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Evolution of the European Cenozoic Rift System: interaction of the Alpine and Pyrenean orogens with their foreland lithosphere

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Abstract

The evolution of the European Cenozoic Rift System (ECRIS) and the Alpine orogen is discussed on the base of a set of palaeotectonic maps and two retro-deformed lithospheric transects which extend across the Western and Central Alps and the Massif Central and the Rhenish Massif, respectively.

During the Paleocene, compressional stresses exerted on continental Europe by the evolving Alps and Pyrenees caused lithospheric buckling and basin inversion up to 1700 km to the north of the Alpine and Pyrenean deformation fronts. This deformation was accompanied by the injection of melilite dykes, reflecting a plume-related increase in the temperature of the asthenosphere beneath the European foreland. At the Paleocene–Eocene transition, compressional stresses relaxed in the Alpine foreland, whereas collisional interaction of the Pyrenees with their foreland persisted. In the Alps, major Eocene north-directed lithospheric shortening was followed by mid-Eocene slab- and thrust-loaded subsidence of the Dauphinois and Helvetic shelves. During the late Eocene, north-directed compressional intraplate stresses originating in the Alpine and Pyrenean collision zones built up and activated ECRIS.

At the Eocene–Oligocene transition, the subducted Central Alpine slab was detached, whereas the West-Alpine slab remained attached to the lithosphere. Subsequently, the Alpine orogenic wedge converged northwestward with its foreland. The Oligocene main rifting phase of ECRIS was controlled by north-directed compressional stresses originating in the Pyrenean and Alpine collision zones.

Following early Miocene termination of crustal shortening in the Pyrenees and opening of the oceanic Provençal Basin, the evolution of ECRIS was exclusively controlled by west- and northwest-directed compressional stresses emanating from the Alps during imbrication of their external massifs. Whereas the grabens of the Massif Central and the Rhône Valley became inactive during the early Miocene, the Rhine Rift System remained active until the present. Lithospheric folding controlled mid-Miocene and Pliocene uplift of the Vosges-Black Forest Arch. Progressive uplift of the Rhenish Massif and Massif Central is mainly attributed to plume-related thermal thinning of the mantle-lithosphere.

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ECRIS evolved by passive rifting in response to the build-up of Pyrenean and Alpine collision-related compressional intraplate stresses. Mantle-plume-type upwelling of the asthenosphere caused thermal weakening of the foreland lithosphere, rendering it prone to deformation.

Keywords: Crust; Mantle; Lithosphere; Orogeny; Subduction; Rifting; Volcanism

1. Introduction

The European Cenozoic Rift System (ECRIS) is located in the foreland of the Alps and extends over a distance of some 1100 km from the North Sea coast to the Mediterranean (Fig. 1). ECRIS consists of the Rhine and the Massif Central-Rhône Valley rift systems, which are linked by the Burgundy and the

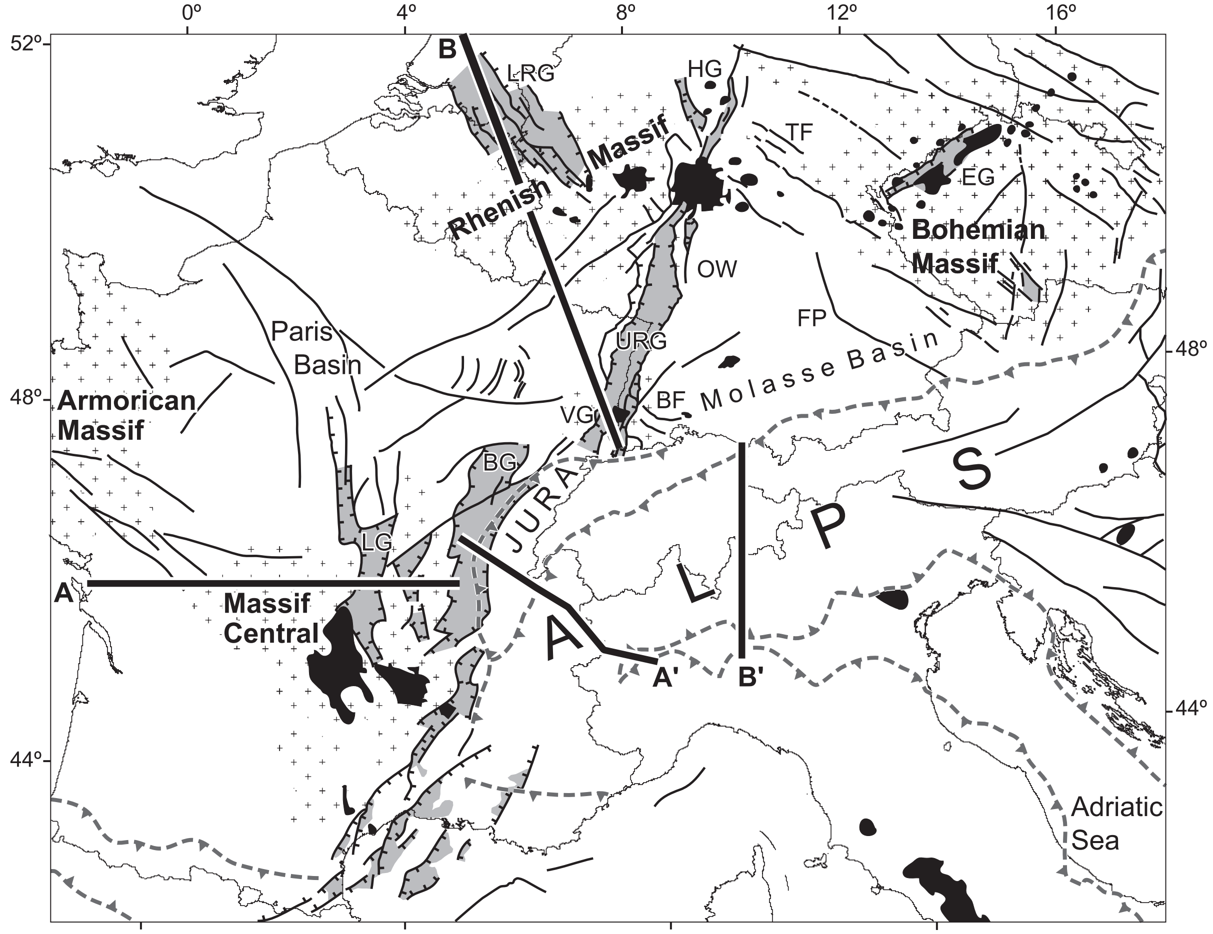


Fig. 1. Location map of ECRIS in the Alpine and Pyrenean foreland, showing Cenozoic fault systems (black lines), rift-related sedimentary basins (light grey), Variscan massifs (cross pattern) and volcanic fields (black). Solid barbed line: Variscan deformation front; stippled barbed line: Alpine deformation 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. Thick lines A–A′ and B–B′ : location of transects given in Figs. 3 and 4.

eastern Paris Basin transfer zones, and includes also the shallow Eger Graben of the Bohemian Massif. The Rhine Rift System, forming the northern part of ECRIS, includes the Upper Rhine, Roer Valley and Hessian grabens. The southern part of ECRIS consists of the grabens of the Massif Central (Limagne, Roanne, Forez), the Bresse Graben and the grabens of the lower Rhône Valley (e.g., Valence, Alès, Manosque, Camargue) and their prolongation into the Western Mediterranean (Ziegler, 1994; Prodehl et al., 1995; Séranne, 1999; Merle and Michon, 2001; Roca, 2001; Sissingh, 2003a). Development of ECRIS began during the late Eocene, mainly by reactivation of late Variscan, Permo–Carboniferous and Mesozoic fracture systems (Schumacher, 2002). By this time, the Alpine deformation front was located over 200 km to the south of the Upper Rhine Graben and some 350 km to the southeast of the Bresse Graben. During the Pliocene, frontal thrusts of the Jura Mountains encroached on the southern margin of the Upper Rhine Graben and the eastern margin of the Bresse Graben. The Manosque and Alès grabens, located in the foreland of the West-Alpine Digne thrust, were partly inverted during the Miocene (Séranne, 1999). The areas of the Roer Valley and Upper Rhine grabens, as well as the Burgundy transfer zone, are seismotectonically active (Giglia et al., 1996; Bonjer, 1997; Hinzen, 2003) and thus correspond to zones of increased seismic hazard (Giardini et al., 2003).

In this contribution, we discuss the development of ECRIS in the context of the Cenozoic evolution of the Alpine and Pyrenean orogens on the basis of a set of palaeotectonic maps and the retro-deformation of two lithospheric transects, which extend across the Western and Central Alps and the Massif Central and the Rhenish Massif, respectively.

2. Construction of lithospheric transects across the Alps and ECRIS

In order to visualize the position of the Rhine and Massif Central-Rhône Valley rift systems relative to the evolving Alpine orogen, we constructed two lithospheric transects. One of these extends from the Po Valley across the Central Alps and the Rhenish Massif to Amsterdam, whereas the other extends across the Western Alps and the Massif Central to the margin of the Aquitaine Basin (Fig. 2). These were step-wise retro-deformed to the late Paleocene (Figs. 3 and 4). Regarding the Alps, the transects and their retro-deformation are based on the interpretation of the NFP-20-East profile (Pfiffner et al., 1997) and the ECORS-CROP profile (Damotte et al., 1990) by Schmid et al. (1996) and Schmid and Kissling (2000), respectively, and on concepts developed by Stampfli et al. (1998, 2002). For the Alps and their foreland, crustal thicknesses as compiled by Dèzes and Ziegler (2002) were used (Fig. 2). The thickness of the thermal lithosphere, as shown for the Upper Rhine Graben area and the Rhenish Massif, is based on Babushka and Plomerova (1992). For the Dutch part of the eastern transect, the lithosphere thickness is poorly constrained. For the Massif Central, lithosphere thicknesses as given by Sobolev et al. (1997) were used. The lithospheric configuration shown for the Western and Central Alps is compatible with the results of high-resolution seismic tomography studies by Lippitsch (2003), and with the results of subsidence modelling, indicating end-Mesozoic thermal lithosphere thicknesses of 100–120 km for Variscan crustal domains (Ziegler et al., 2004). Moreover, we took into consideration that the upper asthenosphere beneath Western and Central Europe is characterized by anomalously low P- and S-wave velocities (Zielhuis and Nolet, 1994; Goes et al., 2000a,b). We also used the results of seismic tomography, imaging low velocity structures rising up from the deep mantle, which apparently feed smaller upper-mantle plumes, such as those welling up beneath the Rhenish Massif and the Massif Central (Granet et al., 1995; Goes et al., 1999; Ritter et al., 2001).

In the area of our transects, the crust is considered as being dominated by felsic rocks. Seismic velocity analyses and the study of xenoliths brought to surface by Cenozoic extrusives indicate that in the Alpine foreland basic rocks occur only in the basal parts of the crust (Downes, 1993; Mengel, 1992; Wittenberg et al., 2000), which in some areas are characterized by densely packed anastomosing reflectors paralleling the crust-mantle boundary. As these reflectors are probably related to mantle-derived basic sills, the laminated lowermost crust, which ranges in thickness from 5 to 10 km, is probably characterized by interlayered mafic rocks and granulite-facies para-

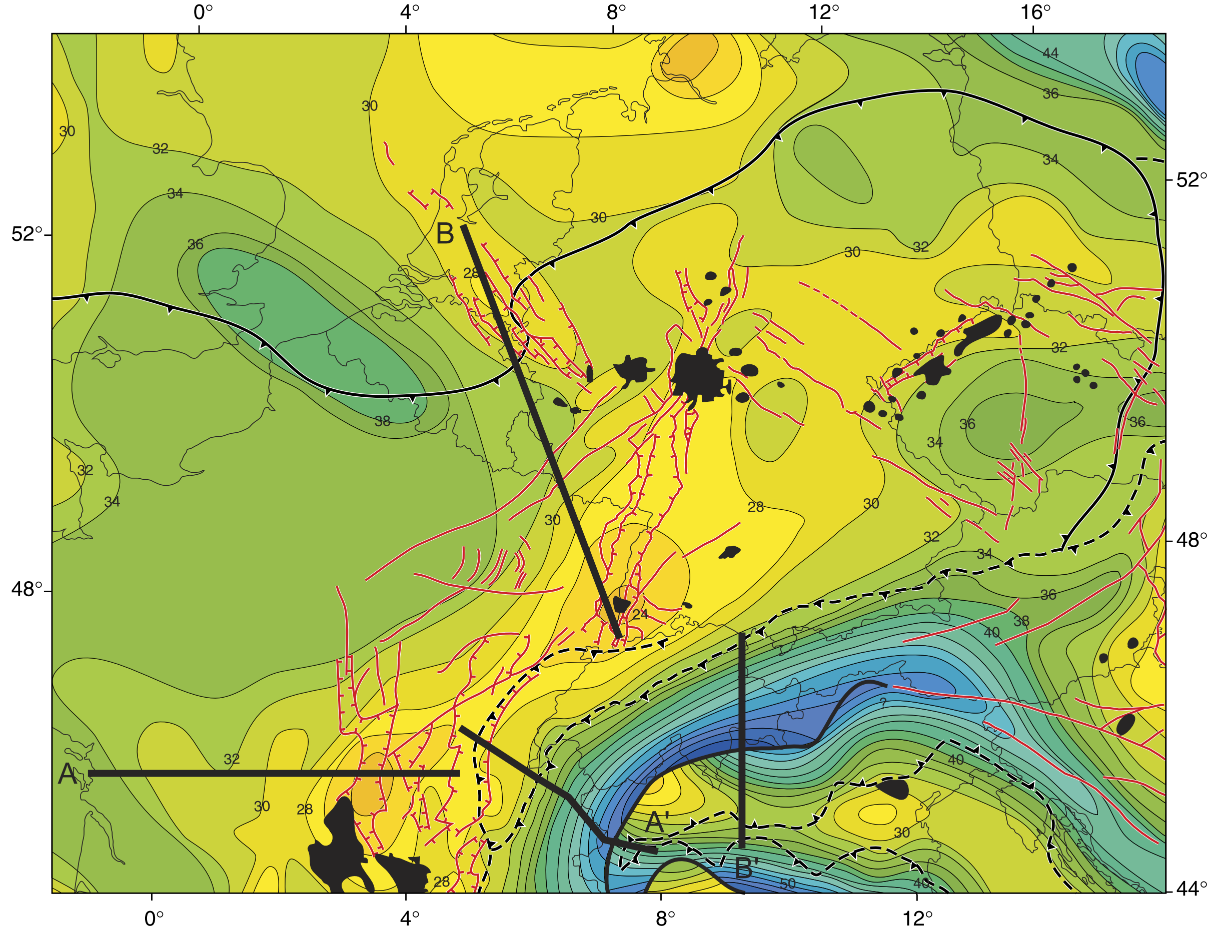


Fig. 2. Depth map of Moho discontinuity, contour interval 2 km (after Dèzes and Ziegler, 2002) with superimposed ECRIS fault systems (thin lines) and volcanic centres (black fields). Thick lines A–A′ and B–B′ location of transects given in Figs. 3 and 4.

gneisses, as seen in the Ivrea Zone (Schmid, 1993; Ziegler et al., 2004). The Permian and younger sedimentary cover of the crust can attain thicknesses of up to 4 km (Ziegler, 1990). The thickness of essentially undeformed Palaeozoic sediments covering the crystalline crust of the Netherlands is uncertain but may reach up to 5 km. In the area of the Rhenish Massif, in which deformed Devonian and Carboniferous sediments attain a thickness of up to 15 km, the upper crust is partly involved in the Rheno-Hercynian thrust belt (Oncken et al., 2000; Ziegler et al., 2004).

