Geologie en Mijnbouw **73:** 99–127, 1994.

Cenozoic rift system of western and central Europe: an overview

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Received 1 July 1993; accepted in revised form 6 October 1993

Key words: plate tectonics, geodynamics, palaeogeography, volcanism, crust, lithosphere, rift models, stress field

Abstract

The Cenozoic rift system of western and central Europe extends over a distance of some 1100 km from the coast of the North Sea to the western Mediterranean; its southern prolongation is formed by the Valencia Trough and a Plio-Pleistocene volcanic chain which crosses the Alboran Sea and the Atlas ranges. Development of this mega-rift was contemporaneous with the Eocene and later phases of the Alpine and Pyrenean orogenies and with the evolution of the Red Sea-Gulf of Suez and Libyan-Pelagian Shelf rift systems. Evolution of the European Cenozoic rift system is thought to be governed by the interaction of the Eurasian and African-Arabian plates and by early phases of a plate-boundary reorganization that may ultimately lead to the break-up of the present continent assembly.

In western and central Europe rifting commenced during the middle and late Eocene; 20–40 Ma later major rift-related domes were uplifted, entailing subsidence reversals of the grabens transecting them. Uplift of the Rhenish Shield can be explained in terms of progressive mechanical and thermal thinning of the lithosphere. The Bohemian Massif, Vosges-Black Forest and Massif Central arches, which are located in the periphery of the Alpine fold belt, are characterized by less pronounced lithospheric thinning; low-velocity mantle-lithosphere anomalies are observed under the Vosges-Black Forest and Massif Central domes; apart from thermal loads, deflection of the lithosphere in response to the build-up of intra-plate horizontal compressional stresses and/or to thrust-loading may have contributed to the uplift of these arches.

Volcanic rocks associated with the Cenozoic rift system of western and central Europe were derived by mixing of partial melts from the convecting asthenosphere and from the mantle-lithosphere; the asthenospheric component shows similarities to the source of ocean island basalts. In the face of limited lithospheric extension, it must be assumed that the upper asthenosphere has a higher than ambient temperature and that the mantle-lithosphere is volatile-enriched. The upper asthenosphere of much of Variscan Europe is characterized by low S-wave velocities, indicating the presence of partial melts. Paleogene development of this anomaly, possibly in conjunction with a reorganization of mantle convection patterns, was accompanied by thermal weakening of the lithosphere, rendering it prone to failure in response to the build-up of intra-plate stresses. The Cenozoic rift system of Europe has many features which are consistent with a 'passive' rift system.

Under the present stress regime, crustal extension is limited to the Roer Valley Graben whereas the Rhine Graben and the Massif Central are subjected to transpressional and transtensional deformation. The entire rift system corresponds to a zone of increased seismic hazard. The Massif Central and the Rhenish Shield are zones of latent volcanic activity.

Introduction

The Cenozoic rift system of western and central Europe consists of the Rhine, Roer Valley and Leine grabens, which straddle the Rhine River and cross-cut the Rhenish Shield, the Eger Graben (Ohre Rift) of the Bohemian Massif, the Bresse, Limagne, Saône, Alès and Gulf of Lions grabens of southeastern France and the Valencia Trough which parallels the east coast of Spain (Fig. 1). This complex rift system extends from the shores of the North Sea over a distance of some 1100 km into the western Mediterranean. From there an alka-

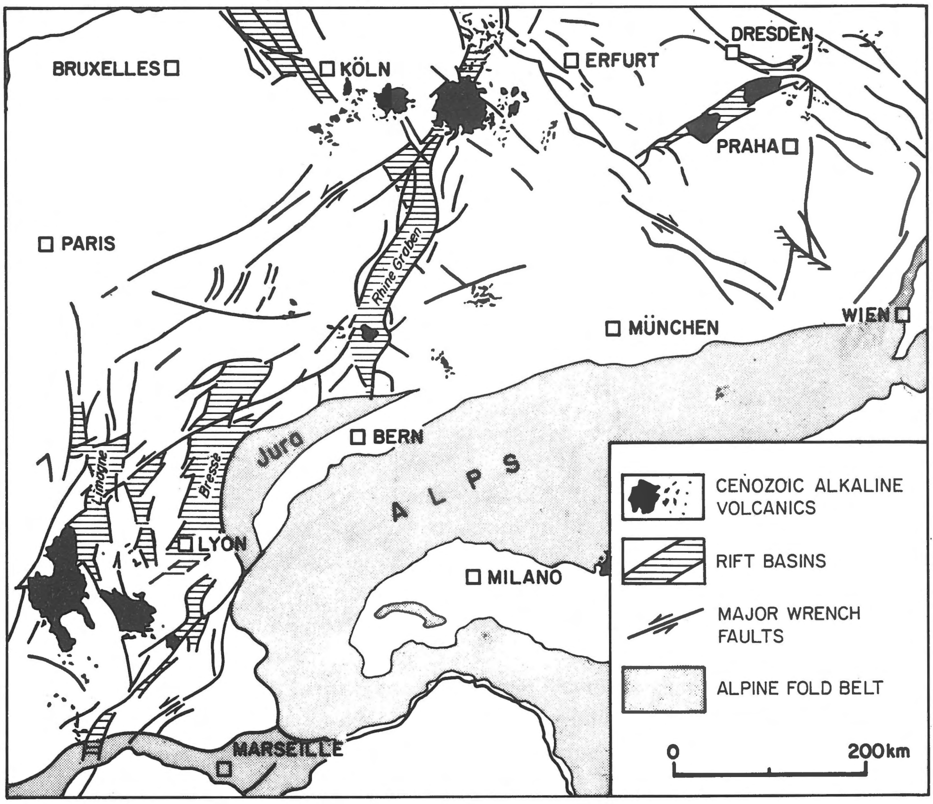

Fig. 1. Structural elements of Cenozoic rift system of western and central Europe (modified after Brousse & Bellon 1983 and Coulon 1992).

line volcanic chain projects southwestwards across the Alboran Sea, the Rif fold belt and the Atlas ranges to the Atlantic coast of northwest Africa and to the Cape Verde Islands; including this volcanic chain, the entire rift system has a length of 3000 km.

The Cenozoic rift system of western and central Europe began to evolve during the middle and late Eocene in the Alpine foreland and propagated during the Oligocene northward and southwards (Figs 2, 3). In the western Mediterranean crustal extension culminated, after a rifting period of only 7 Ma, during the late Aquitanian-early Burdigalian in crustal separation and the opening of the oceanic Algero-Provençal Basin; this involved a counter-clockwise rotation of the Corsica-Sardinia block. During Miocene and Plio-Pleistocene times tectonic and partly also volcanic activity persisted along the different segments of this mega-rift system, albeit under changing regional stress regimes (Fig. 4).

Development of the Cenozoic rift system of western and central Europe was contemporaneous with the Eocene and later phases of the Alpine orogeny, during which the northwestern Alpine foreland was repeatedly subjected to horizontal intra-plate compressional stresses, giving rise to the inversion of Mesozoic extensional basins, for instance in the area of the Western Approaches-Celtic Sea, the English Channel and the southern North Sea. The southern elements of the European Cenozoic rift system cross-cut the Alpine chains of the western Mediterranean domain. Viewed on a broader scale, evolution of the west and central European Cenozoic rift system was broadly contemporaneous with the development of the East African-Red Sea, Libyan and Pelagian Shelf rift systems; its Neogene development was paralleled by back-arc extension governing the subsidence of the Pannonian Basin and the Aegean, Tyrrhenian and Alboran seas. As such the Rhine-Rhône-Valencia and the Red Sea-Libyan rift systems can be considered as forming part of the Neogene Alpine-Mediterranean collapse system (Fig. 4; Ziegler 1988, 1990).

The Cenozoic rift system of western and central Europe hosts major industrial and population centres. It corresponds to a zone of elevated seismic hazard, as

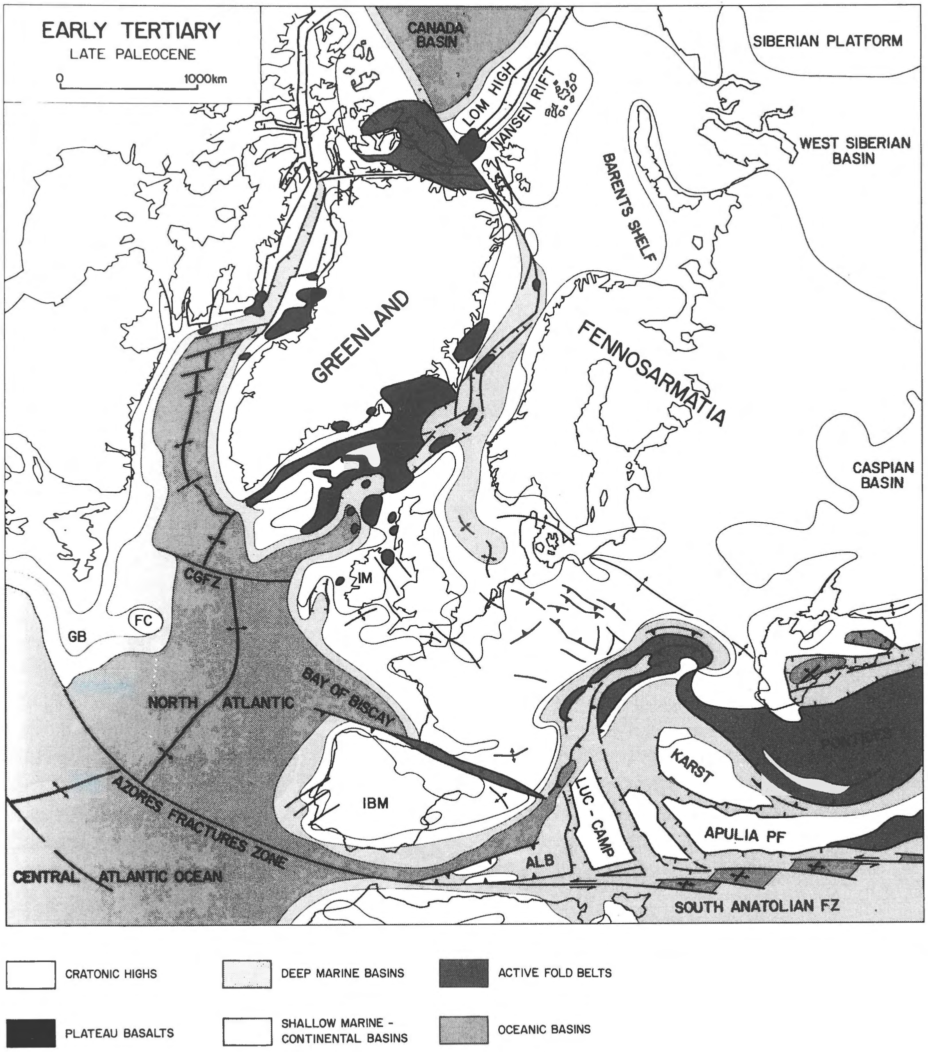

Fig. 2. Paleocene framework of North Atlantic domain. Abbreviations: ALB – Alboran-Kabylia Block, CGFZ – Charlie Gibbs Fracture Zone, FC – Flemish Cap, LUC-CAMP – Lucania-Campania Block.

evident by historical earthquakes such as the one that destroyed in 1356 the city of Basel (southern end of Rhine Graben), the 1477 and 1490 earthquakes near Clermont-Ferrand (Limagne Graben) (Illies & Greiner 1978; Weber 1980; Ahorner 1993; Delouis et al. 1993) and the 1992 earthquake of Roermond. Latent volcanic activity presents an additional hazard in the Eifel and the northern parts of the Massif Central.

The following discussion concentrates on the European onshore parts of this mega-rift system.

