The European Cenozoic Volcanic Province is not caused by mantle plumes

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Abstract

The European Cenozoic Volcanic Province (ECVP) comprises several small volcanic areas. Its formation is often genetically linked to one or several mantle plumes, but there is no convincing evidence for this. The primary observations from the ECVP contradict every prediction of the classical plume hypothesis and require major adaptations of that model to explain them. The mantle plume model thus is not an acceptable explanation for ECVP formation. A more likely explanation is that the volcanism is related to Alpine subduction processes that affect stress, deformation and flow in the European continental crust and underlying shallow mantle, coupled with the influence of local lithospheric conditions.

1. Introduction

The current geodynamic setting of Central and Western Europe is dominated by the continental collision of the European and the Adriatic plates during the Alpine orogenesis. In the northern Alpine forelands extensive lithospheric rift systems developed (*e.g., Ziegler*, 1990), closely linked both spatially and temporally to the Alpine orogeny. Sedimentary basins developed under extensional conditions in central Europe, and their formation was accompanied locally by melt intrusion and extrusion (Figure 1). The main graben structure, the NNE–SSW trending Rhine graben, is essentially amagmatic with the exception of Kaiserstuhl volcano in its southern part (*Wilson & Downes*, 1991). Magmas extend over France (the Massif Central), central Germany (Eifel, Westerwald, Vogelsberg, Rhön), the Czech Republic (the Eger graben) and SW Poland (Lower Silesia), a region ~1,200 km wide. This volcanic region is called the European Cenozoic Volcanic Province (ECVP) in this webpage.

The predominantly intraplate magmatism of the ECVP has been attributed by different authors to a variety of causes (*e.g.*, Lustrino & Carminati, 2007, and references therein). These include one or more mantle plumes and a partial melt layer representing a "fossil plume head" from which small plumes (also termed "baby plumes" or "finger plumes") rise. Nevertheless, the province has essentially none of the characteristics expected for classical Morgan-type mantle plumes (Morgan, 1971). A source related to Alpine subduction processes coupled with local lithospheric conditions offers a much less astonishing explanation for European Cenozoic volcanism than do mantle plumes.

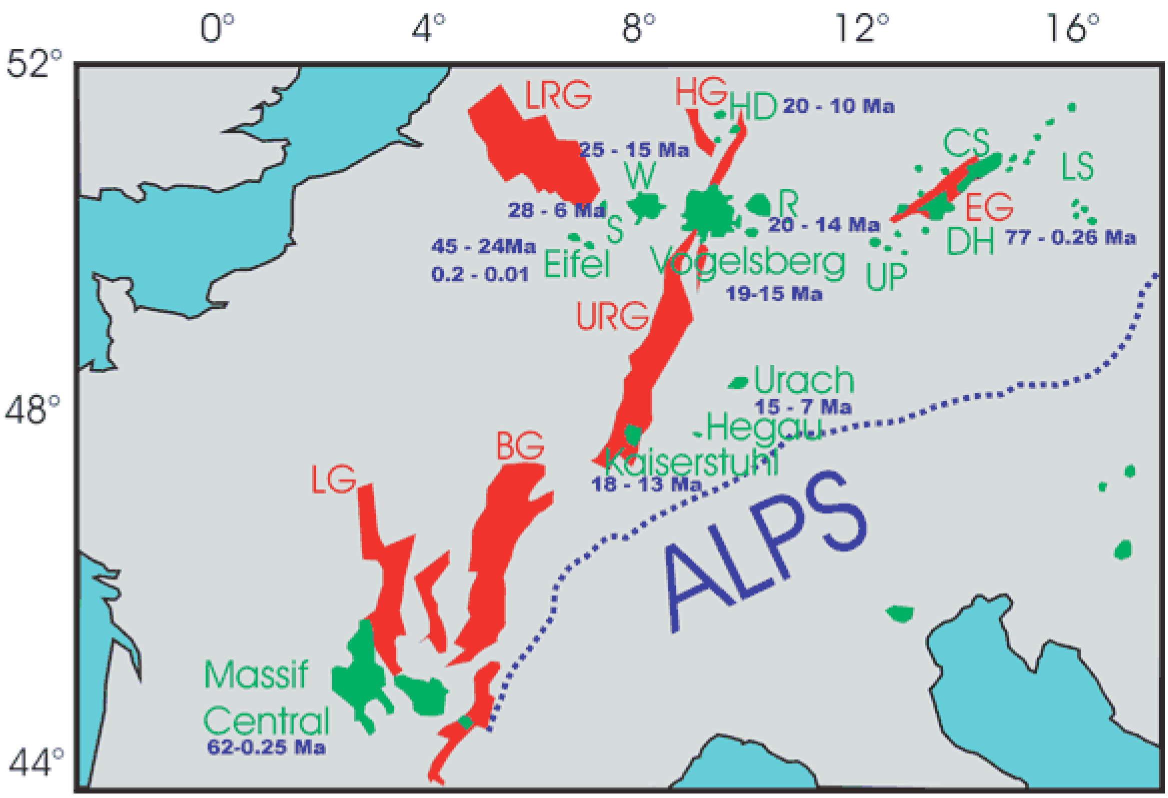

Figure 1: Map of the Cenozoic volcanic rocks of central Europe (green) and rift-related sedimentary basins (red) in the Alpine foreland. Volcanic sub-areas: S: Siebengebirge; W: Westerwald; HD: Hessian Depression; R: Rhön/Heldburg; UP: Upper Palatinate; DH: Doupovské Hory; CS: Ceskè Stredohon; LS: Lower Silesia. Rift systems: LG: Limagne Graben; BG: Bresse Graben; URG: Upper Rhine Graben; LRG: Lower Rhine (Roer Valley) Graben; HG: Hessian grabens; EG: Eger (Ore) Graben. Age data from Abratis et al. (in press), Lustrino & Wilson (2007) and references therein.

