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Discussion

Reply to comments by L. Michon and O. Merle on "Evolution of the European Cenozoic Rift System: interaction of the Alpine and Pyrenean orogens with their foreland lithosphere" by P. Dèzes, S.M. Schmid and P.A. Ziegler, Tectonophysics 389 (2004) 1–33☆

Pierre Dèzes*, Stefan M. Schmid, Peter A. Ziegler

University of Basel, Department of Earth Sciences Bernoullistr. 32 CH-4056 Basel, Switzerland

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Processes controlling the development and evolution of the European Cenozoic Rift System (ECRIS) have been debated since the early models of Cloos (1936) and Illies (1973, 1975). Since then, a variety of hypotheses have been advanced involving processes such as plume-related active rifting (Neugebauer, 1978), collisional foreland splitting (Sengör, 1976; Tapponnier, 1977; Sengör et al., 1978; Bergerat and Geyssant, 1980; Dewey and Windley, 1988), back-arc rifting (Jowett, 1991) or slab-pull (Stampfli et al., 1998; Merle and Michon, 2001; Michon et al., 2003).

Based on regional compilations that were carried out in the context of the Peri-Tethys (Stampfli et al., 2001, Ziegler et al., 2001), TRANSMED (Cavazza et al., 2004) and EUCOR-URGENT projects, we re-evaluated the viability of the different models that were proposed for the evolution of ECRIS and came to the conclusion that the onshore parts of ECRIS evolved by passive rifting in response to collisional foreland splitting. By contrast, Merle and Michon (2001), Michon et al. (2003) and Michon and Merle (this issue) advocate a slab-pull model.

Whereas geophysical transects across the Alps, supported by tomographic images of their lithospheric configuration (Schmid et al., 2004a), permit to reject a back-arc rift model for the evolution of the on-shore parts of ECRIS, such a model clearly applies to the development of the Gulf of Lyons–Valencia Trough part of ECRIS (Séranne, 1999; Roca, 2001; Roca et al., 2004).

On the other hand, the postulate that slab-pull contributed to the development of the on-shore parts of ECRIS requires a careful review of the evolution of the Western and Central Alps and particularly of the timing of the slab detachment-related Periadriatic magmatic activity. The Periadriatic intrusions, which are closely associated with the suture between the European lower and the Adriatic upper plate, form a

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* Corresponding author.
E-mail address: pierre.dezes@unibas.ch (P. Dèzes).

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discontinuous chain that extends over a distance of some 650 km from Traversella north of Torino (Italy) to the Pohorje Mountains in Slovenia (Von Blanckenburg and Davies, 1995; Schmid et al., 2004b). Whereas intrusive activity commenced in the Adamello Massif around 42 Ma, all other Periadriatic plutons were emplaced during the latest Eocene and early Oligocene (34–28 Ma), coeval with the activation of the dextral transpressional Periadriatic Fault System (Rosenberg, 2004; Stipp et al., 2004). Von Blanckenburg and Davies (1995) assumed that detachment of the subducted south-dipping Central and East Alpine slab, consisting of oceanic Penninic and Valais and continental Briançonnais lithosphere, occurred around 45 Ma simultaneously along the entire Periadriatic lineament. However, detachment of this slab, accompanied by decompression and cooling of HP metamorphic rocks, may have commenced during the middle Eocene (>42 Ma) in the Adamello domain only, from where it presumably propagated west- and eastward, becoming effective in the adjacent Central and Eastern Alps in the course of the late Eocene (>34 Ma) but no later than at the Eocene–Oligocene transition. At about the same time, also the northeast-dipping subduction slab of the Dinarides was detached from the lithosphere (Pamic et al., 2002) (Fig. 1). With the detachment of the subduction slab of the Central and Eastern Alps, slab-pull forces exerted onto the European lower plate relaxed. During the Oligocene to Pliocene northwest-directed convergence of Apulia with Europe, a new, some 180 km long south-dipping slab was inserted into the mantle in the area of the Central Alps. This tomographically imaged slab, which consists of