Tectonic framework

The Cenozoic rifts of southeastern France strike subparallel to obliquely to the deformation front of the Western Alps and transect the structural grain of the Late Palaeozoic Variscan fold belt at a steep angle (Fig. 1). Localization of individual grabens can be related to tensional reactivation of Stephanian-Autunian and Mesozoic fracture systems; these grabens are linked by northeast-trending sinistral transform faults which largely also coincide with reactivated Permo-Carboniferous fracture systems. Subsidence of the fault-controlled Alès, Saône, Bresse and Limagne

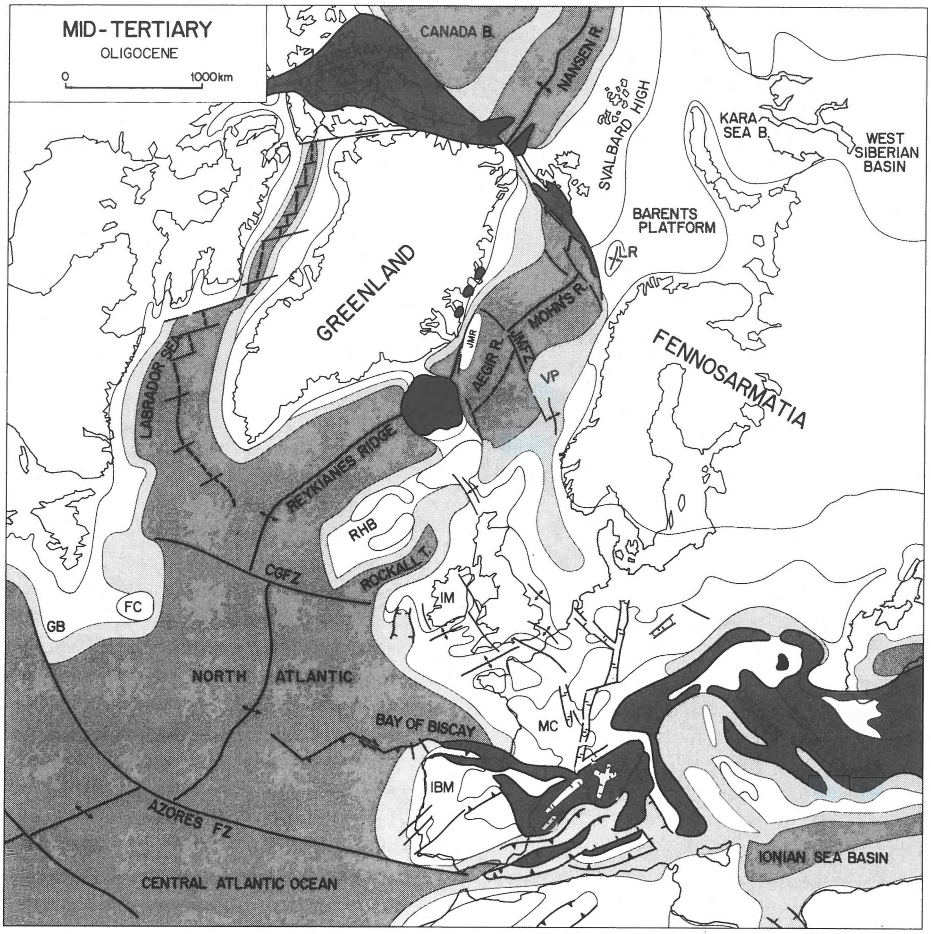

Fig. 3. Late Oligocene framework of North Atlantic domain. Abbreviations: CGFZ – Charlie Gibbs Fracture Zone, FC – Flemish Cap, GB – Grand Banks, IBM – Iberia Meseta, IM – Irish Massif, JMR – Jan Mayen Ridge, LR – Loppa Ridge, MC – Massif Central, RHB – Rockall-Hatton Bank, VP – Vøring Plateau.

basins commenced during the Eocene (Debrand-Passard et al. 1984; Blès et al. 1989). Their evolution was accompanied and in part preceded by volcanic activity that commenced during the late Paleocene, became important during the Miocene and peaked during the Plio-Pleistocene, particularly in the area of the Massif Central where the last volcanic eruption occurred about 4000 years ago (Chenevoy 1974; Brousse & Bellon 1983; Maury & Varet 1980; Downes 1987; Blès et al. 1989).

The northern end of the Bresse Graben is structurally connected with the southern end of the Rhine Graben by a broad, sinistral northeast-striking transform fault system (Laubscher 1970; Rat 1974, 1978; Bergerat 1977). Localization of this so-called Burgundy transfer zone probably involved reactivation of Permo-Carboniferous fracture systems (Laubscher 1986). A somewhat more diffuse transform fault system connects the northern ends of the Limagne and Rhine grabens and crosses the eastern parts of the Paris Basin (Coulon 1992).

The NNE-trending Rhine Graben bifurcates in the area of Frankfurt into the Leine Graben (sensu lato), which maintains the strike direction of the Rhine Graben, and the northwest-trending Roer Valley Graben. Reactivation of Permo-Carboniferous and

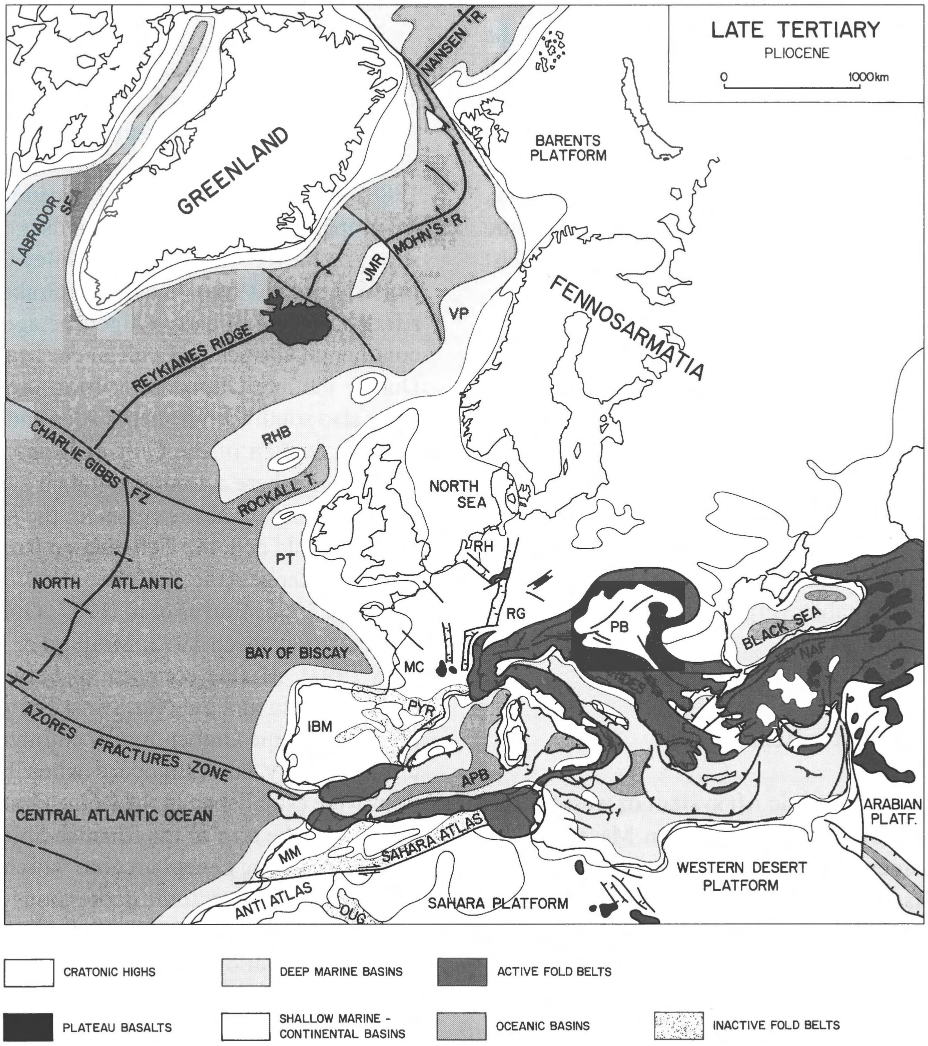

Fig. 4. Pliocene framework of North Atlantic domain. Abbreviations: APB – Algero-Provençal Basin, FC – Flemish Cap, GB – Grand Banks, IBM – Iberia Meseta, IM – Irish Massif, JMR – Jan Mayen Ridge, MC – Massif Central, MM – Morocco Meseta, NAF – North Anatolian Fault, PB – Pannonian Basin, PT – Porcupine Trough, PYR – Pyrenees, RG -- Rhine Graben, RH – Roer Valley Graben, RHB – Rockall-Hatton Bank, VP – Vøring Plateau.

Mesozoic fracture systems played an important role in the localization of these grabens. The Roer Valley Graben dies out northwards in the coastal areas of the Netherlands (Zagwijn 1989; Geluk 1990; Zijerveld et al. 1992) and does not link up with the Mesozoic North Sea rift system which was not tensionally reactivated during Cenozoic times (Ziegler 1990). The Leine Graben looses its identity northward at the edge of the North German Basin (Vinken 1988).

Middle and Late Cretaceous and Paleocene dyke intrusions in the Rhine Graben area were probably triggered by wrench deformations in conjunction with mid-Cretaceous tensional tectonic activity in the area of the Channel and Weald basins and wrench-faulting in the Paris Basin, and by Senonian and mid-Paleocene compressional deformation of the West and Central Netherlands and Lower Saxony basins and of the Bohemian Massif (Ziegler 1990). Subsidence of the Rhine Graben system commenced during the late Eocene. Volcanism was more or less continuous throughout the Cenozoic but peaked during the Miocene and again during the Quaternary. The area of the Rhine-Leine-Roer Valley triple junction was the principal centre of volcanic activity; however, smaller volcanic centres occur up to 100 km to the east of the Rhine Graben, such as in the Rhön area in the north and in the Urach and Hegau areas in the south. On the Rhenish Shield, straddling the triple junction of the Rhine, Roer Valley and Leine grabens, Neogene volcanism spreads over a broad, east-west trending zone, some 250 km wide. During the Quaternary, volcanic activity concentrated on the Eifel area where the last eruption occurred about 11 000 years ago at the Laacher See volcano (Fig. 1; von Eller & Sittler 1974; Illies et al. 1974; Illies 1974, 1978, 1981; Lippolt 1983; Mertes & Schminke 1983).

The ENE-trending Eger Graben is superimposed on the Permo-Carboniferous Central Bohemian-Krkonose-Intra Sudetic Basin. Its development was preceded by Paleocene dyke intrusions and accompanied by important Oligocene, Miocene and Plio-Pleistocene volcanic activity that affected also areas outside the Eger Graben (Figs 5–7; Malkovsky 1980, 1987; Suk et al. 1984).

Evolution of the rift system

In the area of the Cenozoic rift system of western and central Europe, the boundary between Mesozoic and Cenozoic sediments corresponds to a regional erosional hiatus. This break in sedimentation coincides with the Senonian and Paleocene compressional intra-plate deformations which developed in response to the build-up of a stress field that was characterized by northward directed trajectories of the principal horizontal compressive axes; development of this stress field is thought to be related to collisional coupling between the Alpine and Pyrenean orogens and their northern forelands (Fig. 2; Ziegler 1988, 1990). This stress regime relaxed towards the end of the Paleocene and gradually gave way to a new NW-SE oriented extensional stress system (Bergerat 1987a, 1987b; Blès et al. 1989; Lacombe et al. 1990).

In the Alès, Saône, Limagne and Bresse graben system, earliest syn-rift sediments are dated as middle *Eocene* and in the Rhine Graben as late Eocene; these sediments consist of lacustrine shales and carbonates and fringing clastics which give way upward to an evaporitic sequence (Fig. 5). Distal marine incursions reached these rifted basins during the latest Eocene and early Oligocene from the West and Central Alpine foreland basin (Debrand-Passard 1984; Blès et al. 1989; Rat 1974; Sittler 1969a, 1969b; Doebl & Teichmüller 1979; Duringer & Gall 1993).

During the early *Oligocene*, rifting propagated northward and breached the Rhenish Shield (Fig. 6). This led to the development of a narrow sea-way linking the Alpine foreland basin via the Rhine and Leine grabens with the Northwest European Basin (Vinken 1988). Late middle to late Oligocene northwestward rift propagation into the Netherlands is evident from the stratigraphic record of the Roer Valley Graben (Zagwijn 1989; Geluk 1990). In the Eger Graben, the first syn-rift deposits are of early Oligocene age and consist of continental clastics and volcanics (Malkovsky 1987). During the late Oligocene, rifting propagated apparently also southward from the Alès and Saône grabens into the domain of the Gulf of Lions and the Valencia Trough where a complex graben system came into evidence; these grabens cross-cut the structures of the Pyrenean fold belt, the Celt-Iberian Range and Catalonian Cost Ranges (Fig. 3; Cohen 1980; Thomas & Gennesseaux 1986; Burrus et al. 1987; Gorini et al. 1993; Banda & Santanach 1992; Maillard & Mauffret 1993; Bois 1993). Marine connections between the Alpine foreland basin and the Northwest European Basin via the Rhine-Leine Graben remained intermittently open until the early late Oligocene when brackish conditions were established in the Rhine Graben. During the Oligocene, the area of the Rhenish Shield corresponded to a low-lying peneplain onto which transgressions encroached along tectonic depressions and also across the Ardennes (Fig. 6), as indicated by the erosional remnants of shallow-marine and coastal sands (Kadolsky et al. 1983; Meyer et al. 1983; Zöller 1983; Gramann & Kockel 1988). Temporary marine connections may also have been established during the Oligocene between the Alpine foreland basin and the Paris Basin via the graben systems of the Massif Central.

