Chapter 1 –
LATE- TO POST-HERCYNAN EVOLUTION
IN WESTERN-CENTRAL EUROPE

1.1 Introduction

The Hercynian mountain belt is part of a large Paleozoic mountain system, 1000 km broad and 8000 km long, extending from the Caucasus to the east, to the southern Appalachian and Ouachita Mountains in North America to the west, and the Mauritanides in west Africa to the Bohemian Massif in Poland and Czechoslovakia, and Greece and Turkey in south-eastern Mediterranean area. In western Europe, remnants of the chain occur in Spain and Portugal, within the Pyrenees chain, Corsica, French Central massif, Sardinia, Alps and Calabria-Peloritani Orogen (CPO; Fig. 1.1).

Fig. 1.1: Simplified geologic map of European Hercynian Massifs. IB=Iberian Massif, AM=Armorica Massif, MC=Massif Central, VM=Vosges Massif, BF=Black Forest Massif, AA=Aar Batholith, AB= Alboran ; BB=Bernina Batholith, TB=Tauern Batholith, BM=Bohemian Massif, H=Harz Massif, TESZ=Trans European Suture Zone (*Carrigan et al., 2005; modified*).

The Hercynian belt resulted from the accretionary processes developed between 480 and 290 Ma (*Burg et al., 1987; Matte, 1986, Ledru et al., 1989; Matte, 1991, 2001*), that caused the disappearance of numerous oceanic basins and the docking of small and intermediate continental plates to the *Laurentia-Baltica* block, after the closure of the Iapetus Ocean and before the definitive Carboniferous collision between *Gondwana*, to the south, and *Laurussia* (*Laurentia* + *Baltica*), to the North, with the subsequent formation of the *Pangea* supercontinent.

Although the existence of *Pangea* is universally accepted, the identities and geometries of the different colliding blocks are still debated (e.g., *Stampfli and Borel, 2002; Muttoni et al., 2003*).

The Hercynian chain was extensively eroded before the Permian and was dismembered during the Mesozoic plate motion and the opening of the Atlantic ocean, that nowadays divides it into two branches (*Matte, 1991*).

The chain is described by some authors as a system formed by a central ridge and two distinct bordering fold belts (e.g., *Ahrendt et al., 1978*; *Lorenz and Nicholls, 1984*). The central ridge consists of Precambrian and lower Paleozoic rocks affected by HT-LP metamorphism and syn- to post-tectonic granitic plutons. The main period of metamorphism and granitic emplacement ended in the late Carboniferous. In the fold belts the ages of folding and overthrusting decreases with increasing distance from the ridge axis. The presence of these two distinct lateral belts, according to Lorenz and Nicholls (*1984*) would be related to the presence of two different subduction systems that caused the closure of old oceanic basins.

1.2 Plate tectonic models

Plate tectonic models proposed to reconstruct the Paleozoic geodynamic evolution of the Hercynian terranes are numerous. According to Matte (*1991, 2001*), the Hercynian evolution would be related to the presence and migration of continental micro-plates located between the two supercontinents *Laurussia* and *Gondwana*. On the basis of paleomagnetic and paleobiostratigraphic data (*Scotese and McKerrow, 1990*), two are the main micro-plates recognized: *Avalonia* and *Armorica*. These plates are generally considered to have been detached from *Gondwana* during the early

Ordovician and docked to *Baltica* and *Laurentia* before the Carboniferous collision between the two super continents.

However, *Avalonia* according to Matte (*1991*, *2001*) was detached from *Gondwana* during the middle Ordovician opening of the *Rheic Ocean (Nance et al., 2010 and references therein)* and drifted northward independently from *Armorica*; then it was docked to *Baltica* reaching an equatorial position in the Silurian time (*Torsvik, 1998*).

Conversely, *Armorica* remained close to Gondwana in the early Ordovician and then drifted northward *1)* opening a large ocean (*Van der Voo, 1979; Tait et al., 1997*) or *2)* remaining more or less close to *Gondwana* from Ordovician to Devonian times (*Scotese and Golonka, 1992; Torsvik, 1998*) until the definitive collision with *Avalonia*, causing the closure of the *Rheic Ocean.* On the other hand, according to Linnemann et al. (*2004*) Armorica was never detached from west Africa.

However, successively, in the lower – mid Devonian, closure of the *Galicia-Southern Brittany ocean (Matte, 2001)*, caused the final continental collision with *Gondwana* (Fig. 1.2).

The two main oceanic basins, *Rheic* and *Galicia-Southern Brittany* closed by opposite subduction as indicated by the occurrence of HP/UHP metamorphism (430 and 360 Ma) on both sides of the European chain.

After collision, two ocean basins remained open on both sides of suture, the *Paleotethys* and the *Theic oceans*. In late Carboniferous – early Permian, *Gondwana* rotated clockwise relative to *Laurasia* widening the *Paleotethys* and closing the *Theic ocean* between Africa and the southern Appalachians (*Arthaud and Matte, 1977; Bard, 1997; Matte, 2001 and reference therein;* Fig. 1.2).

