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Geodynamic evolution of a wide plate boundary in the Western Mediterranean, near-field *versus* far-field interactions

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Abstract – The present-day tectonic setting of the Western Mediterranean region, from the Pyrénées to the Betics and from the Alps to the Atlas, results from a complex 3-D geodynamic evolution involving the interactions between the Africa, Eurasia and Iberia plates and asthenospheric mantle dynamics underneath. In this paper, we review the main tectonic events recorded in this region since the Early Cretaceous and discuss the respective effects of far-field and near-field contributions, in order to unravel the origin of forces controlling crustal deformation. The respective contributions of mantle-scale, plate-scale and local processes in the succession of tectonic stages are discussed. Three periods can be distinguished: (1) the first period (*Tethyan Tectonics*), from 110 to 35 Ma, spans the main evolution of the Pyrenean orogen and the early evolution of the Betics, from rifting to maximum shortening. The rifting between Iberia and Europe and the subsequent progressive formation of new compressional plate boundaries in the Pyrénées and the Betics, as well as the compression recorded all the way to the North Sea, are placed in the large-scale framework of the African and Eurasian plates carried by large-scale mantle convection; (2) the second period (*Mediterranean Tectonics*), from 32 to 8 Ma, corresponds to a first-order change in subduction dynamics. It is most typically Mediterranean with a dominant contribution of slab retreat and associated mantle flow in crustal deformation. Mountain building and back-arc basin opening are controlled by retreating and tearing slabs and associated mantle flow at depth. The 3-D interactions between the different pieces of retreating slabs are complex and the crust accommodates the mantle flow underneath in various ways, including the formation of metamorphic core complexes and transfer fault zones; (3) the third period (*Late-Mediterranean Tectonics*) runs from 8 Ma to the Present. It corresponds to a new drastic change in the tectonic regime characterized by the resumption of N-S compression along the southern plate boundary and a propagation of compression toward the north. The respective effects of stress transmission through the lithospheric stress-guide and lithosphere-asthenosphere interactions are discussed throughout this period.

Keywords: mantle flow / convection / collision / subduction / lithosphere / Mediterranean / back-arc extension / slab retreat / slab tearing / Pyrénées / Betic Cordillera / Rif / Liguro-Provençal Basin / Alboran Sea / Tyrrhenian Sea

Résumé – Évolution géodynamique d'une limite de plaque diffuse en Méditerranée occidentale, interactions à courte et longue distance. Le contexte tectonique actuel de la Méditerranée occidentale, des Pyrénées aux Bétiques et des Alpes à l'Atlas, résulte d'une évolution géodynamique complexe en 3-D impliquant les interactions entre les plaques Afrique, Eurasie et Ibérie et le manteau asthénosphérique sous-jacent. Dans cet article, nous présentons une revue critique des principaux événements tectoniques survenus

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dans la région depuis le Crétacé supérieur pour discuter les contributions respectives des processus locaux et lointains, dans le but de mieux comprendre les forces contrôlant la déformation crustale. Les contributions respectives des processus à l'échelle du manteau, à l'échelle des plaques et les processus régionaux sont discutées. Trois périodes doivent être distinguées : (1) la première période (Tectonique Téthysienne), de 110 à 35 Ma, recouvre toute l'évolution de l'orogène pyrénéen et l'évolution précoce des Bétiques et du Rif, depuis le rifting jusqu'au raccourcissement maximum. Le rifting entre Ibérie et Europe et la formation de nouvelles limites de plaques convergentes et compressives dans les Pyrénées et les Bétiques et la propagation de contraintes compressives jusqu'à la Mer du nord sont placés dans le cadre des plaques Eurasie et Afrique portées par la convection mantellique à grande échelle ; (2) la deuxième période (Tectonique Méditerranéenne), de 32 à 8 Ma, est plus typiquement méditerranéenne avec une contribution dominante du recul des panneaux lithosphériques plongeants et des flux mantelliques associés. Construction des chaînes de montagnes et ouverture des bassins arrière-arc sont dominés par cette dynamique de retrait et de déchirure des panneaux plongeants et par le flux asthénosphérique qui en résulte. Les interactions 3-D entre les différents morceaux des panneaux plongeants sont complexes et la croÛte accommode les flux mantelliques de façons variées, incluant la formation de metamorphic core complexes et de zones de transfert ; (3) la troisième période (Tectonique Tardi-Méditerranéenne) débute à ~8 Ma et se poursuit jusqu'à aujourd'hui. Elle correspond à un nouveau changement drastique de la dynamique de subduction associé à une reprise de la compression N-S au travers de la limite de plaque méridionale et une propagation de la compression vers le nord et l'est. Les effets respectifs de la transmission des contraintes le long du guide de contraintes qu'est la lithosphère et des interactions lithosphère-asthénosphère sont discutés pour chacune de ces trois périodes.

Mots clés : flux mantellique / convection / collision / subduction / lithosphère / Méditerranée / extension arrière-arc / retrait du panneau plongeant / déchirure du panneau plongeant / Pyrenées / Cordillère Bétique / Rif / Bassin Liguro-Provençal / Mer d'Alboran / Mer Tyrrhénienne

1 Introduction: different scales and different periods

The Mediterranean region is the heir of a former wider ocean, the Neo-Tethys. The Mediterranean mountain belts and post-Eocene back-arc basins result *in fine* from the convergence of Africa and Eurasia (Dercourt *et al.*, 1986; Ricou *et al.*, 1986; van Hinsbergen *et al.*, 2019). This region has been studied in great detail since more than a century and its tectonic history has been reconstructed since the Triassic (Dercourt *et al.*, 1986, 1993; Ziegler, 1999; van Hinsbergen *et al.*, 2019; Angrand *et al.*, 2020). Subduction dynamics changed around 35–30 Ma in this region, from a typical Tethyan pattern with the formation of mountain belts along the south margin of Eurasia, to a characteristic Mediterranean dynamics with fast slab retreat and formation of tight arcs (Jolivet and Faccenna, 2000). Tight arcs and fast-opening back-arc basins, coeval with the continuing formation of mountain belts near the subduction trench, is indeed a characteristic of the Mediterranean (Malinverno and Ryan, 1986; Royden, 1993; Jolivet *et al.*, 1994; Carminati *et al.*, 1998a; Jolivet *et al.*, 1998; Jolivet and Faccenna, 2000; Wortel and Spakman, 2000; Faccenna *et al.*, 2001a; Spakman and Wortel, 2004; Faccenna *et al.*, 2013b). The Western Mediterranean shows this type of evolution with the addition of a complex 3-D configuration where several orogenic systems (Alps, Pyrénées, Apennines, Maghrebides, Betics) interfere in time and space (Réhault *et al.*, 1984, 1987; Carminati *et al.*, 1998a, 1998b; Faccenna *et al.*, 2001a, 2001b, 2004; Rollet *et al.*, 2002; Vignaroli *et al.*, 2009). Detailed reconstructions of the post-Eocene kinematics are available in the literature (Dewey *et al.*, 1989; Rosenbaum *et al.*, 2002a, 2002b; Schettino and Turco, 2006; Mantovani *et al.*, 2009; van Hinsbergen *et al.*, 2014, 2019; Mantovani *et al.*, 2020; Romagny *et al.*, 2020) that agree on the first order displacements but differ on the driving mechanisms. The respective influence of slab retreat and extrusion are for instance debated. This 3-D complexity seen in the crust in fact reflects the complexity of slab interactions at depth. A wealth of recent studies across the Africa-Eurasia plate boundary zone, including those of the *Orogen* project, shed light on these interactions through time. We review in this paper the recent findings with a special attention to the outcome of the *Orogen Project* and discuss their implications on the various drivers, near-field and far-field, of the orogens and basins of the Western Mediterranean.

2 Geodynamic context of Western Mediterranean orogens and basins

Plate movements and deformations are to a first order controlled by the variable degree of coupling between the lithosphere and the asthenospheric mantle convecting underneath (Fig. 1). Plates attached to a subducting lithosphere are the fastest and they are mostly driven by slab pull (Uyeda and Kanamori, 1979; Ricard and Vigny, 1989; Ricard *et al.*, 1989; Conrad and Lithgow-Bertelloni, 2002; Coltice *et al.*, 2019). The role of plumes in these displacements is debated, whether they are a passive response to subduction or an active player (plume push) (Forte *et al.*, 2010; Becker and Faccenna, 2011; Moucha and Forte, 2011; Glisovic *et al.*, 2012; Faccenna *et al.*, 2013a, 2014). This classical question about the respective roles of plume and subducting slabs is possibly not correctly formulated as the sub-lithospheric mantle is an incompressible fluid on geological time scales and any displacement somewhere has to be instantaneously compensated by a displacement somewhere else. It has been proposed that mantle

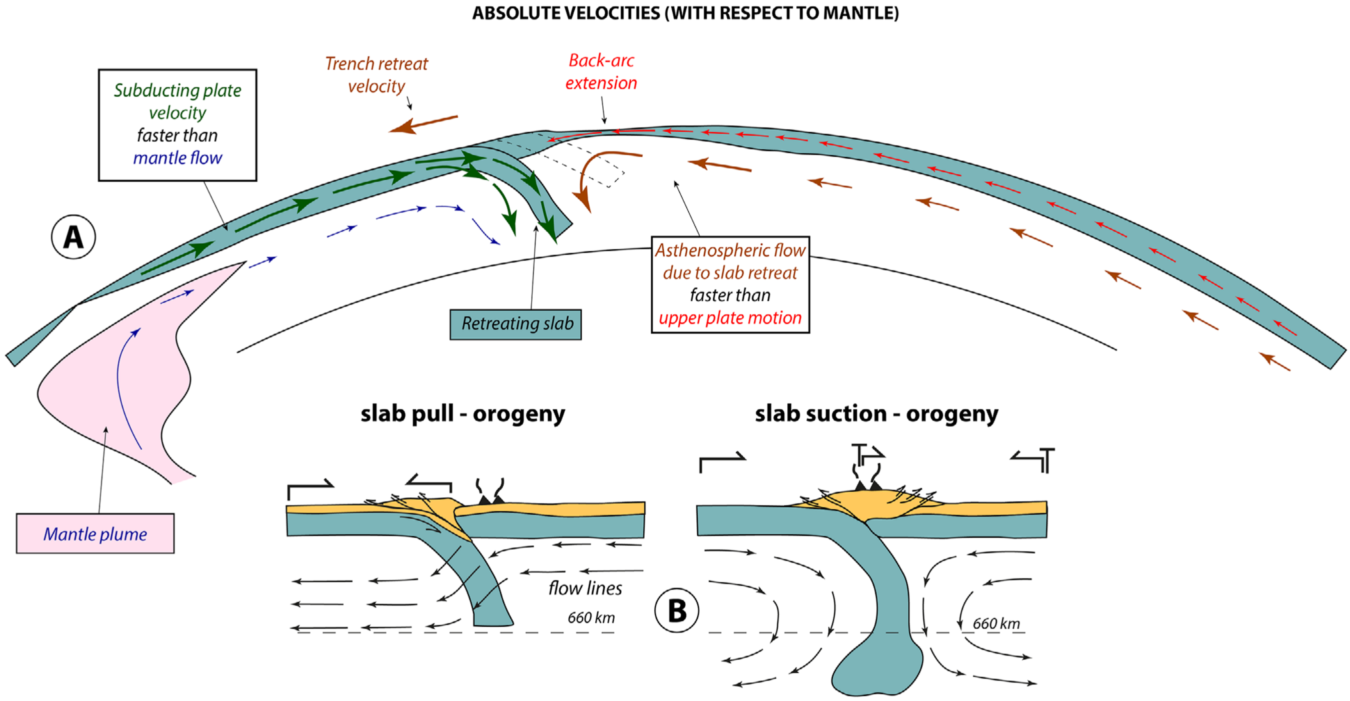

Fig. 1. Plate motion, lithospheric deformation and mantle flow. A. Differential velocities between lithosphere and asthenosphere in plate attached to slabs and overriding plates and associated deformation, after Coltice *et al.* (2019). B Slab-pull *versus* slab-suction orogeny after Faccenna *et al.* (2013a). In slab-pull orogens, the slab is retreating and flow lines in the mantle thus cross the slab, while in slab-suction orogens, the slab is anchored in the lower mantle at the contact between two convection cells.

flow underneath continents, partly due to mantle plumes, can be an active force when it pushes on irregularities of the base of the lithosphere (Stoddard and Abbott, 1996; Alvarez, 2010; Koptev *et al.*, 2015, 2017, 2019). When, instead, continents are carried by plates that are not attached to a subducting lithosphere, their deformation is, on the opposite, largely driven by the mantle flowing underneath and this is particularly the case on the edges of continents above retreating subduction zones (Coltice *et al.*, 2019).

Orogeny is moreover controlled by two types of forces, body-forces and boundary-forces (Molnar and Lyon-Caen, 1988; England and Houseman, 1989). Boundary-forces arise from lithospheric plate interactions, divergence, convergence or strike-slip. Stresses are transmitted through the lithospheric stress-guide over large distances (Elsasser, 1968) but plate motions are also the visible part of mantle convection (McKenzie, 1969). Continents are passively transported on the convective mantle, but they can locally modify the convective regime (Burov and Gerya, 2014; Koptev *et al.*, 2015, and references therein). The complex geometry of plate boundaries can lead to drastic changes in the stress field during convergence and explain the succession of tectonic stages with variable characteristics (Bellahsen *et al.*, 2003). Far-field events can then have direct effects on the dynamics of an orogen in this context.

Body-forces, on the other hand, arise from lateral density contrasts within the lithosphere, in which case they control the surface kinematics and thus the distribution of stresses within the plates, hence interfering with boundary forces (England and Houseman, 1989). For instance, crustal thickening due to shortening leads to an increase of the potential gravitational energy stored within the crust and favors crustal collapse or spreading (Molnar and Lyon-Caen, 1988; England and Houseman, 1989; Platt and Vissers, 1989; Vissers *et al.*, 1995). But crustal shortening is also associated with subduction of the lithospheric mantle, and the slab below the accretionary wedge has an opposite effect to crustal thickening, pulling the topography downward (Faccenna *et al.*, 2013a). Besides, lithospheric thinning due to rifting or delamination also leads to an increase of the gravitational potential energy and thus favors further extension (Bird, 1979, 1991).

The complexity of the interplay of body-forces and boundary-forces can explain the episodic evolution of orogens with, for instance, the interruption of a shortening continuum or modifications of stress regime when the lithospheric root detaches or delaminates, setting a new geodynamic frame-work. This already complex interactions can be further modulated by the variable thickness and rheology, thus the thermal age of the continental lithospheres involved in orogenic processes (Burov, 2011; Mouthereau *et al.*, 2013). Such complexity becomes even more important in 3-D when the thickened crust can escape laterally as observed in the India-Asia collision zone with the lateral eastward escape of the thick Tibetan crust (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1976; Tapponnier *et al.*, 2001; Royden *et al.*, 2008; Yin, 2010). When a continent clogs a subduction zone, it can induce a detachment or a tear of the subducting slab, which can have drastic consequences on the geometry of convection underneath and on crustal deformation (Capitanio, 2014; Sternai *et al.*, 2014; Capitanio *et al.*, 2015; Sternai *et al.*, 2016). The case of the Mediterranean is an example of this behavior. The collision of Arabia with Eurasia induced slab tearing in the Late Eocene-Early Oligocene and a change in the subduction regime with fast roll-back and the opening of back-arc basins (Le Pichon, 1982; Jolivet and Faccenna, 2000; Faccenna *et al.*, 2006; Schildgen *et al.*, 2014).

Slab retreat is a major ingredient of Mediterranean geodynamics. It occurs when the velocity of subduction (understood here as the length of lithosphere sinking into the asthenosphere per time unit) outpaces the velocity of convergence, leading to a retreat of the trench with respect to the overriding plate (Dewey, 1980; Malinverno and Ryan, 1986; Royden, 1993; Funiciello *et al.*, 2003). It generally corresponds to the passive sinking of the subducting plate in the mantle under its own weight. Slab retreat has some very close similarities with delamination in the sense of Bird (1979), as exemplified by the evolution of the Apennines (Jolivet *et al.*, 1998; Piana Agostinetti and Faccenna, 2018) or the Gibraltar Arc system (Angrand *et al.*, 2020; Daudet *et al.*, 2020).

Combining the effects of intra-lithospheric stresses and interactions between convection and lithospheric deformation, Faccenna *et al.* (2013a) have proposed two end-members for orogeny (Fig. 1). *Slab-pull orogens* are driven by a retreating slab confined in the upper mantle and they are mostly asymmetrical (Western Alps), whereas *slab-suction orogens* develop above a slab penetrating the lower mantle, associated with surge of mantle upwelling leading to more symmetrical orogens in general (Himalaya and Tibet). In this interpretation, the first order driver of orogeny is mantle convection. Whether stresses controlling the deformation of continents are transmitted through the lithospheric stress-guide or result from coupling with the convective mantle is thus a difficult problem in 3-D when continental margins and plate boundaries have complex pre-orogenic geometries, like in the Mediterranean realm. However, smaller orogens, such as the Pyrénées, where finite convergence is low (maximum 200 km) do not easily fit in this two end-member model because mantle flow will have a limited effect during shortening and tectonic inheritance of the pre-orogenic rift crustal scale structures is a major ingredient.

The view emphasizing slab retreat as the first order engine of back-arc rifting is, however, not shared by all workers. Alternative mechanisms have been proposed. Doglioni *et al.* (2002) put forward internal deformation of the overriding plate, independent of slab retreat, for the Aegean extension. Armijo *et al.* (1999) argue in favor of rigid extrusion as the main driver of the formation of the North Anatolian Fault and coeval Aegean extension. A recent synthesis by Mantovani *et al.* (2020) also argues in favor of extrusion to explain extension in the Central and Western Mediterranean. The reader is referred to these papers to read more about these alternative models and we deliberately place our discussion in the framework of the school of thought interpreting back-arc basins of the Mediterranean region as the result of slab retreat.

Within this large-scale framework, the Pyrénées (Fig. 2) are a small late Cretaceous to Cenozoic orogen, partly frozen in an early stage of its development, surrounded by oceanic domains, either pre-orogenic (the Cretaceous oceanic domain of the Bay of Biscay) (Jammes *et al.*, 2010b; Chevrot *et al.*, 2018; Lescoutre and Manatschal, 2020) or post-orogenic (the Oligo-Miocene Liguro-Provençal Basin and Algerian Basin). This orogen results from the convergence between Iberia and Europe and on a larger scale from the Africa-Eurasia convergence (Roure *et al.*, 1989; Choukroune *et al.*, 1990; Vergés *et al.*, 1995, 2002; Mouthereau *et al.*, 2014; Ford *et al.*, 2016; Chevrot *et al.*, 2018; Wehr *et al.*, 2018), and it developed together with neighboring orogens such as the Alps, the Iberian Range or the Betic-Rif chain. It is abruptly interrupted in the east where it gives place to the Gulf of Lion passive margin formed during the rifting of the Liguro-Provençal back-arc basin (Réhault *et al.*, 1984; De Voogd *et al.*, 1991; Gorini *et al.*, 1993, 1994; Pascal *et al.*, 1993; Mauffret *et al.*, 1995; Gueguen *et al.*, 1998, Guennoc *et al.*, 2000; van Hinsbergen *et al.*, 2014; Jolivet *et al.*, 2020). The tectonic history of the Pyrénées and, at a larger scale, the Western Mediterranean orogens and basins, from pre-orogenic rifting to post-orogenic extension, thus results from complex 3-D interactions between Africa, Iberia, Europe and the asthenospheric mantle convecting underneath and is thus key to better understand the interplay between forces created by relative plate motion and those due to mantle convection. This paper is a review of these interactions at different scales (near-field *vs* far-field) through time and of the different models proposed so far, in order to extract the main controlling parameters for each period and the triggers of the main changes, inception of rifting in the early Cretaceous, convergence in the late Cretaceous, slab retreat at the end of Eocene and renewed compression in the late Miocene.

2.1 Present-day kinematics

The Western Mediterranean basins belong to a series of back-arc regions formed since the Eocene-Oligocene transition in the Mediterranean region, from the Aegean Sea in the east to the Alboran Sea in the west (Malinverno and Ryan, 1986; Jolivet and Faccenna, 2000). The physiography of the Mediterranean region is essentially shaped by this episode of back-arc extension and dispersion of former mountain belts now forming archipelagos, such as the Cyclades or the islands offshore Tuscany (Jolivet *et al.*, 1998, 2008; Ring *et al.*, 2010; Faccenna *et al.*, 2014). While this dynamics is still active in the Eastern Mediterranean region with extension in the Corinth or Western Anatolian rifts (Armijo *et al.*, 1999; Ford *et al.*, 2007; Rohais *et al.*, 2007; Aktug *et al.*, 2009; Taylor *et al.*, 2011; Ford *et al.*, 2017; Gawthorpe *et al.*, 2018), the Western Mediterranean is for the time being mainly under compression (Deverchère *et al.*, 2003; Billi *et al.*, 2011; Faccenna *et al.*, 2014; Medaouri *et al.*, 2014; d'Acremont *et al.*, 2020). The transition from back-arc extension to N-S compression started some 8 Ma ago and progressively invaded the whole Western and Central Mediterranean, including the South Tyrrhenian Sea, probably very recently (Jolivet *et al.*, 2006; Meijninger and Vissers, 2006; Jolivet *et al.*, 2008; Billi *et al.*, 2011; Augier *et al.*, 2013; Do Couto *et al.*, 2014, 2016; Zitellini *et al.*, 2019; d'Acremont *et al.*, 2020). Whether the Gibraltar and Calabria subductions are still active or not is a debated question (Gutscher *et al.*, 2002; 2017), the stress regime has, however, certainly changed to mostly compressional since the Late Miocene. This can be illustrated by the distribution of focal mechanisms of earthquakes (Fig. 3) showing mostly reverse fault-type earthquakes along the northern coast of Africa and north of Sicily (Meghraoui *et al.*, 1996;

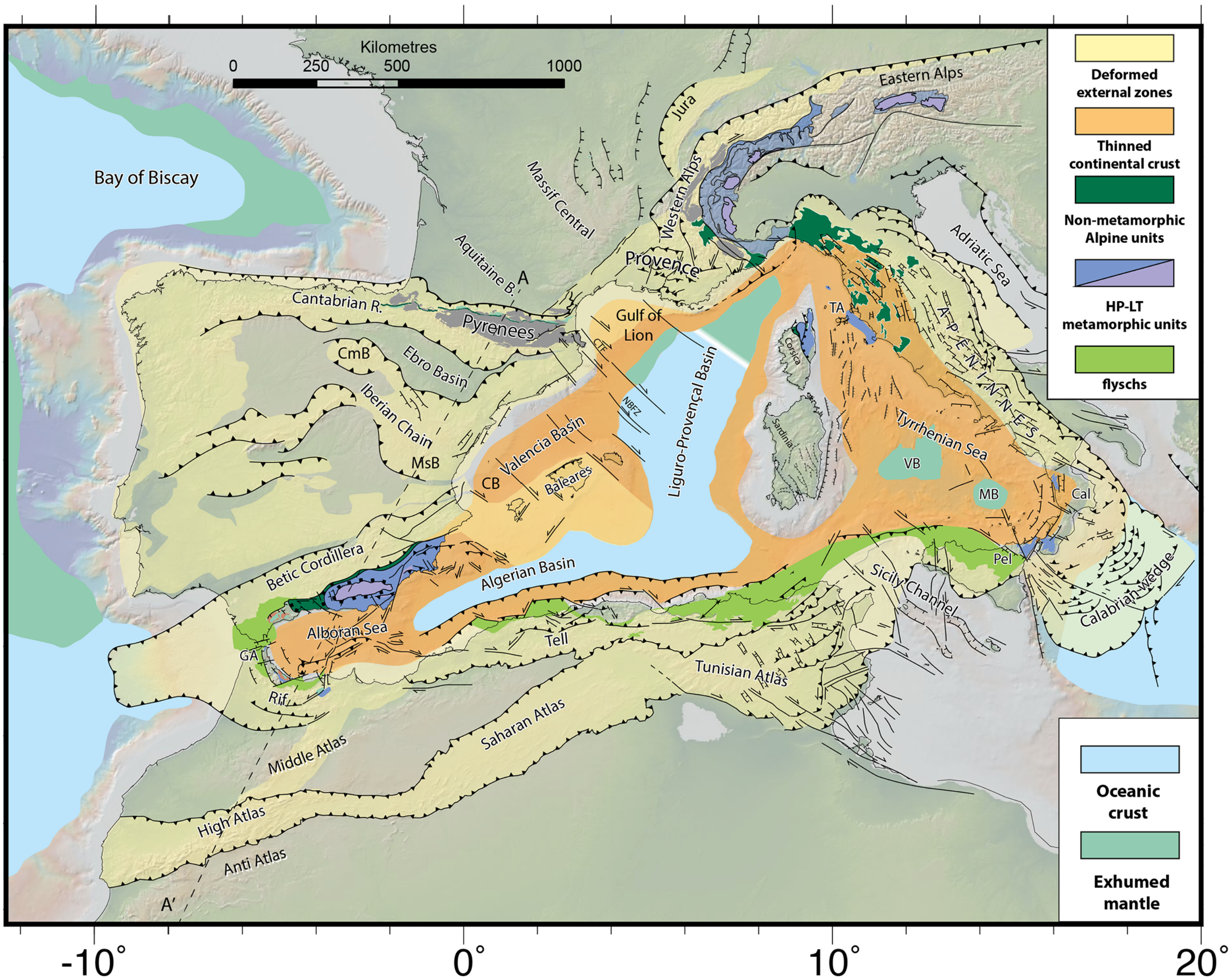

Fig. 2. Tectonic context of the Western Mediterranean region. AA': lithospheric-scale section shown on Figure 22; CB: Columbrets Basin; CmB: Cameros Basin; MsB: Maestrat Basin.