The main tectonic subsidence phases of the Alès, Saône, Bresse and Limagne grabens are of late Eocene-early Oligocene and late Oligocene age (Blès et al. 1989). Similarly, the Rhine Graben subsided rapidly during the late Eocene to early Oligocene (42–31 Ma) and late Oligocene to early Miocene (25–20 Ma) (Villemin & Coletta 1990). Subsidence of the Roer Graben was relatively fast during the Oligocene and slowed down during the Miocene (Zijerveld et al. 1992). In the southern parts of the Rhine Graben, late Eocene and early Oligocene sequences contain major salt and potash deposits which gave rise to the development of diapiric structures. In the Bresse, Limagne and Saône grabens, early and late Oligocene salts are considerably thinner; their partial dissolution caused

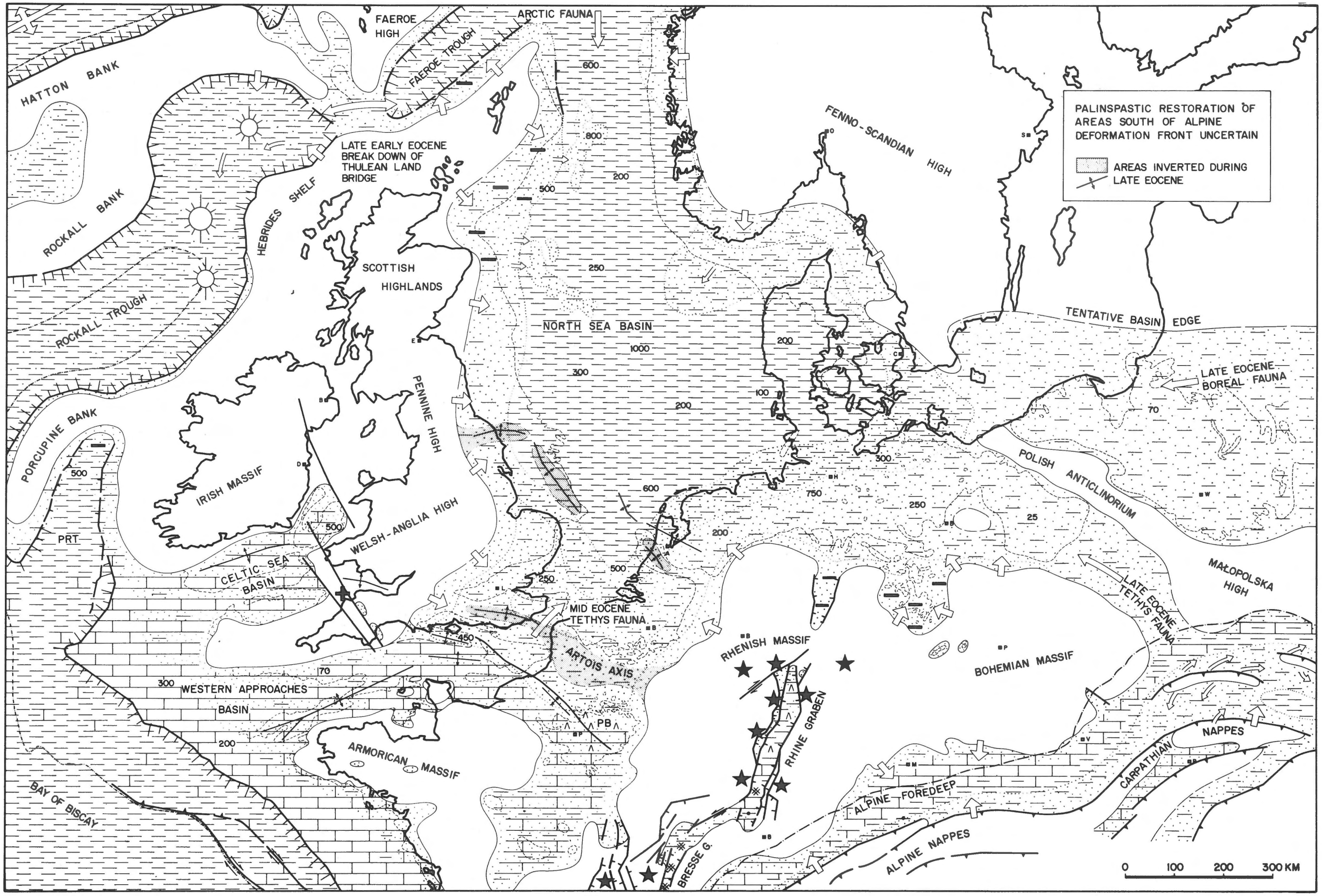

Fig. 5. Eocene palaeogeography of Rhine Graben area (Ziegler 1990); PB Paris Basin, stars: volcanic centers, stippled areas: zones of basin inversion, dashed lines: erosional edge of mapped interval. Thicknesses in meters.

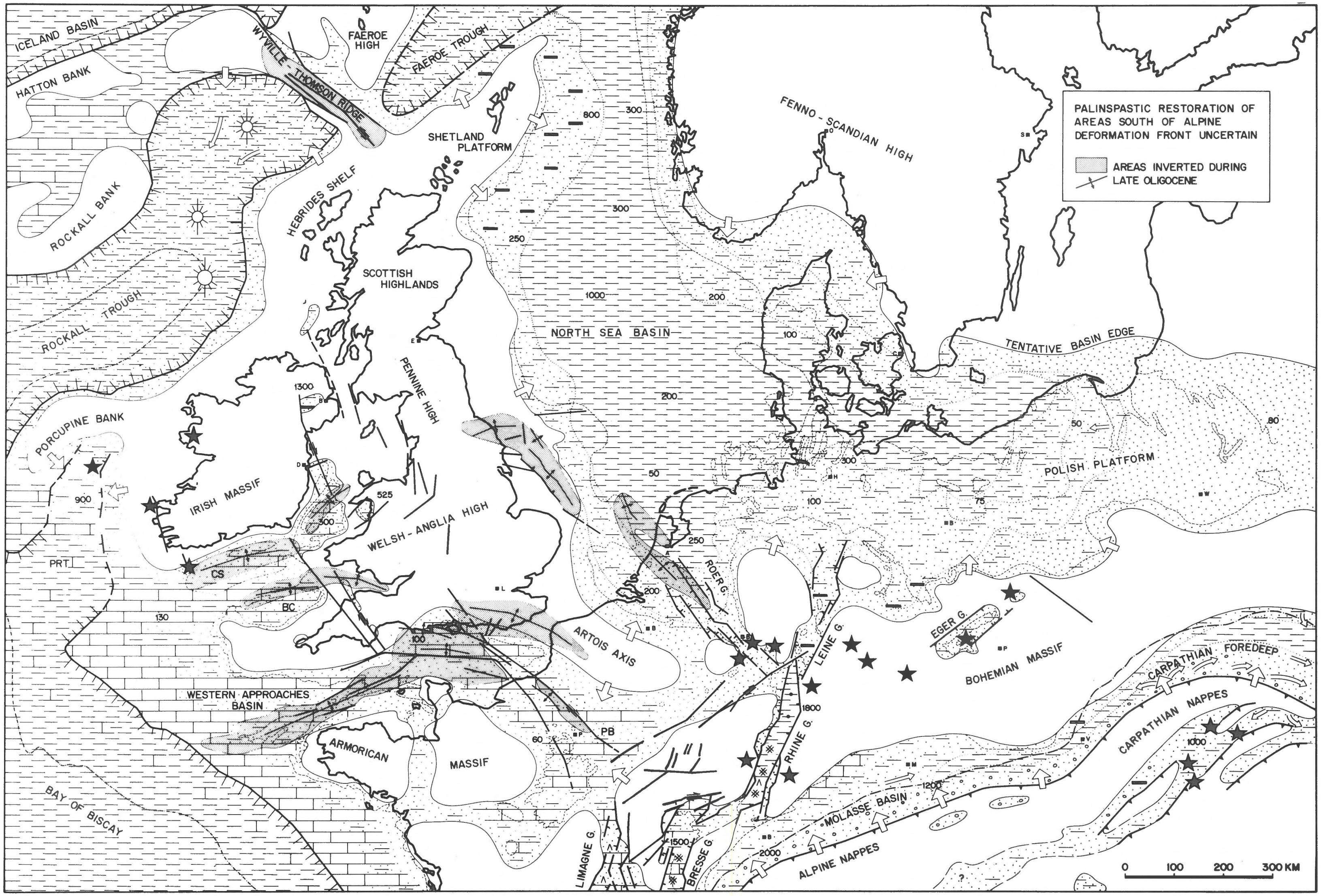

Fig. 6. Oligocene palaeogeography of Rhine Graben area (Ziegler 1990); symbols as Fig. 5; BC – Bristol Channel Basin, CS – Celtic Sea Basin. Thicknesses in meters.

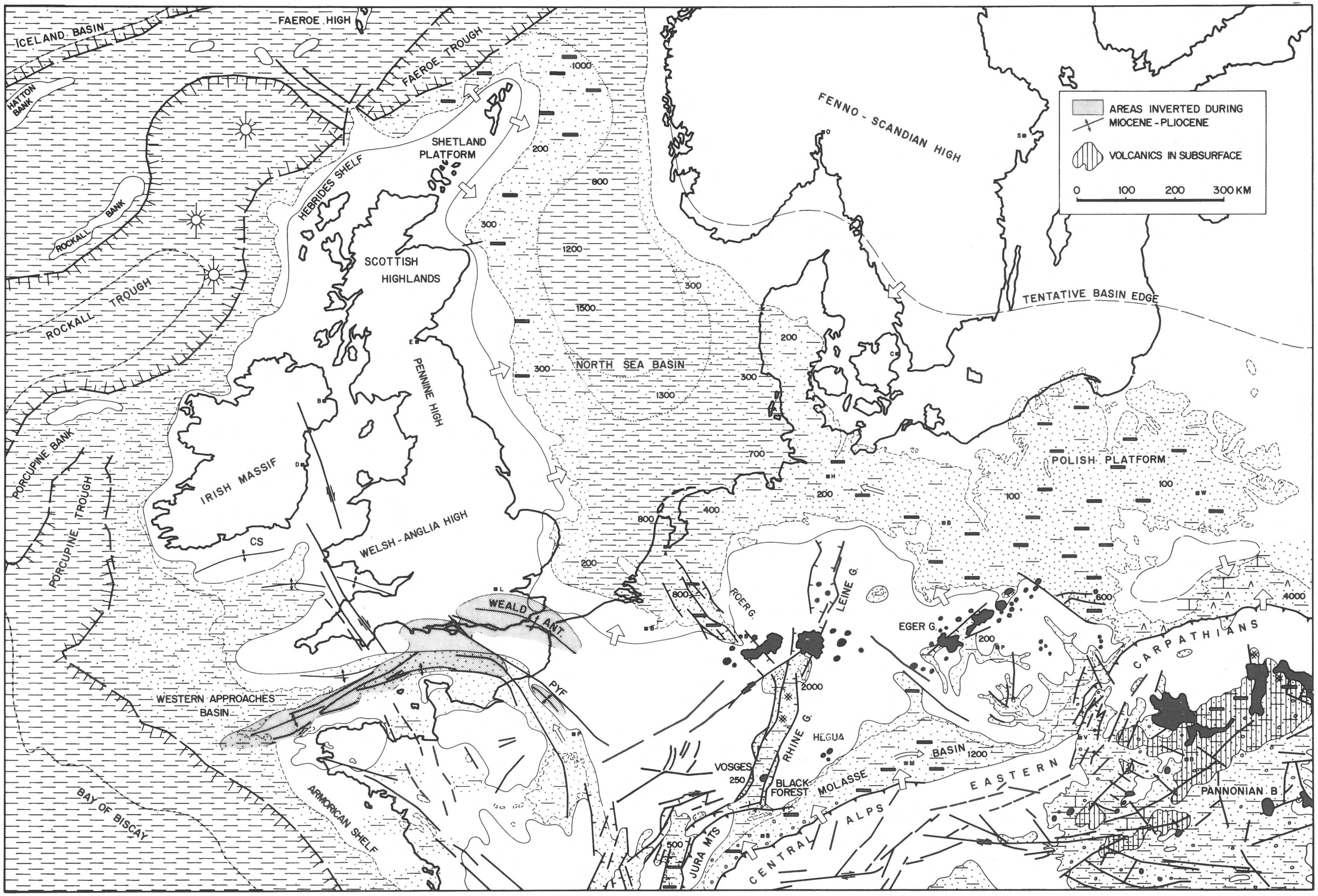

Fig. 7. Mio-Pliocene palaeogeography of Rhine Graben area (Ziegler 1990); symbols same as Fig. 5; black: plateau basalts; CS – Celtic Sea Basin; PYF – Pays-de-Bray Fault. Thicknesses in meters.

local development of collapse structures (Blès et al. 1989; Bergerat et al. 1990).

By the early *Miocene*, marine connections between the Alpine foreland basin, the Paris Basin and the Northwest European Basin were severed in conjunction with the thermal uplift of the Massif Central and the Rhenish Massif (Fig. 7). Despite continued crustal extension across the Rhine and Roer Valley grabens and possibly also the Leine Graben, subsidence patterns of the Leine Graben and the southern parts of the Roer Valley Graben were reversed and their sedimentary fill subjected to erosion. Uplift of the Rhenish Shield was paralleled by a sharp increase in volcanic activity (Lippolt 1983; Mertes & Schminke 1983). During the early and mid-Miocene, last temporary marine incursions reached the Mainz Basin, located at the northern end of the Rhine Graben; these advanced either from the north through the Roer Valley Graben or, more likely, originated from the Alpine foreland basin in the south (Struve 1973; Gramann & Kockel 1988). Thereafter, connections between the Rhine Graben and the Northwest European Basin were permanently interrupted (Fig. 4).

