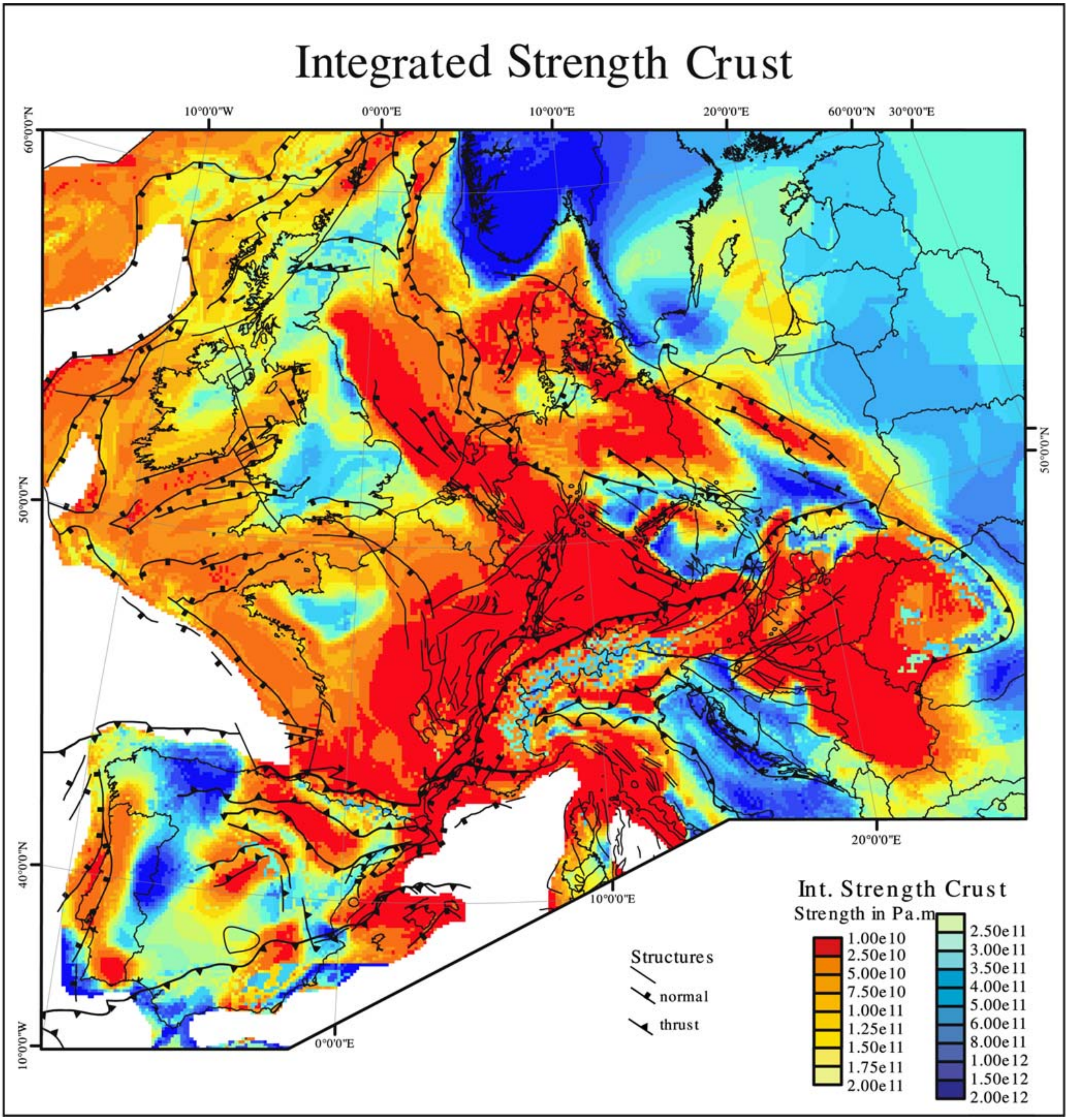

Fig. 4c (Contd.)

beneath coastal France (Dèzes et al. 2004). Our calculations suggest that the thickness of the MSML is of the order of 20 km beneath the Massif Central and the Bresse Graben and that it increases only gradually towards the Atlantic coast and the Alps, thus describing a 500 km wide zone of plume-related thermal weakening of the mantle–lithosphere. In the area of the Massif Central the thickness of the crust ranges between 24 km and 32 km, with lower crustal levels showing evidence of thermal weakening. As across the grabens of the Massif Central and the Bresse Graben total extensional strain by upper crustal faulting does not exceed 7 km, whereas their crustal configuration suggests a three times greater amount of extension, it is postulated that pre-rift crustal thickness were not uniform. This concept is supported by the fact that crustal thicknesses vary considerably in the western parts of the Massif Central that were not affected by Cenozoic rifting.

