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Lithosphere tectonics and thermo-mechanical properties: an integrated modeling approach for EGS exploration in Europe

Things to check

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D. Bonte will work on figures and new heat flow compilation with Lazlo in march

Abstract

For geothermal exploration and production of enhanced geothermal systems (EGS) knowlegde of the thermo-mechanical signature of the lithosphere and crust is important to obtain critical constraints for the crustal stress field and basement temperatures. The stress and temperature field in Europe is subject to strong spatial variations which can be linked to Polyphase extensional and compressional reactivation of the lithosphere, in different modes of deformation. The development of innovative combinations of numerical and analogue modeling techniques is key to thoroughly understand the spatial and temporal variations in crustal stress and temperature. In this paper we present an overview of our advancement developing and applying analogue and numerical thermo-mechanical models to quantitatively asses the interplay of lithosphere dynamics and basin (de)formation. Field studies of kinematic indicators and numerical modeling of present-day and paleo-stress fields in selected areas have yielded new constraints on the causes and the expression of intraplate stress fields in the lithosphere, driving basin (de)formation. The actual basin response to intraplate stress is strongly affected by the rheological structure of the underlying lithosphere, the basin geometry, fault dynamics and interplay with surface processes. Integrated basin studies show that rheological layering and strength of the lithosphere plays an important role in the spatial and temporal distribution of stress-induced vertical motions, varying from subtle faulting to basin reactivation and large wavelength patterns of lithospheric folding, demonstrating that sedimentary basins are sensitive recorders to the intraplate stress field. The long lasting memory of the lithosphere, in terms of lithospheric scale weak zones, appears to play a far more important role in basin formation and reactivation than hitherto assumed. A better understanding of the 3-D linkage between basin formation and basin reactivation is, therefore, an essential step in research that aims at linking lithospheric forcing and upper mantle dynamics to crustal vertical motions and stress, and their effect on sedimentary systems and heat flow. Vertical motions in basins can become strongly enhanced, through coupled processes of surface erosion/sedimentation and lower crustal flow. Furthermore patterns of active thermal attenuation by mantle plumes can cause a significant spatial and modal redistribution of intraplate deformation and stress, as a result of changing patterns in lithospheric strength and rheological layering. Novel insights from numerical and analogue modeling aid in quantitative assessment of basin and basement histories and shed new light on tectonic interpretation, providing helpful constraints for geothermal exploration and production, including understanding and predicting crustal stress and basin and basement heat flow.

1 Introduction

At intermediate depths (<10 km) the Earth presents a huge amount of heat, which allows the extraction of geothermal energy through producing hot water from the subsurface used for heating and electricity production most likely providing about 5-10% of electricity demand in 2050 (e.g. Tester et al., 2007). Enhanced geothermal systems (EGS) is targeted at electricity production from circulating water through high temperatures rocks at depths of approximately 3-6 km EGS involves the fracturing of rock through high injection pressure, improving permeability of the rock for fluid circulations as proved successful in Soultz research project in France (e.g. geothermics Special volume Soultz, 2006). Favourable conditions of tectonic stress are required to allow fracturing the rock with limited injection pressure.

Temperature conditions are critical for the potential of electricity production, as they both influence the thermal power and the efficiency of electricity generation (e.g. DiPippo, 2007). Currently at temperatures of ca 120C, relatively high flow rates of ca 200 L/s are capable to deliver a power of about 3 MWe (Unterhanchung) . On the other hand at much higher temperatures of 150-200C flowrates of 50-100 l/s are sufficient to deliver the same amount of power. are required for electrity production using Given the fact that higher reservoir temperatures both increase efficiency and thermal power it is important to identify prospects with relatively high temperatures, and/or flowrate capable of substaining power production over the lifetime of the geothermal project.

Up till now exploration and production areas have been selected largely on the basis of observations of high near surface temperature gradients (e.g. volcanic areas such as Iceland and Tuscany in Italy, fig 1) and or relative high temperatures assessed in deep boreholes drilled mainly for Hydrocarbon exploration and production (e.g. Soultz).
For improved assessment of the exploration potential of continental regions for geothermal energy we need to look beyond depths of temperatures known from shallow wells and need capability to predict temperatures at depth in areas where no well control is avalaible. For enhanced flow rates for fluid circulation generated through fraccing, we need to be capable to predict critically stressed conditions, and predict the stress field orientation for optimizing well planning during exploitation.

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In this paper we review key thermo-mechanical processes which influence the thermal state and the stress conditions in Europe. In doing so we will first introduce basic concepts on the compositional structure of the lithosphere and its relationship with thermal structure of the earth up to ca 100 km depth. We will show how first order Thermal constraints from surface heat flow and deep subsurface geophysical datasets can be jointly used to build and constrain thermal models of the lithosphere beyond well control. Although first order patterns in predicted thermal structure and near surface thermal gradients fit well with observations. First order patterns can be well explained by both compositional controls and active tectonic processes. Strong local deviations are possible, related to local deviations in compositional properties such as caused by granites and tectonic activity. Next we explain how the thermal and compositional structure of the lithosphere controls the mechanical or socalled rheological properties. Rheological models for Europe are validated relating spatial distributions of low strength with zones of active deformation, as the drivers for earth deformation are far field stresses orginating from plate tectonics. We will demonstrate that thermal and rheological model predictions fit very well with earthquake distribution, partitioning of deformation in Europe derived from geodetic measurements. The combined interpretation of the thermal and rheological state of the lithosphere is in close agreement with independent other geophysical data such as the gravity field.

In more detail validation of rheological models through analogue and numerical modeling of deformation processes over geological timescales, demonstrates that first order rheological models fail to take into account lithosphere and crustal scale weak zones which are inherited from previous deformation. These weak zones/faults play an important role in distributing stress and strain in the upper crust representing the top 10 km in the earth. Detailed geomechanical models linking far field stress state and subbasin fault fabrics allow to validate stress distributions in basins. This type of modeling aids in predicting critically stressed faults. Active faults most likely represent active hydrothermal zones enhancing probability of EGS favourable conditions.

The thermo-mechanical state which is observed for Europe relates to a complex evolution through geological times, involving pervasive extensional and compressional deformation. For selected tectonic settings, marked by a specific deformation style, we discuss the relationship between key-factors in tectonic evolution and geothermal prospectivity. This relationship allows to approach in a rational fashion continent-scale exploration for geothermal resources and building hypothesis for thermal and mechanical characterization at depth.

2 Thermal structure of Europe's lithosphere

2.1 Basic definitions and concepts

The lithosphere may be defined thermally or mechanically. From the viewpoint of mantle convection the lithosphere can be regarded as the upper thermal boundary layer of the convecting mantle (Turcotte and Oxburgh, 1967). In the lithosphere heat is transported primarily by conduction, in contrast to the underlying asthenosphere, where heat is transported primarily by convection. In steady state the lithosphere is characterized by a nearly linear temperature-depth profile. According to the mechanical definition the lithosphere is the outermost shell of the Earth in which stresses can be transmitted on geological time scales (McKenzie, 1967; Cloetingh and Ziegler, 2007). The viscosity of rocks depends strongly on the temperature therefore the mechanical behavior of the lithosphere is related to its thermal condition. At lower temperatures the rocks are solid. As temperature increases the rocks begin to deform in a ductile way, and therefore they are unable to sustain stresses on geological time scales.
Considering its chemical composition the lithosphere consists of a crust and a mantle part. Two types of lithosphere exist on the Earth: oceanic and continental lithosphere. Oceanic lithosphere is produced at the mid oceanic ridges. There, it consists of an about 7 km thick oceanic crust which is made up of crystallized partial melts of the uppermost

mantle. The thickness of the mantle part of the oceanic lithosphere is zero at the mid oceanic ridge. With increasing age and distance from the ridge the thickness of the mantle part of the oceanic lithosphere increases as asthenosphere freezes to the base of the crust. Oceanic lithosphere is being produced and consumed at all times. The oldest oceanic lithosphere on Earth is not much older than about 100 my. In contrast, the present day continents consist largely of Proterozoic continental lithosphere, which has been reworked in many places. The continental crust is chemically highly differentiated, and it has a high content in radioactive elements. The mantle part of the continental lithosphere consists of similar material than the underlying asthenospheric mantle, but it acts like a solid, because of its lower temperature.

There is not a sharp boundary between the lithosphere and the asthenosphere. Between the purely conductive and purely convective regimes a transition zone is located, where temperature tends asymptotically to the fully convective profile.

2.2 Steady state Lithosphere temperatures and thickness

The thickness of the lithosphere depends on the method by which it is determined. The base of the thermal lithosphere is defined by the point where the temperature profile intersects the 1300 °C isotherm. The temperature profile, the geotherm, is always calculated from a thermal model, which specifies the heat transport processes in the lithosphere.