Uplift of the Vosges-Black Forest rift dome, straddling the southern Rhine Graben, started presumably during the early Miocene as indicated by a short-lived intra-Burdigalian hiatus. However, the main uplift phase commenced around 15 Ma during the late middle Miocene, as evident by the deposition of the 'Jura Nagelfluh' in the area of the Jura Mountains; these conglomerates are the erosion product of the former Mesozoic cover of the Vosges and Black Forest basement uplifts (Laubscher 1987, 1992). At about the same time, the southern parts of the Rhine Graben ceased to subside, as indicated by the disconformable contact between its late Oligocene and Pleistocene sedimentary fill; this hiatus testifies to a major reversal in the subsidence pattern of the southern Rhine Graben and the erosion of up to 1500 m of syn-rift sediments (Fig. 8; Brun et al. 1992). The main uplift phase of the Vosges-Black Forest arch was slightly preceded by the development of the Kaiserstuhl shield volcano, located in the southern, axial parts of the Rhine Graben (18–13 Ma), and coincided with the development of the Hegau volcanic field on the southeastern flank of the Black Forest uplift (13–6 Ma; Horn et al. 1972; Baranyi et al. 1976) and of the Jura fold and thrust belt around 13–7 Ma (Laubscher 1992). In contrast, the northern parts of the Rhine Graben continued to subside during the Miocene such that up to 1500 m of shales, sand and minor carbonates and halites were deposited (Illies 1974; Sittler 1969a, 1969b; von Eller & Sittler 1974; Roll 1979). Late Eocene to Miocene crustal extension across the Rhine Graben was paralleled by dextral reactivation of the Permo-Carboniferous South Hunsrück-Taunus border fault zone (Anderle 1987; Schwab 1987) and by sinistral wrench faulting along the Burgundy transfer zone (Bergerat 1977; Rat 1978).

Subsidence of the Saône, Bresse and Limagne grabens was interrupted during the latest Oligocene (?) to middle Miocene, presumably in conjunction with thermal doming of the Massif Central, that was paired with an increase of volcanic activity (Blès et al. 1989; Chenevoy 1974; Rat 1974; Brousse & Bellon 1983).

During the *late Miocene* and *Plio-Pleistocene*, the various segments of the Rhône-Rhine-Roer Valley rift system show major differences in their evolution. This is thought to be a consequence of changes in the regional stress regime and the development of the present stress field which is characterized by northwest directed maximum compressive horizontal stress trajectories (Illies & Greiner 1978; Villemin & Bergerat 1987; Bergerat 1987a, 1987b; Blès & Gros 1991; Cloetingh & Kooi 1992; Müller et al. 1992).

Subsidence of the Bresse Graben resumed again during the late Miocene. Its eastern margin was overridden by the frontal thrust elements of the Jura Mountains during the latest Miocene; these are sealed by poorly dated Pliocene sediments (Chauve et al. 1988; Guellec et al. 1990). At present, the Bresse and Limagne grabens are being uplifted; they contain some 2500 and 3000 m of Cenozoic sediment, respectively. On the other hand, the Saône, Alès and Gulf of Lions grabens apparently ceased to subside at the end of the early Miocene, possibly as a consequence of a stress release at the moment crustal separation was achieved in the Provençal Basin (Rehault et al. 1985); these grabens contain about 2000 m of syn-rift sediments (Blès et al. 1989). In the Massif Central volcanic activity peaked during the Pliocene and Pleistocene and persisted into Holocene times; this was accompanied by strong uplift of the entire Massif Central (Lucazeau & Bayer 1982). Structural data indicate that the grabens of the Rhône Valley and the Massif Central were affected by a late Miocene-Pliocene phase of E-W to SE-NW compression, causing partial inversion of the Alès and the eastward adjacent Manosque half-grabens (Blès et al. 1989; Roure et al. 1992). The present stress field of France is characterized by northwest directed maximum compressive horizontal stress trajectories (Bousquet & Phillip 1981; Blès & Gros 1991; Cornet &

Burlet 1992). Neotectonic analyses indicate that the Massif Central is at present subjected to transtensional deformation (Delouis et al. 1993). Precision leveling shows that uplift particularly of the eastern parts of the Massif Central continues at rates of up to 1.75 mm a^{-1} whereby differential movements between blocks separated by ENE-trending discontinuities can be recognized (Lenotre et al. 1992).

In contrast, the grabens of the Gulf of Lions, which were overstepped by the Miocene and younger passive margin prism of the Provençal Basin, show little evidence for tectonic reactivation (Burrus et al. 1987; Gorini et al. 1993). However, during the Plio-Pleistocene a volcanic chain extending from the Cevennes to the Mediterranean coast came into evidence (Maury & Varet 1980). Quantitative analyses show that subsidence of the Gulf of Lions accelerated during the Plio-Pleistocene, presumably in response to the build-up of the present stress field (Cloetingh & Kooi 1992). On the other hand, the Valencia Trough was tensionally reactivated during the Plio-Pleistocene whereby volcanic activity resumed (Banda & Santanach 1992; Maillard & Mauffret 1993; Torres et al. 1993).

During the Pliocene and Pleistocene the Vosges and Black Forest were further uplifted; this was paralleled by slow intermittent subsidence of the southern parts of the Rhine Graben in which truncated Oligocene sediments are covered by a relatively thin veneer of Quaternary gravels. In contrast, the northern parts of the Rhine Graben continued to subside during late Miocene to Pleistocene times; here, Plio-Pleistocene sediments, accumulating in depositional continuity with Miocene series, locally attain thicknesses of up to 1000 m (Fig. 8; Brun et al. 1992). Under the present, northwest-directed maximum horizontal compressive stress system the Rhine Graben is subjected to sinistral shear. Indeed, small-scale upthrusts, reported from the eastern graben margin north of Karlsruhe, testify to its transpressional reactivation (Illies & Greiner 1976, 1978; Becker et al. 1987; Larroque et al. 1987; Müller et al. 1992). However, continued subsidence of the northern parts of the Rhine Graben must be related to transtensional deformations (Laubscher 1992). Cenozoic sediments reach maximum thicknesses of 3000 m in the northern part of the Rhine Graben and 2000 m in its southern part; the polarity of the graben changes from predominantly west-hading in the south to east-hading in the north at the latitude of Offenburg (Fig. 8).

The Roer Valley Graben, which strikes parallel to the northwest-directed maximum horizontal compressional axis of the present stress field, feathers out to the northwest and looses its identity before it reaches the shores of the North Sea (Illies & Greiner 1978; Ahorner et al. 1983; Zagwijn & Doppert 1978; Zagwijn 1989; Remmelts & Duin 1990). In the axial parts of the Roer Valley Graben, Cenozoic sediments attain maximum thicknesses of some 2000 m (Geluk 1990). Quantitative subsidence analyses show that the Roer Valley Graben subsided slowly during Miocene times and that subsidence rates accelerated during the Pliocene and Pleistocene, presumably in response to the build-up of the present stress field (Zijerveld et al. 1992).

From Miocene until Recent times the Rhenish Shield was progressively updomed. Volcanic activity affected a broad area and shifted during the Quaternary to the Eifel, located west of the Rhine, where the Laacher See volcano erupted 11 000 years before present (Lippolt 1983; Mertes & Schminke 1983; see also Fuchs et al. 1983; Schminke et al. 1990). Precision leveling indicates that the area of the Rhenish Shield, straddling the Rhine River, is presently being uplifted at a rate of 0.4–0.6 mm a^{-1}, with maxima of over 1 mm a^{-1} being reached in the Eifel. In the northern part of the Rhine Graben, present subsidence rates reach maxima of 1 mm a^{-1} between Karlsruhe and Frankfurt. The southern parts of the Rhine Graben are more or less stable (– 0.1 to + 0.2 mm a^{-1}; DKG-ARH 1979; Prodehl et al. 1992). At present, broad areas around the Rhine-Roer Valley graben system and the Eger Graben to the east are seismically active and have been the locus of some rather destructive historical earthquakes (Ahorner et al. 1983; Fuchs et al. 1983; Larroque et al. 1987; Grünthal et al. 1990).

The Eger Graben, which started to subside during the late Oligocene, remained active till sub-recent times; it consists of a string of small, fault-bounded basins, containing some 500 m of Cenozoic sediments, which are separated by major volcanic complexes. After an initial Paleocene phase of melilitic dyke intrusion, the first major episode of volcanism straddled the Oligocene-Miocene boundary; a second phase is dated at $\pm$ 9 Ma and the last phase spanned 2.8 to 0.8 Ma. Basin subsidence generally followed the termintion of each volcanic cycle (Malkovsky 1980, 1987; Suk et al. 1984). Late Miocene marine transgressions, originating from the Carpathian foreland basin, advanced northward through valley systems into the southern parts of the Bohemian Massif, and thus indicate that the area was still located near sea level. The Plio-

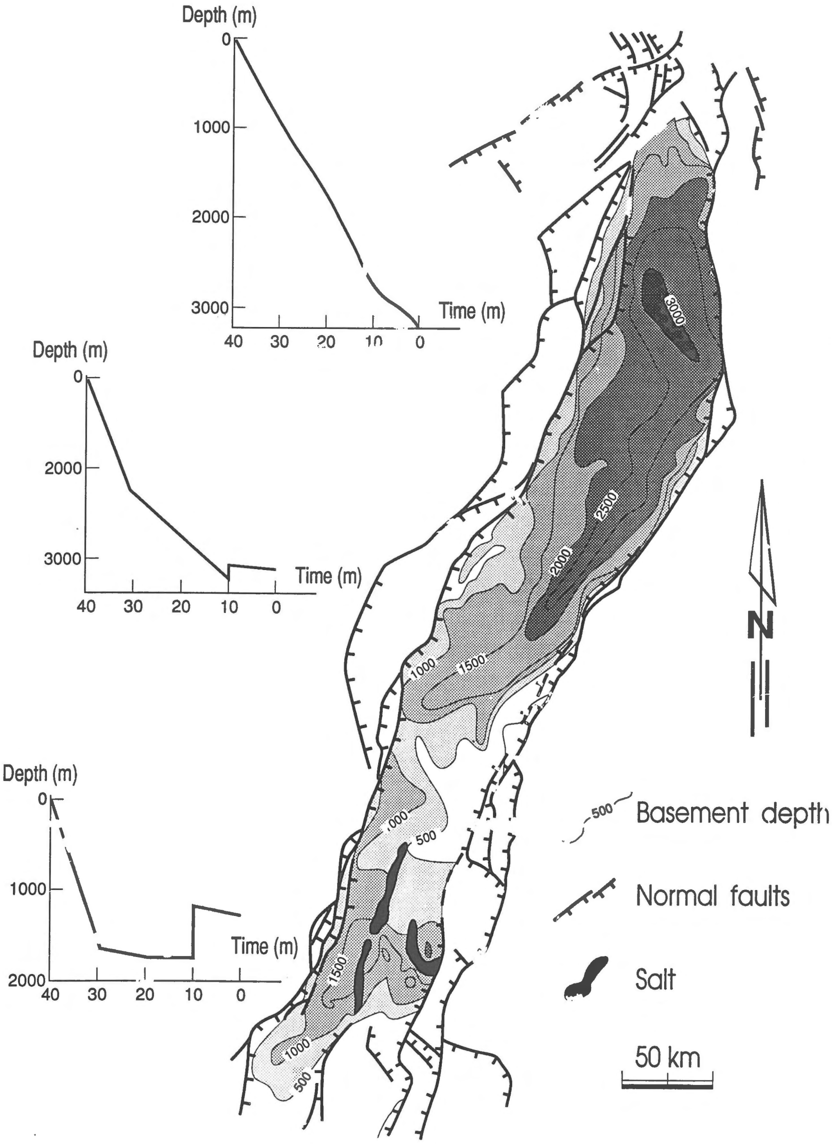

Fig. 8. Isopach map of Tertiary sedimentary fill of Rhine Graben (contour values in meters) and representative Cenozoic tectonic subsidence curves for the northern, central and southern parts of the Rhine Graben (time axis in Ma) (after: Doebl & Obrecht 1974; Brun et al. 1992).

Pleistocene volcanic cycle was associated with doming of the Bohemian Massif. Geodetic data show that the western parts of the Eger Graben are still rising whereas much of the southern parts of the Bohemian Massif are slowly subsiding (Suk et al. 1984). The ENE-striking Eger volcano-tectonic zone extends over a distance of more than 300 km from the southwestern Bohemian Border Zone into Poland. Evolution of the Eger Graben was paralleled by tectonic reactivation of fault systems delimiting the Bohemian Massif to the southwest and northeast (Schröder 1987), presumably involving minor wrench displacements. Similar wrench deformations gave rise to the subsidence of grabens and half grabens in the area of Lower Silesia (Pozaryski 1977) and to the development of the Upper Moravian Graben, which is superimposed on the foreland and the externides of the Carpathian fold belt; its development is probably related to the subsidence of the intra-Carpathian Pannonian-Vienna Basin collapse system (Tari et al. 1992).

The chemistry of volcanic rocks associated with the Cenozoic rifts in the Alpine foreland is typically alkaline and bimodal felsic-mafic; melts were derived by partial melting from the upper asthenosphere and the lower lithosphere; magma mixing, fractional crystallization and crustal contamination played a significant role in the diversity of extruded magmas. In the face of apparently limited lithospheric stretching, it must be assumed that the temperature of the asthenosphere and the volatile content of the mantle-lithosphere were elevated in order to explain the volume of melts generated; based on geochemical criteria, a possible contribution from deep mantle plumes cannot be excluded (Downes 1987; Wilson & Downes 1992). Magmatic activity is at present sub-active as evident by the age of the youngest extrusive rocks, widespread hydrothermal activity and frequent CO_2 and mantle helium emanations (Oxburgh & O'Nions 1987).

