Crustal Evolution of Western and Central Europe

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Abstract:

A new Moho depth map has been assembled for Western and Central Europe and the West-Mediterranean area that is exclusively based on published regional Moho depth maps. Tectonic overlays summarize Caledonian and Variscan tectonic units, Permo-Carboniferous fault systems and magmatic provinces, Mesozoic and Cenozoic rift-wrench systems, areas of intraplate compression, the outlines of Alpine orogens and the distribution of oceanic crust. Based on a comparison of these overlays with the Moho depth map we assess processes that controlled the evolution of the crust in the different parts of Europe through time.
The present-day crustal configuration of Western and Central Europe results from polyphase Late Palaeozoic to recent lithospheric deformation that overprinted the margin of the Proterozoic East-European Craton and particularly the Caledonian and Variscan crustal domains. Following consolidation of the Caledonides, their crustal roots were destroyed in conjunction with Devonian wrench tectonics and back-arc rifting. During the Permo-Carboniferous tectono-magmatic cycle, wrench faulting disrupted the crust of the Variscan Orogen and its foreland and their lithosphere was thermally destabilized. Late Permian and Mesozoic re-equilibration of the lithosphere-asthenosphere system was interrupted by the development of the Arctic-North Atlantic, Tethyan and associated rift systems. During the Alpine orogenic cycle, intraplate compressional stresses controlled basin inversion-related crustal thickening and lithospheric folding, as well as the evolution of the Rhine-Rhône rift system. Variably deep crustal roots characterize the Alpine orogenic chains. Neogene back-arc extension disrupted the eastern Pyrenees, Betic-Balearic, Apennine and Dinarides orogens.

Submitted to
European Lithosphere Dynamic
D.G. Gee & R.A. Stephenson (eds.)
Memoir of the Geological Society, London
11’883 words

“Crustal Evolution of Europe”

Introduction:

The depth-map of the Moho discontinuity presented in figure 1 was constructed by digitally scanning, scaling and assembling published regional Moho maps (see compilation reference list and Dèzes & Ziegler, 2002) and by redrawing the depth contours as vectorized polygons. Most of the maps used for this compilation became available after the publication of earlier Moho compilation maps covering large parts of Europe by Meissner et al. (1987), Ziegler (1990) and Ansorge et al. (1992). The map resulting from our efforts (Fig. 1) gives the depth of the Moho discontinuity for Western and Central Europe and adjacent oceanic domains, but not the true thickness of the crust, as no corrections were applied for the thickness of its sedimentary cover nor for surface topography in elevated areas or for water depths in offshore areas.

The objective of this compilation was to obtain an impression of the present-day crustal configuration of entire Western and Central Europe, including the West-Mediterranean area, and to develop a base for the analysis of processes, which through time, contributed towards the evolution of the crust in the different parts of Europe. To this end, a set of overlays was constructed which summarize the main tectonic elements of the Caledonian and Variscan orogens (Fig. 2), the Stephanian-Early Permian fault systems and magmatic provinces (Fig. 3), the Mesozoic rift and wrench systems (Fig. 4), areas of intraplate compression, the outlines of the Alpine orogens and Cenozoic rift and wrench systems (Fig. 5). In Figures 2 to 5, that also show the present-day distribution of oceanic crust in the Atlantic and Mediterranean domains, these overlays are reproduced together with the Moho depth contour map.

A comparison of Figures 1 and 2 clearly shows that the stable parts of the Proterozoic Fennoscandian - East-European Craton are characterized by Moho depths of up to 48 km (within the frame of our map) whilst in more mobile Phanerozoic Europe Moho depths vary between 24 and 38 km and no longer bear any relation to the Caledonian and Variscan orogens. On the other hand, Moho depths of 20-26 km characterize the Proterozoic Hebridean craton and reflect strong modification of this crustal domain during Mesozoic rifting cycles. By contrast, the Alpine chains, such as the Western and Central Alps, the Carpathians, Apennine and Dinarides, as well as the Betic Cordillera and the Pyrenees are characterized by more or less distinct crustal roots reaching to depths of up to 60 km. Inferring an Alpine crustal model (Stampfli et al., 1998) for the continent-to-continent collisional Caledonian and Variscan orogens, the present crustal configuration of extra-Alpine Phanerozoic Europe implies post-orogenic destruction of their crustal roots and that their crust was repeatedly modified during Mesozoic and Cenozoic phases of rifting and intraplate compression.

In this context it should be kept in mind that the present depth of the Moho discontinuity is not only controlled by the thickness and composition of the crust but also by the thickness of its sedimentary cover, and in off-shore areas, by water depths. So far, we have not yet been able to construct a regional thickness map of the crystalline continental crust for the extra-Alpine domains, as in areas north of the Varsican deformation front the thickness of pre-Permian Palaeozoic sediments is still poorly constrained.

Processes controlling depth of the crust-mantle boundary

During orogenic processes, the Moho discontinuity can be depressed to depths of 60-75 km in conjunction with subduction of continental crust, its imbrication and the stacking of basement-cored nappes, as evident e.g. in the Western and Central Alps and the Pyrenees (Roure et al., 1996; Waldhauser et al., 1998; Schmid et al, 1996, 2004). During nappe emplacement, the foreland lithosphere is deflected in response to its thrust- and slab-loading, accounting for the development of foreland basins and a corresponding depression of the crust-mantle boundary (Ziegler et al., 2002). Depending on convergence rates and the thermal state of the foreland lithosphere, underthrusted continental crust is eclogitized at depths of 55-75 km and assumes densities in the range of ρ = 3.06-3.56, comparable to those of the average mantle (ρ = 3.35; Bousquet et al., 1997; Henry et al., 1997). In the process of this, the P-wave velocity of crustal material increases, depending on its composition, to 8.0-8.4 km^{-sec}, and thus, is transferred across the geophysically defined Moho discontinuity (break-over from $Vp \leq 7.8$ to 8.0-8.2 km^{-sec}) into the lithospheric mantle (Ziegler et al., 1998). This limits depth of orogenic crustal roots as defined by seismic velocities. In this respect, it should be kept in mind that the petrologic and seismic crust-mantle boundary does not always coincide.

Synorogenic thickening of the crust, involving subduction of continental lithospheric mantle and lower crustal material and stacking of upper crustal nappes (see e.g. Schmid et al., 2004; Dèzes et al., 2004), is accompanied by widespread metamorphism of crustal and sedimentary rocks, their metasomatic reactivation, and plutonic activity, as seen in the Variscan and Caledonian orogens. During post-orogenic times, erosional and tectonic unroofing of an orogen, often to former mid-crustal levels, results in the exposure of a newly formed crystalline basement complex.

The post-orogenic re-equilibration of orogenically destabilized lithosphere with the asthenosphere can involve such processes as detachment of subducted lithospheric slabs and passive upwelling of the asthenosphere, thermal thinning and/or partial delamination of the mantle-lithosphere, interaction of mantle-derived melts with a felsic lower crust, and possibly also retrograde metamorphism of eclogitized crustal roots. All these processes contribute towards uplift and erosional unroofing of an orogen, and a corresponding shallowing of the crust-mantle boundary. Moreover, the post-orogenic collapse of an orogen can be accelerated by its wrench- and/or extension-induced tectonic unroofing in response to a re-orientation of the regional stress field (e.g. Devonian collapse of the Arctic-North Atlantic Caledonides, Permo-Carboniferous collapse of the Variscides). Under such conditions, the crust of an orogen can be thinned to 30-35 km within a few 10-20 My after crustal shortening has terminated (e.g. Variscides). Furthermore, depending on the degree of post-orogenic thinning of the lithospheric mantle and the thickness of the crust, long-term thermal subsidence of the lithosphere of former orogens can account for the subsidence of intracratonic basins, involving a gradual depression of the crust-mantle boundary (Ziegler et al., 2004, this volume).

During rifting and wrench faulting the continental crust is thinned by mechanical stretching (McKenzie, 1978). The magnitude and mode of lithospheric stretching (pure- or simple-shear), the depth of the lithospheric necking level, whether or not magmatic processes contributed toward crustal thinning or crustal thickening by underplating, and the degree of thermal thinning of the lithospheric mantle, control the magnitude of syn-rift subsidence of extensional basins, the amount of uplift of their rift flanks and the position of the crust-mantle boundary. Similarly, these processes have a bearing on the magnitude of post-rift subsidence of extensional basins and, thus, on the ultimate depth of the Moho discontinuity at the end of the post-rift re-equilibration of the lithosphere-asthenosphere system (Ziegler & Cloetingh, 2004).