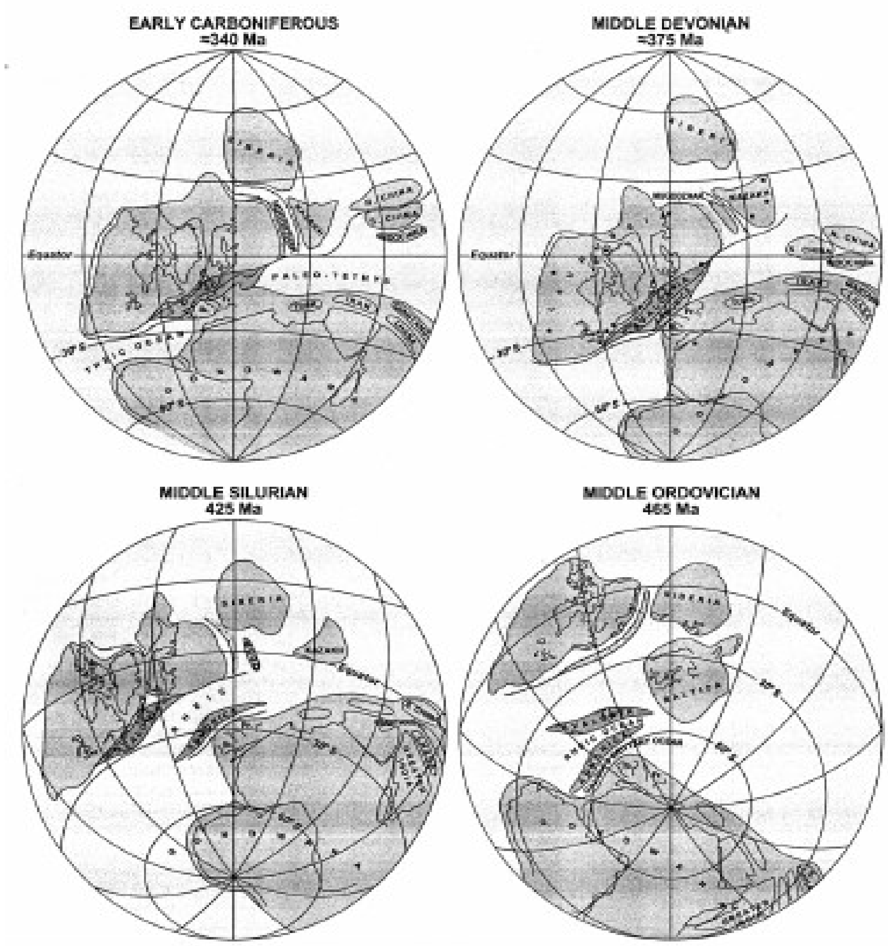

Fig. 1.2: Paleozoic reconstructions from middle Ordovician (465 Ma) to lower Carboniferous (340 Ma; *Matte, 2001*)

Conversely, according to the plate tectonic model proposed by Stampfli and Borel (*2002*; Fig. 1.3), the main continental collision took place during the late Carboniferous. They introduced the concept of "*Hun-terrane*", *European* and *Asiatic,* formed by all the terranes belonging to the northern sector of *Gondwana* and detached from it through the opening of back-arc oceans, following the subduction of the mid-ocean ridge of a former peri-Gondwana ocean, the *Proto-Tethys*. This subduction, during the Ordovician, triggered the break-off of *Avalonia*, that started its northwest drifting towards Laurasia. The opening of Rheic ocean definitely separated Avalonia from Gondwana and the subsequent opening of a back-arc basin, the *Paleo-Tethys*, causes the detachment of the *European Hunic terrane* from the Gondwana super continent.

During the Devonian (380 Ma), the *European Hunic terrane* collided with *Laurussia*, causing the closure of the *Rheic Ocean.*

Finally, *Gondwana* collided with *Laurussia* during the late Carboniferous, causing the consumption of the *Paleo-Tethys* ocean. However, in the Mediterranean area, its final closure did not take place before the upper Triassic.

The northward subduction of *Paleo-Tethys* is considered responsible for the late Carboniferous calc-alkaline magmatism in the Alpine-Mediterranean domain.

Slab roll-back also produced the general collapse of the Hercynian cordillera. Finally, *Neo-Thetys* started to open in the late Carboniferous to early Permian, as a consequence of the increasing slab-pull forces in the *Paleo-Tethys* domain following the subduction of its mid-oceanic ridge below the Eurasian margin. This opening was associated with the drifting of the Cimmerian super-terrane and the final closure of *Paleo-Tethys* in Middle Triassic.

After the final collisional event, between 320 and 280 Ma, extensional tectonics affected the previously thickened lithosphere, by intervening of low-angle normal faults (*Doblas et al., 1994; 1998; Burg et al., 1994*), causing the collapse of the chain and the formation of late Carboniferous to early Permian intracontinental basins (*Menard and Molnar, 1988; Malavieille et al., 1990; Doblas et al.,*1998). The consequent lithosphere thinning and asthenospheric upwelling led to a widespread intrusive and extrusive magmatism that pervaded the central part of the Pangea super-continent during the Permo-Carboniferous (*Finger and Steyer, 1990; 1991; Doblas et al., 1998; Wilson et al. 2004*).