Rift-related uplift of the Massif Central, the Vosges-Black Forest area, the Rhenish Shield and the Bohemian Massif, combined with wrench-related upwarping of the Burgundy area, the Armorican Massif and also of the Ardennes, caused regional truncation of their Mesozoic and Cenozoic sedimentary cover and hence the erosional partial isolation of the Paris Basin (Mégnien 1980) and the South German Franconian Platform (Ziegler 1990).

Crustal and lithospheric configuration

Geophysical data show that the Cenozoic rifts in the Alpine foreland are associated with a marked uplift of the Moho-discontinuity (Fig. 9). Zones of maximum crustal thinning coincide with the Massif Central, the Vosges-Black Forest area and the trace of the northern Rhine-Leine Graben. The Burgundy transfer zone corresponds at the Moho level to a northeasterly trending positive axis. The Eger Graben area is associated with a broad uplift of the Moho. In places the Moho is apparently underlain by an anomalous, low-density, low-velocity upper mantle (Perrier & Ruegg 1973; Coisy & Nicolas 1978; Edel et al. 1975; Fuchs et al. 1981, 1987; Mechie et al. 1983; Gajewski et al. 1987; Prodehl 1989; Prodehl et al. 1992; Zeis et al. 1990; Bois 1993).

Published maps suggest that in the area of the Massif Central, the southern Rhine Graben and the Rhine-Leine-Roer Valley Graben triple junction the asthenosphere/lithosphere boundary is domed up to a level of 50 to 60 km below the surface and that it descends laterally to depths ranging between 80 to 120 km (Babuska et al. 1987; Giese 1983; Prodehl 1989; Prodehl et al. 1992; Suhadolc et al. 1990). Particularly in the case of the southern parts of the Rhine Graben available refraction- and reflection-seismic and gravity data can be interpreted either in terms of a tensional-failure model of lithospheric extension or in terms of a mantle-plume model, both of which could explain the observed crustal doming and the associated gravity anomalies (Fig. 10; St. Müller, pers. comm.). The mantle-plume model proposes that progressive crustal doming is caused by lithospheric extension and thermal thinning of the sub-crustal lithosphere over upwelling mantle diapirs or plumes, possibly involving small-scale convection in the latter. The tensional-failure model assumes that in areas of lithospheric stretching melts have diapirically risen from the base of the lithosphere to the crust/mantle boundary where they spread out laterally upon finding their density equilibrium, thus forming an asthenolith causing doming of the overlying crust (Turcotte 1981; Turcotte & Emerman 1983; see also Wilson & Downes 1992). These rather simplistic models must be considered as end-members which are unlikely to do justice to the true natural complexities. Therefore, it is likely that crustal doming generally involves a combination of both mechanisms.

For the *Massif Central* refraction-seismic and gravity data show that the Moho-discontinuity rises from depths of 30–32 km under its flanks to about 24 km in the area of the Limagne Graben and the adjacent Chaîne des Puys (Fig. 9) and that it is underlain by an anomalously low velocity upper mantle (7.4 km s^{-1}); with increasing depth mantle velocities increase to 8.4 km s^{-1} at about 50 km. This indicates that the central parts of the Massif Central are underlain by an asthenolith that is apparently detached from the upwarped asthenosphere (Perrier & Ruegg 1973; Sapin & Prodehl 1973; Coisy & Nicolas 1978; Lucazeau & Bayer 1982; Bergerat et al. 1990; Prodehl 1989; Prodehl et al. 1992; Bois 1993). Mantle tomography indicates that in the area of the Limagne Graben the asthenosphere/lithosphere boundary is located at a level of about 80 km, descends under the flanks of the Massif Central to depths of about 120 km, dips gently under the western Alps, and reaches values of some 180 km beneath the western Po Valley (Spakman 1990). Moreover, tomographic data show that beneath the Massif Central the upper mantle is characterized by an anomalously low velocity (Granet et al. 1994). Crustal extension across the Massif Central, which has a structural relief of at least 1200 m, gave rise to the subsidence of a complex system of fault-bounded

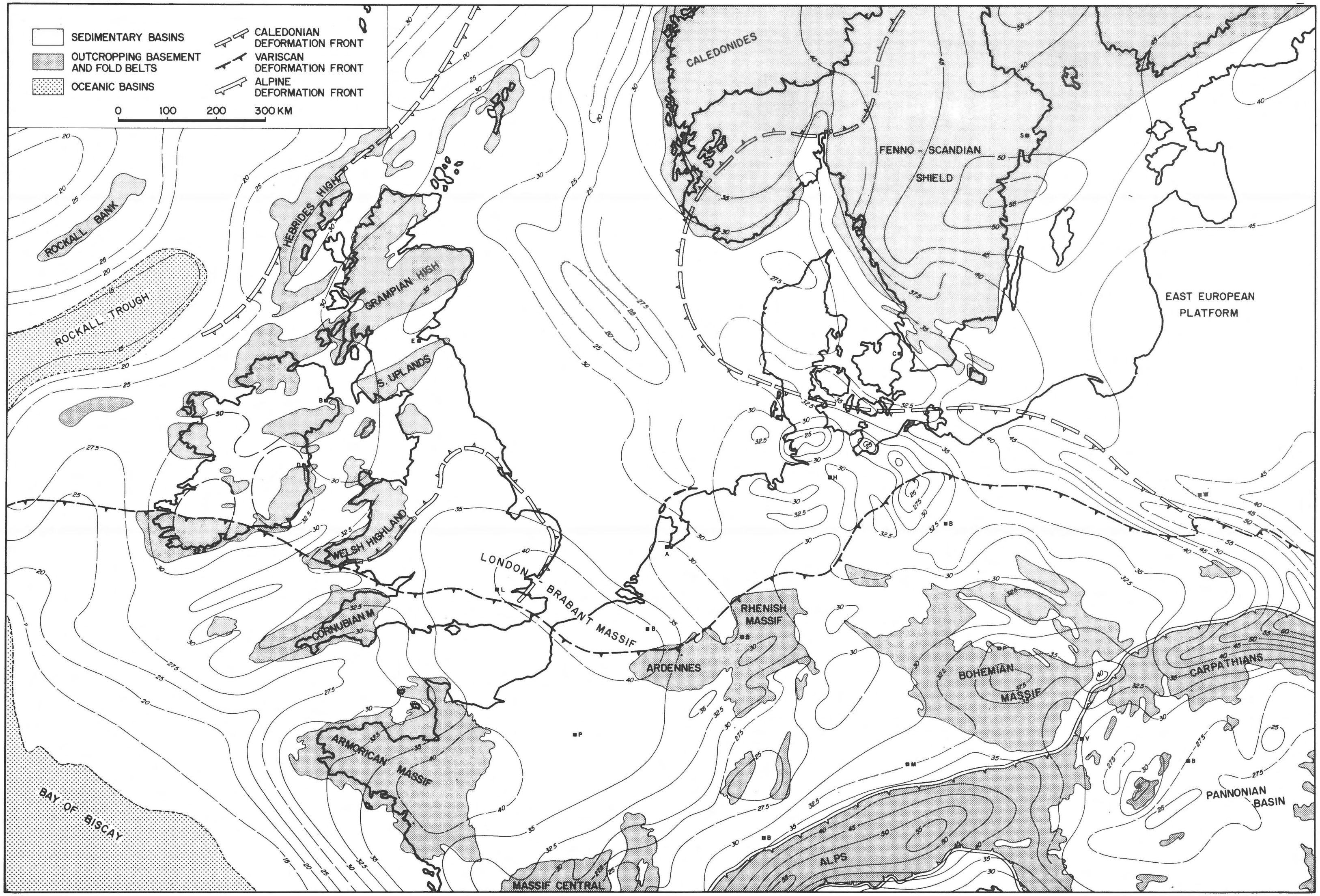

Fig. 9. Depth map of Moho-discontinuity in Rhine Graben area (Ziegler 1990); contour values in kilometers; barbed interrupted black line: Variscan deformation front; barbed interrupted open line: Caledonian deformation front; barbed continuous open line: Alpine deformation front; stippled areas: outcropping Palaeozoic massifs.

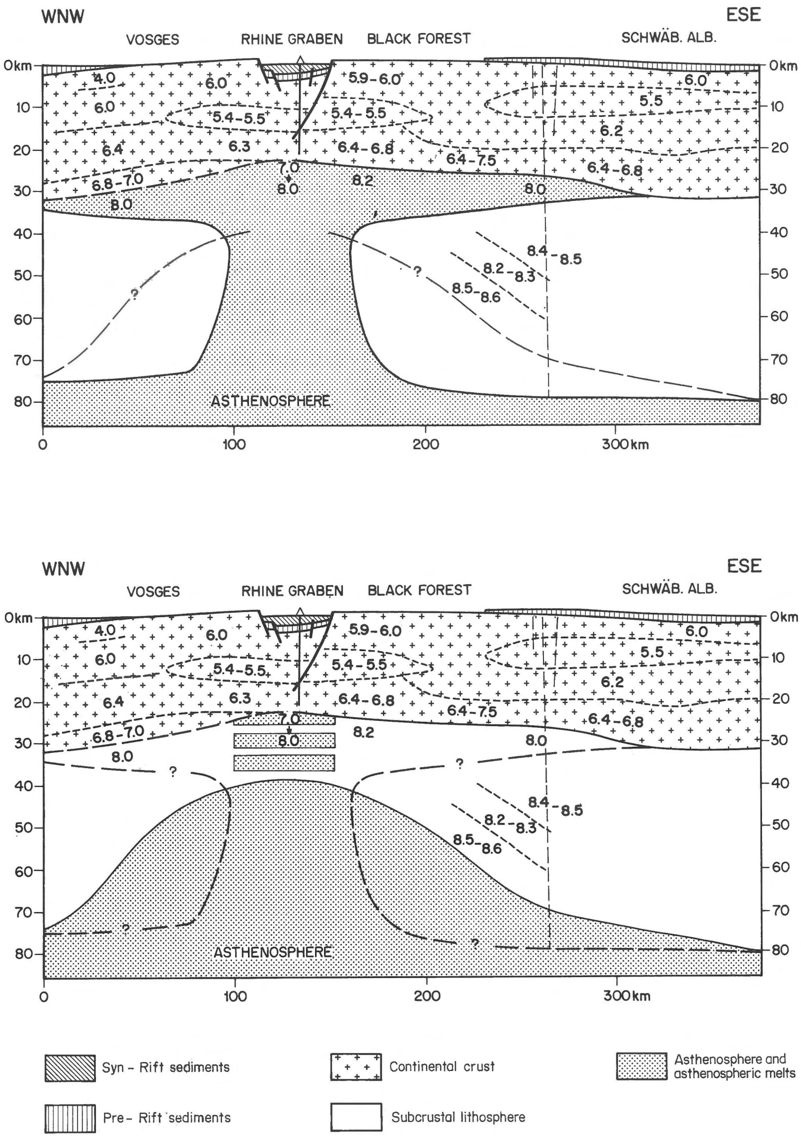

Fig. 10. Lithospheric configuration of southern Rhine Graben, alternative interpretations along Tensional Failure (top) and Mantle Plume (bottom) models (modified after Edel et al. 1975, Prodehl 1989 and Prodehl et al. 1992).

basins amongst which the Limagne, Roanne, Forez and Bresse grabens are the most important. As the amount of upper crustal extension across the Limagne-Bresse graben system has not yet been quantified, we are unable to assess whether the observed crustal thinning results from lithospheric stretching alone or whether magmatic destabilization of the Moho has contributed toward it. Only the Bresse Graben, which is located on the eastern flank of the Massif Central arch, is covered by a deep reflection-seismic profile, which shows clear lamination of the lower crust (Bergerat et al. 1990). Gravity modelling supports the notion that development of the Massif Central dome has to be explained by a combination of the mantle plume and tensional failure models (Blès et al. 1989).

For the *Rhine Graben* area refraction- and deep reflection-seismic data show that the Moho-discontinuity rises from a depth of 28 km under the Franconian Platform to 24–25 km under the graben and descends to 30 km west of the Vosges; towards the north the amplitude of the Moho uplift, which is clearly aligned with the axis of the Rhine Graben, decreases

(Fig. 9). A special feature of the Moho-topography is a secondary, northeasterly striking axis which extends from the Burgundy transfer zone across the Vosges and Black Forest into the area of the Franconian Platform (see Prodehl et al. 1992; Rousset et al. 1993). Subcrustal velocities are in the 8.0–8.2 km s^{-1} range. Within the mantle-lithosphere, low-velocity zones are evident only under the graben and the Franconian Platform. Crustal uplift commences along the flanks of the Vosges-Black Forest arch at a considerable distance away from the zone of crustal extension in the Rhine Graben. Similarly, reflectivity of the lower crust generally increases under the flanks of the Vosges and Black Forest towards the graben margins; however, under the Rhine Graben the reflectivity of the lower crust decreases possibly as a result of energy absorption in the sedimentary fill of the graben (Prodehl et al. 1992; Brun et al. 1992). In the area of the Black Forest, the top of the highly reflective lower crust coincides with the base of a high-conductivity, low-velocity layer that, in itself, is characterized by low reflectivity (Fuchs et al. 1987). A similar, mid-crustal low-velocity zone is also evident under the Vosges (Fig. 10). Low-velocity zones within both the crust and the upper mantle are interpreted as being associated with zones of fluid inclusions, residual thermal anomalies and possible partial melts. One of the most prominent intra-crustal anomalies is observed on the eastern flank of the Black Forest in the area of Urach; this feature corresponds to a major thermal anomaly that coincides with a mid-crustal low-velocity, high-density and high-reflectivity anomaly (Meissner 1986; Gajewski & Prodehl 1987; Gajewski et al. 1987; Fuchs et al. 1987; Lüschen et al. 1987; Prodehl et al. 1992).