During phases of collision-related intraplate compression, inversion of extensional basins and upthrusting of Rocky Mountain-type arrays of basement blocks can lead to crustal thickening and depression of the Moho discontinuity (Ziegler et al, 1998, 2002). Moreover, intraplate compressional stresses can cause broad-scale folding of the lithosphere and, depending on its polarity, depression or uplift of the crust-mantle boundary (Cloetingh et al., 1999).

Caledonian crustal domain

The Caledonides of the British Isles and Scandinavia are thought to have collapsed shortly after their earliest Devonian consolidation in response to orogen-parallel extension, reflecting the activation of the Arctic-North Atlantic megashear (Ziegler, 1989; Braathen et al., 2002). Wrench faulting and rifting controlled the Devonian development of the Orcadian pull-apart basin and uplift of core-complexes, and the subsidence of the Midland Valley and Dublin-Northumberland grabens, respectively. The Dublin-Northumberland graben is superimposed on the Iapetus suture (Fig. 2). During Late Devonian to Namurian times (370-315 Ma), the Midland Valley and Dublin-Northumberland grabens were sites of crustal extension. However, crustal thinning related to the development of these basins, which were partly inverted during the Late Westphalian (310-305 Ma), is not clearly reflected by the present-day depth of the Moho discontinuity (Ziegler, 1989, 1990).

The Mid-European Caledonides, that are exposed in the Ardennes and mark the Rheic suture between the Gondwana-derived East-Avalonia and the composite Armorican-Saxo-Thuringian terranes (Armorican Terrane Assembly; Pharaoh, 1999; Winchester & PACE, 2002; Verniers et al., 2002), were disrupted during the Early Devonian by back-arc extension, controlling the opening of the Rheno-Hercynian Basin and limited sea-floor spreading in its Lizzard and Giessen-Harz sub-basins. By contrast, there is only limited evidence for Devonian tensional reactivation of the North German-Polish Caledonides, which are associated with the southwest- to south-dipping Thor-Tornquist suture along which East-Avalonia was welded to the East-European Craton (Banka et al., 2002; Krawczyk et al., 2002). Following opening of the oceanic Lizzard-Giessen-Harz Basin , its northern passive margin was transgressed and developed into the broad Rheno-Hercynian Shelf that was dominated by carbonate platforms, particularly during the Middle Devonian and Early Carboniferous (Ziegler, 1990). This reflects rapid degradation of the North German-Polish Caledonides and of the central North Sea area where they merge into the Scottish-Norwegian Caledonides. Middle and Late Devonian development of a shallow sea arm, which extended from the Rheno-Hercynian Shelf into the central North Sea, may reflect mild tensional reactivation of the north-western segment of the Thor-Tornquist suture (Ziegler, 1990; Williamson et al., 2002). By

Early Carboniferous times, the thickness of the continental crystalline crust underlying the Rheno-Hercynian Shelf may have ranged up to 35 km in its northern parts and to some 38 km in the area of the London-Brabant Massif, tapering to zero along the northern margins of the oceanic Lizzard and Giessen-Harz basins.

Variscan crustal domain

The Variscan Orogen is delimited to the north by its external Rheno-Hercynian thrust belt (Fig. 2). This thrust belt evolved by imbrication of the crust and sedimentary cover of the Rheno-Hercynian Shelf, which, following Early Carboniferous closure of the oceanic Lizzard-Giessen-Harz Basin was converted into a flexural foreland basin that subsided in response to thrust- and slab-loading (Oncken et al., 1999; 2000). The internal parts of the Variscan Orogen include a number of Gondwana-derived continental terranes, such as the Armorican, Saxo-Thuringian, Bohemian, Moldanubian, Aquitaine-Cantabrian blocks (Ziegler, 1989; Pharaoh, 1999; Franke et al., 2000). Sutures marking the location of subducted oceanic basins delimit these terranes. During the Late Devonian and Carboniferous main phases of the Variscan orogeny, major crustal shortening and subduction of continental lithospheric material was accompanied by widespread high-pressure metamorphism and associated magmatism (Franke et al., 2000; Ziegler et al., 2004).

Intraplate compressional stresses, which were exerted onto the Variscan foreland, are held responsible for the Mid- to Late Carboniferous tensional reactivation of the Arctic-North Atlantic megashear and the onset of rifting in the Norwegian-Greenland Sea area, as well as for the Westphalian partial inversion of Carboniferous rifts on the British Isles (Ziegler, 1989, 1990; Ziegler et al., 2002).

By the end-Westphalian times (305 Ma), when crustal shortening in the Variscan orogen ended, its internal parts were probably characterized by a thermally destabilized lithosphere and 45-60 km deep crustal roots. Deep-reaching subducted lithospheric slabs were probably still attached to the Variscan lithosphere at the Bohemian/Moldanubian suture and the suture between the Armorican and Aquitaine-Cantabrian terranes, the latter being associated with the nappe systems on the Arverno-Vosgian and Ligerian zones (Fig. 2). On the other hand, the Rheno-Hercynian zone was underlain by a thermally stabilized foreland lithosphere that extended as a subduction slab some 200 km beneath the Rheno-Hercynian/Saxo-Thuringian suture. The oceanic parts of this slab, corresponding to the Lizzard and Giessen-Harz basins, had already been detached from the foreland lithosphere during the Early Carboniferous (Ziegler et al., 2004, this volume)

Stephanian-Early Permian tectono-magmatic cycle

At the end of the Westphalian (305 Ma), oblique collision of Gondwana and Laurussia gave way to their dextral translation. Stephanian-Early Permian (305-269 Ma) continued crustal shortening in the Appalachian and Scythian orogens was paralleled by the wrench-induced collapse of the Variscan Orogen. Continental-scale dextral shears, such as the Tornquist-Teisseyre, Bay of Biscay, Gibraltar-Minas and Agadir fractures zones, were linked by secondary sinistral and dextral shear systems (Fig. 3). Together, these overprinted and partly disrupted the

Variscan Orogen and its northern foreland (Arthaud & Matte 1977; Ziegler, 1989, 1990; Coward, 1993; Ziegler & Stampfli, 2001). Significantly, wrench tectonics, both of a transtensional and a transpressional nature, as well as associated magmatic activity, abated in the Variscan domain and its foreland during the late Early Permian (285-269 Ma), in tandem with the consolidation of the Appalachian Orogen (Ziegler, 1989, 1990; Marx et al., 1995; Ziegler et al., 2004).

Stephanian-Early Permian wrench-induced disruption of the rheologically weak Variscan Orogen and of its rheologically much stronger northern foreland was accompanied by regional uplift, wide-spread extrusive and intrusive magmatic activity, peaking during the Early Permian, and the subsidence of a multi-directional array of transtensional trap-door and pull-apart basins in which continental clastics accumulated (Fig. 3). Basins developing during this time span show a complex, polyphase structural evolution, including a late phase of transpressional deformation controlling their partial inversion (Ziegler, 1990). Although Stephanian-Early Permian wrench deformation gave locally rise to uplift of extensional core complexes (Vanderhaeghe & Teyssier, 2001), crustal stretching factors were on a regional scale relatively low, as seen in the Southern Permian Basin that is located in the Variscan foreland and encroaches in its eastern parts on the Rheno-Hercynian thrust belt (Ziegler, 1990; van Wees et al., 2000).

Stephanian-Early Permian wrench deformation of the West and Central European lithosphere apparently caused detachment of the subducted Variscan lithospheric slabs and a general reorganization of the mantle convection system, involving the activation of a system of not-very-active mantle plumes. Upwelling of the asthenosphere induced partial delamination and thermal thinning of the mantle-lithosphere and magmatic inflation of the remnant lithosphere. This was accompanied by the interaction of mantle-derived partial melts with the felsic lower crust. These processes accounted for regional uplift and the destruction of the Variscan orogenic roots. By end-Early Permian times, the Variscan crust was thinned down to 28-35 km on a regional scale, mainly by magmatic processes and its erosional unroofing and only locally by its mechanical stretching. Quantitative subsidence curves, derived from the Late Permian and Mesozoic record of intracratonic sedimentary basins, and their modeling suggest that, at the end of the Early Permian, the thickness of the remnant lithospheric mantle ranged between 10-50 km in the area of the Southern Permian and Paris basins, the Hessian Depression and the Franconian Platform (Ziegler et al., 2004, this volume).

Development of a major magmatic province in northern Germany and Poland, which extends to the southwest into the area of the Saar-Nahe Basin, may be a direct consequence of detachment of the Rheno-Hercynian slab. Related thermal thinning of the lithospheric mantle and magmatic thinning of the crust provided the driving mechanism for the Late Permian and Mesozoic subsidence of the Southern Permian Basin (van Wees et al., 2000). In northeast Germany, the thickness of the crystalline crust decreases from 32 km beneath the northern flank of the Southern Permian Basin to 22 km under its axial parts (Bayer et al., 1999). As this part of this basin was only mildly affected by Early Devonian rifting during the opening of the Rheno-Hercynian back-arc basin (Krawczyk et al., 2002), and is neither underlain by major extensional Permo-Carboniferous basins nor was it overprinted by Mesozoic rifting, the observed crustal thinning has to be attributed to Permo-

Carboniferous magmatic destabilization of the crust-mantle boundary (Ziegler et al., 2004).