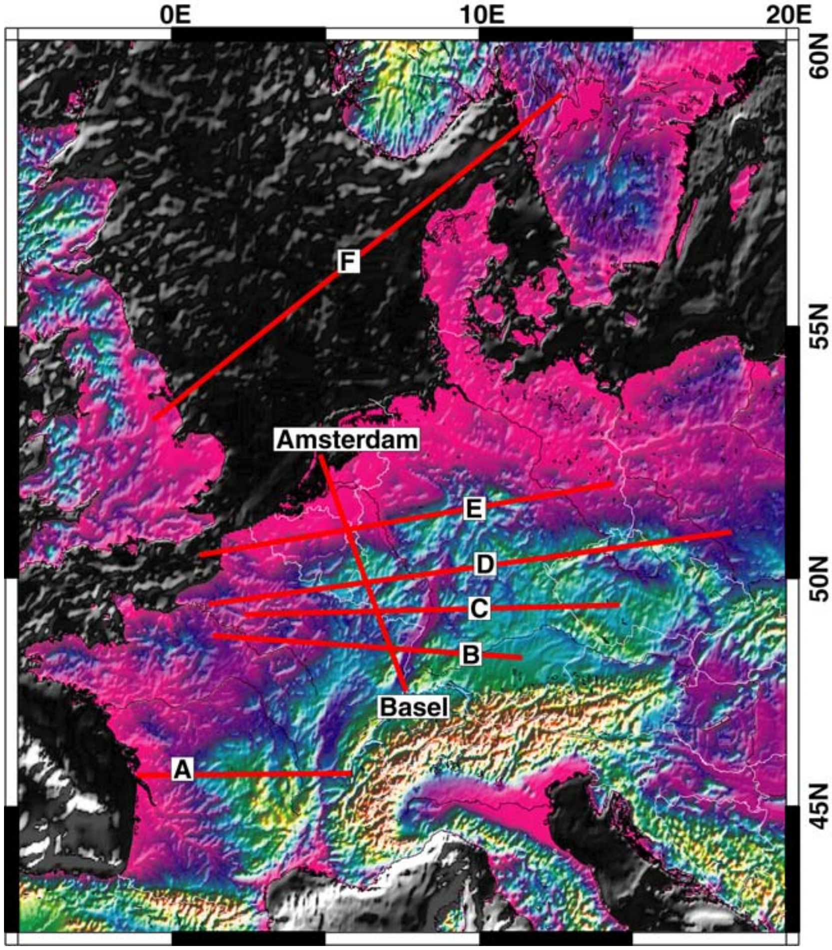

Fig. 5 Topographic map of the Alpine foreland, extracted from ETOPO-2. The red lines denote the location of the rheological cross sections shown in Figs. 6 and 7

Presumably this must be attributed to differential crustal thinning during the Permo-Carboniferous cycle of wrench-faulting and magmatism (Ziegler et al. 2004; Dèzes et al. 2004).

Maxima of lower crustal weakening are located under the Massif Central and the Bresse Graben, corresponding to the surface expression of these structures. Therefore lower crustal flow (Burov and Cloetingh 1997) may have played an important role in the coupling (or decoupling) (Ter Voorde et al. 1998; Gaspar-Escribano et al. 2003) between widespread mantle weakening and localized upper crustal deformation.

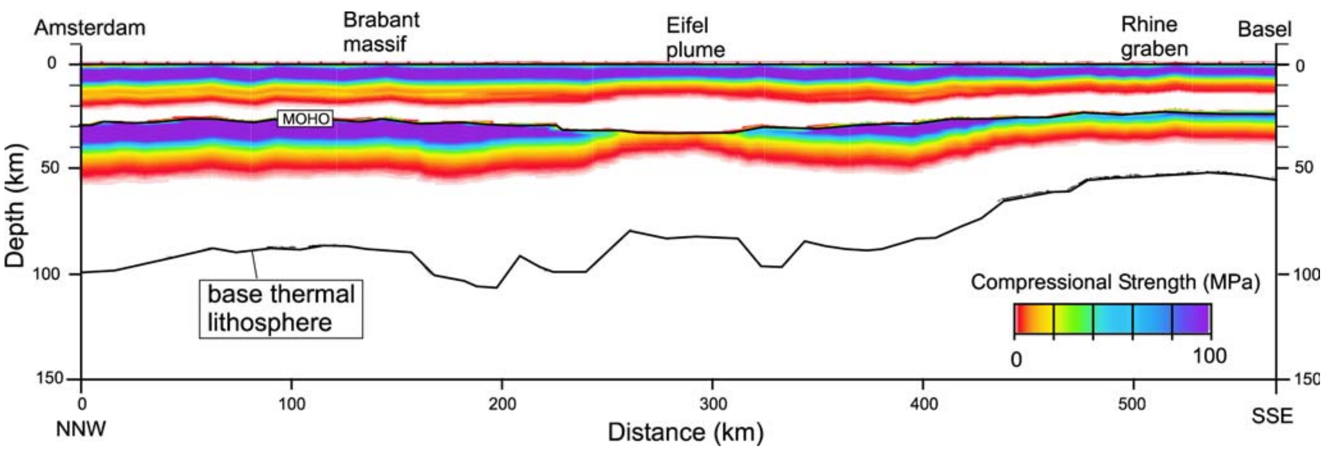

Fig. 6 Rheological cross-section *along* the Rhine graben segments of ECRIS (Amsterdam-Basel). The lower boundary of the mechanically strong part of each intra-lithospheric layer is taken at 2 MPa, and coincides in the color contouring with the transition from red to white. The base of the thermal lithosphere is taken at 1300°C and is indicated by the solid black line