Some primary observations from the ECVP are:

1. The area covered with volcanic rocks is much too small to be classified as a Large Igneous Province (LIP);
2. The spatial distribution of volcanism is not time-progressive;

3. Sgutterome of the volcanic fields are located close to Cenozoic rift systems (*e.g.,* Kaiserstuhl, Vogelsberg), whereas others are located on uplifted basement massifs (*e.g.*, the Westerwald, Eifel and Auvergne);
4. The time-scale of uplift and volcanism is not what is predicted by the mantle plume model;
5. At least two different mantle xenolith types in host magmas show that the subcontinental mantle of the ECVP is heterogeneous. In general, ECVP magmas have trace-element and isotopic characteristics that resemble closely ocean-island basalts (OIB);
6. $^{3}He/^{4}He$ ratios are homogeneous (R/Ra = 6.32 ± 0.39; *Gautheron et al.,* 2005) and lower than MORB values (8 ± 1 Ra; *Allègre et al.,* 1995);
7. No high-Mg (picritic) magmas, often assumed to indicate excess temperatures in the mantle, are reported, and;
8. Low-wave-speed seismic mantle anomalies are detected beneath some, but not all, sub-areas. Where they are observed, they are confined to the upper mantle.

These eight points are discussed in sequence and in detail in Section 3 below.

2. Tectonic setting

The developing Alps and Pyrenees exerted compressional stresses on the European plate during the Paleocene and caused lithospheric buckling and basin inversion out to distances of up to 1700 km north of the Pyrenean and Alpine deformation fronts (*Dèzes et al.,* 2004). This buckling was accompanied locally by the injection of mafic melts into the European foreland crust. The Paleogene geodynamic history of the Alpine orogen was dominated by the subduction of the continental Briançonnais block subduction after subduction of the South Penninic Ocean was complete (well summarized by *Schmid et al.,* 1996). At the end of the Early Eocene the whole Briançonnais zone, along with major parts of the North Penninic Bündnerschiefer/Valais "ocean", were subducted. The following initial continent-continent collision stage (Early Eocene, Ypresian to Oligocene, Rupelian) is defined by the subduction of the Adula nappe–the southern, distal margin of the continent of Europe (*Schmid et al.,* 1996).

During the middle Eocene, the European Cenozoic Rift System (ECRIS) began to form as a result of the north-directed intraplate compressional stresses that resulted from the processes described above. In the north, the ECRIS is represented by the Rhine rift system which comprises the northward-trending Upper Rhine and Hessian (Wetterau, Leine) grabens and the northwest-striking Roer Valley (Lower Rhine) graben. The southern branch of the ECRIS includes the grabens of the Massif Central (Limagne, Forez, Roanne) and the Bresse Graben (*Ziegler & Dèzes,* 2005; 2007). The northern and southern rift systems are thought to involve tensional reactivation of Late Hercynian fractures (*Ziegler*, 1992).

Tectonic events associated with Alpine orogeny, which were important for ECVP magmatism occurred during the Eocene-Oligocene transition. Major parts of the Penninic units, including oceanic crust, Briançonnais, and distal European crust were subducted (*Schmid et al.,* 1996). The European upper continental crust was accreted to the orogenic wedge after the Eocene collision. This excessively thickened the orogenic wedge whilst the detached lower crust of the European foreland was subducted from 36 Ma onward (P.A. Ziegler, pers. comm.). Around 34 Ma the subducting slab of the central Alps, comprising the lower crust and the lithospheric mantle of the distal European margin and the Briançonnais and the lithosphere of the Valais and Penninic oceans, broke off from the European foreland lithosphere (*Dèzes et al.*, 2004, 2005). Slab break-off and tearing scenarios have also been described for Anatolia and Mexico.

3. Features of the ECVP important to source models

3.1 The size of the ECVP

The ECVP covers a maximum total area of 20,000 km^2. This is much smaller than the minimum areas of 50,000 km^2 or 100,000 km^2 that have been suggested for LIPs. Lava thicknesses and volumes are also small. LIPs such as the North Atlantic Igneous Province and the Deccan flood basalts have thicknesses of several to many kilometres. Volcanism in the ECVP, on the other hand, is characterized by small, monogenetic centres (*e.g.*, Eifel), scattered necks and plugs (*e.g.,* the Hessian Depression), dykes (*e.g.*, the Rhön/Heldburg area) and a few central volcano complexes (*e.g.*, Cantal and Vogelsberg). The sizes of the different regions vary from large central volcano systems (*e.g.*, Vogelsberg–ca. 2500 km^2) to small isolated plugs.

3.2 Spatial and temporal patterns

One of the most important mantle plume model predictions, time-progressive volcanism, is violated in the ECVP. The volcanics are mainly restricted to a belt 300 km wide surrounding the northern Alps. A space-time correlation between the igneous sub-areas, that might indicate a plume track, does not exist. For example, magmatic activity in the Eger Graben and the Eifel area started in the Eocene (in the Massif Central, Vosges-Black Forest and Bohemian Massif even as early as the Paleocene), but the major phase of activity in western and central Europe occurred only in the Neogene. Along with intervals of low activity, eruptions continued locally to a few thousand years BP, despite paleogeographic reconstructions that show that activity in these areas started when they were approximately 1000 km south of their present positions (*Torsvik et al.*, 2001). This cannot be explained by one or more relatively stationary mantle plumes, but suggests a phenomenon that traveled with the lithosphere.