Similarly, crustal thinning across the Oslo Graben, along the North Danish Basin (Sorgenfrei Line) and probably also along the Polish Trough must be largely attributed to Permo-Carboniferous tectonic and magmatic processes related to the activation of the Tornquist-Teisseyre-Sorgenfrei Line. In Poland this line reflects wrench-induced reactivation of the suture between the Polish Caledonides and the Proterozoic East-European Craton. However, to the northwest, wrench faulting propagated along the Sorgenfrei Line through the Neoproterozoic Dalslandian crust of southern Scandinavia and terminated in the highly volcanic pull-apart Oslo-Skagerrak Graben (Ziegler, 1990; Banka et al., 2002). Across this graben, upper crustal extension by faulting amounts to about 10-20 km, whereas its crustal configuration suggests 40-60 km of extension. This indicates that interaction of mantle-derived melts with the lower crust caused destabilization of the Moho, and thus contributed significantly to the observed crustal thinning (Ro & Faleide, 1992; Ziegler & Cloetingh, 2004).

The occurrence of extensive Permo-Carboniferous dyke swarms and sills in Scotland, and of Stephanian-Early Permian basins and volcanics in the Irish Sea area and on the Western Shelves (Heeremans et al., 2004), shows that the British Isles were also destabilized during the Permo-Carboniferous tectono-magmatic cycle, possibly contributing to thinning of their crust. By analogy with the Southern Permian Basin, subsidence of the Northern Permian Basin, which occupies much of the central North Sea and extends into northern Denmark, probably corresponds also to a zone of thermally driven Permo-Carboniferous thinning of the lithospheric mantle and crust (Ziegler, 1990).

Of special interest is the major magnetic anomaly that transects the Paris Basin along the trace of the Seine-Loire wrench-fault system in the prolongation of the Sillion Houillier shear zone of the Massif Central (Cavellier et al., 1980; Banka et al., 2002). This anomaly probably reflects emplacement of Permo-Carboniferous basic magmas in the lower crust of the stable Cadomian Armorican block. However, in the area of the Sillion Houillier, which transects the Variscan nappe systems and associated intrusive bodies of the Arverno-Vosgian and Ligerian zones (Ledru et al., 2001), this anomaly is not evident. By contrast, the Sillion Houillier part of this deep crustal wrench zone is characterized by distinct crustal thinning that must be attributed to Permo-Carboniferous magmatic destabilization of the Moho, as this zone was neither overprinted by Mesozoic nor by Cenozoic rifting. Similarly, the northwest-trending axis of crustal thinning which underlies the northeast flank of the Aquitaine Basin and the Limousin must be attributed to Permo-Carboniferous destabilization of the crust-mantle boundary.

We conclude that the Permo-Carboniferous tectono-magmatic pulse had a major impact on the crustal configuration of Western and Central Europe.

Late Early Permian to Early Cretaceous thermal sag basins and rifts

In the Variscan domain and its northern foreland, magmatic activity gradually abated during the late-Early Permian (285-269 Ma) and decay of thermal anomalies, introduced during the Stephanian-Early Permian tectono-magmatic cycle, commenced. This reflects that after the Permo-Carboniferous thermal surge (300-280Ma) the temperature of the asthenosphere had returned to ambient levels (1300 ^{0}C). During the Late Permian and Mesozoic, a new, ever expanding system of intracratonic thermal sag basins developed, nucleating from the Southern and Northern Permian basins (Ziegler, 1990; Ziegler et al., 2004). During the development of this basin system, the formerly elevated crust-mantle boundary was gradually depressed. For instance, in the axial parts of the Southern Permian Basin, the crust subsided during Late Permian to Cenozoic times by as much as 8 km in response to cooling and sedimentary loading of the lithosphere (Scheck & Bayer, 1999).

In large parts of Western and Central Europe, post-Early Permian thermal subsidence of the lithosphere was, however, overprinted and partly interrupted by the Late Permian-Early Triassic onset of a new rifting cycle which preceded and accompanied the step-wise break-up of Pangea. Major elements of this break-up system were the southward propagating Arctic-North Atlantic and the westward propagating Neotethys rift systems (Ziegler & Stampfli, 2001). During the Triassic, a multi-directional rift system developed in Western and Central Europe, major constituents of which are the North Sea rift, the North Danish-Polish Trough, the graben systems of the Atlantic shelves and the Bay of Biscay rift. Development of these grabens, partly involving tensional reactivation of Permo-Carboniferous fracture systems, persisted during the Jurassic and Early Cretaceous and was in some rifts accompanied by major crustal extension and commensurate thinning of the crust (Fig. 4; Ziegler et al., 2001). For instance, across the northern and central parts of the North Sea rift, upper crustal extension by faulting amounted to 30-40 km (Ziegler, 1990; Ziegler & Cloetingh, 2004).

The Norwegian-Greenland Sea rift propagated southward during the Late Permian into the north-western shelves of the British Isles, and during the Triassic into the Central Atlantic domain, whilst the Neotethys rift systems propagated westward into the Bay of Biscay and Northwest Africa and linked up with the Atlantic rift system in the North Atlantic domain (Ziegler, 1988, 1990; Ziegler & Stampfli, 2001). This was accompanied by activation of the Central Iberian rift (Salas et al., 2001).

During the Early Triassic, the North Sea rift, consisting of the Horda half-graben, the Viking, Murray Firth, Central and Horn grabens, was activated and transected the western parts of the Northern and Southern Permian basins whilst the North Danish-Polish Trough transected their eastern parts. Simultaneously, the rift systems of the Alpine domain, the Bay of Biscay and the Western Shelves were activated. The latter included the Porcupine, Celtic Sea and Western Approaches troughs. Crustal extension across the Celtic Sea and Western Approaches troughs was compensated, at their eastern termination, by reactivation of Permo-Carboniferous shear systems controlling the subsidence of the Channel and Wessex basins and intermittent destabilization of the Paris thermal sag basin (Ziegler, 1990; Boldy 1995).

Following late Early Jurassic crustal separation in the Central Atlantic (190-180 Ma) and Middle Jurassic (177-160 Ma) crustal separation in the Alpine Tethys, the evolution of the West and Central European rifts was dominated by northward propagation of the Atlantic rift system (Ziegler, 1988; Ziegler et al., 2001; Stampfli & Borel, 2004). During the Late Jurassic and earliest Cretaceous, accelerated rifting activity is evident in the Western Approaches, Celtic Sea and Porcupine, Rockall-Faeroe troughs and the Bay of Biscay. At the same time, rifting accelerated in the North Sea, focusing on its axial Viking and Central grabens. This was accompanied by the development of sinistral shear systems at the southern termination of the North Sea rift, controlling the subsidence of the transtensional Sole Pit, Broad Fourteens, West Netherlands, Lower Saxony, Sub-Hercynian and Altmark-Brandenburg basins. By contrast, crustal extension across the North Danish-Polish trough apparently waned at the Jurassic-Cretaceous transition (Ziegler, 1990; Kutek, 2001).

In the North Atlantic, crustal separation progressed gradually northwards during the Late Jurassic and Early Cretaceous and by Mid-Aptian times (±110 Ma) crustal separation was achieved in the Bay of Biscay (Ziegler et al., 2001). With this, the grabens on the Western Shelves became inactive and began to subside thermally. Following the Late Jurassic-Early Cretaceous rifting pulse, tectonic activity gradually abated also in the North Sea rift system whilst crustal extension focused on the zone of future crustal separation between Europe and Greenland (Ziegler, 1988, 1990; Osmundsen et al., 2002). With this, post-rift thermal subsidence of the North Sea Basin began.

Late Cretaceous and Paleocene rifting and intraplate compression

During the Late Cretaceous and Palaeocene, rifting activity was centered on the Rockall-Faeroe Trough and the area between the Rockall-Hatton-Faeroe Bank and Greenland. During the Cenomanian-Santonian (98-84 Ma), limited sea-floor spreading may have occurred in the southern parts of the Rockall Trough. During the Campanian-Maastrichtian (84-65 Ma), the Iceland plume impinged on the North Atlantic-Greenland Sea rift system, giving rise to the Palaeocene (65-55 Ma) development of the Thulean flood basalt province that had a radius of more than 1000 km (Ziegler, 1988; Morton & Parson, 1988; Larsen et al., 1999). At the same time, the northern parts of the British Isles and the Rockall-Hatton-Faeroe Bank were thermally uplifted and subjected to erosional unroofing. Mantle-derived melts, underplating and intruding the crust, probably contributed in the area of the Hebrides Shelf and the Rockall-Hatton-Faeroe Bank towards crustal thinning. At the Palaeocene-Eocene transition (55 Ma), crustal separation was achieved between Greenland and Europe to the west of the Rockall-Hatton-Faeroe Bank and in the Norwegian-Greenland Sea (Mosar et al., 2002). With this, volcanic activity terminated on the conjugate margins and became centered on the evolving sea-floor spreading axes and on Iceland (Ziegler, 1988, 1990).

