# Plate tectonics of the Alpine realm

GÉRARD M. STAMPFLI* & CYRIL HOCHARD

*Institute of Geology and Palaeontology, University of Lausanne, Anthropole, CH 1015 Lausanne, Switzerland*

**Corresponding author (e-mail: gerard.stampfli@unil.ch)*

**Abstract:** New field data on the East Mediterranean domain suggest that this oceanic basin belonged to the larger Neotethyan oceanic system that opened in Permian times. A Greater Apulia domain existed in Mesozoic times, including the autochthonous units of Greece and SW Turkey. It also included a united Adria and Apulia microplate since Early Jurassic times. This key information implies that a new post-Variscan continental fit for the western Tethyan area is necessary, where the relationships between the Adriatic, Apulian and Iberian plates are defined with greater confidence. To construct a reliable palinspastic model of the Alpine realm, plate tectonic constraints must be taken into consideration in order to assess the magnitude of lateral displacements. For most of the plates and their different terranes, differential transport on the scale of thousands of kilometres can be demonstrated. This plate tectonic framework allows a better geodynamic scenario for the formation of the Alpine chain to be proposed, where the western and eastern transects have experienced contrasting geological evolutions. The eastern Alps–Carpathians domain evolved from the north-directed roll-back of the Maliac–Meliata slab and translation of the Meliata suture and Austroalpine domain into the Alpine domain. In the western Alps, the changing African plate boundary in space and time defined the interaction between the Iberian–Briançonnais plate and the Austroalpine accretionary wedge.

## Introduction

To construct a plate tectonic scenario for the Alpine domain is not an easy task, as the wealth of information on these regions cannot be grasped, nor properly cited, in a single publication. Therefore here we try to improve on our former attempts at reconstructing one of the most complicated areas of the planet (Fig. 1), by applying plate tectonic concepts as much as possible, and by working on a large enough scale to integrate the Alps into the geodynamic framework of the western Tethyan area, and also integrating information from the Atlantic domain, such as magnetic anomalies (Fig. 2).

Our approach is totally 'non-fixist' – meaning that if two terranes are now juxtaposed they most likely were not so before, and we would like to dedicate this paper to H. Schardt who, more than a century ago (e.g. Schardt 1889, 1898, 1900, 1907), proposed that the Briançonnais domain of the western Prealps was an exotic terrane. Through his 1900 paper 'encore les régions exotiques' (again the exotic regions) one can see that his 'non-fixist' ideas were strongly rejected at that epoch, we expect present geologists will be more open-minded. Schardt's proposal was finally proven correct, and the Briançonnais terrane is a good example of margin duplication in an orogen (Frisch 1979; Stampfli 1993; Stampfli *et al.* 2002).

For a long time, the Tethys was considered as a large and single oceanic space, mostly of Mesozoic age, located between Gondwana and Eurasia. Already in the 1940s and 1950s, a distinction between a Palaeo- and a Neo-Tethys appeared (see references in Sengör 1985; Stampfli & Kozur 2006) and it was recognized that the latter comprised marine Permian and younger strata, whereas the former opened during the Early Palaeozoic. Stöcklin (1968, 1974), following extensive field work in Iran, gave a formal definition of these two large oceanic entities, the Neotethys becoming a Permian to Cretaceous peri-Gondwanan ocean (whose suture was between Iran and Arabia), whereas the Palaeotethys suture was located just north of Iran, thus between the Cimmerian terranes (sensu Sengör 1979) and Eurasia. In that sense, the Palaeotethys separates the Variscan domain from late Palaeozoic Gondwana-derived terranes.

Besides the two large Palaeotethyan and Neotethyan oceanic domains (one replacing the other during the Triassic) many oceanic back-arc type oceans opened just north of the Palaeotethys subduction zone. They are sometimes erroneously considered as Neotethyan because of their Triassic to Jurassic age, but most of these had no direct connection (either geographical or geological) with the peri-Gondwanan Neotethys ocean, and should therefore be called by their local names (e.g. Meliata, Küre, Maliac, Pindos, Huglu,

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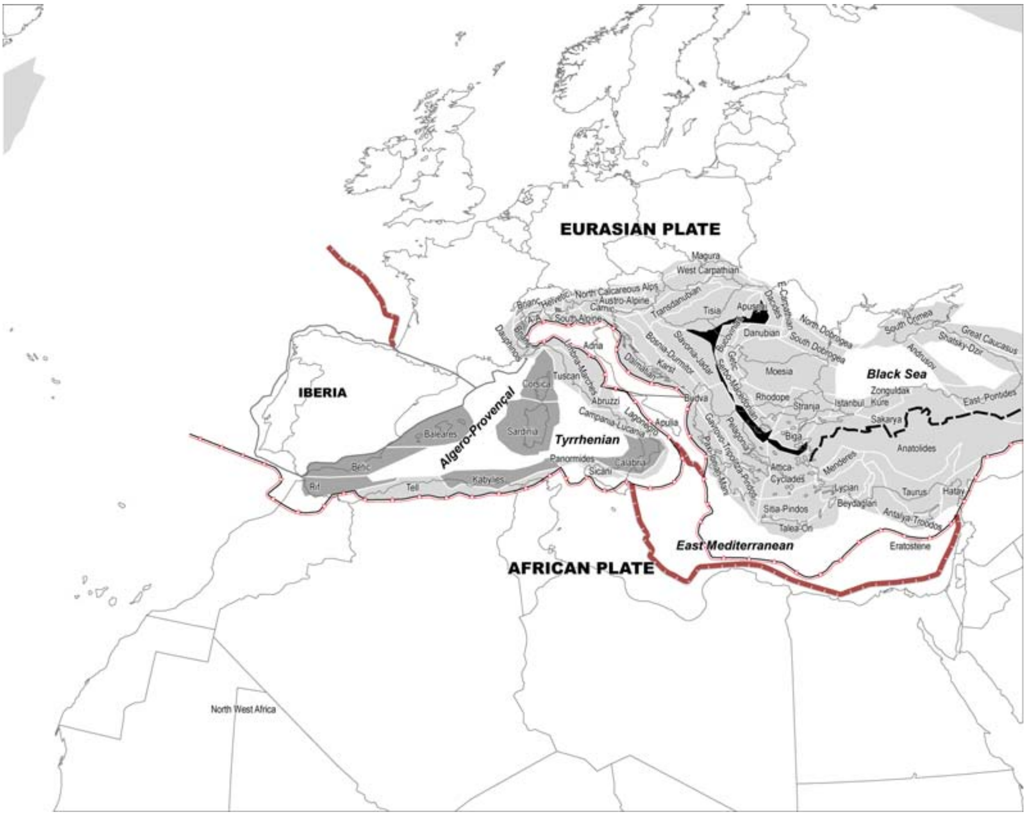


**Fig. 1.** Western Tethys terrane map. Line with triangle marks the limit between the African and Eurasian plates, grey areas are terranes implied in the Alpine orogenic events, darker grey terranes represent fragments of the former Iberian plate. Black area represents the Vardar suture zone, black dash line the Izmir Ankara suture zone. Thick lines with ticks represent the passive margin of respectively the East Mediterranean basin and the Gulf of Biscay.

Vardar) (Stampfli & Kozur 2006). During the break-up of Pangaea, another relatively long, if not wide, oceanic domain appeared, consisting of the Central Atlantic and its eastern extension in the Alpine–Carpathian domain. This Jurassic ocean was named 'Alpine Tethys' (Favre & Stampfli 1992), in order to mark the difference between this relatively northerly ocean (opening in the Variscan domain) and the peri-Gondwanan Neotethys. This Alpine Tethys is made of oceanic segments represented by the Ligurian, Piemont and Penninic/Vahic ophiolitic sequences (e.g. Lagabrielle *et al.* 1984; Liati *et al.* 2005).

The resulting picture of the Tethys realm in Jurassic time is, therefore, quite complex, and made of numerous small oceans and a large peri-Gondwanan Neotethys. Further complexity arose during the convergence stages, as many of these oceanic realms gave birth to new back-arc basins, especially around Turkey and Iran (Moix *et al.* 2008). These are, in most cases, at the origin of the many ophiolitic belts found in the Tethyan realm, whereas older oceanic domains totally disappeared without leaving large remnants of their sea floors (Stampfli & Borel 2002).

## The East Mediterranean Neotethys connection

In the light of the large amount of new data provided by the CROP Atlas (Finetti 2005), the geodynamic evolution of the Adria and Apulia micro-continents has been recently redefined (Stampfli 2005). One of the main issues is the age of the East Mediterranean–Ionian sea basin, and the nature of the sea floor in this area. This, in turn, influences Mesozoic continental re-assembly models.