Volcanic activity ceased in some sub-areas, whereas it continued or started in others with the same chemistry (*e.g., Abratis et al.*, in press). An excellent example of this "jumping" of activity is the area around the Vogelsberg (Figure 2). Volcanic activity ceased at 18 Ma in the Rhön area whereas it continued in the Vogelsberg and in the northern Hessian Depression in the WNW, and started in the Grabfeld, farther SE, at 16 Ma. At present, geologically recent magmatic activity as well as the oldest igneous ECVP rocks (Paleogene), occur at both the W and E ends of the ECVP.

3.3 Tectonic setting of the volcanic fields

ECVP eruptions occurred in two major tectonic settings:

1. the European Cenozoic Rift System, and
2. uplifted Variscan basement massifs (the Massif Central, Rhenish Massif and Bohemian Massif).

Magmas erupted in a volcanic belt from the French Massif Central, through central Germany (Eifel, Westerwald, Vogelsberg, Rhön), to the Eger graben and SW Poland (Figure 1). This belt runs perpendicular to the post-Alpine NNE-SSW rift system of the Upper Rhine graben (*Ziegler*, 1992). The volcanic complex of the Vogelsberg is located at the northern end of the Rhine Graben where it splits into two branches. Other volcanic fields associated with the Cenozoic rift systems are the Eger graben and Hessian depression volcanic areas. Examples of volcanic areas associated with uplifted basement massifs are the Westerwald, Eifel and the Massif Central.

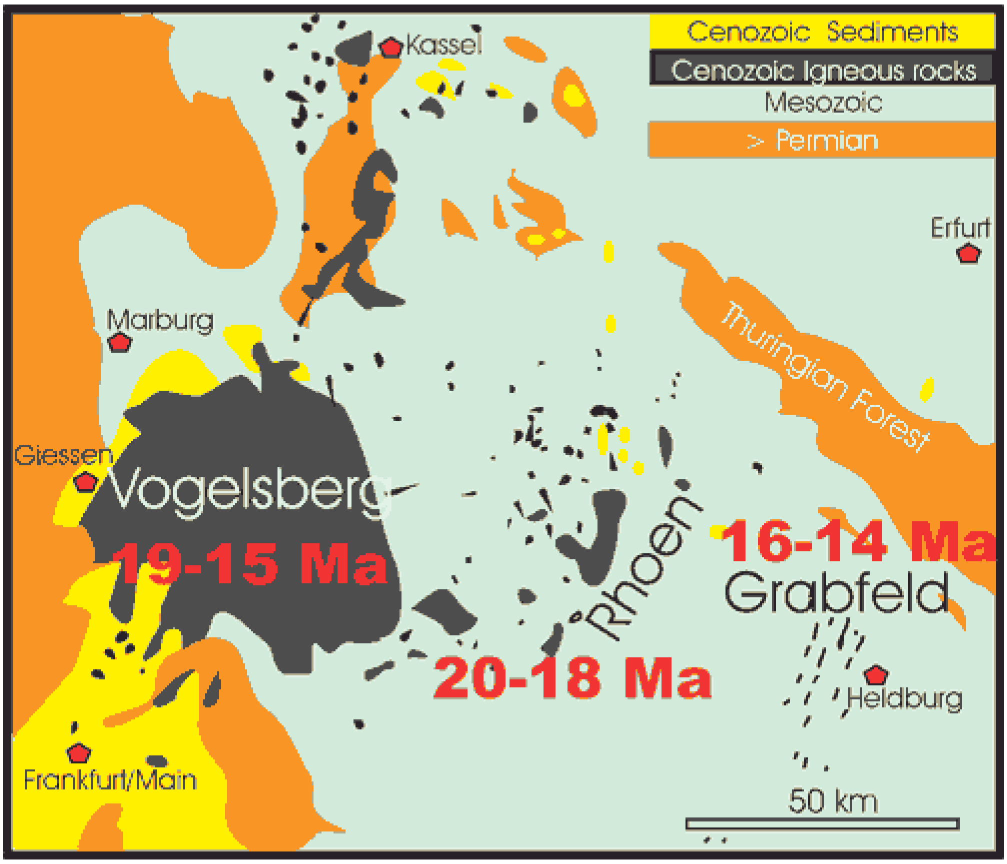

Figure 2: Simplified geological map of central Germany showing the ECVP sub-areas Vogelsberg, Rhön and Heldburg Gangschar. Volcanics and dykes are indicated by black coloured fields. Age data from [Abratis et al. (in press)] and references therein.

In general, the sub-provinces on uplifted areas have been linked to arriving mantle plumes (*e.g.*, *Granet et al.*, 1995; *Ritter et al.*, 2001). However other European coeval uplifted massifs show no Cenozoic magmatic activity (see Section 3.4).

The volcanic areas close to or within the Cenozoic rift systems are clearly the result of decompression melting processes due crustal extension. On the other hand the Vogelsberg, situated at the triple junction of the Upper Rhine Graben and the Hessian grabens (Figure 1), has often been used as a textbook example of a triple junction caused by the impact of a mantle plume on the base of the lithosphere. It has been suggested that the large volume of igneous material at the Vogelsberg requires high mantle temperatures. However, source temperatures higher than for other ECVP sub-areas are not required for these (presumably) higher melt fractions and volumes. Petrogenetically, three different possible scenarios could result in melting:

a. Higher temperatures;
b. Decreasing pressure; and
c. Changes in volatile content.