The consolidation age of the continental crust underlying the Netherlands and the Rhenish Massif ranges between 560 and 400 Ma (Cadomian and Caledonian). South of the Rhenish Massif, the crustal consolidation age of the Variscan internides ranges between 310 and 280 Ma (Variscan). The Western and Central Alps involve Variscan basement that was deformed during a time span of 80–0 Ma.

3. Retro-deformation of lithospheric transects

The present-day transects were retro-deformed into a sequence of conceptual palinspastic transects, presented in Figs. 3 and 4. These depict the stepwise evolution of the Alpine orogen and its foreland from 60 Ma onwards (time scale of Berggren et al., 1995). In these transects, we distinguish between three types of mantle-lithosphere that re-equilibrated with the asthenosphere at different times, namely: (1)

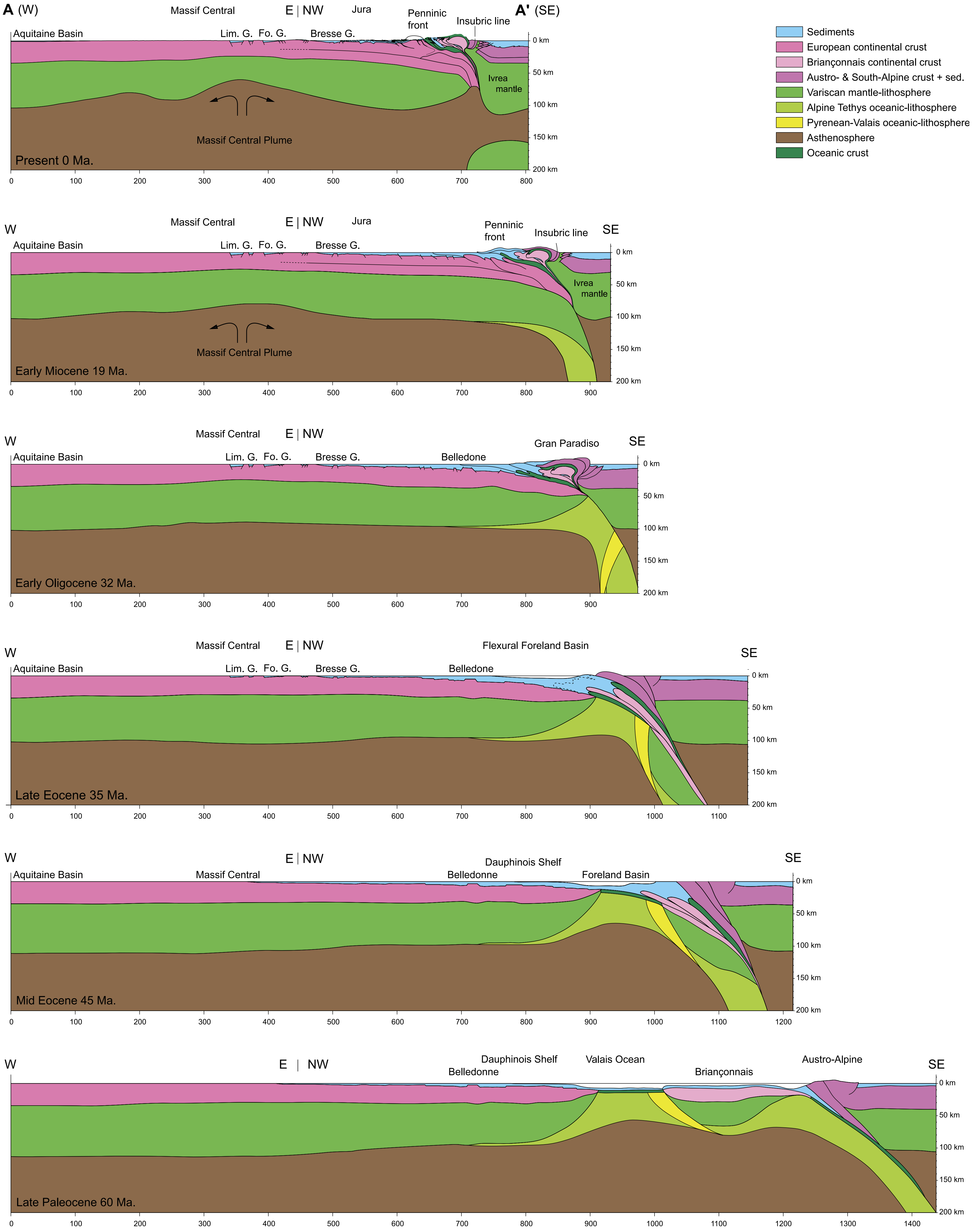


Fig. 3. Lithospheric transect A–A′ across the Western Alps and Massif Central, step-wise restored to 60 Ma, showing conceptual Cenozoic evolution of the lithosphere (for location see Fig. 2). Please refer to the web version of the paper to view this figure in colour

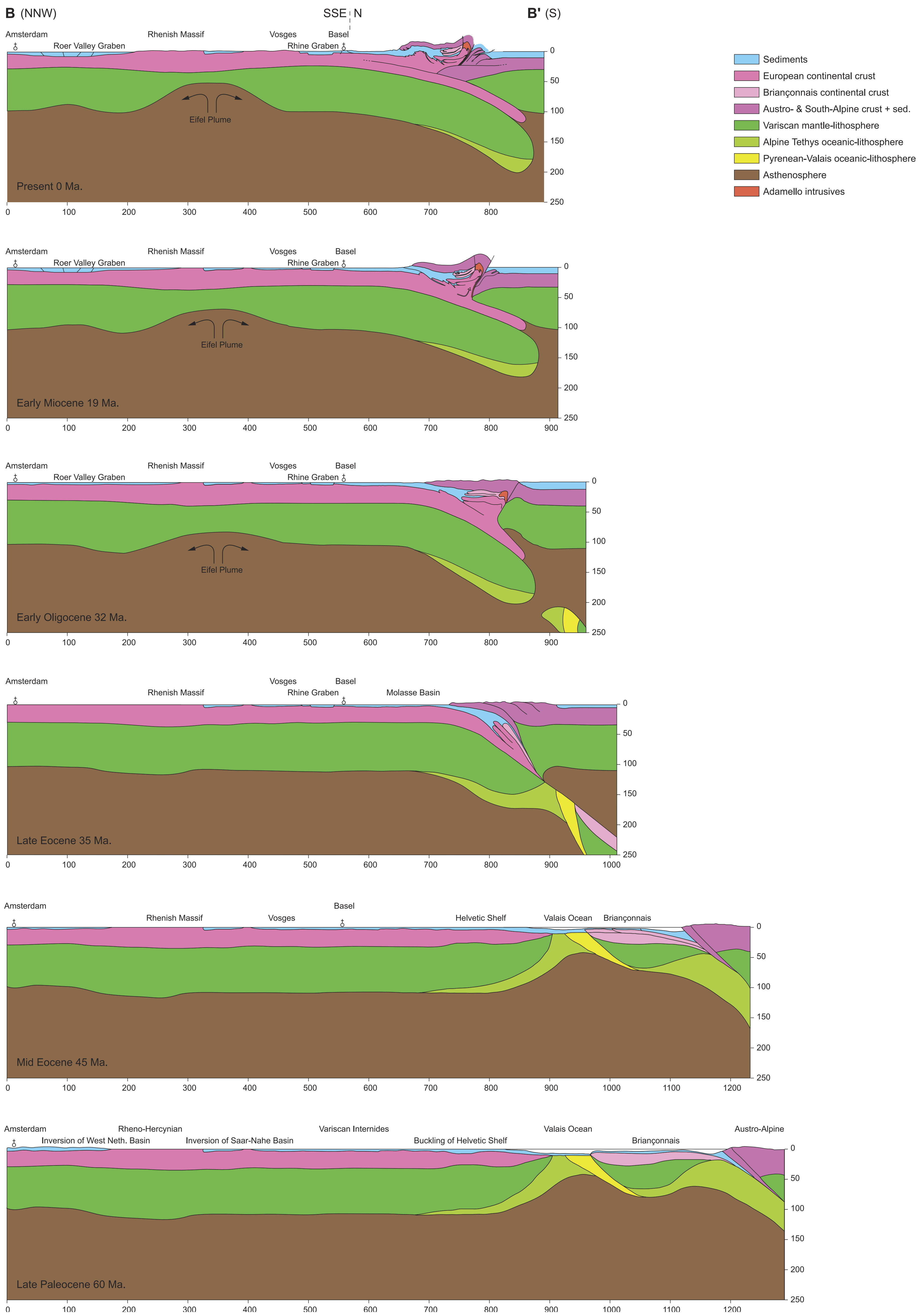


Fig. 4. Lithospheric transect B–B′ across the Central Alps and Rhenish Massif, step-wise restored to 60 Ma, showing conceptual Cenozoic evolution of the lithosphere (for location see Fig. 2). Please refer to the web version of the paper to view this figure in colour.

"Variscan" mantle-lithosphere, post-dating the Variscan orogeny and the Permo–Carboniferous tectono-magmatic event (310–280 Ma; Ziegler et al., 2004); (2) "Alpine Tethys" mantle-lithosphere, formed during and after the Mid-Jurassic opening of the Piemont–Ligurian–Maghrebian Ocean (Alpine Tethys, 160 Ma); and (3) "Pyrenean–Valais Ocean" mantle-lithosphere, formed during and after the early Cretaceous separation of the Iberia–Briançonnais terrane from the Armorican-Provençal margin of France (110–90 Ma). According to Stampfli et al. (1998, 2002), Late Cretaceous opening of the North Atlantic was accompanied by the opening of the Bay of Biscay–Pyrenean Ocean, counter-clockwise rotation of Iberia, and lateral insertion of the Briançonnais terrane into the Alpine Tethys, thus separating its northern Valais Ocean parts from its southern Piemont–Ligurian Ocean parts. Others, such as Frisch (1979) and Schmid et al. (2004), envisage an extension of the Pyrenean Ocean into the Alpine Tethys, which led to the opening of the Valais Ocean in the area of the future Western Alps. In any case, the Valais Ocean was partly floored by Alpine Tethys lithosphere and partly by Pyrenean–Valais Ocean lithosphere.

Calculated subsidence curves indicate that by end-Cretaceous times the lithosphere of the internal parts of the Variscan orogen had equilibrated with the asthenosphere at depths between 100 and 120 km (Ziegler et al., 2004). In the eastern transect and for areas south of the Rhenish Massif, we therefore assumed a Paleocene thermal lithosphere thickness of 100 km, tapering to 80 km in the Valais Ocean. For the Rhenish Massif, a Paleocene lithosphere thickness of 120 km was adopted, since this massif remained a high during late Permian and Mesozoic times (except for the Triassic–early Jurassic Trier Embayment; Ziegler, 1990), and because it is underlain by the Cadomian–Caledonian lithosphere of the Variscan foreland. For the West-Netherlands Basin, which is also superimposed on Cadomian–Caledonian crust, a lithosphere thickness of about 100 km was assumed since it was affected by the Permo–Carboniferous tectono-magmatic event, as well as by Mesozoic rifting. In the western transect, a Paleocene lithosphere thickness of 110 km was assumed for the Massif Central, since it was only marginally transgressed by Mesozoic series. For the Dauphinois Shelf, however, a lithosphere thickness of 100 km, tapering to 80 km at the transition to the Valais Ocean was adopted.

Below we discuss the main steps in the Cenozoic evolution of the Alps, Pyrenees and ECRIS, as summarized in the retro-deformed transects (Figs. 3 and 4) and the supporting palaeotectonic sketch maps (Fig. 5).

4. Step 1: Paleocene (65–54.8 Ma)

Convergence rates between Africa and Europe decreased sharply from up to 20 mm/year during late Cretaceous times to practically zero during the late Maastrichtian and Paleocene (67–55 Ma) (Rosenbaum et al., 2002). Strong mechanical coupling of African and European plates across the Alpine–Mediterranean orogen probably underlies this decrease in convergence rates. During the late Paleocene (60.9–54.8 Ma), a pulse of intense intraplate compression affected not only much of Western and Central Europe but also the East-European Craton and North Africa (Ziegler, 1990; Ziegler et al., 1995, 2001; Nikishin et al., 1999; Stampfli et al., 2001).

In both transects, the southward subduction of the Piemont Ocean (Alpine Tethys) beneath the Austro–Alpine orogenic wedge was apparently completed by late Paleocene times (Figs. 3 and 4). This ocean was originally located between the Apulian and European margins in the case of the Eastern Alps, but between Apulia and the Briançonnais Terrane in the case of the Central and Western Alps (Fig. 5A; Stampfli et al., 1998, 2002; Schmid et al., 2004).

Closure of the Piemont Ocean resulted in the East-Alpine and North-Carpathian domains in the collision of the Austro–Alpine orogenic wedge with the European foreland and the build-up of intraplate compressional stresses in the latter. By reactivation of pre-existing crustal discontinuities, these stresses caused upthrusting of the basement block forming the Bohemian Massif and inversion of the Polish Trough (Fig. 5A). In the Central- and West-Alpine domain, however, the Austro–Alpine accretionary wedge collided with the continental Briançonnais terrane, which still was attached to the Iberian microplate (Stampfli et al., 1998, 2002). Possibly due to subduction resistance of the continental Briançonnais terrane compressional stresses were transmitted through it and the still open Valais Ocean into the northward adjacent Helvetic Shelf and the domain of the future Rhine Rift System. These stresses induced gentle broad-scale warping and uplift of the European lithosphere, causing the development of a regional erosional unconformity, and, by reactivation of pre-existing faults, inversion of the West-Netherlands and the Saar-Nahe basins (Fig. 5A). The most distal Paleocene inversion structures are located in the central North Sea, some 1700 km to the northwest of the contemporaneous Alpine collision front (Ziegler, 1990; Ziegler et al., 1998, 2002; de Lugt et al., 2003).

Similarly, in the foreland of the Western Alps, the Dauphinois Shelf and the Massif Central were uplifted and subjected to erosion during the Paleocene (Ziegler, 1988, 1990), presumably in response to compressional intraplate stresses emanating from the continent–continent collisional Pyrenean orogen (Vergés and García-Senez, 2001; Andeweg, 2002) and sinistral shear motions related to the incipient detachment of the Briançonnais terrane from the Iberian microcontinent. Related far-field intraplate stresses are held responsible for early inversion movements in the Paris Basin and Channel area (Fig. 5A; Ziegler, 1990; Ziegler et al., 2002).