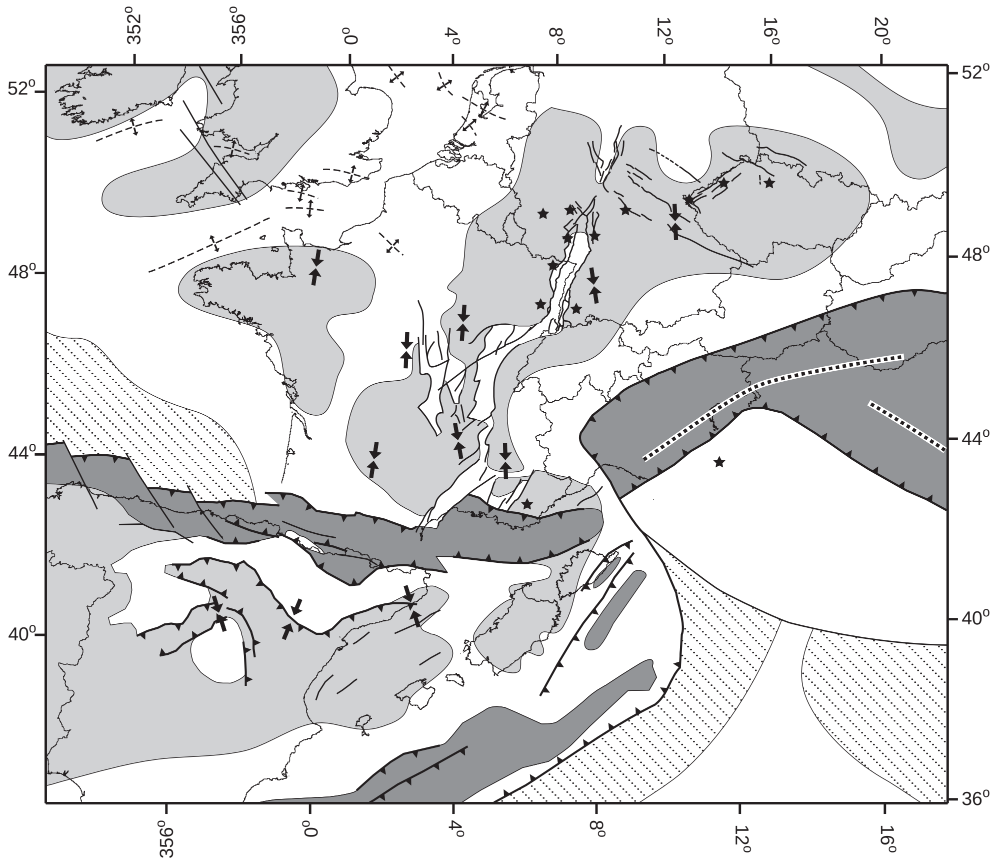

Fig. 1. Late Eocene palaeotectonic sketch map of ECRIS area showing trace of slab detachment along the Alpine and Dinarides Periadriatic lineaments (modified after Dèzes et al., 2004). Legend: dark grey: orogens, light grey: areas of non-deposition, white: sedimentary basins, stippled: oceanic basins, stars: volcanism, arrows: maximum horizontal compressional stress direction, dotted lies: trace of slab detachment.

European continental mantle-lithosphere and lower crust, hardly exerted slab-pull forces onto the European lower plate, owing to its limited length and relatively gentle dip and low density (Schmid et al., 2004a). Consequently, it is unlikely that slab-pull forces exerted by this secondary Central Alpine subduction slab contributed toward the Oligocene to recent evolution of ECRIS.

Whereas our transect B crosses the Penninic suture of the Central Alps in the area of the early Oligocene Bergell pluton, transect A crosses the Penninic suture of the Western Alps some 15 km to the southwest of the early Oligocene Traversella intrusion, the westernmost Periadriatic intrusion. Unlike for the Central Alps, we postulate that the West Alpine subduction slab remained attached to the lithosphere and was not detached from it before the Mio–Pliocene transition. This late detachment is indicated by the mantle tomographic image of the Western Alps that shows a narrow slab window at a depth of about 110 km separating the Alpine lithosphere from a deep reaching subducted slab (Schmid et al., 2004a). In view of Eocene to Miocene oblique subduction and a progressively increasing curved geometry of the West Alpine subduction slab (see Fig. 5 of Dèzes et al., 2004), possibly involving its disruption, we doubt that its slab-pull forces were sufficiently large to control the evolution of the Rhône Valley and Massif Central grabens.