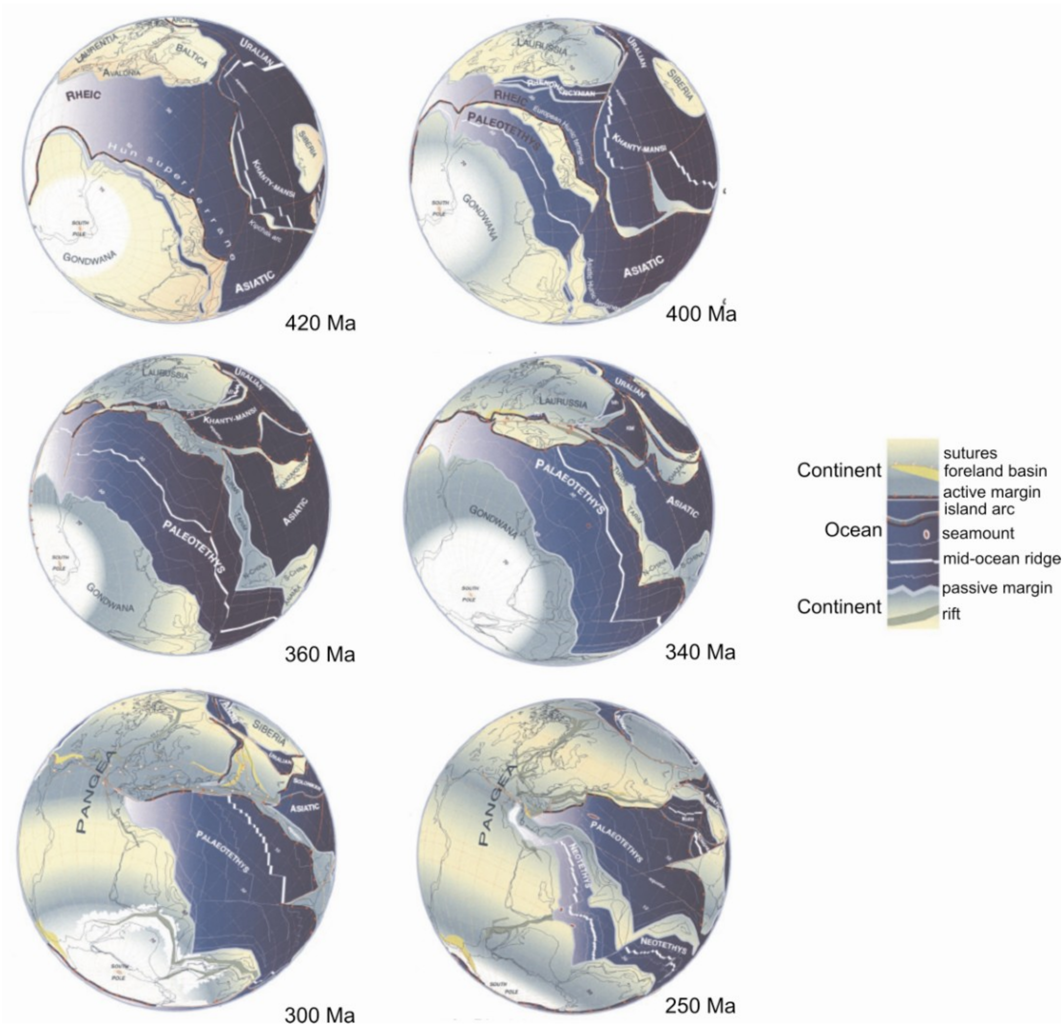

Fig. 1.3: Plate tectonic model proposed by *Stampfli and Borel (2002)* for the Silurian-Permian, with Europe fixed in its present day position

Concerning the tectono-magmatic evolution of the Hercynian chain, Doblas et al. (*1994*) proposed an interpretation for the central Spanish sector of the chain that could be extended to other European sectors. According to them, the tectonic evolution of the chain involved four main tectono-magmatic stages (D1, D2, D3, D4) and three metamorphic events (M1, M2, M3):

- D1, characterized by prevailing compressional conditions, covers the time span from middle Devonian to early Carboniferous. It caused crustal thickening (ca. 50 Km) and the emplacement of small granitic plutons deriving from crustal anatexis processes. A metamorphic event (M1) is associated to this stage and it is characterized by relatively low geothermal gradients, a prograde character and medium P/T conditions (6 - 8 Kbar and 551° C) of Barrovian type (*Arenas et al., 1991*);

- D2, covers a lower to middle Carboniferous time span and is characterized by increasing extensional conditions. This stage is associated with the main granitoid magmatism; it coincides with the M2 metamorphic event developed under higher geothermal gradients and low P/high T conditions, later climaxing at 4-5 Kbars and 750° C (*Bellido et al. 1981; Arenas et al., 1991*).

- D3, took place during the middle Carboniferous -lower Permian and consists of a pure extensional event responsible for the gravitational collapse of the previously over-thickened orogenic belt, through the development of detachment faults.
The climax of the M2 metamorphic event took place during this stage at the same time of widespread intrusion of acidic dykes.