This problem was already reviewed in some detail in previous publications (Stampfli 2000; Stampfli *et al.* 2001*b*); depending on the authors, the East Mediterranean–Ionian sea basin is regarded

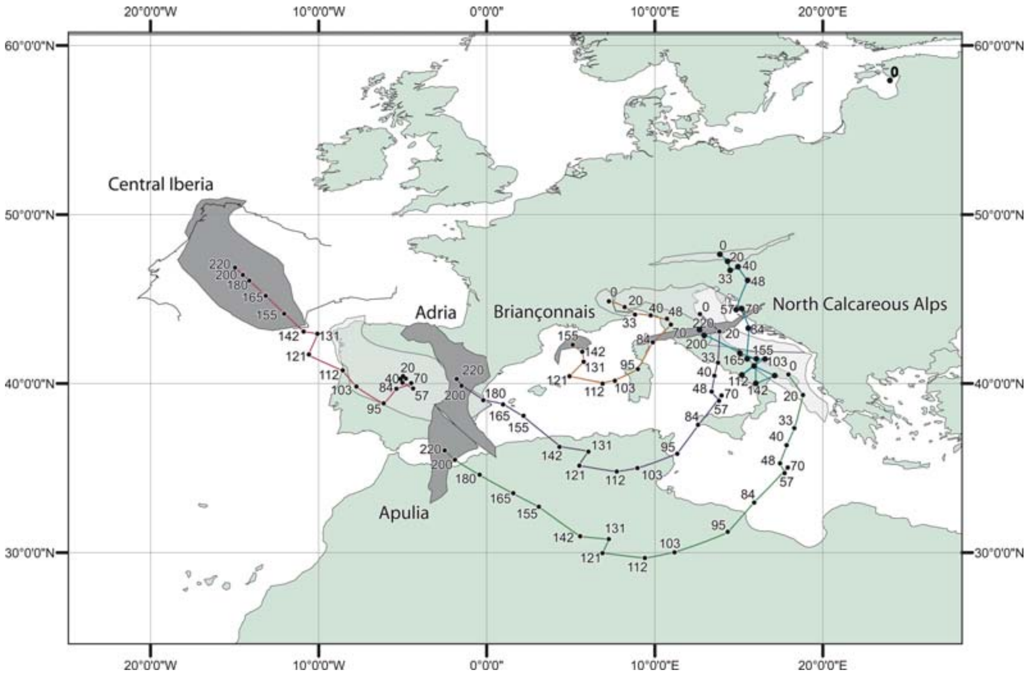


**Fig. 2.** Wander path of some terranes involved in the Alpine orogen. Central Iberia, Apulia, Adria, the western part of the Briançonnais and the North Calcareous Alps are shown in dark grey in their 220 Ma position, and in light grey in their present-day position. Their related velocities are shown in Table 1.

as already opening in the Late Palaeozoic (Vai 1994) or as late as the Cretaceous (e.g. Dercourt *et al.* 1985, 1993). Most people would regard this ocean as opening in the Late Triassic or Early Jurassic (e.g. Garfunkel & Derin 1984; Sengör *et al.* 1984; Robertson *et al.* 1996) and therefore possibly related to the Alpine Tethys–Central Atlantic opening.

A new interpretation (Stampfli *et al.* 2001*b*), showing that the East Mediterranean domain corresponded to a westward extension of the Permian Neotethys, is supported by a large volume of geological and geophysical data. We proposed a Middle to Late Permian onset of sea-floor spreading in the eastern Mediterranean basin, concomitant with the opening of the Neotethys eastward, and the northward drift of the Cimmerian continents since late Early Permian. This model also implies a late closure of the Palaeotethys (Middle to Late Triassic) on a Mediterranean transect (Stampfli *et al.* 2003; Stampfli & Kozur 2006).

Thus, the original position of Apulia with respect to Africa can be well determined by closing the Ionian sea by a *c.* 40° rotation of Apulia around a point located north of Tunisia (Finetti 2005). This represents the total rotation of Apulia from Late Carboniferous to Late Triassic, but a more complicated scheme should be envisaged when considering that Apulia belonged for a while to the drifting Cimmerian continent, whose rotation point changed in time. Therefore, a slightly different placing and rotation of Apulia is shown in the following reconstructions.

## The Apulia–Adria problem and the Late Triassic fit

The late Triassic reconstruction shown in Figure 3 incorporates all the terranes/geological elements shown on the present-day map (Fig. 1). The main problem regarding this model is the position of Adria and Apulia. The present-day shape of these two terranes (present-day Italy) cannot be fitted between Iberia and Africa without some important deformation in classical fits and remains a major problem at the centre of the western Tethyan realm (e.g. Wortmann *et al.* 2001; Schettino & Scotese 2002).

The continuity between the active subduction zone under Greece and the outer Dinarides (Wortel & Spakman 1992; de Jonge *et al.* 1994) shows that there is a possible plate limit between Apulia s.l. and the autochthonous terrane of

**Table 1.** *Velocities and distances to rotation poles of peri-Alpine terranes (see the paths in Fig. 2). Ages are expressed in Ma, and velocities correspond to the centroids velocities and are expressed in cm/a*

| Initial age | Final age | Adria | | Apulia | | Central Iberia | | North Calcarous Alps | | West Briançonnais | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | Velocity (cm/a) | Distance to pole (°) | Velocity (cm/a) | Distance to pole (°) | Velocity (cm/a) | Distance to pole (°) | Velocity (cm/a) | Distance to pole (°) | Velocity (cm/a) | Distance to pole (°) |
| 220 | 200 | 0.26 | 27.71 | 0.44 | 31.92 | 0.16 | 32.58 | 0.52 | 27.48 | 0.00 | 2.58 |
| 200 | 180 | 0.75 | 28.15 | 0.84 | 31.79 | 0.00 | 2.86 | 0.16 | 16.86 | 0.00 | 2.58 |
| 180 | 165 | 0.70 | 36.57 | 1.53 | 29.68 | 0.76 | 33.57 | 0.64 | 52.10 | 0.00 | 42.54 |
| 165 | 155 | 1.33 | 25.06 | 1.53 | 29.18 | 0.75 | 36.80 | 0.89 | 16.44 | 0.00 | 42.54 |
| 155 | 142 | 2.16 | 34.30 | 2.35 | 37.65 | 2.26 | 38.86 | 1.23 | 8.11 | 0.52 | 14.30 |
| 142 | 131 | 1.38 | 32.06 | 1.51 | 35.08 | 0.62 | 5.46 | 0.98 | 40.73 | 0.58 | 11.03 |
| 131 | 121 | 1.08 | 29.94 | 0.97 | 26.68 | 1.48 | 40.58 | 1.26 | 59.34 | 1.18 | 30.71 |
| 121 | 112 | 2.33 | 26.24 | 2.73 | 31.14 | 2.24 | 25.06 | 0.89 | 9.15 | 1.92 | 21.30 |
| 112 | 103 | 1.36 | 10.96 | 2.00 | 16.27 | 1.37 | 11.04 | 1.60 | 17.50 | 0.70 | 5.63 |
| 103 | 95 | 2.89 | 12.92 | 4.05 | 18.25 | 2.29 | 10.20 | 1.00 | 13.78 | 1.73 | 7.69 |
| 95 | 84 | 1.98 | 28.64 | 2.28 | 33.39 | 1.10 | 15.35 | 1.83 | 51.40 | 1.73 | 24.65 |
| 84 | 70 | 1.63 | 17.62 | 2.07 | 22.58 | 0.59 | 6.29 | 0.96 | 12.29 | 1.13 | 21.55 |
| 70 | 57 | 0.25 | 66.92 | 0.24 | 62.41 | 0.27 | 83.39 | 0.23 | 5.04 | 0.00 | 44.67 |
| 57 | 48 | 0.73 | 35.00 | 0.75 | 35.71 | 0.71 | 23.91 | 2.25 | 19.27 | 0.41 | 18.71 |
| 48 | 40 | 1.45 | 23.15 | 1.69 | 27.32 | 0.37 | 6.64 | 1.18 | 6.09 | 1.18 | 8.02 |
| 40 | 33 | 1.42 | 21.43 | 1.69 | 25.83 | 0.17 | 6.09 | 0.58 | 3.47 | 0.88 | 1.74 |
| 33 | 20 | 1.45 | 27.12 | 1.65 | 31.19 | 0.07 | 9.93 | 0.50 | 6.74 | 0.55 | 2.75 |
| 20 | 0 | 0.74 | 36.68 | 0.76 | 37.89 | 0.00 | 40.57 | 0.27 | 4.69 | 0.40 | 1.58 |

Greece. However, we regard this feature as recent and as having no bearing on the fact that a Greater Apulia plate existed in Mesozoic times, in which all the autochthonous portion of Greece was included (PIM, Tor), as well as the Bey–Daglari (Bdg) of SW Turkey, representing the northern margin of the East Mediterranean basin (Moix *et al.* 2008). The apparent present plate limit (between Ap and PIM, Fig. 1) comes from the fact that the Hellenic orogenic/accretionary wedge is oblique with respect to former palaeogeographic domains. It is therefore still colliding with Apulia on an Albanides transect while still subducting the East Mediterranean sea floor on a Greek transect.

The CROP seismic lines through the Adriatic domain (Finetti 2005) clearly show that there is no major tectonic accident cutting Italy into two units, at least since the Jurassic. Thus, Italy must have reached its present configuration between the Triassic and Middle Jurassic. Palaeomagnetic data show that the Apulian plate s.l. (Italy) suffered relatively little rotation in regard to Africa since the Triassic (e.g. Channell 1992, 1996; Muttoni *et al.* 2001).

Our basic hypothesis is that the Apulian part of Italy was definitely an African promontory (Argnani 2002) from Middle Triassic to Recent times, without much displacement with respect to Africa, and that the Adriatic and Apulian microplates were welded in an Eocimmerian collision phase during the Middle–Late Triassic, when both units became part of the African plate.

The need to cut Italy into two microplates comes from the fact that in a Triassic Pangaean reconstruction, there is very little room to insert the present form of Italy in its proper place. In order to solve this dilemma, we had to reconsider the fit of Iberia with Europe, as well as the size and position of the Alboran domain microplates. We made a much tighter fit of these elements with Europe, following similar previous proposals (Srivastava & Tapscott 1986; Srivastava *et al.* 1990). Still, there was not enough room to put the entire present length of Italy in its proper position. The conclusion was that Italy was shorter in Late Triassic times than it is now (or already in the Late Jurassic). A few hundred kilometres were gained through major phases of rifting affecting Greater Apulia since the Triassic and related to the break-up of Pangaea and the opening of the Alpine Tethys [Alpine Lombardian basin in Italy and Ionian rift system affecting the Hellenic domain (Stampfli 2005)].

The reason for widespread Jurassic rifting affecting Italy was that for the Pangaean break-up to succeed, the Atlantic rift system had to join an active plate limit located far to the east in the Neotethyan domain. Thus, all possible ways to break through the Alpine–Mediterranean lithosphere were tried and many resulted in aborted rifts. This pervasive Jurassic rifting finally gave birth, in the Alps, to passive margins, flanked by aborted rifts that became rim basins (e.g. Subbriançonnais, Helvetic-Dauphinois domains, Lombardian basin, Subbetic basin, Magura basin), and a narrow oceanic strip dominated by mantle denudation (Stampfli & Marchant 1997; Rampone & Piccardo 2000). This can be regarded as forced rifting through an already thinned lithosphere, and effectively all this rifting took place under water, with little isostatic/thermal rebound of the rift shoulders.