Deverchère *et al.*, 2003; Medaouri *et al.*, 2014). Extension, however, is still active within the Apennines, especially in the regions of highest elevation (Lavecchia, 1988; D'Agostino *et al.*, 1998; Collettini and Barchi, 2002; Ghisetti and Vezzani, 2002; Collettini and Barchi, 2004) and also in the internal zones of the internal part of the Southern French-Italian Alps (not show on Fig. 3 because magnitudes are too small, only symbolized by a double-headed arrow) (Sue *et al.*, 1999; Calais *et al.*, 2002; Delacou *et al.*, 2004; Walpersdorf *et al.*, 2018). The present-day stress pattern is quite simple, as shown by several types of indicators (focal mechanisms, *in-situ* measurements, boreholes) in the World Stress Map (Zoback, 1992; Heidbach *et al.*, 2018) with σ_{hmax} trending NNE-SSW in most regions whatever the local tectonic regime, extensional, compressional or strike-slip (Fig. 4). At the scale of the Mediterranean, present-day displacements measured with space geodesy (Fig. 5) are clearly dominated by the fast counterclockwise rotation of Anatolia and the faster displacement of the Hellenic Arc (McClusky *et al.*, 2000; Reilinger *et al.*, 2006; Serpelloni *et al.*, 2007; Reilinger *et al.*, 2010; Faccenna *et al.*, 2014). When zooming on the Western Mediterranean (Fig. 5), two regions are moving significantly with respect to Eurasia, the backbone of the Italian peninsula and Sicily on the one hand, moving northward or north-northeastward, and the Gibraltar Arc and part of northwest Africa, moving westward, on the other hand (Serpelloni *et al.*, 2007; Pérouse *et al.*, 2012). This recent kinematic regime was established progressively from the Late Miocene.

2.2 Mediterranean orogens and back-arc basins

The Pyrénées essentially formed during the late Cretaceous and Eocene and were partly dismantled in their eastern part after the Eocene (Choukroune, 1989; Choukroune *et al.*, 1990; Mouthereau *et al.*, 2014; Teixell *et al.*, 2018; Espurt *et al.*, 2019; Jolivet *et al.*, 2020). They are exemplary of these complex interactions of boundary forces and body forces, with a transition from post-rift to contraction at ~84 Ma, a direct consequence of the relative motion of Africa and Eurasia (Fig. 6). Some forces are locally originated (near-field) or find

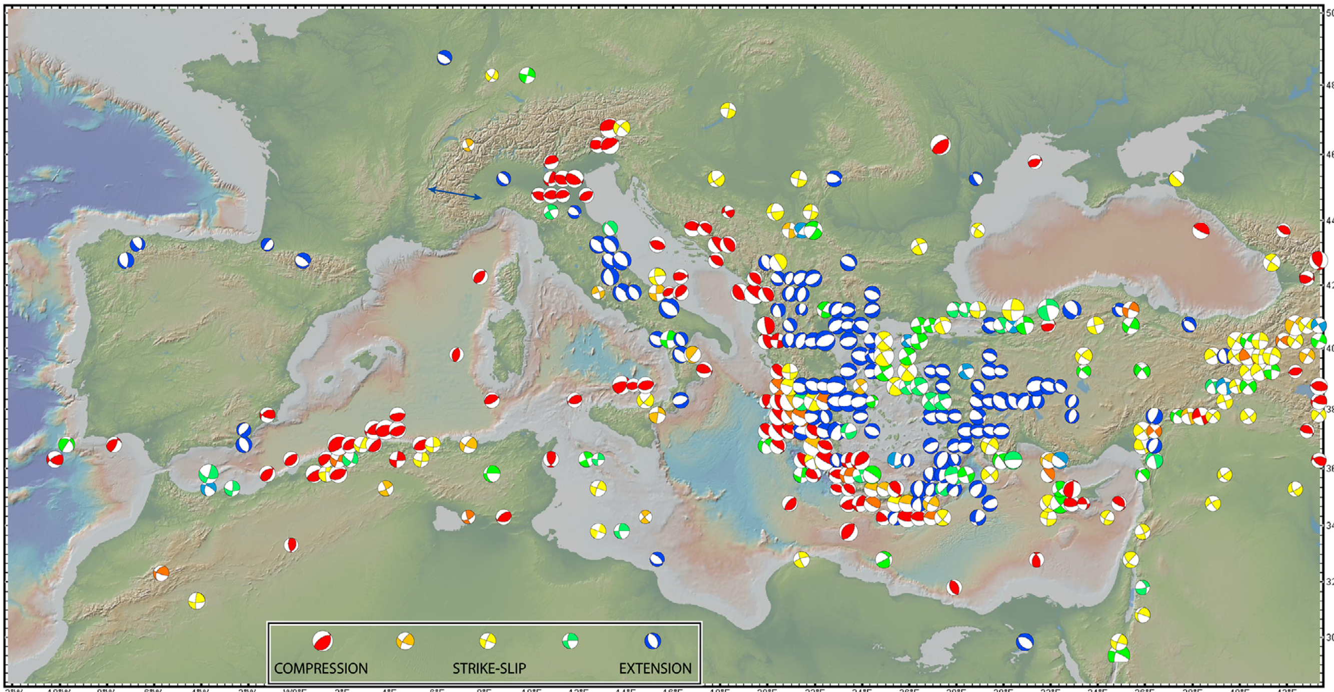

Fig. 3. Focal mechanisms of earthquakes in the Mediterranean region after Faccenna *et al.* (2014) showing the compressional context of the Western Mediterranean. The blue double-headed arrow shows the direction of extension recorded by GPS measurements in the southern part of the Alps, after Walpersdorf *et al.* (2018).

their source in distant processes (far-field). These various forces can be described as a function of scale (Fig. 7).

The Mediterranean region is a puzzle of basins and orogens (Figs. 6 and 8) sharing a common timing. The Eastern Mediterranean oceanic crust is the last remnant of the former Tethys Ocean (Dercourt *et al.*, 1986; Frizon de Lamotte *et al.*, 2011; van Hinsbergen *et al.*, 2019) (Fig. 9). Its age is debated, with large variations from the Late Cretaceous to the Paleozoic (Dercourt *et al.*, 1986; Stampfli and Borel, 2002; Frizon de Lamotte *et al.*, 2011; Granot, 2016; van Hinsbergen *et al.*, 2019; Tugend *et al.*, 2019). We consider in this paper an early to Middle Jurassic age for the Ionian basin after Late Triassic to early Jurassic rifting based on the work of Tugend *et al.* (2019). From the Late Cretaceous, the overall convergence of Africa and Eurasia led to the formation of mountain belts above north-dipping subduction zones (Dewey *et al.*, 1989; Rosenbaum *et al.*, 2002a, Rosenbaum *et al.*, 2002b) except in the Alps where subduction was toward the south (Dercourt *et al.*, 1986; Handy *et al.*, 2010; Schmid *et al.*, 2017). In the Eastern Mediterranean, the Hellenides and Taurides formed above the subduction of the Vardar and Pindos Oceans and intervening continental blocks, the so-called Pelagonian and Apulian paleogeographic domains (Aubouin, 1959; Aubouin *et al.*, 1962; Blake *et al.*, 1981; Bonneau and Kienast, 1982; Dercourt *et al.*, 1986; Ricou *et al.*, 1986). The Alps and Carpathians formed above the south-dipping subduction of the Ligurian and Valais oceanic domains below Apulia (Schmid *et al.*, 1997; Oberhänsli *et al.*, 2004; Schmid *et al.*, 2004, 2008; Handy *et al.*, 2010; Schmid *et al.*, 2017). A classical interpretation is that the Western Mediterranean orogens (Fig. 10) derive partly from the northward subduction of the Mesozoic oceanic lithosphere of the Ionian Sea underneath the so-called AlKaPeCa (from Alboran, Kabylies, Peloritani, Calabria) block (Fig. 6) and the southward subduction of a small partly oceanic domain sandwiched between Europe and the AlKaPeCa block (Bouillin *et al.*, 1986; Michard *et al.*, 2002; Chalouan *et al.*, 2008; Leprêtre *et al.*, 2018). Vergés and Fernandez (2012) however, proposed an alternative interpretation with two subduction zones with opposite dip separated with a transform fault. A solution intermediate between Michard *et al.* (2002) and Vergés and Fernandez (2012) was proposed in the recent reconstructions of Romagny *et al.* (2020). The initial position in the Western Mediterranean domain (its distance from Iberia) is, however, debated and Daudet *et al.* (2020) place it closer to Iberia than previously published models. The transition in space toward the Alps is unclear because a large part of this system is now buried below the passive margin of the Gulf of Lions and deformed by the Oligo-Miocene Liguro-Provençal back-arc basin (Fig. 2).

From 30–35 Ma, subduction dynamics abruptly changed and all subduction zones started to retreat, forming the Aegean Sea, Liguro-Provençal Basin and Alboran Sea extensional basins in the overriding plate (Figs. 6 and 10), thus dispersing the former mountain belts and forming metamorphic core complexes such as the Cyclades, the Tuscan Archipelago or the Sierra Nevada (Lonergan and White, 1997; Carminati *et al.*, 1998a; Jolivet *et al.*, 1998; Jolivet and Faccenna, 2000; Wortel and Spakman, 2000; Faccenna *et al.*, 2001a; Rosenbaum *et al.*, 2002a; Schettino and Scotese, 2002; Faccenna *et al.*, 2004; van Hinsbergen *et al.*, 2014; Jolivet *et al.*, 2016b). The detailed reconstructions produced by Romagny *et al.* (2020) in the framework of the Orogen project illustrate this evolution for the Western Mediterranean (Fig. 10). Above these retreating subduction zones, slab-pull-type orogenic wedges (Faccenna

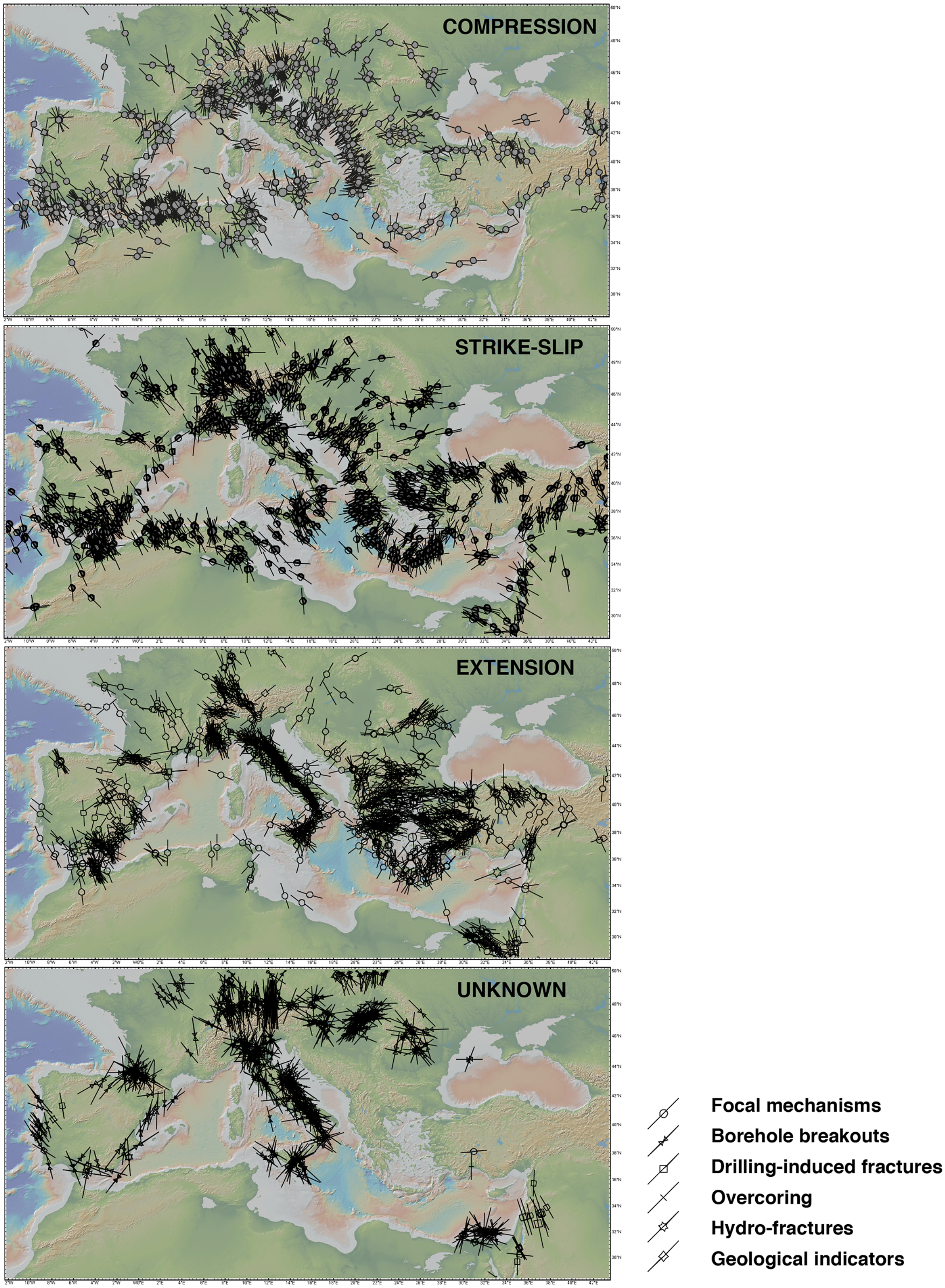

Fig. 4. Extracts of the World Stress Map for the Mediterranean region (Heidbach *et al.*, 2018) showing the NNW-SSE direction of the main horizontal compression in the Western Mediterranean region and surroundings. Whatever the deformation regime (extension, compression or strike-slip) σ_{hmax} always show the same constant trend.

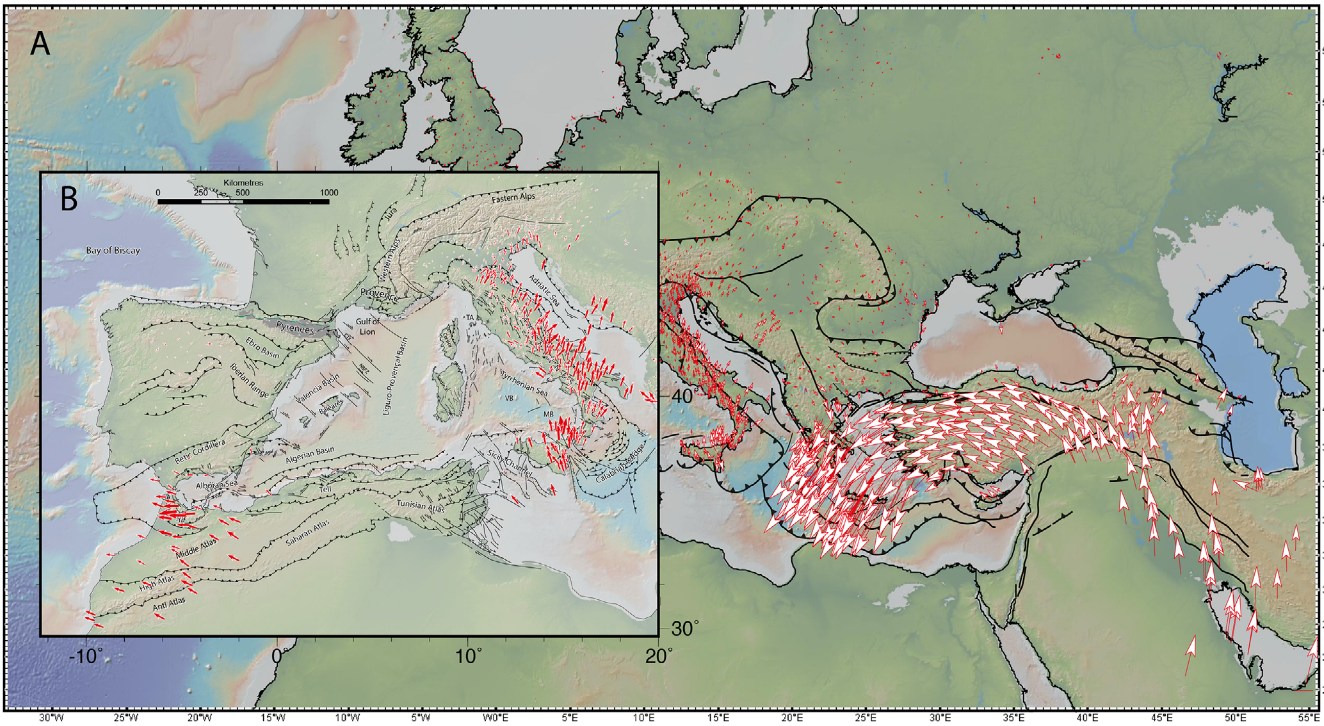

Fig. 5. Current displacements in the Mediterranean region (A) based on GPS measurements: data from Nevada Geodetic Laboratory database and from Faccenna *et al.* (2014). B. Zoom on the Western Mediterranean region.

et al., 2013a) continued to form by progressive accretion of external zones while the orogenic crust was then thinned in the back-arc domains (Jolivet *et al.*, 1994, Jolivet *et al.*, 1998; Jolivet and Brun, 2010). The Apennines, the Peloponnese and Cretan thrust faults formed in this context while the Tyrrhenian Sea and the Aegean Sea were opening (Jolivet *et al.*, 1994, Jolivet *et al.*, 1998, Jolivet *et al.*, 2003, Jolivet *et al.*, 2013). Similarly, the Betic-Rif arc was formed while the Alboran Sea was opening in the overriding plate (Platt and Vissers, 1989; Comas *et al.*, 1992; Platt, 1993; Martinez-Martinez and Azañon, 1997; Comas *et al.*, 1999; Platt *et al.*, 2003b, Platt *et al.*, 2013; Crespo-Blanc *et al.*, 2016).

This post-35 Ma evolution was associated with large displacements and rotations of crustal blocks, amounting to 800 km eastward motion for the present-day Calabrian arc, for instance, above the retreating Ionian slab and a minimum of 400 km westward for the Gibraltar arc (Dewey *et al.*, 1989; Rosenbaum *et al.*, 2002a; van Hinsbergen *et al.*, 2014; Romagny *et al.*, 2020). The nature of the retreating slab, however, is different in the two cases. The slab subducting underneath Calabria is oceanic while that underneath Gibraltar is probably mostly continental. The Oligocene and Miocene geodynamic evolution of this region is thus characterized by the concomitant formation of mountain belts above retreating slabs (slab-pull orogens) and back-arc rifting in the overriding plate, with large displacements of the subducting slab and thus fast movements in the asthenosphere underneath (Barruol and Granet, 2002; Lucente *et al.*, 2006; Buontempo *et al.*, 2008; Jolivet *et al.*, 2009; Salimbeni *et al.*, 2018).

Back-arc rifting, however, did not evolve as a continuum and salient periods must be noticed (Carminati *et al.*, 1998a; Faccenna *et al.*, 2001b, Faccenna *et al.*, 2003). After the first rifting at the turn of the Eocene and Oligocene, the basins opened widely until the AlKaPeCa block collided with the north African margin at about 20–15 Ma (Leprêtre *et al.*, 2018; Romagny *et al.*, 2020). From then on, the retreating slab was divided in two parts, one retreating eastward to form the Tyrrhenian Sea and one retreating westward to form the Algerian Basin and the Alboran Sea (Carminati *et al.*, 1998a; Wortel and Spakman, 2000; Spakman and Wortel, 2004). These changes are associated with slab tears at depth. In the Western Mediterranean, the geodynamic context changed once more around 8 Ma when north-south compression resumed in the Betics and the present-day compressional reworking of the North African passive margin was set (Billi *et al.*, 2011). Active compression is also observed along the southern margin of France in the region of Nice (Walpersdorf *et al.*, 2018).

2.3 The Iberia-Europe plate boundary in this context

In this large-scale context, the Pyrénées (Fig. 11) formed by the inversion of a system of Early Cretaceous intra-continental rifts where sub-continental mantle was exhumed by hyper-extension (Lagabrielle and Bodinier, 2008; Lagabrielle *et al.*, 2010; Jammes *et al.*, 2010b; Clerc *et al.*, 2012; Tugend *et al.*, 2014; Teixell *et al.*, 2018). High-temperature and low-pressure (H*T*-L*P*) metamorphism of the Internal Metamorphic Zone (IMZ) of the North Pyrenean Zone (Fig. 11) in Early Cretaceous syn-rift deposits is associated with hyper-extension and mantle exhumation (Ravier, 1959; Kornprobst and Vielzeuf, 1984; Goldberg and Leyreloup, 1990; Lagabrielle and Bodinier, 2008; Clerc *et al.*, 2015b; Chelalou *et al.*, 2016; Clerc *et al.*, 2016; Ducoux *et al.*, 2019). The rifted area goes beyond the Pyrénées and similar inverted

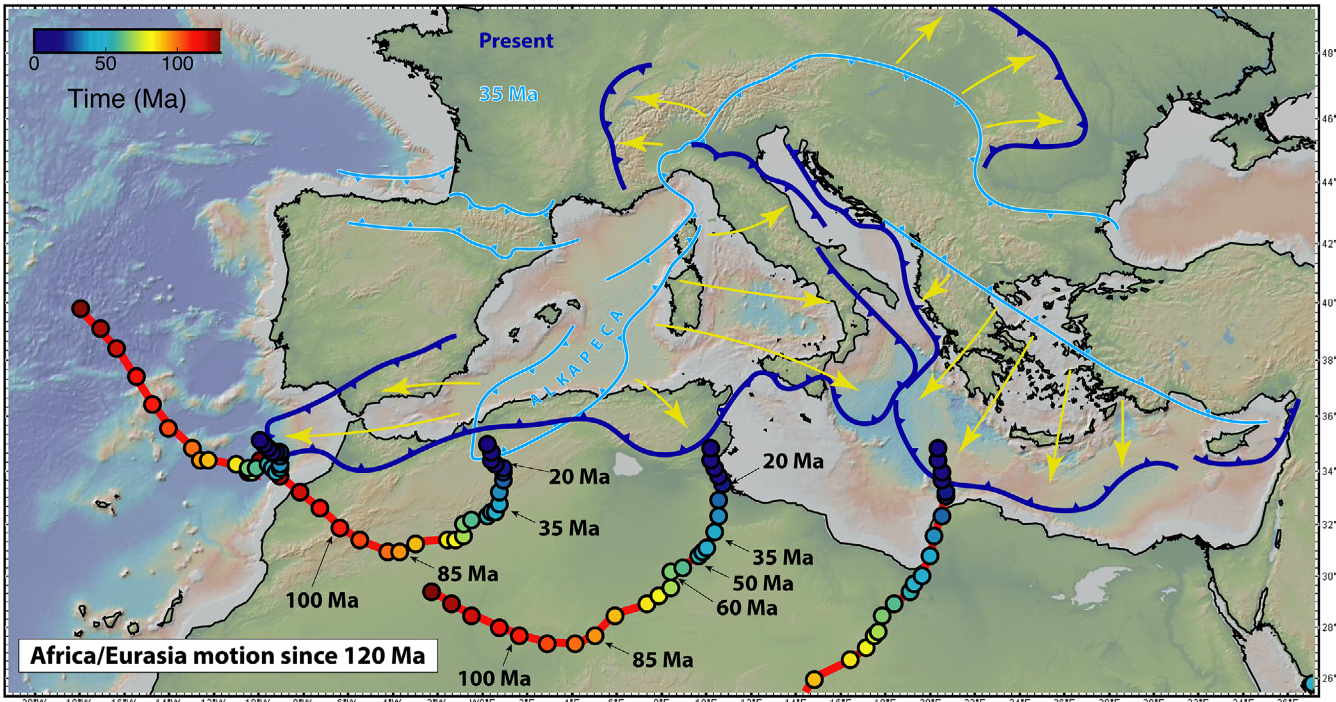

Fig. 6. Africa motion relative to Eurasia and the timing of the successive periods discussed in the paper. Positions of the main trenches and thrusts at 35 Ma and at present and schematic displacements in between.

rifts, the Cameros and Maestrat Basins (Fig. 2), can be found further south within the Iberian chain south of the Ebro Basin (Roca and Guimerà, 1992; Salas and Casas, 1993; Salas *et al.*, 2001; Canérot, 2016; Etheve *et al.*, 2018; Rat *et al.*, 2019). The southwestern part of the Valencia Basin, the Columbrets Basin, is also one of those preserved early rifts (Etheve *et al.*, 2018). Similar rifts are probably hidden below the Gulf of Lion passive margin, as suggested by the Late Cretaceous K-Ar metamorphic age of Paleozoic metasediments recovered at drill hole GLP2 (Fig. 11) (Guennoc *et al.*, 2000) and were overprinted by back-arc extension.