Lithosphere stretching and thinning, coupled with consequentially induced local, small-scale convection at this triple junction, negates the need for high temperatures.

3.4 Vertical motions

Plume models predict uplift of the surface prior to arrival of a plume head (*Campbell*, 2005). A maximum elevation of ~500-1000 m is predicted for a temperature anomaly of ~100°C. Uplift is expected to be followed by subsidence as the plume head spreads beneath the lithosphere, and the impingement site is transported away from the plume stem. The lithosphere is expected to stretch above the arriving plume. Vertical movements associated with plume impingement are predicted to place the lithosphere under stress, and zones of compression and extension are predicted to result above the plume head (*Burov & Guillou-Frottier*, 2005). Whether the resulting rift zones are expected to be elongated, as in the ECRIS and its perpendicular ECVP, is unknown.

The ECRIS is characterized by relatively low crustal stretching factors (*Cloetingh et al.*, 2005), hardly reaching values of 1.15 in the deepest and narrowest parts of the Upper Rhine Graben (*Dèzes et al.*, 2004; *Ziegler & Dèzes*, 2007), but with a distinct uplift of the crust–mantle boundary. Basement uplift in the ECVP area started during the Neogene (*e.g.*, the Massif Central: Early Miocene; Vosges-Black Forest Arch: Middle Miocene) extending over a period of ~20 Ma (*Ziegler*, 1990; 1992), which is 20 to 40 Ma after the beginning of rifting (*Ziegler*, 1992). Stratigraphic observations show that the Rhenish Massif, Eifel, and the Massif Central were still close to sea level during the Oligocene (*Ziegler & Dèzes*, 2005 and references therein) when volcanism had already commenced. These areas were mainly uplifted later, commencing during the Middle Miocene (*Ziegler & Dèzes*, 2007).

Volcanic activity appears to have been synchronous with the Late Oligocene minor uplift of the Rhenish Massif (*Meyer et al.*, 1983). During the Miocene, the Roer Valley graben continuted to subside due to the NW-directed compressional stress field (*Schumacher*, 2002), in contrast with the Rhenish Massif. This included the Hessian grabens, which were gradually uplifted and became the site of increased volcanic activity (*e.g.*, *Jung*, 1999). This can be related to progressive thinning of the mantle lithosphere, and the related induction of small-scale convection. The Bohemian Massif experienced a major phase of volcanism during the Early and Middle Oligocene which was succeeded by subsidence of the NE-striking Eger Graben (*e.g., Ulrych et al., 1999*). It is beneath this region that seismic imaging recently failed to detect a plume-like structure in the mantle (*Plomerová et al., 2007*).

An additional observation relevant to vertical movements in central Europe is that many volcanic constructions were strongly eroded (*e.g., Bücking*, 1916) after volcanic activity ceased. This suggests that uplift still continued after the volcanic activity. *Meyer et al.* (1983) concluded that uplift in the western Rhenish Massif accelerated during the Quaternary while volcanic activity resumed in Eifel (0.7-0.01 Ma). Vertical movement continues to the present-day (see *Meyer & Stets*, 2002; *van Balen et al.*, 2000 in *Ziegler & Dèzes*, 2007).

It is also interesting to note that the uplifted Armorican Massif (NW France) and the Belgian Ardennes have no volcanic activity at all. Clearly uplift in the Alpine foreland has a complex history and pattern and cannot simplistically be explained by one or more mantle plumes. For more discussion of uplift mechanism, see *Ziegler & Dèzes* (2007).

3.5 Petrogenesis

3.5.1 Rock types and melt fractions

In the ECVP basanites, nephelinites, and alkali basalts dominate over tholeiitic basalts (olivine to quartz; Figure 3). The higher partial melting degrees presumed for the tholeiitic basalts (~10%; *Meyer et al.*, 2002) compared with alkali basalts (2-3%; *Meyer et al.*, 2002) cannot automatically be attributed to an abnormally hot mantle. In a high-temperature mantle plume scenario, a relationship in space and time between the alkaline and tholeiitic melts would be expected. *Abratis et al.* (in press) conclude, however, that the different magmas erupted more or less simultaneously within sub-areas. The magmas furthermore erupted continually in the same small area of crust over millions of years on a moving plate. The higher partial melting degrees thus seem to be more easily explained by fertility heterogeneities in the source.

3.5.2 Mantle heterogeneity

Mantle homogeneity beneath a volcanic area can only be directly studied using mantle xenoliths (*e.g., Meyer & Hertogen*, 2007). Two distinct suites of spinel peridotites (*Stosch*, 1987) found in mantle xenoliths from the ECVP testify to a heterogeneous mantle:

1. A high-temperature anhydrous suite–the coarse-grained, high-T anhydrous Ib type. This suite is characterized by the absence of hydrous minerals. The texture of these xenoliths shows a uniform grain size, suggesting that they formed in the lithosphere under constant P-T conditions over millions of years. They are spinel harzburgite and olivine clinopyroxenites (*Witt-Eickschen, 1993*). The pyroxenites are also indicators of a heterogeneous mantle, since they most likely represent veins in the lherzolitic lithospheric mantle (*Witt-Eickschen & Kramm, 1998*). Geochemically these xenoliths range from REE chondritic to moderately LREE depleted lherzolites, to moderately LREE enriched harzburgites;

2. A low-temperature hydrous suite–the tabular-mosaic hydrous Ia type. These xenoliths contain pargasitic amphibole in addition to olivine, orthopyroxene, clinopyroxene, spinel and the breakdown products of spinel. The hydrous Ia xenoliths show strong enrichment of LREE and other incompatible elements (*Schmidt et al.*, 2003).