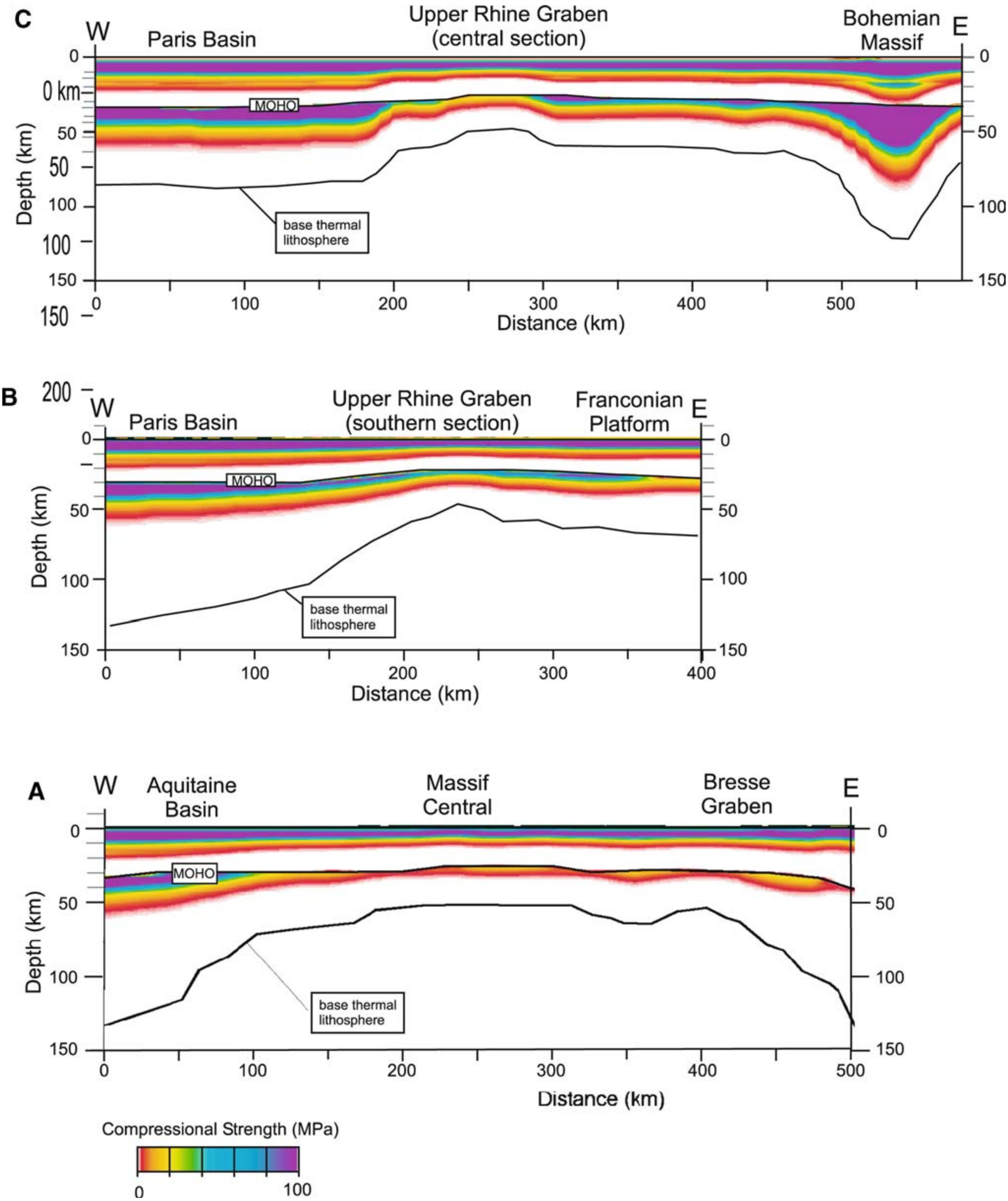

Fig. 7 Rheological cross-section(s) *across* the main grabens of ECRIS. (**a**) Bresse Graben; (**b**) southern part of the URG (Freiburg-Colmar area); (**c**) central part of the URG; (**d**) triple junction of the URG, Lower Rhine Graben, and Hessian Grabens; (**e**) Roer Valley and Hessian Grabens; (**f**) North Sea Basin. See Fig. 5 for the location of the cross-sections. See Fig. 6 for conventions