Thus, the final pre-collisional length and geometry of the Adriatic plate would have been established by Middle–Late Jurassic times only (Figs 4 & 5). Deformation of this plate interior during Alpine times also took place, but only at a small scale as shown by the seismic data, and despite the fact that most of its border was strongly involved in subduction and/or crustal shortening.

The larger Apulia domain had already gone through major rifting phases in the Permian and Triassic, even leading to the opening of small oceans between Greater Apulia and the Austroalpine (AA) domain. This finally led to the subduction of large amounts of continental crust material due to the negative buoyancy of former rift zones and attenuated margins, and regions on which large-scale ophiolite obduction had taken place. On a Hellinides transect this was calculated to be in the order of 900 km of subducted sub-upper-crust continental lithosphere (van Hinsbergen *et al.* 2005), and from our reconstructions a similar amount can be calculated through a Dinarides–Balkan transect (e.g. compare Figs 3 & 12).

## Construction of the models

The reconstructions shown in Figures 3 to 12 are based on a tight pre-Pangaea break-up Permian fit as explained above. From the Early Jurassic onward they are based on magnetic anomalies from the Central Atlantic. Plate tectonic concepts have been systematically applied to our palinspatic models of the western Tethys, moving away from pure continental drift models, not constrained by plate limits, to produce a model which is increasingly self-constrained. In this approach, first explained in Stampfli & Borel 2002, inter-dependent reconstructions are created from the past to the present. Except during collisions, plates are moved step by step, as single rigid entities. The only evolving elements are the plate boundaries, which are preserved and follow a consistent geodynamic evolution through time and an interconnected network through space. Hence, lithospheric plates are constructed by adding to, or removing oceanic

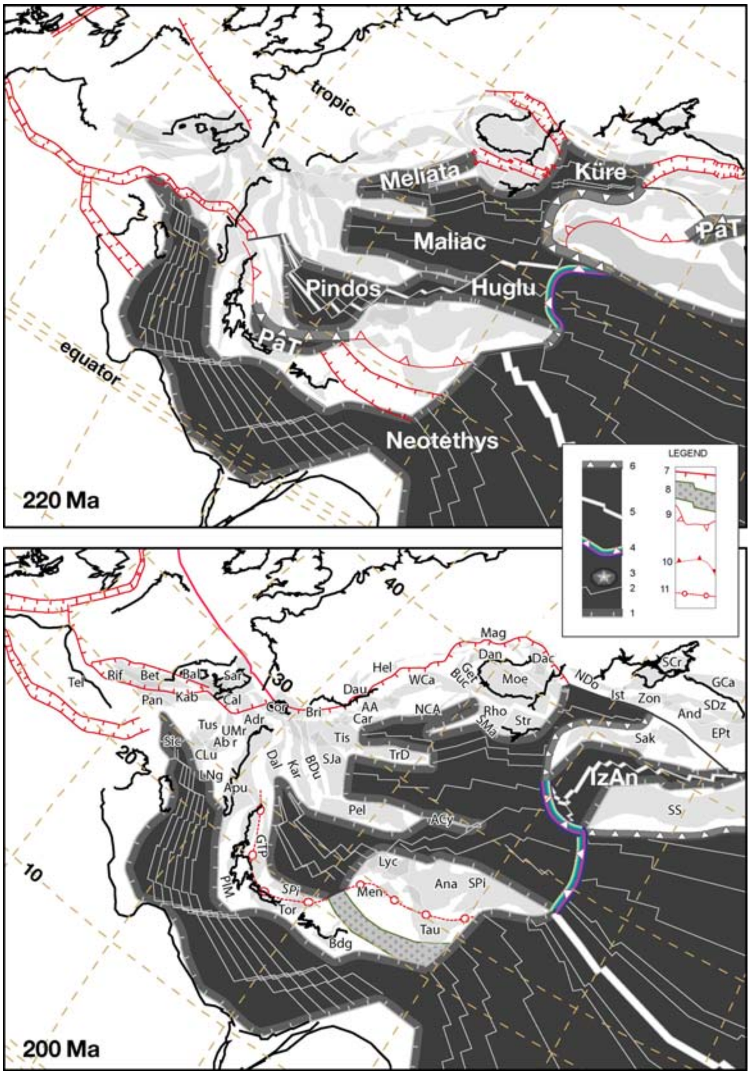


**Fig. 3.** 220–200 Ma. *Late Triassic*. As the Palaeotethys (PaT) subduction reached its final stage, slab roll-back along its northern margin accelerated and was marked by the opening of successive oceanic back-arc basins before the final closure of Palaeotethys in late Carnian times. Extension affected southwestern Eurasia and north Africa. A possible link with the north Atlantic rift system existed through the Pyrenean rift. The East Mediterranean part of Neotethys ceased spreading; on its northern margin, Greater Apulia (LNg, Ap, PIM, Tor, Bdg, Tau) represents at this time the westernmost part of the Cimmerian terranes, detached now from the Iranian blocks (SS). *Triassic-Jurassic boundary*. At this time Palaeotethys was completely closed. The Central Atlantic rift widened but had difficulty finding a way to link with a plate limit to the east. The closure of Palaeotethys south of the Küre basin generated the southward subduction of the latter. Slab roll-back, both in Küre and the Neotethys, allowed the opening of the Izmir-Ankara (IzAn) ocean. See Figure 1 for abbreviation. All reconstructions are in spherical equidistant projection, centred 20E20N. *Symbols*: 1: passive margin; 2: magnetic or synthetic anomalies; 3: seamount; 4: intraoceanic subduction; 5: mid-ocean ridge; 6: active margin; 7: active rift; 8: inactive rift (basin); 9: collision zone; 10: thrust; 11: suture. Oceanic lithosphere in black. Abbreviations for Figures 3 to 12: AA, Austro–Alpine; Abr, Abruzzi; ACy, Attica–Cyclades; Adr, Adria; Ana, Anatolides; And, Andrusov; AnT, Antalya; APr, Algero–Provençal; Ap, Apulia; Apu, Apuseni; Bal, Baleares; Bdg, Beydaglari; BDu, Bosnia–Durmitor; Bet, Betic; Big, Biga; Bri, Briançonnais; BS, Black Sea; Buc, Bucovinian; Bud, Budva; Cal, Calabria; Car, Carnic; CLu, Campania–Lucania; Cor, Corsica; Dac, Dacides; Dal, Dalmatian; Dan, Danubian; Dau, Dauphinois; DoN, Dobrogea North; DoS, Dobrogea South; ECa, East-Carpathian; EPt, East-Pontides; Era, Eratostene; GCa, Great Caucasus; Get, Getic; Gos, Gosau; GTP, Gavrovo-Tripolitza-Pindos; Hat, Hatay; Hel,

material (symbolized by synthetic isochrones) from major continents and terranes.

In recent years we changed our tools and moved into GIS softwares and built a geodynamic database to support the reconstructions, and the model was, and still is, extended to the whole globe (Hochard 2008). An example of this new approach can be found in Ferrari *et al.* (2008).

Most geodynamic/geological constraints and data used for the reconstructions can be found in our previous publications (Stampfli 2000; Stampfli *et al.* 2001*a*, *b*; Stampfli & Borel 2002, 2004). A new global terrane, or rather geological elements map, was established following the approach outlined in Ferrari *et al.* (2008). Part of this map is presented in Figure 1, where only the elements displaced during the Alpine orogenic event are shown. The present-day outlines of the elements should be regarded as geographic markers, their shape having little to do with their original shape, generally most larger, excepted for elements having gone through large-scale Tertiary extension, such has Corsica or Sardinia, or the Cyclades elements. Due to the GIS database, many geodynamic/kinematic information can be derived from the reconstructions; one of the most interesting is the velocity map of the moving terranes, and their wander paths in respect to a fixed Europe (Fig. 2), which can be established here with great precision. This quantitative information can then be confronted to major structural patterns and kinematic indicators in the field; they show the changing direction and speed of convergence in space and time.

## The geodynamic evolution of the larger Alpine area

We shall review some of the major steps of the peri-Alpine evolution used to constrain the reconstructions; Palaeotethys ocean evolution was recently reviewed in Stampfli and Kozur (2006) and is not repeated here. We also do not discuss alternative models, quite numerous, and not often taking into consideration plate tectonics principles, but always very useful for the geological constraints they offer.

### *The Jurassic ocean: Alpine Tethys, Central Atlantic and Vardar (Figs 3, 4 & 5)*

The results of field-work on the Canary Islands and in Morocco (Favre *et al.* 1991; Favre & Stampfli 1992; Steiner *et al.* 1998) indicate that the onset of sea-floor spreading in the northern part of the Central Atlantic occurred in the Toarcian. Similar subsidence patterns between this region and the Lombardian basin (Stampfli 2000) led us to propose a direct connection between these areas. The Lombardian basin aborted in Middle Jurassic times (Bertotti *et al.* 1993) as it could not link up with the nearby oceanic Meliata-Maliac domains whose already cold lithosphere was rheologically considerably stronger than the surrounding continental areas. Therefore, the Alpine Tethys rift opened to the north of the Meliata basin, separating Adria and the Austro-Carpathian domain from Europe (Bernoulli 1981). Thermal subsidence of areas flanking the Alpine Tethys commenced in the Aalenian in the west (Briançonnais margin: Stampfli & Marchant 1997; Stampfli *et al.* 1998, 2002) and in the Bajocian further to the east (Helvetic and Austroalpine margin: Froitzheim & Manatschal 1996; Bill *et al.* 1997). The Alpine Tethys ocean spreading was considerably delayed with respect to the Central Atlantic; very slow spreading gave birth to a limited amount of oceanic crust, the oceanic area being dominated by continental mantle denudation. A larger transform Maghrebide ocean linked the central Atlantic and the Alpine Tethys, and was also characterized by delays in thermal subsidence (e.g. Rif area, Favre 1995; Stampfli 2000).