During the Turonian-Santonian (93.5-83.5 Ma), Africa began to converge with Europe in a counter-clockwise rotational mode (Rosenbaum et al., 2002). Resulting space constraints within the Tethyan belt caused activation of new subduction zones that controlled the gradual closure of the Alpine Tethys and the Bay of Biscay (Stampfli et al., 2001). Commencing in the late Turonian (±90 Ma), compressional stresses were exerted on the northern Tethyan shelves of the Eastern Alps and Carpathians, inducing inversion of Mesozoic tensional basins and upthrusting of basement blocks by reactivation of pre-existing crustal discontinuities. The

Senonian pulse of intraplate compression (89-70 Ma), which affected the North Danish-Polish Trough, the Bohemian Massif, the Brandenburg-Altmark, Sub-Hercynian, Lower Saxony, West Netherlands and the Sole Pit basins, as well as the southern parts of the North Sea rift, can be related to compressional stresses that were projected from the Alpine-Carpathian orogenic wedge through the oceanic lithosphere of the Alpine Tethys into the lithosphere of Western and Central Europe. The more intense Palaeocene phase of intraplate compression (65-55 Ma), which affected about the same areas, as well as the Tethyan shelves of the Western and Central Alps, the Paris Basin and the Channel area, probably marked the collision of the Alpine orogenic wedge with its East Alpine-Carpathian foreland and with the Briançonnais terrane in the West and Central Alpine domain (Ziegler et al., 1998; Dèzes et al., 2004). This phase of foreland compression, during which a Rocky Mountain-type array of basement blocks was upthrusted in the Bohemian Massif and the Polish Trough was deeply inverted, involved also broad lithospheric folding and accelerated subsidence of the North Sea Basin. Strong inversion of the Polish Trough caused thickening of its crust and the development of an up to 50 km deep Moho keel that is offset to the northeast with respect to the inversion axis as defined at supracrustal levels. Similarly, imbrication of the basement of the Bohemian Massif entailed crustal thickening (Fig. 5; Ziegler, 1990; Ziegler et al., 1995, 2002; Bayer et al., 1999; Jensen et al., 2002).

Convergence rates between the Africa-Arabian and European plates decreased sharply from as much as 20 mm/y during the Late Cretaceous to practically zero during the late Maastrichtian and Paleocene (70-55 Ma) (Rosenbaum et al., 2002). Presumably this resulted from strong collisional coupling of the Africa-Arabian and European plates across the Alpine-Mediterranean orogen. This gave rise to the Paleocene pulses of intense intraplate compression which affected not only Western and Central Europe but also the East-European Craton and North Africa (Ziegler et al., 2001; Nikishin et al., 2001).

Opening of North Atlantic, Cenozoic rifting and interaction of the Alpine Orogen with the European foreland

With the early Eocene onset of sea-floor spreading between Greenland and Europe (55.9-53.3 Ma; Mosar et al., 2002), post-rift thermal subsidence of the Rockall-Hatton-Faeroe Bank, the Rockall-Faeroe Trough and the shelves of northwest Ireland and Scotland commenced. However, during the late Eocene and Oligocene reorganization of sea-floor spreading axes in the Norwegian-Greenland Sea, the Atlantic shelves of the British Isles were destabilized by minor wrench faulting in the prolongation of the Iceland ridge and the Charlie Gibbs fracture zone, causing the development of inversion structures and the subsidence of small pull-apart basins in the Irish Sea area (Fig. 5; Ziegler, 1990; Boldreel & Andersen, 1998; Mosar et al., 2002).

Convergence rates between Africa and Europe gradually increased during the Eocene and Oligocene (55-23.8 Ma), but decreased again during the early Miocene (Rosenbaum et al., 2002). Thrust-loaded deflection of the foreland of the Western, Central and Eastern Alps, as well as of the Carpathians, commenced during the Eocene. Eocene to middle Miocene emplacement of the East-Alpine and Carpathian nappe systems was not accompanied by further intraplate compressional deformation of their foreland, thus reflecting mechanical decoupling of these orogens from their forelands. By contrast, late Eocene-early Oligocene and late Oligocene-early Miocene inversion pulses evident in the Celtic Sea, Western Approaches, Channel, Wessex, Paris, Sole Pit, Broad Fourteens and West Netherlands basins reflect transmission of compressional stresses from the evolving West and Central Alpine Orogen into its foreland and thus their mechanical coupling (Ziegler, 1990, Ziegler et al., 2002; Dèzes et al., 2004). The present deep crustal roots of the Central and Eastern Alps evolved in response to continued underthrusting of the foreland after detachment of the subducted oceanic Alpine-Tethys lithospheric slab towards the end of the Eocene (Schmid et al., 1996; Stampfli et al., 1998). By contrast, the subducted slab of the Western Alps remained attached to the lithosphere until early Pliocene times (Dèzes et al., 2004; Schmid et al., 2004). Oligocene and later underthrusting and subduction of little attenuated foreland lithosphere, combined with the development of an upper plate mantle back-stop, accounted for increasing mechanical coupling of the West and Central Alpine orogenic wedge with its foreland at crustal and lithospheric mantle levels (Ziegler and Roure, 1996).

Evolution of the Pyrenees commenced during the Campanian (80 Ma) and lasted until the early Miocene (±20 Ma), involving northward subduction of the Iberian lithosphere under Europe and southward subduction of the oceanic crust of the Bay of Biscay under Iberia (Vergéz & Garcia-Senez, 2001). During the Paleocene and Eocene, foreland compression controlled the evolution of the Languedoc-Provençal fold-and-thrust belt whilst thrust-loaded subsidence of the Aquitaine and Ebro foreland basins commenced. During the late Eocene and Oligocene, the Ebro foreland basin became isolated in response to inversion of the Mesozoic Central Iberian and Catalan Coast Range rifted basins (Fig. 5; Salas et al., 2001; Ziegler et al., 2002).

Development of the European Cenozoic rift system (ECRIS), which extends from the Dutch North Sea coast into the West-Mediterranean, commenced during the late Eocene (Fig. 5). Its southern elements are the Valencia Trough, the graben systems of the Gulf of Lions, and the northerly striking Valence, Limagne and Bresse grabens; the latter two are superimposed on the Massif Central and its eastern flank, respectively. The Burgundy Transfer Zone links these grabens with the southern end of the northerly striking Upper Rhine Graben. A further, though more diffuse transform fault system links the northern ends of the Limagne and Upper Rhine grabens and crosses the eastern parts of the Paris Basin. Northward, the Upper Rhine Graben bifurcates into the northwest-trending Roer Graben and the northerly-trending Hessian grabens, which transect the Rhenish Massif. The northeast striking Ohre (Eger) Graben, which cuts across the Bohemian Massif, forms an integral part of ECRIS (Ziegler, 1994; Dèzes et al., 2004).

Tensional reactivation of Permo-Carboniferous and Mesozoic shear systems played an important role in the localization of ECRIS. The on-shore parts of ECRIS are associated with a distinct and broad shallowing of the crust-mantel boundary, that can be only partly attributed to Cenozoic rifting as upper crustal extension across the Upper Rhine Graben and the grabens of the Massif Central does not exceed 7 km (Fig. 5; Dèzes et al., 2004). Evolution of ECRIS was accompanied by the development of major volcanic centers in Iberia, on the Massif Central, the Rhenish

Massif and the Bohemian Massif, particularly during Miocene and Plio-Pleistocene times (Wilson & Bianchini, 1999). Mantle tomography images beneath ECRIS a system of upper asthenospheric low velocity anomalies, interpreted as plume heads that have spread out above the 410 km discontinuity (Goes et al., 1999; Spakman, 2004; Sibuet et al., 2004). From these anomalies secondary, relatively weak plumes presently rise up beneath the Eifel (Ritter et al., 2001) and Massif Central (Granet et al., 1995), but not beneath the Vosges-Black Forest arch (Achauer & Masson, 2002). These upper asthenospheric anomalies presumably developed during the Paleocene, following activation of the Northeast Atlantic and Iceland mantle plumes that rise up from the core-mantle boundary (Hoernle et al., 1995; Bijwaard and Spakman, 1999), and subsequently evolved further. This is compatible with volcanic activity in the ECRIS area that commenced during the Paleocene and persisted into the Quaternary (Dèzes et al., 2004). As through time a shift in areas of main volcanic activity can be observed, it is likely that the supply of partial melts through secondary upper mantle plumes was not steady state but pulsating and entailed a shift in their location. In the ECRIS area, this plume activity caused thermal weakening of the lithosphere, thus rendering it prone to deformation, but was not the driving mechanism of rifting. ECRIS is generally considered to have evolved in response to passive rifting that was mainly controlled by compressional stresses originating at the Alpine and Pyrenean collision zones (Dèzes et al., 2004).