- D4, represents the last deformational stage, developed during the lower – middle Permian and is characterized by generalized EW- oriented extensional stress, that will reach the climax during the Triassic time, and by E-dipping high-angle normal faults and conjugated sets of strike-slip faults. This episode correspond to the "*late Variscan*" event of Arthaud & Matte (*1977*).

1.3 The Hercynian magmatism

The peculiarity of the European Hercynian belt resides in the extreme abundance of granitoids and migmatites (*Zwart, 1967*), emplaced during the time span 370-280 Ma (upper Devonian – lower Permian; *Matte, 1986*), and indicating intense heat transfer processes and partial melting of the crust.

Five main groups of granitoids have been recognized (*Finger et al., 1997*):

1. *Upper Devonian – lower Carboniferous (370 – 340 Ma) "Cordilleran" I-type granitoids,* mainly represented by tonalites and granodiorites. They often have hornblende and occur in association with diorites and gabbros. They can be interpreted as volcanic arc granites, being related to early oceanic subduction. Models involving mantle sources and AFC processes may be feasible.
2. *Early Carboniferous (340 Ma) deformed S-type granite/migmatite associations*, peraluminous, considered syn-collisional crustal melts.

3. *Late Visean – early Namurian (340-310 Ma) S-type and high-K I-type granitoids,* mainly granitic in composition and rarely deformed. It represents the most abundant group probably related to the post-collisional extension and magmatic underplating. Most of these granitoids formed through high-T fluid-absent melting in the lower crust. Partial melting events in the middle crust produced a number of high-T/low-P, S- and I-type diatexites and some S-type granite magmas.

Four sub-groups have been detected:

a) moderately peraluminous S-type granites with some muscovite or cordierite;
b) highly potassic, weakly peraluminous, sometimes metaluminous, K-feldspar megacrystic biotite-granitoids. This group also includes monzonites;
c) Mg-K rich plutons. They are commonly Hbl-bearing melagranites with transitions into monzonite and syenite. Typically, they are associated with small stocks and dykes of coeval mafic to intermediate rocks;
d) leucocratic, near-minimum peraluminous (S-type) granites, commonly fine-grained and with primary muscovite.

4. *Westfalian – early Permian (310-290 Ma),* p*ost-collisional, epizonal 1-type calc-alkaline granodiorites and tonalites*. Such late I-type plutons could be related to renewed subduction along the southern fold belt flank, and/or to extensional decompression melting near the crust/mantle boundary. Post-collisional mantle or slab melting may have occurred in connection with remnant subduction zones below the orogen undergoing thermal relaxation and dehydration.

5. *Late Carboniferous to Permian (300-250 Ma) leucogranites.* The rocks are generally weakly peraluminous or metaluminous. Many of these rocks are similar to sub-alkaline A-type granites. Potential sources for this final stage of plutonism could have been melt-depleted lower crust or lithospheric mantle. On Pearce-type discrimination diagrams *(Pearce et al., 1984*), the rocks plot mostly in the "within-plate-granite" field.

The Hercynian magmatism and in particular the post-collisional magmatism (Carboniferous - Permian), including the major volume of emplaced magmas, is still subject of debate. In particular, still debated are the mechanisms and sources involved in the magma production.

The large variability of geochemical compositions of the produced magmatic rocks could be correlated to a number of factors: composition of parental magmas, type and composition of involved sources, operating mechanisms during the magma ascent able to mask or cancel the original features of melts (e.g., fractional crystallization, magma mixing, crustal contamination, AFC).

Generally, the sources variably invoked to explain this widespread and voluminous magmatic event are the lithospheric and asthenospheric mantle, the continental crust and the old oceanic slab that, according to some authors (e.g., *Visonà, 1982; Lorenz and Nicholls, 1984; Stille and Buletti, 1987; Di Battistini et al., 1988; Finger and Steyrer, 1990, 1991; Bonin et al., 1993; Hsu, 1994; Traversa et al., 2003*), would play a dominant role especially in the production of the calc-alkaline magma.

Nevertheless, several authors reject such contribution since subduction is considered too old (upper Devonian) to influence the post-collisional magmatic activity.

In such a context, two are the main tectono-magmatic models proposed: the first suggests that magmatism originated in response to post-orogenic lithospheric extension, without any relation with subduction processes (*Voshage et al., 1990; Boriani et al., 1992; Quick et al., 1992; Dal Piaz, 1993; Cocherie et al., 1994; Sinigoi et al., 1994; Rottura et al.,1998; Cortesogno et al., 2004a*); the second refers the contrast between orogenic geochemical affinity and post-orogenic to anorogenic setting of the Permian calc-alkaline magmatism to mantle source contamination during previous Hercynian subduction (*Cortesogno et al., 1998; Janousek et al., 2000; Vrana and Janousek, 2006*). Indeed, calcalkaline magma can be produced by decompression melting of an asthenospheric mantle source previously metasomatized by subduction-related fluids/melts (*Johnson et al., 1978; Cameron et al., 2003*) or of a continental lithospheric mantle previously modified by subduction (*Hawkesworth et al., 1995; Wilson et al., 1997*). Besides, calcalkaline compositions have been interpreted in literature also as the result of interaction between mantle-derived melts with continental crust, through concurrent wall-rock assimilation and fractional crystallization (*Stille and Buletti, 1987; Innocent et al., 1994; Rottura et al., 1998; Cannic et al., 2002*). In all these cases, the presence of calcalkaline magmatism does not automatically imply the presence of coeval subduction.