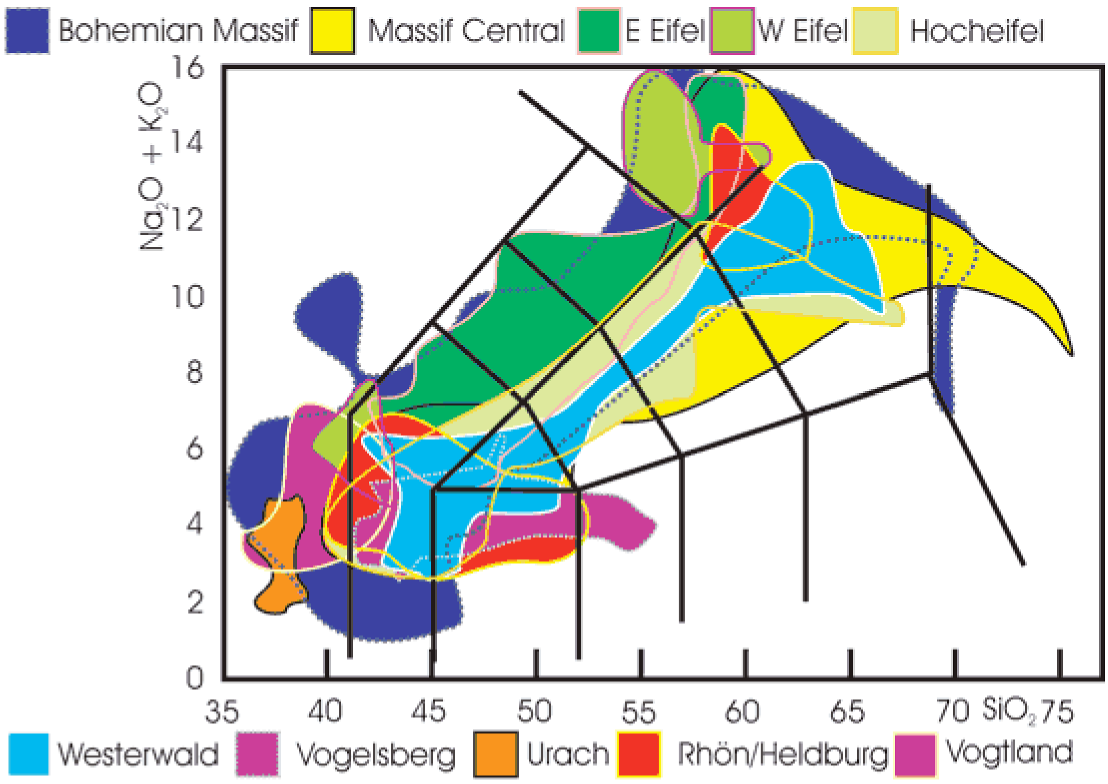

Figure 3: Total alkali–silica (TAS) diagram which outlines the compositional variability of ECVP rocks. The corresponding geochemical data are published in Lustrino & Wilson (2007) in addition to some personal (RM) unpublished data.

Based on P-T data it has been postulated that the amphibole-bearing xenoliths came from just below the Moho, and that the anhydrous xenoliths came from greater depths. Following the earlier ideas of *Sleep* (1984) and *Fitton & Dunlop* (1985), a heterogeneous mantle containing lherzolite layers with layers/veins of subducted eclogitic material has been suggested, and termed "marble cake" mantle, by *Allègre & Turcotte* (1986).

3.5.3 The OIB paradox, and an asthenospheric or lithospheric source

The origin of intraplate magmas has been petrogenetically linked to different sources:

1. Partial melts of a metasomatized lithospheric mantle (see, for example, Metasomatic OIB page);
2. Melts from a metasomatized asthenospheric mantle source, and;

3. Combined asthenospheric and lithospheric mantle sources (*e.g.*, *Lustrino & Wilson,* 2007 and references therein).

There are diverse opinions on this issue, which may result in part from the difficulties in distinguishing the trace-element and isotopic characteristics of deep-mantle sources from those acquired by the involvement–either as melt source or as contaminant–of the lithospheric mantle and/or the continental crust. The OIB-like geochemistry of ECVP magmas has been presented as "proof" of a deep mantle plume source (*e.g., Wörner et al.,* 1986; *Hoernle et al.*, 1995; *Wedepohl & Baumann*, 1999; *Haase et al.*, 2004). However, as *Fitton* (2007) recently pointed out, OIB and OIB-like basalts are widespread throughout the oceans and at many localities where there is no evidence otherwise for, or expectation of, mantle plumes, and where other observations rule them out to a high degree of certainty. Such observations include very small magma volumes and continual volcanism as the locality is transported for thousands of kilometres by plate movements. Many other continental rift systems exhibit similar behaviour and magma compositions *e.g.*, the East African rift and the western United States. See also Origin of OIB pages.

3.6 ^{3}He/^{4}He geochemistry

ECVP magmas contain enriched isotopic OIB signatures of mainly HIMU type, mixed in variable proportions with EM-1 (*e.g.*, Vogelsberg) or EM-2 (*e.g.,* Massif Central; Figure 4). This isotopic composition clearly identifies the magmas as OIB-like. However, the occurrence of OIB and OIB-like magmas cannot be petrogenetically unambiguously explained using traditional geochemical reasoning. Other tracers for a deep-mantle or core-mantle-boundary origin have thus been proposed such as noble gases–in particular ^{3}He/^{4}He ratios higher than the common value of 8 ± 2 Ra generally attributed to MORB (*e.g., Allégre et al.,* 1995). This postulate has nevertheless been challenged (see Helium Fundamentals, Noble Gases, Pt-Os and Osmium & Tungsten pages).