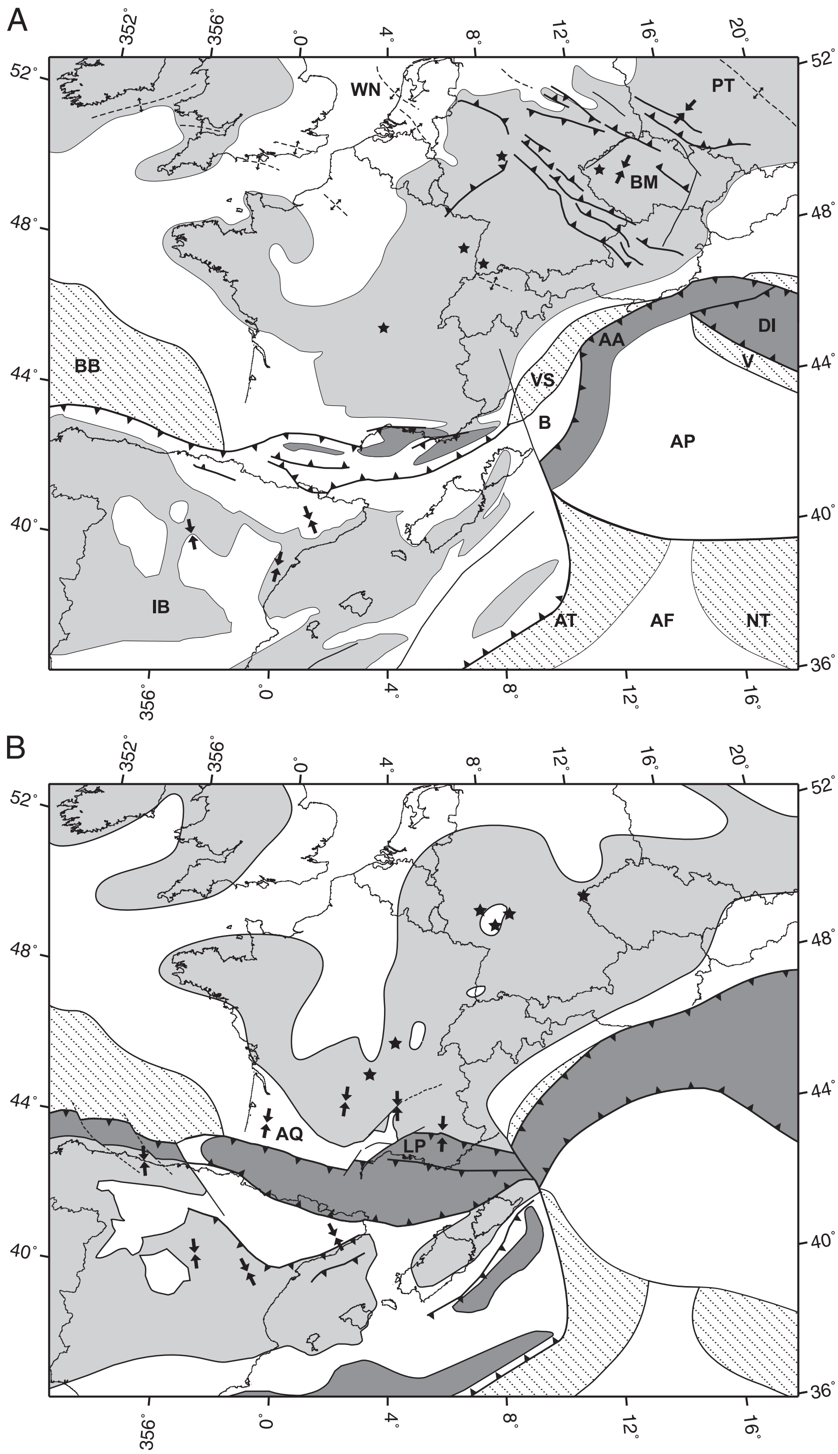
The Paleocene compressional deformation of the Alpine and Pyrenean forelands was accompanied by the injection of nephelinite and olivine melilite dykes in the area of the Massif Central, the Vosges-Black Forest, the Rhenish Massif and the Bohemian Massif. These magmas were derived by very low degree partial melting from an asthenospheric source and the lithospheric thermal boundary layer (Wilson et al., 1995; Ziegler et al., 1995; Michon and Merle, 2001; Keller et al., 2002). As such, this suggests that in these areas the temperature of the asthenosphere had increased to above ambient levels, presumably in response to activation of an asthenospheric upwelling system, or alternatively, of not-very-energetic mantle plumes (Granet et al., 1995; Sobolev et al., 1997; Goes et al., 1999; Ritter et al., 2001; Ziegler et al., 2004). Activation of this plume system was coeval with the development of the Iceland (Ziegler, 1990; Bijwaard and Spakman, 1999) and Northeast Atlantic plumes (Hoernle et al., 1995). All this indicates that the development of a broad thermal anomaly at the base of the West and Central European lithosphere had commenced during the Paleocene, causing its weakening and, thus, rendering it prone to deformation (Zielhuis and Nolet, 1994; Ziegler et al., 1995; Goes et al., 2000a,b).

5. Step 2: early and middle Eocene (54.8–37.0 Ma)

During the early Eocene (52.4 Ma), convergence rates between Africa and Europe gradually increased again; only by early Miocene times (19.2 Ma) did they decrease once more (Rosenbaum et al., 2002). Consequently, the Alpine orogenic wedge converged rapidly in a northerly direction with the European foreland during the Eocene (Schmid and Kissling, 2000).

In the Central-Alpine domain, imbrication of the sedimentary cover and upper crust of the Briançonnais terrane, as well as of the Valais Ocean sediments began during the early and middle Eocene (Schmid et al., 1996). This was accompanied by the subduction of the middle and lower crust, and also of the mantle-lithosphere of the Briançonnais (Stampfli et al., 1998). Sediment subduction presumably accounted for mechanical decoupling of the Alpine orogenic wedge from its foreland, as evidenced by the relaxation of compressional stresses in continental Europe at the transition from the Paleocene to the Eocene (Ziegler, 1990; Ziegler et al.,

Fig. 5. Palaeotectonic sketch maps of ECRIS area. (A) Late Paleocene, (B) middle Eocene, (C) late Eocene, (D) late Oligocene, (E) early middle Miocene, (F) Pliocene–Quaternary (modified after: Ziegler, 1990; Séranne, 1999; Sissingh, 2001, 2003b; Andeweg, 2002; Stampfli et al., 2002). Legend: dark grey: orogens, light grey: areas of non-deposition, white: sedimentary basins, stippled: oceanic basins, stars: volcanism, arrows: maximum horizontal compressional stress direction (after: Bergerat, 1987; Blès and Gros, 1991; Schumacher, 2002), thick dashed line: axis of lithospheric fold. Abbreviations: AA: Austro–Alpine orogen, AF: Africa, AP: Apulia, AT: Alpine Tethys (Ligurian-Maghrebian Ocean), AQ: Aquitaine Basin, B: Briançonnais, BB: Bay of Biscay, BM: Bohemian Massif, CS: Celtic Sea, DI: Dinarides, GV: Gulf of Lions-Valencia rift, IB: Iberia, LP: Languedoc-Provençal fold belt, PB: Provençal Basin, PT: Polish Trough, V: Vardar Ocean, VS: Valais Ocean, WA: Weald-Artois axis, WN: West-Netherlands Basin, WP: Western Approaches.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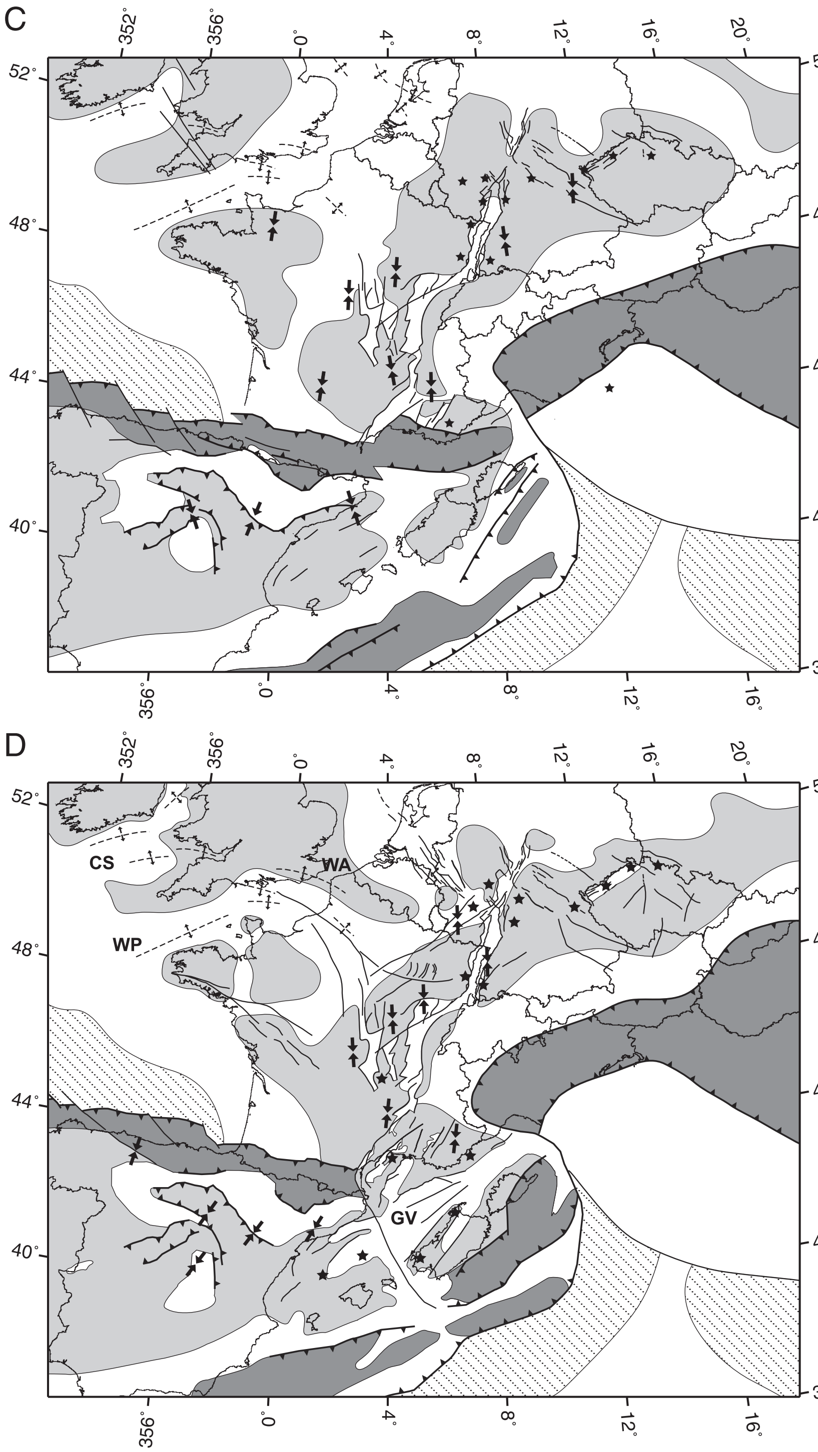

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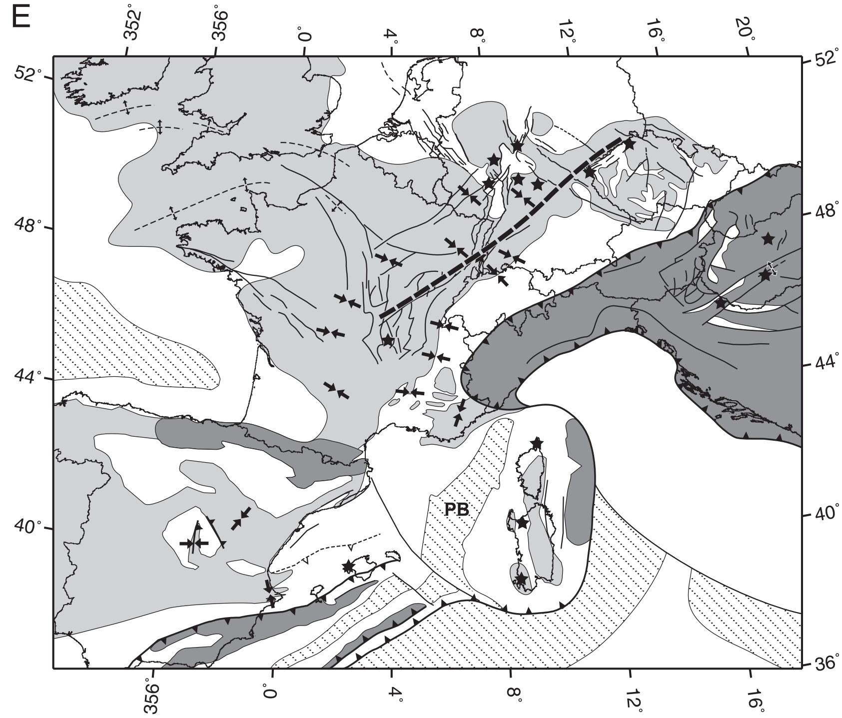

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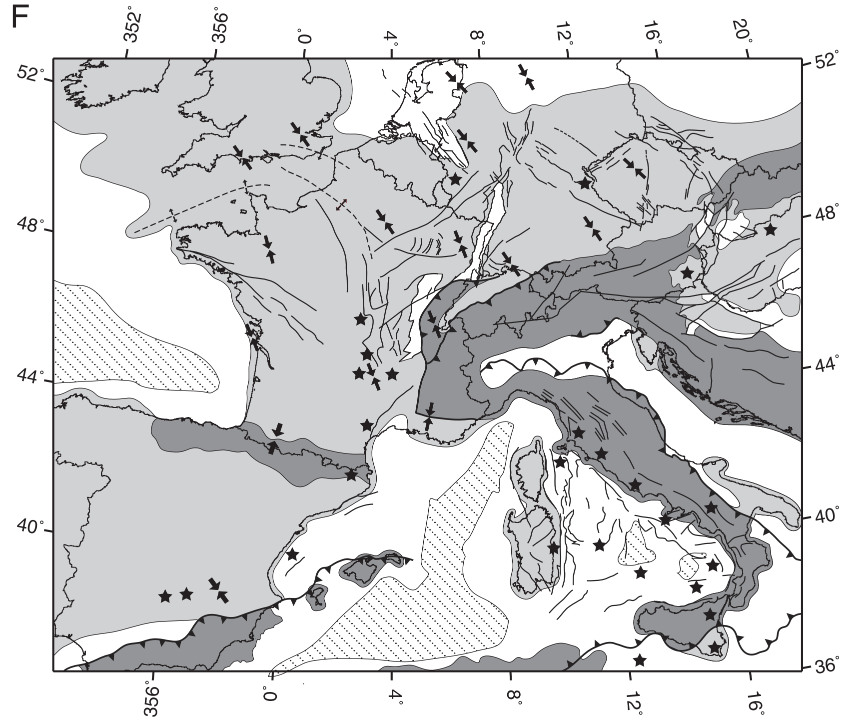

Fig. 5 (*continued*).

2002). By late middle Eocene times (Bartonian 41 Ma), flexural subsidence of the Helvetic Shelf commenced (Burkhard and Sommaruga, 1998), reflecting the onset of its slab- and thrust-loading in the Central Alpine transect (Fig. 4).

In the West-Alpine domain (Fig. 3), subduction of the southern parts of the Briançonnais terrane began during the early Eocene (Bucher et al., 2003) whilst sedimentation persisted in its northern parts (Stampfli et al., 2002). This was followed by detachment of upper crustal slices from the Briançonnais whereas its lower crust and mantle lithosphere were subducted. Note that in the French–Italian Alps sinistral oblique subduction of the Valais Ocean beneath the northerly moving Briançonnais upper crustal slices commenced at the same time (Bucher et al., 2003; Fügenschuh and Schmid, 2003). Flexural subsidence of the Dauphinois Shelf began in the middle Eocene (late Lutetian, ±45 Ma; Lickorish and Ford, 1998; Sissingh, 2001), presumably in response to its slab-loading and possibly also thrust-loading by the advancing West-Alpine orogenic wedge (Fig. 3).