Within the Alpine domain, there is a fundamental difference between the AA–Carpathian and Western Alps systems. The AA–Carpathian evolution was rooted in the dynamics of the Triassic back-arc basins located to the south (Meliata–Maliac domain). These back-arc basins were shortened in conjunction with the opening of the Central Atlantic and rotation of Africa with respect to Europe. Subsequent slab roll-back of the subducting Küre, Maliac–Meliata oceanic lithosphere induced opening of the Vardar back-arc ocean, which by Late Jurassic times had completely replaced the pre-existing oceanic basins (Fig. 5).

**Fig. 3.** (*Continued*) Helvetic; Ion, Ionian; Ist, Istanbul; Kab, Kabylies; Kar, Karst; LNg, Lagonegro; Lig, Ligurian ocean; Lom, Lombardian; Lyc, Lycian; Mag, Magura; Men, Menderes; Moe, Moesia; NCA, North Calcareous Alps; NDo, North Dobrogea; Pan, Panormides; Pel, Pelagonia; Pen, Penninic, Vahic ocean; Pie, Piemontais ocean; PIM, Paxi-Ionian-Mani; Pyr, Pyrenean ocean; Rho, Rhodope; Rif, Rif; SA, South Alpine; Sak, Sakarya; Sar, Sardinia; SCr, South Crimea; SDz, Shatsky–Dzir; Sic, Sicani; SJa, Slavonia–Jadar; SMa, Serbo–Macedonian; SPi, Sitia–Pindos; Sre, Srednogorie; SS, Sanandaj–Sirjan; Str, Stranja; SBe, Sub–Betic; Tau, Taurus; Tel, Tell; Tis, Tisia; Tor, Talea–Ori; TrD, Transdanubian; Tro, Troodos; Tus, Tuscan; Tyr, Tyrrhenian; UMr, Umbria–Marches; WCa, West Carpathian; Zon, Zonguldak–Küre.

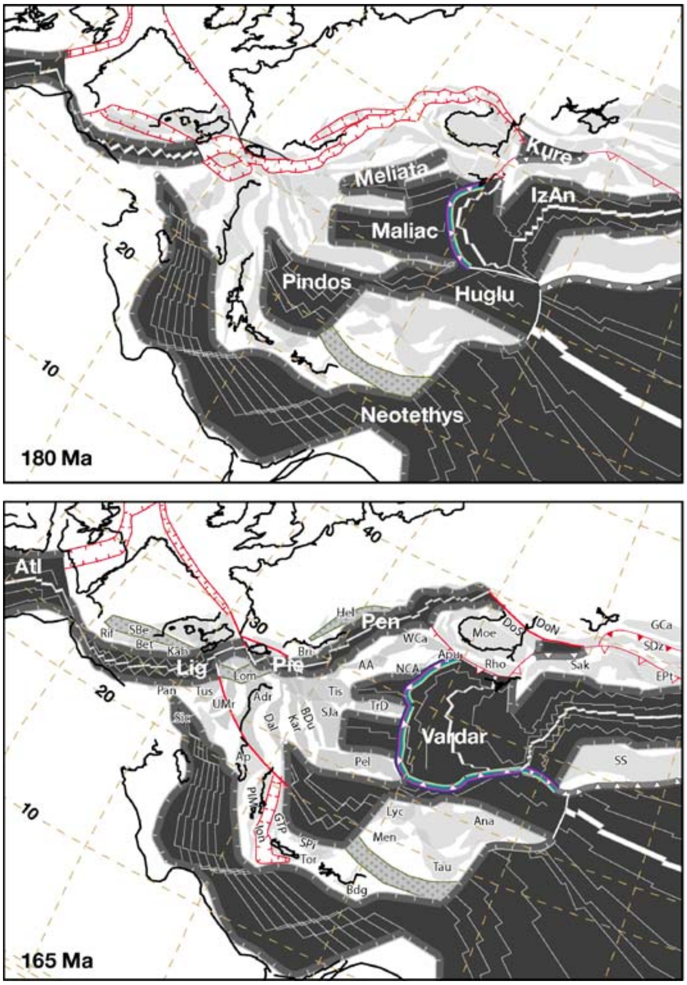


**Fig. 4.** 180–165 Ma. *Toarcian; Bathonian*. Spreading is now active in the Central Atlantic and the segment of the Alpine Tethys located south of Iberia (Lig). Eastward in the Carpathians domain, the Alpine Tethys successfully opened in the Bajocian as it was able to connect with a plate limit through the Dobrogean transform (DoS, DoN) and from there to the active subduction zone around the Sakarya plate. The Izmir–Ankara ocean is a back-arc of the Neotethys, whereas opening of the Vardar corresponds to subduction progradation from the Küre domain toward the Maliac domain. Accelerating roll-back of the Maliac sea floor generated northwestward spreading of the Vardar basin in a scenario of intra-oceanic subduction. Closure of the Küre basin was entering a collisional stage around the Rhodope promontory. Around the Alpine Tethys, aborted rifts become rim basins (sub-Betic, Helvetic–Vocontian, Lombardian) and rifting is still active from Italy to Greece (Ionian zone).

Continued rotation of Africa provoked ridge failure in the Vardar and large-scale Late Jurassic ophiolitic obduction onto the Dinaride–Hellenide passive margin of the Pelagonian terrane (e.g. Laubscher & Bernoulli 1977; Dercourt *et al.* 1986). Roll-back of the oceanic Maliac-Meliata slab was a centrifugal phenomenon that controlled successive collision of the Vardar arc with all passive margins surrounding the Meliata–Maliac basin. Accompanying the obduction on the western Dinaride margin in Late Jurassic times, the northeastern part of the Vardar arc-trench system collided with the northern Meliata passive margin (the Northern Calcareous Alps: NCA, Fig. 4) (Bernoulli 1981), the Western Carpathian domain (Csontos & Vörös 2004) and the Rhodope. In the latter, remnants of the Vardar arc are found in northern Greece and Bulgaria as tectonic klippen (Bonev & Stampfli 2003, 2008). Closure of the Balkan rift system between Moesia and the Rhodope (Figs 5 & 6) controlled

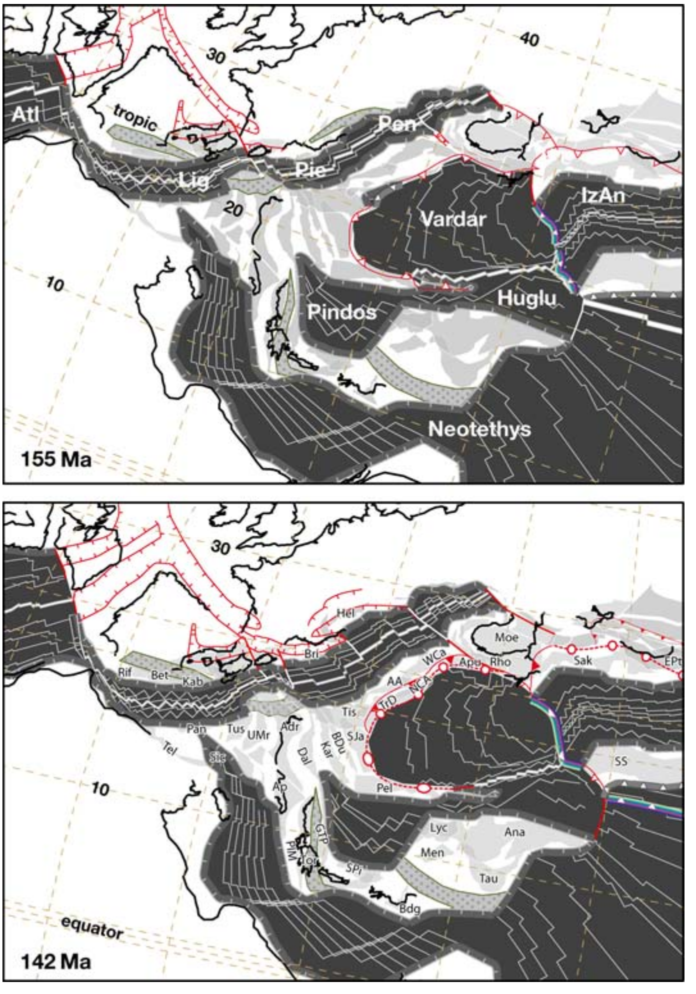


**Fig. 5.** 155–142 Ma. *Kimmeridgian; Berriasian*. Spreading in the Alpine Tethyan ocean now reached the Carpathian domain. In the process, Moesia was detached from Europe by only a few hundreds of kilometres, the point of rotation of Gondwana being located close to Moesia. Northward extension of the Central Atlantic triggered active rifting between Iberia and Newfoundland, and the northern limit of Iberia was affected by rifting extending eastward into the Briançonnais domain, through the Pyrenees and Provence. Extension is affecting also the Helvetic-Dauphinois rim basin. The Küre ocean was totally closed; collision of its arc-trench system with the Rhodope was causing the first phases of the Balkan orogeny, accompanied by inversion of former rift zones. Roll-back of the Meliata–Maliac slab allowed rapid westward expansion of the Vardar back-arc basin, its arc trench system colliding with the Pelagonian and Dinaric (SJa,Tis) landmass. This is accompanied by east–west shortening in the Maliac–Vardar domain due also to the anti-clockwise rotation of Gondwana with respect to Europe. We regard this event as creating a change in spreading direction in the western Neotethyan domain, at the origin of the mid-ocean ridge failure, south of the Sanandaj–Sirjan block (SS).

subsequent development of the Balkan orogen, accompanied by large-scale Early Cretaceous northward nappe emplacement and metamorphism (Georgiev *et al.* 2001). This circum-Vardar orogenic event commenced in the Middle Jurassic and was sealed by Albian to Cenomanian molasse-type sediments in the Balkans (Georgiev *et al.* 2001), and mid-to late Cretaceous platforms in the Hellenides–Dinarides.