In contrast to the high ^{3}He/^{4}He signatures reported from volcanic regions such as the Deccan, Afar and Yellowstone (*Basu et al.*, 1993; *Marty*, 1993; *Dodson et al.*, 1997), ^{3}He/^{4}He ratios from the ECVP are all lower than for MORB (R/Ra = 6.32 ± 0.39; *Gautheron et al.*, 2005). The isotopic ratios in each sub-area tend to be uniform and homogeneous, with maximum values, for example, of R/Ra = 6.73 ± 0.11 for Eifel and 6.91 ± 0.3 for the Massif Central (*Gautheron et al.*, 2005). In addition, the ^{3}He/^{4}He ratios do not vary as a function of distance from postulated "baby plume" centres.

Gautheron et al. (2005) concluded that helium, neon and argon ratios argue against a lower-mantle plume as the driving force for ECVP magmatism, but support rift-related melting of the lithospheric mantle. *Dunai & Baur* (1995) explained Massif Central and Eifel helium systematics by 1-2% of recycled continental crust in the mantle source. They calculated model ages ranging from 350 Ma to 800 Ma for this contamination process. They concluded that contamination of the ECVP mantle source occurred during and/or shortly prior to the Variscan orogeny. The Variscan orogeny lasted from the Upper Devonian to the Lower Carboniferous (~375 to ~330 Ma). An alternative orogeny could be the Cadomian orogeny which took place in the Late Proterozoic and Early Cambrian (~570 to ~540 Ma).

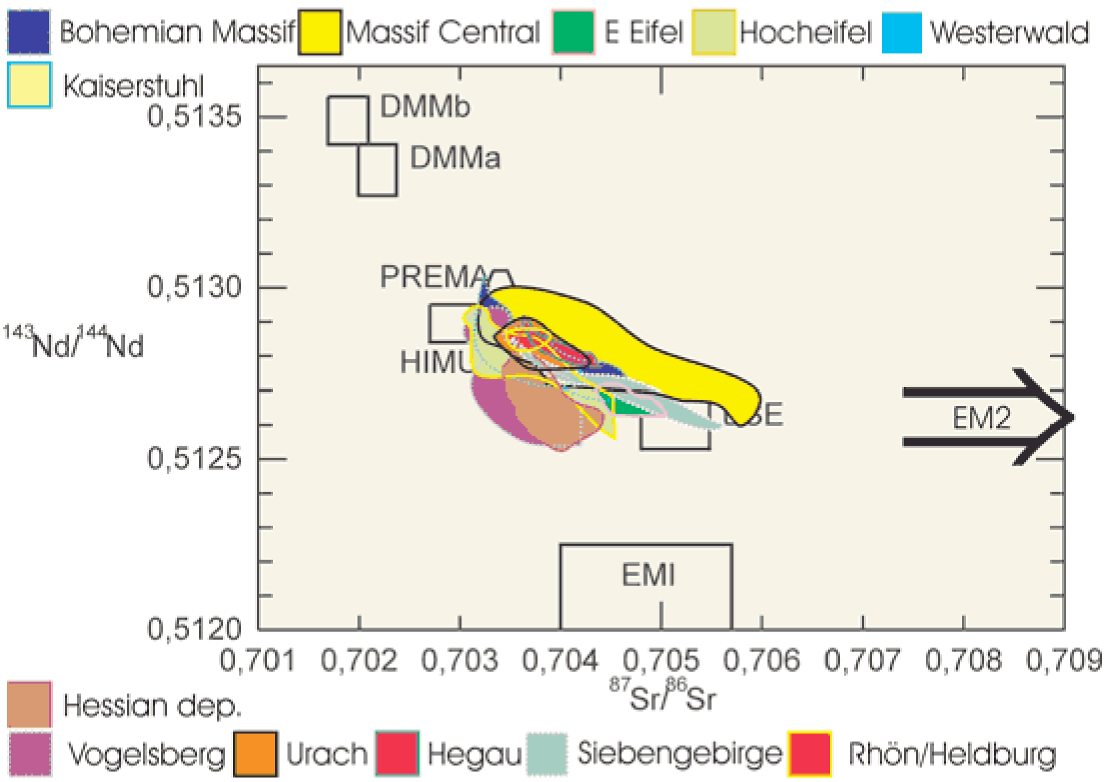

Figure 4: $^{143}Nd/^{144}Nd$ vs. $^{87}Sr/^{86}Sr$ diagram for ECVP magmas. Data from Lustrino & Wilson (2007) in addition to some personal (RM) unpublished data. Fields from Hart & Zindler (1989). Click here or on figure for enlargement.

3.7 Source temperature indicators

Throughout the entire ECVP no picritic magmas, commonly cited as evidence for high source tempeatures, have been found. Rocks with relatively high Mg contents (up to 16 wt.%; *e.g., Meyer et al.*, 2002) have been reported in the Rhön for primitive basanites. These rocks are not picrites, however, because their Na_2O+K_2O contents exceed 3 wt.%. The petrogenesis of these rocks does not require anomalously high temperatures.

Adopting the mantle plume model, *Ritter et al.* (2001) explains the maximum tomographic *P*-wave-speed anomaly of ~2% down to depths of 400 km by a 150–200°C temperature anomaly in the mantle. However alternative scenarios involving the presence of partial melt or other fluid, mineralogy and bulk-rock chemistry of the mantle can also explain this anomaly (*Lustrino & Carminati*, 2007). *Keyser et al.* (2002) explains the –5% *S*-wave-speed anomaly between 31-170 km depth below Eifel as a temperature anomaly of 100°C plus ~1% of partial melt. Low or absent temperature anomalies are more consistent with the observed absence of picrites.