In the Pyrenees, continued early and middle Eocene crustal shortening (Vergés and García-Senez, 2001; Andeweg, 2002) was accompanied by the main deformation phase of the Languedoc–Provençal thrust belt (Roure and Colletta, 1996) and the further development of the flexural Aquitaine foreland basin (Le Vot et al., 1996) (Fig. 5B).

In the area of the Upper Rhine, Bresse and Valence grabens, and on the Massif Central and Bohemian Massif, isolated, shallow depressions developed during the middle Eocene. In these basins, generally thin, fine-grained fluvio-lacustrine sediments accumulated that are separated by a distinct hiatus from the late Eocene earliest syn-rift deposits (Sissingh, 1998, 2001, 2003a). It is uncertain whether development of these depressions can be attributed to a middle Eocene initial rifting phase that involved necking of the lithosphere at a shallow level (Kooi et al., 1992) and that was controlled by the build-up of a northerly directed stress field.

Scattered volcanic activity occurred during the early and middle Eocene on the Massif Central (Michon and Merle, 2001), the Rhenish Massif (Lippolt, 1983) and the Bohemian Massif (Ulrych et al., 1999).

6. Step 3: late Eocene (37–33.7 Ma)

In the Central-Alpine domain, the bulk of the Briançonnais terrane and the Valais Ocean lithosphere had been subducted by late Eocene times, and shortening of the Helvetic Shelf commenced whilst upper crustal slices of the Briançonnais and Valais Ocean were incorporated into the orogenic wedge (Fig. 4). This was paralleled by further late Eocene thrust- and slab-loaded subsidence of the Helvetic Shelf and rapid widening of the flexural Molasse foreland basin (Schmid et al., 1996; Stampfli et al., 1998; Burkhard and Sommaruga, 1998). Note that during the middle and late Eocene the evolving orogenic wedge of the Alps advanced with respect to the European foreland by as much as 350 km (±35 mm/year; comparable to Miocene roll-back rates of the Apennine slab, Doglioni et al., 1999). This reflects rapid subduction of the Valais Ocean and the toe of the European margin in response to accelerated northward convergence of the Central- and East-Alpine orogenic wedge, culminating in its head-on collision with the Helvetic Shelf.

This essentially north-directed Eocene motion of the East- and Central-Alpine orogenic wedge was accompanied by continued sinistral oblique convergence of the West-Alpine orogenic wedge with the Dauphinois Shelf, involving stacking of crustal slices derived from the Briançonnais into nappes, and subduction of the corresponding lower crust and mantle-lithosphere (Fig. 3). Moreover, at this stage, the oceanic crust of the Valais Basin became imbricated and incorporated into the orogenic wedge (Schmid and Kissling, 2000; Bucher et al., 2003). During the late Eocene, gentle thin-skinned compressional deformation of the flexurally subsiding Dauphinois Shelf (Lickorish and Ford, 1998) testifies to weak collisional interaction of the West-Alpine orogenic wedge with its foreland.

Progressively increasing subduction resistance of the European lithosphere was coupled with imbrication of the basement of the distal Helvetic Shelf (Subpenninic Adula nappe: Schmid et al., 1996) and a renewed build-up of northerly directed intraplate compressional stresses in the foreland of the Central Alps (Bergerat, 1987). We propose that these stresses controlled the late Eocene early rifting phases of the Upper Rhine Graben, involving transtensional reactivation of Late Variscan and Permo–Carboniferous crustal discontinuities (Fig. 5C; Sissingh, 1998; Schumacher, 2002). At the Eocene–Oligocene transition, far-field compressional stresses governed the second inversion phase of the West-Netherlands Basin (Ziegler, 1990; de Lugt et al., 2003). Volcanic activity associated with this initial rifting stage of the Upper Rhine Graben (Lippolt, 1983; Keller et al., 2002) indicates gradually increasing activity of the Rhenish mantle plume (Ritter et al., 2001). Similarly, volcanic activity increased during the late Eocene on the Bohemian Massif (Ulrych et al., 1999).

Also the Valence, Massif Central and Bresse grabens, as well as the Burgundy transfer zone linking the latter with the Upper Rhine Graben, were activated during the late Eocene under a northerly directed compressional stress field. Development of this rift and wrench system involved tensional and transtensional reactivation of Permo–Carboniferous fracture systems (Bergerat, 1977, 1987; Michon and Merle, 2001; Sissingh, 2001). Significantly, there is no evidence for late Eocene volcanic activity on the Massif Central (Michon and Merle, 2001). Contemporaneous development of the Alès, Manosque and Camargue grabens in the southern Rhône Valley, which are superimposed on the Provençal thrust belt, involved transtensional reactivation of Mesozoic extensional fault systems (Sanchis and Séranne, 2000). Activation of the Massif Central-Rhône Valley rift system is attributed to north-directed intraplate compressional stresses, which originated in the Pyrenean continental collision zone, rather than in the Alps (Fig. 5C; Ziegler, 1994; Séranne, 1999; Merle and Michon, 2001; Andeweg, 2002). However, we doubt that slab-pull forces exerted by the east-dipping West-Alpine subduction slab interfered constructively with these Pyrenean stresses, as postulated by Merle and Michon (2001) and Michon et al. (2003), since in the Western Alps subduction of continental lower crust and mantle-lithosphere is oblique and, thus, unlikely to give rise to major slab-pull forces.

7. Step 4: Oligocene (33.7–23.8 Ma)

At the Eocene–Oligocene transition, convergence of the West-Alpine orogenic wedge with its European foreland changed to a northwesterly direction due to westward movement of the Apulian plate and parts of the Alpine orogen (Fig. 5D; Steck, 1984, 1990; Schmid and Kissling, 2000; Ceriani et al., 2001). At the same time, the subducted lithospheric slab of the Central and Eastern Alps was detached from the European foreland lithosphere (von Blanckenburg and Davies, 1995). This slab consisted of the continental lower crust and mantle-lithosphere of the distal parts of the Helvetic margin and of the Briançonnais, as well as of the lithosphere of the Valais and Piemont oceans (Fig. 4; Schmid et al., 1996). After this slab detachment, and related isostatic rebound of the European foreland lithosphere and development of the Apulian mantle backstop, convergence between the Apulian and European plates continued. The ongoing subduction of the continental European lithosphere was accompanied by imbrication of upper crustal slices (e.g., Gotthard Massif) and by back folding and back thrusting of the southern parts of the by now bi-vergent Alpine orogenic wedge (Schmid et al., 1996). This accounted for strong collisional coupling of the Central Alpine orogen with its European foreland, including the Rhine Graben area. Note that by late Oligocene times the Alpine deformation front was located only some 100 km to the south of the city of Basel where the Upper Rhine Graben terminates against the Burgundy transfer zone (Figs. 4 and 5D).

In the Western Alps, where the subducted lithospheric slab remained attached to the European foreland, intense crustal shortening persisted, as evidenced by the early Oligocene (33.7–28.5 Ma) emplacement of a stack of nappes derived from the Valais Ocean, the Briançonnais terrane and the Piemont Ocean on the Dauphinois Shelf, forming the Penninic frontal thrust (Fig. 3; Ceriani et al., 2001; Fügenschuh and Schmid, 2003; Bucher et al., 2003). This was accompanied by back folding of the West-Alpine orogenic wedge and by thrust propagation into the Dauphinois Shelf where it induced inversion of Mesozoic extensional basins (Roure and Colletta, 1996; Lickorish and Ford, 1998). By late Oligocene times, the West-Alpine deformation front was located some 100 km to the east of the Bresse Graben (Fig. 5D).

During the Oligocene, north-directed compressional stresses were projected from the still active central Pyrenean collision zone (Vergés and García-

Senez, 2001) into the European foreland. These stresses, together with those acting on the foreland of the Central Alps, played a dominant role in controlling the Oligocene main extensional stage of ECRIS (Fig. 5D; Bergerat, 1987; Villemin and Bergerat, 1987). In the area of Rhine rift systems, these "Pyrenean" stresses interfered constructively with north-directed compressional stresses exerted by the Central Alps on their northern foreland (Schumacher, 2002). By contrast, subsidence of the northeast trending Eger Graben, in which late Oligocene and early Miocene sediments attain maximum thicknesses of 300–400 m, commenced only towards the end of its Oligocene main magmatic pulse and was probably controlled by an indigenous stress field, related to the collapse of a thermal dome (Malkovsky, 1987; Adamovic and Coubal, 1999).

During its Oligocene main rifting phase, the Upper Rhine Graben propagated northward in Rupelian times into the Hessian and Roer Valley grabens (Fig. 5D; Ziegler, 1990; Schumacher, 2002; Michon et al., 2003; Sissingh, 2003b). Similarly, the grabens of the Massif Central and Rhône Valley subsided rapidly during the Oligocene and coalesced. This rifting activity was coupled with further sinistral movements along the complex Burgundy transfer zone (Bergerat, 1977; Ziegler, 1994; Merle and Michon, 2001) and activation of the more diffuse eastern Paris Basin transfer zone (Coulon, 1992). During the Oligocene, intermittent marine connections were established between the Alpine foreland basin and the North German Basin via the Rhône Valley and Bresse grabens and the Rhine rift system (Sissingh, 1998, 2001, 2003b). The Rhenish Massif and the Massif Central were still located close to sea level during Oligocene times, as indicated by the occurrence of remnants of marginal marine sediments on the former (Ziegler, 1990; 1994; Sissingh, 2003b) and by repeated marine incursions into the grabens of the latter (Merle et al., 1998; Michon and Merle, 2001; Merle and Michon, 2001; Sissingh, 2001).

Increasing volcanic activity in the area of the Rhine-Roer-Hessian graben triple junction, and late Oligocene gradual uplift of the Rhenish Massif can probably be attributed to intensifying activity of the Rhenish mantle plume and associated thermal thinning of the mantle-lithosphere (Lippolt, 1983; Jung, 1999; Ritter et al., 2001). On the other hand, on the Massif Central, rifting was accompanied only from the late Oligocene onward by scattered volcanic activity in its northern parts (Michon and Merle, 2001). By contrast, the Bohemian Massif experienced during the early and middle Oligocene a major phase of volcanism, which preceded the subsidence of the north-easterly striking Eger Graben (Malkovsky, 1987; Adamovic and Coubal, 1999; Ulrych et al., 1999).

During the late Rupelian and Chattian, the graben systems of the lower Rhône Valley propagated southward across the eastern, by now inactive parts of the Pyrenean orogen into the Gulf of Lions and along coastal Spain into the area of the Valencia Trough. This rift system remained active until late Aquitanian times (21.5 Ma) when crustal separation was achieved and the oceanic Provençal Basin began to open. Significantly, the grabens of the lower Rhône Valley became inactive upon crustal separation in the Provençal Basin. Development of the rather short lived Gulf of Lions–Valencia Trough rift system was driven by back-arc extension and eastward rollback of the West-Alpine Tethys subduction slab (Ligurian–Maghrebian Ocean), which dipped beneath Corsica–Sardinia and the Balearic islands (Fig. 5D; Séranne, 1999; Roca, 2001).

During the late Oligocene, lithospheric shortening compensating for Africa–Europe convergence, was gradually transferred from the Pyrenean collision zone to the Corsica–Sardinia–Balearic arc-trench system. Consequently, the "Pyrenean" north-directed intraplate compressional stresses gradually relaxed in the European foreland. On the other hand, late Oligocene–early Miocene back-arc extension, related to rollback of the Corsica–Sardinia–Balearic arc-trench system, controlled the evolution of the Gulf of Lions–Valencia rift and presumably contributed towards the subsidence of the lower Rhône Valley grabens (Séranne, 1999; Andeweg, 2002).

Turning back to the main elements of ECRIS, we conclude that a "Pyrenean" stress field, in which syn-collisional stresses emanating from the Pyrenees and Central Alps interfered constructively, essentially controlled the Oligocene main extensional stage of the Rhine and Massif Central–Rhône Valley rift systems. However, crustal extension across ECRIS demands the principle compressive stress axis to be oriented vertically, as indeed mapped by Bergerat

(1987), rather than sub horizontally as the collision-related foreland stresses. This asks for the interference of these foreland stresses with stresses controlled by gravitational forces, such as, for example, the load of upwelling mantle plumes. Conversely, we question that slab-pull forces exerted by the southeast- to east-dipping West-Alpine subduction system contributed to the extensional subsidence of the Bresse and Valence grabens and those of the Massif Central (Stampfli et al., 1998; Merle and Michon, 2001; Michon et al., 2003), which strike sub parallel to the contemporaneously active West-Alpine deformation front. From a kinematic point of view, the moderate cumulative E–W extension across the Rhine and Massif Central–Rhône Valley rift systems (in the order of 5–7 km, see below), demands a minor westward escape of areas located to the west of ECRIS. In this respect, it is noteworthy that the late Oligocene–early Miocene pulse of basin inversion, which is evident in the Celtic Sea, Western Approaches, Channel and Paris basins, can be attributed to a minor westward rotation of France in response to extensional strain achieved across ECRIS, as well as to the build-up of northwest-directed intraplate compressional stresses which were exerted by the Western Alps onto their foreland (Fig. 5D and E; Ziegler, 1990; Ziegler et al., 1995, 1998, 2002).

8. Step 5: Miocene (23.8–5.3 Ma)

Convergence rates between Africa and Europe were still high during the earliest Miocene, but abruptly decreased during the Burdigalian (19.5 Ma) (Rosenbaum et al., 2002) as the Alpine orogen progressively encroached on un-stretched European crust. In response to continued convergence and strong collisional coupling between the Alpine orogenic wedge and its European foreland, imbrication of the external massifs of the Western and Central Alps commenced during the Burdigalian (Mugnier et al., 1990; Schmid et al., 1996; Fügenschuh and Schmid, 2003). However, unlike in the Central Alps, where imbrication of the foreland basement is restricted to the upper crust (Fig. 4), a major lower crustal ramp-flat structure developed beneath the Western Alps, causing uplift of the entire orogenic wedge and its rapid exhumation (Fig. 3; Schmid and Kissling, 2000; Fügenschuh and Schmid, 2003). Imbrication and uplift of the Central- and West-Alpine external massifs was accompanied by the propagation of thin-skinned thrusts into the Helvetic and Dauphinois shelves and ultimately into the domain of the Jura Mountains (Schmid et al., 1996; Burkhard and Sommaruga, 1998; Lickorish and Ford, 1998; Philippe et al., 1996, 1998).

Increased collisional coupling between the Alpine orogen and its northern and western forelands, starting in Burdigalian times, had clear repercussions on the evolution of the Rhine and Massif Central–Rhône Valley rift systems, as will be discussed below (Fig. 5E).

During the Miocene, crustal extension across the Rhine, Roer and Hessian grabens continued under a northwest-directed stress field (Villemin and Bergerat, 1987; Schumacher, 2002), whilst the triple junction area of these grabens was gradually uplifted and became the site of increased volcanic activity (Lippolt, 1983; Jung, 1999; Sissingh, 2003b). The latter can be attributed to progressive thermal thinning of the mantle-lithosphere and thermal expansion of the remnant lithosphere above a relatively narrow, more intense asthenospheric thermal anomaly with a radius of about 100 km, corresponding to the Rhenish plume (Ritter et al., 2001).