Schaltegger (*1997*) reveled that late Hercynian magmatic activity developed in several distinct short-lived pulse separated by longer periods of inactivity; each episode spanned a few million of years and was characterized by a distinct geochemical signature and geodynamic setting. On the basis of geochronological data obtained from the widespread granitoids of the Central Hercynian Belt, he recognized four different main magmatic pulses developed between 340 – 270 Ma, triggered by mantle upwelling and thermal relaxation of the lithosphere, controlled in space and time by tectonic factors. Such episodic character of the Hercynian magmatism, according to the author would represent an argument against a subduction-related melt formation that should be regarded as a continuous process.

The ages of the Hercynian magmatism are not the same everywhere; Carrigan et al. (*2005*) supplied their U-Th-Pb zircon and monazite ages obtained for representative Hercynian intrusions in the Balkan sector of the orogen (Central Bulgaria). They provide ages in the range of ~315–285 Ma, thus younger than the ages of the post-collisional intrusions in the central European Hercynian Massifs (~340–270 Ma; i.e., Bohemian *(Finger et al., 1997; Klotzli et al., 2001; Gerdes et al., 2003; Janousek et al., 2000; Janousek and Gerdes, 2003)*, Black Forest (*Schaltegger, 1997, 2000*) and Vosges (*Schaltegger et al., 1996; 1999; Schulmann et al., 2002*) although some younger intrusions may also occur.

Within the intra-Alpine massifs, both relatively old (~340–320 Ma; *Schaltegger and Corfu, 1992*) and young (~310–290 Ma; *Schaltegger and Corfu, 1992; Schaltegger, 1994; von Quadt et al., 1994*) granitoids are present, and finally, intrusions in the Iberian Massif appear to be similar to the intrusions of the Balkans (~325–290 Ma; *Bea et al., 1999*; *Fernandez-Suarez et al., 2000*).

The intrusion ages for the Sardinia Corsica batholith, mainly formed by quartz–diorites, monzogranites and prominent leuco-monzogranites, with calc-alkaline affinity, are very similar to those obtained for the central European Hercynian Massifs since ranging between 330–275 Ma (*Cortesogno et al., 1998*, and references therein).

Geochronological data for the OCP granitic intrusions, mainly formed by calcalkaline granodiorites, granites and minor tonalites, from metaluminous to strongly

peraluminous (*Rottura et al., 1990, 1993*), gave tighter variation ranges, supplying ages between 315 – 270 Ma *(e.g., Del Moro et al., 1982; Graessner et al., 2000; Fiannacca et al., 2008).*
According to Carrigan et al. (2005), this diachronism may indicate that collision of adjacent crustal blocks culminated at earlier times in the central and northern portions of the orogen, and moved to east, west, and south as the evolution of the orogen progressed.

Post-collisional magmatism thus produced huge volume of plutonic and volcanic rocks as well as widespread Permo-Carboniferous dyke swarms, of variable thickness and orientation, genetically related to extensional tectonics coupled with asthenospheric upwelling or rising mantle plumes, potentially able to lead to break-up of continental lithosphere (*Ernst and Buchan, 1997; Wilson et al. 2004*).
Products of this so various post-collisional magmatism are widely found in the northern and southern sector of western Europe (Fig. 1.4). Such occurrences are reported from the Pyrenean Chain (e.g., *Innocent et al., 1994; Lago et al., 2004*), the French Massif Central (e.g., *Ledru et al., 2001; Perini et al., 2004*), the Bohemian Massif (e.g., *Janousek et al., 2000; Ulrych et al., 2006*), the Spanish Central System (e.g., *Perini et al., 2004; Orejana et al., 2008*), the Sardinian-Corsica Domain (e.g., *Traversa et al., 2003; Bonin, 2004*) and the Calabria-Peloritani Orogen (*Rottura et al., 1990, 1993; Cirrincione et al., 1995*).