During the late Eocene, the Limagne, Valence, Bresse, Upper Rhine and Hessian grabens began to subside in response to northerly-directed compressional stresses that reflect collisional interaction of the Pyrenees and the Alps with their foreland (Merle & Michon, 2001; Schumacher, 2002; Dèzes et al. 2004). During their Oligocene main extensional phase, these originally separated rifted basins coalesced and the Roer and Ohre grabens came into evidence. During the late Oligocene, rifting propagated southward across the Pyrenean Orogen into the Gulf of Lions and along coastal Spain in response to back-arc extension that was controlled by eastward roll back of the Alpine Tethys subduction slab that dipped beneath the Corsica-Sardinia-Balearic-Betic arc system. By late Aquitanian times (21.5 Ma), crustal separation was achieved in the West-Mediterranean Basin, the oceanic Provençal-Ligurian Basin began to open, and the grabens of southern France and the Massif Central became inactive (Séranne, 1999; Roca, 2001). By contrast, the Upper Rhine and Roer Valley grabens remained tectonically active until the present under a northwest-directed compressional stress field that developed during the Miocene (Dèzes et al., 2004). By end-Oligocene times, the area of the triple junction of the Upper Rhine, Roer and Hessian grabens was uplifted and magmatic activity on the Rhenish Shield increased, probably accompanied by plume-induced thermal thinning of the mantle-lithosphere. By mid-Miocene times (±18 Ma) the Massif Central, the Vosges-Black Forest Arch and, slightly later, also the Bohemian Massif was uplifted. This was accompanied by increased mantle-derived volcanic activity (Ziegler. 1994; Merle & Michon, 2001). At the Moho level, a broad anticlinal feature extends from the Massif Central via the Burgundy Transfer Zone and the Vosges-Black Forest into the Bohemian Massif. Development of this arch, which was paralleled by imbrication of the External Massifs of the Alps, can be attributed to folding of the lithosphere in response to the build-up of collision-related compressional stresses at mantle-lithospheric levels in the Alpine foreland (Dèzes et al., 2004). This concept is compatible with the lack of lithospheric thinning beneath the Vosges-Black Forest Arch (Achauer & Masson,

2002) and the lithospheric configuration of the Bohemian Massif (Babushka & Plomerova, 2001), Uplift of this lithospheric fold entailed partial erosional isolation of the Paris Basin (Ziegler et al., 2002).

Under the present northwest-directed stress regime, which had intensified during the Pliocene, the Upper Rhine Graben is subjected to sinistral shear whereas the Roer Graben is extending nearly orthogonally (Dèzes et al., 2004). Moreover, the North Sea Basin is experiencing a Plio-Pleistocene phase of accelerated subsidence and a related depression of the Moho that can be attributed to stress-induced downward deflection of the lithosphere (van Wees & Cloetingh, 1996). Similarly, lithospheric folding probably contributes to the continued uplift of the Fennoscandian Shield (Cloetingh et al., 2004). The present stress field reflects a combination of forces related to collisional interaction of the Alpine Orogen with its foreland and Atlantic ridge push (Gölke et al., 1996).

Alpine Orogens and West-Mediterranean Basins

The Alpine orogenic belts are characterized by variably deep crustal roots. The deepest roots are associated with the Pyrenees (Choukroune et al., 1990; Vergés & Garcias-Senez, 2001), the Alps (Waldhauser, 1998), the northern Apennines (Finetti et al., 2001) and the Dinarides (Skoko et al., 1987), reflecting insertion of continental foreland crust into the mantle and the development of mantle back-stops, involving an offset of the upper and lower plate crust-mantle boundaries (Roure et al., 1996). By contrast, the Betic Cordillera of Spain, the North African Tellian-Maghrebian chain and the Carpathians are characterized by shallower crustal roots or their absence (Fig. 1; see also Cavazza et al., 2004). This can be variably attributed to slab detachment, back-arc extension and early stages of post-orogenic extensional collapse of these orogens.

For instance, in the Central Alps, crustal shortening persisted after the late Eocene detachment of the subducted oceanic Alpine-Tethys slab, accounting for the insertion of a secondary, 120 km long slab, consisting of continental lower crust and lithospheric mantle, into the asthenospheric mantle (Schmid et al., 1996, 2004; Dèzes et al., 2004). As at depths of 55-60 km subducted crustal material entered the eclogite stability field, its P-wave velocity increased to velocities typical for the mantle, and thus, by crossing the Moho discontinuity, limited the seismic depth of the crustal roots (Bousquet et al., 1997; Stampfli et al., 1998).

On the other hand, the Betic-Balearic-Corsica Orogen, which was activated during the Late Cretaceous (±85-80 Ma, Faccenna et al., 2001), was disrupted by late Oligocene-early Miocene back-arc extension, culminating in Burdigalian (±18 Ma) detachment of the Kabylian arc from the orogen. This marked the onset of opening of the oceanic Algerian Basin (Roca, 2001) that was compensated by progressive subduction of the Alpine Tethys (Doglioni et al., 1999a). Following Langhian collision of the Kabylian arc with the North African margin, the subducted Tethys slab was detached, as evidenced by widespread bimodal magmatism, whilst compressional deformation of the Maghrebian-Tellian systems persisted intermittently into the Pleistocene, involving inversion of the Mesozoic Atlas rift system and commensurate crustal thickening (Vergés & Sabat, 1999; Carminati et al., 1999; Frizon de Lamotte et al., 2000; Ziegler et al., 2002; Spakman, 2004).

Following opening of the Provençal-Ligurian Basin and collision of the Corsica-Sardinia accretionary wedge with the Apulian passive margin, commencing in the North during the mid-Oligocene and progressing in time southward, the internal parts of the evolving Apennine orogenic belt were disrupted from the late Miocene onward by back-arc extension governing opening of Tyrrhenian Basin that is partly floored by denuded mantle (shown in Fig. 1 as oceanic crust) (Séranne, 1999; Mauffret & Contrucci, 1999; Faccenna et al., 2001; see also TRANSMED Transect III in Cavazza et al., 2004). Controlling mechanisms are seen in delamination, roll-back, deformation and partial detachment of the subducted Alpine Tethys slab from the Apulian lithosphere that was accompanied by a high-K calc-alkaline to shoshonitic magmatism (Carminati et al., 1998; Doglioni et al., 1999b; Wilson & Bianchini, 1999; Lucente & Speranza, 2001; Argnani & Savelli, 2001; Faccenna et al., 2001; Spakman, 2004).

After the Eocene collisional main deformation phase of the Dinarides, continued northward movement of the Apulian block caused dextral transpressional reactivation of the Sava-Vardar suture during the early Oligocene, triggering detachment of the subducted lithospheric slab and extensive shoshonitic magmatism (Pamic, 2002; Pamic et al., 2002). During the Miocene, eastward extrusion of the Alpine-Carpathian Block and roll-back of the Carpathian subduction system was accompanied by continued crustal shortening in the Carpathians and wrench deformation of the internal Dinarides and the Pannonian domain, controlling the subsidence of transtensional and pull-apart basins (Horvàth, 1993; Frisch et al., 1998; Tari & Pamic, 1998; Fodor et al., 1999). This was coupled with intense thinning of the orogenically destabilized crust and lithospheric mantle of Pannonian Basin, involving upwelling of the asthenosphere (Tari et al., 1999; Cloetingh & Lankreijer, 2001).

Summary and Conclusions

Depending on convergence rates and the thermal state of the foreland lithosphere, crustal roots of active continent-continent collisional orogens can extend to depths of 55-60 km (Alps) or even to 75 km (Himalayas). At these depths the subducted continental crust becomes eclogitized, assumes densities and velocities comparable to those of the mantle and, thus, is transferred across the seismically defined Moho-discontinuity, and seismically appears to form part of the lithospheric mantle (Bousquet et al., 1997; Henry et al., 1997). By analogy with modern orogens, the Caledonian and Varsican orogens were presumably characterized, prior to their post-orogenic collapse, by variably deep-reaching crustal roots.