Along the NCA margin, elements of the AA micro-continent were scraped off and incorporated into the accretionary wedge, to form the different internal units of the AA–Carpathian orogen (Kozur 1991; Plasienka 1996; Faupl & Wagreich 1999; Csontos & Vörös 2004). This event was accompanied by Early Cretaceous HP–LT metamorphism (e.g. Thöni & Jagoutz 1992; Ivan 2002). Subsequently, the enlarged AA orogenic

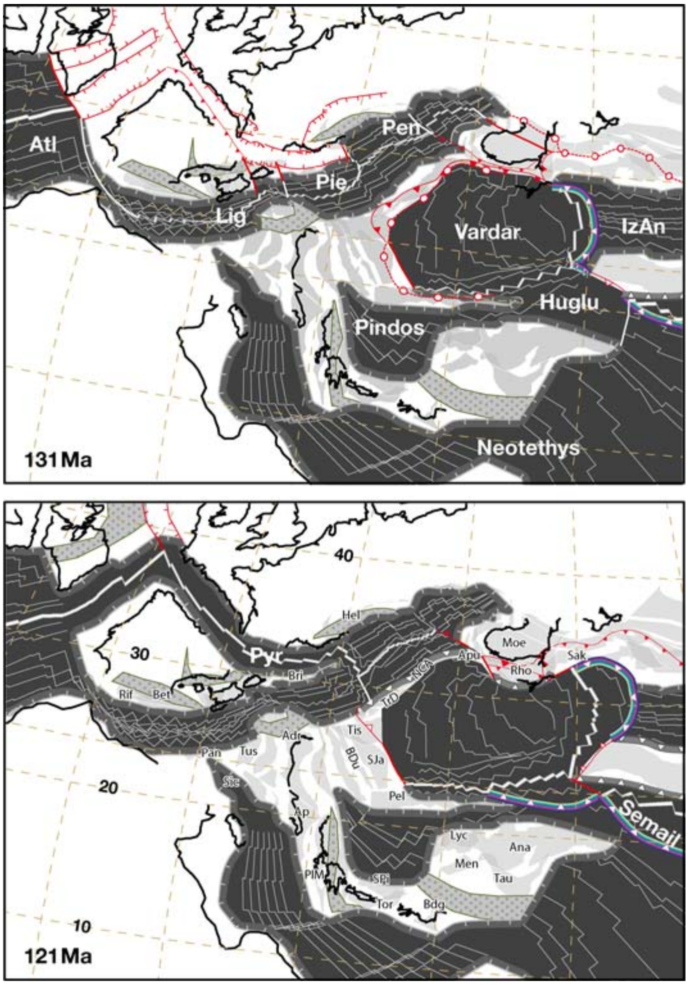


**Fig. 6.** 131–121 Ma. *Hauterivian; Aptian*. Accelerating anti-clockwise rotation of Gondwana was responsible for the obduction of part of the Vardar mid-ocean ridge system onto the Pelagonia, Dinaride (SJa) and Tisia blocks. Collision of the Vardar arc-trench system with the AA block was taking place at this time, detaching the future NCA domain from its basement (internal AA). Collision of the Vardar arc-trench system continued also in the Balkan, where parts of the Rhodope cover and basement were thrust northward . The major changes affecting the Neotethyan domain brought to an end the opening of the Izmir–Ankara back-arc basin system. The Izmir–Ankara slab started retreating eastward, allowing the opening of a new supra-subduction spreading centre. In the western Neotethys, ridge failure generated a new intra-oceanic subduction zone along the former spreading centre; this new oceanic domain will eventually obduct onto Arabia (e.g. Semail ophiolites). The Iberian plate was totally detached from Laurasia, whereas spreading stopped in the Ligurian-Piemont part of the Alpine Tethys. In the eastern Penninic segment of the Alpine Tethys, subduction progradation brought the exotic AA terranes onto the Alpine Tethys sea floor. Orogenic processes were soon to come to an end in the Balkan (sealed by Albian molasses), whereas the Vardar intra-oceanic system extended southward and eastward.

wedge began to move over the eastern segment of the Alpine Tethys (Figs 6 & 7) (Penninic–Vahic ocean), to finally collide with its northern passive margin (Helvetic domain s.l., Magura rim basin) thus forming the present Eastern Alps and Carpathian orogen (Wortel & Spakman 1993). This process involved continuous slab roll-back that had commenced during the Carnian in the Küre domain in Turkey (Stampfli & Kozur 2006) and continued into the Neogene period in the Eastern Carpathians.

The Cretaceous AA accretionary wedge was affected by large-scale collapse in the Late Cretaceous (Figs 8 & 9), accompanied by the development

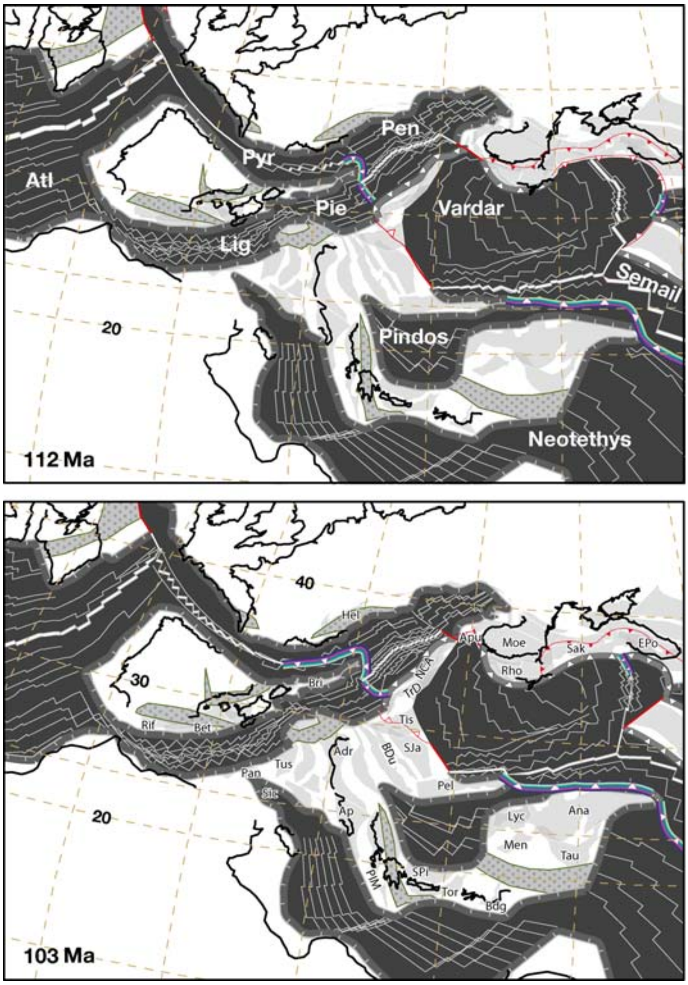


**Fig. 7.** 112–103 Ma. *Albian*. The absence of magnetic anomalies between the M0 and C34 anomalies reduces the constraints on the plate tectonics evolution during the intervening 37 Ma. However, the reconstructions are strongly constrained by the transform movement between the African and Indian plates during that time, leaving very little place for speculation. The Pyrenean 'ocean' opened and closed rapidly due to the rotation of Africa. The Briançonnais peninsula becomes an upper plate promontory slowly advancing onto the Penninic ocean. A flexural basin developed along the northern margin of the Izmir–Ankara ocean, with emplacement of ophiolitic mélanges on the Sakarya–East Pontides domain, preceding subduction reversal. The southern Vardar and Semail intra-oceanic back-arc system was expanding southward following slab retreat of the Huglu and Neotethys oceans.

of the Gosau basins (e.g. Faupl & Wagreich 1999). Before that, an obstacle prevented such a collapse, and we propose that spreading lasted in the Penninic–Vahic ocean until mid-Cretaceous times, the collapse of the accretionary wedge being only possible when the mid-oceanic ridge had been subducted. Recently dated gabbros in the Monte Rosa nappe have given Cenomanian ages (Liati & Froitzheim 2006). These gabbros are interpreted by these authors as belonging to the 'Valais ocean', but in view of their structural position, and the necessary east–west convergence between the Iberian plate and the Penninic ocean (Figs 7 & 8), it is more likely that these gabbros belong to the latter and were underplated below the Piemont/Penninic Zermatt–Saas ophiolites.

### *Cretaceous oceans: North Atlantic and the Pyrenean domain*

From magnetic anomalies in the Atlantic domain, at least from the Aptian (M0 magnetic anomaly,

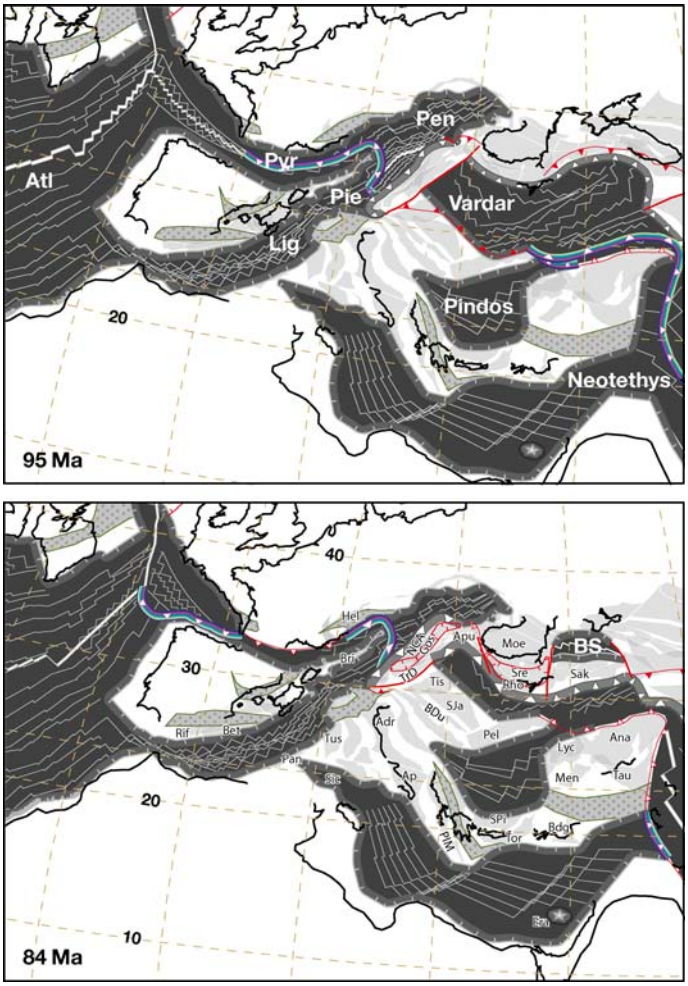


**Fig. 8.** 95–84 Ma. *Cenomanian; Santonian*. The northern limit of the African plate in the Pyrenees became a zone of convergence/subduction, partly extending into the Biscay–Atlantic domain, and generating large-scale inversion from the Pyrenees to the western Alps. The Penninic mid-ocean ridge is subducted and the supra-subduction Gosau rift started to open within the Austroalpine accretionary wedge. East–west shortening in the Alpine-Vardar region brought Adria behind the Austroalpine prism, whereas the north–eastward subducting Vardar remnant ocean generated an active margin setting in the Balkan, accompanied by the opening of the Srednogorie (Sre) and western Black Sea (BS) back-arc basins. The southern Vardar ridge obducted southward on the Anatolide block, and linked eastward with the Troodos–Semail obduction system. Finally, the Anatolian-Tauric plate was nearly totally covered by an ophiolitic-type mélange. The accelerated rotation of Africa narrowed down significantly the Neotethyan domain south of the Semail ocean. The Eratostene seamount could be related to volcanic activity in the Levant during the Cretaceous.