The simplest assumption is that the continental lithosphere is in thermal equilibrium, and the geotherm is steady state. In this case the temperature gradient in sediments and underlying crust and mantle is often approximated assuming a layered steady state geotherm based on surface heat flow ($q_{surface}$, **fig 1**), rock conductivity (k) and heat production (A) (cf. Table 1; Fig. 2). In each layer in the lithosphere marked by fixed heat production A_{layer}, the analytical steady state temperature in the layer is corresponding

$$T(z) = T_{top} + q_{top}\frac{z}{k} - 0.5A_{layer}\frac{z^2}{k} \quad (1)$$

Where T_{top} and q_{top} correspond to the temperature and heat flow at the top of the layer. The heat flow decreases towards the base of the layer or top of the next layers as:

$$q_{base} = q_{top} - A_{layer} * \Delta z_{layer} \quad (2)$$

The actual temperatures extrapolated at depth are sensitive to surface heat flow , and thermal parameters of the rocks (k,A). Knowledge of rock conductivity and heat production is based on measurements in boreholes and values extrapolated from laboratory experiments, taking into account mineralogical, temperature and porosity effects. Heat conducitivity ranges from 1 to 4 W m^{-1} K^{-1}, depending on rock type, generally lower for sediments than hard rock. Heat production is mostly occurring in the sediments and crust. In most models it is assumed that the upper crust in the initial configuration is accounting for a siginificant portion of the surface heat flow, typically in range of 40% measuring between 1 and 2 μW m-3 (Pollack and Chapman).. For the lower crust a significantly lower fixed heat production is used. In the mantle lithosphere there is no heat production. In steady state using typical properties (Table 1) the geotherm in the crust is slightly curved due to the heat production of radioactive elements, and it is linear in the mantle part of the lithosphere (**Fig. 2**).

Sediment and crustal thickness are an important constraint for thermo-mechanical models for the lithosphere as they have strong contrasting thermo-mechanical properties compared to subcrustal mantle material (Table 1). Dèzes and Ziegler (2004) have integrated 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. From the Moho depth map (Fig.3), it is evident that the stable parts of the Precambrian Fennoscandian - East-European craton are characterized by a thick crust and Moho depths ranging up to 48 km, whereas in more mobile Phanerozoic Europe Moho depths range between 24 and 48 km and no longer bear any relation to the Caledonian and Variscan orogens. On the other hand, areas underlain by the Precambrian Hebridean craton are characterized by a crustal thickness in the range of 20 to 26 km, reflecting a strong overprinting by a Mesozoic rifting. By contrast, the Alpine chains, such as the Western and Central Alps, the Carpathians, Apennines, Dinarides, as well as the Betic Cordilleras and the Pyrenees, are characterized by more or less distinct crustal roots with Moho depths attaining values of up to 60 km. The present crustal configuration of Phanerozoic Europe reflects that the crustal roots of the Caledonides and Variscides were destroyed during post-orogenic times, and that their crust was repeatedly modified by Mesozoic and Cenozoic tectonic activities.

Taking surface heat flow and Moho depth and sediment thickness in Europe recent studies have presented maps of the thermally defined base of the lithosphere, marked by the depth at which the geotherm reaches 1300C (Fig. 2. Fig. 3). The lithosphere thickness shows strong variations closely related to combined effect of surface heat flow and thermal properties on temperatures at depth. The European lithosphere shows considerable variations in lithosphere thickness and associated temperatures at depth.

<< Fig. 2 and Fig. 3 approximately here >>

2.3 Validation of Lithosphere temperatures and thickness by geophysical methods

The temperatures extrapolated at great depth in the lithosphere can independently be determined by geophysical methods. The different methods measure variations in different properties of the mantle: e.g. density, elastic moduli and conductivity, which are related to variations in composition, structure, mineral alignment and temperature. Global geophysical methods, such as seismic tomography or magnetotellurics, are able to determine the top of the convective regime, just below the lithosphere base. The difference between the seismic lithosphere, defined as the seismic high velocity region on the top of the mantle, and thermal lithosphere can be up to 40-50 km (Jaupart and Mareschal, 1999).

In particular, spectacular improvements have been made in global travel time tomography. A new model parameterization technique and new 3-D 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, as a follow-up on Bijwaard et al., 1998). 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 the cause for longstanding central European volcanism (Goes et al., 1999). Future developments in tomography will be directed towards a more detailed structure (50 km) of the upper 1000 km, although the whole mantle is involved in the analyses (**Fig. 4**).

Goes et al. (2000a) developed a new inversion strategy which takes advantage of the fact that most of the variation in seismic wave velocities in the upper mantle is caused by variations in temperature. State-of-the-art seismological models and experimental data on the physical properties of mantle rocks are 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).

The most comprehensive overview to date of the lithospheric structure of the major tectonic provinces in Europe is given by Artemieva et al. (2006). The integrated model of the lithospheric thickness (Fig. 5) is based on P-wave seismic tomography models (Spakman, 1990; Kissling and Spakman, 1996; Bijwaard and Spakman, 2000; Piromallo and Morelli, 2003), surface-wave tomography models (Panza et al., 1986; Du et al., 1998; Calcagnile, 1991; Shapiro and Ritzwoller, 2002; Boschi et al, 2004), P-wave residuals (Babuska and Plomerova, 1992), thermal models (Balling, 1995; Cermak and Bodri, 1995; Artemieva, 2003), magnetotelluric studies (Jones, 1984; Adam et al., 1982; Praus et al., 1990) and P-T data for mantle xenoliths (Coisy and Nicolas, 1978; Kukkonen and Peltonen, 1999).

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2.4 large scale spatial variation in heat flow and lithosphere thickness in Europe, related to thermal age

The Precambrian part of Europe (Ukrainian shield, East European Platform, Baltic shield; for location see Fig 1) is characterized by 40-50 km thick crust and 200-250 km thick high-velocity lithosphere. The low surface heat flow values (30-50 mW/m^2) are close to the global average for Precambrian cratons (Nyblade and Pollack, 1993), which have not been tectonically reactivated after its formation. Based on the low heat flow values steady state thermal models predict cold and 170-280 km thick thermal lithosphere (Pasquale et al., 2001; Artemieva, 2003). The high seismic velocity beneath the Baltic shield is only partly due to the low temperature in the upper mantle. Gravity and buoyancy constraints on mantle density (Artemieva, 2003; Kaban et al., 2003) reveal a strong density anomaly, which can be attributed to highly depleted lithospheric mantle. Palaeozoic rifts within the Precambrian part Europe (the Oslo rift and Pripyat-Dnieper-Donets rift) have lithospheric thickness similar to the Variscan belt (100-140 km).

The Phanerozoic lithosphere of Western and Central Europe has uniform crustal thickness of 28-32 km, and the lithospheric thickness is in the range of 80-140 km with the larger values beneath the Proterozoic-early Palaeozoic terranes (the Armorican, Bohemian and Brabant Massifs, and the northern part of the Massif Central). The main tectonic events which profoundly affected the structure of the lithosphere were the Caledonian (500-400 Ma) and the Variscan (430-400 Ma) orogenies involving the triple plate collision of Baltica, Laurentia and Avalonia (Dewey, 1969). The thin crust, in places with seismically laminated lower crust and sharp subhorizontal Moho, that crosses pre-existing terrane boundaries, and a lack of seismic signature in the lithospheric mantle suggest that a large portion of the lower crust and lithospheric mantle have been delaminated during the Palaeozoic orogenies (Ziegler et al., 2004) or following it (Nelson, 1992). Surface heat flow varies between 50 and 70 mW/m^2 resulting in a significant rise in lithospheric temperature compared to the Precambrian lithosphere (Cermak, 1993). Thickness of the thermal lithosphere (70-120 km, Cermak and Bodri, 1995, Zeyen et al., 2002) is in agreement with the thickness of the seismic lithosphere. Compared with the Palaeozoic orogens of Western Europe, the Uralides, which remained intact within the continental interior, have a thick crust (50-55 km) and thick lithosphere (170-200 km thick). Fig 6 (Nordern et al., 2007) shows a detailed modeling study of the transition of lithosphere structure from Western and Central Europe lithosphere to the eastern European Craton, clearly demonstrating the deep lithosphere control on the sharp contrast in surface heat flow and crustal temperatures in these very distinct domains.

<< fig. 6 >>

2.5 Uncertainty in thermal parameters -heat producing granites

Andrea F. can you fill in half a page on the granites in the bohemia massigf and possible a digitial versio of a figure. Note that effect of heat production of granites on lithosphere temperature and thickness has been included in Fig. 2.