Due to shallow local Moho depths, the temperature fluxes are higher in thinned crustal segments (*e.g.,* the Rhine graben). The typical surface temperature flux in Germany is 50-80 mW/m^2. In contrast, higher fluxes ranging from 80 to 120 mW/m^2 have been observed in the

Rhine graben (*Blundell et al.*, 1992). Recent geothermal work shows a higher temperature gradient for Eifel, but not for any other German, Czech or Polish ECVP sub-area.

3.8 Mantle tomography

The classical plume model (*e.g., Morgan*, 1971; *Griffiths & Campbell*, 1990; 1991; *Campbell & Griffiths*, 1990; *Farnetani & Samuel*, 2005) assumes that plumes originate at the core-mantle boundary and have a head-tail structure. Seismic tomography images are expected to show this morphology. Teleseismic tomography experiments have been performed around the French Massif Central and Eifel (*e.g., Granet et al.*, 1995; *Ritter et al.*, 2001; several papers in *Ritter & Christensen*, 2007) and the Eger graben (*Plomerová et al.*, 2007).

Results from the Massif Central showed a low *P*-wave-speed anomaly extending from the surface deep into the upper mantle. The Eifel Plume Project reveals a similar *P*-wave structure, though there is poor agreement with the *S*-wave-speed result. The *P*-wave model shows a columnar low-velocity anomaly beneath the Eifel volcanic area, extending down 400 km into the upper mantle. The *S*-wave model (*Keyser et al.*, 2002), on the other hand, detects no anomaly in the depth range ~170 to 240 km depth. If the *P*-wave anomaly is due to high temperature, then the *S*-wave anomaly should be similar but stronger. The *P*-wave anomaly is strongly attenuating in the lithosphere, but weaker in the mantle (*Ritter et al.*, 2002).

The Eifel lithospheric anomaly has been interpreted as a zone of magmatic intrusions. In contrast, the asthenospheric anomaly has been interpreted as representing high temperature (*e.g., Achauer et al.*, 2003). There is no evidence that either the Eifel or the Central Massif anomalies extend through the transition zone and into the lower mantle. There is thus no evidence that they represent diapirs rising by thermal buoyancy from a thermal boundary layer within or at the base of the deep mantle. They may arise from the transition zone in the upper mantle, between 410 and 650 km depths, but there is no evidence that this is a thermal boundary layer. These anomalies resemble that which underlies Iceland, which has robustly been shown to not extend through the transition zone and into the lower mantle (see Iceland webpages).

Low-wave-speed seismic *P*- and *S*-wave anomalies are often automatically assumed simply to correspond to high mantle temperatures, with no influence from other factors. However, temperature is far from the only physical property that influences seismic velocities. Partial melt and compositional variations can also produce low seismic wave-speed anomalies, and their effects may be stronger. The presence of partial melt is particularly powerful in lowering seismic wave speed. A trace of partial melt, even < 1%, could account for the low-seismic-wave-speed anomalies detected beneath virtually all "hot spots", including the sub-areas of the ECVP. Such a trace of melt could result, for example, from a trace of CO_2 (*e.g., Presnall & Gudfinsson*, 2005).

Recent teleseismic tomography and seismic anisotropy results for the upper mantle beneath the Eger rift do not detect a low seismic wave-speed anomaly beneath that magmatically active ECVP sub-area (*Plomerová et al.*, 2007). It seems that such anomalies exist beneath some sub-areas of the ECVP (*e.g.,* the Massif Central and Eifel), but not beneath other recently active areas with similar geochemistry (*e.g.,* Vosges-Black Forest Arch: *Achauer & Masson*, 2002).

4. Discussion and proposed geodynamic model for ECVP formation

Top-down tectonic processes provide a much more reasonable explanation for ECVP magmatism than bottom-up, thermally driven ones. The observations suggest that several relatively small passive diapiric upwellings occurred at various times beneath the ECVP. Alpine orogenesis and rifting, inducing small-scale flow, is the most likely explanation for the magmatism.

Subduction of the European lower continental crust lithospheric mantle had a major influence on the ECVP The continental crust was detached in the Alpine region, and the lower crust was subducted. Such a process, where the lithosphere including the lower crust is subducted, must clearly have influenced the crustal stress field to the north. The sinking slab before break-off may have introduced extensional stresses in the lower crust and lithosphere beneath the foreland, while collision of the upper crust reflected the compressional tectonic setting of the orogeny. The more ductile lower crust may have been strongly thinned locally in the Alpine foreland (ECVP areas), allowing hot asthenosphee to rise. Later slab break-off would have influenced mantle dynamics 300 km to the north, and small-scale convection would have developed as a result of the ongoing deformation.

Most ECVP sub-areas are perpendicular to the main NNE-SSW trending rift system of the Upper Rhine Valley (Figure 1). That rift has been explained as the result of post-Alpine extension. The Miocene to Pliocene basins in central Europe also record this continental rifting phase. The observation that most Cenozoic rift faults cross-cut older Variscan sutures has often been interpreted as evidence that rift development was related to plume activity (*Ritter*, 1999, and references therein), but the rationale for this is unclear.