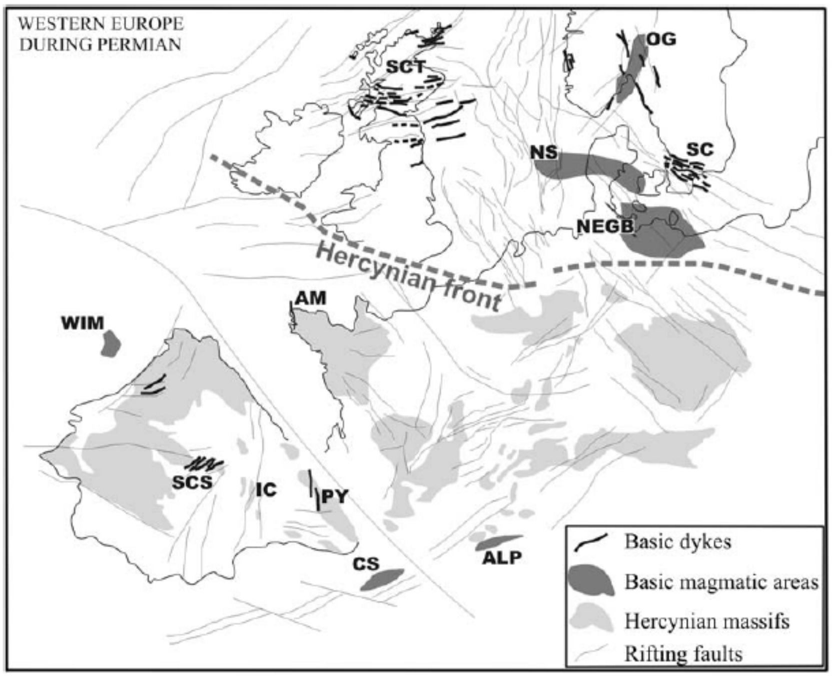

Fig. 1.4: Sketch map after Orejana et al. (2008, and reference therein) showing the location of the main magmatic regions within western Europe during Permian. **WIM:** West Iberian Margin; **P:** Portugal; **PY:** Pyrenees; **IC:** Iberian Chain; **AM:** Armorican Massif; **CS:** Corsica-Sardinia; **ALP:** western and southern Alps; **SCT:** Scotland; **NS:** North Sea; **NEGB:** North East German Basin; **OG:** Oslo Graben; **SC:** Scania.

Early rifting processes are commonly associated with this late Paleozoic magmatic event (*Wilson et al., 2004*), linked to the change in regional stress patterns affecting western and central Europe (*Ziegler, 1990; Ziegler and Cloetingh, 2004*) at the end of orogenic activity in the Hercynian fold belt. The regional uplift, evident in many areas of the Hercynian foreland, has been commonly related to the presence of a widespread thermal anomaly within the upper mantle (i.e. a mantle plume or, possibly, several plumes). Doblas et al. (*1998*) proposed indeed a theory envisaging an upper mantle superplume impinging upon the base of the lithosphere. Such super plume would have been developed over an elliptical area up to 4000 km in diameter stretching from Morocco to the Oslo rift. More recently, this theory was supported by other authors (e.g.: *Nikishin et al., 2002; Perini et al., 2004*).

Elements supporting this theory are, according to Doblas et al. (*1998*): the spatial distribution of the volcanic provinces, the trends of the tholeiitic dyke swarms in the northern sector of Europe (southern Sweden and northern United Kingdom) and the multi-directional pattern of the Permo-Carboniferous fractures within four distinct blocks: northern European, central-west European, Iberian and north-west African (*Ziegler, 1989, 1990; Ernst and Buchan, 1997).*

In some Iberian and African areas, magmatic products show a decrease in age outwards from the central region of the hypothetical Pangean super plume (*Youbi et al., 1995; Lago et al., 1996*). Additionally, in NW Europe and Africa, the magmatic activity often shows a typical evolution of a p*lume-related region*, characterized by initial volcanism with crustal and/or lithospheric mantle signatures, followed by production of HIMU-type melts (*e.g., Cabanis et al., 1990; Innocent et al., 1994; Hofmann, 1997; Neumann et al., 2004; Timmerman, 2004*). Additionally, a high thermal gradient in the central part of the Pangea (Europe and northwest Africa) is indicated by the widespread occurrence in northern Europe (from England to Germany) of hydrocarbons and coal consistent with the preponderant role of plumes in the maturation of hydrocarbon source rocks and the coalification of organic matter (*Larson, 1991*).

1.4 Geochemical signatures of late- to post- Hercynian magmatism

Two different late- to post-collisional volcanic cycles are recognized in western-centrale Europe (e.g., *Cortesogno et al., 1998, 2004a; Bonin et al., 1998; Bonin, 2004; Rottura et al., 1998; Traversa et al., 2003*). The first igneous activity – developed mostly during the Late Carboniferous-Early Permian – produced sub-alkaline low- to high-K calcalkaline, mainly andesitic and rhyolitic, magmas (*Cortesogno et al., 1998, 2004a; Cassinis et al., 2008*), whereas the second – mostly Late Permian-Early Triassic – produced transitional to Na-alkaline magmas (Bonin, 1989; Bonin et al., 1998; Cortesogno et al., 1998; 2004b).

Petrogenetic models and geodynamic interpretations proposed for these magmatic events are still matter of debate. In particular, one of the main object of diatribe concerns the geodynamic significance of the magmas belonging to the first cycle, typically showing convergent plate margin geochemical features (*Lorenz and Nicholls, 1984; Finger and Steyrer, 1990; Dal Piaz and Martin, 1998*), although geological reconstructions suggest a post-collisional context and a transtensional regime, developed within an over-thickened continental crust suffering a gravity collapse of the main chain, associated with the formation of intermontane troughs (*Arthaud and Matte,* *1977; Ziegler, 1993; Cortesogno et al., 1998, 2004a; Lustrino, 2000*). The occurrence of igneous activity with subduction-related geochemical characteristics in post-collisional settings (e.g., far after the end of the oceanic lithosphere subduction) is an anomalous but quite widespread feature, recorded also in younger orogens, like the Alpine Chain (*e.g., Rosenberg, 2004; Lustrino et al., 2011, and references therein*).