3 Large scale spatial variation in heat flow related to Alpine and Cenozoic tectonics

The Palaeozoic Variscan orogens have been significantly reworked and overprinted by the collision of the Eurasian and the African plates (Fig. 1). Consequently the lithospheric structure of the tectonically active areas in the Mediterranean region is highly heterogeneous. The Cenozoic orogens of the Alps, Dinarides and Caucasus formed by continental subduction are characterized by crustal thickness which locally reaches 60-65 km and lithospheric thickness exceeding 150-200 km (Fig. 3). Coeval to the continental collision several smaller fragments of Tethyan ocean domains were subducted during the late Cenozoic to present time. It resulted in the formation of back-arc basins (e.g. the Tyrrhenian, Aegean and Pannonian basins. These basins have thin crust (20-30 km) and thin lithosphere (60-80 km). West and North of the alps, tectonic and magmatic events associated with the formation and development of the Cenozoic European Central Rift System (Ziegler and Dezes, 2006).

The various extensional and compressive Cenozoic settings, which are to large extent active up to the present day show a strong relation with basin and basement temperatures and can well be correlated with spatial varations in heat flow, deep basin and basement temperature and lithosphere thickness.

Quantitative models Below we describe in more detail the Cenozoic evolution of the different tectonic settings including the back-arc basin settings in the Mediterranean and the Pannonian Basin, the Central European Rift System. ,. In particula and their impact on heat flow.

3.1 Central European Rift System

Development of the presently still active European Cenozoic rift system (ECRIS), which extends from the Dutch North Sea coast to the western Mediterranean, commenced during the late Eocene (**Fig. 7, Ziegler et al., 2005**). 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 Upper Rhine Graben which 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 shear systems. Although the on-shore parts of ECRIS are characterized by relatively low crustal stretching factors they associate with a distinct uplift of the crust-mantle boundary. To what extent this feature must be attributed to Cenozoic rifting or whether Permo-Carboniferous processes have contributed to it, remains an open question (Dèzes et al., 2006). Evolution of ECRIS was accompanied by the development of major volcanic centres 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. Similar data are, however, not available for Iberia and the Bohemian Massif. Despite this, the evolution of ECRIS is considered to be a clear phase of passive rifting.

During the Late Eocene, the Limagne, Valence, Bresse, Upper Rhine and Hessian grabens began to subside in response to northerly-directed compressional stresses (**Dezes et al., 2006**) that may be related to collisional interaction of the Pyrenees and the Alps with their foreland (Merle and Michon, 2001; Schumacher, 2002). During their Oligocene main extensional phase these originally separated rifted basins coalesced, and the Roer Valley and Eger graben came into evidence. During the Late Oligocene, rifting propagated southward across the Pyrenean orogen into the Gulf of Lions and along coastal Spain in response to back-arc extension that was controlled by eastward roll-back of the subducted Betic-Balearic slab. By Late Burdigalian times, crustal separation was achieved, the oceanic Provencal Basin began to open and the grabens of southern France became inactive (Roca, 2001). By contrast, the intra-continental parts of ECRIS remained tectonically active until the present, although their subsidence was repeatedly interrupted, possibly in conjunction with stresses controlling far-field inversion tectonics. By End Oligocene times, the area of the triple junction between the Upper Rhine, Roer Valley and Hessian grabens became uplifted, and magmatic activity on the Rhenish Shield increased. By middle-Late Miocene times, the Massif Central, the Vosges-Black Forest arch and slightly later also the Bohemian Massif were uplifted This was accompanied by increased mantle-derived volcanic activity. As at the level of the Moho a broad anticlinal feature extends from the Massif Central via the Burgundy Transfer zone, the Vosges-Black Forest into the Bohemian Massif, uplift of these arches may have involved folding of the lithosphere in response to increased collisional coupling of the Alpine orogen with its foreland. Uplift of the Burgundy transfer zone entailed partial erosional isolation of the Paris Basin.

Under the present northwest directed stress regime (**Fig. 3**), which built up during the Pliocene, the Upper Rhine Graben is subjected to sinistral shear, the Roer Valley Graben is under active extension, whereas the North Sea Basin experiences a late phase of accelerated subsidence that can be related to stress-induced deflection of the lithosphere (Van Wees and Cloetingh, 1996). Similarly, the continued uplift of Fennoscandia is thought to be related to intraplate compressional deformation 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).

In the European Central Rift System (ECRIS), lithospheric thickness is similar to that of the adjacent Palaeozoic Variscan structures (80-120 km), although in some parts it can be as thin as 70-80 km. In the the Eifel region regional and large scale P-wave tomography models show a narrow low-velocity anomaly down to the bottom of the upper mantle (Ritter et al., 2001; Montelli et al., 2004) supporting diapiric upwelling of small scale, finger-like convective instabilities from the base of the upper mantle, which presumably act as the main source for Tertiary-Quaternary volcanism of Western and Central Europe. This volcanism and domal uplift of Variscan basement massifs is spatially and temporally linked to the development of ECRIS (Wilson and Patterson, 2001). Tomographic images for the Upper Rhine Graben portion of the ECRIS (Achauer and Masson, 2002) suggest that the URG originated as a passive rift, with complex regional stress fields associated with the convergence of the Eurasian and African plates and inherited structures playing a controlling role in its evolution (Dèzes et al., 2004).

Beneath the French Massif Central (Granet et al., 1995), both local and global seismic tomographic studies indicate the presence of an upper mantle anomaly with a diameter of 100-300 km that rises from the upper to lower mantle transition zone and involves material 100-200 ^{0}C hotter than the ambient mantle, similar to the Eifel area. Similarly to the Rhenish Massif, some authors explain the low-velocity zone in the upper mantle (60-100 km, Granet et al., 1995) and the Cenozoic volcanic activity in the Massif Central by a plume or mantle diapir (Lucazeau et al., 1984; Cebriá and Wilson, 1995; Sobolev et al., 1996). Others propose asthenospheric flow related to west Mediterranean extension (Barruol and Granet, 2002).

3.2 Back-arc and post-orogenic collapse basins

In the aftermath of Collision of the European and African lithospheric plates resulted in the pening of extensional basins in the Mediterranean region. From west to east, these basins are the Alboran, Ligurian, Tyrrhenian, Pannonian and Aegean basins. It is common in their formation and evolution that they were formed on former orogenic wedges by back-arc extension due to roll-back of subducted lithospheric fragments of the Tethys ocean in the general frame of slow convergence between Africa and Europe (Fig. 8). Within this geodynamic framework, the western Mediterranean was initiated during the late Oligocene as the west-directed Apennines-Maghrebides subduction began to retreat towards southeast (Robertson and Grass, 1995). At this time either the main east-directed Alpine subduction changed polarity to west-directed subduction, because the Alpine-Betic system had by then reached the continental collision stage making subduction more difficult (Doglioni et al., 1998; Gueguen et al., 1998), or the subduction was always going on slowly towards west since 80 Ma and by this time it reached a critical stage and began to accelerate (Facenna et al., 2001). Roll-back of this west directed subduction then induced opening of the Ligurian-Valencia-Alboran basins.

Extension advanced until oceanic crust have been formed in the eastern Alboran and Ligurian basins. In early-middle Miocene times as the subducted slab has reached the 660 km transition zone the roll-back and the extension slowed down (Facenna et al., 2003). Detachment of the subducted slab along northern Africa in Middle Miocene resulted in the shift and renewal of the active subduction zone east of Corsica and Sardinia and formation of the Tyrrhenian basin. Finally, a last slab detachment process along the Apennines (Pliocene-to recent) led to a concentration of the active expansion in the southern Tyrrhenian basin.

Formation of the Carpathian arc and the Pannonian basin took place in a similar tectonic setting from Early Miocene . In Late Oligocene the northward convergence of the Adriatic plate towards the European foreland led to eastward tectonic escape of the Pannonian terrain from the collisional zone (Kovács and Kázmér, 1985; Royden, 1988 v. Stefanescu, 1988, Ratschbacher et al., 1991). In the eastern boundary Tethyan oceanic crust persisted, which was subducting below the Pannonian lithosphere. In Early Miocene the compressional deformation changed to extensional due to the roll-back of the subducted oceanic slab and collapse of the previously thickened crust (Csontos et al, 1992; Horváth, 1993). The main phase of extension was Middle Miocene when several troughs and grabens were opened and connected by strike-slip faults (Royden et al., 1983a). By the end of the period the subductable oceanic lithosphere was consumed and the buoyant European continental crust entering into the subduction zone has blocked the process. The subducted lithosphere detached along the Carpathian arc and sank into the asthenosphere (Spakman and Wortel, 2000). The last remnant of the subducted slab hangs vertically in the Vrancea zone, in the hinge zone of the Eastern and Southern Carpathians, as shown in seismic tomographic images (Wenzel et al. 2002, Martin et al., 2005). Since Pliocene to recent the Pannonian basin is in an initial phase of positive structural inversion. It manifest itself in large scale lithospheric folding and inversion of former extensional faults as thrust faults (Horváth and Cloetingh, 1996). The present tectonic regime is strike-slip and transpressional (Bada).

