Neogene uplift of Variscan Massifs in the Alpine foreland: Timing and controlling mechanisms

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Abstract

The European Cenozoic rift system (ECRIS) and associated fault systems transect all Variscan Massifs in the foreland of the Alps. ECRIS was activated during the Eocene in the foreland of the Pyrenees and Alps in response to the build-up of collision-related intraplate stresses. During Oligocene and Neogene times ECRIS evolved by passive rifting under changing stress fields, reflecting end Oligocene consolidation of the Pyrenees and increasing coupling of the Alpine Orogen with its foreland. ECRIS is presently still active, as evidenced by its seismicity and geodetic data.
Uplift of the Massif Central and the Rhenish Massif, commencing at the Oligocene-Miocene transition, is mainly attributed to plume-related thermal thinning of the mantle-lithosphere. Mid-Burdigalian uplift of the SW-NE striking Vosges-Black Forest Arch, that has the geometry of a doubly plunging anticline breached by the Upper Rhine Graben, involved folding of the lithosphere. Late Burdigalian broad uplift of the northern parts of the Bohemian Massif reflects lithospheric buckling whereas late Miocene-Pliocene uplift of its marginal blocks involved reactivation of pre-existing crustal discontinuities. Crustal extension across ECRIS, amounting to no more than 7 km, was compensated by a finite clockwise rotation of the Paris Basin block, up warping of the Weald-Artois axis and reactivation of the Armorican shear zones. Intermittent, though progressive uplift of the Armorican Massif, commencing in the Miocene, is attributed to transpressional deformation of the lithosphere.
Under the present-day NW-directed compressional stress field, that developed during the early Miocene and further intensified during the Pliocene, the Armorican Massif, the Massif Central, the western parts of the Rhenish Massif and the northern parts of the Bohemian Massif continue to rise at rates of up to 1.75 mm/y whilst the Vosges-Black Forest arch is relatively stable.
Uplift of the Variscan Massifs and development of ECRIS exerted strong controls on the Neogene and Quaternary evolution of drainage systems in the Alpine foreland.

Introduction

In the Alpine foreland, the internal parts of the Variscan Orogen are exposed in the Massif Central, the Armorican and Bohemian Massifs, the Odenwald, Vosges and Black Forest, whereas its external fold-and-thrust belt crops out in the Rhenish Massif and Ardennes. These Variscan arches, which prior to their uplift were covered to variable degrees by Mesozoic platform sediments, are closely associated with the European Cenozoic Rift System (ECRIS) and related fault systems (Fig. 1; Ziegler, 1990). Uplift of these arches, that generally commenced during the Neogene and persisted during the Quaternary, is intricately related to the evolution of ECRIS that was controlled by collision-related compressional intraplate stresses, accompanied by mantle-plume activity (Dèzes et al., 2005).

ECRIS extends from the coast of the North Sea to the Mediterranean over a distance of some 1100 km and transects most of the Variscan massifs (Fig. 1). ECRIS consists of the Rhine and Massif Central-Rhône Valley rift systems, which are linked by the Burgundy and Paris Basin transfer zones, and the shallow Eger (Ohre) Graben of the Bohemian Massif. The Rhine Rift System, forming the northern parts of ECRIS, includes the Upper Rhine Graben that crosscuts the Vosges-Black Forest Arch, and the Roer Valley-Lower Rhine and Hessian grabens that transect the Rhenish Massif. The Massif Central-Rhône Valley rift system, forming the southern parts of ECRIS, consists of the Limagne, Roanne and Forez grabens of the Massif Central, and the Bresse and Valence, Alés, Manosque and Camarque grabens of the Rhône Valley that flank this massif to the East (Ziegler, 1994; Prodehl et al., 1995; Séranne 1999; Merle & Michon, 2001; Sissingh, 2003a; Dèzes et al., 2004). The Armorican Massif is linked to ECRIS by a seismically active system of northwest striking Variscan shears that branches off from the Massif Central (van Vliet-Lanoë et al., 1997; Cloetingh et al., 2005).

The Variscan arches of the Alpine foreland are partly characterized by major topography (Fig. 2). For instance the Vosges and Black Forest rise to 1424m and 1493m, respectively, whereas the alluvial plain of the intervening URG is located at 200-220 m above MSL. Similarly, the Massif Central attains elevations of 1634 m whilst the surface of the Limagne Graben does not exceed 330m. The Bohemian Massif peaks in the Bohemian Forest and the Erzgebirge at 1452 and 1214m, respectively. By contrast, peak elevations of the Rhenish Massif are considerably lower (Taunus 880 m; Hunsrück 816 m, Snow Eifel 697 m), whilst elevations of the low-lying Armorican Massif hardly reach 250 m. Geodetic data show that most of these arches are currently rising at rates of up to 1.75 mm/y with seismic activity being concentrated on the Rhine-Rhône rift system, the Armorican shear zones and the northern parts of the Bohemian Massif (Fig. 3; Cloetingh et al., 2005).

Development of these arches appears to be closely related to the evolution of ECRIS that began during middle and late Eocene times in response to the build-up of compressional stresses that were exerted by the evolving Pyrenean and Alpine orogens on their northern forelands (Dèzes et al., 2004). However, whilst the timing and level of volcanic activity associated with the different arches and the timing of their uplift vary considerably, mechanisms controlling their uplift differ as well. In the following we summarize the evolution of the different arches that are associated with ECRIS, assess mechanisms controlling their uplift, and discuss implications for the development of drainage systems on the European Platform.

Crustal and lithospheric configuration and mantle plumes

The depth map of the crust/mantle boundary, given in Fig. 4, clearly illustrates that ECRIS is superimposed on a broad zone of Moho shallowing that can only be partly attributed to Cenozoic crustal extension which amounts to no more than 7 km across the Limagne-Bresse graben system and the Upper Rhine Graben and tapers to zero at the NW end of the Roer Valley Graben (Dèzes et al., 2004).

The Massif Central Arch is characterized by irregular shallowing of the Moho to as little as 24 km that must be attributed to a combination of Permo-Carboniferous crustal thinning, particularly along the Sillion Houillier, Eocene-Oligocene crustal extension and Neogene regional uplift. By contrast, the Armorican Massif is characterized by crustal thicknesses in the 30-34 km range (Ziegler & Dèzes, 2005). At the level of the Moho, the Vosges-Black Forest Arch coincides with the culmination of a SW-NE trending anticlinal feature that extends from the northern margin of the Massif Central via the Burgundy transfer zone into the northern parts of the Bohemian Massif. The Rhenish Massif, that straddles the triple junction of the Upper Rhine, Hessian and Roer Valley-Lower Rhine grabens, is not associated with a distinct broad shallowing of the Moho, though the different graben axes are clearly expressed by Moho shallowing. The shallow Eger Graben separates the southern parts of the Bohemian Massif, which are characterized by an up to 38km thick crust, from its northern parts in which crustal thicknesses vary between 29 and 32km.

In the wider ECRIS area, the thickness of the lithosphere (Fig. 5) varies between 100-120 km but decrease to as little as 50-60 km beneath the volcanic fields of the Massif Central (Sobolev et al., 1997) and the Rhenish Massif (Fig. 6; Prodehl et al., 1995; Basbuska & Plomerova, 1992) and to 80 km beneath the Eger Graben on the Bohemian Massif (Fig.

7; Babuska & Plomerova, 2001). By contrast, the Vosges-Black Forest Arch is characterized by a 100 km thick lithosphere (Fig. 6a; Achauer & Masson, 2002).