Shortening started in the Late Cretaceous, at around 84 Ma and continued until the Late Eocene on the northern side of the Pyrénées and until the Early Miocene on the southern side (Mouthereau *et al.*, 2014; Bosch *et al.*, 2016; Labaume *et al.*, 2016; Teixell *et al.*, 2018). This early setting of compressional conditions is also recorded in the Betics (Vergés and Fernàndez, 2012; Daudet *et al.*, 2020). The post-orogenic phase, not to be confused with post-orogenic extension that formed the Gulf of Lion rift, started at around 23 Ma and lasted until ~12 Ma with the draping of large alluvial fans at high altitude on top of the northern and southern wedges (Babault *et al.*, 2005; Bernard *et al.*, 2020). Shortening was accommodated by a series of north-dipping and south-dipping thrusts involving the basement in the North Pyrenean Zone and the Axial Zone and mainly the cover in the South Pyrenean Zone above the Triassic evaporites decollement (Choukroune, 1989; Choukroune *et al.*, 1990; Mouthereau *et al.*, 2014; Teixell *et al.*, 2018; Espurt *et al.*, 2019; Grool *et al.*, 2019; Jourdon *et al.*, 2020). The main shortening was accommodated by the northward underthrusting of the Iberian crust and lithospheric mantle below the European lithosphere. The period spanning the late Lutetian and the Bartonian, between 43 and 37 Ma, is a turning-point in the style of shortening across the belt (Waldner *et al.*, 2019). Waldner *et al.* (2019), using compiled thermochronological data, show that, before 40 Ma, shortening was distributed across the belt and the maximum exhumation was recorded in the North Pyrenean Zone, while, after 40 Ma, north-dipping thrusts accommodated the thickening of the Iberian crust and exhumation migrated southward. The temporal changes in the style of shortening further highlights the accretion of different portions of the rifted margin (Jourdon *et al.*, 2019; Ternois *et al.*, 2021). Structural sections across the belt and seismic profiles show this underthrusting below most of the belt (Vergés *et al.*, 1995, Vergés *et al.*, 2002; Chevrot *et al.*, 2018; Teixell *et al.*, 2018; Espurt *et al.*, 2019) (Fig. 11). The same sections also show the presence of mantle bodies exhumed in the early Cretaceous by hyper-extension, high in the present-day crust of the North Pyrenean Zone (Wang *et al.*, 2016). This orogenic structure totally disappears below the eastern Pyrénées where the Moho appears flat and shallows eastward toward the Gulf of Lion passive margin (Chevrot *et al.*, 2018; Diaz *et al.*, 2018) (Fig. 11A, insert). This striking change has been diversely interpreted, even within the Orogen project, either as a major difference in the pre-orogenic template of the belt (Chevrot *et al.*, 2018) or as an effect of post-orogenic extension during back-arc rifting (Jolivet *et al.*, 2020) (Fig. 11). Early foreland basins have formed on the southern (Ebro Basin) and northern (Aquitaine Basin) sides of the belt. The early internal parts of these two basins are now integrated in the belt, especially in the North Pyrenean Zone as Late Cretaceous flysch-type

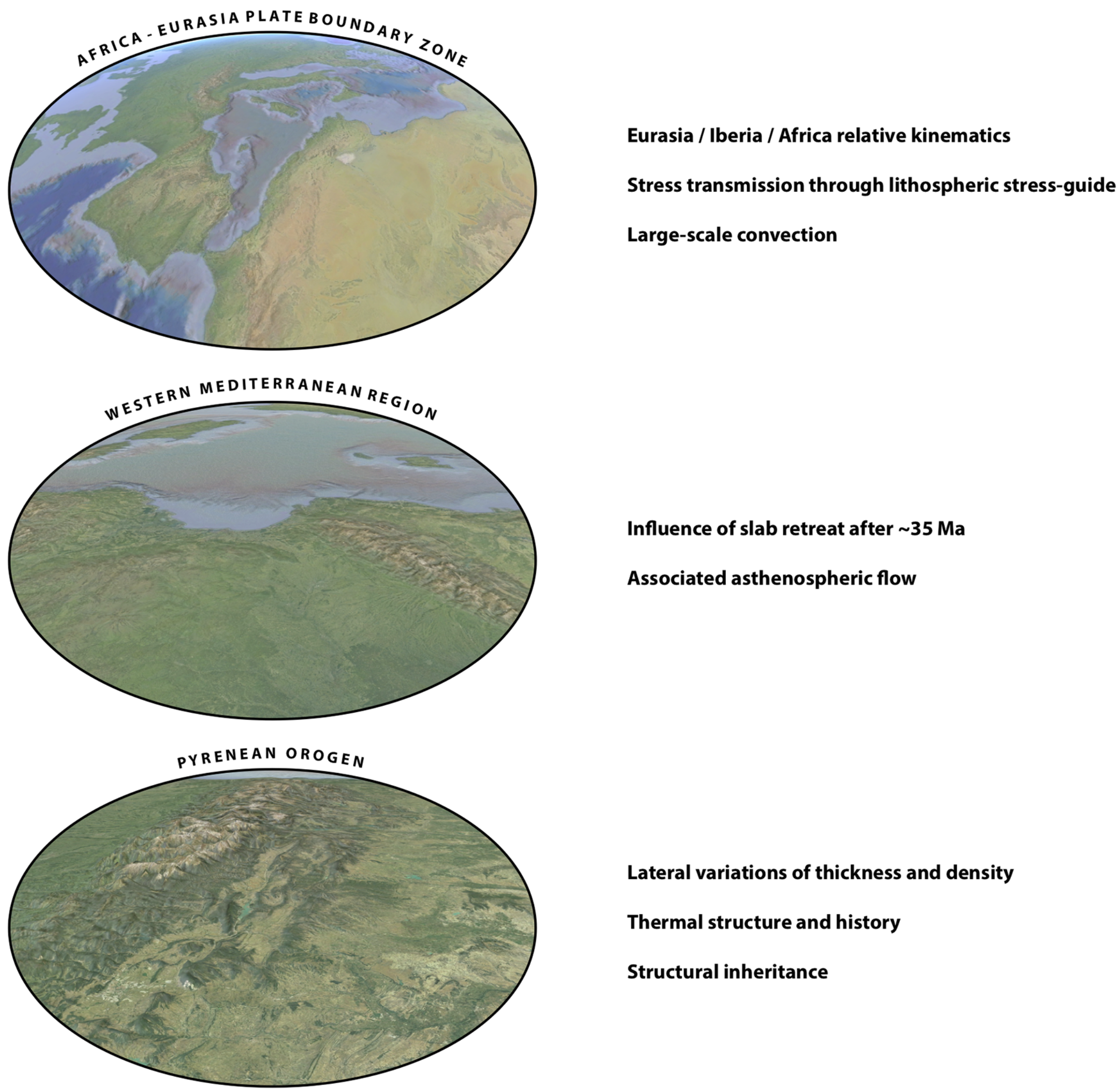

Fig. 7. The different scales and types of forces driving the tectonic evolution of the Western Mediterranean region.

formations (Muñoz, 1992, Muñoz, 2002; Vergés *et al.*, 2002; Ford *et al.*, 2016; Angrand *et al.*, 2018; Grool *et al.*, 2018; Espurt *et al.*, 2019). The basins have then propagated northward and southward.

The transition toward the post-orogenic period is well recorded in Languedoc where syn-tectonic basins formed in a left-lateral transpressional context in the Priabonian during the formation of the West European Rift System, before back-arc extension took over at the end of the Eocene (Séranne, 1999; Séranne *et al.*, 2021). Jolivet *et al.* (2020) proposed that back-arc extension in the Gulf of Lion impacted the eastern part of the range through basal erosion and removal of the upper mantle and lower crust by the asthenospheric flow due to slab retreat. The impact of asthenosphere dynamics beneath the range had already been used as a driver for the late Neogene uplift period around 10 Ma by Gunnell *et al.* (2009), see also Calvet *et al.* (2020) or Huyghe *et al.* (2020). The active extension observed today in the Western Pyrénées seems caused by the juxtaposition of crustal blocks with contrasted density and does not necessarily imply that the system is under extensional boundary conditions (Souriau *et al.*, 2014; Fillon *et al.*, 2021).

2.4 Some key-dates

This short review highlights some key-dates in the evolution of the Western Mediterranean region (Figs. 9 and 12). (1) The early Cretaceous is the period of distributed rifting within Iberia and across the Iberia-Europe boundary zone, (2) the Late Cretaceous sees the first shortening in the Pyrénées, but also in the Alps and the Betics, (3) the Eocene is the climax of shortening, but also of formation of high-pressure and low-temperature (H*P*-L*T*) metamorphic complexes in the continental units of the Alps, Alpine Corsica or the Betics and Kabylies, (4) the end of the Eocene (Priabonian) records the effects of the West European Rift System, (5) from the Oligocene to the early Late Miocene back-arc basins forming above fast retreating slabs and the Eastern Pyrénées are

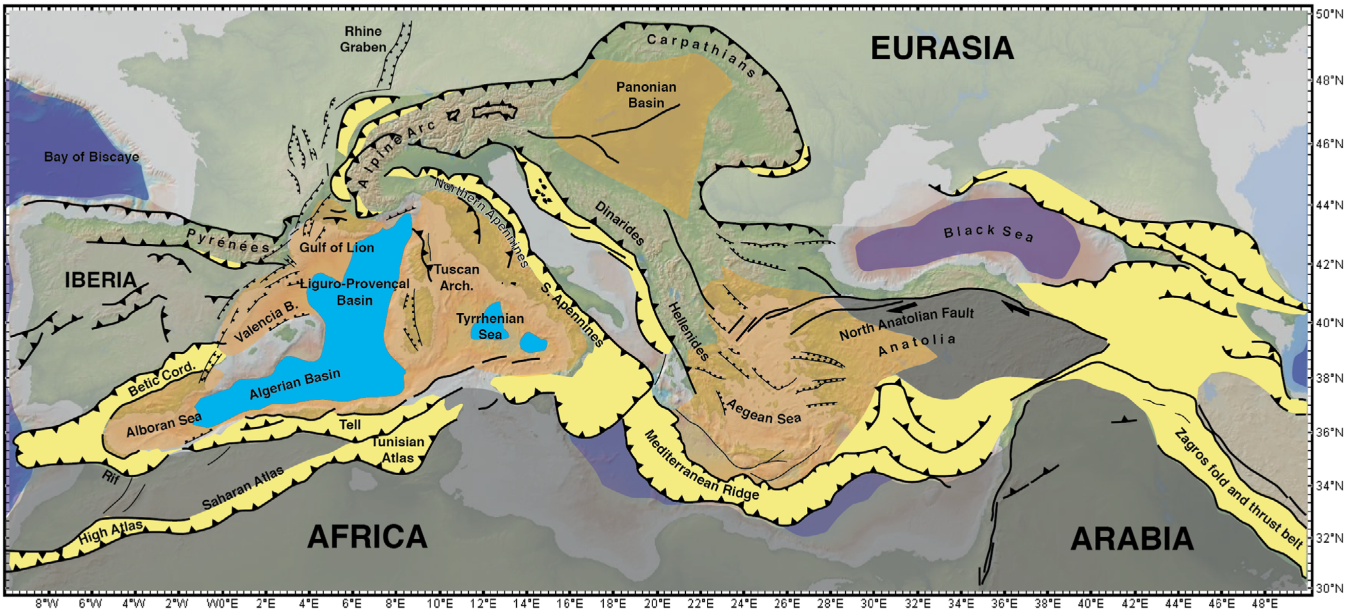

Fig. 8. Tectonic context of the Mediterranean region. Violet: Mesozoic oceanic crust; blue: Cenozoic oceanic crust and/or exhumed mantle; orange: areas affected by Neogene post-orogenic extension.

dismantled during the rifting of the Gulf of Lion, (6) at around 20–15 Ma an episode of tearing separates the retreating slab in two parts, leading to the formation of the Alboran and Tyrrhenian Seas, (7) from the Late Miocene onward north-south compression takes over and the extended orogens are shortened again.

This timing finds echoes in the French-Italian Alps where the Eocene is the major period of formation of blueschists and eclogites in the internal zones and the early Oligocene is characterized by an acceleration of westward thrusting in the external zones (Oberhänsli *et al.*, 2004; Ford *et al.*, 2006; Bousquet *et al.*, 2008; Bellahsen *et al.*, 2014; Bellanger *et al.*, 2014). The timing is also partly similar in the Eastern Mediterranean where the Cycladic Blueschists form in the Eocene, the Aegean Sea back-arc basin starts to rift around 35–30 Ma above the retreating Hellenic slab and a major episode of slab tearing starts at about 15–20 Ma (Bonneau and Kienast, 1982; Maluski *et al.*, 1987; Wijbrans and McDougall, 1988; Wijbrans *et al.*, 1993; van Hinsbergen *et al.*, 2005; Jolivet *et al.*, 2013, Jolivet *et al.*, 2015b). At a larger scale, the timing of the Africa-Eurasia convergence zone also shows the same phases, early Cretaceous extension, Late Cretaceous compression, Eocene compression, a drastic change at around 30 Ma with the collision of Africa and Eurasia and the first rifting in the Gulf of Aden, coeval with the surge of volcanism of the Afar plume (Jolivet and Faccenna, 2000; Faccenna *et al.*, 2013; Jolivet *et al.*, 2016). These coeval events cannot be just coincidences because the Mediterranean region has been a semi-closed system since approximately 35 Ma where internal displacements proceed much faster than the relative motion of Africa and Eurasia at the external boundaries of the system. We now explore these successive periods on various scales, focusing on the Western Mediterranean and the Pyrénées in order to identify the main driving factors of observed deformations, near-field versus far-field.

3 Different scales

We first present a brief review of the main parameters that may play a role in the observed deformation, depending upon scale.

3.1 The scale of the Pyrénées

3.1.1 Kinematic boundary conditions, shortening finite rates and pre-orogenic template

At the scale of the Pyrénées several factors are possibly at play. The first one is the kinematic boundary conditions, whether the Iberian and European plates diverge or converge. A recent reevaluation of the kinematics based on Atlantic magnetic anomalies gives a reliable framework (Macchiavelli *et al.*, 2018) but the precision available on the velocity and direction of relative motion remains debated because of the magnetic quiescence during the Cretaceous Normal Superchron (Aptian to Santonian) and because of uncertainties on the interpretation of the nature of the J oceanic magnetic anomaly in the southern North Atlantic that makes it a doubtful isochron (Nirrengarten *et al.*, 2018). This is mostly true for the rifting period also because the width of the deforming zone is poorly known. A series of rifts have been described from the center of Iberia to the North Pyrenean Zone, but the kinematics and amount of extension accommodated by each of them is poorly constrained (Canérot, 2016; Rat *et al.*, 2019). These Mesozoic rifts had already accommodated significant extension earlier, during the Permian and Triassic (Saspiturry *et al.*, 2019), a fact that is often overlooked, which leads to the overestimation of the relative motion between Iberia and Europe during the Mesozoic rifting. Angrand *et al.* (2020) instead propose reconstructions where the Iberia-Europe plate

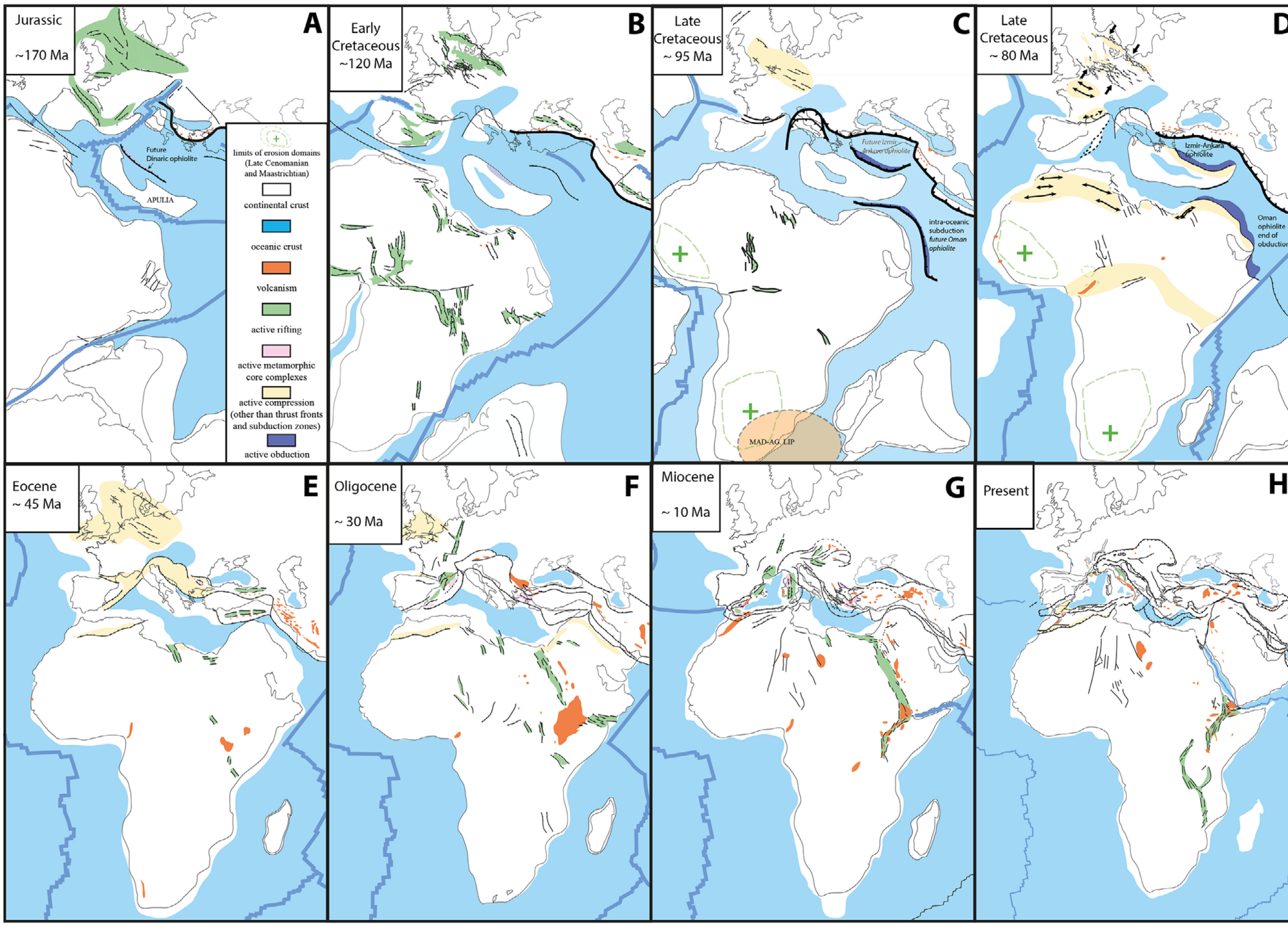

Fig. 9. Reconstructions of the Tethys Ocean from the early Cretaceous to the Present and the evolution of the Mediterranean region in this large-scale framework. This figure is adapted from Jolivet *et al.* (2016a) with the addition of the evolution of northern Europe and the North Sea (Evans *et al.*, 2003).

boundary is a wide deformation zone that does not need a localized left-lateral motion on the Pyrénées, at variance with several available models (Choukroune and Mattauer, 1978; Olivet, 1996; Sibuet *et al.*, 2004; Handy *et al.*, 2010; Jammes *et al.*, 2010; Vissers and Meijer, 2012; Barnett-Moore *et al.*, 2016). Then, the continuation of the rift zone toward the east (toward the Alps) is not precisely known, but Tavani *et al.* (2018), based on a detailed study of the Marguareis extensional domain in the southern Briançonnais and a compilation of observations in the transition zone between the Pyrénées and the Alps, recently concluded that rift basins similar to the North Pyrenean basins, with similar stratigraphy and timing, can be recognized until the South Provence Basin, the Vocontian Basin and the Marguareis basin, with NE-SW transfer faults. During the compressional period, the amount of shortening is more precisely estimated thanks to crustal-scale sections across the Pyrénées (Vergés *et al.*, 2002; Mouthereau *et al.*, 2014; Grool *et al.*, 2018; Teixell *et al.*, 2018; Espurt *et al.*, 2019) but they also come with significant uncertainties and differences. Further south, the exact timing and amount of shortening on the southern boundary of Iberia (Betics) remains to be precisely constrained in the Late Cretaceous (Daudet *et al.*, 2020).

Balancing crustal-scale sections across the Pyrénées is indeed a delicate exercise because the pre-orogenic geometry and the amount of finite shortening are not consensual. Despite the wealth of vintage and more recent seismic data with up-to-date techniques across the belt, some first order features are debated, as illustrated by the recent syntheses of Teixell *et al.* (2018) and Chevrot *et al.* (2018). The geometry of subcontinental mantle at shallow depth below the Mauléon Basin is not similar in all published works, which has drastic consequences when estimating the amount of finite shortening. On the single ECORS profile in the Central Pyrénées or along cross-sections nearby, published amounts of shortening vary between ~90 km and more than 160 km (Muñoz, 1992; Roure *et al.*, 1996; Beaumont *et al.*, 2000; Muñoz *et al.*, 2013; Mouthereau *et al.*, 2014; Grool *et al.*, 2018; Muñoz *et al.*, 2018; Teixell *et al.*, 2018). Then, the amount of apparent shortening in the sedimentary cover of the North Pyrenean Zone or Nappe des Marbres in the Basque-Cantabrian basin taken up by pre-orogenic salt-tectonics is much larger than usually considered. The recent works of Ducoux *et al.* (2019) in the Nappe des Marbres, Menant *et al.* (2016) and Izquierdo-Llavall *et al.* (2020) or Labaume and Teixell (2020) on different sections of the Chainons Béarnais, or Ford and Vergés

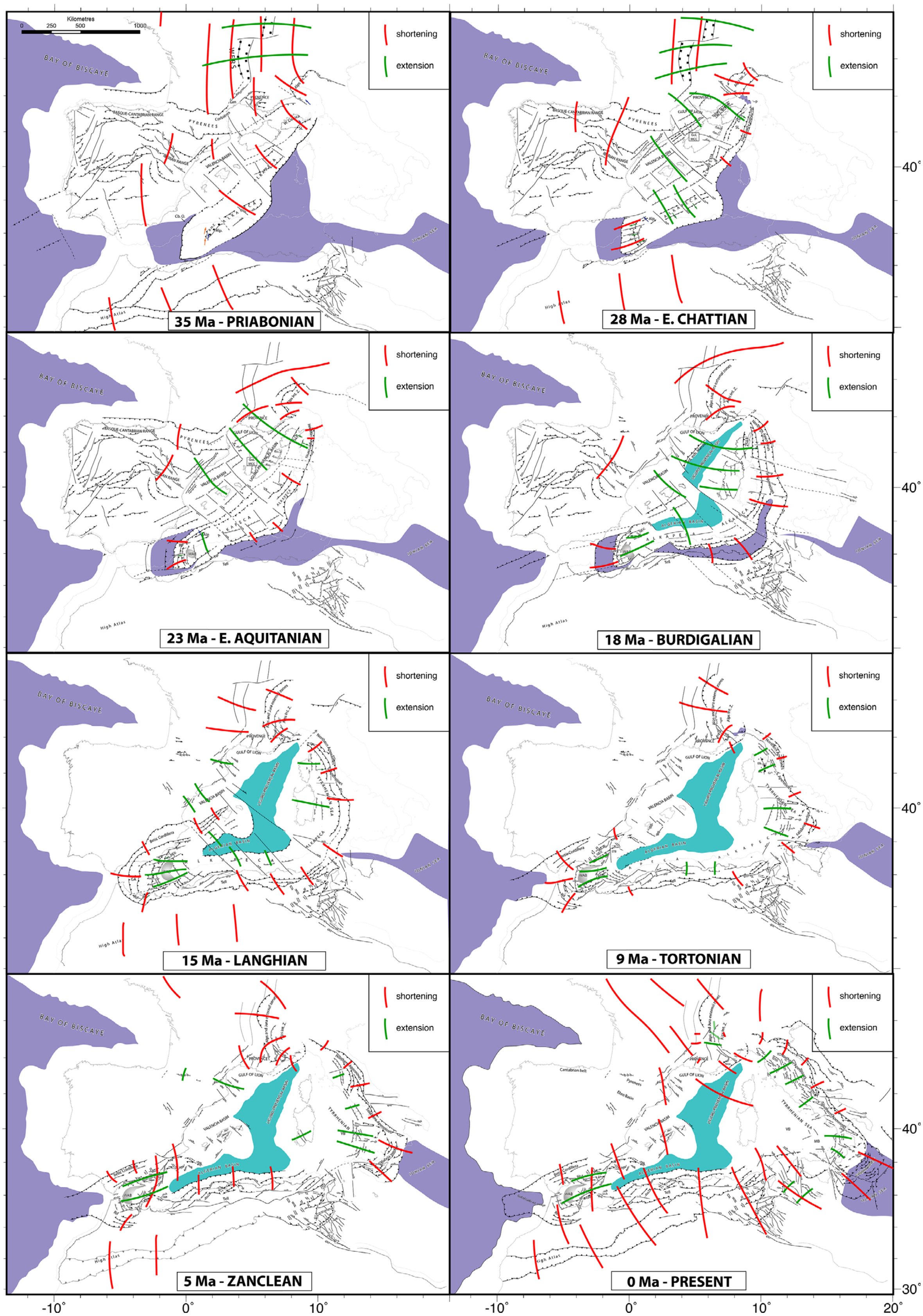

Fig. 10. Schematic evolution of the shortening and extension directions through the reconstructions of Romagny *et al.* (2020) after Bergerat (1987), Séranne (1999), Ziegler and Dèzes (2007), Dèzes *et al.* (2004), Frizon de Lamotte *et al.* (2000, 2008), Walpersdorf *et al.* (2018), Cornet and Burlet (1992) and Baize *et al.* (2013).