The Aegean domain has undergone two successive stages of extension since Oligocene times. In late Oligocene the southward migration of the African slab triggered the gravitational collapse of the crust thickened by nappe stacking in Eocene. This extension has formed core complexes through uplift of the lower middle continental crust to the surface in the central Aegean (Cyclades) during the middle upper Miocene (Lister et al., 1984). Overall crustal extension occurred during the same period within the northern Aegean (Dinter and Royden, 1993) and in southwestern Turkey (Hetzel ez al., 1995a,b). The second stage of Plio-Pleistocene extension is related to the combined effects of the still active retreat of the African slab and of the westward extrusion of Anatolia (4, 10-14, from Tirel). Brittle extension effected mainly the southern and northern Aegean. The Cretan Sea thinning is controlled by the back-arc extension, while the North Aegean extension is due to the combined effects of the extrusion of Anatolia and back-arc extension. Between these two regions, the Cyclades likely behave as a rigid plateau. Formation of the Mediterranean back-arc basins was accompanied by subduction related calc-alkaline magmatism producing large volume of volcanic rocks mainly during the active subduction phase and sporadic alkaline magmatism in the post-collisional phase.

In the Western Mediterranean region the calc-alkaline magmatism became progressively younger from west to east: Oligocene in Provence, Miocene in the Betics-Alboran-Valencia region and Sardinia-Corsica, and Plio-Pleistocene in the eastern coast of the Tyrrhenian basin. In the Pannonian region the main phase of calc-alkaline volcanism was coeval with the extension in middle Miocene. However, in the Eastern Carpathians volcanism lasted until recently, and it is interpreted as a consequence of gradual slab detachment. In the Aegean region the main calc-alkaline volcanic centers of Oligocene-middle Miocene age are found interestingly in NE-Aegean and W-Anatolia. The volcanic activity shifted to Central Anatolia in middle Miocene and terminated only a few hundred thousand years ago. It was probably related to the subduction along the Cyprus arc. At the present day active subduction occurs only beneath Calabria and the Aegean arc resulting in volcanic activity in the Aeolian islands, and in the volcanic islands of the Cretan Sea, respectively.

3.3 Quantitative assessment basin temperatures beyond well control in extensional basins

Steady state models linking heat flow to lithosphere temperatures generally hold well for most of Europe. However in geologically actively deforming regions , marked by active (depth-dependent) kinematics over the last 50 million years, such as subduction in the Alps, subcrustal thermal attenuation by mantle plumes in the Eifel area or rapid erosion and/or sedimentation in the alps and Rhine graben resp, the lithosphere scale temperature assessment need to take into account non steady state kinematic effects.
Kinematic effects are incorporated in numerical lithosphere models by calculating the transient thermal effects of perturbing the initial steady state geotherm of the lithosphere using the deformation history, constrained by the geological record (e.g. burial histories, apitite fission track data, etc). Thermo-kinematic models which take into account the formation of crustal and lithospheric root in the Alps (Willingshofer et al., 2001), the stretching of the lithosphere in the Valencia trough (Zeyen and Fernandez, 1994), Tyrrhenian sea (Spadini and Cloetingh., 1995), Pannonian basin (Tari et al., 1999). The various versions of the lithospheric stretching models were applied to the Mediterranean back-arc basins, Valencia trough, Tyrrhenian sea, Pannonian basin). (Morgan and Fernandez, 1992; Watts and Torné, 1992, Torné et al., 1998), Gulf of Lion (Bessis, 1985; Burrus, 1989; Kooi et al., 1989), Tyrrhenian sea (Spadini and Cloetingh, 1995;), Pannonian basin (Royden et al., 1983; Royden and Dövényi, 1988; Lenkey, 1999), Aegean sea (Le Pichon et al., 1988; Wijbrans et al., 1993).

The most widespread thermo-mechanical models of extensional basin formation are the various lithospheric stretching models. The concept of lithospheric stretching was introduced by McKenzie (1978). In the model of homogeneous stretching it is assumed that the lithosphere is stretched instantaneously in pure shear mode by a certain amount. Due to conservation of mass the lithosphere is thinned by the same amount. During thinning each elementary volume of the lithosphere keeps its original temperature as it is rising to a higher position, resulting in higher geothermal gradient in the lithosphere, and thus, higher surface heat flow. The high temperature in the lithosphere is not stable, the lithosphere begins to cool by heat conduction. Since heat transfer by conduction is a very slow process it takes a few hundred million years for the temperature to reach its original steady state (Fig. 9a).

The stretching model of the lithosphere was widely accepted and used, because it explains the subsidence history of extensional basins. In passive margins and back-arc basins it was observed that during the active tectonic phase an initial, or syn-rift subsidence occurs, which is followed by a long term, exponentially decaying postrift subsidence. The syn-rift subsidence is an isostatic response to thinning of the crust, and the postrift subsidence is due to thermal contraction caused by the cooling of the lithosphere. In later modifications (Royden and Keen, 1980, Fig. 9) the amount of stretching in the crust and mantle lithosphere can be different, which changes the relative magnitudes of the syn-rift and postrift subsidences compared to the uniform stretching model.

The evolution of the temperature in the lithosphere basically depends on the stretching (thinning) factors. In general, the stretching factors are accepted if the model is able to reproduce the subsidence history. If temperature measurements and hydrocarbon maturity indicators (e.g. vitrinite reflectance) are also available from boreholes, then the modeled present day temperature and vitrinite reflectances must also fit to the observed data. Vitrinite reflectance depends on the cumulated heat effected the organic material, therefore it is an indicator of the thermal history. In the oil industry the vitrinite reflectance is modeled by varying the basement heat flow until the modeled vitrinite reflectence fits to the observed values (e.g. Van Wees et al., 2008) However, in an extensional tectonic setting the stretching model predicts the evolution of the temperature in the lithosphere, thus the basement heat flow too, therefore it is evident to couple the two modeling procedure (Horváth et al, 1988, Van Balen et al., 2000; Van Wees et al., 2008). The fit to the thermal observations increase the reliabilty of the stretching model.

In the uniform stretching presented by McKenzie in 1978,high heat flows are associated to regions of ongoing lithosphere extension. However recently it has been demonstrated that heat flows for sedimentary extensional basins should be considerably lower than predicted by the McKenzie model, as the original model does not include the effects of sediment infill which damps the heat flow effect of extension. Furthermore crustal extension results in a reduction of crustal heat production further .lowering the heat flow. The combined effect results in moderate extensional heat flow, which for onshore basins is generally less than the heat flow prior to extension (Van Wees et al., 2008. fig. 11). Such a prediction is clearly not in agreement with strongly elevated heat flows observed in onshore extensional basins in the ECRIS system, and back-arc basins such as the Pannonian Basin. Below we present a number of mechanisms which can cause elevated heat flow in extensional basins, incorporating basin evolution aspects beyond the McKenzie model

Post-orogenic collapse. In the McKenzie model it is assumed that extension occurs on a lithospheric column with its top at sea level. In Back-arc basins sedimentary basins developed on top of orogenic wedges, having the top of the thickened lithosphere column at elevated topography. During the (initial stages of) extension the basin is relatively sediment starved as it remains above sealevel. Heat flow predicted for sediment starved basins (Fig. 10), lacking the damping effect of the sediments result in considerably higher heat flow than sediment filled basins developing on passive margins (e.g. North Sea, Beekman et al., 2000; Van Balen et al., 2000; Van Wees et al., 2008, Fig. 11). Consequently this effect may enhance heat flows considerably in post-orogenic settings.

Subcrustal mantle attenuation: In parts of the ECRIS, and back-arc basins a major contributing factor to elevated heat flow appears to be the fact that thermal attenuation is considerably larger in the mantle than in the crust. Adopting this effect in the extension model ($\beta > \delta$, Figs 10,11). predicted heat flow becomes significantly elevated in an onshore setting (Fig. 11). Various explenations are possible for the subcrustal mantle attenuation, dependening on the tectonic setting. In back-arc settings mantle lithosphere attenuation is can take place faster than crustal stretching because of slab delamination (e.g. Van Wees et al., 2000; Ziegler et al., 2006) or because of complex mantle flow aspects (Fig. 9, lower panel). In the ECRIS impingement of mantle plumes as demonstrated for the URG and MC can cause thermal uplift and thermal alteration of the mantle lithosphere which can be modeled through applying $\beta > 1$, $\delta = 1$ (Cloetingh and Van Wees, 2005).