Mantle tomography images beneath the Massif Central, the Rhine Rift and the Bohemian Massif a system of low velocity anomalies in the lower parts of the upper asthenosphere (Fig. 8). These are interpreted as plume heads that have spread out above the 410 km discontinuity (Goes et al., 1999; Spakman, 2004; Sibuet et al., 2004; Dèzes et al., 2005). From these anomalies secondary, relatively weak plumes presently rise up beneath the Eifel (Ritter et al., 2001) and the Massif Central (Granet et al., 1995). Presumably these deep-seated anomalies started to develop during the Paleocene in conjunction with the activation of the NE Atlantic and Iceland plumes and subsequently evolved further, as evidenced by persisting volcanic activity, particularly on the Rhenish and Bohemian Massifs (Dèzes et al., 2004, 2005). As in time a shift in areas of main volcanic activity can be observed (e.g. decreasing towards the end of the Oligocene on the Bohemian Massif [Ulrych et al., 1999] and Western Rhenish Massif, increasing during the early and middle Miocene on the Eastern Rhenish Massif and shifting in the Quaternary back to the Western Rhenish Massif [Lippolt, 1983; Jung, 1999]; increasing on the Massif Central during the middle Miocene and Pliocene [Michon & Merle, 2001]), it is likely that the supply of partial melts through secondary upper mantle plumes was not steady state but pulsating and entailed a shift in their location.

Broad, and still on-going uplift of the Rhenish Massif is mainly attributed to thermal thinning of its lithospheric mantle and to the buoyant load of the Eifel mantle plume (Fig. 6a; Jung. 1999; Garcia-Castellanos et al., 2000; Ritter et al., 2001). In this respect it should be noted that the zone of lithospheric mantle thinning extends in an SW-NE direction over a distance of more than 200 km across the entire Rhenish Massif (Prodehl et al., 1995), and thus is considerably broader than the modern Eifel plume that has a diameter of some 100 km (Ritter et al., 2001). This is presumably the effect of earlier plume activity in eastern parts of the Rhenish Massif. Similarly, plume-related thinning of the lithospheric mantle apparently contributed significantly to the Neogene and Quaternary uplift of the Massif Central (Fig. 6b; Granet et al., 1995; Sobolev et al., 1997), whereas the Armorican Massif is characterized by lithospheric thicknesses in the range of 120-150 km (Judenherc et al., 2002). For the Bohemian Massif, tomographic data indicate thinning of the lithosphere from 100 km under its northern and southern parts to 80 km under the Eger Graben (Fig. 7; Babuska & Plomerova, 2001). This may be attributed to thermal thinning of the lithospheric mantle during the Oligocene main phase of volcanic activity in the Eger Graben area (Ulrych et al., 1999; Dèzes et al., 2004). Although there are indications for asthenospheric upwelling beneath the Eger Graben, there is no evidence for an active plume penetrating the lithosphere that could underlie the Plio-Quaternary resumption of volcanic activity in this area (Babuska et al., 2005; Ulrych et al., 1999). For the Vosges-Black Forest Arch, seismic tomography indicates a lithospheric thickness of about 100 km (Fig. 6a) and that only very small amplitude low-velocity anomalies occur in the depth range of 67-107 km (Achauer & Masson, 2002). Correspondingly, thermal thinning of the lithospheric mantle cannot be held responsible for the uplift of the Vosges-Black Forest and Armorican arches and perhaps contributed only marginally to the uplift of the Bohemian Massif (Dèzes et al., 2004).

Uplift history of ECRIS arches

Development of ECRIS in the foreland of the Alps and Pyrenees was preceded by a latest Cretaceous-Paleocene phase of intraplate compression (Fig. 9a), involving broad lithospheric buckling (e.g. areas of Massif Central and Vosges-Black Forest) and, by reactivation of pre-existing crustal discontinuities, inversion of Mesozoic tensional basins (e.g. West Netherlands and Lower Saxony basins) and upthrusting of basement blocks (e.g. Bohemian Massif). This caused profound truncation of Mesozoic platform sediments and the development of a regional erosional unconformity (Ziegler, 1990; Dèzes et al., 2004). During the middle and late Eocene initial rifting phase of ECRIS (Figs. 9b, 9c), sedimentation in the evolving graben segments was dominated by fluvial and lacustrine, partly evaporitic environments (Sissingh, 2003a), with the local development of conglomeratic fans (e.g. Upper Rhine Graben: Duringer, 1988), indicating flexural uplift of the graben flanks (Kusznir & Ziegler, 1992; van Wees & Cloetingh, 1996). This suggests that by these times the entire ECRIS region was still located above sea level. During the Oligocene and early Miocene (Fig. 9d & e) intermittent marine transgressions advanced from the North Sea-North German Basin into the Rhine Rift, and during the early Oligocene from the Alpine Foreland Basin into the Bresse and Massif Central grabens (Sissingh 2001, 2003a,b; Berger et al., 2005a). This shows that the different graben segments of ECRIS had subsided close to sea-level with the rift flanks forming emerging, though generally low relief highs, as indicated by the prevalence of fine-grained, low energy sediments accumulating in its grabens (Sissingh, 2003a).

Magmatic activity commenced in the ECRIS area, after a long Mesozoic lull, during the latest Cretaceous and Paleocene with the intrusion of olivine melilite and olivine nephelinite dykes in the Massif Central, the Vosges-Black Forest area and the Rhenish and Bohemian Massifs (Wilson et al., 1995; Lippolt, 1983; Ulrych et al., 1999; Michon & Merle, 2001; Keller et al., 2002). This reflects very low-degree partial melting of the lithosphere/asthenosphere boundary layer in response to a temperature increase of the asthenosphere to above ambient levels (Wilson et al., 1995), presumably in conjunction with activation of the Iceland plume (Ziegler, 1990; Bijwaard & Spakman, 1999) and further development of the NE-Atlantic plume (Hoernle et al., 1995). During the Eocene, volcanic activity persisted in the western parts of the Rhenish Massif and along its southern margin (Lippolt, 1983), as well as on the Bohemian Massif (Ulrych et al., 1999), but was at a low level on the Massif Central (Michon & Merle, 2001) and in the area of the Vosges-Black Forest (Keller, et al., 2002). During the Oligocene, volcanic activity increased on the Rhenish Massif, shifting to its northern and eastern parts (Lippolt, 1983; Jung, 1999). On the Bohemian Massif, volcanic activity peaked during the early and middle Oligocene and decreased during the late Oligocene-early Miocene (Ulrych et al, 1999). By contrast, there is only scattered evidence for late Oligocene volcanic activity on the Massif Central (Michon & Merle, 2001).

By early Oligocene times, the surface of the Massif Central and the Rhenish Massif was still located close to sea level, as evidence by repeated marine incursions into the grabens of the former (Merle et al., 1998; Michon & Merle, 2001; Merle & Michon, 2001; Sissingh, 2001) and the occurrence of remnants of marine and brackish sediments on the latter (Ziegler, 1990; 1994; Sissingh, 2003b), whilst in the area of the Vosges-Black Forest the rift shoulders were already elevated (Hinsken et al., 2005). By contrast, the Bohemian Massif remained a positive feature throughout the Cenozoic and was subjected to continued weathering and denudation (Malkovsky, 1979). The Armorican Massif remained close to sea level during Oligocene to early Pliocene times (Ollivier-Pierre et al., 1993; van Vliet-Lanoë et al., 1998a).