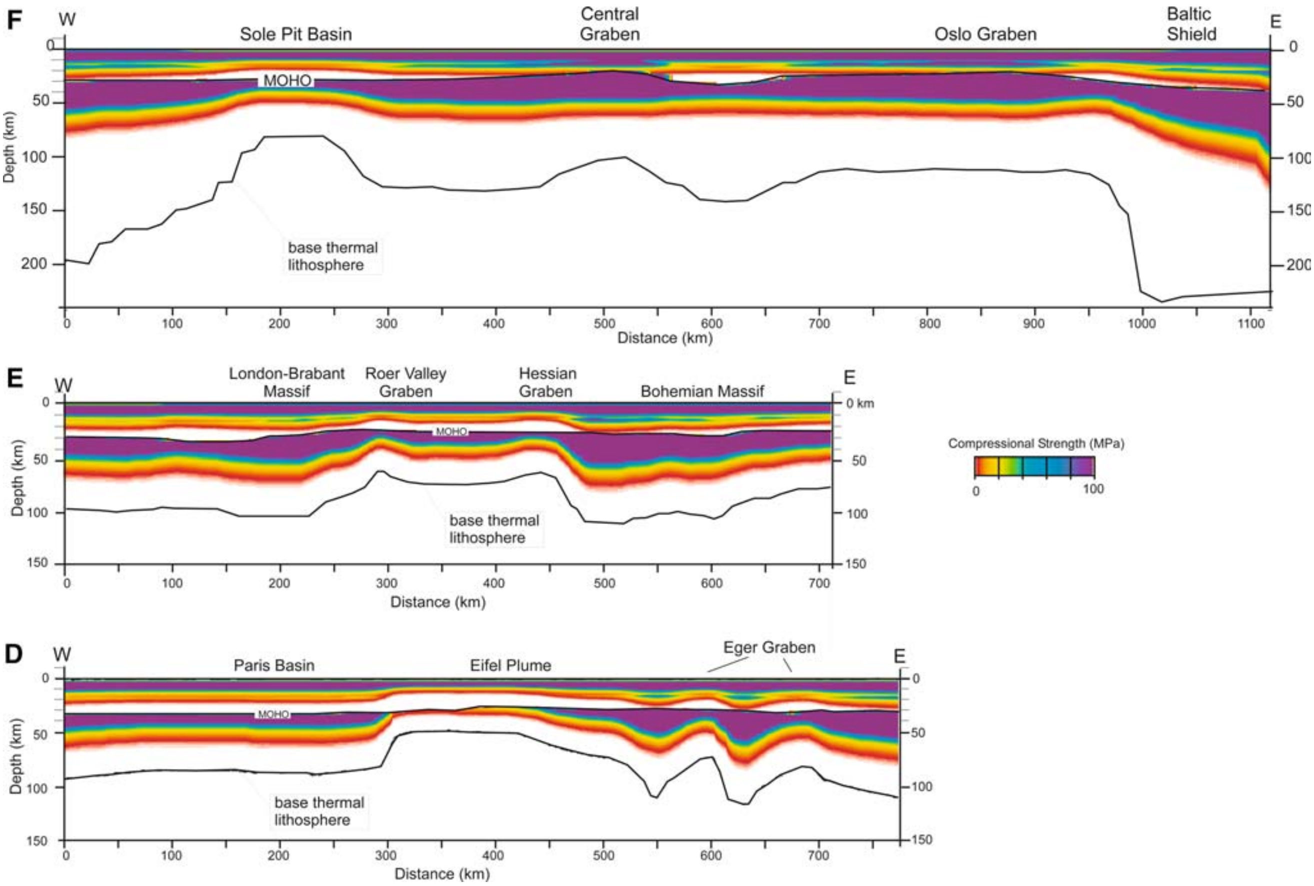

Fig. 7a–f (Contd.)

Rheological transect through the southern part of the URG

Figure 7b gives a transect that extends from the Paris Basin across the Vosges, the URG in the Colmar—Freiburg area, and the Black Forest and terminates on the Franconian Platform. Other studies have shown that the thermal thickness of the lithosphere decreases from some 120 km or perhaps more than 150 km under the Paris Basin (Goes et al. 2000a, b) to about 100 km beneath the URG (Achauer and Masson 2002) and increases again to some 120 km in the area of the Franconian Platform (Babuska and Plomerova 1992).

It should be noted that the transect displayed in Fig. 7b crosses the Urach thermal and magmatic anomaly on the Franconian Platform. This thermal anomaly has probably led to the local reduction of the thickness of the thermal lithosphere, visible in the eastern part of the transect.

A striking feature of our rheological transect is the distinct asymmetry in the thickness of the MSML to the east and west of the URG. Our rheological calculations indicate that beneath the Paris Basin the MSML extends to a depth of about 60 km, shallows to about 35 km beneath the URG and reaches beneath the Franconian Platform depths of 40–35 km.

In the URG, weakening of the lower crust appears to be restricted to a much narrower zone as compared to the broad extent of mantle–lithospheric weakening. The width of the lower crustal weakened zone coincides roughly with the width of the zone of upper crustal faulting. This is in agreement with the results of a teleseismic tomography study of the lower crust in the URG (Fig. 8; Lopes Cardozo and Granet 2003).

Rheological transect across the central part of the URG

The transect given in Fig. 7c extends from the Paris Basin into the southern parts of the Bohemian Massif and crosses the URG along the northern margin of the Vosges-Black Forest arch. Tomographic data indicate that the thickness of the lithosphere decreases from 120 to 150 km under the Paris Basin (Goes et al. 2000a, 2000b) to 70–80 km beneath the Rhine Graben and increases to about 120 km along the margin of the Bohemian Massif (Babuska and Plomerova 1992, 2001). The crust–mantle boundary is located at depths of about 35 km beneath the Paris Basin, rises to 27 km under the Rhine Graben and descends gradually to 32 km at the margin of the Bohemian Massif (Dèzes et al. 2004).