Subduction and extension derived melts thickening the crust: Thermal attenuation in the mantle can cause pervasive magmatism as melt occurs when temperatures are in excess of the mantle solidus (McKenzie and Bickle, 1988). Melts accumulate as underplates at the base of the crust and can cause basalt extrusions at the surface (White and McKenzie, 1995). The kinematic models presented here do not allow to model detailed kinematic effects of such volcanism, which can make up to 10-50% of the crustal thickness and results in permanent uplift, which should be accounted for in a rift model. The uplift effect and large scale thermal attenuation can be taken into account using a crustal stretching value which is less than subcrustal stretching ($\beta > \delta$), which is similar to the effect of subcrustal mantle attenuation. Subduction related magmatism can also result in apparent thickening of the crust($\beta > \delta$), provided magmatic volumes correspond as reflected by large deep seeted volcanic provinces such as in Tuscany. Local scale thermal attenuation, which can involve volcanism in the crust will result in much more elevated heat flow as tectonically predicted by lithosphere kinematics. These secondary effects may explain patterns of heat flow up to 400 mW m-2 in Tuscany, related to volcanism at crustal levels.

Example from the Pannonian Basin

The subsidence, thermal and maturation history of the Jász-I well from the Pannonian basin is presented in Fig.12 as an example (Horváth et al., 1988). The well penetrated the

Paleozoic basement at a depth of 3637 m. The thick line on the left panel starting from the surface at 17 Ma and reaching 3637 m at the present shows the accumulation of the Neogene and Quaternary sediments. The other lines starting from the surface from left to right show the thicknesses of progressively younger sediments. The water depth was calculated as the difference between the predicted basement depth and the thickness of sediments. The subsidence rate in the first 5 Ma was higher than the sediment accumulation rate resulting in relatively deep water of 500 m. This is in agreement with the sedimentological observations (Juhász, 1994) and interpretation of seismic sections (Vakarcs et al., 1994), which show that the basin was filled up by a delta system prograding into the basin from the peripheral areas. The dashed and dotted lines indicate the evolution of temperature and vitrinite reflectance through time, respectively. The temperature-depth and vitrinite reflectance-depth profiles calculated for the present are in good agreement with the observations. The best fit model of the Jász-I well is obtained assuming that the lithospheric extension started at 17 Ma and lasted till 11 Ma and it caused thinning of the crust and the mantle lithosphere by a factor of 1.8 (β_c) and 100 (β_m), respectively. The large mantle thinning factor means that the mantle was practically removed during stretching. Horváth et al. (2006) interpret the large mantle thinning that during the tectonic escape of the Pannonian block the crust extruded directly onto the asthenosphere. Since 11 Ma cooling of the lithosphere occurs causing thermal subsidence.

As the model reproduces the subsidence history and fits to the thermal data, we assume that it predicts correctly the present day temperature distribution in the lithosphere (Fig. 12).

3.4 Orogenic belts and flank uplift: Exhumation and surface heatflow

Collisional orogenic belts such as the European Alps portray a two-stage tectono-thermal evolution during which stacking of crustal units lead to burial of rocks (stage I) and subsequent denudation of the mountain belt by erosion and tectonics (stage II) results in exhumation of rocks back to the surface. Both, burial and exhumation of rock units perturb the thermal structure of the lithosphere transiently, but with opposite sign. While burial by for example thrust imbrication leads to downward deflection of isotherms, exhumation through normal faulting or erosion has the opposite affect and may result in up-warped isotherms (Fig. 13). Controlling parameter for the magnitude of the thermal perturbation is the velocity at which rocks are buried or exhumed (e.g. Mancktelow and Grasemann, 1997).

In the Eastern Alps post-collisional exhumation of the metamorphic core commenced during the Late Oligocene and resulted in the formation of fault-bound metamorphic domes, the Tauern and Rechnitz domes. A wealth of structural and stratigraphic data suggests that rock exhumation was driven by a combination of orogen-parallel extension and erosion (e.g. Frisch et al., 1998). Exhumation rates deduced for the western Tauern Window were high (~ 4 mm a^{-1}) during the Late Oligocene to early Miocene. Since the

middle Miocene exhumation rates decreased rapidly to about 1 mma^{-1} for the time period between 15 and 10 Myr and to ca. 0.2 mm a^{-1} for the last 10 Myrs (Fügenschuh et al. 1997). Rapid exhumation of the Tauern Window occurred under isothermal conditions due to the upward advection of heat. The resulting thermal perturbation resembles that of a thermal dome with strong lateral gradients (Genser et al., 1996). In accord with the geochronological data, Sachsenhofer (2001) estimated the paleo-heat flow at the end of the rapid exhumation phase, based on vitrinite reflectance data from syn-orogenic strata, to be in the order of 175-210 mWm^{-2} for the Tauern Window region. This exhumation related thermal anomaly decayed from the Middle Miocene onward resulting in a present-day heat flow in the order of 60-70 mWm^{-2} what is still higher when compared to the neighboring regions in the north (<55 mWm^{-2}) (Sachsenhofer, 2001), and the south (<60 mWm^{-2}) though dada coverage in the Southern Alps is poor (della Vedova et al., 2001). These findings have been supported by 2d thermal-modelling of Willingshofer and Cloetingh (2003) along the TRANSALP deep seismic line (TRANSALP Working Group, 2002), where the predicted present-day up-warping of the isotherms in lower and mid-crustal levels as well as the presence of condensed isotherms in upper crustal levels probably reflects the decay of a more pronounced early Miocene thermal perturbation (Fig. 13). This interpretation of the Tauern thermal anomaly is at variance with a more recent thermal modeling study along the same transect in which Vosteen et al., (2006) argue, based on 1d-thermal modeling, that the calculated deviation of the temperature from steady state (+ 3.5 K in 1 km and 15 K at 5 km depth, respectively) is merely the consequence of slow exhumation (0.3 mma^{-1}) during the past 14 Myrs.
As illustrated for the example of the Eastern Alps, quantifying the rates of mass movement through the crust and lithosphere in terms of rock exhumation is essential for a better understanding of its bearings on the past and present-day thermal structure of the lithosphere.

PARAGRAPH MARLIES OVER CCR / CS heat flow, UPDATE FIG 14 Appropriately

4 Active intraplate deformation and Thermo- Mechanical structure of Europe's Lithosphere

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, Cloetingh et al., 2007), that are accompanied by accelerations of subsidence and sedimentation rates.

Over the last few years, seismicity studies and geomorphologic evidence demonstrate the important contribution of neotectonics to the topographic evolution of intraplate Europe including France, the Netherlands, Germany, Spain and Hungary and Romenia (Fig. 1).

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 et al. 2001) and modeling studies (e.g. Bada et al. 1998; Goelke 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 firstorder 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, mostly for the Northwest European foreland (Cloetingh et al., 2005, 2006, 2007; Tesauro et al, 2008).

4.1 Basic concepts of lithosphere strength

The strength of continental lithosphere is controlled by its depth-dependent rheological structure (Fig. 15) in which the thickness and composition of the crust, the thickness of the mantle–lithosphere, the potential tem- 761 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 agedependent 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 extensional, thrust- and wrench-faults, can cause significant weakening of the otherwise mechanically strong upper parts of the crust (Van Wees and Stepenson, 1995, Ziegler et al., 1995; 1998; Van Wees and Beekman, 2000). 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, Fig. 16). 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 (Fig. 17), 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; Worum et al., 2004). 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.

4.2 Model setup

Drawing on the newly compiled Moho map of Europe of Dezes 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), Cloetingh et al. (2005) 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. 15) (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, taking surface heat flow as boundary constraint (cf. Fig. 2). 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.

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 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; Pe` rez-Gussinye´ 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 strainrates, 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.

4.3 Lithospheric strength maps for intraplate Europe

Figs. 20 and 21 show the integrated compressional strength of the entire lithosphere of Western and Central Europe. As evident from Figs. 20 and 21, 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 ofWestern 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. 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. 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. The lateral strength variations of Europe's intraplate lithosphere are primarily caused by variations in the mechanical strength of the mantle–lithosphere (MSML) (Cloetingh et al., 2005).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).

Increased seismic activity is associated with the ECRIS (Fig. 22), 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 crustalscale faults under the present compressional stress field of the Alpine foreland (Fig. 2). On a much broader scale, seismicity (Fig. 1; Grunthal et al. 1999) and stress indicator data (Muller et al. 1992; Tesauro et al. 2005, see also Golke 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. 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). For regions outside the ECRIS, Cloetingh and Van Wees (2005) have hown that impingement of mantle plumes has caused a Late Cenozoic strength inversion facilitating intraplate reactivation of major fault zones bordering Paleozoic massif areas, instead of North Sea Basin inversion which abated in Miocene times.