Massif Central

At the end of the early Oligocene, when marine incursions into the grabens of the Massive Central ceased, lacustrine conditions prevailed during the late Oligocene and early Miocene until they gave way to fluvial conditions at the transition to the middle Miocene when the north-directed Loire, Allier and Cher drainage system came into evidence (Fig. 10). After an early Burdigalian unconformity, only minor late-early and middle Burdigalian fluvial sediments were deposited in the Limagne Graben, probably representing the last syn-rift sediments (Merle et al., 1998; Michon & Merle, 2001; Sissingh, 2001). On the Massif Central, volcanic activity increased dramatically during the middle Miocene (14 ma), shifted to its southern parts and peaked during the late Miocene, whereas in its northern parts volcanism was interrupted during the late-middle Miocene (12 Ma) (Michon & Merle, 2001). Gradual, though slow uplift and northward tilting of the Massif Central commencing during the late Oligocene, becoming effective in the middle Miocene was presumably driven by plume-related thinning of the lithospheric mantle (Michon & Merle, 2001). During the Pliocene (5.5 Ma), volcanic activity resumed in the northern parts of the Massif Central where it peaked between 4 and 1 Ma, whereas in its southern parts major volcanism peaked for a second time during the late Pliocene and early Quaternary (3.5-0.5 Ma). After 0.5 Ma, volcanic activity gradually decreased to the present sub-active level (Michon & Merle, 2001). From about 3.5 Ma onwards, uplift of the Massif Central accelerated and still continues at rates of up to 1.75 mm/year with differential movements occurring between blocks delimited by ENE-trending fractures (Lenôtre et al., 1992). Under the present NW-directed stress field, the Massif Central is subjected to transtensional and transpressional deformation (Nicolas et al., 1990; Delouis et al., 1993). Progressive uplift and northward tilting of the Massif Central apparently counteracted transtensional subsidence of its graben systems, thus impeding accumulation of thick Pliocene and Quaternary sequences, as seen in the Bresse Graben (Dèzes et al., 2004).

Middle Miocene to recent uplift of the Massif Central can be mainly attributed to plume-related thermal thinning of the lithospheric mantle (Fig. 6b), thermal expansion of the remnant lithosphere and the buoyant load of the mantle-plume impinging on it (Granet et al, 1995; Dèzes et al., 2004). Uplift of this arch played a major role in the development of the Loire, Allier, Cher and Rhône drainage system (Fig. 10; Sissingh, 2001).

Armorican Massif

During the Mesozoic, the western parts of the Armorican Massif formed a low-relief high that was subjected to profound weathering whilst its eastern parts were overstepped during the Late Cretaceous. During the Paleocene and Eocene, the entire massif formed a positive feature on the southern margin of which marine transgressions encroached during the early and middle Eocene (Durand & Esteoule-Choux, 1974). Early Oligocene, transtensional reactivation of pre-existing NW-SE trending Variscan shear systems led to the opening of a narrow marine channel that transected the Massif and in which up to 120 m of clays and carbonates were deposited (Ollivier-Pierre et al., 1993). Repeated transgressions encroached on the Armorican Massif during the late Miocene and early Pliocene along partly fault-controlled valleys (Fig. 11; van Vliet-Lanoë et al., 1998a). The late Pliocene and Quaternary uplift of the Armorican Massif was accompanied by strong reactivation of its essentially NW-SE trending crustal discontinuities, as evidenced by differential river incision and river captures (Bonnet et al., 1998; Brault et al., 2001).

A diffuse belt of seismic activity extends from the Massif Central into the Armorican Massif, with earthquakes rarely exceeding 4.0 (Fig. 3). Present-day differential uplift rates of the order of 0.5-1.1 mm/y can be related to transpressional and transtensional reactivation of the Palaeozoic Armorican shear zones (Nicolas et al., 1990; van Vliet-Lanoë et al., 1997; Bonnet et al., 1998; Lenôtre et al., 1999; Cloetingh et al., 2005) and buckling of the lithosphere under the prevailing NW-directed stress field (Müller, 1997).

Rhenish Massif

The NNE striking Hessian grabens (HG) transect the eastern parts of the Rhenish Massif in the prolongation of the Upper Rhine Graben (URG) whereas the NW striking Lower Rhine-Roer Valley Graben (LRG-RVG) transects its western parts (Fig. 1). The triple junction of these grabens is located in the Mainz-Frankfurt area. Although the sedimentary fill of graben segments crossing the Rhenish Massif was partly eroded during the Miocene and younger uplift of this arch, erosional remnants still permit to reconstruct their evolution (Sissingh, 2003b).

Fission-track data indicate that Cretaceous uplift and cooling of the Rhenish Massif (120-80 Ma) had ended during the Campanian (Glasmacher et al., 1998) whilst along its southern and northern margins the Saar-Nahe, and the West Netherlands-Roer Valley and Lower Saxony basins, respectively, were inverted during the Coniacian-Santonian and late Paleocene phases of intraplate compression (Fig. 9a). However, the resulting topography was rapidly degraded, as evidenced by the transgression of late Paleocene and Eocene series across the inversion axes of the West Netherlands and Lower Saxony basins (Ziegler, 1987, 1990; de Lugt et al., 2003; Michon et al., 2003; Worum et al., 2005).

During the early Oligocene the URG propagated northward into HG and LRG-RVG (Sissingh, 2003b; Michon et al., 2003). This paved the way for intermittent Oligocene to middle Miocene marine connections between the URG and the North Sea-North German Basin. During the early Oligocene, when a sea way extended from the North German Basin via the HG into the URG, transgressions advanced across the graben shoulders (Rhön, Westerwald and Mainz basins) and linked up via the Mosel area with the Paris Basin (Fig. 12a). This indicates that by this time much of the Rhenish Massif still formed a low-lying peneplain, except for the volcanic Eifel area (45-24 Ma). During the late Oligocene, marine communications between the North German Basin and the URG via the HG were interrupted when volcanic activity increased on the eastern parts of the Rhenish Massif (Rhön 26-11 Ma, Westerwald 28-5 Ma, Siebengebirge 28-6 Ma). However, a new marine gateway opened during the latest Oligocene between the North Sea Basin and the URG via the LRG. Along this axis, intermittent marine incursions advanced again into the URG during the Aquitanian, Burdigalian and Langhian (Sissingh, 2003b; Berger et al., 2005b). Communications with the Paris Basin via the Mosel depression were closed towards the end of the Oligocene when uplift of the Western Rhenish Massif commenced and the Mosel drainage system began to develop (Fig. 12b). Uplift of the eastern parts of the Rhenish Massif accelerated during the late Oligocene and early Miocene with the activation of volcanic activity in the Vogelsberg (24-9 Ma) and the Hessian Depression (20-8 Ma). Whilst in the eastern parts of the Rhenish Massif volcanic activity gradually abated in the course of the late Miocene, volcanism resumed in its western parts during the Quaternary (Eifel 0.7-0.01 Ma) (Lippolt, 1983; Jung, 1999). Late Miocene to Recent domal uplift of the Rhenish Massif, centred on its western parts, amounted to 400-500 m (Demoulin et al., 2005). This was accompanied by deep entrenchment of such antecedent rivers as the meandering Meuse, Mosel and Lahn and the somewhat more linear Middle Rhine (Negendank, 1983; Westaway, 2002a). Of this uplift, about 250 m can be attributed to the last 0.8 Ma (Meyer & Stets, 2002). By contrast, the eastern parts of the Rhenish Massif remained relatively stable during Pliocene-Quaternary times. Geodetic data show uplift rates of up to 1.6 mm/y for the western parts of the Rhenish Massif, decreasing to zero in its eastern parts (Mälzer et al., 1983).

Uplift of the Rhenish Massif, commencing in the early Miocene and persisting to the present, can be attributed to plume-related thermal thinning of the lithospheric mantle (Fig. 6a; Prodehl et al., 1995), thermal expansion of the remnant lithosphere and the buoyant load of the present-day Eifel plume (Garcia-Castellanos et al., 2000; Ritter et al., 2001). Crustal-scale folding and/or reactivation of Variscan thrust faults (Ahorner, 1983; Hinzen, 2003) under the presently prevailing northwest-directed compressional stress field (Müller et al., 1997) presumably plays a contributory role (Dèzes et al., 2004). Progressive uplift of the Rhenish Massif and Ardennes caused entrenchment of the antecedent rivers Mosel, Lahn, Rhine and Meuse (Fig. 2), the sedimentary load of which was deposited in the Roer Valley Graben and Rhine-Meuse delta (van Balen et al., 2000).