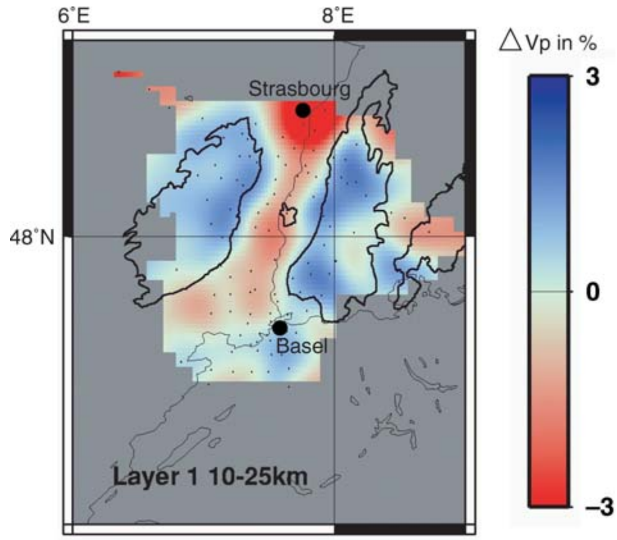

Fig. 8 Relative P-wave velocity image for the lower crust (depth range 10–25 km) of the URG (from Lopes Cardozo and Granet 2003). As shown by Brun et al. (1992) on the basis of deep seismic reflection data, the lower crust is significantly affected by the Cenozoic rifting. The low velocities beneath the graben show the effect of rifting on the velocity structure. The crystalline basement of the Vosges (west) and Black Forest (east) massifs (after Lopes Cardozo and Granet 2003) result in high velocities under the rift shoulders

Our rheological transect indicates weakening of the lower crust in the area of the URG and the Franconian Platform that is paralleled by marked thinning of the MSML, particularly under the Upper Rhine Graben and to a lesser extent under the Franconian Platform. As extensional strain across the URG does not exceed 7 km, Cenozoic thermal thinning of the mantle–lithosphere must have been an important contributing factor. Significantly, the strength of the lower crust and the thickness of the MSML increase sharply along the margin of the Bohemian Massif.

Rheological transect through the triple junction of the Upper and Lower Rhine and Hessian grabens

Figure 7d presents a transect that extends from the Paris Basin across the triple junction of the Upper and Lower Rhine and the Hessian grabens near Frankfurt into the area of the Eger Graben on the Bohemian Massif. The thermal thickness of the lithosphere decreases from 120 km to 150 km under the Paris Basin to 50–60 km beneath the Rhenish Massif (Babuska and Plomerova 1992; Prodehl et al. 1995) and reaches about 80 km beneath the Eger Graben (Babuska and Plomerova 2001). Thinning of the lithosphere beneath the area of the Rhenish triple junction is attributed to plume-related thermal thinning of the mantle lithosphere, with lithospheric extension playing a subordinate role (Dèzes et al. 2004).

Our rheological transect displays very prominent lateral variations in the thickness of the MSML, but also in lower and upper crustal strength. The base of the MSML rises abruptly from about 60 km beneath the eastern part of the Paris Basin to about 30 km in the area of the Rhenish triple junction and gradually descends eastward towards the Bohemian Massif to about 70 km. In the area of the Eger Graben our model shows rapid and major lateral variations in MSML thickness.

A strong reduction in strength of crustal and mantle–lithospheric layers in the area of the Rhenish triple junction is attributed to the presence of finger shaped upper mantle plumes rising up to the base of the lithosphere (Fig. 9, Ritter et al. 2001). It should be noted that in this area the zones of upper mantle and lower and upper crustal weakening closely coincide, thus forming vertical cylindrical structures. Similarly, a mantle plume