4.4 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 two regional rheological lithosphere-scale transects across the Massif Central and the Bresse graben, the Rhine Rift System and its flanks (Fi.g 21). More sections are presented in Cloetingh et al. (2005)

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

Figure 21, section A, 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 765 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 (De` zes 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 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 triple junction of the Upper and Lower Rhine and Hessian grabens

Figure 21, section D, 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 (Dezes 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 (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

4.5 Effective Elastic thickness, regional gravity and Geodetic validation

Fig. 23, 24, 25→ Fred Beekman

EPSL Tesauro et al

EPSL Tesauro et al. 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).

5 Crustal stress and strain controls in Neotectonic extension in the ECRIS: natural laboratory studies

In the previous section we have shown that large parts of the ECRIS and back arc basins such as the Panonian Basin are marked by decoupled lithosphere dynamics, marked by relatively strong upper crust and upper mantle.

Neotectonically active tectonic structures in the ECRIS have been extensively studied in the recent years by various European (e.g. ENTEC) and National research initiatives focused on geohazard assessment (e.g. Cloetingh et al., 2005; Cloetingh et al., 2006).

Multi-discilinary research on selected natural laboratories in the ECRIS have provided key insights in the relative importance of deep crustal scale faults (> 5km) controlling neotectonic deformation, and showing a close relationship with deformation partitioning in basins and seismicity. These controls are particularly relevant for exploring active tectonic fabric at intermediate depth (<= 5km). These are particular relevant to geothermal exploration as active fault zones are proximal to (shear) failure stresses, facilitating hydraulic fracturization. Furthermore active fault zones appear to be marked by natural permeability on geological timescales (e.g. Gartrell et al., 2006), which may retain permeable on timescales of geothermal production. Below we demonstrate key insights for two subareas in the ECRIS system located in the LRG and URG (Fig. 7).

5.1 Lower Rhine Graben

The Lower Rhine Graben (LRG), which forms the northwestern segment of the European Cenozoic Rift System, extends from the margins of the Rhenish Massif to the Dutch North Sea coast (Fig. 26). Its main elements are the Erft half graben in the German Lower Rhine Embayment and the Roer Valley Rift System (RVRS) of the Netherlands (Fig. 27). The late Oligocene and younger RVRS is superimposed on the West Netherlands basin that underwent a complex evolution, involving several Mesozoic and Paleogene extensional and inversion phases. (Zijerveld et al., 1992; Geluk et al., 1994; Dirkzwager et al., 2000; Van Balen et al., 2005). During these Late Palaeozoic faults were repeatedly reactivated (Ziegler, 1990, 1992; Dirkzwager et al., 2000; Houtgast et al., 2002).
As shown in Figure 27a, the RVRS structurally consists from SW to NE of the Campine Block, the Roer Valley Graben (RVG) and the Peel Block (Dirkzwager et al., 2000). Subsidence of the RVG was controlled by multi-stage evolution of the main bounding fault zones, including the Peel Boundary fault zone (PBFZ), the Feldbiss Fault Zone (FFZ), the Veldhoven Fault and the Rijen Fault. Remarkably, the subsidence of the basin is almost balanced by the offsets along these fault zones (Houtgast and Van Balen 2000; Michon et al., 2003). The Oligocene and younger syn-rift sediments attain a thickness of 1700 m. In its central, deepest parts, the RVG has the geometry of a half-graben that is bounded in the NE by the Peel Boundary fault zone (Fig. 27b). Towards the NW and SE

the RVG shallows progressively. In its SE parts, the RVG has the geometry of a symmetric graben, with the bounding fault zones having about equal throws. In the southeastern continuation of the RVRS, the main extension is accommodated in the assymetric Erft Graben that is, however, not the direct continuation of the RVG but corresponds to a separate structural element (Geluk et al., 1994; Klett et al., 2002; Michon et al., 2003).
The RVRS started to subside during the late Oligocene under a northerly-directed compressional stress regime, controlling WNW-ESE directed oblique extension across the evolving graben At the transition to the Miocene, the compressional stress regime permutated to a NW-directed one, causing nearly orthogonal NE-SW extension across the RVRS). This stress regime persisted until the present, although its magnitude apparently increased during the Pliocene, as evidenced by a subsidence acceleration of the RVRS. In the recent past, fault movements, some of which may have been accompanied by earthquakes, gave rise to the development of distinct fault-scarps. Some of these have been investigated in the framework of paleoseismologic research (Camelbeeck and Meghraoui, 1998; Camelbeeck et al., 2001; Houtgast et al., 2003a,b, 2005). At present, the RVRS is seismotectonically active, and thus, presents a zone of increased seismic hazard, as highlighted by the 1992 Roermond earthquake (M_L = 5.8) (Fig. 27a, Plenefisch and Bonjer, 1997; van Eck and Davenport, 1994; Hinzen, 2003).

Earthquake activity in the LRG

Since the early studies of Ahorner (1983) on seismicity and faulting in the LRG much progress has been made owing to a multidisciplinary approach that integrates German, Belgian and Dutch studies on seismicity (EUCOR-URGENT; Dost and Haak, 2005; Camelbeeck et al., 1994) and high resolution digital elevation models (Cloetingh and Cornu, 2005) with the results of trenching (Houtgast et al., 2004; Meghraoui et al., 2000), high-resolution reflection-seismic data recorded on rivers (e.g. Cloetingh, 2000; Dèzes et al., 2004) and detailed analyses of industrial reflection-seismic and well data (Geluk et al., 1994; De Mulder et al., 2003), as well as geomechanical modeling of fault reactivation (e.g. Dirkzwager et al., 2000; Worum et al., 2004). Geomechanical study results presented in Fig. 28, clearly demonstrate that border faults are marked by relative weakness compared to default rheology, facilitating localization of strain and seismicity on the major graben bounding faults.

These studies, which are based on subsurface data that were not yet available at the time Ahorner (1983) carried out his initial studies, reveal the prolongation of neotectonically active faults into the vulnerable coastal lowlands of the Netherlands (Fig. 27). These active faults largely coincide with Base Miocene faults. (Fig. 27 inset) (Geluk et al. 1995; De Mulder et al., 2003; Worum et al., 2004; Cloetingh and Cornu, 2005).

5.2 Upper Rhine Graben

Studies carried out in the framework of ENTEC (e.g. Cloetingh et al., 2006) addressed particularly the southern parts of URG (Fig. 18a), an area of increased seismic hazard (Fig. 1), as for instance evidenced by the 1356 earthquake that severely damaged the city of Basel. Despite dedicated research, the seismic source of this historical earthquake (strike slip, thrust or normal faulting, reactivation of Oligocene or Permo-Carboniferous faults) has not yet be unequivocally identified (Meyer et al., 1994; Nivière and Winter, 2000; Meghraoui et al., 2001; Müller et al., 2002; Lambert et al., 2005). Furthermore, it is not clear whether on-going deformation of the North-Alpine foreland at convergence rates of about 1 mm/y or less (Müller et al., 2002; Ustaszewski et al., 2005b) is partitioned between the crystalline basement (including Permo-Carboniferous troughs) and its sedimentary cover along rheologically weak Middle and Upper Triassic evaporite layers (Müller et al., 1987). During the Pliocene, shortening in the Jura Mountains propagated north-westward and northward and encroached during Late Pliocene times on the southern margin of the URG (Nivière and Winter, 2000; Giamboni et al., 2004). This late phase of Jura Mountain folding was accompanied by a change from previously "thin-" to "thick-skinned" deformation (Philippe et al., 1996; Becker, 2000; Dèzes et al, 2004). Solving these problems is a key issue in assessing the seismic hazard potential of the southern URG area that requires knowledge on fault kinematics during the geological past.

ENTEC research in the southernmost URG concentrated on detailed mapping of basement faults and kinematic reconstructions throughout time, integrating available geophysical data with results of structural field studies and geomorphologic observations.

Crustal extension across the southern URG accounts for a net stretching factor of 1.2 (Villemin et al., 1986), corresponding to a total extensional strain of 6-7 km (Brun et al., 1992). During the late Eocene and Oligocene, deformation was concentrated on the NNE-striking main border faults, which obliquely cross-cut Palaeozoic structures (Sittler, 1969). During the early Miocene, deformation progressively migrated towards the interior of the evolving graben as initial E-W directed extension rotated counter-clockwise to a nearly NE-SW directed one (Behrmann et al., 2003; Bertrand et al., 2005). The modelling study discussed below covers parts of the southern URG and its shoulders in the Colmar and Freiburg-Offenburg area (Fig. 29).

The rifting history of the southern URG was analyzed by applying numerical modelling techniques, based on finite element methods and contact mechanics (Cloetingh et al., 2006). Both forward and backward models were carried out to address two major aspects of rifting processes, namely the kinematics of extension and fault propagation. The forward model aimed at defining the evolution of faulting during the rifting phase, and at analyzing the relationship between the strike of faults and the extension direction (orthogonal versus oblique extension) (Fig. 29). The backward model focused on the kinematics of rifting in the southern parts of the URG. Retro-deformation of this graben segment helped to define the finite amount of extension that occurred across it, the potential contribution of strike-slip deformation to observed displacements, and the cumulate amount of subsidence and possible post-rift uplift (Fig. 23).