Fig. 7) to the Maastrichtian (Fig. 9) (Stampfli & Borel 2002) and most likely up to the Thanetian (anomaly C25, Fig. 9), it is observed that Iberia rotated with Africa and Apulia–Adria, without appreciable North–South shortening (Fig. 2); it also implies that spreading stopped in the western branch of the Alpine Tethys. Starting in the Early Cretaceous (Figs 6 & 7) the Pangaean break-up had increasing difficulty in linking up eastward with another plate boundary. From divergent, the African plate limit in this region had become convergent, due to intra-oceanic subduction in the Neotethys and shortening affecting the Vardar region. Also, the African plate was now separated from South America and India and started to rotate more on itself. As a result, the Pyrenean-Biscay ocean opened, with a rifting phase starting in the Oxfordian, and the onset of ocean spreading in the Aptian for the Portuguese–Galician ocean and Pyrenean area, and in the Albian for the Gulf of

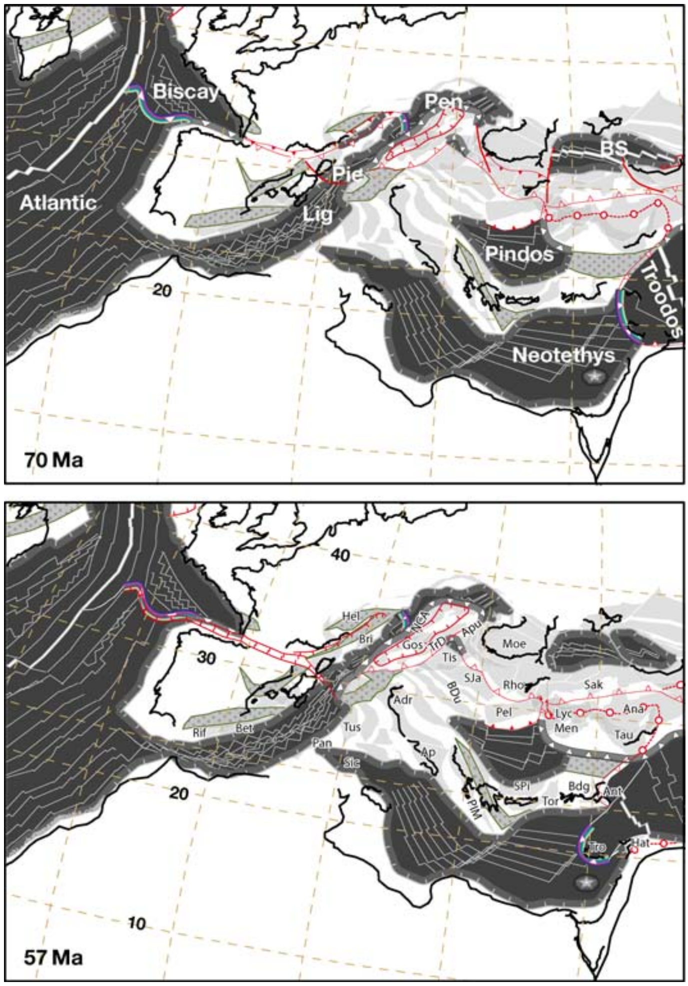


**Fig. 9.** 70–57 Ma. *Maastrichtian; Thanetian*. In the Alps, the northern sinistral Adriatic plate boundary was extending westward into the Piemont ocean, defining a temporary Corsica-Briançonnais plate, which started subducting southward beneath Adria. This was also allowing the AA prism to collapse westward and the supra-subduction Gosau rift to continue to open. North–south shortening was taking place along the northern boundary of the Iberian plate before a slight phase of extension in the Palaeocene. East–west shortening was still very active in the Alpine and Vardar domains, passing into continental subduction. Roll-back of the remnant oceanic slab allows the opening of the narrow east Black Sea back-arc basin. Counter-clockwise rotation of Africa was responsible for the obduction of the Semail ocean on the Arabia margin (from Oman to Syria, Hatay), whereas intra-oceanic subduction persisted in front of the Troodos plate.

Biscay, as clearly shown by peri-Iberian subsidence curves (Borel & Stampfli 1999; Stampfli *et al.* 2002) and a clear mid-Cretaceous thermal event in the Pyrenees (e.g. between 112 and 97 Ma; Schärer *et al.* 1999).

By the end of the Santonian (84 Ma, Fig. 8), the break-up between North America and Greenland took place, and in the Campanian the Biscay spreading aborted (Olivet 1996). Closing of the Pyrenean domain already began during the opening of the Gulf of Biscay (Vergés & Garcia-Senz 2001), due to the accelerated rotation of the Iberian plate together with Africa (Fig. 8).

It is uncertain how wide the Pyrenean 'ocean' was, and whether it was limited to mantle denudation as indicated by the Palaeozoic lherzolites at Lherz (Fabries *et al.* 1998; Lagabrielle & Bodinier 2008). In order to control the geometry of the plate limit, a tentative ridge is put on the reconstructions for the Pyrenean 'ocean', but as was the case in large parts of the Alpine Tethys (see above), we think that only mantle denudation took place in this domain. A minimum extension of 60–80 km can be calculated from restored cross-sections through the Pyrenees (Vergés & Garcia-Senz 2001). However, from our reconstructions we come to a larger amount (*c.* 200 km), similar to the present northern Red Sea, where indeed no sea-floor spreading has yet taken place. Obviously, most of the distal margins have been subducted/eroded in the Pyrenean domain, as is the case in the Alps.

The rotation of Iberia finally placed the Briançonnais peninsula, by intra-oceanic subduction of the Piemont/Penninic ocean, beside the distal Helvetic margin. This created a repetition of the European margin in the western Alps domain (Figs 8 & 9) (Ringgenberg *et al.* 2002). The space between these two similar margin segments is generally referred to as the Valais 'ocean', a domain that formerly belonged to the Piemont (Alpine Tethys) ocean, trapped by the eastward displacement of the Iberian–Briançonnais block during the opening of the North Atlantic. The Middle Jurassic age (161 Ma) of potential parts of the Valais (Piemont) sea floor from the Misox zone was recently confirmed (Liati *et al.* 2003). As in the Ligurian and Piemont domains, where mantle rocks associated to Alpine ophiolites have shown Permian and even Precambrian ages (Rampone & Piccardo 2000), the Versoyen magmatic complex in the internal Valais domain also has Permian ages, but was clearly reheated during the Cretaceous (110–100 Ma). But it is not yet clear if this complex moved with the Briançonnais, or if it belonged to the toe of the Helvetic margin (Fügenschuh *et al.* 1999). The presence of the Cretaceous thermal event there would place them as part of the northern Briançonnais margin. Other gabbros in the Monte Rosa nappe have been dated as Cenomanian (Liati & Froitzheim 2006), but as discussed above, we would rather see them as pertaining to the Penninic ocean.

## *Along strike shortening*

A relatively large remnant of the Vardar ocean subducted northeastward under Moesia during the Late Cretaceous, as evidenced by the large Srednogorie volcanic arc of the Balkans and the Late Cretaceous opening of the Black Sea (Nikishin *et al.* 2003), representing the third generation of back-arc opening in that region. This subduction zone and its extension eastward up to Iran, represents a major slab pull force moving Africa in that direction until the closure of the Vardar in the Late Maastrichtian–Palaeocene. This is clearly confirmed by the change of convergence direction and velocity decrease of the African plate at that time (Fig. 2), also related to the synchronous peri-Arabian ophiolitic obduction (Pillevuit *et al.* 1997) and slab detachment.

This NE directed subduction of the African plate in the Cretaceous, brought Africa–Iberia far to the east, inducing a relative westward escape of the AA wedge into the Piemont oceanic corridor. The southern margin of the Piemont ocean (Southern Alps–AA domain of the western Alps) was affected by tectonic movements since the Coniacian–Santonian, as evidenced by the onset of flysch deposition in the Piemont (Gets and Dranse flysch, Caron *et al.* 1989; Ligurian, Argnani *et al.* 2006 and Lombardian basins, Bernoulli & Winkler 1990). We relate such tectonism to large-scale sinistral strike-slip movements that affected the boundary between Adria and the Alpine domain.

These very large-scale lateral movements are well known also in the eastern Alps (Trümpy 1988, 1992) and finally placed Adria–Tizia south of the AA accretionary prism. In this process, the Piemont oceanic lithosphere was progressively detached from the northern margin of Adria and subducted beneath it. Frontal pieces of the western Adria–AA margin were dragged into the subduction zone, as evidenced by the Late Cretaceous–Early Palaeocene HP–LT metamorphism recorded in the Sesia domain (Oberhänsli *et al.* 1985; Rubatto 1998).