Consequently, and as apparent from Fig. 1, we cannot support the slab-pull model of Michon and Merle (2001) and Michon et al. (2003). Therefore, we further explored the collisional foreland-splitting model by combining the structural and stratigraphic record of ECRIS with the evolution of the Alps (Schmid et al., 2004a,b), Pyrenees and Iberia (Roca et al., 2004; Andeweg, 2002).

We are fully aware of the latest Cretaceous and Paleocene pulse of intraplate compression that affected the Alpine and Pyrenean foreland. We have indeed discussed it, referring to Ziegler (1990), Ziegler et al. (1995, 1998, 2002) and De Lugt et al. (2003). In the ECRIS area, this deformation phase involved, apart from the compressional/transpressional reactivation of pre-existing crustal discontinuities (basin inversion, up-thrusting of basement blocks), also lithospheric buckling. Fission-track data indicate, however, for the Vosges and Black Forest, only a weak Late Cretaceous–Paleocene cooling trend (Link et al., 2004; B. Fügenschuh, personal communication 2004), and for the Rhenish Massif, that its Cretaceous uplift and cooling (120–80 Ma) ended during the late Senonian (Glasmacher et al., 1998). Whereas the Massif Central was only gently warped up during the Paleocene development of its regolithic (Terres Rouges, Sidérolithique) cover (Séranne et al., 2002), Senonian and Paleocene uplift of the Bohemian Massif was controlled by up-thrusting of an array of major basement blocks (Ziegler, 1990). Our concept of a Paleocene plume-related temperature increase of the asthenosphere in the ECRIS area is compatible with (a) the intrusion of very low degree partial melts in the Massif Central, Vosges-Black Forest, Rhenish Massif and Bohemian Massif which were derived from the lithospheric thermal boundary layer, (b) the indicated low amplitude of Paleocene lithospheric folds, and (c) the latest Cretaceous and Paleocene activation of the NE Atlantic and Iceland plumes (Ziegler, 1990). Although we agree with Michon and Merle (this issue) that palaeo-stress analyses are unable to discriminate between Paleocene and Eocene stress axes, we adopted the data published by Bergerat (1987), Villemin and Bergerat (1987), Andeweg (2002) and others and integrated them in our palaeo–tectonic maps which retrace the evolution of ECRIS and the Alps and Pyrenees. As during the Paleocene much of the ECRIS area was located above the erosional base-level, it is difficult to determine whether, apart from the fault systems of the Bohemian Massif, the Lower Saxony and West Netherlands basins and the Hunsrück border zone, other faults in the ECRIS area were also reactivated during this phase of intense intraplate compression (Fig. 5A of Dèzes et al., 2004). Certainly, there is no stratigraphic evidence for Paleocene graben development.

We re-emphasize that stress directions given in our palaeotectonic maps (Fig. 5 of Dèzes et al., 2004) correspond to the maximum horizontal compressional stress axes, which are not necessarily identical with the sigma-1 stress axes. This is particularly the case during the Oligocene main extension of the ECRIS areas, where sigma-1 was vertical. Although magmatic activity in the ECRIS area was at a low level during the Eocene, it nevertheless is in evidence and increased during the Oligocene particularly on the Rhenish and Bohemian Massifs. During these times, much of the ECRIS area remained above sea level except for its graben systems that began to subside differentially during the late Eocene. We take this as an indication for progressive, albeit slow thermal weakening and thinning of the lithosphere, thus rendering it prone to deformation. During the late Eocene, and particularly during the Oligocene, consistently northerly-directed collision-related compressional stresses built up again in the foreland of the Alps and Pyrenees. These stresses apparently exceeded the frictional strength of suitably oriented pre-existing Late Palaeozoic and Mesozoic crustal– and perhaps also lithosphere–scale faults, causing their tensional/transtensional reactivation, and with this, activation of ECRIS. Extensional strain across ECRIS was compensated by westward escape of the Paris Basin block, involving reactivation of the Armorican shear zone and inversion of the tensional Mesozoic Western Approaches, Channel and Weald basins, representing lithospheric weakness zone. We agree with the comments by Michon and Merle (this issue), that accelerated plume-related thermal thinning of the mantle lithosphere started to control uplift of the Rhenish Massif and Massif Central only from the latest Oligocene and the late Miocene onward, respectively, and thus did not materially contribute to the late Eocene and Oligocene tensional subsidence of ECRIS.