Upper crustal extension across the Rhine Graben, as estimated from reflection-seismic data, ranges from 5 to 7 km (Laubscher 1970; Sittler 1969a; Doebl & Teichmüller 1979; Villemin et al. 1987; H. Durst pers. comm.). The crustal configuration of the southern Rhine Graben, as defined by refraction-seismic data and the deep ECORS-DEKORP reflection profiles, indicates an extension value of 17 km for the lower crust (Brun et al. 1992), provided it is assumed that the pre-rift crust had a uniform thickness and that its volume was preserved during the rifting process. This discrepancy between stretching values determined at upper crustal levels and at the base of the crust is thought to be indicative of a rift-induced destabilization of the crust/mantle boundary (Ziegler 1992, 1993).

The *Vosges-Black Forest* dome has a structural relief of some 2500 m and, in an east-west direction, a width of over 300 km. It straddles the Rhine Graben almost symmetrically. Manifestations of surface volcanic activity are confined to the area that is characterized by an elevated crust/mantle boundary, but are concentrated on the eastern flank of the dome. Petrological data indicate that volcanic rocks, both within the rift zone and on its flanks, are derived from a source that is located at depths of 70–100 km, corresponding to the base of the lithosphere (J. Keller pers. comm.; Wilson & Downes 1992); this is in keeping with the geophysically derived lithospheric configuration of the southern Rhine Graben area (Prodehl et al. 1992).

According to teleseismic data the asthenosphere/lithosphere boundary is not significantly uplifted in the area of the Vosges-Black Forest arch. However, the mantle-lithosphere shows a complex structure including small-scale low-velocity anomalies which are aligned to depths of 65 km with the graben axis and at the depth range of 75–125 km display an easterly trend; these can be interpreted as being related to the intrusion of partial melts during the middle and late Miocene doming stage of the southern Rhine Graben area (Prodehl et al. 1992; Glahn & Granet 1992). Although this data, as well as gravity modelling, contradicts the presence of a massive, low-density body in the uppermost mantle beneath the southern parts of the Rhine Graben, as previously suggested by the studies of Kahle & Werner (1980), uplift of the Vosges-Black Forest dome appears to be at least in part related to inflation of the mantle-lithosphere by partial melts that have risen from the asthenosphere/lithosphere boundary (tensional failure model, Fig. 10). However, in order to explain the elevation of this arch, buoyancy effects of hot, upwelling asthenosphere must be invoked (Lyon-Caen & Molnar 1989; Rousset et al. 1993). In this respect it is interesting to note that results of seismic mantle tomography indicate for the southern parts of the Rhine Graben the presence of a significant low-velocity zone in the 150–200 km depth range (Spakman 1990). Although the position of the asthenosphere/lithosphere boundary is poorly defined in the area of the Vosges-Black Forest arch, it must be assumed that it has at least some relief, particularly in the area of its northeast trending Moho axis.

For the central parts of the *Rhenish Shield* arch, which straddle the Rhine-Roer-Leine graben triple junction, geophysical data indicate that the asthenosphere/lithosphere boundary is uplifted to a level of 45–50 km and that an anomalous low-velocity upper mantle extends to depths of about 200 km. The mantle lithosphere has a remnant thickness of 15–20 km in the crestal parts of this arch and appears to thicken rapidly towards its flanks (Prodehl et al. 1992). Crustal thicknesses vary between 30 and 35 km. The crust and mantle-lithosphere display a complex structure, including several low-velocity and high-conductivity zones. Lamination of the lower crust appears to be confined to the area of Cenozoic volcanic activity and uplift, as seen on the DEKORP-1 and 2-N reflection profiles (DEKORP Research Group 1990). For the Rhenish Shield rift arch, upper crustal extension values cannot be quantified due to intense erosion of syn-rift sediments that were deposited during the Oligocene in the southern parts of the Roer Valley and Leine grabens prior to their uplift. Crustal thinning is spread over a wide area whereas the axis of the Moho culmination follows the trace of the Leine Graben and projects northward to the margin of the Northwest European Basin (cf. Figs 7, 9; Mechie et al. 1983; Larroque et al. 1987; Prodehl 1989). A discrete domal uplift of the Moho, as associated with the Vosges-Black Forest arch, is not evident. Rift-induced thermal uplift of the Rhenish Shield, which commenced at the beginning of the Miocene, is difficult to quantify but probably lies in the range of 400 to 650 m, as indicated by the present level of the base Tertiary horizon, defined by erosional remnants of Oligocene sediments (Meyer et al. 1983).

Geophysical data suggest that doming of the Rhenish Shield was governed by progressive thinning of the mantle-lithosphere (mantle plume model dominant), whereas the intrusion of melts into the lithosphere may have played a more important role in the uplift of the Vosges-Black Forest arch (tensional failure model?). In this context it must be kept in mind that if the rate of thermal uplift (in response to thermal thinning of lithosphere and/or replacement of an asthenolith) exceeds the rate of isostatic subsidence of an axial graben (in response to lithospheric extension), the subsidence of the rift can be reversed. At present the southern parts of the Leine Graben are elevated to a level of some 500 m, whereas the southern parts of the Rhine Graben have returned, after a late Miocene to Pliocene phase of uplift and erosion, to the erosional base-level during the Quaternary. Uplift of the Rhenish Shield still continues, while the southern Rhine Graben area appears to be stable.

In the non-volcanic *Roer Valley Graben*, upper crustal extension by faulting amounts to some 600 m whereas its crustal configuration suggests a stretching value of 1 to 2 km. This discrepancy may be explained by the complex evolution of the Roer area which had undergone Permo-Triassic and late Jurassic-early Cretaceous stretching events and a latest Cretaceous-Paleocene phase of basin inversion prior to the Oligocene onset of rifting (Geluk 1990; Remmelts & Duin 1990; Zijerveld et al. 1992). Unlike the Leine Graben, the Roer Valley Graben is seismically very active and continues to subside in response to sinistral transtensional deformation of the crust (Geluk et al. 1994).

The area of the *Eger volcano-tectonic zone* coincides with a relatively narrow positive gravity anomaly which is superimposed on a broad uplift of the crust/mantle boundary; however, the lithosphere has apparently not been thinned substantially and the lithosphere/asthenosphere boundary is located at a depth of 90 to 100 km (Suk et al. 1984; Malkovsky 1987; Babuska et al. 1987). A tensional-failure model may apply for the Eger Graben for which neither detailed crustal profiles nor estimates of the magnitude of upper crustal extension are available. The area corresponds to a zone of increased seismic activity (Kárník et al. 1991); precision leveling indicates that the region of the Doupovské Hory stratovolcano is being uplifted at rates of up to 2 mm.a^{-1} (Suk et al. 1984).

Overall, the crustal and lithospheric configuration of the Rhine Graben rift system speaks in favour of a depth-dependent, pure-shear extensional model in which lower crustal thinning is largely confined to the relatively narrow zones of upper crustal extension by faulting, and attenuation of the sub-crustal lithosphere is spread over a considerably wider area.

Reflection seismic data and the focal depths of earthquakes indicate that in the Rhine Graben the brittle/ductile transition zone is located at a depth of 15–20 km, coinciding with the top of the lower crust which appears to deform by ductile shear (Bonjer et al. 1984; Brun & Gutscher 1992). On the other hand, simple-shear deformation of the crust, involving lower crustal detachment horizons, may have played a significant role in the development of the nearly 200 km wide extensional field of the Massif Central (Bergerat et al. 1990). In the process of crustal thinning, interaction of mantle-derived melts with the lower crust appears to have played an important role. Magmatic inflation of the mantle-lithosphere, including emplacement of asthenoliths, apparently contributed to the uplift of the Vosges-Black Forest, Massif Central and possibly also the Bohemian Massif. Significant thermal thinning of the mantle-lithosphere has apparently taken place in the areas of the Massif Central and the Rhenish Shield.

Geodynamic processes

The Eocene and younger tensional and compressional intra-plate deformations of the northwestern Alpine foreland are broadly synchronous and contemporaneous with the main and late phases of the Alpine orogeny, and at least in part with the main folding phases of the Pyrenees and the Atlantic-related wrench deformations of the British Isles (Ziegler 1990). In the western Mediterranean, Cenozoic rifting post-dates the Pyrenean orogeny but is synchronous with the development of the Betic-Balearic thrust belt (Banda & Santanach 1992; Roca & Desegaulx 1992; Torres et al. 1993; Bois 1993).

Micro-tectonic analyses indicate that the stress field affecting northwest Europe, as well as the entire Alpine-Mediterranean orogenic system, changed repeatedly during Cenozoic times (Letouzey & Trémolière 1980; Bergerat 1987b; Blès et al. 1989; Lacombe et al. 1990). This can be related to changes in the convergence patterns between Africa-Arabia and Europe that are reflected in the sea-floor spreading rates in the different segments of the Atlantic Ocean, the Norwegian Greenland Sea and the Indian Ocean, as well as to early phases of a fundamental plate-boundary reorganization that may ultimately lead to the break-up of the present continent assembly. In the light of this, it is unlikely that a simple, unified kinematic regime, as proposed by Dewey & Windley (1988), could explain the sequence and patterns of Cenozoic deformations that are observed in the Alpine foreland of western and central Europe (see also Tapponnier 1977; Ziegler 1988, 1990).

Interplay of foreland compression and rifting

By analogy with the Senonian and mid-Paleocene compressional deformations of the northern and northwestern Alpine foreland (Fig. 2), the late Eocene to Miocene inversion of wrench- and rift-induced Mesozoic basins in the southern North Sea, the Channel and the Celtic Sea-Western Approaches areas, as well as the Mio-Pliocene wrench deformation of the Armorican Massif (Figs 5–7), are probably the effect of an intermittent build-up of horizontal compressional stresses which were transmitted from the Alpine collision front into the foreland. The observed systematic westward shift of compressional foreland deformations during late Eocene to Pliocene times is presumably the expression of an increasingly important dextral translation component in the late Alpine convergence of Africa-Arabia and Europe (Ziegler 1988). Stress related to the collision of Iberia and Europe probably interfered with stress transmitted from the Alpine collision zone, at least during the late Senonian to earliest Miocene Pyrenean orogeny, and might have played a significant role in the Eocene reactivation of the Permo-Carboniferous fracture systems controlling the localization of the Cenozoic rifts in the western Alpine foreland (Fig. 3). However, as the late Oligocene rifts of the Gulf of Lions transect the eastern part of the Pyrenean fold belt, their development cannot be related to stresses originating from the Pyrenean collision front.

In this context, it should be noted that the late Eocene-early Oligocene phase of foreland compression, which is even expressed in the Central North Sea, coincides with the initial rifting phases of the Alès, Saône, Bresse, Limagne and Rhine grabens (Fig. 5). Subsidence of these rifts was governed by northerly directed trajectories of maximum horizontal compression, whereas northwestward directed ones induced the inversion of, for instance, the Celtic Sea Trough. During the early Oligocene phase of compressional deformation of the West Netherlands Basin, the Rhine Rift propagated northward through the Leine Graben area to the margins of the Northwest European Basin (Fig. 6). Mid-Oligocene temporary relaxation of compressional foreland stresses is indicated by a remission/slow-down of inversion movements and reduced subsidence rates of the Rhine Graben. During the late Oligocene-early Miocene phase of foreland compression, which affected mainly the Channel area and the Western Approaches Trough, the Rhine Graben underwent a second phase of rapid subsidence and propagated northwestward, as indicated by the subsidence of the Roer Valley Graben (Fig. 6). On the other hand, the Saône, Limagne and Bresse grabens apparently ceased to subside during the early Miocene. Following the relaxation of these stresses, the Bresse Graben subsided again during the late Miocene and Pliocene. Stresses causing late Miocene folding of the essentially thin-skinned Jura Mountains had apparently no major effect on the Bresse Graben, although its eastern margin was overridden by the frontal thrust elements of this fold belt (Fig. 7; Guellec et al. 1990). During the Miocene and Pliocene, the northern parts of the Rhine Graben and the Leine Graben were uplifted and subjected to erosion. As a consequence of the Pliocene build-up of the present stress field, that is characterized by northwest-directed trajectories of maximum horizontal compression, the Bresse Graben ceased to subside, the Rhine Graben is now subjected to sinistral shear deformation, whereas the Roer Valley Graben continues to subside. The same stress field induced a broad-scale negative deflection of the lithosphere in the area of the North Sea Basin, as indicated by the Plio-Pleistocene acceleration of its subsidence rates (Ziegler 1990; Cloetingh & Kooi 1992).