(2020) in the eastern Pyrénées show that a significant part of the apparent shortening shown by the deformation of the Mesozoic cover is in fact due to early salt-tectonics during the rifting episode and partly before the onset of the H*T*-L*P* Pyrenean metamorphism. Wicker and Ford (2021) also show that early salt tectonics can be recognized also in the southern part of the Provence fold and thrust belt and that the famous Beausset klippe (Bertrand, 1887) could be interpreted as a megaflap of the Bandol diapir. This must significantly reduce the overall finite shortening across the belt calculated on balanced cross-sections and the kinematic consequences have not yet been fully evaluated. Ford and Vergès (2020) show in addition that a significant component of sinistral strike-slip tectonics should be considered during rifting.

The width of the Pyrenean rift is similarly variable and the least known figure is the width of the exhumed mantle domain during rifting that varies from 15 km (Jammes *et al.*, 2010a, Jammes *et al.*, 2010b; Lagabrielle *et al.*, 2010) to 50 km (Mouthereau *et al.*, 2014), not mentioning the 300 km of a true oceanic domain proposed by Vissers and Meijer (2012). The timing of the first shortening of paleo-margins is also variable in the literature, from 50 Ma to 83 Ma (Mouthereau *et al.*, 2014; Teixell *et al.*, 2018), and evidence for the initiation of flexural basins in the northern and southern forelands all point to the Campanian, thus around 83 Ma or even earlier as soon as the Coniacian and Santonian (Bilotte, 1985; Grool *et al.*, 2018; Andrieux *et al.*, 2021). The first contact between the two necking zones during shortening dates back to the Campanian (Mouthereau *et al.*, 2014). Issautier *et al.* (2018) rather consider that these early deformations correspond to buckling of the European lithosphere under the compressional regime imposed by the Africa-Eurasia convergence. Finally, the significance of the flat and shallow Moho below the eastern Pyrénées is also discussed, whether it results from the shortening of a different pre-orogenic geometry compared to the western Pyrénées (Chevrot *et al.*, 2018) or to the result of post-orogenic extension during the rifting of the Gulf of Lion (Wehr *et al.*, 2018; Jolivet *et al.*, 2020).

All these first order uncertainties make the relations between the Africa-Eurasia convergence through time, which is more precisely constrained, and the tectonic evolution of the Pyrénées, difficult to assess. It is particularly difficult to be conclusive on the amount of shortening in the Eastern Pyrénées where the influence of post-orogenic thinning during the rifting of the Gulf of Lion is major. If we consider the 125 km estimated by Vergès *et al.* (2002), it is not much different from the average values proposed for the Central Pyrénées. East of the Catalan transfer zone, below the present-day Gulf of Lion passive margin, the amount of pre-Oligocene shortening is unknown. A gradient of shortening from the Cantabrian Range and Western Pyrénées (~90–100 km) to the Central Pyrénées (~120 km) seems instead more consensual (Teixell *et al.*, 2018) and in agreement with prediction of large scale kinematic models (Macchiavelli *et al.*, 2018). Note, however, that the kinematic model of Macchiavelli *et al.* (2018) does not include any intraplate shortening within Iberia. It is also logical given the transition from the Bay of Biscay to the orogen, but these figures should be revised once the effect of early salt tectonics is subtracted. Moreover, the width of the early Cretaceous rifts is unknown and restoring the section at crustal scale is thus difficult.

One major parameter at this scale is the east-west variation of the pre- to post-orogenic structural template. Given the uncertainties on the amount of shortening briefly discussed above, reconstructing the pre-orogenic geometry of the margins is not well constrained. Tugend *et al.* (2014) tried to circumvent this difficulty by mapping the rift domains within the Pyrénées and compare with the nearby margins of the Bay of Biscay. They come up with a rift with constant width across the entire belt, closing westward below the Cantabrian Range and relayed northward by the Bay of Biscay rift closing eastward. Lescoutre and Manatschal (2020) further show how the geometry and interactions in space of these two rift systems can explain the present-day geometry of the Western Pyrénées and the transition with the Cantabrian Range. Cadenas *et al.* (2021, in prep.) and Miro (2020) further show that the rifting should be divided in two main stages after the first Triassic episode: a first Late Jurassic to Barremian stage, mainly distributed south of the present-day Pyrénées and a second Aptian-Cenomanian stage more focused in the Pyrénées.

An alternative approach is to consider the distribution of the Internal Metamorphic Zone (IMZ) high-temperature metamorphism. Earlier papers described a gradient of temperature from west to east in the North Pyrenean rifts where the mantle is exhumed (Clerc and Lagabrielle, 2014; Clerc *et al.*, 2015b), but more recent studies show a more complex situation. The westward decrease of the maximum recorded temperature is clear from the Boucheville Basin to the Chainons Béarnais, but maximum temperatures as high as those of the Boucheville Basin are also recorded in the Nappe des Marbres near the western end of the belt, south of the Cinco Villas Massif (Ducoux *et al.*, 2019; Ducoux *et al.*, 2021a). How does this distribution of maximum temperatures exactly fit the reconstruction of the early Cretaceous rift is not precisely known, but the asymmetry of the thermal structure suggests that the rift was also asymmetric, an observation compatible with current models of mantle exhumation below low-angle detachments (Lagabrielle *et al.*, 2019a, 2019b; Lescoutre *et al.*, 2019; Ducoux *et al.*, 2019, 2021a; Lescoutre and Manatschal, 2020, Saspiturry *et al.*, 2020). The triangular shape of the Bay of Biscay oceanic domain has suggested that the amount of extension was larger in the west than in the east, but this gradient is not clear either, and the lack of syn-rift kinematic indicators in the field does not help. The direction of extension during rifting is often deduced from that of supposed transfer faults such as the Pamplona Fault, which existence and significance is debated (Tugend *et al.*, 2014; Saspiturry *et al.*, 2019; Lescoutre and Manatschal, 2020). The NNE-SSW direction of extension can also be deduced from the geometry of the Arzacq basin (Masini *et al.*, 2014; Saspiturry *et al.*, 2019; Issautier *et al.*, 2020; Ducoux *et al.*, 2021b). Frasca *et al.* (2017), however, suggest a significant component of sinistral strike-slip tectonics during the formation of rift basins from the Early Cretaceous to the early Late Cretaceous, based on an analysis of Mesozoic supra-salt syn-tectonic sedimentation in the eastern Pyrénées. The strike-slip component is suggested by the arrangement of salt walls in map view related to basement faults.

3.1.2 Crustal root and body forces, rheological issues

Crustal thickening during the Eocene has increased the potential gravitational energy stored in the Pyrenean crust, partly balanced by the lithospheric root. If no oceanic slab is visible in tomographic models underneath the Pyrénées, a crustal root is nevertheless present (Chevrot *et al.*, 2014; 2018) that applies a load on the crust and prevents the formation of a higher relief. Below the eastern Pyrénées, this root is not present and several possibilities have been discussed, either a different amount of finite shortening, different distribution of strain and a different pre-orogenic template or a removal of the root during the rifting of the Gulf of Lion (Chevrot *et al.*, 2018; Jolivet *et al.*, 2020).

These different boundary and body forces apply on a lithosphere with a given rheology and a given pre-orogenic tectonic history and inheritance is thus important to consider, especially in the early history of mountain belts (Manatschal *et al.*, 2021). The nature of the deep crust during the formation of the orogen is thus an important question to address. Whether weak or strong, it will react differently. The continental basement of the Iberian and European lithospheres has been strongly mobilized during the Variscan orogeny until the exhumation of granulitic domes during late-orogenic extension (Cochelin *et al.*, 2017; Saspiturry *et al.*, 2019). Granulitic slices are observed in the IMZ associated with peridotites. They are granulitic paragneisses, catazonal marbles, intermediate and basic charnockites equilibrated at 8 kbar and 750–800°C associated with a Hercynian thermal anomaly around 300 Ma (Vielzeuf and Kornprobst, 1984). The late- to post-orogenic evolution of the Variscan orogen in the Pyrénées was characterized by distributed extension and the exhumation of migmatitic gneiss-cored metamorphic core complexes with intense magmatism (Denèle *et al.*, 2007, 2008, 2009; Cochelin *et al.*, 2017), a situation similar to that of the Aegean region during the Oligocene and Miocene (Jolivet and Brun, 2010). The lithology of the lower crust underneath the Pyrénées and more specifically the ratio between resistant granulites and weaker gneiss and migmatites is unknown. It is, however, likely that the middle crust and part of the lower crust after the end of the collapse of the orogen were highly heterogeneous essentially made of granites and gneiss and the lower crust locally made of granulite and mafic material underplated during the Permian after the end of the Variscan orogeny as suggested by the study of the Sondalo gabbroic complex in the Central Alps (Petri *et al.*, 2017). This suggests a rather weak bulk rheology for the middle crust, opening the possibility of a decoupling during Pyrenean shortening. Bellahsen *et al.* (2019) and Airaghi *et al.* (2020) further show that fluid circulation during pre-orogenic rifting episodes since the Permian have significantly modified the mineralogical content of the upper crustal basement, with intense sericitization, and thus weakened the resistance of the upper crust, a possible explanation for the observed distributed deformation during Pyrenean shortening. The early Cretaceous rifting episode was locally associated with mantle exhumation and serpentinisation, but the intensity of this serpentinisation differs between peridotite bodies (Clerc *et al.*, 2014) that were exhumed within the upper crust during rifting. Such weak levels might have nevertheless been used as decoupling levels during convergence (Tugend *et al.*, 2014; Manatschal *et al.*, 2021).

Models of formation of magma-poor passive margins suggest that weakening of the crust and weakening of the upper mantle by serpentinisation during extension can lead to lower crust and mantle exhumation with low-angle extensional shear zones (Lavier *et al.*, 1999, 2019). Although large uncertainties remain on the mechanical stratification of the pre-convergence crust, it was thus probably in average rather weak with several decoupling levels and a weak upper crust. The rapid onset of convergence after the end of the rifting episode in the Santonian (see Mouthereau *et al.*, 2014 for a review) or Eocene (Vacherat *et al.*, 2017) suggests that the lithosphere was still hot and thus weak when compression started. This relative weakness helped the progressive localization of a new plate boundary in the Pyrénées during the Late Cretaceous, as described by Dielforder *et al.* (2019). From a situation where compressional stresses were recorded across a wide zone from the north of Africa to northern Europe, compressional deformation progressively localized within the former rift systems, i.e. the Pyrénées and the Iberian Chain. The degree of Late Cretaceous strain localization in the future Betic-Rif orogen is unknown, but Daudet *et al.* (2020) show that a foreland basin was already present there in that period and subsided further around 50 Ma.

The formation of a crustal root by underthrusting of the Iberian crust underneath the Pyrénées as well as the presence of mantle rocks high up in the nappe stack introduce sharp lateral density gradients and thus a complex distribution of body forces. Souriau *et al.* (2014) show that the extensional earthquakes within the Central Pyrénées are intimately associated with contacts between blocks of different densities and that it might not necessarily reflect the regional stress regime. Depending upon the completeness of metamorphic recrystallization in the subducted portion of Iberian crust and the intensity of serpentinisation in mantle bodies incorporated in the orogenic wedge, the distribution of densities can significantly vary, even more so considering the along-strike changes in the overall structure of the orogenic wedge shown by receiver-function profiles (Chevrot *et al.*, 2018). Using homogeneous layers and the 3-D modelling software GeoModeller, Wehr *et al.* (2018), however, successfully reproduce the distribution of gravity anomalies over the Pyrénées. This may suggest either that the subducted Iberian crust has not been eclogitized nor otherwise intensely metamorphosed during underthrusting, or that the depth of interfaces should be modified for the deep parts of the wedge where the deep root is present. Wehr *et al.* (2018) furthermore show that a strong low-density anomaly is required in the upper mantle and/or lower crust to account for the negative Bouguer anomaly below the eastern Pyrénées. They conclude that a high-density lower crust is missing below this region. Dufréchou *et al.* (2018) also conclude to the absence of high-density lower crust below the eastern Pyrénées, except locally where the lower crust could have been eclogitized. As discussed later, rifting of the Gulf of Lion might be responsible for the removal of the lower and associated upper mantle from below the eastern part of the belt, which could also explain the fast cooling and exhumation period around 30 Ma documented by low-temperature thermochronology (Jolivet *et al.*, 2020). One point of discussion is the extent of this lower crust removal process during the Oligo-Miocene rifting, how far did it reach westward underneath the Pyrénées. The interpretation of Jolivet *et al.* (2020) is that the westernmost affected region is the region under which the crustal root disappears on the seismic profiles, between profiles C and D (Fig. 11)

3.2 The scale of the Western Mediterranean

3.2.1 Tectonic drivers at the scale of the Western Mediterranean, convergence vs slab retreat, asthenosphere vs lithosphere

Before the major change of subduction dynamics at 35–30 Ma, the relative motion of Africa and Eurasia was first divergent before 90–84 Ma and convergent afterward in the Western Mediterranean realm (Fig. 6). After 32 Ma, while the northward motion of Africa toward Eurasia continued at the same pace or even slower, the retreat of the slab subducting below Sardinia and Corsica became the dominant kinematic feature. All reconstructions (Gueguen *et al.*, 1998; Rosenbaum *et al.*, 2002a; Jolivet *et al.*, 2003; Lacombe and Jolivet, 2005; Romagny *et al.*, 2020; Mantovani *et al.*, 2020) show arcs moving outward at high velocities (from 3 to more than 10 cm/yr), outpacing the overall convergence (Figs. 6, 10 and 13). Motion paths on Figures 13 and 14 show that the fast displacements of arcs during slab retreat are dominant over the velocity of Africa-Eurasia convergence and the absolute motion of the two plates.

After 32 Ma, most of the deformation in the Western Mediterranean was then no longer under the control of convergence but of slab retreat and associated asthenospheric flow. The thick crust of mountain belts (Provence, Alpine Corsica, Internal Apennines, Calabria) formed during earlier stages was then thinned in the back-arc domain, while new ones formed in the vicinity of the retreating subduction zones and were progressively dilacerated during the migration of the thrust front (Jolivet *et al.*, 1998; Jolivet and Faccenna, 2000; Faccenna *et al.*, 2001a; Rosenbaum *et al.*, 2002a; Jolivet *et al.*, 2003; Lacombe and Jolivet, 2005; van Hinsbergen *et al.*, 2014; Romagny *et al.*, 2020).

One of the main questions is the transition in space from the Pyrénées to the extending domain in the east (Gulf of Lion) and from the Alps to the Apennines which migrated in opposite directions after 35–30 Ma (Fig. 10). Similarly, at around 16 Ma, the slab retreating southward carried the AlKaPeca blocks in contact with the North African margin, and a new major slab tearing event divided it in two parts, each retreating in opposite directions, westward and eastward, forming the opposite Calabrian and Gibraltar subduction zones. The width of the slab portions were progressively reduced and the velocity of retreat accordingly increased, outpacing even more the convergence (Wortel and Spakman, 1992; van der Meulen *et al.*, 1998; Faccenna *et al.*, 2005; Govers and Wortel, 2005; Jolivet *et al.*, 2021) and a large component of strike-slip motion is recorded along the northern and southern margins of the Alboran domain and north of Sicily (Fig. 13).

The case of the Pyrénées is, however, peculiar. While the Eastern Pyrénées where actively dismantled by the underlying asthenospheric flow related to slab retreat (Jolivet *et al.*, 2020) or other asthenosphere dynamics (Gunnell *et al.*, 2009; Huyghe *et al.*, 2020; Calvet *et al.*, 2021), the Central and Western Pyrénées were still recording shortening until the Late Oligocene or the base of the Miocene (21 Ma) (Jolivet *et al.*, 2007; Mouthereau *et al.*, 2014; Labaume *et al.*, 2016; Muñoz *et al.*, 2018). This shows that the extensional stress regime imposed by the retreat of the Apennines slab did not affect the whole of the Pyrenean orogen and that compressional stresses were still transmitted through the lithospheric stress-guide across the Africa-Iberia-Europe plate boundary zone.

After the end of this episode of back-arc extension, N-S compression resumed in the Betics around 8 Ma and finally most of the Western Mediterranean was under N-S compression and locally E-W extension (Meghraoui *et al.*, 1986; Yelles *et al.*, 2009; Strzerzynski *et al.*, 2010; Billi *et al.*, 2011; Meghraoui and Pondrelli, 2012; Medaouri *et al.*, 2014; Soumaya *et al.*, 2018; Zitellini *et al.*, 2019; Strzerzynski *et al.*, 2021) as shown by the current seismic activity (Figs. 3 and 4). The Africa-Eurasia convergence is certainly one of the main drivers in this new situation, leading to the localization of shortening along the northern margin of Africa and reactivation of the former passive and strike-slip margins of the Alboran domain, coeval with a second episode of shortening all along the High Atlas mountain range (Frizon de Lamotte *et al.*, 2000; Jolivet *et al.*, 2006; Frizon de Lamotte *et al.*, 2008; Lanari *et al.*, 2020a, Lanari *et al.*, 2020b). The northward asthenospheric flow due to large-scale convection is also an important driver of crustal deformation, especially around the slab dipping vertically under the western Betics and the Alboran Sea as recently suggested by Spakman *et al.* (2018) and Capella *et al.* (2019).

3.2.2 The scale of the Eurasia-Africa plate boundary

At the scale of plates (Fig. 9), the relative motion of Africa and Eurasia is an important driver of crustal deformation. Whether the relative motion is divergent (before 84 Ma) or convergent (after 84 Ma), extensional or compressional stresses will be transmitted through the lithospheric stress-guide. Plates are, on the other hand, carried by the convecting mantle underneath. In the Late Cretaceous, compression is recorded across a very large domain from the center of the African plate to Northern Europe, this period culminating with the obduction of ophiolites in the eastern Mediterranean and Middle East. The transmission of stresses through the lithosphere can be modulated by the pattern of convection underneath, whole-mantle convection or local convection due to slab retreat in the Mediterranean (Jolivet *et al.*, 2016a). Plate motion and deformation being driven by slab pull and/or basal drag by the convecting mantle underneath (see Fig. 1 and related text), any change in the deep convection pattern will reflect in the relative and absolute motions of plates. All along the former Tethyan Ocean, from Indonesia and the Himalayas, all the way to the Western Mediterranean, India, Arabia and Africa are carried by the asthenospheric conveyor belt toward the north faster than the slow motion of the Eurasian plate, leading to convergence (Becker and Faccenna, 2011; Faccenna *et al.*, 2013b). This convergence decreases from east to west from more than 5 cm/yr to less than 1 cm/yr (Fig. 6). One important aspect at this scale is the change in the direction of convergence during the Cretaceous, from a large component of strike-slip motion toward pure convergence, perpendicular to the Africa-Eurasia plate boundary zone (Fig. 6) and the associated changes in the convection pattern.

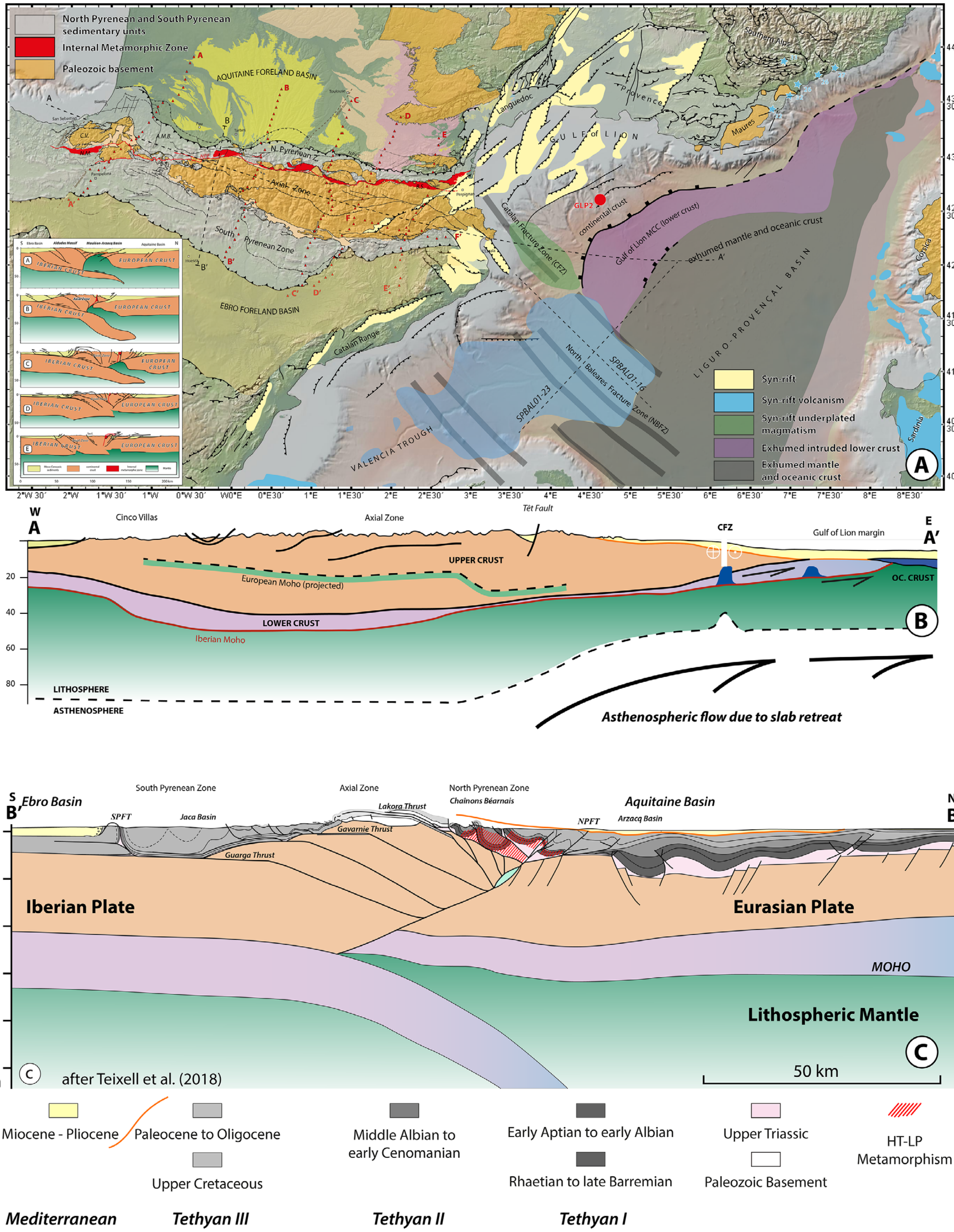

Fig. 11. A. Tectonic map of the Pyrénées, Gulf of Lion and northeastern Valencia Basin. Aligned red triangles show the receiver-functions profiles of Chevrot *et al.* (2018) interpreted in the inset (from A to E, eastward). Dashed lines offshore show the position of the seismic lines of Figure 17. The dashed line along the strike of the Pyrénées reaching the Gulf of Lion margin is the section shown below. AMB: Arzacq-Mauléon Basin; BS: Boucheville Syncline; CV: Cinco Villas Massif; NMU: Nappe des Marbres Unit; GLP2: Golfe du Lion profond drillhole #2. B. Along-strike section of the Pyrénées and Gulf of Lion margin. CFZ: Catalan Fracture (Transfer) Zone. C. Across-strike section of the Central Pyrénées, modified after Teixell *et al.* (2018) showing the different stages discussed in the paper.