Bohemian Massif

Starting in the late Turonian and culminating during the Paleocene, the Rocky Mountain-type array of basement blocks forming the Bohemian Massif was upthrusted, involving reactivation of Late Palaeozoic and Mesozoic crustal discontinuities (Fig. 9a; Malkovsky, 1987; Ziegler, 1990; Wagner et al., 1997; Ziegler et al., 1998; 2002). This was accompanied by the injection of olivine melilite and olivine nephelinite dykes (79-49 Ma) (Adamovic & Coubal, 1999; Ulrych et al., 1999). Subsequently the Bohemian Massif was subjected to profound weathering and erosion, resulting in the development of a peneplain on which only locally thin late Eocene-early Oligocene fluvial and lacustrine clastics were deposited, whilst its southern flank was overstepped by marine series of the Alpine-Carpathian foreland basin (Ziegler, 1990; Wagner, 1998).

During the early Oligocene the main volcanic episode of the Eger volcano-tectonic zone commenced, lasted until the early Miocene (42-20 Ma) (Adamovic & Coubal, 1999; Ulrych et al., 1999), and preceded the main subsidence phase of the Eger Graben (Malkovsky, 1987). During the latest Oligocene to Burdigalian, when volcanic activity had gradually abated and a northerly-directed drainage system had developed within the Bohemian Massif, up to 500 m of lacustrine and fluvial clastics, including coals, accumulated in the Eger volcano-tectonic zone under a mildly tensional setting (Fig. 13a; Malkovsky, 1975, 1979). Extension intensified, however, after sedimentation in the Eger Graben had ended around 18 Ma. During the middle and late Miocene, the Eger Graben and the northern parts of the Bohemian Massif were uplifted and subjected to erosion, presumably in response to lithospheric folding that was accompanied by a late Miocene compressional phase (Adamovic & Coubal, 1999). During the early Langhian (early Badenian, ±16-15 Ma) temporary marine incursions advanced northward from the Alpine-Carpathian foreland basin along river valleys into the southern and eastern parts of the Bohemian Massif (Fig. 13b), reflecting that these were still located close to sea level, and that the intra-Bohemian watershed had shifted northward (Malkovsky 1979; Suk, 1984). During the late Langhian-early Serravallian (±15-13 Ma), compressional reactivation of the Bohemian Massif fault systems commenced and persisted into the Quaternary, causing disruption of the pre-existing peneplain, and by uplift of its marginal Thuringian-Bohemian and Bavarian Forest, Erzgebirge, Lusatian, Sudetic and Moravian blocks the gradual development of its present physiographic relief (Figs. 13c,d). In the process of this, remnants of early Langhian marine deposits were uplifted in the southern parts of the Bohemian Massif to as much as 600 m above MSL (Suk, 1984). Moreover, in the Upper Austrian Molasse Basin that flanks the Bohemian Massif to the south, some syn-flexural Oligocene normal faults were compressionally reactivated during late Miocene-Pliocene times (Wagner, 1996, 1998; Gunzenhauser et al., 1997).

In the domain of the Eger volcano-tectonic zone volcanism had decreased during the middle Miocene (16-12 Ma) but intensified again during the late Miocene and Pliocene (11.4-3.95 Ma) and lingered on to 0.26 Ma (Adamovic & Coubal, 1999; Ulrych, 1999). In the Eger Graben, uplift and minor extension resumed after the deposition of Pliocene fluvial clastics (Malkovsky, 1979).

Pliocene and Quaternary uplift of the Bohemian Massif exerted a strong control on the development and deep incision of its present-day mainly north-directed drainage system that, as compared to its middle Miocene drainage system (Fig. 13b), entailed an important south-eastward shift of the watershed between the Danube and North German-Polish drainage systems (Fig. 13d; Tyracek, 2001). For instance, terrace systems show that the river Ohre incised since 3.5 Ma (late Pliocene) by as much as 180 m into the sediments of the Eger Graben (Westaway, 2002a), whilst in the area of Prague, the river Vltava terraces indicate 160 m of incision during the last 1.9 Ma. The river Labe (Elbe) cuts in up to 380 m deep gorges across the essentially Oligocene Ceské Stredohori volcanic complex and through Cretaceous sandstones at the eastern termination of the Erzgebirge (Tyracek et al., 2004). Geodetic data indicate that under the presently prevailing northwest- to north-directed compressional stress regime (Müller et al., 1997) the northwestern parts of the Bohemian Massif are being uplifted whereas its central Cretaceous Basin appears to gently subside (Frischbutter & Schwab, 1995).

Mid-Miocene to recent uplift of the Bohemian Massif can be attributed to lithospheric buckling and the reactivation of pre-existing crustal discontinuities in response to the build-up of collision-related intraplate compressional stresses (Ziegler et al., 2002).

Vosges-Black Forest Arch

The Vosges-Black Forest Arch (VBFA), which is transected by the north-easterly striking URG, has at the top-basement level in a N-S direction an amplitude of some 2.5 km and a wavelength of 250 km with a steeper southern and a gentler northern flank (Fig. 6a). At the level of the Moho-discontinuity, this arch coincides with the culmination of a SW-NE striking anticlinal feature that extends from the Massif Central towards the Bohemian Massif (Fig. 4). Across the culmination of the VBFA the width of the URG narrows from about 60 km in the area of Strasbourg to 35 km north of Mulhouse from where it widens out southward to 63 km in the area of Basel. However, independent of the graben width the total extensional strain across the URG remains constant, not exceeding about 6-7 km (Brun et al., 1992; Durst, 1991; Dèzes et al., 2004; Hinsken et al, 2005).

Subsidence of the southern parts of the URG commenced during the middle Eocene and persisted into the early Miocene when it was interrupted only to resumed during the late Pliocene. During its middle Eocene (Barthonian) to earliest Oligocene (early Rupelian) initial rifting phases, the southern parts of the URG were occupied by a cyclically sediment starved and overfilled, intermittently hypersaline lacustrine basin, the depocentre of which coincided with the narrowest part of the graben, the so-called Potash Basin north of Mulhouse. Into this basin, conglomeratic fans began to prograde from the gradually uplifting rift flanks (Duringer, 1988). These coarse, torrential conglomerates consist primarily of Mesozoic components, reflecting progressive erosion of the sedimentary cover of the Vosges and Black Forest that formed the flexurally uplifted rift flanks of the URG. The occurrence of crystalline components in early Oligocene conglomerates along the western margin of the Potash Basin indicates that by this time the basement of the Vosges was exposed, whereas on the conjugate eastern margin erosion had cut only down into Lower Triassic series (Hinsken et al., 2005). As in this area the thickness of Mesozoic pre-rift sediments preserved beneath the syn-rift sediments of the URG is of the order of 1000 km (Lutz & Cleintuar, 1999), and Eocene syn-rift sediments attain a thickness of some 1000 m, the throw on the western border fault system of the URG must have already exceeded 2000 m. As to the north and south of the Potash Basin Eo-Oligocene conglomerates contain exclusively Mesozoic components, flexural uplift of the rift flanks apparently decreased with increasing graben width, reflecting extensional strain partitioning over wider areas (Hinsken et al., 2005).