Dèzes et al. (2004) have demonstrated that from 18 Ma onward lithospheric folding controlled uplift of the Vosges-Black Forest arch and its continuation towards the Massif Central and Bohemian Massif. As the amplitude of the Neogene Vosges-Black Forest arch is of the order of 2.5 km, its uplift apparently caused decompressional partial melting of the lithospheric thermal boundary layer and the upper asthenosphere, the latter being characterized by an above ambient temperature, as evidenced by a magmatic surge spanning 18–7 Ma. That in the area of the Vosges-Black Forest arch the amplitude of Mio–Pliocene lithospheric folding was indeed considerably larger than during the Late Cretaceous–Paleocene is documented by fission-track data that indicate accelerated Miocene cooling (Link et al., 2004).

Development of the Vosges-Black Forest arch, starting around 18 Ma, coincided with the onset of imbrication and gradual uplift of the crystalline Aare, Mont Blanc, Beldonne and Pelvoux external massifs, which accelerated after 10 Ma, as evidenced by fission-track data (Fügenschuh and Schmid, 2003). This indicates that during the early Miocene, collisional coupling between the Alpine orogenic wedge and the foreland had increased at mantle-lithospheric and crustal levels. Combined with the end-Oligocene decay of compressional stresses originating at the Pyrenean collision zone and the onset of sea-floor spreading in the Provençal Basin (21.5 Ma), this explains the early Miocene reorientation of the stress field in the Alpine foreland under which the grabens of the Rhône Valley and Massif Central became inactive whilst the Upper Rhine and Roer Valley Grabens remained active.

Sue and Tricart (2002) suggest that in the internal parts of the Western Alps extension commenced before the end of the Oligocene and persisted to the present, while shortening continued in the frontal parts of the orogen; they attributed this extension to the development of a deep crustal indenter (ramp anticline) or to buoyantly rising crust. Whereas during the Oligocene nappe stacking and back-folding of the orogen, its internal parts were uplifted and started to cool, early Miocene development of a lower crustal ramp-flat structure in the European lower plate, followed and accompanied by imbrication of the external massifs, caused exhumation and extension of the overlying nappe stack. By contrast, tensional reactivation of the frontal Zone Houillière (Briançonnais) thrust and overall uplift of the Western Alps, starting around 5 Ma (Fügenschuh and Schmid, 2003), has been attributed by Schmid et al. (2004a) to the detachment of the subducted slab from the lithosphere.

Mantle tomography images beneath ECRIS a system of upper asthenospheric low velocity anomalies, interpreted as plume heads which have spread out above the 410 km discontinuity (Goes et al., 1999; Spakman, 2004; Sibuet et al., 2004), and from which secondary, relatively weak plumes presently rise up beneath the Eifel (Ritter et al., 2001) and Massif Central (Granet et al., 1995). Presumably these deep-seated anomalies started to develop during the Paleocene and subsequently evolved further, as evidenced by persisting volcanic activity, particularly on the Rhenish and Bohemian Massifs. As in time a shift in areas of main volcanic activity can be observed (e.g. decreasing on the Bohemian Massif towards the end of the Oligocene, increasing on the Rhenish Massif during the early Miocene and Pliocene and on the Massif Central during the late Miocene and Pliocene), it is likely that the supply of partial melts through secondary upper mantle plumes was not steady state but pulsating and entailed a shift in their location.

We are firmly of the opinion that development of the on-shore parts of ECRIS was governed by collision-related compressional stresses rather than by slab-pull forces. By contrast, slab roll-back controlled the evolution of the offshore parts of ECRIS. In the ECRIS area, mantle plume activity, starting in the Paleocene, caused progressive thermal weakening and thinning of the lithosphere, thus rendering it prone deformation, but was not a driving mechanism of rifting.

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