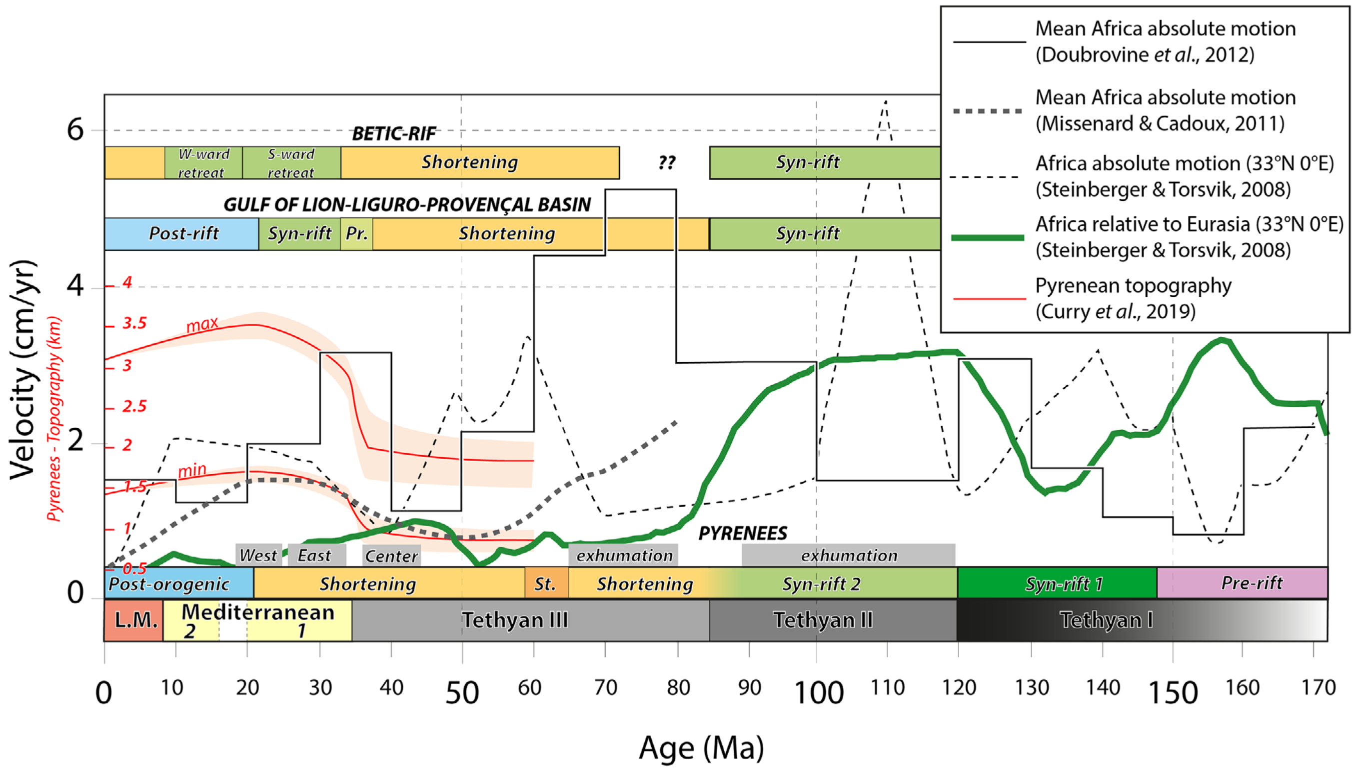

Fig. 12. Timing of various kinematic events in and around the Western Mediterranean region compared to the Pyrénées and the three main periods distinguished in this paper, *Tethyan Tectonics*, *Mediterranean Tectonics* and *Late-Mediterranean Tectonics* (L.M.). Mean Africa absolute motion after Doubrovine *et al.* (2012) and after Missenard and Cadoux (2012), Africa absolute motion for a point located at 33°N and 0°E after Steinberger and Torsvik (2008), motion of Africa relative to Eurasia for the same point after Steinberger and Torsvik (2008), topographic evolution of the Pyrénées after Curry *et al.* (2019).

In the following we discuss these interactions between near-field and far-field stresses for each of the main periods of the geodynamic evolution of the Africa-Eurasia plate boundary, from the early Cretaceous to the Present. We have grouped the different events discussed above and below into three first-order periods (Figs. 11 and 12); (i) *Tethyan Tectonics*: from early Cretaceous to Eocene, which corresponds to the pre-Mediterranean stage, before the inception of fast slab retreat, this stage is further divided in Tethyan I, II and III to depict the progressive rifting and subsequent shortening episodes (ii) *Mediterranean Tectonics*: from 35–30 to 8 Ma, which is the stage where slab retreat and back-arc extension are mostly active, the typical Mediterranean dynamics and (iii) *Late-Mediterranean Tectonics*: from 8 Ma to the Present, corresponding to the progressive return to a pre-Mediterranean situation where compressional stresses are not diverted by slab retreat and are again transmitted from Africa to Eurasia.

4 *Tethyan Tectonics*, from Early Cretaceous to Eocene

Before the beginning of slab retreat around 32 Ma, the main geodynamic process at work was the relative motion of Africa and Eurasia, divergent before 84 Ma, convergent afterward. The Pyrénées started to form at about 84 Ma and the climax of crustal thickening in the Pyrénées across the range is recorded in the Eocene (Bosch *et al.*, 2016; Mouthereau *et al.*, 2014; Teixell *et al.*, 2018; Waldner *et al.*, 2019). The transition from extension to shortening is abrupt in the Pyrénées and is characterized by a change in the sedimentation regime within the North Pyrenean Zone and in the northern foreland with the development of the Aquitaine foreland basin and a tilt of the Axial Zone before the Campanian in the Coniacian-Santonian (Bilotte, 1985; Biteau *et al.*, 2006; Ford *et al.*, 2016; Rougier *et al.*, 2016; Grool *et al.*, 2018, Andrieux *et al.*, 2021; Issautier *et al.*, 2018), which is compatible with the change in the trajectory of Africa relative to Eurasia (Fig. 6). If the geometry of the belt and the rheological properties of the lithosphere (elastic thickness) are strongly impacted by the heritage from the pre-orogenic rifting episode (Angrand *et al.*, 2018; Espurt *et al.*, 2019; Manatschal *et al.*, 2021), the overall tectonic evolution is also dependent on far-field stresses and on the progressive localization of a new zone of deformation during the Late Cretaceous (Dielforder *et al.*, 2019).

4.1 Late Jurassic to Cenomanian rifting

After the Permo-Triassic rifting, two episodes of rifting are recorded in Iberia and around (Vielzeuf and Kornprobst, 1984; Puigdefabregas and Souquet, 1986; Vergés and Garcia-Senz, 2001; Lagabrielle and Bodinier, 2008; Miró, 2020; Miró *et al.*,

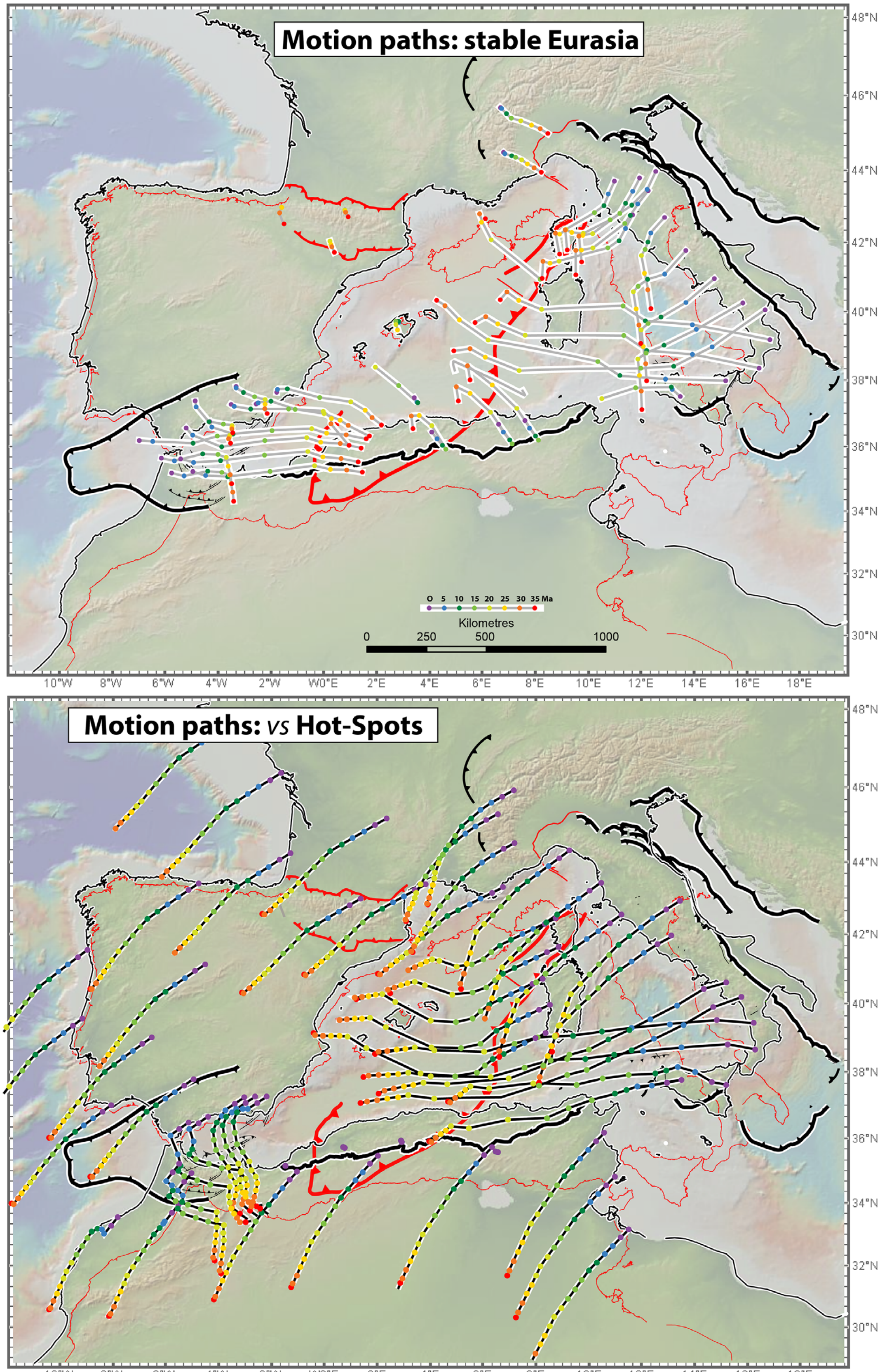

Fig. 13. Motion paths of a series of points within the Western Mediterranean back-arc region calculated from the reconstructions of Romagny *et al.* (2020). Upper: stable Eurasia; Lower: with respect to hotspots.

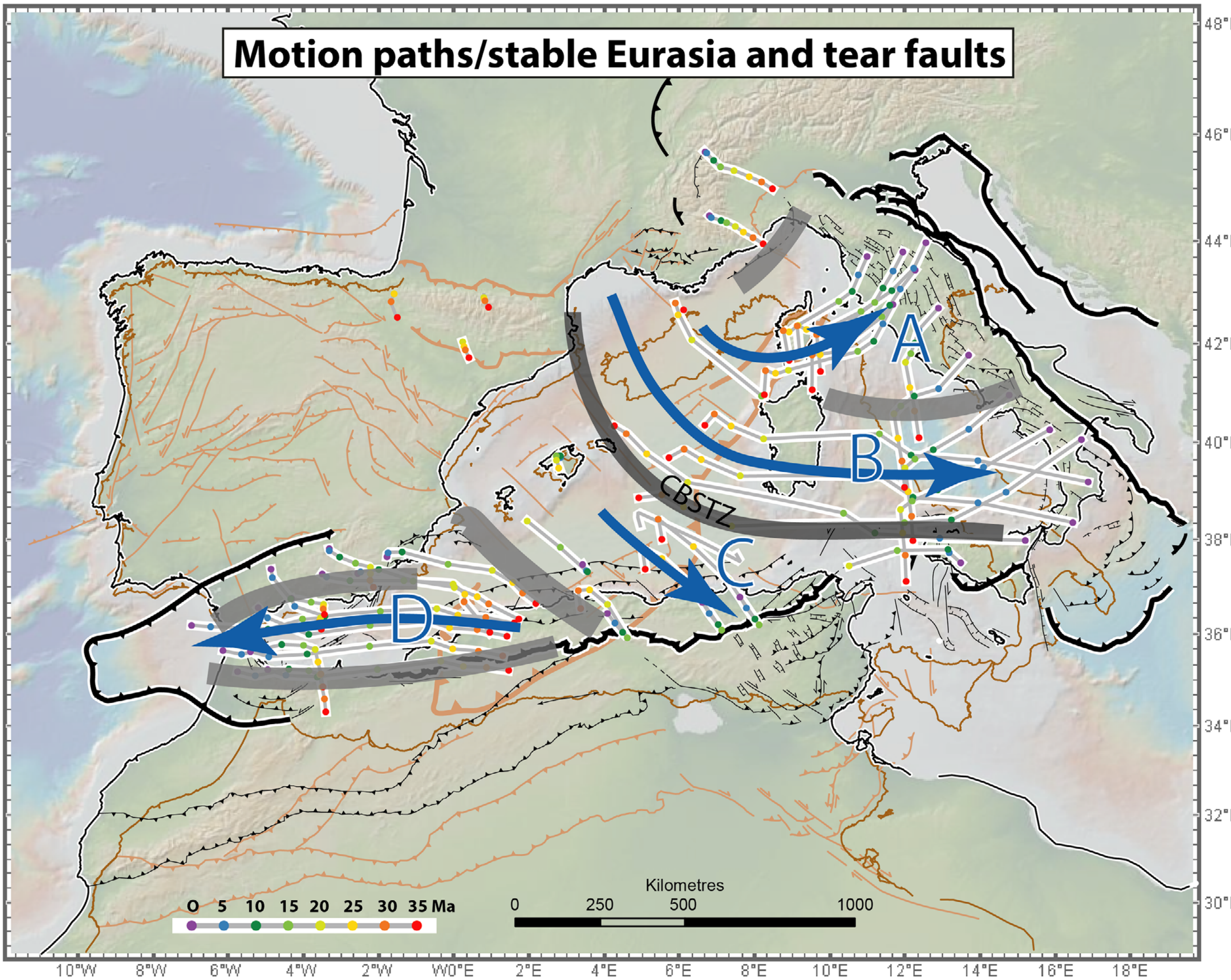

Fig. 14. Motion paths and transfer faults separating coherent domains after the kinematics of Romagny *et al.* (2020). CBSTZ: Catalan Baleares Sicily Transfer Zone (*e.g.* Paul Fallot fault of ancient works).

2020). The first episode spans the Late Jurassic and the early Cretaceous (Neocomian) (see Figs. 11 and 12) are recorded within Iberia in the Basque-Cantabrian basin (Miró, 2020; Miró *et al.*, 2020), the Cameros basin and the Maestrat basin (Rat *et al.*, 2019), the Pyrénées (Tavani *et al.*, 2018) and the offshore Columbrets basin (Etheve *et al.*, 2018). At broader scale this rifting stage is also recorded within the westernmost Iberian plate (Alves *et al.*, 2009; Alves and Abreu Cunha, 2009; Soares *et al.*, 2012). The second episode occurs from the Late Aptian to the Cenomanian and is more focused on the Pyrénées (Jammes *et al.*, 2010a, Jammes *et al.*, 2010b; Masini *et al.*, 2014; Tugend *et al.*, 2014). The transition zone toward the Alps shows similar rift basins (Tavani *et al.*, 2018). The Betic Cordillera records continuous rifting spanning the two periods associated with salt tectonics (Pedrera *et al.*, 2020). This distributed rifting and then focalization on the Pyrenean range is actually observed over most of the African continent (Guiraud *et al.*, 2005; Frizon de Lamotte *et al.*, 2015) as well as northern Europe (Ziegler, 1990; Evans *et al.*, 2003; Kley and Voigt, 2008). It corresponds to the opening of the Atlantic ocean and the period of oblique divergence between Africa and Eurasia (Fig. 6). It is also noticeable that the southern active margin of Eurasia above the subduction of the Neo-Tethys is characterized by the formation of back-arc basins at this period (Dercourt *et al.*, 1986; 1993).

4.2 Late Cretaceous compression and strain localization, lithospheric vs larger-scale

The first shortening in the Western Mediterranean region is coeval with the first convergence between Africa and Eurasia (Fig. 6). Dielforder *et al.* (2019) propose that the progressive localization of a new "plate-boundary fault" in the Pyrénées could explain the observed tectonic evolution in the range and the foreland basin, as well as the variable propagation of compressional stresses at long distance in the foreland, until the North Sea (Ziegler, 1988, Ziegler, 1990; Guillocheau *et al.*, 2000; Evans *et al.*, 2003; Dèzes *et al.*, 2004; Ziegler and Dèzes, 2007; Bourgeois *et al.*, 2007; Grool *et al.*, 2018) and Central Europe grabens (Kley and Voigt, 2008). They note that, while

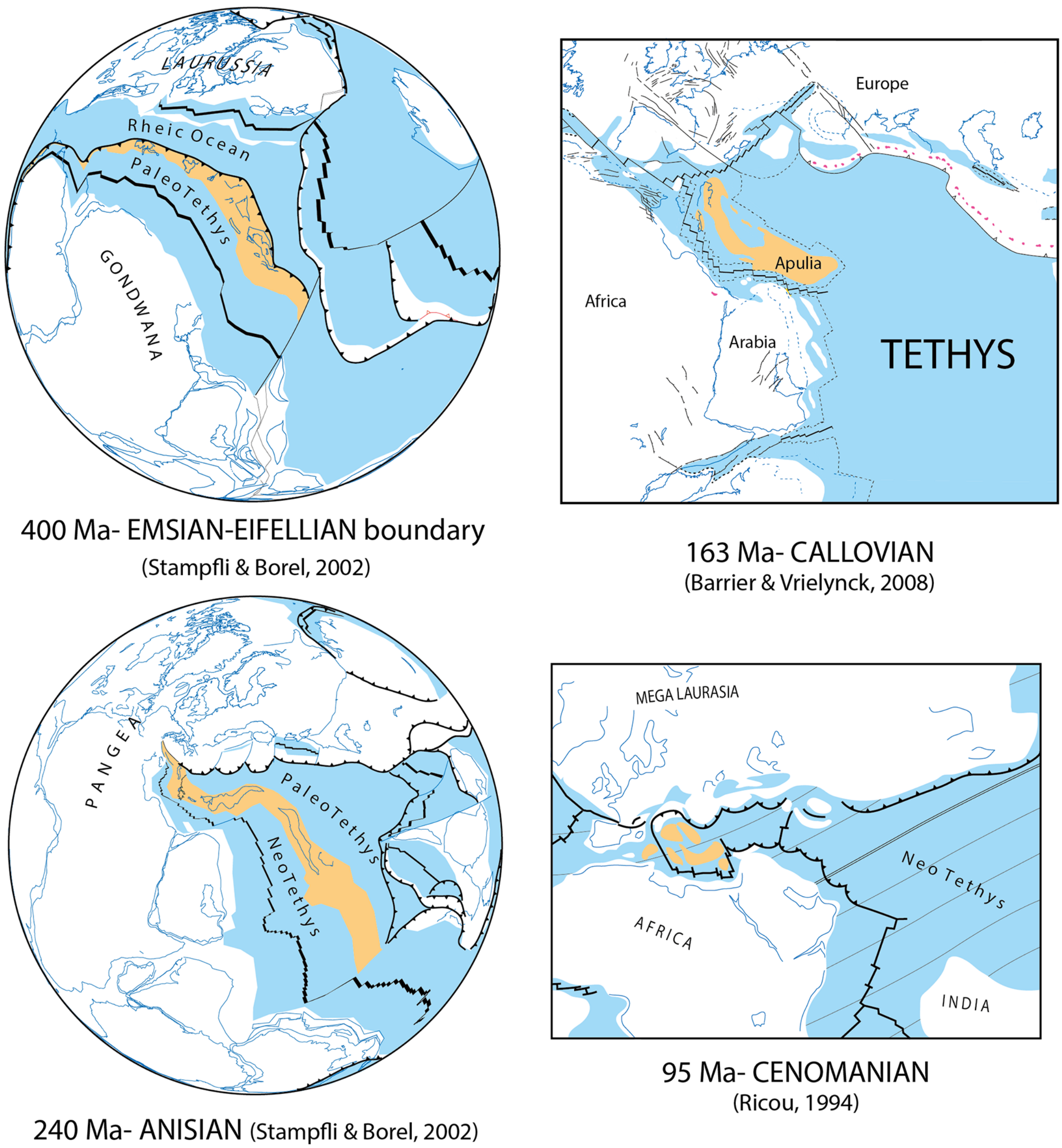

Fig. 15. Four reconstructions of the Tethys Ocean at four stages, Devonian, Anisian, Callovian and Cenomanian after Ricou (1994), Barrier and Vrielynck (2008) and Stampfli and Borel (2002). These reconstructions show the rifting of continental blocks (orange color) away from Gondwana or Africa crossing the ocean to ultimately collide with the southern margin of Eurasia. The most recent of such blocks is the Arabian plate.

rifting was active in the North Sea at the same time as in the Pyrénées, compressional reactivation of the rifted basins is also coeval in the Pyrénées and in the North Sea, until the end of the Cretaceous (Evans *et al.*, 2003).

Except for uplift zones around the British Isles in the Selandian, the Paleocene is a relatively quiet period in the north with little or no-subsidence recorded in the foreland basin around 60 Ma (Sinclair *et al.*, 2005; Mouthereau *et al.*, 2014; Ford *et al.*, 2016), at a time of slow overall convergence (Macchiavelli *et al.*, 2018). Dielforder *et al.* (2019) propose that the progressive formation of a new fault momentarily caused an increase of compressional stresses that could propagate far in the foreland and then a release once the fault had localized. The origin of compressional stresses is thus to be found in the convergence of Africa and Eurasia in this case, but the variations through time of the stress regime within the foreland is also a consequence of the intrinsic mechanical behavior of the weak European lithosphere.

A look at the larger picture moreover reveals a parallel evolution on a much larger scale (Jolivet *et al.*, 2016a; Mouthereau *et al.*, 2021) (Fig. 9). The Early Cretaceous rifting observed in the Pyrénées, the Bay of Biscay and within Iberia is part of a wide extensional domain that encompasses a large part of Africa and Western Europe, a series of rifts forming at this period within much of the northern half of Africa while the Southern Atlantic Ocean opens. The extended domain includes Iberia and Northern Europe until the North Sea. On the other hand, the Late Cretaceous compression is also recorded on a much wider realm, from the inverted rifts in Africa to the North Sea. It starts in Western Europe as soon as the Late Turonian and Coniacian (Ziegler, 1990; Voigt *et al.*, 2006; Lasseur, 2007; Grosheny *et al.*, 2015) thus before the initiation of shortening in the Pyrénées. It leads to the obduction of oceanic lithosphere on the continental margins of Apulia (Izmir-Ankara suture zone) and Africa (Oman ophiolite). This period culminates with the so-called Santonian compression that is felt all over Africa (Guiraud *et al.*, 2005). The Eocene is also a period of compression over a wide region from the High Atlas to the Rif-Betic arc, the Pyrénées, the Alps and further north, before the subduction regime changed in the Mediterranean and back-arc basins started to form. The relatively slow Africa-Eurasia convergence during the Paleocene (Rosenbaum *et al.*, 2002a, 2002b) (Fig. 12) could also partly explain the low recorded deformation at this period. Moreover, the effect of the establishment of the Icelandic plume around 64 Ma on the stress regime in Western Europe (White and McKenzie, 1989; Nadin *et al.*, 1997; Nielsen *et al.*, 2002) should also be looked at.

4.3 Plate motion/deformation and convection

At large scale, the respective roles of the lithosphere as a stress guide and of the convective mantle underneath are still to be explored. The long-term evolution of the Paleo-Tethys and Neo-Tethys shows the repetition of a similar scenario with large continental blocks detaching from the main southern continent and migrating northward across the Tethys to finally collide with Eurasia (Fig. 15) (Jolivet *et al.*, 2016a). This same scenario is repeated from the Devonian to the Present, the most recent of such blocks being Arabia (Jolivet and Faccenna, 2000; Bellahsen *et al.*, 2003; Faccenna *et al.*, 2013b). This behavior, typical of the Tethyan E-W oceanic domain, signs the permanence of a similar engine at the scale of the mantle and it emphasizes the cooperative work of subduction zones (Tethyan subductions) and large-scale plumes (Vaughan and Scarrow, 2003; Moucha and Forte, 2011; Glisovic *et al.*, 2012; Koptev *et al.*, 2019).

The large-scale convection is strongly coupled with the motion of plates at the surface and thus the Africa-Eurasia relative motion. When a plate is attached to a subducting slab, it is powered by the slab pull force and when it is not attached to a slab it is powered by the mantle flowing underneath (Coltice *et al.*, 2019) (Fig. 1). The Tethyan context is, however, complex because the shapes of margins are not simple and collision does not happen everywhere at the same period (Dercourt *et al.*, 1986, Dercourt *et al.*, 1993; Jolivet and Faccenna, 2000; Bellahsen *et al.*, 2003), thus creating lateral variations of the slab pull force along the plate boundary and slowing down portions of continents that are then more easily coupled to the flow of mantle underneath.