The Late Permian-Early Triassic phase, characterized by basic-intermediate Na-alkaline magmas up to evolved compositions (e.g., hawaiitic to A-type rhyolitic dykes in Sardinia and Corsica; *Atzori and Traversa, 1986; Bonin, 1989; Bonin et al., 1998; Traversa et al., 2003*), is instead mostly interpreted as a typical intra-plate event preluding to continental break-up and subsequent formation of the Tethys basins (e.g., *Stampfli et al., 2002*).

However, Permian magmatism does not show everywhere the same geochemical affinity. Figure 1.5 shows the time distribution of late to post-Hercynian basic magmatism of different magmatic affinity (calc-alkaline, alkaline and tholeiitic) within western Europe. Calc-alkaline rocks are totally absent within NW Europe and clearly confined to lower Permian or older times within SW Europe. Besides, they do not coexist with alkaline rocks in the internal zone of the orogen, with the exception of Corsica-Sardinia (*Cocherie et al. 2005)*. A regional tholeiitic event is also recorded, around 295 Ma, in NW Europe (*Heeremans et al. 2004*), while minor volumes of tholeiitic magmas were produced in SW Europe (e.g.: Spanish Central System (*Orejana et al., 2008*); Pyrenees (*Lago et al., 2004*), and not earlier than the late Triassic.

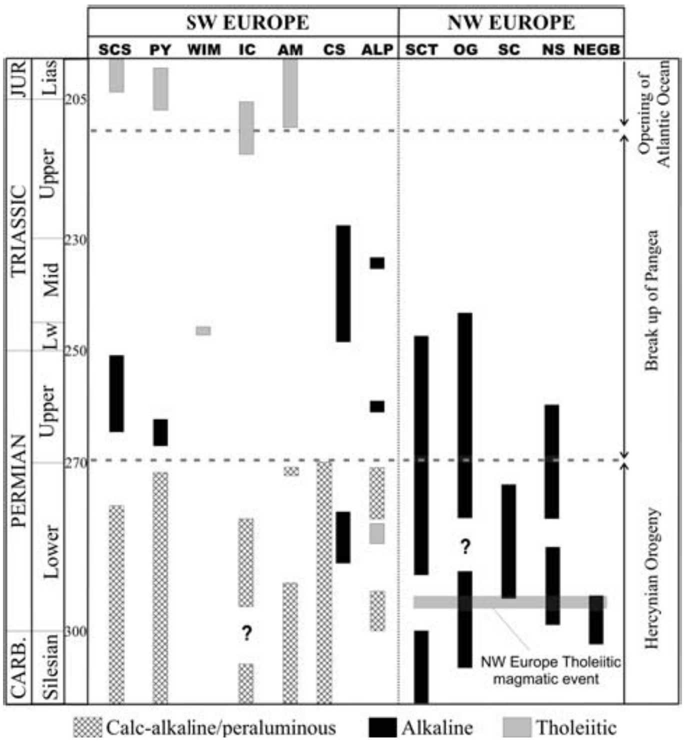

Fig. 1.5: Chronology of the main Permian-Triassic basic magmatism in western Europe (after *Orejana et al., 2008 and references therein*). PY: Pyrenees; WIM: West Iberian Margin; IC: Iberian Chain; AM: Armorican Massif; CS: Corsica-Sardinia; ALP: western and southern Alps; SCT: Scotland; NS: North Sea; NEGB: North East German Basin, OG: Oslo Graben and SC: Scania. Tholeiitic magmatic event in NW Europe according to Heeremans et al. (2004)

1.4.1 Late to Post-Hercynian magmatism in the Calabria-Peloritani Orogen

Late to post-collisional magmatism also occurred in CPO producing intrusive bodies mainly formed by granodiorites, granites and minor tonalites. Two are the magmatic suites recognized in CPO: a main metaluminous to weakly-peraluminous calcalkaline group, and a less extensive suite of strongly peraluminous rocks (*Rottura et al., 1990, 1993; Fiannacca et al., 2008*). The first group, representing about the 70% of the exposed plutonic rocks, show a broad compositional range (~48–70 wt. % SiO_2), with tonalites and granodiorites being the dominant rock types. The strongly peraluminous rocks lack basic to intermediate lithotypes (~67–76 wt. % SiO_2) and contain the typical paragenesis of two micas ± Al-silicates. The metaluminous to weakly-peraluminous granitoids have been interpreted as resulting from the interaction of mantle-derived magmas with lower-crustal melts (*Schenk, 1980; Rottura et al., 1990;*