## *The Pyrenean cycle*

The 'Valais trough', as recognized in the western Alps today, is actually the remnant of trapped Piemont sea floor and of the toe of the Helvetic margin (see above, Stampfli *et al.* 2002). The 'Valais ocean' (as defined in Stampfli 1993) was located south of France and we refer to it here as the 'Pyrenean ocean' to avoid confusion between Valais ocean and Valais trough. No direct traces of this ocean have been found so far because its suture was located exactly where the Algero–Provençal ocean re-opened in Oligo–Miocene times (Roca 2001). A large part of the southern margin of the Pyrenean ocean was the Briançonnais peninsula (Figs 5 & 6); its northern margin was the Corbières–Provençal domain from the Pyrenees to the Maures-Estérel massifs. Elements of the Pyrenean margin of the Briançonnais are found in the Galibier region of the French Alps (Toury 1985), well known for its Late Jurassic Brèche du Télégraphe. Recent investigation there (Luzieux & Ferrari 2002) showed that a pull-apart type basin rapidly deepened under the continental crust deformation (CCD) in Late Jurassic times. This area is regarded as the most external Briançonnais element known so far, unless the Versoyen complex is also seen as a part of this margin (see above). Its conjugate northern Provençal margin area is characterized by important erosion during the Oxfordian (rift shoulder uplift) and the development of Albian basins deepening southward towards the ocean, followed by the accumulation of thousands of metres of upper Cretaceous clastics in a northward migrating fore-deep type basin (Debrand-Passard & Courbouleix 1984). This records a southward closure of the Pyrenean 'ocean' on a Provençal transect, whereas a northward subduction is usually proposed on a Pyrenean transect (Vergés & Garcia-Senz 2001), also seen in Sardinia (Barca & Costamagna 1997, 2000), but again a southward subduction took place westward in the Biscay ocean (Olivet 1996). We propose here a uniform southward subduction during the late Cretaceous Pyrenean phase (Figs 8 & 9), replaced in the Pyrenees by northward subduction of the Iberian indenter during the second Eocene Pyrenean phase (Fig. 10). This dominating last event gave to the Pyrenees its final lithospheric structure, as seen on the ECORS transect. In between the two phases, and according to the Atlantic magnetic anomalies, the Palaeocene position of Iberia seems to retreat slightly (10–20 km) from Europe. This has created some large-scale extension superimposed on the Late Cretaceous orogenic

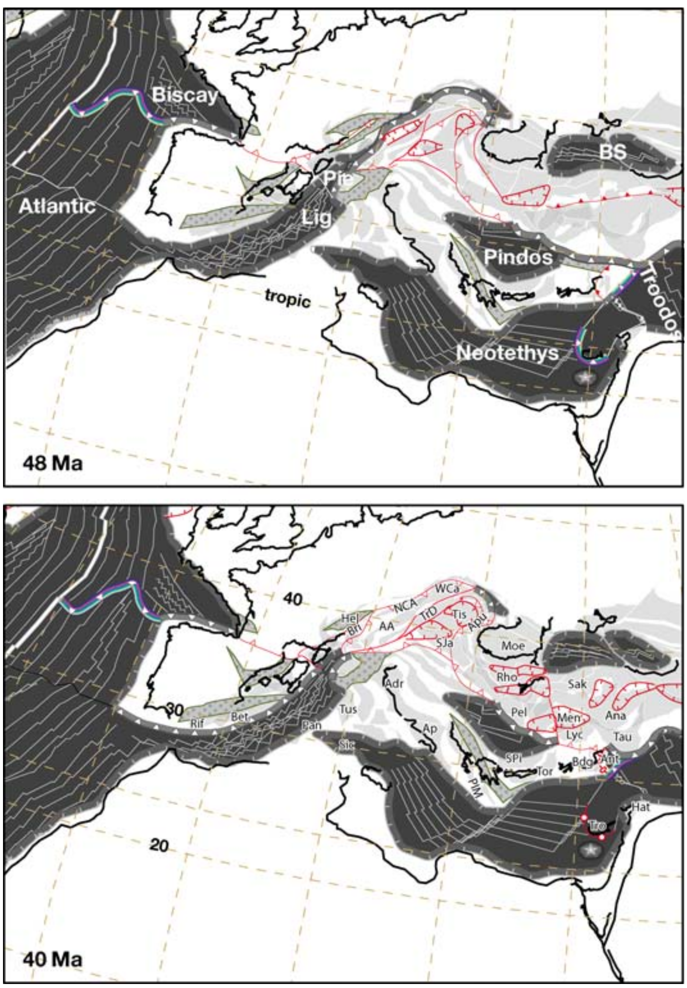


**Fig. 10.** 48–40 Ma. *Lutetian-Bartonian*. The Briançonnais terrane was entering into collision with the AA. The latter has closed a large part of the Alpine Tethys. Behind the prism, extension generated large rift zones, such as on the Tisia terrane (Pannonian basin). Similarly, the subduction of the Pindos and Troodos oceanic slabs, triggered the opening of numerous rifts and associated core-complexes in Turkey, in the Cyclades and in the Balkans. In the Pyrenees, retro-wedge thrusting is now very active; collision of the Iberian and European domains triggered the subduction of the remnant Alpine Tethys south of Iberia.

wedge and is responsible for the deposition of the Palaeocene red beds found all along the Pyrenean chain (Bilotte & Canérot 2006).

The Pyrenean orogen can be followed from the present Pyrenees eastward to southern France (Provence), and continues in the Alps in the form of a large-scale uplift of the Helvetic margin and local inversion of the Jurassic tilted blocks, well expressed by the deposits of the Niesen Flysch (Ackermann 1986) (mainly Maastrichtian) and Meilleret Flysch (Middle Eocene) (Homewood 1974), sedimented on a structured Mesozoic basement. Recent investigations have shown that similar turbiditic deposits of Late Cretaceous age are also found in ultra-Dauphinois units of the French Alps, such as the Pelat units (De Paoli & Thum 2008) and the already known Quermoz unit (Homewood *et al.* 1984). These so-called flysches clearly predate the Alpine syn-collisional event in the Helvetic domain, characterized by the deposition of the classical Alpine flysch sequences not before the earliest Oligocene (Kempf & Pfiffner 2004). These Cretaceous turbiditic deposits clearly point to an orogenic event affecting the distal Dauphinois–Helvetic margin and were deposited in the eastern prolongation of the Provençal foreland basin. On a Provence transect, the closure of the remnant Pyrenean basin certainly took place at the end of the Cretaceous phase (Figs 8 & 9), whereas in the western Alps, this closure was not total, as witnessed by the continuing deposit of coarse clastic turbidites in the Valais trough up to the Late Eocene (Bagnoud *et al.* 1998). This suggests the presence of a subduction/inversion zone between the Valais basin and the Niesen–Quermoz–Pelat foreland basin, both basins being separated by a relief where basement was outcropping (the so-called Tarine cordillera).

Eastward, the uplift of the Bohemian massif (e.g. Tanner *et al.* 1998) was certainly related to the same late Cretaceous inversion event and triggered the deposition of the Rheno–Danubian turbiditic sequence. Along the European margin it is interesting to note the similarity of facies between these Rheno–Danubian turbiditic deposits and the Valais trough sequence (the Valais trilogy) from Albian to Late Cretaceous (Stampfli 1993, and references therein). These deep-water clastic facies were located along the toe of the European margin, often referred to as the North Penninic basin (Figs 7 & 8). The presence of contourites and strong and changing current directions along the basin (Hesse 1974) suggests a connection-with major oceanic domains. So the Valais trough, together with the Pyrenean–Biscay ocean, must be regarded as connections between the Eastern Alpine Tethyan realm and the north Atlantic ocean during the Cretaceous.

## *The Alpine cycle*

After several phases of rifting, as described above, and a new thermal subsidence stage developing during the Cretaceous, the Alpine region entered a phase of convergence between the African and European plate (Fig. 11). The tectonic evolution of the western Alps started with the formation of an accretionary prism related to the closure of the Alpine Tethys where different geological objects, corresponding to different stages of accretion, can be recognized:

- the Adriatic back-stop, comprising an aborted Jurassic rifted basin (Lombardian basin);
- the oceanic accretionary prism of the Piemont/Penninic ocean (the western Alps portion of the Alpine Tethys), including crustal elements from the former toe of the southern passive margin (lower AA elements);
- accreted material of the Briançonnais terrane derived from the Iberian plate; and
- accreted material of the Valais trough, representing the toe of the European passive margin; and
- accreted material of the former European continental margin and rim basin (Dauphinois–Helvetic domain).

In time, one passes from the oceanic accretionary prism to the formation of the orogenic wedge (Escher *et al.* 1996; Pfiffner *et al.* 1997; Ford *et al.* 2006) that we place after the detachment or delamination of the subducting slab in the Early Oligocene (e.g. Stampfli & Marchant 1995; von Blanckenburg & Davies 1995). The resulting heat flux allowed some more units to be detached from the European continental slab, triggering large-scale subduction of continental material (Marchant & Stampfli 1997; Ford *et al.* 2006) and Oligocene to Pliocene overthrusting of the most external units, such as:

- the external Variscan massifs and their cover;
- the Subalpine fold and thrust belts;
- the North Alpine foreland basin (molassic basin); and
- the Jura mountains fold belt.