The observed contemporaneity of compressional and extension periods over a wide domain encompassing a large part of Africa and western Europe has been explained by changes in the convection pattern with variations of the intensity of upwellings below South and West Africa and variations of the subduction regime below the northern margin of the Neo-Tethys Ocean (Jolivet *et al.*, 2016a). In this model, the Late Cretaceous compressional period recorded in the Pyrénées and the obduction in Oman are the result of the penetration of the Tethyan slab through the boundary between the upper and lower mantle at a time of faster mantle upwelling in the south.

4.4 Subduction-related metamorphism and subduction dynamics

From the Late Cretaceous to the Eocene, H*P*-L*T* metamorphism has been recorded in most mountain belts except in the Pyrénées where syn-rift H*T*-L*P* metamorphism is instead observed (Ravier, 1959; Goldberg and Leyreloup, 1990; Clerc *et al.*, 2015b; Ducoux *et al.*, 2019). The absence of blueschists and eclogites in the Pyrénées can be due to the absence of a true subduction (Chevrot *et al.*, 2014). In the Alps, the internal Apennines (Tuscan Archipelago) and the Betics instead, blueschist and eclogite-facies metamorphic overprints are recorded in large oceanic and/or continental units (Chopin, 1984; Goffé and Chopin, 1986; Goffé *et al.*, 1989; Bousquet *et al.*, 1997; Rossetti *et al.*, 1999a, Rossetti *et al.*, 1999b; Oberhänsli *et al.*, 2004; Bousquet *et al.*, 2008; Angiboust *et al.*, 2009; Bianco *et al.*, 2019; Agard, 2021). The oldest H*P*-L*T* metamorphism is recorded in the Eastern Alps with the Koralpe eclogites in the late Early Cretaceous (~100–110 Ma, Miller and Thöni, 1997). Otherwise, most of the blueschists and eclogites were formed in the Eocene (Oberhänsli *et al.*, 2004; Bousquet *et al.*, 2008). Ultra-high-pressure metamorphism is found in the Alps (Chopin, 1984) and in the Edough Massif (Algeria) (Bruguier *et al.*, 2017). Available ages attributed to the peak of pressure along a given transect vary between ~45 Ma and 34–38 Ma and are nowhere younger than 34 Ma (Duchêne *et al.*, 1997b; Rubatto *et al.*, 1997; Vitale Brovarone and Herwartz, 2013; Bruguier *et al.*, 2017; Bessière, 2019; Bessière *et al.*, 2021; Angiboust and Glodny, 2020, see also a synthesis in Agard, 2021). It is not always clear whether these ages really correspond to the peak of pressure, but they are nevertheless the youngest ages obtained from these tectonic units. In the Betics, the H*P*-L*T* metamorphism of the Alpujarride have been recently dated around 38 Ma for the eastern and central parts (Bessière, 2019; Bessière *et al.*, 2021). In the Nevado-Filabride also, 40 Ma seems the age of the peak of pressure (Li and Massonne, 2018; Bessière *et al.*, 2021). 34 Ma correspond to the very end of the compressional period in Alpine Corsica (Beaudoin *et al.*, 2020) and inception of rifting in the Liguro-Provençal basin and it falls just before the drastic change in the shortening regime in the French Alps with the first involvement of the External Crystalline Massifs and the flysch-to-molasse transition, associated with a retreat of the European slab (Vignaroli *et al.*, 2009). We thus see a relation between the change of subduction regime at about 35–32 Ma in the Western Mediterranean and the end of H*P*-L*T* metamorphism, especially the fast exhumation of the UH*P* units in the Alps and Edough Massif.

5 *Mediterranean Tectonics*, between 32 and 8 Ma

We divide this period of fast slab retreat in two episodes, before and after the inception of several major slab tears around 20–15 Ma (Fig. 12).

5.1 From 32 to 20–15 Ma

This first period sees the opening of the Liguro-Provençal Basin and fast rotation of Corsica and Sardinia and the first stages of formation of the Alboran Sea at the expense of the Internal Zones of the Betics and Rif (Fig. 10), see Romagny *et al.* (2020) for detailed reconstructions. The Pyrenean orogenic wedge continues to form above the south Pyrenean front, while its eastern part is dismantled by rifting in the Gulf of Lion and leaves place to the passive margin (Jolivet *et al.*, 2020). The extensional deformation reworking former mountain belts during slab retreat has been described previously; it is characterized by low-angle normal faults and associated basins. They are observed in the Betic Cordillera, in Calabria, in Tuscany and the Tuscan archipelago (Elba, Monte Cristo, Giglio, Gorgona) as well as in Corsica (Platt and Vissers, 1989; Jolivet *et al.*, 1990; Keller and Pialli, 1990; Jolivet *et al.*, 1991; Crespo-Blanc *et al.*, 1994; Crespo-Blanc, 1995; Martinez-Martinez and Azañon, 1997; Jolivet *et al.*, 1998; Rossetti *et al.*, 2001; Platt *et al.*, 2003a; Collettini and Holdsworth, 2004; Rossetti *et al.*, 2004; Platt *et al.*, 2013; Beaudoin *et al.*, 2017). They are also observed within the Apennines with active normal faults with low-angle geometry (Collettini and Barchi, 2002; Brogi *et al.*, 2003; Collettini and Barchi, 2004; Pauselli *et al.*, 2006; Pauselli and Ranalli, 2017). A coeval migration of extension and magmatism is observed from Corsica toward the Apennines from the Miocene to the Present with east-dipping low-angle normal faults and exhumation of metamorphic core complexes (Jolivet *et al.*, 1998). Present-day extension is active in the Apennines west of the water divide.

5.1.1 Pre-drift position of Sardinia

One still pending question is the pre-rift fit of Sardinia (Fig. 16). It is important for the understanding of the dynamics of rifting and also for the pre-rift kinematic restoration of the Pyrénées and their connection with the Provence fold-and-thrust belt. The sharp bend of the North Pyrenean Thrust front across the Corbières may suggest that the entire crustal wedge was also bent, thus questioning the nature of the pre-rift basement below the Gulf of Lion passive margin. The continuity of the Pyrénées toward the east south of Provence before the Oligocene rifting has been discussed for a long time (Gorini *et al.*, 1994; Mascle *et al.*, 1994; Séranne *et al.*, 1995; Séranne, 1999; Tavani *et al.*, 2018). It is classically thought that a significant relief of Paleozoic basement existed south of Provence that was the backstop of the Provence fold-and-thrust belt. The main arguments (Arthaud and Séguret, 1981; Gorini *et al.*, 1994; Guennoc *et al.*, 2000; Bestani *et al.*, 2015; Espurt *et al.*, 2019) revolve around the presence of Pyrenean thrusts within the pre-rift basement on offshore seismic profiles as well as onland (Cap Sicié), the absence of any significant pre-rift sedimentary cover on top of the basement on seismic profiles in the Gulf of Lion, suggesting a period of aerial erosion, the pattern of paleocurrents showing a source of detritus south of Languedoc (BRGM *et al.*, 1974; Christophoul *et al.*, 2003; Vacherat *et al.*, 2017) and the presence of Late Cretaceous conglomerate with a southern provenance in the southern part of Provence (Hennuy, 2003). The detailed offshore study of Fournier *et al.* (2016) shows the regional

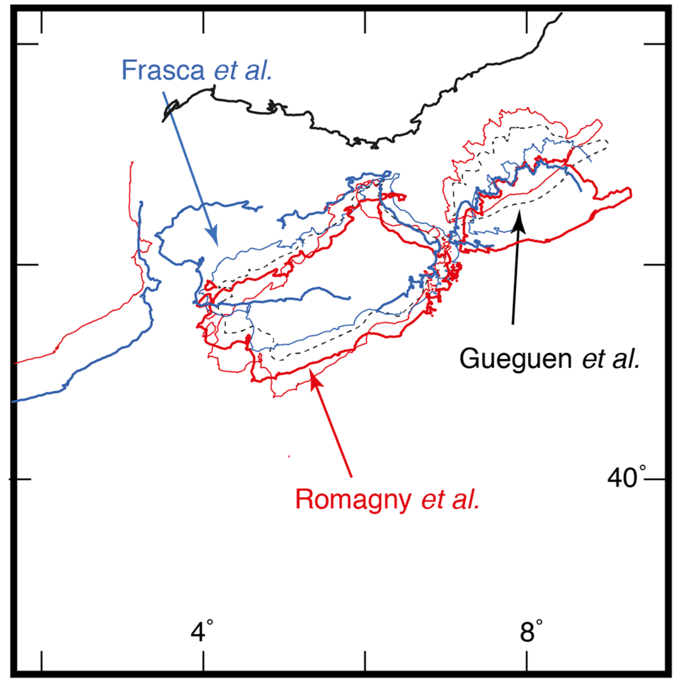

Fig. 16. Different options for the pre-rift fit of Sardinia before the rifting of the Gulf of Lion and opening of the Liguro-Provençal basin, after Romagny *et al.* (2020), Frasca *et al.* (2017) and Gueguen *et al.* (1998).

extension of the Cap Sicié Pyrenean thrust carrying a large Paleozoic basement unit and its Mesozoic cover, which could be the source of the detrital sediments shed over Languedoc with an eastern provenance. This conclusion is consistent with the findings of Ternois *et al.* (2019) who document the uplift of the Agly massif during the Late Cretaceous.

Before the rotation of Corsica and Sardinia, the Provence fold-and-thrust belt was the foreland of the Alpine Corsica accretionary wedge (Vially and Tremolières, 1996; Lacombe and Jolivet, 2005; Bestani *et al.*, 2015). This question has been recently addressed by Romagny *et al.* (2020). If one considers Corsica-Sardinia as a rigid block following the earlier works of Arthaud & Matte (1977), rotating it back to its pre-rift position leaves a "hole" at the emplacement of the Gulf of Lion. The 23 Ma reconstruction of Gueguen *et al.* (1998) shows the situation at the end of rifting before the fast rotation and formation of oceanic crust (Fig. 16). The space left between Sardinia and the Languedoc coastline corresponds to the rifted crust underlying the Gulf of Lion margin and the amount of extension during rifting is not reconstructed. The solution chosen by Romagny *et al.* (2020) is to rotate back the rigid Corsica-Sardinia block so that the crust in the Gulf of Lion returns to a thickness of 30 km, which corresponds to the present-day thickness below the coastline, based on the balanced cross-section of Jolivet *et al.* (2015a). Angrand *et al.* (2020) have adopted a similar reconstruction of the pre-drift position of Sardinia. Ford *et al.* (2020) have chosen an initial thickness of 40 km instead and they must then rotate more Sardinia than Corsica. They thus divided Sardinia in two blocks moving along a sinistral strike-slip fault during rifting. The two solutions are consistent with the possibility of a

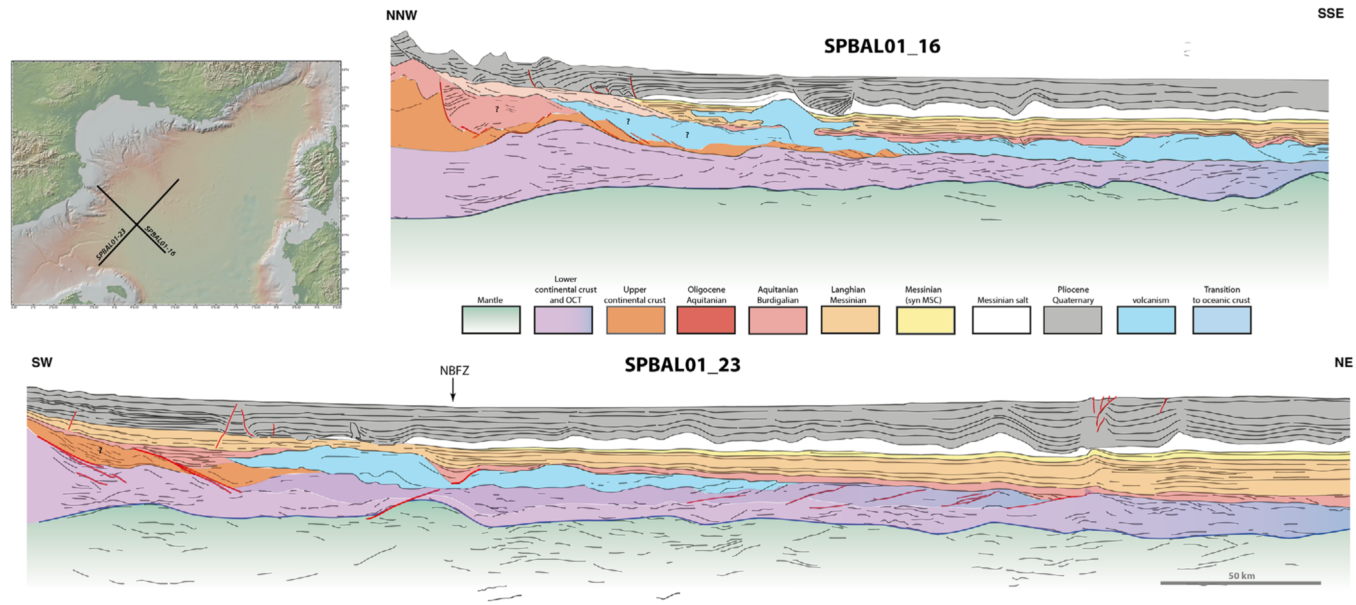

Fig. 17. Two interpreted seismic lines across the transition zone between the Valencia basin and the Gulf of Lion, after Jolivet *et al.* (2020) and Maillard *et al.* (2020).

subaerial erosion before rifting or at the time of rifting, but Ford *et al.* (2020) keep a thicker crust during a longer period. The difference in rotation angle is small and keeps within the error bars of paleomagnetic data and the additional motion of southern Sardinia does not involve much rotation. The accretionary wedge south of Provence in the Eocene involved Alpine and Variscan Corsica (Lacombe and Jolivet, 2005) but the thickness of the crust at that time is unknown. The absence of a thick foreland basin in Provence is not consistent with a thick crust. Both solutions are, however, consistent with the available data set and solve the solution of the "hole" at the emplacement of the Gulf of Lion but it is difficult to assess which fit is the best (see also Advokaat *et al.*, 2014).

5.1.2 Depth-dependent extension during back-arc rifting

The question of the amount of extension during the rifting of the Gulf of Lion has long been debated. Bessis (1986) and Burrus (1989) observed that the amount of thermal subsidence during the post-rift period is compatible with a larger amount of stretching than shown by the moderate extension deduced from normal faults seen on seismic profiles. They also showed that fast thermal subsidence started very early after the end of rifting. Later studies based on seismic reflection profiles such as the ECORS experiment confirmed the intense thinning, and refraction data showed that the distal part of the margin is characterized by an abnormal crust with higher seismic velocities than normal continental crust, without the typical pattern of oceanic crust (Pascal *et al.*, 1993; Chamot-Rooke *et al.*, 1999; Gailler *et al.*, 2009; Bache *et al.*, 2010; Moulin *et al.*, 2015). Based on the interpretation of wide-angle seismic profiles across the Liguro-Provençal Basin (Sardinia cruise 2006), Gailler *et al.* (2009) discussed the nature of the crust below the distal margin with two opposed hypotheses, either lower continental crustal material or a mixture of serpentinized mantle with lower crustal material. Bache *et al.* (2010) confirm that the amount of extension deduced from the observation of upper crustal normal faults can account for only a part of the finite stretching. Jolivet *et al.* (2012, 2015a) used an industrial seismic profile across the Gulf of Lion passive margin to argue in favor of exhumed lower continental crust and upper mantle from below the upper crust by the activity of low-angle detachments dipping toward the continent (Fig. 11). Based on further analyses of the Sardinia experiment, Moulin *et al.* (2015) and Afilhado *et al.* (2015) also argued in favor of exhumed lower crustal material for the distal margin. The presence of this exhumed lower crustal material on both the Provence and Sardinia sides led them to conclude that an asymmetric model with a single detachment is unlikely.

Additional information recently came from the analysis of a series of reflection profiles across the eastern part of the Valencia Basin in the transition zone with the Gulf of Lion. Granado *et al.* (2016) also interpreted the distal margin as made of lower crustal material extracted from below the margin by low-angle detachments. More recently, Jolivet *et al.* (2020) and Maillard *et al.* (2020) showed that the eastern part of the Valencia Basin is characterized by ductilely stretched lower continental crust with evidence of low-angle shear zones and covered with a thick pile of volcanic material emplaced during rifting (Fig. 17). The presence of this volcanic province had already been proposed earlier by Mauffret *et al.* (1995) and Maillard and Mauffret (1999). It makes a striking difference with the Gulf of Lion margin where little volcanic material is observed. Canva *et al.* (2020), however, showed that the prominent Catalan magnetic anomaly is best explained by the underplating of mafic material (gabbros) underneath the crust in the vicinity of the Catalan transfer zone (Figs. 18 and 19). The eastern Valencia Basin and the Gulf of Lion are thus both characterized by the exhumation of ductile lower crust during

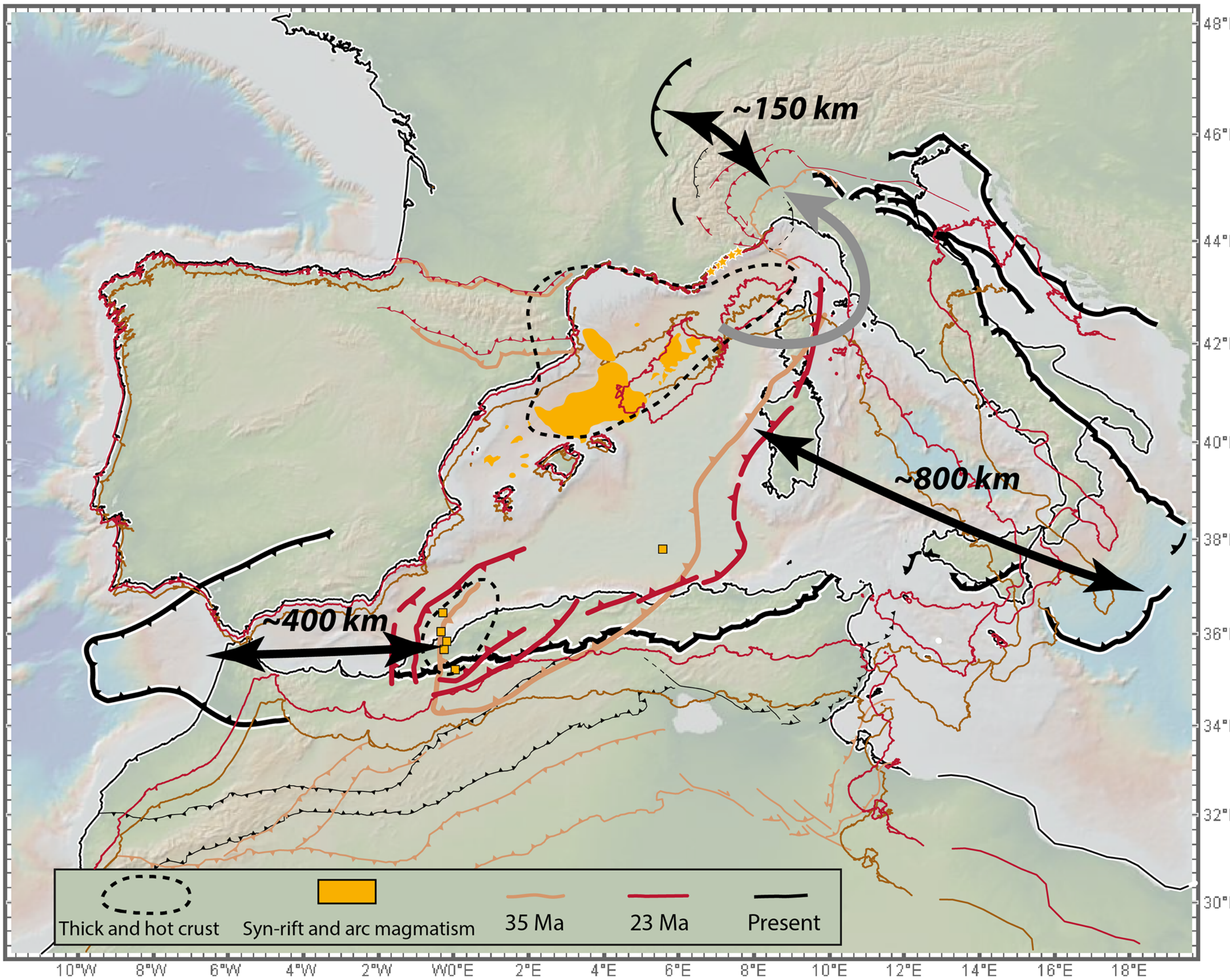

Fig. 18. Comparison of the positions of the subduction zone at 35 Ma, 23 Ma and present and the estimated displacements since 35 Ma. Orange color represents the extent of syn-rift volcanism, including the orange arrows along the southern coast of France that stand for the adakitic outcrops dated from the latest Eocene and early Oligocene (Réhault *et al.*, 2012; Jolivet *et al.*, 2020; Maillard *et al.*, 2020). Thick dashed lines show the limits of thick and hot crust at the time of rifting. The grey curved arrow symbolizes the toroidal flow due to the retreat of the Apennines slab (Vignaroli *et al.*, 2009).

rifting with the addition of intense volcanic activity in the Valencia Basin. The eastern Valencia Basin margin and the Gulf of Lion thus fall in the group of hot margins in the sense of Clerc *et al.* (2015a, 2016, 2017) and Jolivet *et al.* (2018), with differential extension in the upper and lower crust and shearing along the crust-mantle interface.

These recent findings led Jolivet *et al.* (2020) to reinterpret (Fig. 11) the significance of the recent receiver-function experiment across the Pyrénées (Chevrot *et al.*, 2018; Diaz *et al.*, 2018). A series of 5 profiles across the strike of the belts was recorded from west to east with the addition of a longitudinal profile in the Eastern Pyrénées. The three western profiles show a clear "orogenic" structure with the underthrusting of the Iberian crust below the Pyrénées and a very shallow Moho below the North Pyrenean Zone, especially below the Mauleon-Arzacq Basin way above the Iberian Moho. The shallow Moho is interpreted by Chevrot *et al.* (2018) as an inheritance from the Early Cretaceous rifting event, the thin crust being passively transported by Pyrenean thrusts, which is confirmed by the results of the Maupasacq passive seismic experiment (Lehujeur *et al.*, 2021). Toward the east, the situation changes drastically and this orogenic structure is lost with a shallower Moho and no clear image of the underthrusting of Iberia underneath the chain. Chevrot *et al.* (2018) interpreted this difference as inherited from the pre-orogenic template, namely two different geometries of the early Cretaceous rift. A parallel study showed that this region is characterized by an anomalous lithosphere that has lost its lower crust (Wehr *et al.*, 2018). No observation is available to constrain the timing of the removal of the lower crust. An alternative interpretation of the receiver-function profiles was then proposed by Jolivet *et al.* (2020). The missing lower crust below the eastern Pyrénées would have been extracted from below the belt and juxtaposed with the distal part of the Gulf of

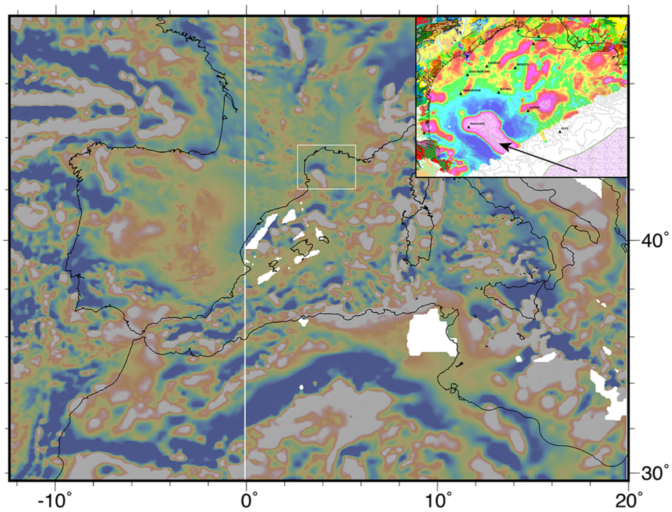

Fig. 19. Magnetic map of the Western Mediterranean (EMAG2– 2013 –made from GeoMapApp; [Ryan *et al.*, 2009]) and a zoom on the vertical gradient of the magnetic anomalies (reduced to the pole) in the Gulf of Lion, after Canva *et al.* (2020).