The crustal roots of the Irish-Scottish-Scandinavian, North German-Polish and Mid-European Caledonides were presumably destroyed in conjunction with Early Devonian post-orogenic wrench faulting and back-arc rifting. Rifting during Late Devonian and Carboniferous times further disrupted the Caledonides of the British Isles. The crustal roots of the Variscan Orogen were destroyed during the wrench-dominated Permo-Carboniferous tectono-magmatic cycle, in the course of which the crust thinned to 28 to 35 km. Simultaneously, the lithosphere of the Variscan foreland was destabilized by wrench faulting and magmatic processes. In the North Sea area, similar to the British Isles, crustal thicknesses were probably quite variable prior to the onset of Mesozoic rifting. Late Permian and Mesozoic rifting, affecting the area of the Atlantic shelves, the North Sea, the Tethys shelves and to a lesser degree the Sorgenfrei-Tornquist-Teisseyre line, caused important crustal thinning that progressed to crustal separation and the Mid-Jurassic opening of the Alpine Tethys, the Early Cretaceous opening of the North Atlantic, and the Mid-Cretaceous opening of the Bay of Biscay and the Valais Trough. Impingement of the Iceland plume, immediately preceding end-Paleocene crustal separation between Greenland and Europe, caused further crustal thinning by magmatic destabilization of the crust-mantle boundary. On the other hand, Eocene to recent development of ECRIS was apparently associated with less intense crustal thinning in the domain of Rhine-Rhône rift system, but progressed by early Miocene times to crustal separation and limited sea-floor spreading in the Provençal-Ligurian Basin. Back-arc extension, controlled by eastward roll back of the Alpine-Tethys slab governed the opening of the oceanic Provençal-Ligurian and Algerian basins and subsidence of the Tyrrhenian Basin, involving mantle denudation. The oceanic floor of the Ionian Sea represents a remnant of Permo-Triassic Neotethys Ocean (Stampfli et al, 2001; Ziegler et al., 2001; Stampfli & Borel, 2004).

The frequently observed discrepancy between the magnitude of upper crustal extension by faulting and the amount of extension derived from the crustal configuration of a rift zone speaks either for syn-rift magmatic destabilization of the crust-mantle boundary (e.g. Oslo Graben) or for non-uniform pre-rift crustal thicknesses (e.g. North Sea Rift, Upper Rhine Graben; Ziegler & Cloetingh, 2004).

Regarding the present-day crustal configuration of extra-Alpine Europe, we would like to further comment on some of the salient features of Figure 1. Broad zones of crustal thinning, that characterize the Rockall-Hatton-Faeroe Bank, the Atlantic shelves of the British Isles, France and Iberia, as well as the North Sea, must be largely attributed to Mesozoic crustal extension, although crustal thicknesses were probably not uniform prior to the onset of Mesozoic rifting. Moreover, in the area of the Hebrides Shelf and the Rockall-Hatton-Faeroe Bank Paleocene plume-related thermal destabilization of the crust-mantle boundary, combined with erosional unroofing of the crust in response to its thermal doming, probably contributed towards crustal thinning. However, Late Cretaceous and Cenozoic post-rift thermal subsidence of these Mesozoic extensional basins caused a gradual depression of the Moho, amounting for instance in the Central North Sea to over 4 km. On the other hand, thinning of the Neoproterozoic Dalslandian crust in the North Danish Trough resulted partly from Mesozoic crustal stretching and partly from Permo-Carboniferous destabilization of the crust-mantle boundary. Crustal thinning in the Oslo Graben must be exclusively attributed to Permo-Carboniferous rifting and magmatic destabilization of the crust-mantle boundary. Similarly, the regional Moho uplift that is still associated with the area of the Southern Permian Basin, despite its Late Permian to Cenozoic subsidence by as much as 8 km, must be attributed to Early Permian magmatic destabilization of the crust-mantle boundary. The SW-NE trending broad zone of Moho shallowing, which is associated with the intracontinental part of ECRIS, probably reflects a combination of (1) Permo-Carboniferous (predominantly magmatic) crustal thinning, (2) Cenozoic crustal extension, (3) erosional unroofing of the crust in response to Neogene plume-induced doming of the Rhenish Massif and the Massif Central, and mid-Miocene-Pliocene lithospheric folding controlling uplift of the Vosges-Black Forest arch.

Latest Cretaceous and Paleocene intraplate compressional deformation probably controlled crustal thickening in the Polish Trough and in the Bohemian Massif and to a lesser extent also in the inverted basins of Denmark, Germany and the Netherlands. Eocene and Oligocene inversion of the Central Iberian and Catalan Coast Ranges resulted in crustal thickening, whilst late Oligocene-early Miocene rifting of the Valencia Trough caused thinning of the previously thickened crust.

The present-day crustal configuration of Western and Central Europe resulted from polyphase Late Palaeozoic to recent deformation of the lithosphere that overprinted the margin of the Proterozoic East-European Craton and, particularly, the Caledonian and Variscan crustal domains. In an effort to explain the crustal configuration of a given area, the total sum of processes that affected it through time must be taken into consideration.

Acknowledgements

This paper is a contribution to EUROPROBE by the European EUCOR-URGENT Project (Upper Rhine Graben: Evolution and Neotectonics). P. Dèzes acknowledges financial support by a University of Basel ELTEM grant. Critical and constructive comments by Wolfgang Franke and Irina Artemieva to a draft of this paper are gratefully acknowledged. Special thanks go to David Gee for enthusiastically leading the EUROPROBE Project that was instrumental in integrating the Earth Science communities of Western and Eastern Europe.

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Text figures and enclosure

Fig. 1 Crustal thickness map of Western and Central Europe (after Dèzes & Ziegler, 2002)

Fig. 2 Caledonian and Variscan structural element superimposed on the crustal thickness map

Fig. 3 Permo-Carboniferous fault systems and magmatic fields superimposed on the crustal thickness map

Fig. 4 Mesozoic rift and wrench fault systems superimposed on the crustal thickness map

Fig. 5 Oligocene and younger rift and wrench systems and magmatism, superimposed on the crustal thickness map, showing areas of latest Cretaceous and Cenozoic compressional intraplate deformation