Discussion of forward model

Qualitative results show that deformation is mainly concentrated on contact zones, the border faults, while the central part of the graben remains less deformed. However, in case of oblique extension, deformation is not necessarily restricted to the border faults: a narrow band of high strain and brittle behaviour develops in the centre of the graben along its axis (Fig. 29) (Cornu and Bertrand, 2005a). This zone is the likely location of subsequent faults that develop during oblique rifting. For this segment of the URG, the narrow zone of high strain and brittle behaviour closely fits the surface trace of the Rhine River Fault (Fig. 30).

It is, however, not possible to propose from this model the sense of displacement on a newly formed fault along the graben axis, as most of the vertical displacement is accommodated along the border faults. Moreover, because the zone of high brittle behaviour is rather vertical, one would expect strike-slip motion combined with a normal component. A plausible rifting scenario for the URG could be, as proposed by Bertrand et al. (2005), that initially the border faults accommodated nearly orthogonal extension, and that, as the extension axis rotated counter-clockwise, new faults developed inside the graben. The models suggest that one of the most important features is the Rhine River fault that probably accommodated a significant amount of strike-slip movement whilst most of the normal displacement was taken up by the Main Border and associated faults.

Discussion of backward model

Displacement components along the 3 axes provide information on rifting processes in the URG. As previously suggested by forward modelling and Cornu and Bertrand (2005b), the majority of deformation is initially accommodated along the border faults (Fig. 30a for heave and 30c for throw) with only a minor part being distributed within the graben itself. This confirms the model of Behrmann et al. (2003) and Bertrand et al. (2005) which suggests that deformation first concentrates on the main border faults whereas localized deformation occurs within the graben during later rifting stages.

Despite strict boundary conditions and a purely orthogonal rifting scenario, components of strike-slip motion have been identified along all faults. Strike-slip components range from a few tens to about 100 meters, as in the previous simple backward model (Cornu and Bertrand, 2005b). In case of partly oblique rifting, it is likely, that the URG accommodated a significant component of strike-slip motion. At this point, we are unable to quantify the cumulate strike-slip deformation component as no trace of strike-slip deformation has yet been identified in the field.

Although the main border faults accommodated the bulk of deformation, the Rhine River fault played an important role in the evolution of the URG. This fault, despite having a relatively low heave owing to its steep dip, accommodated a significant throw, and marks the boundary between the shallow eastern and the deep western part of the southern URG. Each fault block, although cut by secondary faults that accommodate smaller displacements, behaves more or less as a single block.

Summarizing, the forward model provides new insight into the possible faulting history of the URG. It clearly shows that 1) whatever the extension direction, deformation is mainly accommodated along the border faults, and 2) the observed fault pattern can only be reproduced under conditions of oblique extension.
The backward model is in good agreement with the results of previous studies (Behrmann et al., 2003; Cornu and Bertrand, 2005b; Bertrand et al., 2005) which show that 1) the maximum deformation occurs along the border faults, and 2) maximum subsidence is centred on the south-western part of the graben. In addition, the direction and magnitude of observed strike-slip values are compatible with those of a simple 4 blocks model (Cornu and Bertrand, 2005b). Although these lateral motions are mainly a function of fault orientation, an oblique extension component is required for the development of the observed fault pattern.
From the backward and forward models it appears that opening of the URG involved a component of oblique extension, and that the central Rhine River fault played a major role during the rifting history. Therefore, the Rhine River fault deserves further attention in future seismotectonic studies as it cuts densely populated and industrialized areas.

6 Geothermal prospectivity

The most favourable conditions for geothermal systems prevail in the active volcanis areas. Proven high enthalpy geothermal systems exist along the western coast of Italy, in the Tuscan (and Latium??) region related to Late Quaternary volcanism, and in the Aegean region in Milos and Nikonos islands related to recent volcanism. Medium enthalpy reservoirs can be found in Western Turkey, in the Simav region related to Quaternary basaltic volcanism. The southernmost segment of the volcanic arc in the Eastern Carpathians also has some geothermal potential as the last eruption occurred 20 ky ago, though the existence of any reservoir is not proven.
In other places EGS and UGS can be found. The circum-Mediterranean region tectonically is the most active area in Europe as shown by numerous earthquakes (Fig..), and horizontal and vertical movements of smaller lithospheric blocks relative to each other (McKenzie, 1972, Le Pichon et al, Mueller et al., ...Cloetingh et al., 2005). The active fault zones may provide permeable pathways for groundwater flow to penetrate to large depth. The potentially available reservoir host rocks coupled with high heat flow characterizing the region (Fig...), make the circum-Mediterranean region favourable for the development of EGS (e.g Soults). Large fault zones located in areas with extensional stress field are the most perspective places (Italian penninsula, Aegean region including western Anatolia, Fig. xx), because due to the extensional stress the cracks and fractures are more likely open than in case of compressional stress. Faults in strike-slip tectonic regime might be also prospectous as the termination of strike-slip faults, crossing of faults may result in local extension and open fractures (e.g. geothermal field of in California, or potential in the Pannonian basin).
High enthalpy UGS can be found in deeply buried fractured, karstified carbonatic rocks in the Pannonian basin. Such rocks are found in 4-5 km depth overlain by Neogene

sediments. The buried carbonatic reservoirs comprise closed systems with water temperature of 200 C, and overpressure higher than 30 Mpa.
EGS can be developed in the molasse sediments of the foreland of the Alps and Carpathian arc. Subduction of the European lithosphere beneath the Alps and Carpathians during the Tertiary and Neogene, respectively, resulted in 5-10 km deep foreland basins filled with clastic sediments. Though the porosity of the sediments decreases with depth due to compaction, applying hydraulic fracturing may increase the porosity and permeabilty to create EGS.

Back-arc extension and extensional collapse: basins (passive+volcanism) possible in Volcanic areas with calc-alkaline volcanism (Lardello). Greeks Milos island is also in backarc setting. Must be special volume on Milos in geothermics in 1999 or so.

7 References

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Figure captions

Fig. 1. Surface heat flow and tectonic provinces in Europe. To be recompiled and combined in one figure. Tectonic features need to be outlined: ECRIS=European central rift system, URG: upper rhine graben, PB: pannonian basin; RVG: roer valley graben; MC: Massif Central. TS: Tyrenean Sea. New compilation of heat flow by D. Bonte to be prepared in Budapest with Lazlo. Add locations of ongoing geothermal operations

Fig. 2 Construction of a conduction dominated continental lithosphere geotherm from the steady state relation between surface heat flow, stratified conductivity, and heat production in crust and mantle lithosphere . For various values of lithospheric thickness (60, 90,120, 150 km).Crustal thickness is 30 km. Lithosphere parameters after Van Wees et al., 2007, assuming 40% of the surface heat flow is caused by radiogenic heat production in the upper crust.
The surface heat flow corresponds to 100,74, 60, 51 mW m^{-2}. The red curve shows a geotherm of a 120 km thick lithosphere with the effect of abnormal high heat production in the upper crust as a consequence of granite intrusions such as observed in the Bohemian Massif marked by heat flow values up to 107 mW m^{-2} (Förster and Förster, 2000),

Fig. 3 Steady state lithosphere thickness derived from surface heat flow and compositional input of the crust and mantle lithosphere (cf. Table 1, Fig. 2). Panel a) crustal thickness b) surface heat flow and c) lithospheric thickness. From where?

Fig. 4 Seismic velocity anomalies at 100km under Europe for P waves (top left) and S waves (bottom left). Panels on the right show temperatures at 100km depth estimated from the P and S velocity anomalies. The assumption that all velocity anomalies can be attributed to variations in temperature appears to be reasonable in view of the similarity in thermal structure obtainedfrom P- and S-wave velocities (after Goes et al., 2000).

Fig. 5 Integrated model of lithospheric thickness in Europe, based on seismic, thermal, MT, electromagnetic and gravity interpretations. In general, a direct comparison of lithospheric thickness values, constrained by different techniques, is not valid, as they are based on measurements of diverse physical parameters. The difference between the thicknesses of 'seismic' and 'thermal' lithosphere can be up to 40–50 km (Jaupart & Mareschal 1999), which approximately corresponds to the thickness of the transition zone between pure conductive and pure convective heat transfer (from Artemieva, 2006)

Fig. 6 detailed analysis of transition of Western European lithosphere of 100 km thickness to lithosphere thickness of 200 km in East Eurpean Craton. Interpreted lithosphere thickness is based on integrated modeling, taking into account surface heat flow measurements, detailed compositional model of sediments and crust and information from tomographic data (From Norden et al, 2007)

Fig. 7. Location map of the European Cenozoic 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, Upper Rhine Graben; OW, Odenwald; VG, Vosges (after De` zes et al., 2004).