Lion margin (Fig. 11). The change from west to east of the deep structure of the Pyrénées would then be the result of post-orogenic extension during the formation of the Gulf of Lion margin, instead of an heritage from the early Cretaceous rifting stage. This scenario has the additional advantage to explain the fast cooling episode recorded in the Eastern Pyrénées around 30 Ma at the time of rifting based on low-temperature thermochronology data (Morris *et al.*, 1998; Fitzgerald *et al.*, 1999; Waldner *et al.*, 2019; Daril, 2018; Bernard *et al.*, 2019; 2020). More precisely Gunnell *et al.* (2009) constrain this episode between 25 and 35 Ma and Maurel *et al.* (2002) around 26–27 Ma. Milesi (2020) also documents a period of exhumation after 35 Ma and the exhumation of the footwall of the Têt fault after 30 Ma. The removal of the upper mantle from below the belt led to an uplift and the subsequent exhumation of the lower crust led to subsidence, which is recorded by the erosional unconformity seen on seismic profiles below the post-rift deposit (Bache *et al.*, 2010; Jolivet *et al.* (2015a, 2015b)). The removal of upper mantle can be driven by basal shearing and it can also be a thermal consequence of slab retreat. The presence of a prominent erosion surface below the Aquitanian marine deposits contemporaneous of hyper-extension and final mantle exhumation (Bache *et al.*, 2010; Jolivet *et al.*, 2015a) shows that the entire margin was in subaerial conditions during rifting and was thus supported by a thin and hot lithosphere (Fig. 18). In addition, density reduction resulting from serpentinisation could help maintaining the surface of the lithosphere close to sea level (Huyghe *et al.*, 2020). The Aquitaine foreland basin has recorded this transition toward post-orogenic extension with a hiatus and an erosional unconformity between the Chattian and the transgressive Aquitanian, that could be coeval with the erosional surface seen on seismic profiles of the Gulf of Lion margin (Ford *et al.*, 2016; Ortiz, 2019; Calvet *et al.*, 2021; Ortiz *et al.*, 2020). Low-T thermochronology data support the exhumation of the Eastern Pyrénées during the Late Oligocene (Fitzgerald *et al.*, 1999; Maurel *et al.*, 2002; Daril, 2018; Milesi, 2020) and the formation of planation surfaces, now at high elevation until the Late Miocene, also possible effects of mantle dynamics and/or magmatism during the early stages of extension (Gunnell *et al.*, 2009; Monod *et al.*, 2016; Calvet *et al.*, 2021; Huyghe *et al.*, 2020).

The main engine of crustal deformation during this period is the retreat of the slab subducting below the AlKaPeCa blocks and Corsica-Sardinia at the back. Finite displacements of retreating arcs amount to 800 km during this time interval, which implies that the asthenospheric mantle below the back-arc region has also migrated by the same amount, thus inducing a significant flow below the extending lithosphere. One consequence is that the effects of slab retreat must be felt far away from the retreating trench in the back-arc region (Barruol and Granet, 2002; Lucente *et al.*, 2006; Jolivet *et al.*, 2009, 2020).

5.1.3 Transfer zones

The large rigid rotation of the Corsica-Sardinia block and the emplacement of oceanic crust and/or exhumed mantle in the Liguro-Provençal Basin implies the existence of large transfer zones accommodating differential rotation and extension between the Gulf of Lion and the eastern Pyrénées coastline and further south the Valencia Basin and Minorca (Figs. 10 and 14) (Maillard and Mauffret, 1999, Maillard and Mauffret, 2013; Pellen *et al.*, 2016). Such transfer faults have been drawn on maps for kinematic reasons, but they had so far not been characterized by any specific structure, except for their link with an intense volcanic activity in the Valencia Basin (Maillard and Mauffret, 1999). Recent studies (Canva *et al.*, 2020; Maillard *et al.*, 2020) with new seismic lines and the study of magnetic anomalies show these transfer zones with more details and the kinematic reconstructions (Romagny *et al.*, 2020) show the different domains separated by the transfer zones (Fig. 14) (see also Jolivet *et al.*, 2021, for a synthesis of these transfer zones in the whole Mediterranean realm and Ford *et al.*, 2021 for detailed reconstructions of the junction between the Pyrénées and Languedoc). The north-eastern part of the Valencia Basin until the transition with the Gulf of Lion shows a thick sequence of volcanic rocks mainly emplaced during the rifting and also afterward until the Messinian (Maillard *et al.*, 2020). This volcanic sequence rests on top of exhumed lower crust showing evidence for low-angle shear zones and ductile deformation. Some specific features are observed when crossing the putative transfer zones such as localized grabens and their sedimentary infill, as well as lower crust and mantle domes. These features show a transtensional deformation with a significant extensional component associated with the dextral motion during rifting and drifting. The Catalan magnetic anomaly shows a trend parallel to the Catalan Transfer Zone and the work of Canva *et al.* (2020) (Fig. 19) has shown that it can be modelled with the underplating of mafic rocks (gabbros) along the trace of the anomaly. Dextral motion with a significant extensional component and magmatic additions at depth thus characterize these transfer zones.

5.1.4 Interactions of slabs at depth

During this period, the Alboran domain was located eastward of its present position and the Alpujarride metamorphic complex was being exhumed from below the Malaguide-Alpujarride Contact (MAC), a large-scale detachment with a top-to-the NE kinematics (Lonergan and Platt, 1995; Platt *et al.*, 2005; 2013), until the docking of the AlKaPeCa block with the northern margin of Africa. H*P*-L*T* metamorphic units of the Alpujarride were exhumed contemporaneously all the main blueschists units around the Western Mediterranean, including the Alps, Alpine Corsica, Calabria and the Kabylies. 38–34 Ma is the time of the last peak pressure recorded in all these H*P*-L*T* terranes (see above) (Fig. 20). Afterward, lower-pressure blueschists are still recorded east of Corsica in the early Miocene in the islands of Gorgona or Elba during the eastward retreat of the slab (Jolivet *et al.*, 1998; Rossetti *et al.*, 1999a, Rossetti *et al.*, 1999b, Rossetti *et al.*, 2004; Bianco *et al.*, 2019).

The case of the Betics requires some discussion. The age of the peak pressure conditions in the Nevado-Filabride Complex is debated. Originally considered Eocene like in the Alpujarride (Monié *et al.*, 1991; Augier *et al.*, 2005a, Augier *et al.*, 2005b), it was then assigned to the Early Miocene (until 16–17 Ma) based on dates obtained with the Lu-Hf method on garnet (Platt *et al.*, 2006), the U-Pb method on zircons (López Sánchez-Vizcaíno *et al.*, 2001; Gómez-Pugnaire *et al.*, 2012) or Rb-Sr isochrons (Kirchner *et al.*, 2015). More recently, however, U-Pb ages around 40 Ma were obtained on monazite (Li and Massonne, 2018). Bessière *et al.* (2021) then consider that these Early Miocene ages correspond to the late exhumation of these units and that the peak of pressure was indeed Eocene. The end of exhumation of well-preserved H*P* or UH*P* units around 38–34 Ma thus occurred just before or at the time of inception of back-arc rifting in the Gulf of Lion and Alboran Sea and the beginning of slab retreat. Afterward, only lower pressure conditions were attained with locally H*P*-L*T* conditions such as in the Tuscan archipelago (Jolivet *et al.*, 1998; Rossetti *et al.*, 1999a, 1999b; Bianco *et al.*, 2015, 2019). This suggests that the inception of slab retreat changed the dynamics of subduction channels where these H*P*-L*T* metamorphic rocks were formed (Jolivet *et al.*, 2003). This is quite straightforward for the areas directly impacted by back-arc extension, but more surprising for the Alps where this period instead corresponds to the westward propagation of the thrust front and the westward overthrusting of the External Crystalline Massifs at around 32 Ma, before the stacking of the Subalpine domain in the middle Miocene (Ford *et al.*, 2006; Bellahsen *et al.*, 2014; Bellanger *et al.*, 2015). It also corresponds to an evolution of the flexural basin with the flysch-to-molasse transition and an increase of the volume of sediments shed in these basins (Kuhlemann *et al.*, 2002; Kuhlemann and Kempf, 2002).

These coeval evolutions of the back-arc domain and of the Alps is symptomatic of the possible interactions of the Alpine and Apennines slabs at depth. As proposed by Vignaroli *et al.* (2008; 2009), the eastward retreat of the Apennines slab might have forced a counter clockwise toroidal flow of the asthenospheric mantle trapped beneath the Apenninic slab (Figs. 14 and 18), which has then pushed the east-dipping European slab westward, thus opening the subduction channel and favoring the exhumation of H*P*-L*T* metamorphic rocks and accelerating the rate of subduction in the external zones of the French-Italian Alps. The change of subduction regime would then have induced changes in the dynamics of mountain belts all around the Western Mediterranean.

While the Alps, the Eastern Pyrénées or the Betics show sharp changes at around 35–32 Ma, the Western and Central Pyrénées simply record continuing southward propagation of thrusts in the Ebro foreland basin until the early Miocene (Jolivet *et al.*, 2007; Mouthereau *et al.*, 2014; Bosch *et al.*, 2016; Labaume *et al.*, 2016; Teixell *et al.*, 2018;). The peak of exhumation recorded in the Eastern Pyrénées is around 30 Ma (Daril, 2018). It is younger in the west, around 25–20 Ma with also a more recent deformation in the south (Bosch *et al.*, 2016). The central Pyrénées would show a continuous sequence of deformation (Mouthereau *et al.*, 2014). This longitudinal evolution of the Pyrénées could possibly be also a consequence of the mantle flow due to slab retreat underneath that would have eroded the western part only after the eastern part, but the geometry of the belt at depth in the center and west does not comfort this hypothesis. The effect of rifting in the Gulf of Lion was thus seemingly not felt in the western part of the belt where compressional stresses transferred through the lithospheric stress-guide from the Africa-Iberia plate boundary were predominant. This suggests a maximum distance in the influence of slab retreat and associated asthenospheric flow on the overriding plate.

During this period, slab retreat was the primary engine of crustal deformation in a wide domain including the immediate back-arc regions and their hinterlands, including part of the Eastern Pyrénées and also the Alps, which dynamics was strongly modified by the interactions between slabs.

5.2 Between 20–15 and 8 Ma

During this later period, slab tearing goes on with a focalization on two narrower slabs (one retreating westward – Gibraltar, one eastward – southern Tyrrhenian Sea) after an episode of tearing following the collision of AlKaPeCa with the northern margin of Africa (Figs. 10 and 13). An additional ingredient should be considered to explain the change in the dynamics of subduction around 15 Ma. Based on analogue experiments, Faccenna *et al.* (2001a, 2003) suggested that the interaction of the slab with the upper-lower mantle transition zone had slowed down the retreat during a short period, stopping the opening of the Liguro-Provençal and rotation of the Corsica-Sardinia block and initiating the rifting of the Tyrrhenian Sea. An alternative could be that the collision of AlKaPeCa with Africa had stopped the rotation of Sardinia. The width of the Apennines slab progressively narrows through time and the velocity of retreat consequently increases (Carminati *et al.*, 1998a, 1998b). The most recent slab is indeed very narrow and the velocity of retreat was high (10 cm/yr in average) during the Pliocene (Royden *et al.*, 1987; Patacca and Scandone, 1989; Patacca *et al.*, 1990; Faccenna *et al.*, 2005; Guillaume *et al.*, 2010).

The western part of the torn slab retreats westward, forming the Alboran Sea and transporting the West Alboran Basin (Crespo-Blanc *et al.*, 2016; Do Couto *et al.*, 2016). This westward motion of the Alboran domain is limited by two

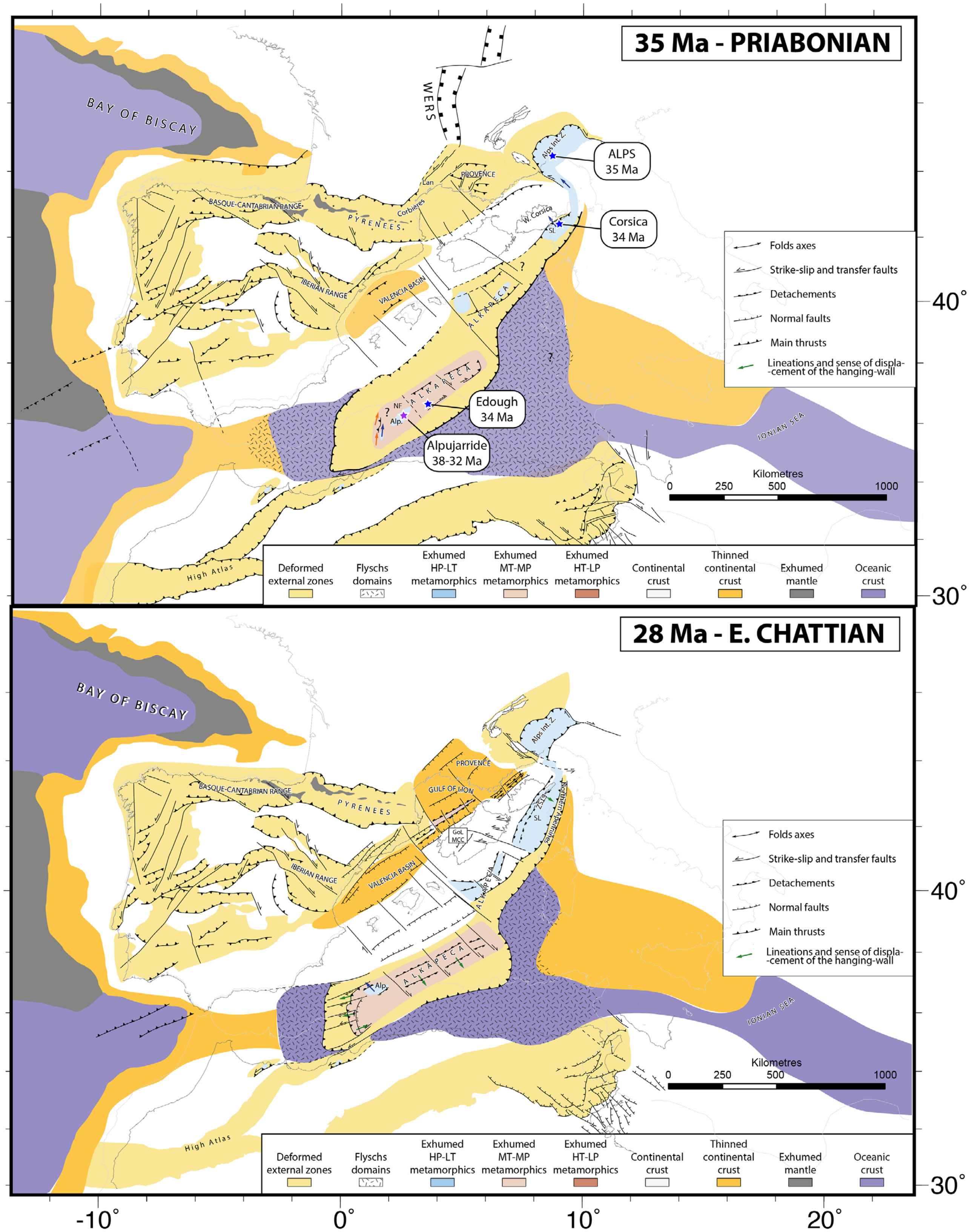

Fig. 20. Reconstructions (Romagny et al., 2020) at 35 Ma and 28 Ma of the Western Mediterranean at the transition between compression and extension in the Gulf of Lion region and right after the beginning of back-arc extension. Ages of the UHP-LT and HP-LT peak pressure conditions are shown on the 35 Ma stage, after Gebauer et al. (1997), Rubatto et al. (1997), Duchêne et al. (1997a), Vitale-Brovarone and Herwartz (2013), Bruguier et al. (2017) and Bessière (2019).

transfer zones where the relative motion is mostly strike-slip, dextral on the Betic side and sinistral on the Rif side (see also Romagny et al., 2020; Jolivet et al., 2021). Retreat of the Alboran slab proceeds along the northern margin of Africa and the strike-slip component is distributed through wide transfer zones on the northern and southern margins of the Alboran basin. This strike-slip component is accommodated by E-W trending strike-slip faults but mostly by E-W extension on low-angle normal faults and ductile detachments, the largest of which being the Filabres Shear Zone that has exhumed the Sierra Nevada-Sierra de los Filabres MCC (Jabaloy et al., 1993; Augier et al., 2005a). These domes are elongated parallel to the main direction of shearing, making them a-type domes (Jolivet et al., 2004) that can be modelled numerically in 3-D within transtensional shear zones (Le Pourhiet et al., 2012).

In the same period, oceanic crust was emplaced in the Algerian Basin (Mauffret et al., 2004; Booth-Rea et al., 2007; Driussi et al., 2015) and strong crustal thinning affected the Alboran Basin and the West Alboran Basin formed by passive subsidence above the steeply-dipping slab following its westward migration (Do Couto et al., 2016).

During this period the main engine was again slab retreat which direction was almost perpendicular to the Africa-Eurasia convergence. Because the slab progressively unzipped westward until the longitude of the Gibraltar Strait, this situation might also lead to a decrease of compressional stresses across the Alboran Sea and thus a progressive relaxation of compressional stresses across Iberia and the Pyrénées, the latter entering a post-orogenic stage.

6 *Late-Mediterranean Tectonics*, from 8 Ma to the Present

8 Ma was a major change in the tectonic evolution of the Alboran region with the slowing down of the westward retreat of the slab beneath Gibraltar and a transition from dominant E-W extension to dominant N-S shortening, although a component of NE-SW extension is still felt today west of the Sierra Nevada (Galindo-Zaldıvar et al., 2003; Pérez-Peña et al., 2010). The N-S shortening is accompanied by the formation of large crustal-scale folds in the Eastern Betics that amplify the MCCs exhumed during the preceding period (Sierra Nevada, Sierra Alhamilla, Sierra de Gador) (Weijermars et al., 1985; Sanz de Galdeano and Vera, 1992; Meijninger and Vissers, 2006; Augier et al., 2013; Janowski et al., 2017) and by the initiation of the left-lateral Trans-Alboran Shear Zone and associated magmatism (Hernandez et al., 1987; de Larouzière et al., 1988; Stich et al., 2006; Estrada et al., 2017; Lafosse et al., 2018; d'Acremont et al., 2020; Lafosse et al., 2020) (Fig. 2). This N-S shortening is progressively felt all over the North African margin and it corresponds today to the compressional earthquakes and active faults mapped offshore (Deverchère et al., 2003; Billi et al., 2011; Martínez-García et al., 2017; d'Acremont et al., 2020; Lafosse et al., 2020) (Figs. 3, 4 and 10). The direction of shortening is N-S or NNW-SSE in the south (High Atlas) and the rate of uplift increases after 6 Ma (Frizon de Lamotte et al., 2000; Benaouali-Mebarek et al., 2006; Babault et al., 2008; Lanari et al., 2020a, 2020b) and becomes more NW-SE in the north (France) (Cornet and Burlet, 1992; Dèzes et al., 2004;

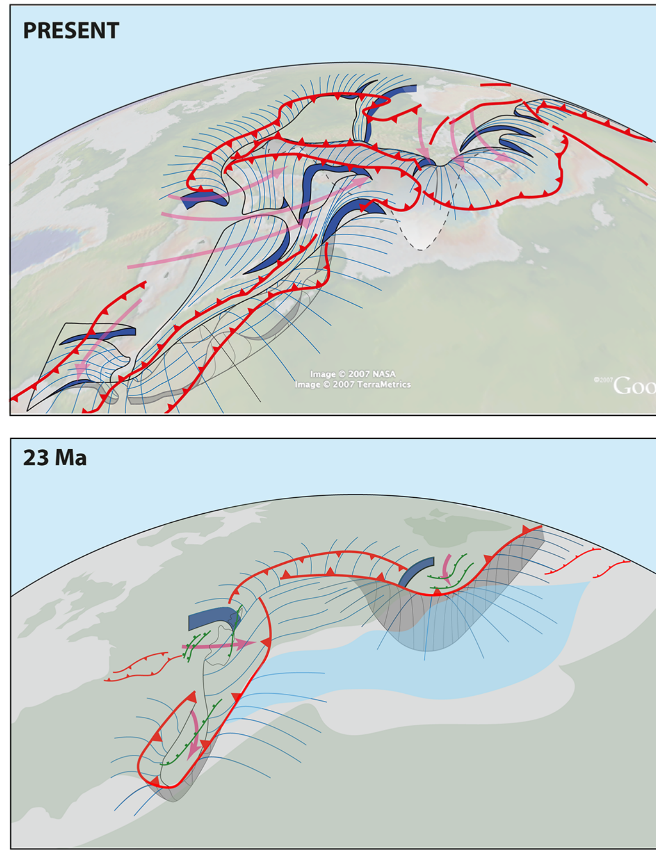

Fig. 21. Two schematic reconstructions in 3-D of the Mediterranean region showing the geometry of slabs during and after the back-arc extension and slab retreat episode, completed from Jolivet et al. (2009).

Baize et al., 2013). Not all the uplift of the Atlas is a consequence of shortening and significance part is due to mantle upwelling there (Babault et al., 2008; Missenard and Cadoux, 2012). Shortening is radial in the frontal zones of the Apennines. Extension remains active with frequent earthquakes in the internal Apennines all the way to the crestline (Amato et al., 1993; D'Agostino et al., 1998) and it is also active in the internal parts of the southern Alps (Sue and Tricart, 1999; Sue et al., 1999; Delacou et al., 2004; Walpersdorf et al., 2018; Sternai et al., 2019). It is also active in the southeast Tyrrhenian Sea with fast slab retreat until the end of the Pliocene (Sartori et al., 2004; Prada et al., 2014, Prada et al., 2018). Whether it is still active today is debated. Zitellini et al. (2019) recently brought observations in favor of a compressional reactivation of the southern Tyrrhenian Sea during the Pliocene, while Gutscher et al. (2017) describe active deformation in the Calabrian accretionary wedge, suggesting that subduction is still active.

Figure 10 shows the evolution of the stress regime in the Western Mediterranean region. The post-8 Ma configuration is similar to the pre-32 Ma one except for the southern Tyrrhenian Sea where compression is more recent (Zitellini et al., 2019). The main changes are recorded in the west where Africa is fully recoupled with northern Europe. The whole Western Mediterranean is now dominated by N-S compression, returning to the pre-32 Ma situation, before slab retreat started.

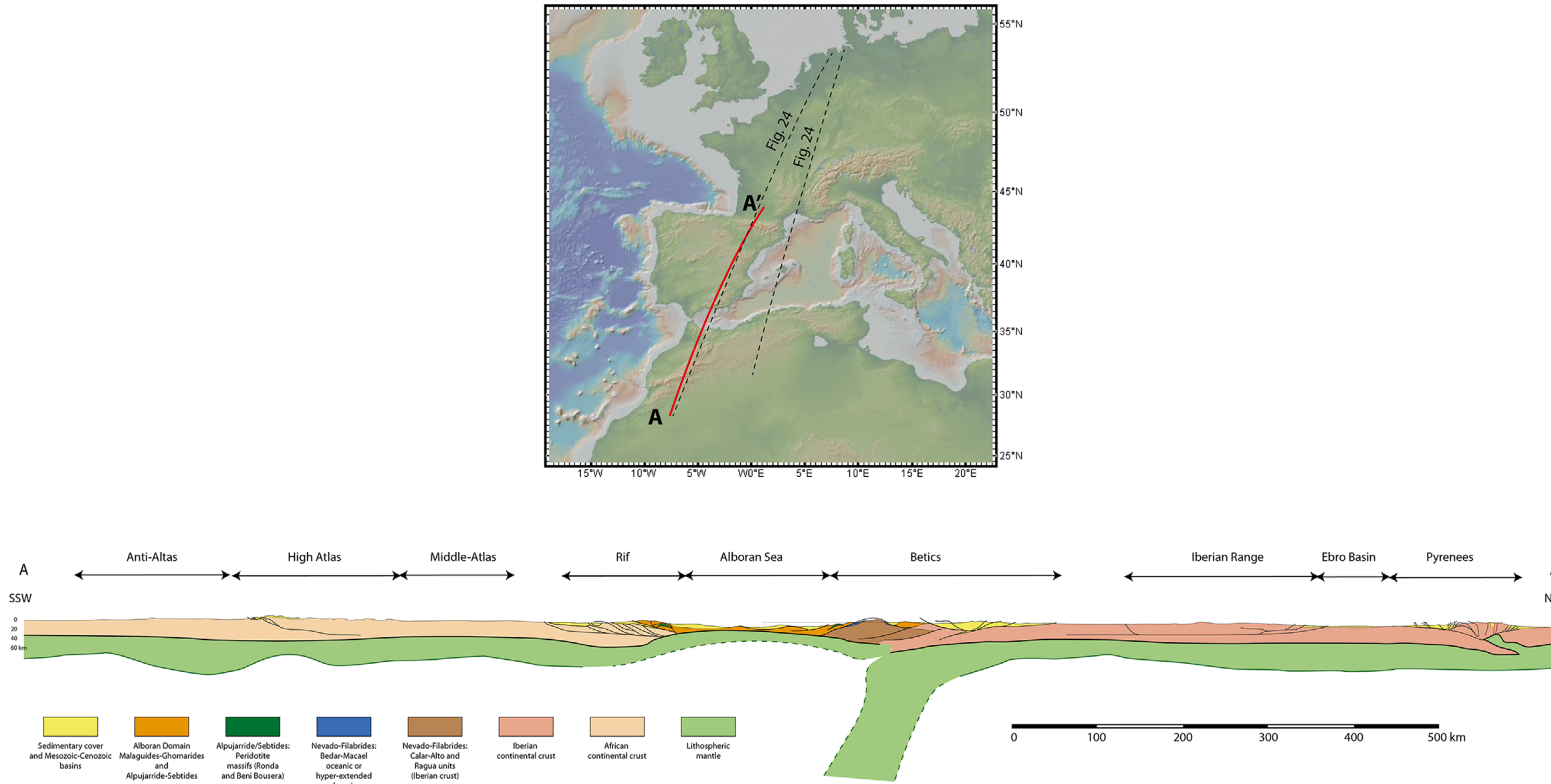

Fig. 22. Lithospheric-scale cross section from the Atlas to the Aquitaine Basin across the Alboran region and the Pyrénées after Zeyen *et al.* (2005), Diaz *et al.* (2016), Mancilla *et al.* (2018), Chevrot *et al.* (2018), Teixell *et al.* (2018), Frizon de Lamotte *et al.* (2000) and Quintana *et al.* (2015).