Development of rift-related arches

Early Miocene doming of the Rhenish Shield and middle to late Miocene and Pliocene uplift of the Massif Central and the Vosges-Black Forest post-dates the onset of rifting by some 20 to 40 Ma, respectively. Uplift of these arches coincides with progressive uplift of the Alpine external massifs, involving imbrication of the European foreland crust, and in part also with the deformation of the Jura Mountains and the West Alpine Dauphiné zone (Laubscher 1992). In the face of this, it cannot be excluded that the build-up of compressive stresses, transmitted from the Alpine collision front into the foreland, caused buckling of the foreland-lithosphere, its positive deflection in areas were its elastic strength was weakened by rifting, and the reactivation of pre-existing crustal discontinuities. This concept is in keeping with the results of structural analyses which indicate that the area of the Massif Central and the grabens of the Rhône Valley were affected by a late Miocene-early Pliocene phase of E-W to SE-NW directed compression, as indicated, for instance, by the partial inversion of the Oligocene Alès and the eastward adjacent Manosque half-grabens (Blès et al. 1989; Blès et al. 1991; Roure et al. 1992); contemporaneous wrench deformations are also evident in the Armorican Massif and Paris Basin and underlie further inversion of the Western Approaches Trough (Fig. 7; Ziegler 1990). Similarly, the Plio-Pleistocene accelerated subsidence of the North Sea Basin must be explained in terms of a stress-induced lithospheric deflection (Cloetingh & Kooi 1992).

An alternative model is envisaged by Laubscher (1987, 1992), who interprets the southern Rhine Graben and Massif Central arches as flexural foreland bulges which developed in conjunction with the early Miocene Helvetic and the late Miocene Jura phases of the Alpine orogeny in areas where the elastic strength of the lithosphere was reduced by rifting; this gave rise to a greater deflection of the lithosphere, its inelastic yielding (Waschbusch & Royden 1992) and inflow of hot asthenosphere that may have been enhanced by progressive thrust-loaded down-flexing of the subducting foreland slab. In this context, it should be noted that volcanic activity in the area of the Vosges-Black Forest arch peaked during its early uplift phases (18–17 Ma) and waned during the folding of the Jura Mountains (13–7 Ma); however, the major subsidence reversal of the southern Rhine Graben (10 $\pm$ 2 Ma) does coincide closely with the Jura folding phase (Laubscher 1992; Villemin et al. 1987).

Both of the above-discussed processes are only indirectly rift-related. However, they may have contributed to a varying degree to the uplift of the Vosges-Black Forest and Massif Central arches and possibly also of the Bohemian Massif by causing positive deflections of the lithosphere, thereby contributing to additional decompressional melting near its base, resulting in an increase of the thermal load on the lithosphere. These arches are characterized by a structural relief in the range of 1000–2000 m and a considerably smaller degree of lithosphere thinning than seen in the Rhenish Shield. In this respect, the north-easterly trending Moho uplift which is associated with the Burgundy transform zone and which crosses the Vosges-Black Forest arch (Prodehl et al. 1992) is of particular interest, as it may have originated either as a compressional or a flexural lithospheric deflection.

Overall, it must be kept in mind that uplift of the different arches of the Cenozoic rift system was not synchronous. Neither is the level and timing of the associated surface volcanic activity comparable, nor can the same uplift mechanisms be invoked for all of them. For instance, uplift of the Rhenish Shield clearly pre-dates uplift of the Vosges-Black Forest arch; volcanism ceased in the southern parts of the Rhine Graben about 13 Ma ago whereas in the Massif Central it peaked between 7–1 Ma and is now sub-active; similarly volcanism is sub-active on the Rhenish Shield. Uplift of the Rhenish Shield, the Massif Central and the northwestern parts of the Bohemian Massif continues at present, whereas the Vosges-Black Forest arch appears to be more or less stable.

For the development of the Rhenish Shield arch preference is given to the mantle-plume hypothesis, which invokes progressive lithospheric stretching, upwelling and decompressional melting of the asthenosphere and at some point small-scale convective flow in the rising mantle diapir, causing accelerated thermal thinning of the sub-crustal lithosphere and progressive doming of the rift zone. This mechanism, in combination with lithospheric buckling related to the present stress field, may be responsible for the continued uplift of the Rhenish Shield. In this respect it should be kept in mind, that earthquake focal mechanisms indicate that Plio-Pleistocene uplift of the Ardennes involves compressional reactivation of Hercynian thrust faults (Camelbeeck & van Eck 1994).

Uplift of the Massif Central, Vosges-Black Forest and Bohemian Massif arches was clearly preceded by crustal extension that was followed at a later stage by positive deflection of the lithosphere in response to such processes as compression-induced buckling, thrust-load-induced flexure and development of a melt diapir, or a combination thereof. The volcanic record of these arches indicates that at the base of the lithosphere partial melting and melt segregation had occurred already during their initial rifting stages. However, volcanism peaked during the uplift stages of the respective arches, suggesting accelerated diapiric ascent of melts, permeating the mantle-lithosphere and/or accumulating near the base of the crust to form an asthenolith, and intruding the crust. This was accompanied by thermal expansion of the lithosphere, contributing to the uplift of these arches (Villemin et al. 1987). Melt migration took presumably place via a complex zone of intermittently active feeder channels and not via a coherent asthenospheric plume as shown schematically in Fig. 10 (tensional failure model). Asthenoliths are thought to consist of a mixture of indigenous lithospheric material and partial melts, derived from near the base of the lithosphere. Melt diapirism appears to have played an important, if not dominant role during the uplift of the Massif Central; this process is probably still active today (Lucazeau & Bayer 1982). On the other hand, melt diapirism played probably a less important role in the uplift of the Vosges-Black Forest arch. Since this dome has apparently ceased to rise, it must be assumed that the underlying mantle-lithosphere and asthenospheric anomalies are not further evolving and have started to cool (Prodehl et al. 1992; Glahn & Granet 1992; Rousset et al. 1993). For the Bohemian Massif, which is characterized by a broad Moho uplift underlying its northern parts, a relatively thick lithosphere ($\pm$ 90 km), young volcanic activity and persistent uplift, geophysical data available to the author do not permit to assess the mechanisms of its post-Miocene arching.

Geodynamic models

A number of geodynamic models have been advanced in an attempt to explain the evolution of Cenozoic rifts in the Alpine foreland of western and central Europe. These range from hot-spot-driven 'active' rifting models (Cloos 1939; Duncan et al. 1972; Burke et al. 1973) to 'passive' rifting models which propose that lithospheric extension is driven by far-field stresses. Such stresses may be related, for instance, to the collisional coupling of the Alpine orogen and its foreland, causing splitting of the latter (Molnar & Tapponnier 1975; Sengör et al. 1978; Bergerat & Geyssant 1980; Dewey & Windley 1988); alternatively, changes in the geometry of the Alpine subduction zones may be associated with the development of back-arc rifts (Jowett 1991). On the other hand, Ziegler (1988) postulates that during the late Cretaceous and Cenozoic phases of the Alpine orogeny lateral motions between the colliding African-Arabian and Eurasian plates contributed to the evolution of rifts in the Alpine forelands and that gradual assertion of a new cycle of plate boundary reorganization may ultimately cause the break-up of the present continent assembly.

Hot-spot model

The hot-spot model of 'active' rifting proposes that rift systems evolve in response to deviatoric tensional stresses developing in the lithosphere over mantle-plumes which rise from the lower mantle or the core-mantle boundary. Such rift systems should be characterized by early doming of the rift zone, relatively minor crustal extension in response to arching of the lithosphere and by major volcanic activity whereby primitive mafic rocks bear the distinct geochemical and isotopic signature of ocean island basalts (Artemjev & Artyushkov 1971; Dewey & Burke 1974; Bott & Kusznir 1979; Bott 1992; Wilson 1989).

The evolution of the Saône-Limagne-Bresse-Rhine rift system shows that crustal extension, accompanied by relatively minor volcanic activity, preceded by 20 to 40 Ma the development of discrete arches that could be related to the activity of localized deep-seated mantle plumes. Some branches of the European Cenozoic rift system are characterized by a low level of volcanic activity (e.g. Rhine Graben) or are totally a-volcanic (e.g. Bresse and Roer Valley grabens). Moreover, the amount of upper crustal extension across the Rhine Graben and the Massif Central is significantly larger than could result solely from uplift of a thermal dome. Correspondingly, uplift of these arches cannot be considered as the driving mechanism of rifting but is rather a secondary effect of lithospheric extension (Ziegler 1993). Finally, the shift of volcanic activity in time and space does not conform with the drift pattern of the European craton over a system of deep-seated, stationary mantle plumes. However, in the face of limited crustal extension, it must be assumed that during the rifting stage the asthenosphere was characterized by an elevated potential temperature and/or that an anomalous amount of volatiles was available in order to account for the volume of melts extruded and injected into the lithosphere, for instance in the area of the Massif Central, the Rhenish Shield and the Eger Graben (Wilson 1993).

In this context it should be noted that, in contrast to the stable Precambrian East European Platform, which is characterized by a thick crust (40–55 km) and lithosphere (90–190 km), low heat flow and an upper asthenosphere having high shear-wave velocities, large parts of the metastable Palaeozoic craton of western and central Europe are characterized by crustal and lithospheric thicknesses of less than 35 and 90 km, respectively, an elevated mantle heat flow and an upper asthenosphere displaying anomalously low shear-wave velocities (Cermak 1982; Suhadolc et al. 1990; Ansorge et al. 1992; Zielhuis 1992). The magnitude of this asthenospheric low-velocity anomaly, which is restricted to depths of less than 200 km, cannot be explained alone by an increased temperature of the upper asthenosphere, but requires the presence of partial melts in order to be compatible with the observed surface heat flow (Zielhuis 1992; Ansorge et al. 1992). This suggests that partial melting of volatile-enriched mantle source regions, having lower than normal solidus temperatures, has occurred (Wilson 1993). Therefore, the observed anomaly must be interpreted in terms of either the head of a large, not very energetic mantle-plume, or of partial melting of volatile-rich mantle-lithosphere which is underlain by an asthenosphere having a somewhat higher than ambient temperature (> 1300° C), possibly forming part of an upwelling and out-flowing branch of the sub-lithospheric mantle convection system.

The geochemical and Sr-Nd-Pb isotopic characteristics of primitive mafic volcanic rocks associated with the Cenozoic rifts of western and central Europe suggest that they were derived by mixing of partial melts originating from the convecting asthenosphere and from the mantle part of the lithosphere (Wilson & Downes 1992; Wilson 1993). The geochemical signature of the lithospheric component varies significantly across the European rift system, reflecting compositional differences between the different lithospheric terranes making up the Variscan fold belt (Wilson & Downes 1992; Rosenbaum et al. 1993). The asthenospheric component appears to be rather more homogenous across the province and has geochemical similarities to the source of HIMU oceanic island basalts (subducted oceanic lithosphere; Hart & Zindler 1989; Wilson 1989); this possibly reflects the involvement of an isotopically distinct mantle plume component in the petrogenesis of the magmas. This component could be derived either from a Neogene mantle plume or system of plumes, or could equally well reflect a Permo-Carboniferous or Mesozoic plume input into the thermal boundary layer at the base of the lithosphere; however, the simplest model is that it is related to Cenozoic plume activity (Wilson pers. comm. 1993). Alternatively, rather than invoking discrete mantle plume activity, Cenozoic development of an upwelling and out-flowing branch of the mantle convection system, entraining previously subducted oceanic material, could be envisaged (see below; Hart & Zindler 1989).

The evolution of sedimentary basins shows that the lithosphere/asthenosphere system of western and central Europe was repeatedly disturbed during Late Palaeozoic, Mesozoic and Cenozoic times. Following the late Westphalian consolidation of the Variscan orogen, its deep lithospheric roots were destroyed during the Stephanian-Autunian phase of wrench tectonics and widespread magmatism, whereby the Moho-discontinuity was regionally re-equilibriated at a depth of about 30–35 km. During the Triassic the lithosphere of much of the area that was later affected by Cenozoic rifting had thermally relaxed to the degree that the surface of the crust subsided below the erosional base-level and sedimentation resumed in progressively expanding areas. The evolution of Mesozoic basins was governed by rift- and wrench-tectonics, entailing a renewed destabilization of the lithosphere, which culminated in mid-Jurassic opening of the western Tethys and Cretaceous opening of the North Atlantic ocean. In western and central Europe, Mesozoic rifting activity was accompanied by relatively minor magmatic activity only; the mid-Jurassic central North Sea arch is an exception (Hendrie et al. 1993). The onset of Cenozoic rifting was preceded by late Cretaceous and Paleocene phases of major intra-plate compressional deformations, which can be related to the collision of Iberia and of the Alpine orogen with the metastable platform of western and central Europe, and also by the early Eocene onset of sea-floor spreading in the Norwegian-Greenland Sea (Ziegler 1988, 1990). During the Variscan orogeny and the post-Variscan period of lithosphere stabilization to a thickness of some

100–120 km, a significant amount of crustal material may have become incorporated into the mantle part of the lithosphere by means of phase transitions and by delamination of eclogitic material initially emplaced as basaltic underplate to the lower crust.