During its Oligocene-early Miocene rifting phase, the entire URG subsided continuously (Roll, 1979; Villemin et al., 1986). During the late Rupelian marine conditions were established in the URG that persisted until early Chattian times. Renewed temporary marine transgressions occurred during the late Chattian and latest Chattian-early Aquitanian, originating from the North Sea-North German basins (Sissingh, 2003a; Hinsken et al., 2005). This shows that the surface of the URG had subsided below sea level whilst the prevalence of mudstones over sands reflects a generally low level of coarser grained clastics influx from the uplifted rift flanks (Doebl, 1970; Derer et al., 2003).

During the Neogene and Quaternary, fault-controlled subsidence of the northern parts of the URG continued without interruption, whereas its southern parts were uplifted and subjected to erosion during late Burdigalian to early Pliocene times (Roll, 1979; Durst, 1991; Lutz & Cleintuar, 1999; Schumacher, 2002; Derer et al., 2003; Dèzes et al., 2004; Haimberger et al., 2005). A corresponding erosional unconformity is clearly imaged by high-resolution reflection-seismic lines that were recorded on the river Rhine under the auspices of the EU-INTERREG III MoNit (Nitrate) and the EUCOR-URGENT Projects. These lines show that the hiatus across this unconformity gradually decreases northward from the Potash Basin and that 20 km north of Karlsruhe sedimentation was continuous during the Miocene and Pliocene (Fig. 14). The age of this unconformity is stratigraphically constrained as about mid-Burdigalian (± 18 Ma; end upper Hydrobia beds; Roll, 1979; Dèzes et al., 2004). This unconformity is southward progressively overstepped by late Burdigalian and younger beds. Significantly, these do not show major convergence in their onlap geometry nor any evidence for the development of additional unconformities within them, suggestive of progressive or intermittent uplift of the southern parts of the URG. This indicates that this unconformity developed in response to a discrete tectonic event of relatively short duration. Industrial reflection-seismic data show that development of this intra-Burdigalian unconformity involved regional uplift of the axial parts of the southern URG and transpressional reactivation of its border faults (Lutz & Cleintuar, 1999; Rotstein et al., 2005a,b).

This intra-Burdigalian unconformity is interpreted as reflecting rapid uplift of the VBFA (Roll, 1979; Laubscher, 1992; Dèzes et al., 2004), the crestal parts of which straddle the Potash Basin, located in narrowest parts of the southern URG (Hinsken et al., 2005). To what extent this arch continued to rise during middle Miocene to Pliocene times is difficult to assess (see above). From the VBFA arch, coarse conglomerates were shed southward into the area of the future Jura Mountains, the oldest remnants of which are dated as Serravallian (14 Ma) (Berger et al., 2005). These conglomerates contain in their lower parts mainly Jurassic carbonate components whereas upward Triassic Buntsandstein and Permian porphyrite components become dominant, reflecting progressive erosional unroofing of the VBFA (Kälin, 1997; Kemna & Becker-Haumann, 2003). Similarly, conglomerates were shed from the Black Forest southeastward onto the Franconian Platform, commencing in the late Aquitanian and intensifying during the Burdigalian to Serravallian (17-21 Ma; Müller et al., 2002). Furthermore, fission-track data indicate for the Vosges and Black Forest an accelerating cooling trend from early Miocene times onward that, provided it is not a modelling artefact, can be attributed to their uplift and erosional unroofing (Link et al., 2004; Timar-Geng et al., submitted)). Therefore, it is likely that after its rapid mid-Burdigalian uplift the VBFA continued to rise slowly in response to its isostatic adjustment to erosional unloading (removal of some 1000 m of Mesozoic sediments and deep truncation of basement rock across its culmination; Roll, 1979).

Although crustal extension across the URG continued during Miocene times, as evident by persisting fault-controlled subsidence of its northern parts, subsidence of its southern parts was apparently over-compensated by the uplift of the VBFA. This resulted in a net uplift of the southern parts of the URG from which at least some 1000-1500 m of syn-rift sediments were eroded during late Burdigalian to early Pliocene times. During the late Pliocene sedimentation resumed in this part of the URG and continues to the present. This suggests that by late Pliocene times the surface of the southern URG was located near the erosional base level and that its fault-controlled extensional subsidence began to outpace the slow uplift rates of the VBFA (Roll, 1979; Dèzes et al., 2004). Geodetic data indicate for the Black Forest a pattern of slow uplift of horst and slow subsidence of graben structures at rates rarely exceeding 0.25 mm/year (Müller et al., 2002).

Rapid uplift of the VBFA around 18 Ma was accompanied by volcanic activity within the URG and on its eastern flanks, spanning 18-7 Ma, that was essentially confined to the SW-NE striking crest of this arch (Kaiserstuhl 18-16 Ma; Hegau 15-7 Ma; Urach 16-11 Ma; Jung, 1999; Keller et al., 2002). This magmatic surge involved low-percentage partial melting of a predominantly asthenospheric source and magma segregation at depths of 100-70 km, at the base of the lithosphere and within its thermal boundary layer (with a possible contribution from deeper sources, J. Keller personal communication). As tomographic data indicate beneath the VBFA the occurrence of only very small amplitude low-velocity anomalies in the depth range of 67-107 km (Achauer & Masson, 2002), it must be assumed that potential minor thermal loads that may have contributed to the late Burdigalian to Tortonian uplift of this arch have since then largely decayed.

Uplift of the VBFA around 18 Ma clearly preceded the late Miocene to early Pliocene thin-skinned folding phase of the Jura Mountains (10-9 to 4 Ma), which around 4 Ma gave way to thick-skinned folding that continues to the present (Laubscher, 1986; Philippe et al., 1996; Becker, 2000; Ustaszewski et al., 2005). With the folding of the Jura Mountains, the drainage system on the southern flank of the VBFA was deflected westward, and from 4.2 Ma onward was joined by the palaeo-Aare river, that previously had debouched into the Danube and now flowed westward along the thrust front of the Jura Mountains into the Bresse Graben (Fig. 15; Sundgau Gravels). This illustrates that the southern parts of the URG were still located above the erosional base level with a watershed occurring in the Kaiserstuhl area. However, around 2.9 Ma, the paleo-Aare river was deflected into the URG, presumably in response to its tensional subsidence. During the late Pliocene-Quaternary, up to 240 m thick fluvial deposits accumulated in the southern parts of the URG in fault-controlled depressions (Roll, 1979), whilst the mid-Pliocene Sundgau gravels were folded along the Jura Mountains thrust front (Müller et al., 2002; Giamboni et al., 2004a,b). During the late Pliocene (2.9-1.65 Ma) the URG formed the sediment sink for the entire clastic load of the river Rhine, reflecting that sediment supply and generation of accommodation space by its extensional subsidence were in balance. However, at the Pliocene-Quaternary transition (1.65 Ma) sediment supply to the URG outpaced its subsidence rates, as evidenced by the first arrival of Alpine components in the LRG (Boenigk, 2002; Dèzes et al., 2004).

Intra-Burdigalian rapid uplift of the VBFA is attributed to lithospheric folding in response to the build-up of northwest-directed intraplate compressional stresses at crustal and lithospheric mantle levels, reflecting increasing collisional coupling between the Alpine orogenic wedge and its foreland during the early imbrication phases of the Alpine external crystalline massifs (Fig. 9e; Ziegler et al., 2002; Dèzes et al., 2004, 2005; Schmid et al, 2004). In the area of the VBFA, the build-up of compressional stresses at crustal levels is evidenced at the southern end of the URG by the End-Aquitanian (20.5 Ma) transpressional reactivation of pre-existing basement discontinuities (Laubscher, 2003), and by the latest Aquitanian-intra Burdigalian transpressional reactivation of border faults outlining the north-easterly striking southern parts of the URG (Rotstein et al., 2005a). The Burdigalian to Tortonian (18-7 Ma) volcanism, that is associated with the uplift of the VFBA by up to 2 km, can be attributed to decompressional partial melting of the lithospheric thermal boundary layer and the upper asthenosphere, with the latter being characterized by an above ambient temperature. This interpretation is compatible with an about 100 km thick lithosphere in the VBFA area (Babuska & Plomerova, 1992), the absence of a mantle plume rising up into its lithosphere (Achauer & Masson, 2002), and the observed generally low level of pre-Miocene volcanic activity (Jung, 1999; Keller et al., 2002).