This renewed compression has been proposed as the trigger for the onset of the Messinian Salinity Crisis as it would have progressively severed the connections between the Atlantic Ocean and the Mediterranean (Jolivet *et al.*, 2006). Spakman *et al.* (2018) and Capella *et al.* (2019) further proposed that the recent evolution is a consequence of the dragging of the slab hanging below the Western Alboran region by the absolute motion of Africa and Eurasia. This proposition is based upon a 3-D numerical modelling of the behavior of the hanging slab (Chertova *et al.*, 2014) and a good fit of the outcome with observations of active displacements and deformation in the Alboran realm. Then, the resumption of N-S shortening is not restricted to the western Alboran region. It instead affects the whole North African margin, the whole of the Atlas from west to east and progressively most of the western Mediterranean except for the Apennines (Frizon de Lamotte *et al.*, 2000, 2008). An additional mechanism must then be looked for (see below).

7 Discussion, far-field versus near-field interactions

We now summarize and discuss further these evolutions in terms of forces, attempting at separating the near-field and far-field interactions, which are due to lithospheric-scale or crustal-scale transmission of stresses *versus* deeper contributions, mainly mantle flow due to large-scale convection and/or slab retreat. Numerical models of mantle convection with lithospheric plates show that slabs play a major role in mantle flow. Coltice *et al.* (2019) show that plates attached to a subducting slab are the fastest and they are mainly powered by slab-pull (Fig. 1). The lithosphere in this case moves faster than the mantle underneath. Plates not attached to a subducting slab are moved and deformed by the mantle flowing faster underneath. In plate convergence areas, slab retreat then becomes a major driver of deformation of the overriding plate. The Mediterranean region during the Mediterranean Tectonics stage is typical of the latter behavior and crustal deformation of back-arc region is driven to a large extent by the flow of mantle due to slab retreat (Fig. 21).

We focus our discussion on several lithospheric sections running from the Atlas Mountains to the North Sea (Figs. 22 and 23). Figure 22 shows a section from the Atlas to the Pyrénées showing crustal and lithospheric thickness variations based the works of Zeyen *et al.* (2005), Diaz *et al.* (2016), de Lis Mancilla *et al.* (2018), Chevrot *et al.* (2018), Teixell *et al.* (2018), Frizon de Lamotte *et al.* (2000, 2008) and Quintana *et al.* (2015). A detailed lithospheric section of the whole Western Europe is also proposed by Mouthereau *et al.* (2021) based on seismic reflection lines and informed by geophysical constraints for lithosphere thickness. Figure 23 is more schematic and includes the section of Figure 22 in a larger framework, all the way to the North Sea with additional data from Cloetingh *et al.* (2009, 2010). These two sections show that the compressional stresses that lead to the formation of the Rif, the Betics, the Iberian Range and the Pyrénées were set on a rather thin lithospheric lid carrying a crust of normal thickness. Oligo-Miocene extension has thinned the crust and the lithosphere in the Alboran Sea and its margins afterward. The thickest lithosphere is found at the extremities of the system, below Northern Europe and North Africa. Compression was felt as far as the North Sea at a distance that is equivalent to that between the Pyrénées and the Atlas.

On Figure 23, we show the different types of forces that may have controlled this tectonic evolution. Compression arising from the convergence between Africa and Eurasia (black arrows) has been transmitted horizontally through the lithospheric stress-guide. In addition to convergence, some more horizontal compression can be provided by the Alpine collision and by the Iceland plume. Crustal thickening has locally modified the lateral distribution of densities, generating body forces (blue arrows) that can lead to crustal spreading. Probably not significant in the Pyrénées, this type of body forces has played an important role in the Miocene evolution of the Alboran region (Platt and Vissers, 1989; Vissers *et al.*, 1995). The slab now dipping underneath the Gibraltar Arc has a complex history of southward-then-westward retreat that has largely controlled the deformation all around the Alboran Sea (green and orange arrows). This slab dynamics includes the asthenospheric flow created by slab retreat (orange arrows). The large-scale convection carrying the large plates northward also played a significant role (red arrows).

We have then constructed a series of cross-sections showing the evolution of this region through time from the early Cretaceous (110 Ma) to the Present (Fig. 24) emphasizing the role of each type of forces. We now discuss the interactions of these different forces (Figs. 9, 10 and 24).

During the Tethyan Tectonic stage, the Western Mediterranean region lies between the the westernmost Tethys (*lato sensu*) and the actively opening Atlantic Ocean and it reacts to external solicitations due to the lithospheric-scale interactions between Africa, Eurasia, Adria/Apulia and Iberia and the interactions with the convecting mantle underneath. From the Late Jurassic to the Early Cretaceous (Tethyan I), the entire domain from West Africa to Northern Europe is under extension. The Pyrenean and Iberian rifts open during this period and Apulia rifts away from Africa in the Late Triassic and early Jurassic. This distributed extension is coeval with the opening of Central Atlantic and from the Lower Cretaceous also the South Atlantic. The African plate is driven both by the slab-pull in the northern Tethys subduction zones and by the plume in the south. Extension is caused simply by the oblique divergence between Africa and Eurasia and/or by the flow of mantle underneath Africa that leads to the separation of Apulia from Africa. This pattern is permanent during the whole history of the Tethys, from the PaleoTethys to the Neo-Tethys and even the present situation (Fig. 15). The long-term presence of a plume above the Tuzo large low-shear velocity province (Burke and Torsvik, 2004; Burke *et al.*, 2008) could explain this permanent behavior, typical of the Tethys Oceans across time (Fig. 15).

Similarly, the generalized compression observed in the Late Cretaceous (Tethyan II) from Africa to northern Europe must be due to a large-scale process ultimately leading to the southward obduction of oceanic crust on the northern margins of Apulia (Izmir-Ankara suture zone) and Africa (Oman ophiolite), along several tens of thousands of kilometers, at least from the Aegean region to Oman and probably farther east in the future Himalaya (Jolivet *et al.*, 2016a). This event can simply be a consequence of the convergence between Africa and Eurasia that started some 84 Ma ago, but evidence

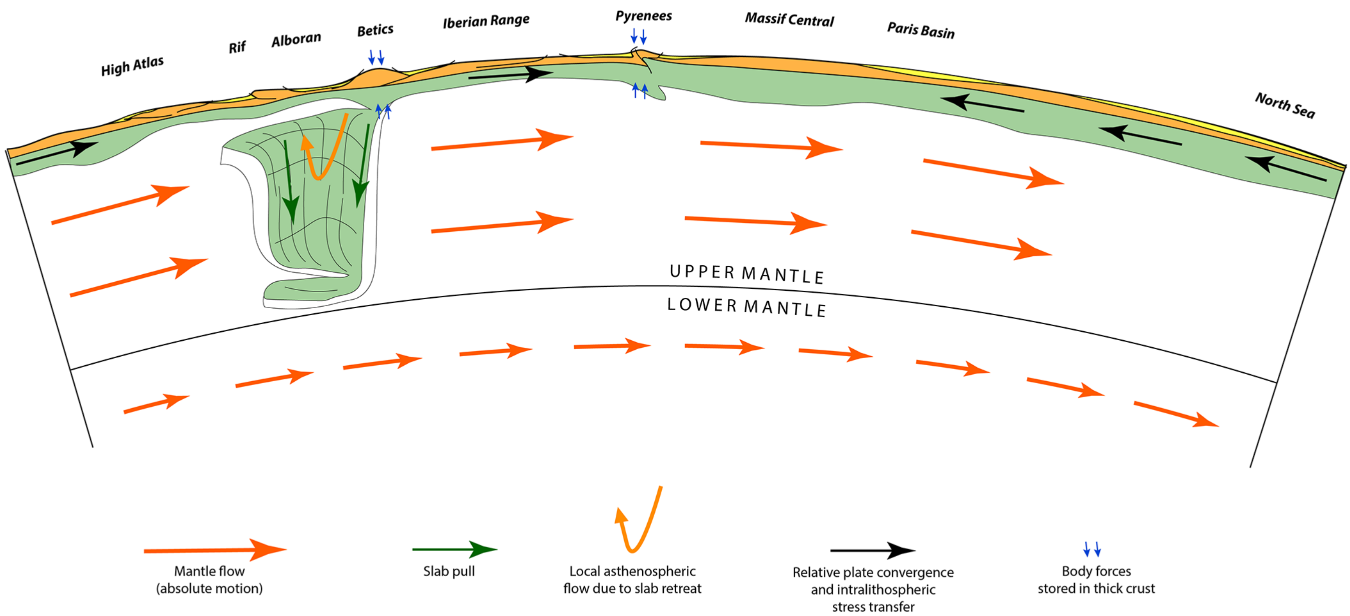

Fig. 23. Schematic section at lithospheric scale (topography not to scale) from the High Atlas to the North Sea showing the various forces driving the kinematics and deformation.

for compression and obduction started earlier around 100 Ma, 84 Ma being the culmination of shortening (Guiraud *et al.*, 2005; Jolivet *et al.*, 2016a). An alternative solution is that this stress regime change was partly due to the interaction between the Tethys oceanic slab and the upper-lower mantle transition zone, which in turn induced the kinematic change. The nature of the subducting lithospheres and especially the behavior of the thick lithosphere of cratonic areas is also an important point to consider (Mouthereau *et al.*, 2021). In this framework, the initiation of compression in the Pyrénées would be a part of a larger-scale process involving the whole mantle, and the kinematic change a consequence of these deep interactions. The progressive localization of strain along the Pyrenean thrusts has modified the distribution of stress and strain through time and space (Dielforder *et al.*, 2019) but would then be a second order feature, at this large scale.

After 84 Ma (Tethyan III), compression in the Western Mediterranean lasts until 35 Ma. In the Eocene, compression is recorded from the High Atlas in the south to the North Sea. The whole system from North Africa to the North Sea is coupled. The Pyrénées and the Betic-Rif orogens form in this context of generalized shortening over a large distance, coeval with the Dinarides and the Hellenides further east. In the eastern Mediterranean and Middle East, the Tethyan subduction creates back-arc basins in the overriding plate, suggesting a different subduction behavior. The observed compression in the west certainly owes much to the compressional stresses transmitted through the lithospheric stress guide between Africa and Eurasia. But, given the scale of the domain recording this compression, a larger-scale cause (mantle) is likely.

The Priabonian is a transitional period with continuing shortening in the Pyrénées and some limited extension in the West European Rift System. It represents also the last record of UH*P* metamorphic rocks and the beginning of fast exhumation of all HP-LT metamorphic units, from the Alps to the Betics, suggesting a change of subduction dynamics. Whether the rifting in the West European Rift System is related to this change in subduction dynamics or a more local process is an open question. Merle and Michon (2001) proposed that this episode of rifting could result from the downward pull of the Alpine lithospheric slab inducing a counter flow of the asthenosphere and extensional stresses in the subducting lithosphere. In subsequent periods (Mediterranean Tectonics), compression is recorded only along the front of mountain belts forming above retreating slabs, including the Alps. The rest of the Western Mediterranean is under extension, powered by the retreat of the slab. Shortening progressively stops in the Pyrénées and the eastern part of the belt is dismantled and replaced by the Gulf of Lion passive margin (Jolivet *et al.*, 2020). An alternative interpretation is to consider that the compressional deformation has migrated southward to concentrate in the Betics and the Rif with a progressive reduction of shortening in the Pyrénées (Daudet *et al.*, 2020).

The slab first retreats southward and then westward and eastward after a major slab tear consecutive to the docking of the AlKaPeCa with the North Africa Margin some 20 Ma ago (Leprêtre *et al.*, 2018; Romagny *et al.*, 2020). The counter clockwise rotation of the Corsica-Sardinia block and the progressive subduction of Adria in a retreating trench are coeval with the formation of the Liguro-Provençal and Tyrrhenian basins. The retreat of the Apennines slab induces a counterclockwise toroidal flow under the slab and this flow in turn pushes the Alpine slab westward, inducing the propagation of the thrust front and the exhumation of the External Crystalline Massifs, as well as compressional stresses in the Alpine foreland (Vignaroli *et al.*, 2008). This situation lasts until the Late Miocene, around 8 Ma, when compression resumes and Africa and Eurasia are coupled again in the Western Mediterranean. Extension still prevails until very recently only where back-arc extension is still active, namely the South Tyrrhenian Basin as well as the Apennines.

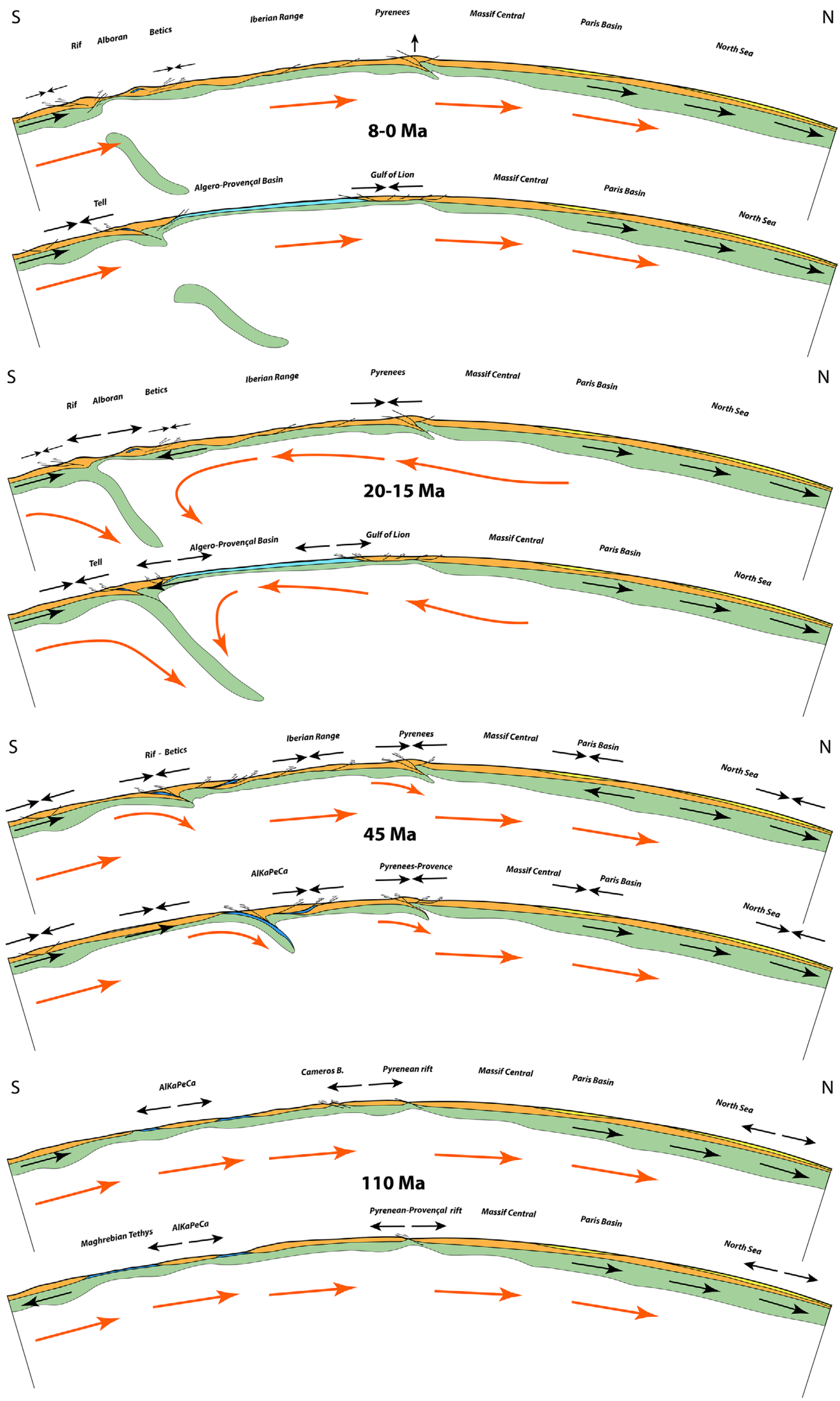

Fig. 24. Evolution of two cross-sections (locations on Fig. 22) from 110 Ma to 8 Ma showing the succession of stages with the driving forces.

After the Late Miocene and until Present (Late Mediterranean Tectonics), the stress regime in the Western Mediterranean returns to a situation quite similar to what it was in the Eocene prior to the initiation of slab retreat with N-S compression everywhere from the Atlas to the North Sea. This suggests that the slab retreat and related asthenospheric flow had decoupled Africa and Eurasia in this region as the flow of mantle was much faster than the Africa-Eurasia convergence and controlled the deformation in the back-arc region. It is only once the rate of retreat had significantly decreased in the west that N-S compression was felt again (Fig. 25). Dragging of the slab by the absolute motion of Africa and Eurasia (Spakman *et al.*, 2018) adds its effects to control the observed stress and strain field in this region. The situation in the Pyrénées is partly unclear at this period. Geomorphological markers and thermochronology show evidence for uplift after 10 Ma, which causes are not obvious. It does not seem to be associated with a clear renewal of compression. A change in the mantle has been proposed by Gunnell *et al.* (2009) and Calvet *et al.* (2021), but this period should be studied in more details to unravel this important question.

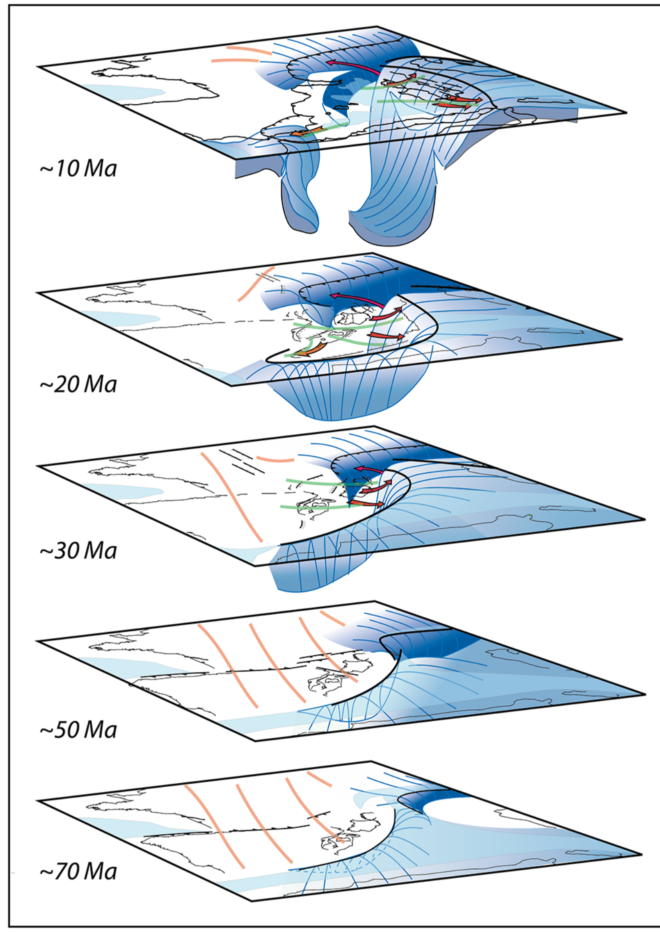

Fig. 25. Synthesis of the evolution of the Western Mediterranean slabs before, during and after slab retreat and the directions of shortening and extension. Modified from Vignaroli *et al.* (2008, 2009).

8 Conclusion

This review shows the evolution of the geodynamic framework of the Western Mediterranean since the Early Cretaceous and the respective contributions of local (near-field) and distant (far-field) forces. We distinguish three periods that reflect changes in the source of driving forces: (i) *Tethyan Tectonics* (150–35 Ma) when all deformations are driven by the relative motion of large plates and the large-scale convection underneath, from rifting to convergence. This period is further divided in *Tethyan I* (first rifting), *Tethyan II* (second rifting) and *Tethyan III* (convergence and shortening). (ii) *Mediterranean Tectonics* (35–8 Ma) when all deformations are under the control of slab retreat and (iii) *Late-Mediterranean Tectonics* (8–0 Ma) when slab retreat progressively stops and the situation return to the Tethyan stage with stress transmission across the whole plate boundary zone, from Africa to Europe, and large-scale mantle convection is again the main driver.

During the pre-35 Ma (*Tethyan Tectonics*) period the Western Mediterranean orogens are controlled by the large-scale relative motions of Africa, Iberia and Eurasia and the mantle flowing underneath (large-scale convection). Early Cretaceous extension is distributed between Africa and the North Sea with a focalization within a wide plate boundary encompassing the future Rif-Betics arc, the future Iberian ranges and the Pyrenean rift basins. It is a consequence of the opening of the Atlantic Ocean and Bay of Biscay and of the long-term evolution of the Tethys Ocean involving the northward flow of mantle due to large-scale convection. From the Late Cretaceous, the relative motion of African and Eurasia becomes convergent and compression ensues. It is recorded from the reactivated early Cretaceous African rift basins all the way to the North Sea. This phase of compression culminates in Africa at 84 Ma and it leads to the obduction of ophiolite nappes on the northern margins of Apulia and Africa. The scale of this compressional domain and the specific characteristics of large-scale ophiolite obduction suggests that whole mantle is involved in this change of regime and models involving modifications of the convection pattern and slab behavior have been proposed. In this large-scale framework, deformation progressively localizes in the Pyrénées that then act as a new convergent plate boundary. This situation lasts until the Late Eocene with shortening distributed from the Atlas, the Alboran region, the Pyrénées and further north until the North Sea.

Between 35 and 8 Ma (*Mediterranean Tectonics*), the situation changes drastically when the western Mediterranean slabs start retreating. The velocity of retreat is faster than the velocity of relative motion between Africa and Eurasia and faster than the absolute velocities of both plates. The flow of mantle resulting from slab retreat thus dominates the deformation in the Western Mediterranean. It proceeds in two main steps, the first one between 35 and 20 or 15 Ma, and the second step after 15 Ma. During the first stage, one main slab retreats southeastwards and the Liguro-Provençal Basin opens, and during the second stage, after a main slab tearing event, two slab portions retreat in opposite directions, eastward for the Calabrian Arc and westward for the Gibraltar Arc. During these periods, the mountain belts formed earlier during the Late Cretaceous and Eocene are partly dismantled. The eastern part of the Pyrénées is replaced by the Gulf of Lion rifted margin and asthenospheric flow due to slab retreat induces depth-dependent extension with the extraction of the lower crust and mantle from below the belt. In the Betic-Rif orogen, the internal zones (Alboran domain) are extended and metamorphic core complexes form, first with a N-S direction of extension, then E-W after 20 Ma. The Eocene nappe stack of Alpine Corsica is also reworked by extension during the rotation of the Corsica-Sardinia block and extension then migrates eastward until its present situation in the Apennines, following the retreat of the Apennines slab. At about 35 Ma, the maximum burial is recorded in HP-LT metamorphic terranes from Corsica to the Betics and the peak pressure then decreases (Tuscan archipelago) during the migration of orogenic wedges following slab retreat. This change of P-T conditions results from the change of subduction regime from compressional to extensional. The interactions of slabs at depth have consequences on the tectonic evolution of the Western Alps where the thrust front migrates westward.

From 8 Ma onward (*Late-Mediterranean Tectonics*) the stress regime progressively changes again back to the pre-35 Ma situation and the progressive recoupling between the African and Eurasian lithospheres with compression observed from the Atlas to France. This evolution coincides with the end or slowing down of slab retreat on both the Calabrian and Gibraltar subduction zones. The dragging of the slab by the mantle flow driving the absolute motion of Africa and Eurasia adds its effects on the specific deformation of the Gibraltar Arc.

During this long evolution from the Early Cretaceous to the Present, deformation in the Western Mediterranean region has been mainly controlled by the Africa-Eurasia convergence and the flow of mantle due to large-scale convection, except between 35 and 8 Ma when slab retreat and associated mantle flow were dominant drivers and partly prevented the compressional stresses due to convergence to be transmitted from Africa to Eurasia. The observation of a compressional structures in the Central and Western Pyrénées, as well as in the Corbières until the Early Miocene (Bosch *et al.*, 2016; Labaume *et al.*, 2016; Teixell *et al.*, 2018; Parizot *et al.*, 2021), shows that some stress transmission was locally possible within the lithosphere.

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