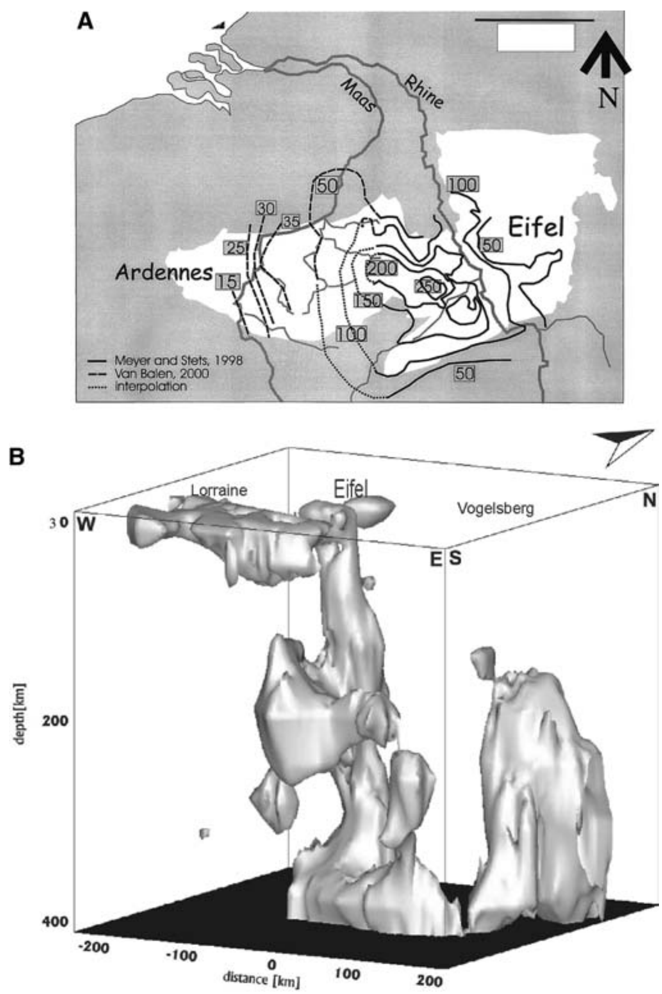

Fig. 9 (**a**) Uplift pattern of the Ardennes and Rhenish Massif (contours in meters) during the last 800,000 years as derived from a compilation of measures of river incision. Modified after Garcia-Castellanos et al. (2000); data sources: Meyer and Stets (1998) and Van Balen et al. (2000). (**b**) 3D representation of the P-wave low-velocity anomaly beneath the Eifel region. The 400 km×400 km wide model represents the –1% v_p anomaly in the upper mantle. (after Ritter 2004)

may also be present beneath the Bohemian Massif (U. Achauer, pers. comm. 2003), thus accounting for the modeled lithospheric strength variations in the area of the Eger Graben. The Rhenish and Bohemian Massif are both characterized by an anomalous topography and important volcanic fields. Of special interest is the pronounced weakening of the lithosphere along the margins of the Bohemian Massif that are characterized by increased seismicity (Fig. 1).

Rheological transect through the Roer Valley and Hessian Grabens

Figure 7e shows a transect that extends from the London-Brabant Massif across the Roer Valley and the Hessian Grabens, and the Harz Mountains and ends at the southern margin of the North German Basin. Beneath the northern parts of the Bohemian Massif the thermal lithosphere thickness is of the order of 100–120 km (Babuska and Plomerova 1993, 2001), decreases to 50–60 km under the northern parts of the Rhenish Massif (Prodehl et al. 1995) and may increase to over 150 km in the area of the London-Brabant Massif (Goes et al. 2000a, b).

Our rheological transect illustrates strong thermal weakening of the mantle–lithosphere and lower crust beneath and between the Roer Valley and Hessian Grabens, and a relative strong mantle lithosphere under the London-Brabant Massif and the northern parts of the Bohemian Massif. Areas of increased lower crustal strength occur on the London-Brabant Massif and in the northern parts of the Bohemian Massif.

Rheological transect through the North Sea Basin

The rheological transect given in Fig. 7f extends from southeastern England across the North Sea to coastal Sweden (Fig. 5) and crosses the Mesozoic rifted Sole Pit Basin, Central Graben and North Danish Basin, which are separated by the Mid-North Sea and Rinkøbing Highs. All three Mesozoic extensional basins were partly inverted during the latest Cretaceous and Paleocene, with the Sole Pit Basin being affected by further inversion movements during the Eocene-Oligocene (Ziegler 1990, 1994; Ziegler and Cloetingh 2004). Along this transect, the thermal thickness of the lithosphere is poorly constrained. However, along the TOR transect, which parallels the NE part of our transect, lithospheric thicknesses of some 200 km are recorded in southern Sweden; these decrease sharply across the Sorgenfrei-Tornquist zone to about 120 km beneath the Ringøbing-Fyn High and across the Trans-European suture to 75–100 km beneath the North German Basin (Cotte et al. 2002; Plomerova et al. 2002).

Our rheological transect indicates an overall weakening of the mantle lithosphere beneath the North Sea Basin, which is attributed to its destabilization during the Permo-Carboniferous tectono-magmatic cycle (Ziegler et al. 2004) and again during Triassic to Early Cretaceous rifting (Ziegler 1990; Ziegler and Cloetingh 2004). A marked increase in the thickness of the MSML is evident across the Sorgenfrei-Tornquist zone towards the Baltic Shield, whereas the thickness of the MSML decreases beneath the Sole Pit Basin.

Unlike in the other rheological profiles presented in this paper, the lower crust appears to have regained its strength in the area of the Central Graben and North Danish Basin, owing to a long phase of post-rift lithospheric cooling (Pascal et al. 2002), whereas it is still weak beneath the Sole Pit Basin, presumably as a consequence of its Paleogene inversion. Lateral upper crustal strength variations can be attributed to thermal blanketing effect of the thick post-rift sediments in the North Sea Basin.

Seismicity

Increased seismic activity is associated with the URG and the Roer Valley Graben, the Armorican shear zone and the Massif Central, as well as with the Eger Graben. This cannot be directly attributed to the activity of mantle plumes impinging to the attenuated lithosphere of the Massif Central, the Rhenish Massif and the Bohemian Massif (Wilson and Patterson 2001) but rather to the reactivation of Cenozoic and older crustal-scale faults under the present compressional stress field of the Alpine foreland (Fig. 2). On a much broader scale, seismicity (Fig. 1; Grünthal et al. 1999) and stress indicator data (Müller et al. 1992; Tesauro et al. 2005, see also Gölke and Coblentz 1996) demonstrate, that active compressional deformation continues in the Alpine foreland also, outside the different segments of ECRIS. Zones of concentrated seismic activity correspond to areas of crustal contrast between the Cenozoic rifts and their surrounding platform areas, as well as to areas of crustal contrast in the rifted northeastern Atlantic margins. In general, earthquakes are associated with pre-existing faults, such as those bounding the Bohemian Massif and transecting the Armorican Massif. Recent GPS data indicate that the highest deformation rates are associated with the ECRIS, with strain rates being on the order of $10^{-16}\ s^{-1}$(Tesauro et al. 2005). The simultaneous occurrence of thrust faulting, normal faulting and strike slip faulting mechanisms in the Alpine foreland supports a stress distribution dominated by heterogeneous crustal structures, including weak zones therein (Handy and Brun 2004).