Uplift of the Vosges-Black Forest Arch commenced during the Burdigalian (Laubscher, 1992). At top-basement level, and in a N–S direction, this arch has an amplitude of about 2.5 km and a wavelength of 200 km with a steeper southern and a gentler northern flank. At the Moho level, this arch forms the culmination of a broadly southwest–northeast trending anticlinal feature that extends from the Massif Central via the Burgundy transfer zone towards the Bohemian Massif (Figs. 2 and 5E; Dèzes and Ziegler, 2002). As the Vosges–Black Forest Arch is not associated with mantle-lithospheric thinning (Achauer and Masson, 2002), its development must be related to folding of the entire lithosphere. Conglomerates shed from this arch southward into the domain of the Jura Mountains (Kemna and Becker-Haumann, 2003) suggest that its erosional unroofing commenced around 14 Ma ago (Serravallian). Moreover, uplift of this arch was accompanied by relatively minor volcanic activity within and outside the Upper Rhine Graben spanning 18–7 Ma

(Jung, 1999). Magma segregation occurred 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). Uplift of the Vosges–Black Forest Arch, which started prior to and continued during the late Miocene to early Pliocene phase of thin-skinned folding of the Jura Mountains (10–9 to 4 Ma; Laubscher, 1986, 1992; Philippe et al., 1996; Becker, 2000), was coupled with uplift of the southern parts of the Upper Rhine Graben and deep truncation of its Oligocene–early Miocene syn-rift sedimentary fill prior to the late Pliocene resumption of sedimentation (Roll, 1979; Villemin et al., 1986; Sissingh, 1998; Schumacher, 2002). A corresponding erosional unconformity is clearly imaged by high-resolution reflection-seismic lines recorded on the Rhine River under the auspices of the EUCOR-URGENT Project. These lines show that the hiatus across this unconformity gradually decreases north of the Kaiserstuhl and that in the northern parts of the Upper Rhine Graben sedimentation was continuous during the Miocene and Pliocene. So far, the age of this unconformity has not yet been stratigraphically calibrated more closely than about mid-Burdigalian (±18 Ma; end upper Hydrobia beds; Roll, 1979). End-Aquitanian (20.5 Ma) transpressional reactivation of pre-existing basement discontinuities at the southern end of the Upper Rhine Graben (Laubscher, 2003) argues for the build-up of compressional stresses at crustal levels that may have heralded the uplift of the Vosges–Black Forest Arch. Whereas Miocene and early Pliocene crustal extension across the Upper Rhine Graben controlled continued subsidence of its northern parts, uplift of its southern parts, related to lithospheric folding, apparently overcompensated for their extensional subsidence.

Marine connections between the Upper Rhine Graben and the Alpine Foreland and North German basins were severed during the Burdigalian and end-Langhian, respectively (Sissingh, 1998, 2003b) in response to lithospheric folding in the area of the Burgundy transfer zone and the Vosges-Black Forest and to thermal doming of the Rhenish Massif. Therefore, from the Serravallian onward the erosional base level in the Upper Rhine Graben was controlled by the elevation of tectonic sills, which separated it from the adjacent basins.

Early Miocene NW–SE extension across the Eger Graben intensified after the 18 Ma end of sedimentation (Adamovic and Coubal, 1999) and was accompanied by progressive uplift of the northern parts of the Bohemian Massif, presumably in response to lithospheric folding. At the same time, volcanic activity in the Bohemian–Silesian zone decreased gradually (Ulrych et al., 1999).

In 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 southwards (Séranne, 1999; Sissingh, 1998, 2001, 2003a). Development of this unconformity may be related to lithospheric folding along the Burgundy transfer zone that is characterized by an anticlinal uplift of the Moho discontinuity (Lefort and Agarval, 1996; Dèzes and Ziegler, 2002). Moreover, uplift and northward tilting of the Massif Central commenced also during the Burdigalian, as evidenced by the development of its north-directed drainage system. 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 deposits (Merle et al., 1998; Michon and Merle, 2001; Sissingh, 2001).

On the Massif Central, volcanic activity increased dramatically during the middle Miocene (14 Ma), shifted to its southern parts where it peaked during the late Miocene (9–6 Ma), whereas in its northern parts volcanism was interrupted during the late middle Miocene (12 Ma) (Michon and Merle, 2001). This probably reflects increased activity of the Massif Central mantle plume (Sobolev et al., 1997). From the Burdigalian onward the gradual, though slow uplift of the Massif Central was probably mainly driven by plume-related thinning of the mantle-lithosphere and related thermal expansion of the remnant lithosphere (Michon and Merle, 2001) with lithospheric folding playing a subordinate role.

In the Valence Graben, late Burdigalian and younger series seal Oligocene extensional faults (Philippe et al. 1998). The lower Rhône Valley grabens became inactive with the late Aquitanian (21.5 Ma) onset of sea-floor spreading in the Provençal Basin that lasted until 16.5 Ma (late Burdigalian; Séranne, 1999; Roca, 2001). During the

Miocene, the Manosque and Alès grabens were partly inverted (Roure and Colletta, 1996; Sanchis and Séranne, 2000) in response to the build-up of west-directed compressional stresses originating in the Alps (Blès and Gros, 1991).

Late Miocene further inversion of the Paris, Channel and Western Approaches basins testifies to the build-up of northwest-directed collision-related compressional stresses in the West-Alpine foreland (Fig. 5E; Ziegler, 1990). These stresses, as well as thermal and lithospheric folding-related uplift of the Massif Central, impeded further subsidence of the Limagne, Bresse and Valence grabens. Middle and late Miocene lithospheric folding, controlling uplift of the Vosges–Black Forest Arch, the Burgundy transfer zone and contributing to the uplift of the Massif Central, entailed northwestward tilting of the Paris Basin, erosion of the sedimentary fill, particularly along its southeastern margin, and the gradual development of its modern drainage system (Fig. 5E; Mégnien, 1980).

9. Step 6: Pliocene–Quaternary (5.3–0 Ma)

The deep lithospheric configuration of the Alps has recently been imaged by high-resolution 3D seismic tomography (Lippitsch, 2003). The Central Alps are characterized by a subducted slab, which is still attached to the lithosphere and extends to depths of 150 km or even to 200 km. This slab is separated by a 50–150-km wide slab-window from the deep-reaching Valais Ocean–Briançonnais–Alpine–Tethys slab that was detached from the lithosphere at the end of the Eocene (Fig. 4). Correspondingly, the slab, which is still attached to the Alpine lithosphere, consists of continental mantle-lithospheric and lower crustal material that was subducted during Oligocene to recent times. Its length corresponds closely to the 180 km of post-Eocene crustal shortening in the Central Alps (Schmid et al., 1996). By contrast, beneath the Western Alps a narrow slab window is evident at a depth of about 110 km (Fig. 3), separating the Alpine lithosphere from a long, deep-reaching slab which presumably consists of the subducted oceanic Piemont and Valais lithosphere plus Briançonnais and European continental mantle-lithospheric material. This slab was probably detached from the Alpine lithosphere not earlier than during the early Pliocene (5 Ma), as suggested by the activation of major extensional faults in the frontal parts of the Briançonnais nappes (Sue and Tricart, 2002; Fügenschuh and Schmid, 2003).

Crustal shortening in the Western and Central Alps persisted during the Pliocene and Quaternary, as evidenced by continued uplift of their external crystalline massifs (Jouanne et al, 1998; Fügenschuh and Schmid, 2003) and further compressional deformation of the Jura Mountains (Philippe et al., 1996; Giamboni et al., 2004) and the Dauphinois (Martinod et al., 1996; Lickorish and Ford, 1998). Neotectonic activity within the Alps and their foreland is documented by their seismicity (Giglia et al., 1996; Deichmann et al, 2000; Schmid and Kissling, 2000; Sue et al., 2000) and by geodetic data, which indicate for the Dauphinois and French Jura Mountains present-day shortening rates of 2–5 mm/year (Jouanne et al., 1995; Martinod et al., 1996). From about 4 Ma onward, compressional deformation of the Jura Mountains was no longer exclusively thin skinned, but involved also the basement, as indicated by intra-crustal earthquakes (Roure et al., 1994; Philippe et al., 1996; Jouanne et al., 1998; Becker, 2000; Giamboni et al., 2004).

Under the present stress regime the Roer Valley graben is extending nearly orthogonally, whereas the Upper Rhine and lower Rhône Valley grabens are subjected to sinistral transtension (Fig. 5F; Ahorner, 1975; Giglia et al., 1996; Hinzen, 2003). Pliocene and Quaternary transtensional and extensional subsidence the Upper Rhine and Roer Valley grabens, respectively, is documented by the fault-controlled thickness of their Plio–Quaternary sedimentary fill (about 380 m near Heidelberg), the occurrence of active faults, geomorphologic data and by earthquake activity (Ahorner, 1983; Zijerveld et al., 1992; Plenefisch and Bonjer, 1997; Klett et al., 2002; Schumacher, 2002; Michon et al., 2003; Hinzen, 2003). By contrast, there is no evidence for further subsidence of the Massif Central grabens (Michon and Merle, 2001), whereas the Bresse Graben was tensionally reactivated during the Plio–Pleistocene deposition of up to 400 m thick fluvial and lacustrine series (Sissingh, 1998). In the Rhône Valley, there is evidence for minor Pliocene–Quaternary normal faulting (Blès and Gros, 1991).

The present north- to northwest-directed compressional stress regime of Western and Central Europe (Müller et al., 1997) reflects a combination of forces related to continued counter clock-wise convergence of Africa–Arabia with Europe, and consequently collisional interaction of the Alpine orogen with its foreland, and North Atlantic ridge push (Gölke et al., 1996). In the area of the Rhine rift system, the trajectories of the Miocene and present stress field are very similar (Schumacher, 2002). Yet, the magnitude of stresses apparently increased between 3 and 2.5 Ma. This is compatible with a subsidence acceleration of the Roer Valley Graben around 2.5 Ma (Zijerveld et al, 1992; Geluk et al., 1994; Michon et al., 2003). Moreover, the palaeo-Aare river, which between 5 and 4.2 Ma had followed the thrust front of the Jura Mountains before it entered the Upper Rhine Graben at Basel, and which from 4.2 Ma onwards flowed westward into the Bresse Graben (Sundgau gravels), was deflected again into the Upper Rhine Graben around 2.9 Ma (Müller et al., 2002; Giamboni et al., 2004). This may be attributed to a slow-down or the end of up warping of the Vosges–Black Forest Arch and the resumption of tensional subsidence of the southern parts of the Upper Rhine Graben. Geodetic data show 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). In the area of the Rhenish Massif, volcanic activity shifted during the Pliocene and Quaternary towards the Eifel area (Lippolt, 1983). Geomorphologic data indicate that around 0.8 Ma uplift of the Rhenish Massif accelerated, amounting since then in the Eifel area to as much as 250 m (average rate 0.3 mm/year; Garcia-Castellanos et al., 2000; Meyer and Stets, 2002), and still continues today at rates of up to 1.2 mm/year (Mälzer et al., 1983). Uplift of this arch is attributed to the combined effects of the Eifel plume thermal load (radius 50 km; Ritter et al., 2001; Garcia-Castellanos, 2000), thermal thinning of the mantle-lithosphere (Prodehl et al., 1995), and possibly also to crustal-scale folding and/or the reactivation of Variscan thrust faults under the prevailing northwest-directed compressional stress field (Ahorner, 1983; Hinzen, 2003).

High-resolution reflection-seismic data indicate that the northern parts of the Upper Rhine Graben subsided continuously during Miocene to Quaternary times with some faults extending upward through Quaternary deposits. A minor, down-cutting base-Quaternary(?) erosional unconformity, evident in the northernmost parts of the Upper Rhine Graben, may have developed in conjunction with up doming of the Rhenish Massif. Southward, this unconformity disappears towards the Heidelberg depocentre. On the other hand, south of the city of Speyer, late Miocene and Pliocene fluvial and lacustrine sediments progressively overstep the intra-Burdigalian unconformity to the end that sedimentation in the southern parts of the Upper Rhine Graben resumed only during the late Pliocene and Quaternary. Syn-sedimentary extensional faults and local positive flower-structures (Strasbourg transfer zone) were active during the Plio–Quaternary subsidence of the Upper Rhine Graben (see also Illies et al., 1981).

The fact that Alpine detrital components occur for the first time in the Roer Valley Graben at the Plio–Quaternary transition (1.65 Ma, Boenigk, 2002; Heumann and Litt, 2002) indicates that during the late Pliocene (2.9–1.65 Ma) the sedimentary load of the Rhine River was effectively trapped in the Upper Rhine Graben. Sediment supply to the Upper Rhine Graben was apparently in balance with the generation of accommodation space by extensional subsidence and tectonic controls on the erosional base level (uplift of the Rhenish Massif). During the late Pliocene, the Upper Rhine Graben was presumably drained by a northward flowing low energy river (Bingen–Koblenz Rhine) that linked up with the higher energy Moselle river which crossed the Rhenish Massif and debouched into the Roer Valley Graben where the Kieseloolite sands and gravels were deposited (Brunnacker and Boenigk, 1983; Klett et al., 2002; Sissingh, 2003b). At the end of the Pliocene, sediment supply to the Upper Rhine Graben apparently exceeded its subsidence rates. With this, the sediment load and perhaps also the energy of the Bingen–Koblenz Rhine increased, facilitating the transport of Alpine components across the Rhenish Massif into the Roer Valley Graben. During the Quaternary, the position of the erosional base level in the continuously subsiding Upper Rhine Graben was controlled by the balance between the uplift-rate of the Rhenish Massif and the incision rate of the Rhine River. Presently the erosional base level of the Upper Rhine Graben is located 80 m above MSL at Bingen where the Rhine canyon starts to cut across the Rhenish Massif.

In the southern parts of the Upper Rhine Graben, Plio–Quaternary tectonic activity is indicated by folding of the Pliocene Sundgau gravels along the Jura Mountains thrust front (Giamboni et al., 2004), by faults extending through Quaternary deposits of the graben fill, and by the seismicity of the area. Earthquake focal mechanisms indicate that deformation of the upper crust is controlled by a strike-slip to compressional stress regime whilst the lower crust is subjected to extension (Plenefisch and Bonjer, 1997; Deichmann et al., 2000). Transpressional deformation of the upper crust can be attributed to collision-related stresses, which are transmitted from the Alps above an incipient mid-crustal detachment level. By contrast, lower crustal extension may be related to buckling of the mantle-lithosphere, controlling uplift of the Vosges–Black Forest Arch, in response to collision-related stresses transmitted from the Alps through the mechanically strong parts of the mantle-lithosphere. Significantly, earthquakes occur almost down to the Moho but are absent below it (Plenefisch and Bonjer, 1997; Deichmann et al., 2000). In the area of the Vosges-Black Forest Arch, crustal thicknesses range between 24 and 28 km, with the laminated lower crust accounting for 7–10 km (Brun et al., 1992; Dèzes and Ziegler, 2002).

Uplift of the northern Bohemian Massif, which had commenced during the Burdigalian, persisted during the Pliocene and Quaternary and was marked by a renewed volcanic activity (Ziegler, 1990; Ulrych et al., 1999; Michon and Merle, 2001). This uplift is attributed to lithospheric folding (Ziegler et al., 2002; Ziegler and Dèzes, 2005). Lithosphere thicknesses decrease from 100 km beneath the northern and southern flanks of the Bohemian Massif to 80 km in the area of the Eger Graben (Babushka and Plomerova, 2001). This can be mainly attributed to thermal thinning of the mantle-lithosphere during the Oligocene pulse of increased volcanic activity in the area of the Eger Graben. Resulting thermal weakening of the lithosphere presumably caused localization of the postulated lithospheric fold. However, as mantle tomographic data are not yet available for this area, a mantle-plume support of the Bohemian Massif Arch cannot be excluded.