Fig. 8. Outline map of back-arc extensional basins in Mediterranean and Carpathian system.

Fig. 9 Cartoon of the Late-stage scenario of the Pannonian Basin-Carpathians system during the Late Cenozoic (after Horvath and Cloetingh, 1996), illustrating its back-arc setting, involving slab detachment and post-orogenic collapse.

Fig. 10 cartoon of two-layered stretching model, representative for subcrustal mantle attenuation in excess of crustal stretching, either related to mantle plumes as for the ECRIS system or slab detachement in Back-arc basins.

Fig. 11 Synthetic heat flow model results marked by crustal stretching of 1.44, with synrift phase from 220-200 Ma, followed by a 200 My postrift phase, illustrating siginificant thermal effects of rifting up to 30 My after rifting. In uniform stretching onshore basins are predicted to be marked by flat or decreasing heat flow (yellow curve), thermally elevated compared to the theoretical steady state heat flow for the lithosphere thickness (cf Fig. 2). However, back-arc basins developed on orgenic wedges these are marked by elevated heat flow because they can remain relatively sediment starved (pink curve). In addition parts of the ECRIS and most back-arc basins are marked by non-uniform extension ($\beta>\delta$) resulting in strong elevated heat flow (red curve). Please not that heat flow peak arrives later because of time delay in diffusion of heat through the crust.

Fig. 12 Basin Modelling results for the Pannonian Basin. High postrift subsidence and maturity and temeperature data are explained through a two-layered model ($b<d$), in agreement with strong mantle upwelling. Thermal effects of the two-layered stretching have resulted in a marked elevated geotherm and surface heat flow compared to initial geothermal conditions.

Fig. 13 Predicted thermal structure and surface heat flow of the Eastern Alps along the TRANSALP deep seismic line for the present-day crustal configuration. Modified from Willingshofer and Cloetingh (2003).

Fig. 14 Ter Voorde modeling heat flow effects of exhumation related to extension

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Fig.15. 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. Add locations of ongoing geothermal operations, extend to include Iceland. After Cloetingh et al., (2005)

Fig.16 Intraplate stress map for Europe, displaying the present-day orientation of the maximum horizontal stress (SHmax). Different symbols stand for different stress. The length of the symbols represent data quality. Background shading indicates topographic elevation (darker is higher). The map was extracted from the World Stress Map database (World Stress Map, 2004)

Fig. 17. From crustal thickness (top left) and thermal structure (top right) to lithospheric strength (bottom): conceptual configuration of the thermal structure and composition of the lithosphere, adopted for the calculation of 3D strength models. (From Cloetingh et al. 2005)

Fig. 18 Thermo-mechanical prediction of present day geotherm and rheology based on kinematic extension model presented in Fig. 12. Similar results were obtained by Sachsenhofer et al. (1997) and Lankreijer et al. (1997) for the Styrian basin and Danube basin, respectively, (subbasins in the Pannonian basin) assuming steady state temperature in the lithosphere, based on present day heat flows in the order of …. For comparison a rheological profile calculated with a irrealistically low surface heat flow of 50 mW/m^2 and assuming steady state temperature is also shown. The decrease of the lithospheric temperature and surface heat flow would result in considerable strengthening of the lithosphere, not compatible with earthquake depth distributions.

Fig. 19 Depth-dependent rheological model. (upper panel).and integrated strength evolution (lower panel).for the evolution of a rifted upper plate passive margin (after Ziegler et al., 1998, their model).. Area of the strength envelope (upper panel) gives intraplate force (Integrated strength, lower panel) necessary to obtain lithospheric deformation. Integrated strength values for extensional deformation are considerably less than for compression. Initially integrated strength decreases during stretching, but increases considerably ca. 50 Ma after stretching.

Fig.. 20 Strength Map. Integrated lithosphere 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. Rheological rock parameters are based on Carter and Tsenn (1987). The adopted bulk strain-rate is 10-16 s-1, compatible with constraints from GPS measurements (see text). Contours represent integrated strength in compression for total lithosphere. The main structural features of Europe are superimposed on the strength maps (after Ziegler, 1988; Dezes et al., 2004). (From Cloetingh et al. ESR)

Fig. 21. Spatial comparison of crustal seismicity and integrated crustal strength. Earthquake epicentres from the NEIC data center (NEIC, 2004), queried for magnitude N2 and focal depths b35 km. (after Cloetingh et al., 2006)

Fig. 22 Topographic map of the Alpine foreland, extracted from ETOPO-2. The red lines denote the location of the rheological cross sections. (bottom) Rheological cross-sections along the Rhine graben segments of ECRIS (Amsterdam-Basel) and east-west oriented along the Eifel and Rhine graben (section D). 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^{o}C and is indicated by the solid black line. (after Cloetingh et al., 2006)

Fig. 23 a–c. EET of the European lithosphere calculated using the Burov and Diament (1995) approach thickness of the mechanically strong part of the entire lithosphere (a), thickness of the mechanically strong part of the mantle lithosphere (b), thickness of the mechanically strong part of the upper crust (c). The grid shows the areas where upper and lower crust are mechanically coupled.See text for further explanations (after Tesauro et al., 2007).

Fig. 24 Residual mantle anomalies of the gravity field obtained after removal of the crustal effect from the observed field. (a) Total anomaly.(b) Regional component (Lb2000 km) of the residual mantle anomalies correlating with specific tectonic structures. See text for further details.
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/). (after Tesauro et al., 2007)

Fig. 25 strain distribution Europe (after Tesauro et al., 2007)

Fig.26 (a) Topography of the Lower Rhine Graben (LRG) area in colour-coded relief map. Data are from the GTOPO30 global data set. Earthquake epicenters are from the ORFEUS data center, and are shown as white dots, with dot size indicating magnitude. White star gives the location of the 1992 ML=5.8 Roermond earthquake. Red lines depict Base Tertiary faults, yellow lines depict Base Miocene faults. Faults are digitized from Geluk et al. (1995) and De Mulder et al. (2003). Main tectonic structures: RVG=Roer Valley Graben; PB=Peel Block; VB=Venlo Block; CB=Campine Block; EB=Erft Block; SLB=South Limburg Block; PBF=Peel Boundary Fault; FF=Feldbiss Fault. (b) Cross section through the Roer Valley Rift System (after Geluk et al., 1995). Location of profile is indicated by the white line in Fig. 41a.

Fig.27 (a) Base Tertiary structural map of Roer Valley Rift System (RVRS). (b) Oligocene development of the RVR, inferred from the thickness of Late Oligocene sediments, subsidence analyses and distribution of active faults. White arrows indicate extension direction. (c) Miocene-Quaternary development of the RVR, inferred from the thickness of Neogene sediments, subsidence analyses and distribution of active faults. White arrows indicate the extension direction (after Dirkzwager et al., 2006).

Fig. 28 Finite element modelling results for the Roer Valley section for various values for friction angle of the major faults (after Dirkzwager et al., 2000).

Fig. 29 a) Sketch map of the southern part of the Upper Rhine Graben, showing the western and eastern main border faults, and the area covered by the numerical models shown in Fig. 23, denoted by the red rectangle. b) Initial (i.e. pre-rift) geometry of the studied graben segment, used for the forward model. The two lateral blocks correspond to the future graben shoulders (i.e. Vosges and Black Forest Mountains to the west and the east, respectively). Their contact zones with the central block correspond to the border faults delimiting the Upper Rhine Graben. Right Panel: Results of the forward model. Both purely orthogonal and partly oblique extension scenarios were tested. Results are presented in terms of the first invariant of strain E1 (sum of the diagonal terms of the strain tensor): c) orthogonal and d) oblique extension. And in terms of the Drucker-Prager (DP) failure criterion (numerical equivalent of a Mohr-Coulomb criterion, see Appendix B for further details): e) orthogonal and f) oblique extension. Cartesian coordinates are in meters (after Cloetingh et al., 2006)

Fig. 30 Left panel: Construction of initial multi-block domain used for backward modelling (for location see Fig. 22). (a) Present-day geometry of the geological domain. (b) Finite elements domain built from this geological domain. (c) Contact sequence of movement used for backward modelling, the contact borders, (C).are borders containing the “slave” nodes and free borders, (F) those containing “target” nodes (after Cloetingh et al., 2006).

Fig. 30 Right panel: Results of backward model, presented in terms of displacement (m), showing the top surface of all blocks, (a) displacement along the E-W x-axis, (b) displacement along the N-S y-axis, (c) displacement along the vertical z-axis (after Cloetingh et al., 2006).