Lithospheric folding: an important mode of intraplate deformation

Folding of the lithosphere, involving its positive as well as negative deflection, appears to play a more important role in the large-scale neotectonic deformation of Europe's intraplate domain than hitherto realized (Cloet-

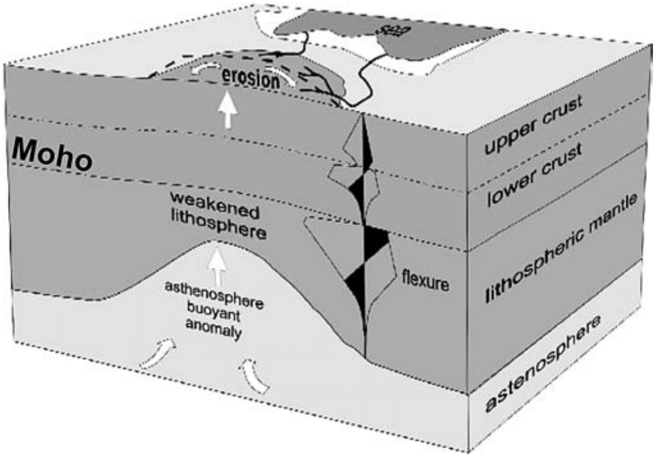

Fig. 10 Elements of the conceptual model adopted for the results presented in Fig. 11. Uplift is assumed to be a flexural response to a buoyant load acting in the base of the lithosphere. Both the amount of load and its extension are assumed to be related to the thermal anomaly at lithospheric depths

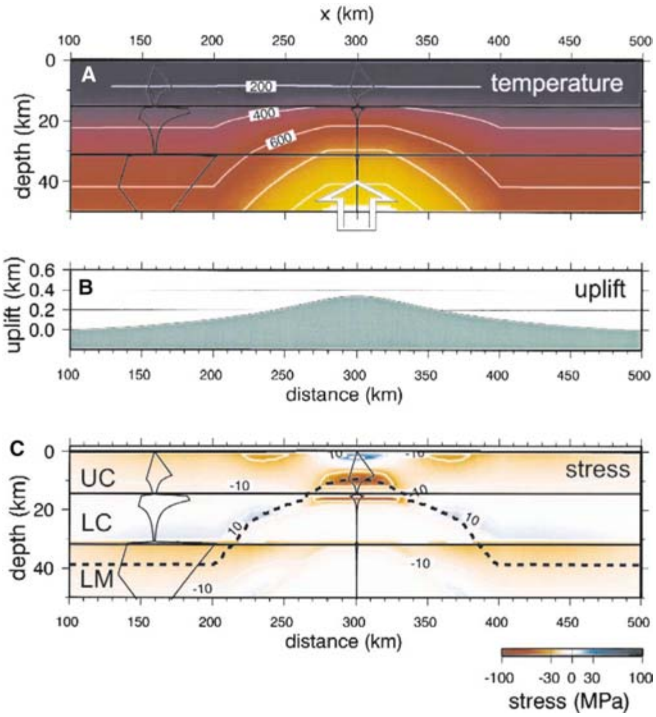

Fig. 11 2D elastic–plastic plate synthetic model. Input temperature distribution (above), calculated uplift (middle panel) and stress distribution (lower panel) predicted in the model assuming a low compression regional setting ($F_x = 1$ TN/m). Positive stresses (in blue) mean extension. The black lines show the yield strength profiles (YSE) representative for both the unperturbed and the anomalous zones. Note that the mantle has nearly no strength in the central area as a result of thermal weakening due to mantle upwelling (see also Fig. 7**d**). The dashed line is the predicted elastic thickness. UC = upper crust; LC = lower crust; LM = lithospheric mantle. Bold lines indicate the base of the Upper and Lower Crust

ingh et al. 1999). The large wavelength of vertical motions associated with lithospheric folding necessitates integration of available data from relatively large areas (Elfrink 2001), often going beyond the scope of regional structural and geophysical studies that target specific structural provinces. Recent studies on the German Basin have revealed the importance of its structural reactivation by lithospheric folding (Marotta et al. 2000). Similarly, the Plio-Pleistocene subsidence acceleration of the North Sea Basin is attributed to stress-induced buckling of its lithosphere (Van Wees and Cloetingh 1996). Moreover, folding of the Variscan lithosphere has been documented for Brittany (Bonnet et al. 2000), the adjacent Paris Basin (Lefort and Agarwal 1996) and the Vosges-Black Forest arch (Dèzes et al. 2004). Lithospheric folding is a very effective mechanism for the propagation of tectonic deformation from active plate boundaries far into intraplate domains (e.g. Stephenson and Cloetingh 1991; Burov et al. 1993; Ziegler et al. 1995, 1998, 2002).