At the Miocene–Pliocene transition, the eastern margin of the Bresse Graben was overridden by the frontal, “Ledonian” thrust sheet of the Jura Mountains by 3.5 km (Fig. 5F; Chauve et al., 1988; Guellec et al., 1990; Roure et al., 1994). The toe of this thrust sheet is sealed by Plio–Pleistocene lacustrine and fluvial clastics, which attain thicknesses of up to 400 m in the south central parts of the graben (Sissingh, 2001). The corresponding accommodation space was provided by a mild tensional reactivation of the Bresse Graben and not so much by its incorporation into the flexural Alpine foreland basin, as postulated by Merle et al. (1998). A tensional model for this late subsidence phase of the Bresse Graben is compatible with the reactivation of its border faults (Rat, 1984) and the distribution and thickness of its Plio–Pleistocene sedimentary fill (Rocher et al., 2003). Moreover, mild reactivation of the complex eastern border fault zone of the graben caused a basin-ward downward deflection of the sole of the Ledonian thrust sheet by some 80 m (Rat, 1984; Chauve et al., 1988). From about 1.4 Ma onwards, the Bresse Graben was tilted to the West, uplifted and subjected to erosion in conjunction with uplift of the northern parts of the Massif Central and further deformation of the Jura Mountains (Sissingh, 1998, 2001; Rocher et al., 2003). Moderate Pliocene and Quaternary extension across the Bresse and Upper Rhine grabens was presumably accompanied by sinistral movements along of the seismically still active Burgundy transfer zone (Giglia et al., 1996).

On the Massif Central, volcanic activity resumed in its northern parts at the transition to the Pliocene (5.5 Ma) 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 a presently sub-active level (Michon and Merle, 2001). During the Pliocene and Quaternary, a volcanic chain developed that extends from the Massif Central southward to the Mediterranean (Maury and Varet, 1980). All this is taken as evidence for increased activity of the Massif Central mantle plume. From about 3.5 Ma onwards, uplift of the Massif Central accelerated (Michon and Merle, 2001) 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 northwest-directed stress field, the Massif Central is subjected to transtensional deformation (Delouis et al, 1993). This stress field probably came into evidence during the Pliocene upon decay of the Miocene west-directed West-Alpine compressional stresses (Blès and Gros, 1991). Progressive plume-related thermal uplift and northward tilting of the Massif Central apparently counteracted transtensional subsidence of its graben system, thus impeding accumulation of thick Pliocene and Quaternary sequences.

10. Magnitude of extensional strain across ECRIS

The magnitude of late Eocene to recent crustal extension across ECRIS is a matter of debate. Differences in estimates largely depend on whether extensional strain is derived from upper crustal faulting in the different grabens or from their crustal configuration (Ziegler, 1994; Merle et al., 1998; Ziegler and Cloetingh, 2004).

Extensional strain derived from upper crustal faulting amounts to about 2 km across the Bresse Graben, 3–4 km across the grabens of the Massif Central and apparently does not exceed 7 km across the Upper Rhine Graben. Across the Roer Valley Graben upper crustal extension diminishes from 4 to 5 km in its south-eastern parts to zero near the Dutch North Sea coast. However, based on uniform pre-rift crustal thickness assumptions, the crustal configuration of the Roer Valley, Upper Rhine and Limagne grabens indicates a two to three times greater amount of extension (Bergerat et al., 1990; Brun et al, 1992; Geluk et al., 1994; Ziegler, 1994; Merle et al., 1998). Note, however, that ECRIS is characterized by a broad belt of Moho shallowing that cannot be exclusively attributed to Cenozoic crustal extension, thermal uplift of the Massif Central and Rhenish Massif and lithospheric folding in the area of the Burgundy transfer zone and the Vosges-Black Forest and Bohemian Massif arches (Fig. 2). Therefore, it is likely that in the ECRIS area crustal thicknesses were not uniform at the onset of Cenozoic rifting, mainly due to lateral variations in the intensity of Permo–Carboniferous crustal thinning (see, e.g., western part of Massif Central; Ziegler and Dèzes, 2005) and Mesozoic crustal extension (e.g., Roer Valley Graben; Zijerveld et al., 1992). Moreover, it cannot be excluded that the Moho was destabilized in areas of intense Cenozoic magmatism (e.g., Rhenish Massif, Massif Central; Ziegler and Cloetingh, 2004).

Giving preference to extension values derived from upper crustal faulting, we assume that crustal extension across the southern and central parts of ECRIS was in the order of only 5–7 km, decreasing to zero at its northern termination. Crustal extension across ECRIS apparently entailed a corresponding rotational westward displacement of France, involving repeated reactivation of northwest trending Variscan and Permo–Carboniferous shears, which transect the Armorican Massif (Van Vliet-Lanoé et al., 1998) and the Paris Basin (Fig. 1; Ziegler, 1990). It is tempting to assume that this displacement was compensated by the inversion of Mesozoic extensional basins in the Channel–Western Approaches–Celtic Sea area and by up warping of the Weald-Artois axis (Fig. 5D–F; Ziegler, 1987, 1990; Butler and Pullan, 1990; Mansy et al., 2003). However, the main inversion pulses of these basins are not fully synchronous with the main subsidence phases of ECRIS. Nevertheless, far-field Pyrenean and Alpine collision-related stresses presumably contributed directly to the inversion of these basins (Ziegler, 1990), just as they did during the extensional subsidence of ECRIS, as discussed above.

11. Conclusions

ECRIS developed by passive rifting in the foreland of the Alps and the Pyrenees in response to the build-up of syn-collisional compressional intraplate stresses, which reactivated pre-existing crustal discontinuities during the early rifting phases. Once these stresses had reached a critical magnitude, westward escape of the French block, flanking ECRIS to the west, allowed for northward and southward rift propagation. Examples of similar syn-orogenic foreland splitting are provided by the development of the late Carboniferous Norwegian–Greenland Sea Rift in the foreland of the Variscan orogen, the Permo–Carboniferous Karoo rifts in the hinterland of the Gondwanan orogen and the Neogene Baikal Rift in the hinterland of the Himalayas (Ziegler et al., 2001; Ziegler and

Cloetingh, 2004). By contrast, evolution of the Gulf of Lions–Valencia Trough rift system, forming the southern extension of ECRIS, was controlled by back-arc extension related to rollback of the Ligurian–Maghrebian Alpine–Tethys slab (Séranne, 1999; Roca, 2001).

ECRIS was activated during the late Eocene and evolved under repeatedly changing stress fields (Bergerat, 1987; Schumacher, 2002; Michon et al., 2003), reflecting changes in the interaction of the Pyrenean and Alpine orogens with their forelands. The Rhine Rift System remained active until the present, whereas the rifting stage of the Massif Central–Rhône Valley grabens ended in the early Miocene (Michon and Merle, 2001).

Volcanic activity in the ECRIS area commenced during the Paleocene, gradually increased during the Oligocene and Miocene and is presently sub-active (Wilson and Bianchini, 1999; Michon and Merle, 2001). Seismic tomography images discrete upper mantle plumes beneath the Massif Central and the Rhenish Massif (Granet et al., 1995; Sobolev et al., 1997; Ritter et al., 2001). Localization of major magmatic activity in the ECRIS area suggests the gradual assertion of these mantle-plumes and related thermal thinning of the mantle lithosphere. Crustal extension played thereby a secondary role in lithospheric thinning, partial melting and melt extraction, although it provided conduits for the ascent of magmas to the surface (Wilson and Downes, 1992; Ziegler et al., 1995). The Paleocene pre-rift injection of melilite dykes in the ECRIS area (Wilson et al., 1995) is interpreted as reflecting the activation of a mantle upwelling system beneath the Alpine foreland. The upper asthenospheric P- and S-wave low velocity anomaly, which is observed in Western and Central Europe to a depth of less than 200 km, requires the presence of partial melts in order to be compatible with the observed surface heat flow (Zielhuis and Nolet, 1994; Hoernle et al., 1995; Goes et al., 2000a,b). Development of this anomaly, which presumably commenced during the Paleocene, and which can be interpreted as the head of mantle-plumes, was associated with an increase in the potential temperature of the asthenosphere. This caused thermal weakening of the foreland lithosphere, thus rendering it prone to deformation (Ziegler et al., 1995).

11.1. Pre-rift stage

Paleocene intraplate compressional deformation of the Western and Central Europe reflects strong collisional mechanical coupling of the East-Alpine–Carpathian orogen with the European foreland, of the Central Alpine orogen with the Briançonnais terrane and of the Pyrenean orogen with its northern foreland. Compressional stresses originating at the Alpine collision zone relaxed in the foreland at the transition to the Eocene, presumably in response to sediment subduction (Ziegler et al., 1995, 2002). Eocene north-directed lithospheric shortening in the Central Alps and sinistral oblique convergence of the Western Alps with their foreland involved subduction of the Briançonnais terrane and the Valais Ocean (Schmid and Kissling, 2000). This was accompanied by a rapid advance of the evolving Alpine orogenic wedge towards the ECRIS area. Slab- and thrust-loaded subsidence of the Helvetic and Dauphinois shelves commenced during the middle Eocene.

11.2. Early rifting stage

During the late Eocene, subduction resistance of the European foreland increased. Consequently north-directed intraplate compressional stresses, originating at the Pyrenean and Alpine collision zones, built-up again, activating ECRIS during the late Eocene (Séranne, 1999) by reactivation of Variscan, Permo–Carboniferous and Mesozoic crustal-scale faults. At the Eocene–Oligocene transition the same stresses controlled far-field basin inversion in the southern North Sea and Western Shelf areas (Ziegler et al., 1995, 1998). Slab-pull forces exerted by the Alpine subduction system probably did not contribute to the late Eocene activation of ECRIS.

11.3. Main rifting stage

The Oligocene main rifting phase of ECRIS was controlled by northerly directed compressional stresses originating at the Pyrenean and Alpine collision zones. Following detachment of the Central- and East-Alpine subduction slab at the Eocene–Oligocene transition (von Blanckenburg and Davies, 1995; Davies and von Blanckenburg, 1995), the Alpine orogenic wedge converged northwestward with its foreland under conditions of increased collisional coupling (Schmid and Kissling, 2000; Ziegler et al., 2002). Therefore, a slab-pull model (Michon et al., 2003) is rejected as a driving mechanism for the Oligocene evolution of the Rhine Rift System. Similarly, it is unlikely that slab-pull forces associated with the oblique West-Alpine subduction system contributed, let alone controlled, crustal extension across the Massif Central–Rhône Valley Rift System, as postulated by Michon and Merle (2001). During the late Oligocene, lithospheric shortening compensating for Africa–Europe convergence was gradually transferred from the Pyrenean collision zone to the Corsica–Sardinia–Balearic arc-trench system due to increasing subduction resistance of the Iberian lithosphere. Correspondingly, intraplate compressional stresses originating at the Pyrenean collision zone gradually decayed in continental Europe during the late Oligocene. At the same time, eastward roll-back of the West-Alpine Tethys subduction slab, which dipped beneath Corsica, Sardinia and the Balearic islands, commenced and controlled by back-arc extension the development of the Gulf of Lions–Valencia Trough rift system and further subsidence of the lower Rhône Valley grabens (Séranne, 1999; Roca, 2001; Andeweg, 2002).

11.4. Late rifting stage

The late rifting stage of ECRIS commenced with the late Aquitanian onset of sea-floor spreading in the Provençal Basin. As a consequence of this, the lower Rhône Valley grabens became inactive (Séranne, 1999). However, during the Burdigalian west- and northwest-directed compressional stresses built up in the Alpine foreland (Bergerat, 1987; Blès and Gros, 1991) in conjunction with imbrication of the Alpine external massifs (Fügenschuh and Schmid, 2003). These stresses controlled the further evolution of ECRIS, involving continued extensional subsidence of the Roer Valley Graben and the northern parts of the Upper Rhine Graben, and, from the mid-Burdigalian onward, uplift of the Vosges–Black Forest Arch and the southern parts of the Upper Rhine Graben owing to lithospheric folding. Relatively minor Miocene magmatic activity in the southern Upper Rhine Graben area (Keller et al., 2002) can be attributed to decompressional partial melting of the asthenosphere and lower lithosphere in response to lithospheric folding controlling uplift of the Vosges–Black Forest Arch, which at the Moho level extends into the Massif Central and the Bohemian Massif (Ziegler and Dèzes, 2005). By contrast, progressive uplift of the Rhenish triple junction is mainly attributed to thermal thinning of the mantle-lithosphere above the Rhenish mantle plume (Jung, 1999; Ritter et al., 2001). Similarly, plume-related thermal thinning of the mantle-lithosphere contributed towards uplift of the Massif Central (Sobolev et al., 1997; Michon and Merle, 2001). Whether plume activity contributed to the mid-Miocene and later uplift of the Bohemian Massif is uncertain (Ziegler et al., 2002). Subsidence of the Massif Central–Rhône Valley grabens terminated during the early Miocene, presumably due to the build-up of west-directed compressional stresses in the West-Alpine foreland (Blès and Gros, 1991) and thermal uplift and tilting of the Massif Central (Michon and Merle, 2001). Following detachment of the subducted West-Alpine lithospheric slab at the Miocene–Pliocene transition (Sue and Tricart, 2002), the Bresse Graben was apparently tensionally reactivated during the Pliocene by north-directed stresses.

11.5. Neotectonic activity

On-going deformation of ECRIS is related to a late Pliocene increase in the magnitude of the present northwest-directed stress field (Müller et al., 1997). This stress field reflects a combination of forces related to continued counter clock-wise convergence of Africa–Arabia with Europe and North Atlantic ridge push (Gölke et al., 1996; Giglia et al., 1996). Note that the Rhine Rift System is not characterized by an indigenous stress field (Plenefisch and Bonjer, 1997). Under the present stress field the Upper Rhine and Roer Valley grabens continue to subside and are sites of increased seismic hazard (Giardini et al., 2003). By contrast, the lower Rhône Valley grabens are only mildly reactivated (Blès and Gros, 1991).

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References

Achauer, U., Masson, F., 2002. Seismic tomography of continental rifts revisited: from relative to absolute heterogeneities. Tectonophysics 358, 17–37.

Adamovic, J., Coubal, M., 1999. Intrusive geometries and Cenozoic stress history of the northern part of the Bohemian Massif. Geolines, Prague 9, 5–14.

Ahorner, L., 1975. Present-day stress field and seismotectonic block movements along major fault zones in Central Europe. Tectonophysics 29, 233–249.

Ahorner, L., 1983. Historical seismicity and present-day microearthquake activity of the Rhenish Massif, Central Europe. In: Fuchs, K., VonGehlen, K., Mälzer, M., Murawski, H., Semmel, A. (Eds.), Plateau Uplift, The Rhenish Shield—A Case History. Springer-Verlag, Berlin, pp. 198–221.

Andeweg, B., 2002. Cenozoic tectonic evolution of the Iberian Peninsula, causes and effects of changing stress fields. Ph.D. Thesis Vrije Universiteit Amsterdam. Netherlands Research School of Sedimentary Geology, Publ. 2020101. 178 pp.

Babushka, V., Plomerova, J., 1992. The lithosphere in Central Europe—seismological and petrological approach. Tectonophysics 207, 141–163.