In the northern Alpine foreland, there is geophysical evidence for localized thermal attenuation of the lithospheric mantle, presumably in response to small-scale convection. *Babuska & Plomerova* (1992) find that lithosphere thickness beneath western and central Europe is uniform and typically 100-140 km. In contrast, it is only ~60 km thick below the SE Rhenish Massif (*Babuska & Plomerova*, 1992), 60-70 km thick beneath the Massif Central (*Sobolev et al.*, 1997) and 80 km thick beneath the Eger Graben (*Babuska & Plomerova,* 2001). The greatest shallowing of the Moho in western and central Europe follows the northern Rhine graben (see Moho map). Crustal thickness beneath the Vogelsberg is less than 30 km. The lower crust is intruded by a strongly reflective zone of basalt dykes at ~20 km depth and crustal underplating may have also occurred (*Braun & Berckhemer*, 1993). Uplift of the asthenosphere–lithosphere boundary may have been caused by local thinning of the ductile lower crust by extension resulting from lower-continental-crust Alpine subduction (Figure 5). The location of the thinned lithosphere is related to older Variscan sutures. For example, the Vogelsberg is located to the NW of the Rheno-Hercynian suture. Buoyant asthenosphere migrated into the thinned lithosphere and probably along the reactivated sutures.

This crustal setting favours the development of small-scale convection. Uplift is expected to result when thick cratonic lithosphere is positioned next to asthenosphere upwelling below a rift (*Vågnes & Amundsen*, 1993; *Kelemen & Holbrook*, 1995). As a result, small-scale convection could enhance melt production during rifting and might be responsible for uplift

(Figure 5). Such uplift would start after development of the rift. This contrasts with the predictions of the mantle plume model, which expects uplift to precede volcanism by several Ma. Partial melting of the mantle in all ECVP areas was likely induced by adiabatic decompression of the asthenosphere. The jumps and onsets of magmatism suggest a common, possibly geochemically similar, fertile mantle source below the sub-areas. For example, in the Rhön, this fertile mantle domain was tapped for ~2 Ma (20-18 Ma), prior to the tapping of a virtually identical reservoir in the Grabfeld area for a comparable period of time (16-14 Ma; Figure 2).

A fertile package in the mantle could arise from old, possibly Variscian, buoyant slab material that was not subducted into the deeper mantle (*e.g.*, a similar situation to that beneath NE China). This slab material thermally equilibrated with the surrounding mantle and the resulting mantle is heterogeneous and contains both depleted and enriched parts. Such material, residing and slowly re-equilibrating in or near the mantle transition zone for a long time, then being drawn up passively in small-scale convection, and melting to a small degree, could also explain the tomographic anomaly beneath Eifel.

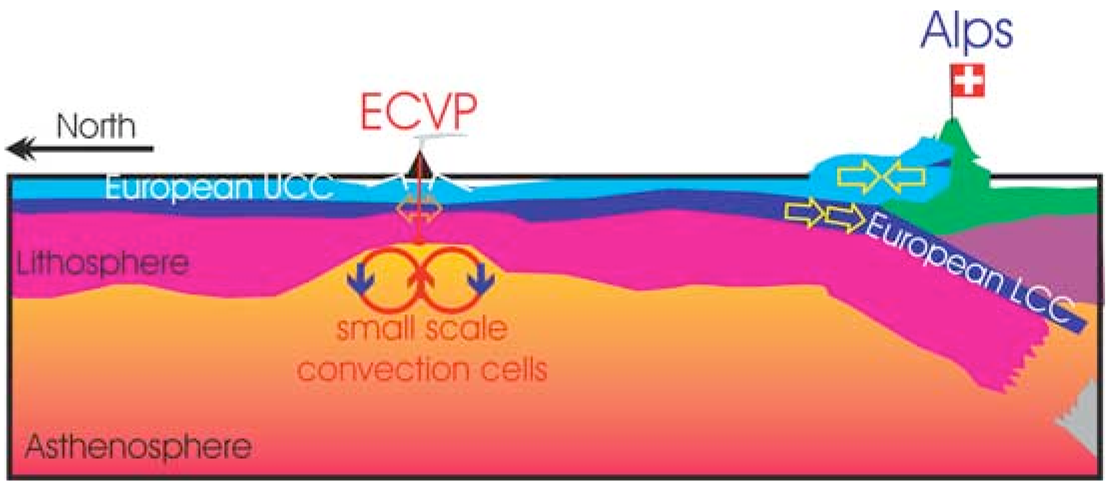

Figure 5: Sketch of the geodynamic model proposed here to explain the origin of the ECVP. Alpine subduction of the European lower continental crust thins the crust in the Alpine foreland at Variscan sutures. This thinning allows the upwelling of asthenosphere and the initiation of small-scale convection below ECVP localities.

Major, trace element and Sr–Nd–Pb isotopic data from ECVP igneous rocks define a common sub-lithospheric mantle source component. This mantle source has geochemical affinities to HIMU oceanic island basalts and is often referred as the European Asthenospheric Reservoir (EAR) and/or the Low Velocity Component (LVC). *Meyer et al.* (2002) proposed metasomatically overprinted sub-continental lithosphere as source. However, the whole mantle may be heterogeneous (*e.g., Meibom & Anderson,* 2004; *Albarède,* 2005) and these heterogeneities may be the source of ECVP magmas. In both models the mantle is a heterogeneous source, but in the second model the “finger plumes” or “baby plumes” observed tomographically may be due to more fertile packages from the mantle transition zone. In response to extensional processes in the continental crust resulting from the Alpine orogenesis, these packages upwell. A similar model has been proposed for Iceland in the North Atlantic Igneous Province (*Korenaga*, 2004). In this model, continental break-up triggered the upwelling of old, deep, large, fertile packages which cause the tomographic signal commonly interpreted as a mantle plume.

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