At the scale of a microcontinent that was affected by a succession of collisional events, Iberia provides a well-documented natural laboratory for lithospheric folding and the quantification of the interplay between neotectonics and surface processes (Cloetingh et al. 2002). An important factor in favor of a lithosphere-folding scenario for Iberia is the compatibility of the thermotectonic age of its lithosphere and the wavelength of observed deformations.

Well-documented examples of continental lithospheric folding come also from other cratonic areas. A prominent example of lithospheric folding occurs in the Western Goby area of Central Asia, involving a lithosphere with a thermo-tectonic age of 400 Ma. In this area, mantle and crustal wavelengths are 360 and 50 km, respectively, with a shortening rate of ~10 mm/yr and a total amount of shortening of 200–250 km during 10–15 Myr (Burov et al. 1993; Burov and Molnar 1998).

Quaternary folding of the Variscan lithosphere in the area of the Armorican Massif (Bonnet et al. 2000) resulted in the development of folds with a wavelength of 250 km, pointing to a mantle–lithospheric control on deformation. As the timing and spatial pattern of uplift inferred from river incision studies in Brittany is incompatible with a glacio-eustatic origin, Bonnet et al. (2000) relate the observed vertical motions to deflection of the lithosphere under the present-day NW–SE directed compressional intraplate stress field of NW Europe (Fig. 2). Stress-induced uplift of the area appears to control fluvial incision rates and the position of the main drainage divides. The area located at the western margin of the Paris Basin and along the rifted Atlantic margin of France has been subject to thermal rejuvenation during Mesozoic extension related to North Atlantic rifting (Ziegler and Dèzes 2005; Robin et al. 2003) and subsequent compressional intraplate deformation (Ziegler et al. 1995), also affecting the Paris Basin (Lefort and Agarwal 1996). Leveling studies in this area

(LeNotre et al. 1999) also point towards its ongoing deformation.

The inferred wavelength of these neotectonic lithosphere folds is consistent with the general relationship that was established between the wavelength of lithospheric folds and the thermotectonic age of the lithosphere on the base of a global inventory of lithospheric folds (Cloetingh and Burov 1996). In a number of other areas of continental lithosphere folding, also smaller wavelength crustal folds have been detected, for example in Central Asia (Burov et al. 1993; Nikishin et al. 1993).

Thermal thinning of the mantle–lithosphere, often associated with volcanism and doming, enhances lithospheric folding and appears to control the wavelength of folds. Substantial thermal weakening of the mantle is consistent with higher folding rates in the European foreland as compared to folding in Central Asia (Nikishin et al. 1993), which is marked by pronounced mantle strength (Cloetingh et al. 1999).

Conclusions

In conjunction with the World Stress Map project and the Task Force Origin of Sedimentary Basins, both sponsored by the International Lithosphere Program (ILP), new databases were developed for the stress field of NW Europe and recent crustal-scale vertical motions. The present-day stress field of NW Europe (Fig. 2) (Müller et al. 1997) could be successfully modeled by taking Alpine collisional coupling and Atlantic ridge-push forces into account (Gölke and Coblentz 1996; Goes et al. 2000b; Ziegler et al. 2002). These studies established a close link between the stress field, late Neogene to Quaternary intraplate deformation, earthquake distribution and topography (see also Fig. 1).

Furthermore, acquisition of high-quality tomographic data (Goes et al. 2000b) permitted imaging of the thermal structure of the sub-lithospheric mantle beneath NW Europe, revealing mantle plumes upwelling beneath the Massif Central (Granet et al. 1995) and the Rhenish Massif (Ritter et al. 2001). In this context, it is noteworthy that studies on the mechanical properties of Europe's lithosphere reveal a direct link between its thermo-tectonic age and bulk strength, whereas inferences from P- and S-wave tomography and thermo-mechanical modeling point to pronounced weakening of the lithosphere in the area of the Massif Central and Rhenish Massif owing to high upper mantle temperatures (Hardebol et al. 2003). Uplift of the Rhenish Massif by as much as 250 m during the last 0.8 My (Fig. 9) (Meyer and Stets 2002) can be directly attributed to the load of the impinging mantle plume and related thermal thinning of the lithosphere (Figs. 10, 11) (Garcia-Castellanos et al. 2000; Dèzes et al. 2004). Although substantial plume-related thermal thinning of the mantle–lithosphere is confined to the Massif Central and the Rhenish Massif, and possibly also the Bohemian Massif, the mantle–lithosphere has been thermally weakened in a broad zone around the graben systems of ECRIS. The resulting reduction of the thickness of the MSML contributed significantly to the weakening of the West and Central European intraplate domain, thus rendering it prone to neotectonic deformation under the present-day compressional stress regime. There is increasing evidence that the European lithosphere responds to intraplate compressional stresses by lithospheric folding (Cloetingh et al. 1999), as evidenced for instance by the Plio-Pleistocene subsidence acceleration of the North Sea Basin and contemporaneous uplift of the Fennoscandian Shield (Van Wees and Cloetingh 1996).

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