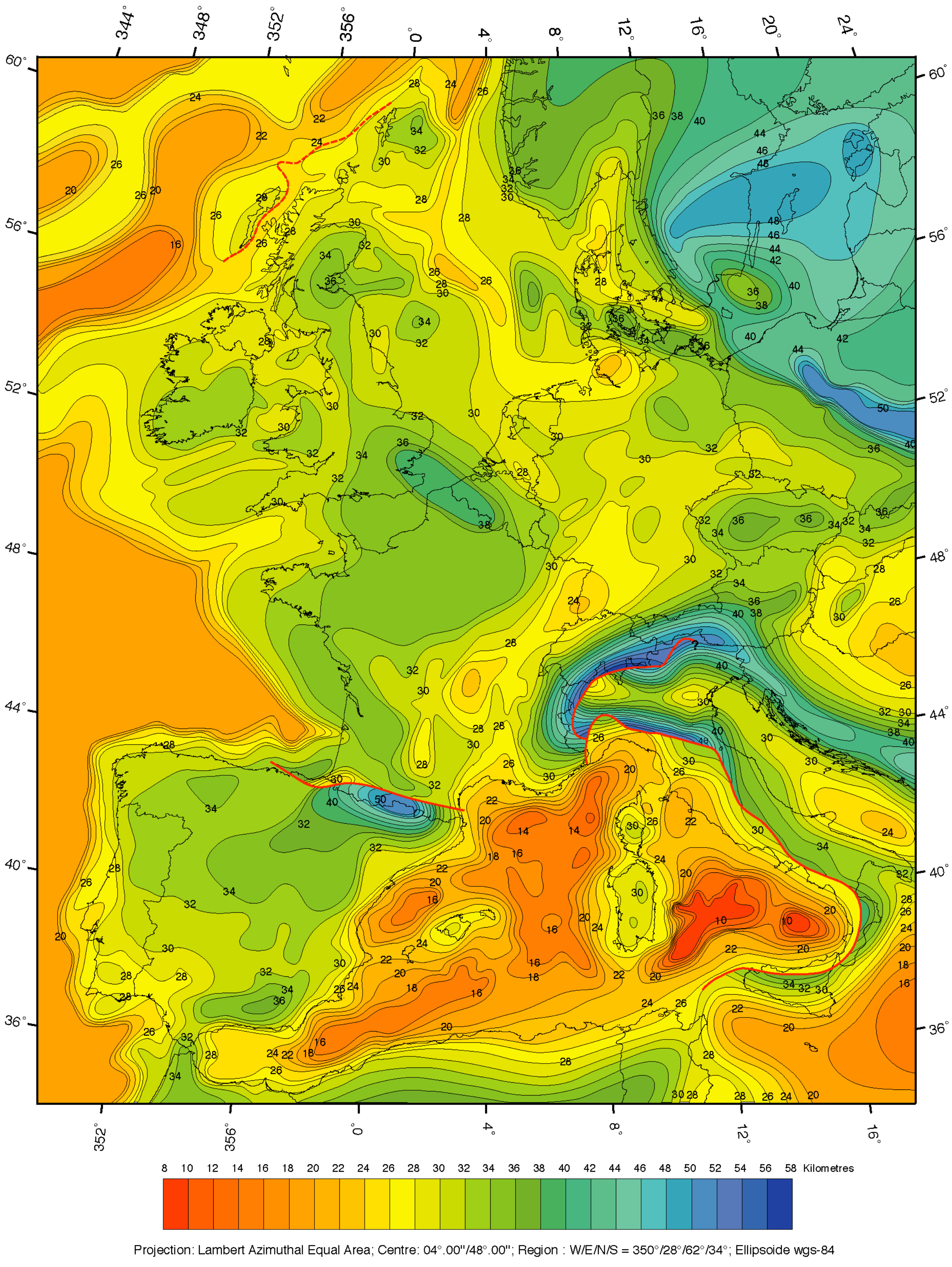
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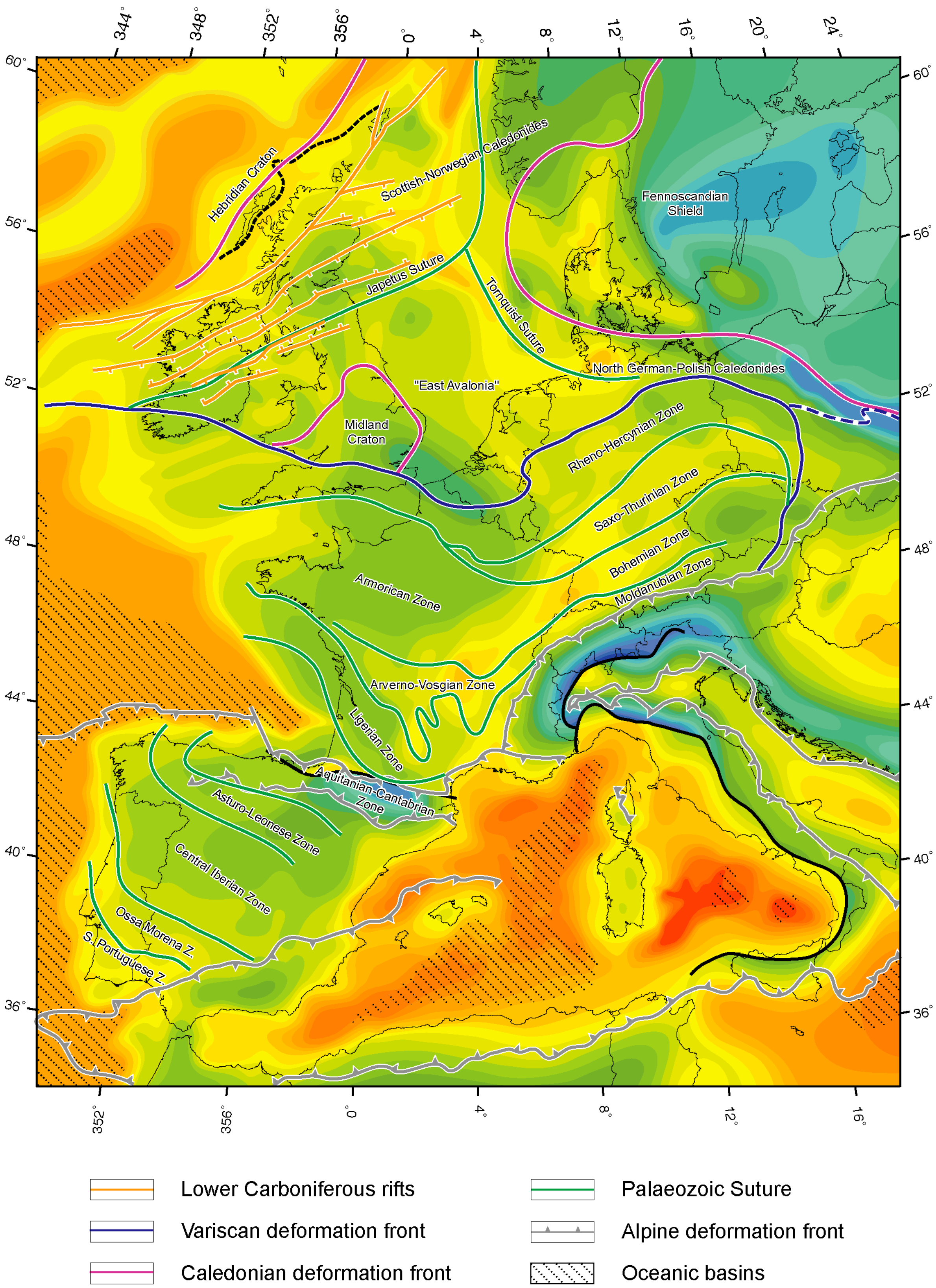
Figure 1


Hebridian Craton
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Tornquist Suture
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North German-Polish Caledonides
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Midland Craton
Rheno-Hercynian Zone
Saxo-Thurinian Zone
Bohemian Zone
Moldanubian Zone
Armorican Zone
Arverno-Vosgian Zone
Ligerian Zone
Aquitanian-Cantabrian Zone
Asturo-Leonese Zone
Central Iberian Zone
Ossa Morena Z.
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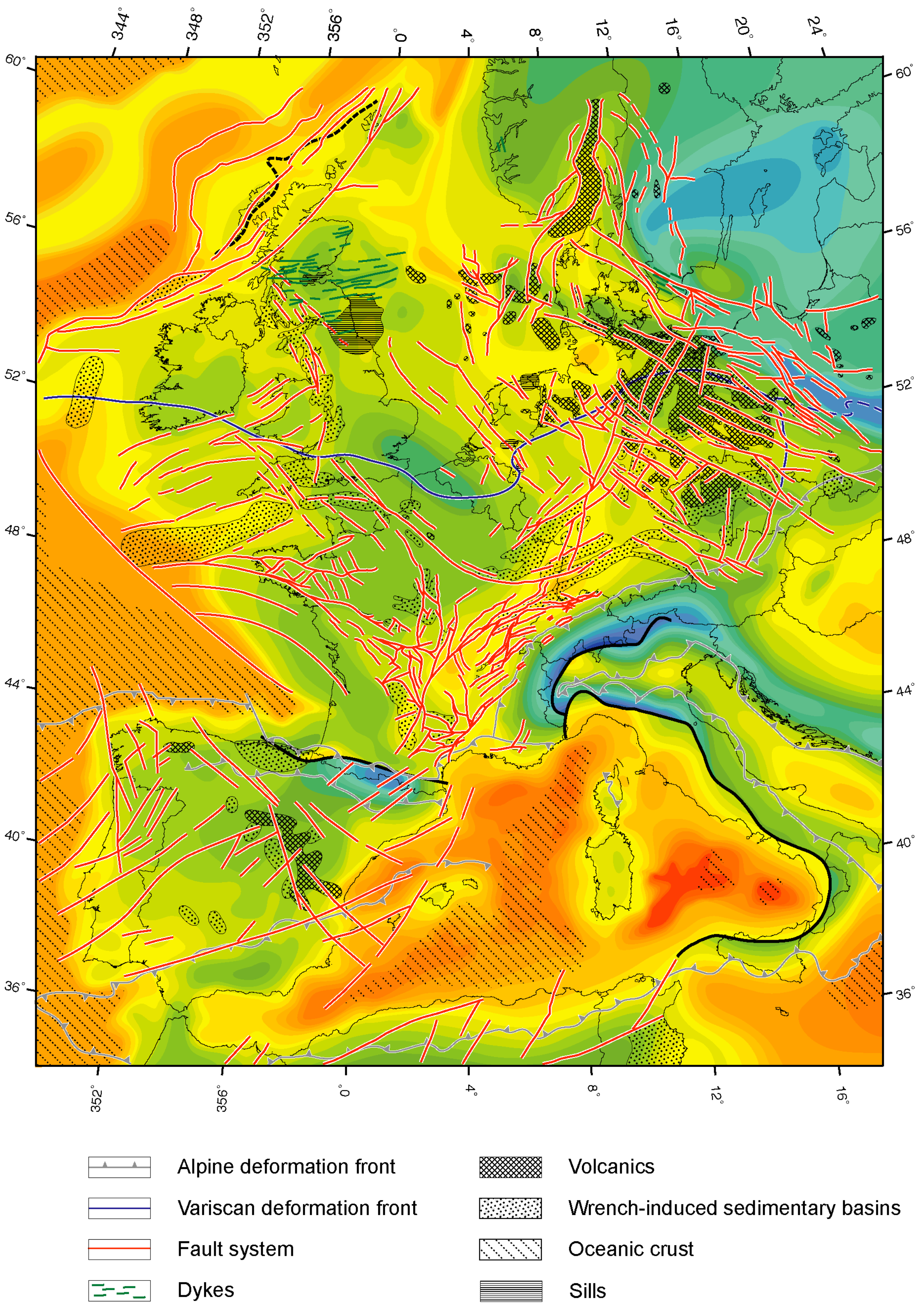
Lower Carboniferous rifts
Variscan deformation front
Caledonian deformation front
Palaeozoic Suture
Alpine deformation front
Oceanic basins