Babushka, V., Plomerova, J., 2001. Subcrustal lithosphere around the Saxothuringian-Moldanubian suture zone—a model derived from anisotropy of seismic wave velocities. Tectonophysics 332, 185–199.

Becker, A., 2002. The Jura Mountains—an active foreland fold-and-thrust belt? Tectonophysics 321, 381–406.

Bergerat, F., 1977. La fracturation de l'avant-pays Jurassien entre le fossé de la Saône et du Rhin, analyse et essai d'interpretation dynamique. Rev. Geogr. Phys. Geol. Dyn. (2)XIX(4), 325–338.

Bergerat, F., 1987. Stress fields in the European Platform at the time of Africa–Eurasia collision. Tectonics 6, 99–132.

Bergerat, F., Mugnier, J.-L., Guellec, S., Trufferrt, C., Cazes, M., Damotte, B., Roure, F., 1990. Extensional tectonics and subsidence of the Bresse basin: an interpretation from ECORS data. Mem. Soc. Geol. Fr. 156, 145–156.

Berggren, W.A., Kent, D.V., Swisker III, C.C., Aubrey, M.P., 1995. Revised Cenozoic geochronology and chronostratigraphy. In: Berggren, W.A., Kent, D.V., Aubray, M.P., Hardenbol, J. (Eds.), Geochronology, Time Scales and Global Stratigraphy Correlation, Soc. Econ. Paleont. Mineral. Spec. Publ., vol. 54 (19), pp. 129–212.

Bijwaard, H., Spakman, W., 1999. Tomographic evidence for a narrow mantle plume below Iceland. Earth Planet. Sci. Lett. 166, 121–212.

Blès, J.-L., Gros, Y., 1991. Stress field changes in the Rhone Valley from the Miocene to the present. Tectonophysics 194, 265–277.

Boenigk, W., 2002. The Pleistocene drainage pattern in the Lower Rhine basin. Neth. J. Geosci. 81, 201–209.

Bonjer, K.-P., 1997. Seismicity pattern and style of seismic faulting at the eastern borderfault of the southern Rhine Graben. Tectonophysics 275, 41–69.

Brun, J.-P., Gutscher, M.-A., DEKORP-ECORS team, 1992. Deep crustal structure of the Rhine Graben from DEKORP-ECORS seismic reflection data. Tectonophysics 208, 39–147.

Brunnacker, K., Boenigk, W., 1983. The Rhine Valley between the Neuwied Basin and the Lower Rhenish Embayment. In: Fuchs, K., Von Gehlen, K., Mälzer, M., Murawski, H., Semmel, A. (Eds.), Plateau Uplift, The Rhenish Shield—A Case History. Springer-Verlag, Berlin, pp. 62–72.

Bucher, S., Schmid, S.M., Bousquet, R., Fügenschuh, B., 2003. Late-stage deformation in a collisional orogen (Western Alps): nappe refolding, back-thrusting or normal faulting? Terra Nova 15, 109–117.

Burkhard, M., Sommaruga, A., 1998. Evolution of the western Swiss Molasse basin: structural relations with the Alps and the Jura belt. In: Mascle, A., Puigdefàbregas, C., Luterbacher, H.P., Fernandez, M. (Eds.), Cenozoic Foreland Basins of Western Europe, Geol. Soc., London, Spec. Publ. vol. 134, pp. 279–298.

Butler, M., Pullan, C.R., 1990. Tertiary structures and hydrocarbon entrapment in the Weald Basin of southern England. In: Hardman, R.F.P., Brooks, J. (Eds.), Tectonic Events Responsible for Britains Oil and Gas Reserves, Geol. Soc., London, Spec. Publ., vol. 55, pp. 371–391.

Ceriani, S., Fügenschuh, B., Schmid, S.M., 2001. Late-stage thrusting at the "Penninic Front" in the Western Alps between Mont Blanc and Pelvoux massifs. Int. J. Earth Sci 90, 685–702.

Chauve, P., Martin, J., Petitjean, E., Sequeiros, F., 1988. Le chevauchement du Jura sur la Bresse. Données nouvelles et réinterpretation des sondages. Bull. Soc. Geol. Fr. 8 (IV), 861–870.

Coulon, M., 1992. La distension oligocène dans le nord-est du bassin de Paris (perturbation des directions d'extension et distribution des stylolites). Bull. Soc. Geol. Fr. 163, 531–540.

Damotte, B., Nicholich, R., Cazes, M., Guellec, S., 1990. Mise en oeuvre, traitement et presentation du profil plaine du Po-Massif Central. In: Roure, F., Heitzmann, P., Polino (Eds.), Deep Structure of the Alps, Mem. Soc. Geol. Fr., vol. 156, pp. 65–76.

Davies, J.H., von Blanckenburg, F., 1995. Slab breakoff: a model of lithosphere detachment and its test in the magmatism and deformation of collisional orogens. Earth Planet. Sci. Lett. 129, 85–102.

Deichmann, N., Ballarin Dolfin, D., Kastrup, U., 2000. Seismizität der Nord- und Zentralschweiz. NAGRA Technischer Bericht, Wettingen. 00-05. 93 pp.

Delouis, B., Haessler, H., Cisternas, A., Rivera, L., 1993. Stress tensor determination in France and neighbouring regions. Tectonophysics 211, 413–437.

De Lugt, I.R., van Wees, J.D., Wong, Th.E., 2003. The tectonic evolution of the southern Dutch North Sea during the Paleogene: basin inversion in distinct pulses. Tectonophysics 373, 141–159.

Dèzes, P., Ziegler, P.A., 2002. Moho depth map of Western and Central Europe. World Wide Web Address: http://www.unibas.ch/eucor-urgent.

Doglioni, C., Gueguen, E., Harabaglia, P., Mongelli, F., 1999. On the origin of west-directed subduction zones and applications to the western Mediterranean. In: Durand, B., Jolivet, L., Horvàth, F., Séranne, M. (Eds.), The Mediterranean Basins: Tertiary Extension within the Alpine Orogen, Geol. Soc., London, Spec. Publ., vol. 156, pp. 541–561.

Downes, H., 1993. The nature of the lower continental crust of Europe: petrological and geochemical evidence from xenoliths. Phys. Earth Planet. Inter. 76, 195–218.

Frisch, W., 1979. Tectonic progradation and plate tectonic evolution of the Alps. Tectonophysics 60, 121–139.

Fügenschuh, B., Schmid, S.M., 2003. Late stage of deformation and exhumation of an orogen constrained by fission-track data: a case study in the Western Alps. GSA Bull. 115, 1425–1440.

Garcia-Castellanos, D., Cloetingh, S., van Balen, R., 2000. Modelling the middle Pleistocene uplift in the Ardennes-Rhenish Massif: thermo-mechanical weakening under the Eifel. Global Planet. Changes 27, 39–52.

Geluk, M.C., Duin, E.J.T., Dusar, M., Rijkers, R.H.B., van den Berg, M.W., van Rooijen, P., 1994. Stratigraphy and tectonics of the Roer Valley Graben. Geol. Mijnb. 73, 129–141.

Giamboni, M., Ustaszewski, K., Schmid, S.M., Schumacher, M., Wetzel, A., 2004. Plio–Pleistocene deformation in the southern Upper Rhine Graben area: geological and geomorphological evidence for on-going transpressive reactivation of Paleozoic and Paleogene faults. Int. J. Earth Sci. 93, 207–223.

Giardini, D., Jménez, M.J., Grünthal, G., 2003. European–Mediterranean seismic hazard map 1:5'000'000. IGCP Project 382 SESAM. European Seismological Commission.

Giglia, G., Capponi, G., Crispini, L., Piazza, M., 1996. Dynamics and seismotectonics of the West-Alpine arc. Tectonophysics 267, 143–175.

Goes, S., Spakman, W., Bijwaard, H., 1999. A lower mantle source for Central European volcanism. Science 286, 1928–1930.

Goes, S., Loohuis, J.J.P., Wortel, M.I.R., Govers, R., 2000a. The effect of plate stresses and shallow mantle temperatures on the tectonics of northwestern Europe. Global Planet. Change 27, 23–38.

Goes, S., Govers, R., Vacher, P., 2000b. Shallow mantle temperatures under Europe from P and S wave tomography. J. Geophys. Res. 105, 11153–11169.

Gölke, M., Cloetingh, S., Coblentz, D., 1996. Finite-element modelling of stress patterns along the Mid-Norwegian continental margin, 62° to 68°N. Tectonophysics 266, 33–53.

Granet, M., Wilson, M., Achauer, U., 1995. Imaging mantle plumes beneath the French Massif Central. Earth Planet. Sci. Lett. 136, 199–203.

Guellec, S., Mugnier, J.-L., Tardy, M., Roure, F., 1990. Neogene evolution of the western Alpine foreland in the light of the ECORS data and balance cross-section. Mem. Soc. Geol. Fr. 156, 165–184.

Heumann, G., Litt, Th., 2002. Stratigraphy and paleoecology of the late Pliocene and early Pleistocene in the open-cast mine Hambach (Lower Rhine Basin). Neth. J. Geosci. 81, 193–199.

Hinzen, K.-G., 2003. Stress field in the Northern Rhine area, Central Europe, from earthquake fault plain solutions. Tectonophysics 377, 325–356.

Hoernle, K., Zhang, Yu.-S., Graham, D., 1995. Seismic and geochemical evidence for large-scale mantle upwelling beneath the eastern Atlantic and Western and Central Europe. Nature, London 374, 34–39.

Illies, J.H., Baumann, H., Hoffers, B., 1981. Stress pattern and strain release in the Alpine foreland. Tectonophysics 71, 157–172.

Jouanne, F., Ménard, G., Darmendrail, X., 1995. Present-day vertical displacements in the north-western Alps and southern Jura: data from levelling comparisons. Tectonics 14, 606–616.

Jouanne, F., Genaudeau, N., Ménard, G., Darmendrail, X., 1998. Estimating present-day displacement fields and tectonic deformation in active mountain belts: an example from the Chartreuse Massif and the southern Jura Mountains, Western Alps. Tectonophysics 296, 403–419.

Jung, S., 1999. The role of crustal contamination during evolution of continental rift-related basalts, a case study from the Vogelsberg area Central Germany. Geolines 9, 48–58.

Keller, J., Kramel, M., Henjes-Kunst, F., 2002. $^{40}Ar/^{39}Ar$ single crystal dating of early volcanism in the Upper Rhine Graben and tectonic implications. Schweiz. Mineral. Petrogr. Mitt. 82, 121–130.

Kemna, H.A., Becker-Haumann, R., 2003. Die Wanderblock-Bildung im Schweizer Juragebirge südlich von Basel: Neue Daten zu einem alten Problem. Eclogae Geol. Helv. 96, 71–83.

Klett, M., Eichhorst, F., Schäfer, A., 2002. Facies interpretation from well logs applied to the Tertiary Lower Rhine basin fill. Neth. J. Geosci. 81, 167–176.

Kooi, H., Cloetingh, S., Burrus, J., 1992. Lithospheric necking and regional isostasy at extensional basins: Part 1. Subsidence and gravity modelling with application to the Gulf of Lions margin (SE France). J. Geophys. Res. 97, 17553–17571.

Laubscher, H.P., 1986. The eastern Jura: relations between thin-skinned and basement tectonics, local and regional. Geol. Rundsch. 75, 535–553.

Laubscher, H., 1992. Jura kinematics and the Molasse Basin. Eclogae Geol. Helv. 85, 653–675.

Laubscher, H., 2003. The Miocene dislocations in the northern foreland of the Alps: oblique subduction and its consequences (Basel area Switzerland–Germany). Jahresber. Mitt. Oberrhein. Geol. Ver. N.F. 85, 423–439.

Lefort, J.P., Agarval, B.N.P., 1996. Gravity evidence for Alpine buckling of the crust beneath the Paris Basin. Tectonophysics 258, 1–14.

Lenôtre, N., Garnier, T., Delfau, M., 1992. Mouvements verticaux actuels dans le sud-est du Massif Central. 14e RST, Toulouse, 13–14 April, 1992. 2 pp. (abstract).

LeVot, M., Biteau, J.J., Masset, J.M., 1996. The Aquitaine Basin: oil and gas production in the foreland of the Pyrenean fold-and-thrust belt. New exploration perspectives. In: Ziegler, P.A., Horvàth, F. (Eds.), Structure and Prospects of Alpine Basins and Forelands. Peri-Tethys Mém. 2, Mem. Mus. Natl. Hist. Nat., Paris, vol. 170, pp. 159–171.

Lickorish, W.H., Ford, M., 1998. Sequential restoration of the external Alpine Digne thrust system, SE France, constrained by kinematic data and synorogenic sediments. In: Mascle, A., Puigdefàbregas, C., Luterbacher, H.P., Fernandez, M. (Eds.), Cenozoic Foreland Basins of Western Europe, Geol. Soc., London, Spec. Publ., vol. 134, pp. 189–211.

Lippitsch, R., 2003. Lithosphere and upper mantle P-wave velocity structure beneath the Alps by high-resolution teleseismic tomography. Ph.D. Thesis ETH14726.

Lippolt, H.J., 1983. Distribution of volcanic activity in space and time. In: Fuchs, K., Von Gehlen, K., Mälzer, M., Murawski, H., Semmel, A. (Eds.), Plateau Uplift, The Rhenish Shield—A Case History. Springer-Verlag, Berlin, pp. 112–120.

Malkovsky, M., 1987. The Mesozoic and Tertiary basins of the Bohemian Massif and their evolution. Tectonophysics 137, 31–42.

Mälzer, H., Hein, G., Zippelt, K., 1983. Height changes in the Rhenish Massif: determination and analysis. In: Fuchs, K., VonGehlen, K., Mälzer, M., Murawski, H., Semmel, A. (Eds.), Plateau Uplift, The Rhenish Shield—A Case History. Springer-Verlag, Berlin, pp. 165–176.

Mansy, J.-L., Manby, G.M., Averbuch, O., Everaerts, M., Bergerat, F., Van Vliet-Lanoe, B., Lamarche, J., Vandycke, S., 2003. Dynamics and inversion of the Mesozoic basin of the Weald-Boulonnais area: role of basement reactivation. Tectonophysics 373, 161–179.

Martinod, J., Jouanne, F., Taverna, J., Ménard, G., Gamond, J.-F., Darmendrail, X., Notter, J.-C., Basil, C., 1996. Present-day deformation of the Dauphiné Alpine and Subalpine massifs (SE France). Geophys. J. Int. 127, 189–200.

Maury, R.C., Varet, J., 1980. Le volcanisme tertiaire et quaternaire en France. In: Autrand, A., Dercourt, J. (Eds.), Evolution géologique de la France. Mém. B.R.G.M., vol. 107. Edition du BRGM, Orléans, pp. 137–189.

Mengel, K., 1992. Evidence from xenoliths for the composition of the lithosphere. In: Blundell, D., Freeman, R., Müller, St. (Eds.), A Continent Revealed, The European Geotraverse. Cambridge Univ. Press, pp. 91–102.

Merle, O., Michon, L., 2001. The formation of the West-European Rift: a new model exemplified by the Massif Central area. Bull. Soc. Geol. Fr. 172, 213–221.

Merle, O., Michon, L., Camus, G., de Goer, A., 1998. L'extension oligocène sur la transversal septentronale du rift du Massif central. Bull. Soc. Geol. Fr. 109, 615–626.

Michon, L., Merle, O., 2001. The evolution of the Massif Central rift: spatio-temporal distribution of the volcanism. Bull. Soc. Geol. Fr. 172, 201–211.