The distribution of earthquakes shows that the URG and the VBFA are tectonically still active (fig. 3). In the area of the VBFA, focal mechanisms indicate for upper crustal levels a strike-slip to compressional stress regime whilst the lower crust is subjected to extension. Transpressional deformation of the upper crust can be attributed to collision-related stresses, which are transmitted from the Alps above a mid-crustal detachment level. By contrast, lower crustal extension may be partly related to buckling of the lithospheric mantle in response to stresses transmitted from the Alps through its mechanically strong upper parts, maintaining the elevation of the VBFA. Significantly, earthquakes occur almost down to the crust-mantle boundary, but are absent below it (Plenefisch & Bonjer, 1997; Deichmann et al, 2000). This suggests that deformation of the mechanically strong part of the lithospheric mantle controls the on-going deformation of the VBFA.

Uplift of the VBFA and crustal extension in the URG had severe repercussions on the development of the drainage system in the northwestern Alpine foreland, as evidences by its Pliocene switching from debouching via the Danube into the Black Sea, then via the

Bresse Graben into the Mediterranean, and finally via the Rhine Rift into the North Sea (Fig. 15).

The Vosges-Black Forest Arch: a non-cylindrical lithospheric fold?

At the Moho level, the VBFA forms a doubly plunging, SW-NE striking non-cylindrical anticline that extends from the northern parts of the Bresse Graben onto the Franconian Platform and towards the Bohemian Massif, the culmination of which coincides with its intersection with the URG. In the area of the VBFA, development of this structure is rather closely constrained as ±18 Ma. In the northern parts of the Bresse Graben, which interferes with the Burgundy transfer zone, an intra-Burdigalian (±18 Ma) erosional unconformity is observed that is overstepped by generally thin Serravallian and younger series; the hiatus across this unconformity decreases southward (Séranne, 1999; Sissingh, 2001, 2003a). Development of this unconformity can be attributed to lithospheric folding along the Burgundy transfer zone that coincides with an anticlinal uplift of the Moho discontinuity (Lefort & Agarval, 1996) leading directly into the VBFA. Moreover, in the NE-ward projection of the latter, uplift of the Eger volcano-tectonic zone commenced also around 18 Ma and was accompanied by the subsidence of its axial graben (Adamovic & Coubal, 1999). Thus, the uplift timing of the different segments of this Moho structure is consistent (Dèzes et al., 2004).

Laubscher (1992), who recognized the SW-NE trending Moho structure that is associated with the VBFA interpreted it as the Jura-phase flexural foreland bulge of the Alpine Orogen, and suggested that it developed between 14 Ma and at least 11 Ma. Moreover, he postulated that its culmination in the VBFA should be attributed to thermal weakening of the lithosphere beneath the southern parts of the URG, owing to mantle upwelling during its Oligocene-early Miocene main rifting phase, as modelled by Werner & Kahle (1980; see also Prodehl et al., 1995). However, in the face of tomographic data, which refute major thermal thinning of the lithosphere in the area of the VBFA (Achauer & Masson, 2002), the timing of its uplift (±18 Ma) that preceded folding of the Jura Mountains by at least 8 My, and the low level of volcanic activity prior to the mid-Burdigalian surge (Jung, 1999; Keller et al., 2002), we cannot support the hypothesis of Laubscher (1992) and prefer to attribute the development of this structure to lithospheric folding (Ziegler et al., 2002; Dèzes et al., 2005).

In this respect, development of non-cylindrical flexural-slip folds in thin-skinned fold-thrust belts offers an interesting analogue to our proposed lithospheric folding model. Field observations and numerical modelling suggest that syn-kinematic erosional breaching of the elastic beam controlling the geometry of an evolving flexural-slip fold causes, by strain concentration, the development of a non-cylindrical fold, the culmination of which is transected by a deep gorge (Fig. 16; Simpson, 2004a,b). This phenomenon has been observed in such active thrust belts as the external parts of the Zagros (Fig. 16b; Oberlander, 1985), the Northern Tien Shan facing the Junggar Basin (Fig 16c; NW China; Avouac et al., 1993), and the central Apennines (Alvarez, 1999). In the arcuate Jura Mountains, pre-existing transverse faults appear to have had a similar effect on fold geometries in so far as the axial culminations of arc-parallel striking anticlines frequently coincide with the interference point of compressionally reactivated transverse structures.

Applying this model to lithospheric scales, we hypothesize that extension-related weakening of lithosphere may have a similar effect on the geometry of a superimposed lithospheric fold, provided the graben strikes at an angle of 45^0 or less to the trajectories of the controlling compressional stress field, as in case of the URG and the VBFA. Significantly, the culmination of the VBFA coincides with the narrow Potash Basin segment of the URG and thus with a zone of extensional strain concentration (present ß = 1.14; Hinsken et al., 2005), in which faults and ductile shear zones cut through the entire crust (Fig. 17; Brun et al., 1992) and earthquakes occur even close to crust-mantle boundary (Fig. 18; Plenefisch & Bonjer, 1997). In this segment, Paleogene-early Miocene syn-rift flank uplift was presumably considerably greater than in the northward and southward adjacent parts of the URG (Roll, 1979). This may have actually contributed to the localization of the VBFA.

In the face of pre-rift crustal thicknesses of about 30 km, the crust contributed significantly to the strength of the lithosphere in the area of the VBFA (Fig. 19; Ziegler & Cloetingh, 2004). However, the faults of the URG weakened the crust substantially whilst stretching and related thermal perturbation of the lithosphere presumably reduced the strength of the lithospheric mantle. Correspondingly the strength of the lithosphere of the URG was considerably lower as compared to flanking areas. During lithospheric folding this had apparently a similar compressional strain concentration effect as erosional breaching of the controlling elastic beam has on an evolving thin-skinned flexural-slip fold. It is planned to test the viability of this concept by interactive analogue and numerical modelling and to report on it at a later date.

Interestingly, the Permo-Carboniferous fault systems that underlie the Burgundy Transfer Zone, and which were reactivated during the development of the URG and Bresse Graben, trend normal to the compressional stress field that controlled the development of the VBFA. In view of the considerably lower relief of the lithospheric fold in the area of the Burgundy Transfer Zone, these faults apparently did not have the same effect as those of the URG. Moreover, Cenozoic transtensional faulting in this area accounts for a very low stretching factor of about ß = 1.02 (Hinsken et al., 2005).

Discussion and conclusions

All major Variscan arches in the Alpine foreland are more or less closely associated with graben structures and fault systems that form part of ECRIS (Fig. 1). This pertains also to the Armorican Massif, the Variscan shear systems of which were reactivated in response to a finite clock-wise westward rotation of the Paris Basin block during the subsidence of the Rhine and Rhône rifts that persists to the present (Fig. 21; Dèzes et al., 2004; Cloetingh et al., 2005; Tesauro et al., 2005).

Extensional strain and flank uplift

Extensional strain across ECRIS, as derived from upper crustal faulting, amounts to about 2 km for the Bresse Graben, 3-4 km for the grabens of the Massif Central (Bergerat et al., 1990), and is in the range of 5-7 km for the URG (Brun et al., 1992; Hinsken et al., 2005). Across the LRG-RVG, upper crustal extension diminishes from 4-5 km in its southern parts to zero near the Dutch North Sea coast (Geluk et al., 1994).