Fornelli et al., 1994), whereas the strongly peraluminous granitoids have been interpreted either as typical S-type granites, with sedimentary source rocks (*D'Amico et al., 1982; Rottura et al., 1990; Fiannacca et al., 2008*), or as magmas with mixed mantle-crust origin (*Rottura et al., 1991, 1993*). All the plutonic rocks were likely emplaced in an extensional regime, during late- to post-collisional phases in the frame of the Hercynian Orogeny (e.g., Caggianelli et al., 2007; Angì et al., 2010)

Early geochronological data for the timing of Hercynian magmatism in southern CPO, gave ages spanning the Paleozoic-Mesozoic boundary (from ~298 ± 5 to ~270 ± 5 Ma; Rb-Sr whole-rock and mineral ages, zircon U-Pb ages; *Borsi and Dubois, 1968; Borsi et al., 1976; Schenk, 1980; Del Moro et al., 1982*). More recently, strongly peraluminous granites (Cittanova granite) intruding the Serre batholith in its south-western sector, have been dated by ID-TIMS monazite at 303±0.6 Ma (*Graessner et al., 2000*) falling in the range of ~304-300 Ma obtained for the strongly peraluminous magmatism in the whole CPO (ID-TIMS monazite and xenotime ages and SHRIMP zircon ages; *Graessner et al., 2000; Fiannacca et al., 2008*).

Late to post-Hercynian dyke magmatism also occurred in the CPO but it is poor known and published data are very few and mostly concerning the age of calcalkaline dykes emplacement in the Sila Massif (ca. 265 – 295 Ma; U-Pb zircon ages (*Liotta et al., 2008, Festa et al., 2010*), and Rb-Sr muscovite age (*Festa et al., 2010*). Recently, Barca et al. (2010) studied the alkaline to transitional dyke magmatism occurring in the Triassic metasedimentary succession of the San Donato unit (northern Calabria), and Liberi et al. (2011) supply their U-Pb zircon ages of ~224 -296 Ma for gabbros of northern Catena Costiera, classified as tholeiites with MORB affinity. Both magmatic events, according to the authors, would be related to the break-up stages of the Pangea, evolving to rifting associated with the opening of the Jurassic Tethys.

On the other hand, literature data about dyke magmatism of the Serre Massif are very scarce and little is known about its serial affinity and tectonic framework. Only few authors describe the dyke occurrences: Colonna et al. (*1973*) recognize widespread dykes, quite variable in lengths and thickness, crosscutting, with mostly paraconcordant relationships, both phyllites and paragneisses, and the granitoids of Serre upper crust and batholith, respectively. De Vivo et al. (*1992*) describe them as porphyritic felsic dykes and finally, Borsi et al. (*1976*) provide a Rb-Sr biotite age, measured on one of these dykes, yielded a value of 228 ± 7 Ma.

Recently, Angì et al. (2010) recognize the local occurrence of late- to post-Hercynian, usually discordant, weakly deformed, leucogranite dykes, in turn cut by later undeformed ones, crosscutting the mylonitic foliation of the paragneiss of Mammola Complex and clearly representing a post-shear magmatic intrusion stage.

1.4.2 Post-collisional magmatism in the Sicilian - Maghrebian chain

A Mesozoic magmatic activity is largely documented by numerous eruptive products enclosed in various sedimentary sequences of the Sicilian - Maghrebian chain. In particular, submarine eruptive rocks are commonly exposed in western and central Sicily (*Lucido et al., 1978; Montanari, 1987a; Catalano et al., 1984*) and several studies have been carried out, from the 50's to the 90's, about the stratigraphic, paleontological and petrographic features of both volcanic rocks and sedimentary host rocks (*Warman and Arkell, 1954; Christ, 1960; Wendt, 1963; Mascle, 1979; Vianelli, 1964, 1968a, 1968b, 1970; Lucido et al., 1978; Grasso and Scribano, 1985; Grasso et al., 1993*).

The chronology of this magmatic activity is differently reconstructed. Lucido et al. (*1978*) described the magmatism affecting western Sicily, and in particular the Sicano basin and recognized three main phases. The first one took place during the Lias and produces the alkaline magmatic rocks. The second phase is referred to the Lias-Dogger transition. During this phase the geochemical affinity of produced magmas is not univocally determined. Indeed, magmatic products erupted in the marginal zones of the basin have a transitional character, between the tholeiitic and the alkaline series; on the contrary, magmas erupted along the axial portions clearly show a tholeiitic affinity.

During the Cretaceous, although according to the authors the extension continued, magmatic rocks were not produced until the Eocene time, when alkaline magmas were emitted (third phase).

Montanari (*1987b*), based on stratigraphic studies, recognized three different phases:

1. a lower Carnian alkaline phase;
2. a transitional to tholeiitic phase, dated to the Dogger – Malm transition;

3. an upper Oligocene alkaline phase.

In conclusion, it is now generally accepted to consider the entire magmatic cycle, mainly consisting into three main phases: late Triassic, middle Jurassic and late Cretaceous. Additionally, a general agreement exists in considering this magmatic activity as linked to the distensive tectonics that, during the Jurassic, led to the opening of the Tethys ocean (*Lucido et al., 1978; Catalano et al., 1984*).