Michon, L., van Balen, R.T., Merle, O., Pagnier, H., 2003. The Cenozoic evolution of the Roer Valley rift system integrated at a European scale. Tectonophysics 367, 101–126.

Nikishin, A.M., Ziegler, P.A., Stephenson, R.A., Ustinova, M.A., 1999. Santonian to Palaeocene tectonics of the East-European craton and adjacent areas. Bull. Inst. Roy. Sci. Nat. Belg., Sci. Terre, 69-Supp. A, 147–159.

Mégnien, C., 1980. Synthèse Géologique du Bassin de Paris Mém. BRGM, 102.

Meyer, W., Stets, J., 2002. Pleistocene to Recent tectonics in the Rhenish Massif (Germany). Neth. J. Geosci. 81, 217–221.

Mugnier, J.-L., Guellec, S., Ménard, G., Roure, F., Tardy, M., Vialon, P., 1990. A crustal scale balanced cross-section through the external Alps deduced from ECORS profile. In: Roure, F., Heitzmann, P., Polino, R. (Eds.), Deep Structure of the Alps, Mem. Soc. Geol. Fr., vol. 156, pp. 203–216.

Müller, B., Wehrle, V., Zeyen, H., Fuchs, K., 1997. Short-scale variations of tectonic regimes in the Western European stress province north of the Alps and Pyrenees. Tectonophysics 275, 199–219.

Müller, W.H., Naef, H., Graf, H.R., 2002. Geologische Entwicklung der Nordwestschweiz, Neotektonik und Langzeitszenarien Zürcher Weinland. NAGRA Technischer Bericht, Wettingen. 99-08, 237 pp.

Oncken, O., Plesch, A., Weber, J., Ricken, W., Schrader, S., 2000. Passive margin detachment during arc-continent collision (Central European Variscides). In: Franke, W., Haak, V., Oncken, O., Tanner, D. (Eds.), Orogenic Processes: Quantification and Modelling in the Variscan Belt, Geol. Soc., London, Spec. Publ., vol. 179, pp. 199–216.

Pfiffner, O.A., Lehner, P., Heitzmann, P., Mueller, S., Steck, A. (Eds.), 1997. Deep Structure of the Alps: Results from NFP 20. Birkhäusert, Basel. 380 pp.

Philippe, Y., Colletta, B., Mascle, E., 1996. The Jura fold-and-thrust belt: a kinematic model based on map-balancing. In: Ziegler, P.A, Horvàth, F. (Eds.), Structure and Prospects of Alpine Basins and Forelands. Peri-Tethys Mem. 2, Mem. Mus. Natl. Hist. Nat., Paris, vol. 170, pp. 235–261.

Philippe, Y., Deville, E., Mascle, A., 1998. Thin-skinned inversion tectonics at oblique basin margins: example of the western Vercors and Chartreuse subalpine massifs (SE France). In: Mascle, A., Puigdefàbregas, C., Luterbacher, H.P., Fernandez, M. (Eds.), Cenozoic Foreland Basins of Western Europe, Geol. Soc., London, Spec. Publ., vol. 134, pp. 239–262.

Plenefisch, T., Bonjer, K.-P., 1997. The stress field in the Rhine Graben area inferred from earthquake focal mechanisms and estimations of frictional parameters. Tectonophysics 275, 71–97.

Prodehl, C., Mueller, St., Haak, V., 1995. The European Cenozoic Rift System. In: Olsen, K.H. (Ed.), Continental Rifts: Evolution, Structure, Tectonics, Developments in Geotectonics, vol. 25. Elsevier, Amsterdam, pp. 133–212.

Rat, P., 1984. Une approche de l'environnement structural et morphologique du Pliocène et du Quaternaire bressans. Geol. Fr. 3, 185–195.

Ritter, J.R.R., Jordan, M., Christensen, U.R., Achauer, U., 2001. A mantle plume below the Eifel volcanic fields, Germany. Earth Planet. Sci. Lett. 186, 7–14.

Roca, E., 2001. The Northwest Mediterranean Basin (Valencia Trough, Gulf of Lions, and Liguro-Provençal basins): structure and geodynamic evolution. In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), Peri-Tethys Memoir 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins, Mem. Mus. Natl. Hist. Nat., Paris, vol. 186, pp. 617–706.

Rocher, M., Chevalier, F., Petit, C., Guiraud, M., 2003. Tectonics of the Northern Bresse region (France) during the Alpine cycle. Geodin. Acta 16, 131–147.

Roll, A., 1979. Versuch einer Volumenbilanz des Oberrheingrabens und seiner Schultern.. Geol. Jahrb., Reihe A Heft 52 (82 pp.).

Rosenbaum, G., Lister, G.S., Duboz, C., 2002. Relative motion of Africa, Iberia and Europe during Alpine orogeny. Tectonophysics 359, 117–129.

Roure, F., Colletta, B., 1996. Cenozoic inversion structures in the foreland of the Pyrenees and Alps. In: Ziegler, P.A., Horvàth, F. (Eds.), Structure and Prospects of Alpine Basins and Forelands. Peri-Tethys Mem. 2, Mem. Mus. Natl. Hist. Nat., Paris, vol. 170, pp. 173–209.

Roure, F., Brun, J.-P., Colletta, B., Vially, R., 1994. Multiphase extensional structures, fault reactivation, and petroleum plays in the Alpine foreland Basin of southeastern France. In: Mascle, A. (Ed.), Hydrocarbon and Petroleum Geology of France, Europ. Assoc. Petrol. Geosci., Spec. Publ., vol. 4. Springer-Verlag, Berlin, Heidelberg, New York, pp. 245–268.

Sanchis, E., Séranne, M., 2000. Structural style and tectonic evolution of a polyphase extensional basin of the Gulf of Lion passive margin: the Tertiary Alès basin, southern France. Tectonophysics 322, 219–242.

Schmid, S.M., 1993. Ivrea Zone and adjacent South Alpine basement. In: von Raumer, J.F., Neubauer, F. (Eds.), Pre-Mesozoic Geology of the Alps. Springer-Verlag, Berlin, pp. 567–583.

Schmid, S.M., Kissling, E., 2000. The arc of the Western Alps in the light of geophysical data on deep crustal structure. Tectonics 19, 62–85.

Schmid, S.M., Pfiffner, O.A., Froitzheim, N., Schönborn, G., Kissling, E., 1996. Geophysical–geological transect and tectonic evolution of the Swiss–Italian Alps. Tectonics 12, 1036–1064.

Schmid, S.M., Fügenschuh, B., Kissling, E., Schuster, R., 2004. Tectonic map and overall architecture of the Alpine orogen. Eclogae Geol. Helv. 97, 93–117.

Schumacher, M.E., 2002. Upper Rhine Graben: the role of pre-existing structures during rift evolution. Tectonics 21, 10.1029/2001TC900022 (6-1 - 6-17).

Séranne, M., 1999. The Gulf of Lion continental margin (NW Mediterranean) revisited by IBS: an overview. In: Durand, B., Jolivet, L., Horvàth, F., Séranne, M. (Eds.), The Mediterranean Basins: Tertiary Extension within the Alpine Orogen, Geol. Soc., London, Spec. Publ., vol. 156, pp. 15–36.

Sissingh, W., 1998. Comparative Tertiary stratigraphy of the Rhine Graben, Bresse Graben and Molasse Basin: correlation of Alpine foreland events. Tectonophysics 300, 249–284.

Sissingh, W., 2001. Tectonostratigraphy of the West Alpine Foreland: correlation of Tertiary sedimentary sequences, changes in eustatic sea-level and stress regimes. Tectonophysics 333, 361–400.

Sissingh, W., 2003a. Stratigraphic framework of the European Cenozoic Rift System: a visual overview. Geol. Ultraiectina, University of Utrecht, Spec. Publ. 2 (atlas).

Sissingh, W., 2003b. Tertiary paleogeogrephic and tectonostratigraphic evolution of the Rhenish triple junction. Palaeogeogr. Palaeoclimatol. Palaeoecol. 196, 229–263.

Sobolev, S.V., Zeyen, H., Granet, M., Achauer, U., Bauer, C., Werling, F., Altherr, R., Fuchs, K., 1997. Upper mantle temperatures and lithosphere–asthenosphere system beneath the French Massif Central constrained by seismic, gravity, petrologic and thermal observations. Tectonophysics 275, 143–164.

Stampfli, G.M., Mosar, J., Marquer, D., Marchant, R., Baudin, T., Borel, G., 1998. Subduction and obduction processes in the Alps. Tectonophysics 296, 159–204.

Stampfli, G.M., Mosar, J., Favre, P., Pillevuit, A., Vannay, J.-C., 2001. Permo-Mesozoic evolution of the western Tethys realm: the Neo-Tethys East-Mediterranean connection. In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), Peri-Tethys Mem. 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins, Mem. Mus. Natl. Hist. Nat., vol. 186, pp. 51–108.

Stampfli, G.M., Borel, G.D., Marchant, R., Mosar, J., 2002. Western Alps geological constraints on western Tethyan reconstructions. J. Virtual Explor. 8, 77–106.

Steck, A., 1984. Structures de déformation tertiaires dans les Alpes centrales (transversale Aar-Simplon-Ossola). Eclogae Geol. Helv. 77, 55–100.

Steck, A., 1990. Une carte des zones de cisaillement ductile des Alpes Centrales. Eclogae Geol. Helv. 83 (3), 603–627.

Sue, C., Tricart, P., 2002. Widespread post-nappe normal faulting in the Internal Western Alps: new constraints on arc dynamics. J. Geol. Soc. (Lond.) 159, 61–70.

Sue, C., Martinod, J., Tricart, P., Thouvenot, F., Gamond, J.-F., Fréchet, J., Marinier, D., Glot, J.-P., Grasso, J.-P., 2000. Active deformation in the inner Western Alps inferred from comparison between 1972-classical and 1996-GPS geodetic surveys. Tectonophysics 320, 17–29.

Ulrych, J., Pivec, E., Lang, M., Balogh, K., Kropacek, V., 1999. Cenozoic intraplate volcanic rocks series of the Bohemian Massif: a review. Geolines, Prague 9, 123–129.

Van Vliet-Lanoé, B., Laurent, M., Halléguet, B., Margerel, J.-P., Chauve, J.-J., Michel, Y., Moguedet, G., Trautman, F., Vautier, S., 1998. Le Mio–Pliocène du Massif armoricain. Données nouvelles. C. R. Acad. Sci. Paris, Earth Planet. Sci. 326, 333–340.

Vergés, J., García-Senez, J., 2001. Mesozoic evolution and Cainozoic inversion of the Pyrenean rift. In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), Peri-Tethys Mem. 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins, Mem. Mus. Natl. Hist. Nat., vol. 186, pp. 187–212.

Villemin, T., Bergerat, F., 1987. Tectonique cassante et paléocontraintes tertiaires de la bordure nord-est du fossé rhénan. Oberrhein. Geol. Abh. 14, 63–87.

Villemin, T., Alvarez, F., Angelier, J., 1986. The Rhine-graben, extension, subsidence and shoulder uplift. Tectonophysics 128, 47–59.

Von Blanckenburg, F., Davies, J.H., 1995. Slab breakoff: a model for syncollisional magmatism and tectonics in the Alps. Tectonics 14, 120–131.

Wilson, M., Bianchini, G., 1999. Tertiary–Quaternary magmatism within the Mediterranean and surrounding regions. In: Durand, B., Jolivet, L., Horvàth, F., Séranne, F. (Eds.), The Mediterranean Basin: Tertiary Extension within the Alpine Orogen, Geol. Soc., London, Spec. Publ., vol. 156, pp. 141–168.

Wilson, M., Downes, H., 1992. Mafic alkaline magmatism associated with the European Cenozoic Rift System. Tectonophysics 208, 172–182.

Wilson, M., Rosenbaum, J.M., Dunworth, E.A., 1995. Melilites: partial melts of the thermal boundary layer? J. Petrol. 32, 181–196.

Wittenberg, A., Vellmer, C., Kern, H., Mengel, K., 2000. The Variscan lower continental crust: evidence for crustal delamination from geochemical and petrological investigations. In: Franke, W., Haak, V., Oncken, O., Tanner, D. (Eds.), Orogenic Processes: Quantification and Modelling in the Variscan Belt, Geol. Soc., London, Spec. Publ., vol. 179, pp. 401–414.

Ziegler, P.A., 1988. Evolution of the Arctic–North Atlantic and the Western Tethys. Am. Assoc. Petrol. Geol. Mem. 43. 198 p.

Ziegler, P.A., 1990. Geological Atlas of Western and Central Europe, Shell Internat. Petrol. Mij., Dist. Geol. Soc. Publ. House, Bath, 2nd ed. 239 pp. and 56 encl.

Ziegler, P.A., 1994. Cenozoic rift system of Western and Central Europe: an overview. Geol. Mijnb. 73, 99–127.

Ziegler, P.A. (Ed.), 1997. Compressional intra-plate deformation in the Alpine foreland. Tectonophysics, vol. 137, 420 pp.

Ziegler, P.A., Cloetingh, S., 2004. Dynamic processes controlling evolution of rifted basins. Earth-Sci. Rev. 64, 1–45.

Ziegler, P.A., Dèzes, P., 2005. Crustal evolution of Western and Central Europe. In: Gee, D., Stephenson, R. (Eds.), European Lithosphere Dynamics, Mem. Geol Soc. Lond. (accepted for publication).

Ziegler, P.A., Cloetingh, S., vanWees, J.-D., 1995. Dynamics of intra-plate compressional deformation: the Alpine foreland and other examples. Tectonophysics 252, 7–59.

Ziegler, P.A., vanWees, J.D., Cloetingh, S., 1998. Mechanical controls on collision-related compressional intraplate deformation. Tectonophysics 300, 103–129.

Ziegler, P.A., Cloetingh, S., Guiraud, R., Stampfli, G.M., 2001. Peri-Tethyan platforms: constraints on dynamics of rifting and basin inversion. In: Ziegler, P.A., Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), Peri-Tethys Mem. 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins, Mem. Mus. Natl. Hist. Nat., vol. 186, pp. 9–49.

Ziegler, P.A., Bertotti, G., Cloetingh, S., 2002. Dynamic processes controlling foreland development: the role of mechanical (de)coupling of orogenic wedges and forelands. In: Bertotti, G., Schulmann, K., Cloetingh, S. (Eds.), Continental Collision and the Tectono-Sedimentary Evolution of Forelands, Europ. Geophys. Soc. Stephan Mueller Spec. Publ., vol. 1, pp. 29–91.

Ziegler, P.A., Schumacher, M.E., Dèzes, P., vanWees, J.-D., Cloetingh, S., 2004. Post-Variscan evolution of the lithosphere in the Rhine Graben area: constraints from subsidence modelling. In: Wilson, M., Neumann, E.-R., Davies, G.R., Timmerman, M.J., Heeremans, M., Larsen, B.T. (Eds.), Permo–Carboniferous Magmatism and Rifting in Europe, Geol. Soc., London, Spec. Publ., vol. 223, pp. 289–317.

Zielhuis, A., Nolet, G., 1994. Deep seismic expression of an ancient plate boundary in Europe. Science 265, 79–81.

Zijerveld, L., Stephenson, R., Cloetingh, S., Duin, E., van den Berg, M.W., 1992. Subsidence analysis and modelling of the Roer Valley Graben (SE Netherlands). Tectonophysics 208, 159–171.