Figure 2



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Alpine deformation front
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Wrench-induced sedimentary basins
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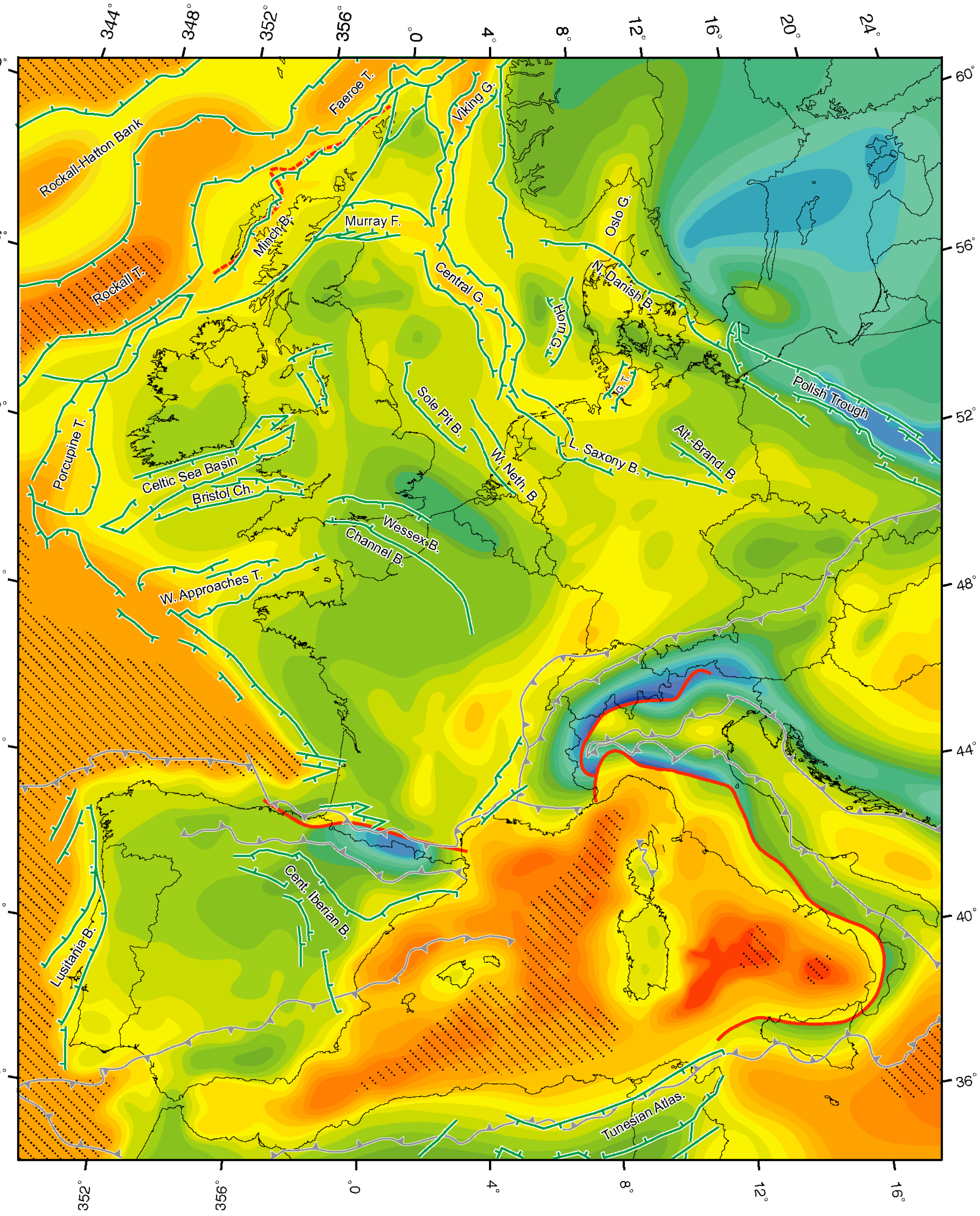
Sills

Figure 3

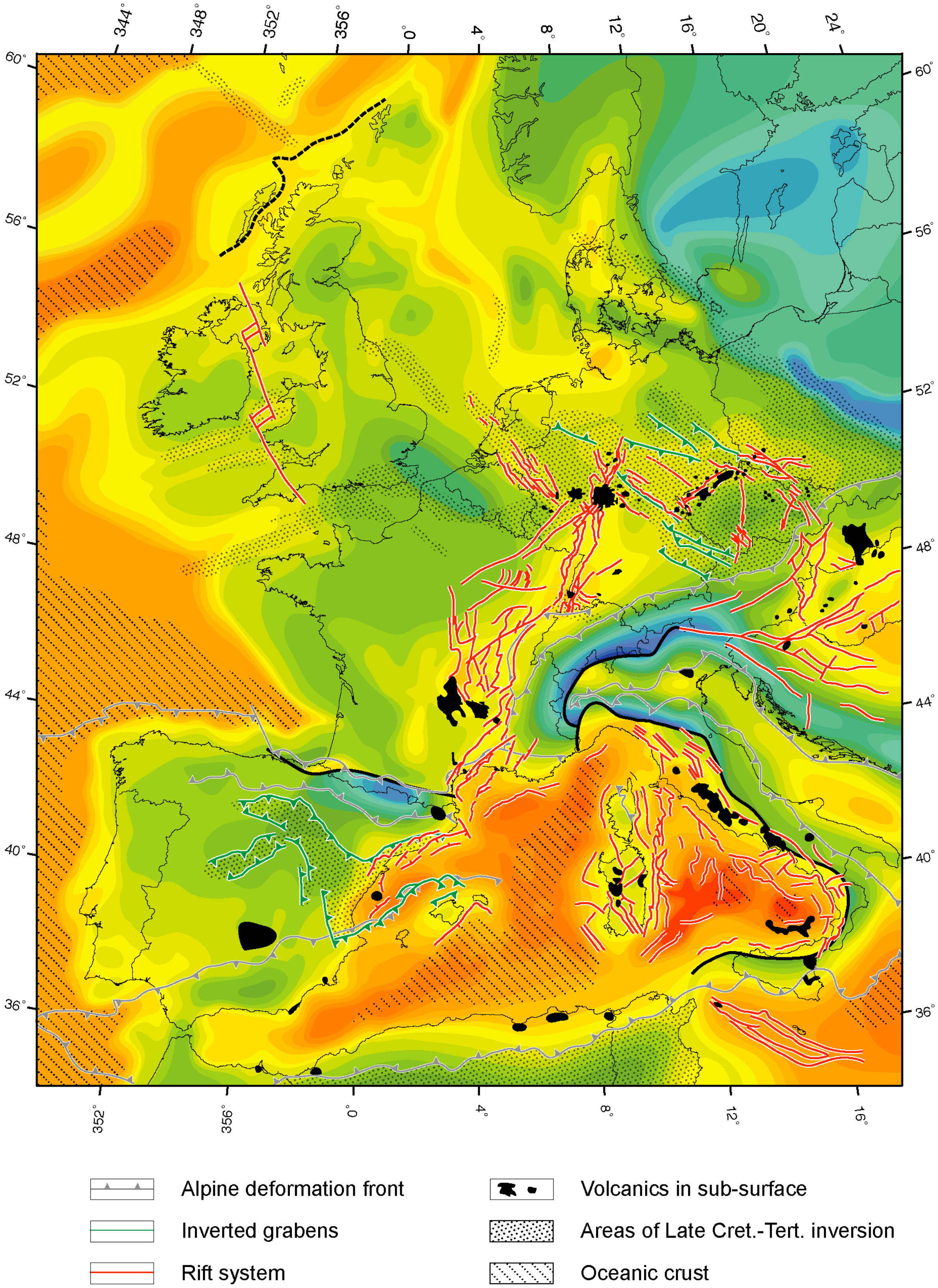


Mesozoic rifts & wrench faults

Alpine deformation front

Oceanic basins

Figure 4



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Rift system
Volcanics in sub-surface
Areas of Late Cret.-Tert. inversion
Oceanic crust

Figure 5