These late events were accompanied by retro-wedge thrusting of the southern Alps on the Po Plain, a flexural basin located above the former Jurassic Lombardian basin. The retro-wedge becomes wider eastward, forming the Southern Alps thrust belt (Schönlaub & Histon 2000; Schmid *et al.* 2004). This younger S-verging thrust system linked two north-dipping subduction systems: the Adriatic–Dinaride continental subduction to the east, and the subduction of the remnant Alpine Tethys under the Iberian plate in the west (Figs 10 & 12). In the Adriatic–Dinaride subduction zone, a large part of the northern Apulian promontory was subducted under the AA domain (Schmid *et al.* 2004; Kissling

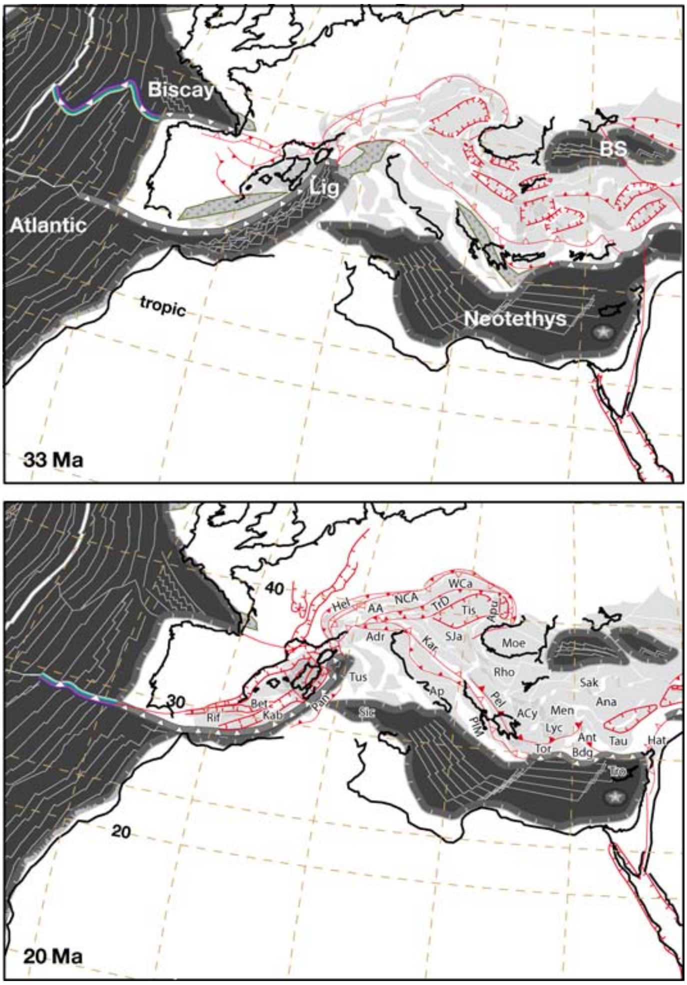


**Fig. 11.** 33–20 Ma. *Rupelian; Burdigalian.* The Alps have entered the main orogenic phase. The Briançonnais was subducted; parts of its cover was detached and accreted to the orogenic prism. Slab roll-back in the Carpathians allowed the Pannonian basin to enlarge and many other rifts and core-complexes were still active in the Balkans, Greece and Turkey until the Pindos was totally closed. Then a collision of the Hellenic orogenic prism took place with the Greater Apulia block (PIM) and triggered the subduction of the East Mediterranean basin (eastern part). Nowadays, this subduction front has reached the Apulian block (Fig. 1). On the Iberian side, the collapse of the active margin onto the remnant Alpine Tethys segment has induced large-scale rifting in the upper plate and the detachment of the Corso–Sardinian block. This rift system link northward with the Bresse–Rhine graben system. A nearly continuous south-vergent subduction/collision front runs from the southern border of Iberia to Turkey, and the north-vergent Alpine prism will progressively become less active.

*et al.* 2006), whereas the northwestern corner of Adria developed as an indenter (Handy *et al.* 2005).

Since the Miocene, the Alps and Carpathian wedges moved mainly due to remnant roll-back of the European plate. The weakly buoyant lithosphere under the rim basins present in that margin allowed further subduction to take place. However, in the western Alps, the Adriatic indenter was pushed against the orogenic prism as it was attached to the still converging African plate, creating maximum shortening in the western Alps.

As we have seen above, the southward subduction of the Alpine Tethys ocean is related to the history of the Meliata–Maliac and Vardar domain, and was inherited in the western Alps from the pre-existing northward vergence of the AA accretionary wedge. This northward vergence is unique in the whole Alpine and Tethyan domain, where most orogens are south vergent.

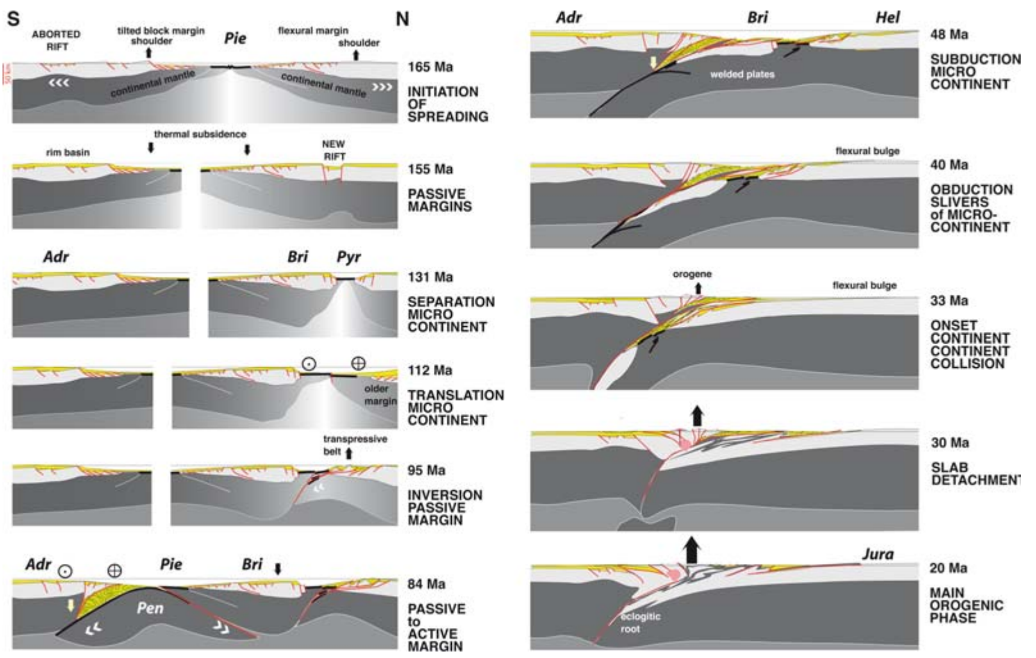


**Fig. 12.** North–south palinspatic cross-sections through the western Alpine segment. The Briançonnais block (Bri, on the reconstruction) is fixed; the other segments move in and out of the picture. Arrows show major uplift and subsidence events.

The change in vergence of Alpine Tethys subduction is found at the connecting region between the Alps and the northern Apennines, south of Corsica. It must be emphasized here that the Penninic Alpine accretionary prism is older (Late Cretaceous–Eocene) than the Apenninic one (Oligocene–Pliocene); actually one started when the other one stopped. During this process, the Apenninic prism re-mobilized parts of the Piemont–Penninic prism as exotic elements (e.g. the Bracco ophiolitic ridge, Elter *et al.* 1966; Hoogenduijn Strating 1991). The Alpine prism collided with the Iberian plate in Corsica (Malavieille *et al.* 1998) in Eocene times (Figs 10–12). The remnant Alpine Tethys oceanic domain (Ligurian basin) south of the Iberian plate started to subduct northward in the Late Eocene (producing HP metamorphism dated at 25–21 Ma, Michard *et al.* 2006) in an ongoing process of shortening between Europe and Africa and following continent-continent collision in the Pyrenees.

On the Miocene reconstruction (Fig. 12), nearly all of this remnant ocean is already subducted. This SE-directed roll-back of the Tethyan slab triggered the opening of the Algero–Provençal back-arc basin starting in the Oligocene (Fig. 12). Thus, the Oligocene volcanism of Sardinia can be regarded as subduction related (e.g. Monaghan 2001). The Alpine accretionary prism stranded in the eastern border of Corsica started to collapse backward (eastward), following the build-up of the new Apenninic accretionary prism; these Piemont–Ligurian units are now found in an upper structural position, whereas they were in a lower one during the Alpine collision. The northern and southern Apenninic prisms collapsed into pre-existing depressions, the Lombardian rift in the north and the Ionian sea oceanic corridor (the western most part of the East Mediterranean Neotethys ocean) in the south, with its still active back-arc opening (Savelli 2000; Argnani & Savelli 2001). In the eastern part of the East Mediterranean basin, subduction had already started in the late Eocene (Fig. 12); it then prograded westward until its present day position in Greece (Fig. 1).

## Conclusions

The proposed reconstructions require a tight fit of the Iberian–Alboran microplates with respect to North America and Europe, and also imply a long lasting rifting phase in the Atlantic regions now separating these domains. The final position of the Apulia–Adria plate and its geometry close to the present-day one, was reached through Jurassic rifting events, which finally gave birth to the narrow Alpine Tethys. The latter is dominated by stretching of the lithospheric mantle of its borders. On its western side, the Alpine area is largely influenced by the opening Atlantic ocean, and the northward extension of the latter in Early Cretaceous times. On its eastern side, the Alpine domain is dominated by the evolution of the supra-subduction Vardar ocean. The latter collided with its border in latest Jurassic–early Cretaceous times. Its northern border consisted of the AA micro-continent, not buoyant enough to resist subduction. Large portions of its crust and cover were underplated and transported northward onto the eastern Alpine Tethys (Penninic–Vahic ocean).

It is then clear that two specific geodynamic scenarios presided at the evolution of the Alpine belt. The western branch was affected by a second phase of rifting, detaching the Iberian plate and its Briançonnais promontory from Europe, finally duplicating the southern European margin in the western Alps. The eastern branch is dominated by an advancing large accretionary prism that included the AA blocks. The prism was collapsing forward but also westward into the oceanic Piemont part of the western branch. This was made possible by large-scale east–west shortening, related to the kinematics of the African plate with regards to Europe (Fig. 2). Thus, the Vardarian oceanic back-stop of the AA prism was replaced in time by the continental larger Adriatic back-stop. The Piemont oceanic space was, at the same time, forced to subduct southward in front of the advancing AA prism. At the close of the Cretaceous, all the major elements are in place, and after the Palaeocene lull, following the change of wander path of Africa, they were thrust over each other to form the Alpine orogenic wedge.

If this wedge is tectonically relatively cylindrical, its components are not closely related. The Alpine orogen is a clear example of juxtaposition of far travelled tectonic elements and cannot be understood from a 'fixist' point of view.

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