In the area of western and central Europe, mantle convection patterns presumably changed repeatedly through time, particularly at the end of the Variscan orogeny, during the opening of the Tethys and Atlantic oceans and again in conjunction with the Cretaceous development of subduction zones in the Tethys domain and the evolution of the deep reaching Alpine roots (Brousse & Bellon 1983; Ziegler 1988, 1990; Ansorge et al. 1992). In view of the generally low level of volcanic activity associated with Mesozoic extensional basins, it is speculated that the upper asthenospheric anomaly of western and central Europe came into evidence towards the end of the Cretaceous and evolved further during the Cenozoic. This notion is in keeping with the occurrence of Paleocene pre-rift dykes (alkali basalts, nephelinites, melilites) in the area of the Massif Central, the Rhine Graben and the Bohemian Massif (Maury & Varet 1980; Brousse & Bellon 1983; Horn et al. 1972; Suk et al. 1984); their emplacement coincides with the compressional intra-plate deformations of the Alpine-Pyrenean foreland. Development of the upper asthenospheric S-wave anomaly, involving a temperature increase of the upper asthenosphere, may be related to the development of an upwelling and out-flowing cell of the convecting mantle beneath the Alpine foreland which evolved in conjunction with the progressive roll-back of the subducting foreland slab and thrust-loaded downflexing of the foreland lithosphere (cf. Laubscher 1992). The resultant temperature increase of the upper asthenosphere presumably entailed the onset of partial melting of volatile-enriched phases in the lithosphere/asthenosphere thermal boundary layer, convective thinning of the lithosphere and a gradual upward displacement of the asthenosphere/lithosphere boundary (Latin et al. 1990; Schmeling & Marquart 1991; Wilson 1993). Thermal thinning of the lithosphere and its infiltration by partial melts was accompanied by a gradual rise in lithospheric geotherms, causing weakening of the lithosphere, thus rendering it more prone to deformation in response to far-field stresses (Cloetingh & Banda 1992). The localization of major magmatic activity in areas affected by Eocene and younger lithospheric extension and doming suggests a rift-related control on further partial melting and melt extraction, the development of melt diapirs and of magmatic pathways to the surface (Wilson & Downes 1992).

Under such a scenario, the Cenozoic European rift system conforms rather to a 'passive' than to an 'active' rift system (Sengör & Burke 1978; Ziegler 1992, 1993; Wilson 1993), though plume-related weakening of the lithosphere may have played a significant role.

Collisional foreland-splitting model

The concept of collisional foreland splitting was developed by Molnar & Tapponier (1975) who related the development of the Baikal rift to the Himalayan collision of India and Asia. Subsequently this 'impactogen' model was applied to the Cenozoic rift system of Europe (e.g. Sengör 1976; Tapponnier 1977; Sengör et al. 1978; Bergerat & Geyssant 1980; Dewey & Windley 1988).

Regarding the hypothesis that collision-related stresses could be responsible for the development of the Roer, Rhine, Bresse, Limagne, Saône and Gulf of Lions grabens, it must be kept in mind that particularly the Gulf of Lions, Saône and Bresse grabens strike nearly normal to the Pyrenean and sub-parallel to oblique to the West Alpine orogenic fronts, respectively (Figs 1–4). Eocene to early Oligocene development of these grabens could possibly be related to the collision of the Pyrenean orogen with its northern foreland (Blès et al. 1989). However, this model is not compatible with the late Oligocene southward propagation of the Rhine-Rhône rift system into the West Mediterranean area, where it is superimposed on the off-shore parts of the Pyrenees and the compressionally deformed Celt-Iberian and Catalonian Coast Ranges, as well as on the Corsica-Sardinia block. Rifting activity persisted in the Rhine-Bresse graben system even during the Aquitanian-Burdigalian opening phase of the oceanic Algero-Provençal Basin. Oligocene and Miocene inversion of Mesozoic grabens in the Celtic Sea-Western Approaches and Channel areas, paralleled by transpressional deformations along the Pays-de-Bray fault (Figs 6, 7), is thought to result from the build-up of compressional stresses in the Alpine foreland in response to its collisional coupling with the West Alpine orogen. Furthermore, the Plio-Pleistocene reactivation of the Valencia Graben (Banda & Santanach 1992), the tensional collapse of the Alboran Sea and the development of a chain of alkaline volcanoes crossing the Rif fold belt and the inverted Middle and High Atlas, can hardly be reconciled with the model of collision-related foreland splitting (Ziegler 1988).

The Plio-Pleistocene volcanism of the Rhenish Shield, the Bohemian Massif and the Massif Central, together with the development of the Valencia Trough-Trans Atlas volcanic chain, testifies to continued activity along this mega-rift system which transects the West Mediterranean Alpine fold belts and their compressionally deformed forelands.

Crustal extension across the Valencia-Rhône-Rhine rift system, which feathers out to the north along the southern margin of the Northwest European Basin, was presumably accompanied by a small-scale clockwise rotation of the Armorican-Paris Basin Block relative to the Franconian-Bohemian Block. This rotation may have played a role in the Oligocene and younger transpressional uplift of the Ardennes-Artois-Wealden axis and the inversion of the Channel, Celtic Sea and Western Approaches basins, as well as in the broad upwarping of the British Isles. Although these compressional intra-plate deformations are partly contemporaneous with the subsidence of the Rhine-Rhône rift system, it is doubtful whether crustal extension across the latter provided the driving force for the development of the former, as postulated by Gillchrist et al. (1987). More likely foreland compressional deformations and rifting were caused by far-field stresses resulting from plate interaction. For instance, compressional stress systems developing in conjunction with the Pyrenean collision of Iberia and Europe may have contributed towards the Eocene to early Oligocene development of the Cenozoic rift of western Europe. Neither their subsequent evolution nor the late Neogene resumption of rifting activity in the West Mediterranean domain, can be explained by such a model as by this time the Pyrenees had become tectonically inactive. Furthermore, the dominance of sinistral motions along the transform zones linking the different segments of the European Cenozoic rift system (Fig. 1) is not compatible with the Neogene dextral translation between Africa-Arabia and Europe which gave rise to the development of the intra-Alpine shear system (Figs 4, 7; Ziegler 1988, 1990).

Back-arc rifting model

The back-arc model proposed by Jowett (1991) for the Cenozoic rift systems of Europe assumes that the subducted lithospheric slabs associated with the Western, Central and Eastern Alps dip westwards and northwards, respectively, under the Alpine foreland. The dynamics of back-arc rifting are explained in terms of secondary mantle convection cells developing under the overriding plate, imparting on it a positive deflection and exerting on the base of its lithosphere horizontal tensile shear stresses, thus causing the development of foreland rifts. Post-collisional steepening of the subducted slab, resulting in a decrease in collision-related tangential compressional stresses projected into the foreland and an increase in convective mantle velocities, giving rise to an increase in tensile shear stresses exerted on the base of the lithosphere, are thought to cause an acceleration of back-arc rifting activity. Upon detachment of the subducted slab from the thickened lithosphere, the inherent body forces of the latter will contribute to its further extension (Bott 1990; Bott et al. 1990).

In the case of the Rhine-Rhône rift system, the applicability of the back-arc rifting hypothesis can be rejected on the basis of refraction and seismic tomography data which all indicate that the subducted lithospheric slabs associated with the Western and Central Alps dip eastwards and southwards, respectively, beneath the Alps (Nicolas et al. 1990; Spakman 1990; Blundell et al. 1992). However, the back-arc model may find application in the case of the late Oligocene-early Miocene development of the Valencia Graben, which evolved in a backc-arc position relative to the Kabylian-Calabrian arc (Fig. 3). Significantly, volcanic rocks extruded during the initial rifting stage of the Valencia Trough have a calc-alkaline composition whereas volcanics extruded during its Plio-Pleistocene reactivation are alkaline (Banda & Santanach 1992; Maillard & Mauffret 1993; Torres & Bois 1993).

Plate-reorganization model

Viewed on a much broader scale, the late Eocene-Oligocene beginning of rifting activity in western Europe coincides with the onset of crustal extension in the Red Sea-Gulf of Suez area (Evans 1988; Ott d'Estevou et al. 1989; Favre & Stampfli 1992). The Oligocene-Miocene northward and southward propagation of the European Cenozoic rift system was paralleled by the evolution of the East African-Red Sea-Gulf of Suez rift system, the Dead Sea wrench zone and the Libyan rift system. During the Plio-Pleistocene rifting propagated northward onto the Pelagian Shelf, through the Strait of Messina and onto the internides of the Apennines (Jongsma et al. 1985; Montenat et al. 1991; Doglioni 1991; Serrini 1990), whilst the Rhine-Rhône rift propagated southwestward across the West Mediterranean area and through northwest Africa to the Cape Verde Islands (Fig. 4; Ziegler 1988, 1990).

Evolution of these mega-rift systems was contemporaneous with major crustal shortening in the different segments of the Alpine-Mediterranean orogen, resulting from the interaction of the converging African-Arabian and Eurasian plates. During the late Cretaceous and early Paleocene phases of the Alpine orogeny, Africa-Arabia converged with Europe in a counter-clockwise rotational mode, whereby differential sinistral movements between them decreased; these ceased altogether during the latest Paleocene to earliest Eocene in conjunction with the opening of the North Atlantic, the Norwegian Greenland Sea and the Eurasian Basin. Based on the analysis of seafloor magnetic anomalies of the Atlantic and Indian oceans, Africa-Arabia and Europe converged during the Eocene in an approximately north-south direction; however, during the Oligocene and early Miocene their convergence changed to a dextral oblique one. Mid-Miocene and Pliocene deformation patterns of the Alpine fold belts suggest, that dextral motions between Africa-Arabia and Europe continued to play an important role and probably persist to the present, as indicated by earthquake focal mechanisms (Figs 2–4; Ziegler 1988).

Dextral translation of Europe relative to Africa-Arabia during the late phases of the Alpine orogeny may have contributed to the development of the Rhine-Rhône-Valencia and the Red Sea-Libyan rift systems and may have played an important role in the opening of the Provençal Basin (Laubscher 1974). However, evolution of these mega-rifts, and particularly of the Red Sea-East African rift system, is difficult to explain alone in terms of the interaction of the Africa-Arabian and European cratons, but heralds, in the author's view, the beginning of a fundamental re-organization of plate boundaries and the gradual assertion of a new kinematic regime that could ultimately lead to the break-up of the current continent assembly. In this respect, general upwelling of the mantle under Africa and its radial outflow may play an important role (Pavoni 1992, 1993). Under such a scenario, it is likely that compressional stresses, related to the collision of Africa-Arabia and Europe, interfered with tensional stresses which governed the development of these new rift systems. In time and space, one or the other process became dominant, resulting in a certain pulsation of tensional and compressional tectonic activity (Ziegler 1988, 1990).

Conclusions

The Cenozoic rift system of western and central Europe has many features which are consistent with a 'passive' rift model. However, its magmatic record indicates the involvement of a geochemically distinct asthenospheric component in the petrogenesis of primary magmas that shows some similarities to the source of plume-related oceanic island basalts. Available seismic tomography data suggest that the mantle beneath western and central Europe may be somewhat hotter than normal. On the basis of the limited amount of lithospheric extension and the volume of magmas generated within the province, this must indeed be the case, unless the bulk of the magmas is derived from volatile-enriched domains within the Variscan mantle-lithosphere (Wilson 1993). Therefore, it must be assumed that the potential temperature of the convecting asthenosphere increased during the Paleogene, possibly in conjunction with a reorganization of the mantle convection patterns. This may have caused a weakening of the lithosphere, thus rendering it more prone to failure in response to the build-up of intraplate stresses.

During the late Eocene and early Oligocene initial development phases of the European rift system, the build-up of principal horizontal compressive stress axes projecting from the Pyrenean and possibly also the Alpine collision front into the foreland may have played a role in the tensional reactivation of Permo-Carboniferous and Mesozoic fracture systems, thus controlling the localization of its individual grabens; in time these coalesced to form a more or less continuous system of rifted basins. Late Oligocene-early Miocene southward propagation of this rift system across the West-Mediterranean fold belts, accompanied by opening of the oceanic Algero-Provençal Basin, and Plio-Pleistocene reactivation of the Valencia rift, is taken as an indication of the gradual assertion of a new kinematic regime which may herald the break-up of the present continent assembly. This view is reinforced by the contemporaneous development of the East African-Red Sea and Libyan-Pelagian Shelf rift systems and the Plio-Pleistocene northward propagation of the latter into the domain of the Apennines.

Subsidence of the various segments of the European Cenozoic rift system was governed by repeated changes in the controlling stress field, presumably reflecting an interplay between collision-related foreland-compressional stresses, regional tensional stresses and rift-related thermal loads. Doming of a rift zone can be induced either by progressive lithospheric thinning (Rhenish Shield) and/or the intrusion of melts to intra-lithospheric levels (Massif Central), as well as by positive deflections of the lithosphere in response to the build-up of horizontal intra-plate compressional stresses and/or thrust-loading of the Alpine foreland (Vosges-Black Forest, Massif Central?). Development of rift-related arches can entail uplift of the grabens which transect them and erosion of their sedimentary fill, even in the face of continued crustal extension. There is circumstantial evidence of destabilization of the Moho discontinuity during the process of rifting.

Earthquakes, neotectonic deformations, hydrothermal activity, frequent emanations of CO_2 and mantle helium, as well as the age of the youngest volcanic activity indicate that the Cenozoic rift system of Europe is still active and corresponds to a zone of increased hazard.

Acknowledgements

The author is indebted to KNMI for sponsoring his participation in the Veldhoven workshop on the Roermond Earthquake and to Shell Internationale Petroleum Mij. for preparing the text figures and releasing them for publication. Thanks are extended to Profs. S. Cloetingh (Amsterdam), H-P. Laubscher (Basel), M. Wilson (Leeds) and F. Wenzel (Potsdam) for their constructive comments on an earlier version of this manuscript.

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