Similar to other rifts, and in the face of a lithospheric necking level within the strong upper part of the lithospheric mantle (van Wees & Cloetingh, 1996), the magnitude of rift shoulder uplift depends in ECRIS largely on the graben width and the number of faults over which the extensional strain is partitioned (Kusznir & Ziegler, 1992). This is particularly evident by the level to which Mesozoic sediments have been eroded on the flanks of the URG, the width of which varies between about 30 km in its northern, 60 km in its central, and between 35 and 63 km in its southern parts, whilst the extensional strain remains constant (Fig. 20). For instance, on the Odenwald Block, that is associated with a major fault marking the eastern margin of the narrow axial northern graben segment (ß = 1.15), Variscan basement is exposed at elevations of up to 692 m (Meier & Eisbacher, 1991), whereas in the much wider central graben segment (ß = 1.11), Middle and Late Triassic series are still preserved on the rift shoulders at elevations not exceeding 500 m, with the alluvial plain of the URG being located at about 100 m.

This suggests that syn-rift flexural uplift of rift flanks played only a subsidiary role in the development of the different arches that are associated with ECRIS. This is also compatible with the fact that in the area of the Rhenish Massif the flanks of the HG and LRG were overstepped during Oligocene-early Miocene times by partly marine sediments (Sissingh, 2003b).

In view of the above, plume-related thermal perturbation of the lithosphere, as well as its deformation in response to the build-up of collision-related compressional intraplate stresses, appear to be the primary mechanisms that controlled the uplift of basement arches in the ECRIS area.

Thermal thinning of lithospheric mantle versus intraplate compression

The level and timing of magmatic activity varies considerably on the different basement arches of the ECRIS area. Plume-related thermal thinning of the lithospheric mantle is indicated for the Massif Central and the Rhenish Massif (Figs. 5 & 6) and may, combined with thermal inflation of the remnant lithosphere and the buoyant load of mantle plumes (Garcia-Castellanos et al., 2000), underlay their progressive uplift that commenced at the Oligocene-Miocene transition. Similarly, thinning of the lithospheric mantle beneath the northern parts of the Bohemian Massif (Fig. 7) may be attributed to the Oligocene main pulse of volcanic activity in the Eger volcano-tectonic zone, uplift of which commenced, however, only after magmatic activity had decreased during the early Miocene (Babuska & Plomerova, 2001; Ulrych et al., 1999). Therefore, thermal mechanisms are unlikely to have contributed to the mid-Miocene and younger uplift of the Bohemian Massif, nor to the Burdigalian uplift of the VBFA and the late Pliocene-Quaternary uplift of the non-volcanic Armorican Massif, for all of which mantle tomography shows that they are not associate with active mantle plumes penetrating the lithosphere (Babuska et al., 2005; Achauer & Masson, 2002; Judenherc et al., 2002).

Intra-Burdigalian uplift of the VBFA, that extends SW-ward into the northern parts of the Massif Central and to the NE towards the Bohemian Massif (Figs. 4 & 9d), was controlled by lithospheric folding that presumably contributed by decompression of the asthenosphere and lithospheric thermal boundary layer to a rather short-lived magmatic pulse. The underlying build-up of intraplate compressional stresses at crustal and lithospheric mantle levels reflects increasing collisional coupling of the Alpine Orogen with its foreland during the initial imbrication and uplift phases of the Alpine external crystalline massifs (Dèzes et al., 2004; Schmid et al., 2004).

Similarly, progressive uplift of the northern parts of Bohemian Massif, beginning in Burdigalian times, can be related to lithospheric buckling in response to the build-up of intraplate compressional stresses at lithospheric levels (Ziegler et al., 2002; Dèzes et al., 2004), and may have contributed by decompression of the lower lithosphere and asthenosphere to the late Miocene-Pliocene phase of volcanic activity in the Eger zone (Ulrych et al., 1999). On the other hand, late Miocene to Quaternary differential uplift of the marginal highs of the Bohemian Massif, such as the Erzgebirge, Lusatian, Sudetic, Thuringian-Bohemian and Bavarian Forest and Moravian blocks, involved reactivation of pre-existing crustal discontinuity in response to the build-up of compressional stresses at crustal levels. The late Burdigalian and younger compressional deformation of the Bohemian Massif can be attributed to increased collisional coupling of the East-Alpine Orogen with its foreland, owing to disruption of the south-dipping East-Alpine subduction slab beginning around 20 Ma, subsequent eastward escape of the ALCAPA block, and the onset of northward subduction of Adriatic lithosphere beneath the European foreland (Schmid et al., 2004).

Present-day stress field

The present-day NW-directed compressional stress field of the Alpine foreland (Müller et al., 1997) reflects a combination of Alpine collisional and Atlantic ridge-push forces (Gölke et al., 1996). This stress field came into evidence during the early Miocene and intensified further during the Pliocene between 3 and 2.5 Ma (Dèzes et al., 2004). This is evidenced by a subsidence acceleration of the nearly orthogonally extending LRG-RVG (Geluk et al., 1994; Michon et al., 2003) and accelerated stress-induced down flexing of the North Sea-North German Basin (van Wees & Cloetingh, 1996). Under the present stress field the Armorican Massif, the Massif Central, the western parts of the Rhenish Massif and the northern parts of the Bohemian Massif continue to rise at rates ranging up to 1.75 mm/y whilst the Vosges-Black Forest Arch is relatively stable. Earthquake focal mechanisms indicate transtensional and transpressional deformation of the Massif Central and the Armorican Massif (Nicolas et al., 1990; Delouis et al., 1993), transpressional to compressional reactivation of Variscan thrusts in the Ardennes and orthogonal and oblique extension for the LRG-RVG and URG, respectively (Ahorner 1983; Hinzen, 2003; Plenefisch & Bonjer, 1997). Under this scenario it is likely that tangential compressional stresses contribute to the continued uplift of the Massif Central, the Armorican Massif, the western Rhenish Massif and the Bohemian Massif. Geodetic data indicate that westward rotational escape of the Paris Basin continues at rates of some 0.7 mm/y (Fig. 21; Cloetingh et al., 2005; Tesauro et al, 2005). This motion, which was initiated already during the late Eocene, was compensated by inversion of Mesozoic extensional basins in the Channel-Western Approaches and Celtic Sea area, by up warping of the Weald-Artois axis, resulting in intermittent opening and closure of the Dover Strait, and by deformation of the Armorican Massif (Figs. 9 & 11; Ziegler, 1987; 1990; Butler & Pullan, 1990; van Vliet-Lanoë et al., 1998a,b; Mansy et al., 2003; Gibbard & Lewin, 2003). However, as the main inversion phases of these basins are not fully synchronous with the main subsidence phases of ECRIS, far-field Alpine collision-related stresses presumably played a contributing role (Ziegler, 1990; Dèzes et al., 2004).

Lower crustal flow and isostasy

Accelerated Pliocene-Quaternary uplift of the Variscan Massifs and related development of fluvial terrace systems has been attributed by Westaway (2002b) to isostatic movements involving lower crustal flow along lateral pressure gradients that developed in response to erosional unroofing of these massifs, sediment accumulation in adjacent depocentres, waxing and waning ice loads, and sea-level fluctuations.

Theoretical considerations suggest that in the presence of wet, felsic rheologies and crustal thicknesses of 30 km, the lower crust may under certain circumstances deform by ductile flow, whereas in the presence of dry rheologies no flow occurs. In nature, crustal flow plays a significant role only during the development of core complexes. These result from high strain extension of 45-60 km thick, orogenically destabilized crust (elevated heat flow) that gives rise to the development of large lateral pressure gradients (Bertotti et al., 2000; Ziegler & Cloetingh, 2004). At normal crustal thicknesses (30-35 km) and in the absence of major extension, lower crustal flow may only occur in the presence of partial melts. In this respect it should be noted that in the ECRIS area pre-rift crustal thicknesses were in the range of 30-35 km, that during the development of its graben systems the lowermost crust deformed in the domain of the brittle/ductile transition zone, and that there is no evidence for lower crustal thickening beneath its grabens, indicative for lower crustal flow (Fig. 17). Present crustal thicknesses of the Variscan Massifs exceed 30 km only in the Armorican Massif and the southern parts of the Bohemian Massif (Fig. 4). Furthermore, preservation of the Variscan orogenic crustal fabric, as imaged by deep reflection-seismic lines (e.g. Meissner & Bortfeld, 1990; see also Ziegler et al., 2004), raises doubts on on-going crustal flow.

Furthermore, it must be kept in mind that the primary isostatic compensation level for lateral changes in crustal thickness, density and related topography, lithospheric thickness and surface loads (e.g. ice-loading and –unloading, water and sediment loading), resides in the sub-lithospheric upper mantle, the viscosity and density of which is essentially temperature-dependent (Bott, 1982; Artyushkov, 1983; Watts, 2001; Turcotte & Schubert, 2002; Pelltier, 1989, 2004).

Therefore, we postulate that the crustal and lithospheric scale deformations which underlie the observed accelerated Pliocene-Quaternary uplift of the Variscan Massifs and the subsidence of the North Sea-North German Basin, are controlled by a combination of intraplate compressional stresses, reflecting increasing collisional coupling of the Alpine orogen with its foreland, and plume-related thermal thinning of the lithospheric mantle, with glacial isostatic movements playing an overprinting role in areas that were intermittently covered by major ice sheets.

Drainage systems

Overall, Neogene and Quaternary uplift of the Variscan Massifs in the Alpine foreland, combined with the evolution of ECRIS, exerted strong controls on the development of the modern drainage systems of the West- and Central-European platform (Salpeteur et al., 2005). Salient examples are i) in the Basel area, the Pliocene changes in the drainage divide between rivers flowing into the Black Sea, the Mediterranean and the North Sea that were controlled by folding of the Jura Mountains, slow uplift of the VBFA and subsidence of the URG, ii) on the Bohemian Massif, the mid-Miocene northward shift of the drainage divide between the Paratethys and the North-German Basin and its late Pliocene southward shift were controlled by lithospheric buckling and by transpressional reactivation of the marginal blocks of this massif, respectively, iii) the middle Miocene development of the rivers Meuse and Maas was controlled by uplift of the Vosges-Black Forest Arch, involving lithospheric folding, whilst their incision into the gradually rising Rhenish-Ardennes Massif, commencing in the late Miocene, was mainly governed by thermal attenuation of its lithospheric mantle, and iv) the Miocene and younger development of the Paris Basin drainage system was controlled by thermal uplift of the Massif Central and Rhenish Massif, folding of the lithosphere in the area of the Burgundy Transfer Zone and transpressional lithospheric buckling of the Armorican Massif.

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Text figures captions

Fig. 1 Location map of ECRIS in the Alpine and Pyrenean foreland, showing Cenozoic fault systems (black lines), rift-related sedimentary basins (light grey), Basement arches (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. Cartographic representation were made using GMT (Wessel & Smith, 1991)

Fig. 2 DEM of ECRIS area with superimposed ECRIS fault systems

Fig. 3 DEM of ECRIS area with superimposed ECRIS fault systems and distribution of earthquake epicentres (based on USGS NOAA NEIC Earthquake Data Base "Hypocenter Data Files")

Fig. 4 Depth map of Moho discontinuity, contour interval 2 km (after Ziegler & Dèzes 2005) with superimposed ECRIS fault systems (thin lines) and volcanic centres (black fields).

Fig. 5 Tentative lithosphere thickness map (after Babuska & Plomerova; 1992) with superimposed ECRIS fault system and trace of lithospheric transects given in Figs. 6 and 7.

Fig. 6 Lithospheric transects across a) the Central Alps and Rhenish Massif and b) the Western Alps and Massif Central (Dèzes et al., 2004)

Fig. 7 Lithospheric profile through NW part of Bohemian Massif (Babuska & Plomerova, 2001)

Fig. 8 Tomographic P-wave velocity cross-sections, showing mantle structure in the ECRIS area. Note prominent low velocity anomalies above 410 km discontinuity (Goes et al., 1999).

Fig. 9 Palaeotectonic sketch maps of ECRIS area for a) late Paleocene, b) mid Eocene c) late Eocene, d) late Oligocene, e) early-middle Miocene and f) Plio-Pleistocene. Legend: dark grey: orogens; light grey: areas of non-deposition; white: sedimentary basins; stippled: oceanic basins; stars: volcanism; arrows: maximum horizontal compressional stress direction; thick dashed line: axis of lithospheric fold (Dèzes et al., 2004).

Fig. 10 Palaeogeographic setting of Massif Central area during a) early Oligocene and b) middle Miocene. (Sissingh, 2001).

Fig. 11 Late Miocene and Pliocene palaeogeographic setting of Armorican Massif. Note repeated opening and closure of the Dover Strait (Van Vliet-Lanoë, 1998a).

Fig. 12 Palaeogeographic setting of the Rhenish Massif during a) the late Oligocene and b) middle Miocene. (Sissingh, 2003b).

Fig. 13 Palaeogeographic setting of the Bohemian Massif during a) the late Oligocene-early Miocene, b) the middle Miocene, c) late Miocene and d) Pliocene, compared to present-day drainage system (Malkovski, 1979).

Fig. 14 High-resolution reflection-seismic line recorded of the river Rhine showing mid-Burdigalian unconformity (courtesy INTERREG III MoNit Project) and onlap pattern of overstepping series. Note young extensional faults cutting Pliocene and Quaternary series.

Fig. 15 Pliocene evolution of the drainage pattern at the southern end of the Upper Rhine Graben (Giamboni et al., 2003a,b)

Fig. 16 a) Schematic diagram showing transverse river intersecting the culmination of a doubly plunging fold structure (Simpson, 2004).
b) Map view of calculated topography development of a elastic-plastic plate under unidirectional compression and progressive shortening, subjected to river incision (contours 100 m) (Simpson, 2004).

Fig. 17 Crustal-scale cross-section through southern Upper Rhine Graben (Brun et al., 1992).

Fig. 18 Depth distribution of earthquake hypocentres in the southern Upper Rhine Graben (Plenefisch & Bonjer, 1997).

Fig. 19 Depth-dependent rheological models for dry and wet, unextended cratonic lithosphere and stretched, thermally attenuated lithosphere, assuming a quartz/diorite/olivine mineralogy; a) unextended, cratonic lithosphere with crustal and lithospheric mantle thicknesses of 30 and 70 km, respectively; b) extended, thermally destabilized cratonic lithosphere with crustal and lithospheric mantle thicknesses of 20 and 38 km, respectively (Ziegler & Cloetingh, 2004).

Fig. 20 DEM and Geological map of the Upper Rhine Graben. Red line on DEM: 100m. elevation, black line: 500m. elevation, blue line:1000m. elevation. (Bundesanstalt für Geowissenschaften und Rohstoffe 1976).

Fig. 21 Calculated velocities of crustal motions in Western and Central Europe based on GPS observations. Black arrows: motion of permament GPS stations. White arrows: motion of virtual points located on average 50 km close to block borders delimited by the Rhine and Rhône rifts and the Armorican shear zone (Tesauro et al., 2005)

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